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USSR Contribution

Permafrost

Second International Conference

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NATIONAL ACADEMY OF SCIENCES

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Permafrost

Second International Conference

July 13-28, 1973

Edited by FREDERICK J. SANGER
with the assistance of PETER J. HYDE

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Preface

At the conclusion of the Second International Conference on Permafrost, the United States Planning Committee and the Canadian Organizing Committee agreed that the Soviet contribution to the conference should be made available in English to the scientists and engineers of the permafrost research community in order to further the international exchange of information among the relevant disciplines. Following the conference, the U.S. Planning Committee, which was administered by the Building Research Advisory Board (BRAB) of the National Research Council, and its Canadian counterpart initiated the effort to raise the funds needed for publication. This effort ultimately proved successful, and the committees wish to thank the sponsors of this publication:

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In addition, the committees are particularly indebted to Frederick J. Sanger, the Honorary Technical Advisor of the U.S. Planning Committee, who served as principal technical editor and without whose dedication and experience publication of this volume would not have been possible, and to Peter J. Hyde, who assisted Mr. Sanger with part VIII. Finally, the committees wish to thank Robert M. Dillon and Claret M. Heider of the BRAB staff

for their assistance in guiding the financial and technical aspects of publication.

The committees believe that this document and its companion volume--*Permafrost, North American Contribution, Second International Conference on Permafrost* (Washington, D.C.: National Academy of Sciences, 1973)--represent a clear statement of the status of permafrost research in 1973. The committees hope that this volume will serve to assist those scientists and engineers who participate in the Third International Conference, to be held in Edmonton, Canada in July 1978, in further advancing the state of permafrost research and in implementing the Resolution of the Second International Conference, which is presented below.

Troy L. Péwé, *Chairman*
United States Planning Committee

J. R. Mackay, *Chairman*
Organizing Committee of Canada

Resolution of the Second International Conference on Permafrost

The Second International Conference on Permafrost notes with satisfaction that representatives from 14 countries have actively participated in its work. These countries are Belgium, Canada, Czechoslovakia, Denmark, France, Hungary, Japan, Mongolia, Norway, Poland, Sweden, the United Kingdom, the United States, and USSR. In the course of preparing for the conference, more than 300 papers totaling 204 printed sheets were published in the USSR, that is, six times more than at the First International Conference.

At the conference a general report was presented in which the progress in geocryology in the last 10 yr is summarized, the present status of the science is reviewed, and future research is planned. Generalizing reports were presented and discussions were held in the following main areas of geocryology:

the operative principles, the thermophysical principles governing the genesis and evolution of the permafrost region;

regional permafrost studies;

the genesis, composition, and structure of permafrost and ground ice;

the physical chemistry and mechanics of frozen earth materials and ice;

groundwater in the permafrost region;

geocryological surveying and forecasting;

the principles of controlling cryogenous processes during the development of areas containing permafrost.

It is noted that during the 10 yr that have passed since the First International Conference on Permafrost was held in the United States, essentially new results have been obtained in all of these subject areas. In the USSR, the United States, Canada, and other countries, more than 100 major works have been published along with a large number of scientific articles on various aspects of geocryology. They reflect the rapid growth of the science of permafrost, which is a consequence of the development of the northern and the alpine regions of northern Eurasia and America. The published results of the studies are being widely applied in the economic development of areas located in the permafrost region of the United States, the USSR, and Canada. Studies of the ancient periglacial zone of various countries have been further developed.

The conference considers that the most urgent problems for which solutions are needed for the further development of geocryology are the following:

Effecting improvements in research methods in order to increase their efficiency by automating the experimental work, using computer techniques, and introducing refinements in scientific instrument design;

obtaining a better understanding of improving the general theoretical principles in underlying the genesis and evolution of the permafrost region;

compiling summary geocryological maps of all of the permafrost areas and the production of a geocryological atlas;

carrying out more detailed studies of the genesis, composition, and structure of the permafrost in various regions;

further elaboration of the scientific principles and practical methods of controlling the physical, chemical, and mechanical properties of frozen, freezing, and thawing ground;

conducting further studies of the problems of the recharge and distribution of groundwater in the various permafrost regions and developing methods of making practical use of this groundwater;

improving geocryological survey procedures and forecasting changes in geocryological conditions through the use of geophysical techniques;

improving upon existing methods and devising new ones for the control of cryogenous processes when developing an area containing permafrost;

extending our knowledge of the mechanics and thermophysics of frozen, freezing, and thawing ground and effecting improvements in methods of building cities and settlements, roads, power supply systems and mining installations, pipelines, communications lines, and other structures in permafrost regions;

formulating the theoretical principles of the laws governing the existence and evolution of the Earth's cryosphere;

studying the interrelation between the cryosphere and the biosphere;

elaborating the scientific principles of environmental protection and the efficient utilization of natural resources in permafrost regions.

It is considered expedient to conduct integrated international research pertaining to environmental protection in the permafrost regions. This includes the exchanging of scientific information, reciprocal visits by specialists, the development of effective procedures for preventing disturbances of the environment, and effecting improvements in the methods of using natural resources.

The conference recommends to the delegations from the USSR, the United States, and Canada that combined working groups be set up that will engage in work aimed at coordinating research and facilitate rapid exchange of information on scientific and practical achievements in the various areas of geocryology between the interested institutions and scientists of the different countries.

The conference considers it expedient to hold the Third International Conference on Permafrost in Canada.

The conference highly values the international role of the *Biuletyn Peryglacjalny* [*Periglacial Bulletin*] (published in Lodz, Poland) in the development of geocryological research and considers that the future activity of this printing agency will be of great importance in the development of international relations, reciprocal exchange of information, and joint research in this field.

The conference expresses its appreciation to the government of the Yakut ASSR; the presidiums of the academies of sciences and the Siberian Division of the USSR Academy of Sciences; the Institute of Permafrost Studies of the Siberian Division of the USSR Academy of Sciences; the Soviet, American, and Canadian Organizing Committees; the Scientific Council on the Cryology of the Earth; and all of the organizations that took part in the convocation and holding of the Second International Conference on Permafrost.

Leader of the Soviet Delegation
Professor P. I. Mel'nikov

Leader of the United States Delegation
Professor T. L. Péwé

Leader of the Canadian Delegation
Professor J. Ross Mackay

Editor's Note on Russian Terms for Soils

Two important terms used in this volume have no English equivalents, so they are explained here. Soil types and their properties are given in the table below.

Supes (*SOO pess*); pl., -esses. This soil is defined in the table and by the figure showing gradation, water content, liquid limit, and plasticity index. Supes is a silt having some clay and sand in it, and more sand than clay.

Suglinok (*Soo GLEE nok*). This is also a silt having more clay than sand with it. Refer to the table and figure for its properties.

Supes obviously comes from "pesok" (sand) and suglinok from "glina" (clay). This may help as a mnemonic.

The void ratio of these two soils is usually between 1 and 1½.

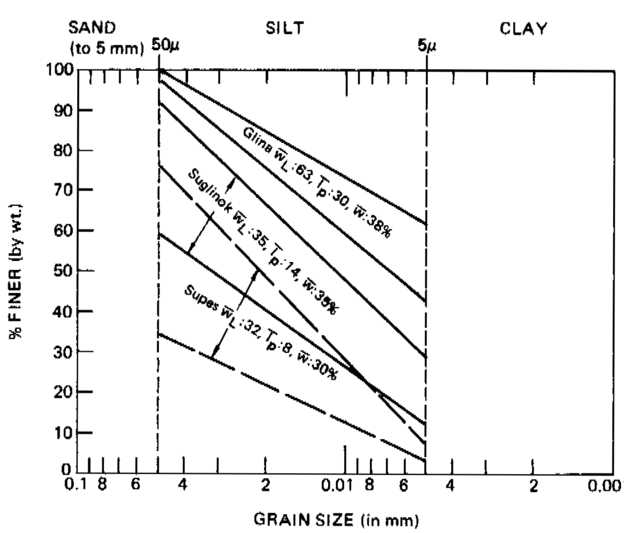
SOIL TYPES

Name	Clay Content % by Wt.	Minimum Diam. of Roll in Plastic Limit Procedure	Liquid Limit	Plasticity Index
Glina (clay)	>30	<1	53-68	23-24 and up
Suglinok	30 to 10	1 to 3	32-38	12-16
Supes	10 to 3	>3	22-38	1-14
Pesok (sand)	<3	N. P.	--	--

It will be remembered that 3 mm is the size for the plastic limit test.

SOIL SIZES

Name	Diam. in mm
Sand	>0.05
Silt	0.05 to 0.005
Clay	<0.005



Silty. In Russian usage, a soil is silty if the proportion of silt-sized particles exceeds that of the sand-sized ones.

Muddy. This word describes a soft highly sensitive (or perhaps thixotropic) soil that has low stability when undisturbed but has negligible strength when remolded.

Frederick J. Sanger
Technical Editor

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Thermophysical Principles of the Formation and Development of the Cryolithosphere

G. V. PERKHAEV and
V. T. BALOBAEV, *Editors*

PROBLEMS OF CRYOLOGY

I. YA. BARANOV   PNIIIS (Industrial and Scientific Research Institute for Engineering Surveys in Construction), Gosstroi USSR

Cryology deals with the cryogenous process, as well as with phenomena and formations that are derived from it. The basis of the cryogenous process is crystallization of common substances capable of changing their aggregate state within strictly defined heat-energy limits and transition from gaseous or liquid states to a solid state and vice versa. The cryogenous process is characteristic not only of our planet but also of other planets in the solar system and other cosmic bodies (from comets and asteroids to accumulations of cosmic dust), where it develops on the basis of one or several substances.

Investigation of the natural cryogenous process and associated phenomena is the doman of cryology. The branch of cryology that deals with the Earth is termed "geocryology." "Planetary cryology" deals with the planets, while "cosmocryology" or "astrocryology" deals with the universe as a whole. Cosmocryology is a branch of astrophysics, which has gained in importance in recent years. Most attention will, of course, be devoted to the study of the cryogenous process on the Earth, since it is of greatest practical importance.

What follows is a brief account of some ideas concerning cryology prepared by the author for the Fifth Meeting of Planetologists in Moscow in 1965. Some of this material was previously published in the *Astronomical Journal*.[1]

THE EARTH

The basis of the cryogenous process on the Earth is crystallization of water under natural conditions. The three-dimensional region of development of the cryogenous process includes parts of the lithosphere, the hydrosphere, and the atmosphere, which interact with each other through material and energy transfer.

Investigation of the cryogenous process and cryogenous and associated phenomena and formations in the lithosphere, on the Earth's surface, in water bodies, and in the atmosphere is the task of permafrost studies, glaciology, hydrology, and atmospheric geophysics (meteorology). Coordinated investigations involving these sciences hold great promise and will improve man's knowledge of the cryogenous process and associated phenomena.

The specific quality of the cryogenous process on the Earth lies in the crystallization of water, a widely distributed substance that penetrates other substances in the surface layers of the Earth and is capable of simultaneous existence in three phases within strictly defined thermal-energetic limits. Under conditions prevailing on other planets and cosmic bodies, substances other than water (e.g., carbon dioxide, methane, ammonia, and hydrogen) can form the material basis of the cryogenous phenomena. Their presence on a planet in a free or a physically bound state depends on such things as the stage and conditions of development of the planet, the cosmic and the planetary parameters, and the distance from the Sun.

With respect to the single process of an aggregate transformation of substances under natural conditions, the cryogenous process represents a special case. This is due to its association with a specific substance--water in liquid and gaseous phases that is capable of transition to a third, solid phase. The latter can undergo transition again to a solid phase via vapor, thus avoiding the liquid phase. The amounts of energy required for the phase transitions water-ice, vapor-ice, and ice-vapor-ice depend on the purity and the freedom of the phases.

The cryogenous process on the Earth is impossible outside the relationship with the initial water phases (gaseous, liquid, and solid) and phase transitions. Being a chemically-active reagent and a substance that penetrates and dissolves solid, plastic, and friable substances and gases, water can partially or completely lose its degree of freedom (i.e., its mobility). Its crystallization point changes in relation to this, and the crystallization process is either limited or is completely stopped (strong solutions).

Dobrovol'skii[2] suggested that within the Earth's system there is a particular region of cooling within which water (or a part of its volume) is in a crystalline state or is capable of acquiring this state. He called this region the cryosphere of the Earth.

V. I. Vernadskii[3] wrote: "The entire region of life, the biosphere, lies within the region of cooling but the latter extends to greater depths into the upper layers of the stratisphere (the region of sedimentary rocks) which forms part of the continents and underlies the adjacent seas. Below the hydrosphere (i.e., the oceans), the region of cooling extends to still deeper layers of the Earth's crust, which under the continents lie at average depths of more than 10 km from the surface. In the hydrosphere, granitic and basic rocks which underlie the oceans are subjected to cooling" (Vol. IV, Book 2, pp. 637-638). At the same time, Vernadskii related the region of cooling to a special form of manifestation of two chemical bodies that play on our planet "an exceptionally important role, i.e., water and carbon dioxide. This is that region of the planet, where three water phases are capable of coexisting: gaseous phase, liquid phase and solid phase, i.e., ice I with a M.P. = 0°C at a pressure of 1 atm. In the same region, two CO_2 phases are present simultaneously, liquid and gas, and the temperature may drop to, or may always be below, the critical temperature of CO_2 (32°C). This region embraces the entire so-called cryosphere proposed by Dobrovol'skii and extends far beyond it" (p. 637).

The energy- and mass-exchange processes that lead to the phase transitions water-ice and ice-ice (via sublimation), and the crystallization itself, form the physical essence of the cryogenous process. The three-dimensional region where crystallization of water is taking place is the cryosphere of the Earth. The cryosphere embraces the parts of the surface layers of the Earth that are in contact with each other and are interrelated by energy and mass exchanges within the framework of the single external thermodynamic system of the Earth.

The large variety of cryogenous and associated phenomena is related to many forms of existence of water in different media, to the degree of freedom of its movement, and to the possibility of undergoing crystallization or mineralization.

Water-bulk Crystallization occurs while water has complete freedom of movement. Under appropriate conditions it is accompanied by a precipitation of salts or an increase in the concentration of solutions (hydrosphere--ice caps and surface water pore-water and ice; lithosphere--ground ice from streams and in various cavities, including those of cryogenous origin).

Precipitated liquid water (meteoric) Crystallization occurs in the atmosphere and is accompanied by formation of suspended crystals, which obey the force of gravity, and by subsequent formation of cryogenous structures on the surface, spreads and freezes on cooled surfaces of various bodies.

Percolating liquid water (pore and capillary-pore water) Crystallization occurs when water movement is restricted by surface tension and the force fields of particles. Water participates in the cryogenous activity.

Film water (loosely or firmly bound) Crystallization occurs in the containing medium (ice, soil, etc.), with or without displacement and involves film- or film-capillary mechanisms of movement and formation of primary ground ice, including ice outside the containing medium ("ice needles," "tetragons," etc.).

The Earth's cryosphere borders on the *region of cryogenous cooling,* which is characterized by temperature below 0°C but where no crystallization of water occurs under the existing natural conditions, either because there is no water or only very small amounts of it (stratisphere, anhydrous rocks) or because of its high mineralization. This region is very extensive both within the continents and the shelf areas of the Northern seas.

A large region of cooling is comprised of the cold water of the oceans and the subpolar regions with temperatures ranging from -1.9° to 0°C. The study of the Earth's cryosphere, the regions of cooling, all classes and types of cryogenous phenomena and formations in the geospheres and on the surface of the continents is a very complex task, which involves investigations of physical phenomena, determinations of interactions of processes, phenomena, and formations, and their effects on the environment, as well as the effects of the environment on them.

The structure of the Earth's cryosphere reflects the thermodynamic conditions prevailing in its near-surface zone. These conditions are determined by a number of factors:

1. Spherical shape of the Earth, the inclination of its axis, and the distance from the Sun (the solar constant),
2. Planetary, geological-tectonic, and geochemical history of development of the Earth's mantles; mode of formation and stability of substances forming the mantles,
3. Three-dimensional distribution of the Earth's mantles: surrounding (atmosphere), forming (lithosphere), and superimposed (hydrosphere),
4. Composition of substances forming the outer mantles of the Earth, their properties, and so forth.

The cryosphere of the Earth consists of three complex components: *cryoatmosphere, cryohydrosphere,* and *cryolithosphere.*

The *cryoatmosphere* consists of three morphostructural zones: upper-mesosphere, middle-stratosphere, and lower-troposphere.[4]

The *mesosphere* is located at altitudes of 50-55 km to 85-90 km. The temperature at its base is 0°C to +10°C and at its upper boundary -90°C to -100°C. The pressure ranges from 10^{-2} mm Hg to 10^{-9} mm Hg. The main cryogenous formation is a very fine ice suspension that forms fine silvery clouds at altitudes of 67-97 km between the latitudes of 45° to 80°. The mesosphere has little water.

The stratosphere is located at altitudes of 17 km to 18 km (equatorial region) and 6 to 8 km (near the poles) to 50 to 55 km. The temperature varies with altitude and ranges from -70°C to 0°C or +10°C (rapid increase above 25 km). The pressure at the base of the zone ranges from 200 mm Hg to 250 mm Hg. Nacreous clouds consisting of an ice suspension occur at altitudes of 22 km to 27 km. The stratosphere has little water also.

The troposphere (up to altitudes of 6 km to 18 km) contains up to 80 percent of the mass of the atmosphere and almost all its water. The water content is highest in the lower part of the troposphere. The 3.5-km layer contains 70 percent water and the 5-km layer 90 percent.[5] The temperature at the base of the troposphere varies within a wide range and decreases with altitude to -70°C. Permanent ice formations of the ice suspension type are present in the upper part of the troposphere at temperatures of -30°C to -70°C. The lower part is the site of dynamic phase transitions, crystallization of liquid (vaporlike) water, ice sublimation, formation of various crystallocondensates in air and on cooling surfaces and accumulative structures on the surface of the Earth. The most important process is cryocondensation.

In actual fact the mesosphere, the stratosphere, and the upper part of the troposphere form a single cryogenous screen that retains water in the atmosphere and prevents it from escaping into space after dissociation in the thermosphere. The most important place among the cryogenous formations related to the atmosphere is occupied by snow.

The region of the cryohydrosphere includes oceans, seas, lakes, and streams formed by free water with different degrees of mineralization. The main types of seasonal and perennial cryogenous formations are surface and subsurface ice. The main process is crystallization of free water.

The circumpolar oceanic regions contain the cold water zone with temperatures as low as -1.9°C, whose thickness reaches several kilometers. The boundary of the cryohydrosphere depends on the conditions of the heat exchange in the troposphere, the sea currents, and other factors. This zone includes the regions of ice sheets, and the latter in turn include the permafrost regions.

The region of the *cryolithosphere* is formed by elements whose development depends in a complex way on the heat exchange between the land masses and their environment. The main types of cryogenous formations are seasonally frozen soils and permafrost, whose maximum thickness reaches 900 m and more. In many regions the frozen soils rest on soils cooled to below 0°C and saturated with mineralized water. The zone of cooled soils reaches 1,300 km to 1,500 m.

In arctic regions and alpine countries, where moisture-saturated air masses circulate toward the Earth's surface, there are glaciers and ice caps, which constitute a special region of development of formations mainly of an accumulative nature.

PLANETS OF TERRESTRIAL TYPE

Mercury

The temperature of the sunny side of the planet varies between 285°C and 437°C and is 330°C on the average, while at night the temperature drops to -120°C. Owing to a low force of gravity (0.36 of the terrestrial force) and a low parabolic velocity (IPI cosmic velocity is equal to 4.2 km/s), Mercury was unable to retain not only volatile, but even certain heavy, gases. Its atmosphere consists mainly of nitrogen (~90 percent) and CO_2 (~10 percent). It is not thick and is very rarefied (1 mm Hg). No water molecules are present since they decompose by solar radiation. Free, as well as physically and chemically bound, water may be present in the interior of the planet. Some volumes of water in the form of vapor may escape into the atmosphere and form temporary crystallocondensates on the dark side.

There may be a polarized cryogenous shadow screen in the atmosphere. Formation of water crystals is evidently impossible. Carbonic acids with or without water may participate in the formation of the cryosphere.

Venus

Venus occupies a special place among the planets of the terrestrial type and differs from Mercury and the Earth by the fact that it is in the pregeological* stage of development.[6]

On its surface the acceleration due to gravity is 0.87 and the parabolic velocity is 0.92 of those values for the Earth. This points to a relatively limited degassing of the planet. The specific characteristic of Venus is the absence of water on its surface, although the presence of free water, as well as vapor and physically bound and chemically bound water, in its lithosphere is beyond doubt. To overcome the atmospheric pressure, water must have a pressure of at least 90 atm to 100 atm.

The surface temperature on the sunny side near the center of the disk reaches 475°C ± 20°C (Venera 7). The temperatures vary with altitude: -75°C at 100 km, -95°C at 150 km, 60°-70°C at 200 km, and 350°C at 250 km.

As is the case on the Earth, the south pole of the planet is colder than the north pole. The atmosphere consists mainly of CO_2 (93-97 percent) and small amounts of inert gases and oxygen. At altitudes of 100 km to 110 km (pressure of 0.6 atm) the water vapor content is 4 to 11 mg/liter and does not reach the saturation value (which may be possible in the higher atmospheric layers).

By analogy with the Earth, the atmosphere of Venus probably contains a cryogenous screen and crystallocondensates of water and CO_2. Judging by a sharply defined outer boundary of the cloud layer, this screen plays the same role as under conditions prevailing on the Earth. The cryoatmosphere is located between the lower overheated air mass and the upper zone of dissociation of water. The presence of a vaporizing crystallic suspension can be expected above the zone of crystallocondensates (up to ~150 km). The limited amount of water in the atmosphere of Venus depends (according to Rosul) on the equilibrium between the supply of water from the interior of the planet and its dissociation in the thermosphere.

As far as the night side of the planet is concerned, the crystallocondensates of water are possible in the circumpolar regions. This is based on the fact that there are polar "caps" of unknown composition, which can be seen with the help of light filters.

*As referred to the Earth.

The Moon

The Moon is vastly different from the Earth, as far as formation of cryogenous structures is concerned, and in this respect it resembles the smaller planets. The acceleration due to gravity on its surface is 0.17, and the parabolic velocity 0.21 (2.38 km/s), of the Earth's values.

The Moon retains large reserves of heat in its interior, and it is the site of tectonic activity. Manifestations of activity at depth, i.e., gas emissions, occur during the periods of maximum and minimum tidal waves in the lunar lithosphere at many points, including large craters. This points to a degassing of the lunar interior by decomposition of gases and their escape into space.

There is hardly any doubt that the interior of the Moon and its subsurface layers contain physically bound and chemically bound water that can be released to the surface together with gases. In the center of the lunar disk the temperature of the surface soil reaches +110°C, while on the night side in the premorning hours it drops to -180°C. Considering that the pressure on the lunar surface is negligible, it is doubtful whether large-scale formation of permafrost is possible on the illuminated surfaces in the equatorial zone. At the existing atmospheric pressure water boils at 0°C. Frozen soils predominate in areas that lie in shadow for long periods of time. Great thicknesses of frozen soils can be formed in the circumpolar regions of the Moon.

Temporary cryogenous structures, mostly frozen (cooled) soils, may be formed on the night side of the Moon. Formation of cryocondensates is possible near the points of emergence of gases that probably contain water vapor. In the regions of deep and stable cooling (the circumpolar regions and shaded areas), the morphostructural types of permafrost may be as follows: upper horizon--temporarily frozen soils; lower horizon--temporary and perennially frozen soils, which serve as a cryogenous screen for the plutonic water.

Therefore, the lunar cryosphere probably consists of three elements: outer, shaded (which includes the atmosphere and part of the lithosphere) and perennial (medium and high latitudes of the lunar lithosphere). Here we may expect to find a layer of daily freezing (cooling) and thawing (warming).

Mars

The mass of the planet is small (0.12 of that of the Earth). The acceleration due to gravity is 0.38, and the parabolic velocity is 0.45 (5 km/s), of those values for the Earth. The atmosphere is therefore rarefied (the pressure on the surface is 9 mbar). If we consider that Mars receives 2.25 times less solar radiation than the Earth, the overall picture becomes sufficiently clear. Considerably lower temperatures and abrupt daily and seasonal temperature variations are characteristic of the planet.[7]

The highest temperatures occur during the day in the equatorial belt (at noon they range from 0°C to +25°C). At night they drop to -40°C or -50°C. On the shaded (night) side the temperature drops to -120°C. We may expect to find the coldest regions in winter near the poles, on mountains that reach altitudes of up to 12 km and on shaded elements of relief. The general situation resembles that on the Moon but is less extreme. Development of perennial cryogenous structures may be expected throughout the lithospehre. Permafrost may be hundreds of meters thick, even on the equator, while on the mountains in temperate and polar regions its thickness may reach several kilometers.[3]

Development of Mars was similar to that of small planets, i.e., it was accompanied by strong degassing of matter. The pressure on its surface (Mariner 4, 6, and 7) does not exceed 9 mbar, which corresponds to the pressure of the Earth's atmosphere at an altitude of 30 km. The cryosphere of Mars is probably as follows: The atmosphere contains a constant cold screen in the form of a sphere, which is thicker in the circumpolar regions and which to some extent prevents the escape of water and CO_2. The atmosphere consists of CO_2 (90 percent) and contains water vapor that forms sparse clouds of ice crystals.

The cryoatmosphere is evidently stable and is subjected to destruction only in the equatorial belt and near the surface in temperate latitudes in the summer.

Evidently the regions of seasonal polar cryocondensation of water should be related to high humidity in the cold regions. The cryogenous crystallocondensates develop because of sublimation. Since the seasonal layer of these formations is thin, the horizontal rate of its destruction in the spring and summer reaches 100 km per day (along the meridian) and occurs at temperatures below zero. The polar "caps" of crystallocondensates sublimate at a slower rate owing to their greater thickness.

Judging by the planetary conditions, water molecules should dissociate at the upper boundary of the cryoatmosphere, and the products of dissociation (H^+, OH^-) should evaporate. This process occurs at a relatively slower rate than on other planets.[8]

According to calculations performed by American scientists, the total water content in the atmosphere of Mars is 1.6 km^3, while in the Earth's atmosphere alone the water vapor content is 12,300 km^3. The moisture content of the atmosphere is controlled by the emergence of water on the surface and by the mechanism of its retention in the atmosphere. These two processes must be in equilibrium. The cryogenous screen in the atmosphere retains water, while the equivalent screen in the lithosphere promotes the emergence of water on the surface. A freer emergence of water in the form of vapor is possible only in some areas of intense postvolcanic and tectonic activity. Crystallization of vapor followed by sublimation and evaporation are possible on the shaded side of Mars.

The zone of cooling of the lithosphere must consist of frozen (loose eolian and residual sediments and bedrock) and frost-fractured (cemented with water and possibly CO_2 ice) soils. Permafrost can be found in eolian beds in depressions of various kinds saturated with groundwater, or condensation water.

Water with various mineral contents may be found in fractures below the permafrost. Thawing (warming) of soils during the day extends to a depth of several centimeters. This process is similar to that in high mountainous regions on the Earth.

Hence, it may be assumed that cryogenous crysstallocondensates based on H_2O and CO_2 are present in the atmosphere of Mars. Furthermore, there may be daily, seasonal, and perennial cryogenous formations on the surface, as well as daily, seasonal, and mainly perennial cryogenous formations in the lithosphere throughout the planet (frozen and frost-fractured soils and permafrost).

GIANT PLANETS

With respect to their planetary parameters and geochemical development, these planets (Jupiter, Saturn, Uranus, and Neptune) differ greatly from the planets of the Earth type. All of them are far removed from the Sun and hence have a limited supply of solar energy.

Jupiter

The mass of the planet is 316.7, the density 0.24, the acceleration due to gravity 2.51, and the parabolic velocity 5.4 (61 km/s), of the Earth values. The planet has a well-developed atmosphere. The outer cloud layer consists of crystalline ammonia; the lower layer presumably consists of ice crystals. Continuous cloud covers are located at altitudes of about 100 km to 150 km.

The atmospheric temperature above the upper cloud layer is about -168°C. Near the surface of the clouds it is about -123°C.

It is almost certain that there is a cryoatmosphere surrounding the planet. According to V. G. Fesenkov, one of the large satellites of Jupiter (Ganymede) is covered with crystalline CO_2.

Saturn

The mass is 95, the density 0.7, the acceleration due to gravity 1.1, and the parabolic velocity 3.3 (37 km/s), of the Earth values. Hydrogen, helium, and traces of methane and ammonia have been found in the atmosphere. The temperature above the cloud layer is lower than for Jupiter. A well-developed ammonia cryoatmosphere is most likely present. No water has been detected so far, but its presence is not unlikely.

The Saturn rings, 113,000 km in width and over 60 km in thickness, comprise a special region of cryogenous formations. These rings consist of ice formed from water that has been centrifuged from the atmosphere.

Uranus and Neptune

These planets are evidently similar to some extent to the aforementioned giant planets. They have their own cryoatmospheres. The atmospheric temperature above the clouds is below -185°C.

The information concerning Pluto is very scanty.

It would be important to summarize the available data on the cryogenous phenomena within the solar systems. The study of astrocryology, which includes various cryogenous processes and formations in space and on the planets of the solar system, is more justified than, for example, the study of astrobotany. Astrocryology could try to solve a number of theoretical problems:

1. The study of cryogenous phenomena is space and in the cryospheres of planets in the solar system from the point of view of astrophysics, thermodynamics, geochemistry, physical chemistry, and other sciences.
2. The structure of the cryospheres of individual planets, the substances they consist of, and transformations of these substances by freezing.
3. The importance of planetary cryospheres in the development of planets and their external thermodynamic regions.
4. The laws of formation and evolution of the cryospheres starting with the cooling stage of circumplanetary space and the appearance of the atmospheres.
5. The dynamics of the cryospheres in relation to external phenomena and tectonic reconstructions on a planetary scale.

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CONTEMPORARY GEOTHERMAL CONDITIONS OF THE EXISTENCE AND DEVELOPMENT OF PERMAFROST

V. T. BALOBAEV, V. N. DEVYATKIN, AND I. M. KUTASOV Institute of Geocryology of the Siberian Department of the Academy of Sciences of the USSR

Permafrost was formed and exists in definite thermal conditions, characterized in the first place by a field of negative temperatures and phase transitions of water into ice. The temperature field in the Earth's crust is gradiental, that is, at any point in the ground there is a heat flow, the resultant vector of which is directed towards the surface, while its value is determined by the nature of the temperature field and the thermal properties of the materials. The temperature field is formed under the influence of the heat exchange on the surface, the composition and structure of the materials, the geothermal flow, local sources of heat (positive or negative), and phase transitions of water. It is irregular in space and changes with time, basically due to change in the heat exchange on the surface.

The determination of the nature of the temperature field as it depends on its determining factors is a direct task of geothermy that is solved theoretically and experimentally. However, in geothermy, as in any other area of geophysics, it is often necessary to solve an inverse problem. All the most important parameters of the temperature field are directly determined as the result of geothermal research, but the causes determining their concrete form remain unknown. The solution to the inverse problem is not unique: To each concrete form of the temperature field there corresponds a varied combination of boundary conditions, internal sources of heat, nonuniformity of the surface, environment and degree of non-steady-state of the processes of thermal conductivity, freezing, and thawing. For this reason, it is possible to increase the reliability of the interpretation of geothermal data only if the amount of experimentally determined parameters is increased. The most important of them is the heat flow--the basic energy indicator of the intensity of the thermal processes in the ground and their variation in time. In the permafrost area, it serves as the main criterion of the steadiness or nonsteadiness of the regime. This follows from Stefan's condition at the phase boundary,

$$q_F - q_T = \pm Q\frac{d\xi}{d\tau}, \qquad (1)$$

where q_T and q_F are the heat flow at the phase boundary in thawed and frozen ground [resp.]; Q is the amount of heat liberated (absorbed) per unit volume in freezing (thawing); ξ is the thickness of the frozen ground; and τ is the time.

It is obvious that the condition $q_F = q_T$ corresponds to a steady regime, $q_F > q_T$ corresponds to a regime of freezing and $q_F < q_T$ corresponds to a thawing regime.[1]

Gradual change in the heat flow with depth indicates both the possible non-steady-state of the heat regime and the possible existence of distributed sources (flows) of heat in conditions of homogeneity of the environment and relief. Considerable horizontal changes in value as a rule indicate inhomogeneity of the environment in this direction, the effect of the relief or of local sources of heat liberation (heat absorption).

Integrated geothermal research undertaken in recent years in the permafrost area, including the determination of the temperature field, heat flows, and thermal properties of earth materials, and also analysis of the existing data, made it possible to clarify a number of important features in the formation and development of frozen series. Factual data accumulated can be divided into three groups.

The first group, which comprises the greater part of geothermal data (Figure 1), obtained by the Institute of Geocryology and partially borrowed from the literature,[2] indicates that the temperature field of permafrost is steady over a considerable part of its zone of distribution. This corresponds to contemporary conditions of the heat exchange on the surface and the subterranean heat flow. None of the curves of distribution of temperature by depth are subjects to any changes in the transition from frozen to thawed states, that is, on the boundary the basic condition of steadiness of the heat regime of frozen ground, $q_F = q_T$, is met. However, smooth changes of the heat flow with depth are observed, for example, in the areas of Ulakhan-Egelyakh and Deputatskii (curves 3 and 4), which have a mountainous relief. Curve 3 shows the temperature of the ground in the valley and Curve 4, on the slope. Redistribution of the heat flow takes place under the effect of relief, and, for this reason, it is greater than the normal value under valleys, close to the surface (curve 3), while it is less under uplands. This can be seen from Table 1, compiled from observations in the Deputatskii area. The undisturbed value of the heat flow is equal to 1.9 $\mu cal/cm^2 \cdot s$.

The steady-temperature field of frozen ground and its thickness are definitely determined by the temperature of the surface, the heat flow, and the thermal conductivity of the materials. The effect of surface temperature on the thickness of the frozen zone can be seen in Figure 1. On the other hand, the fact that it is greater

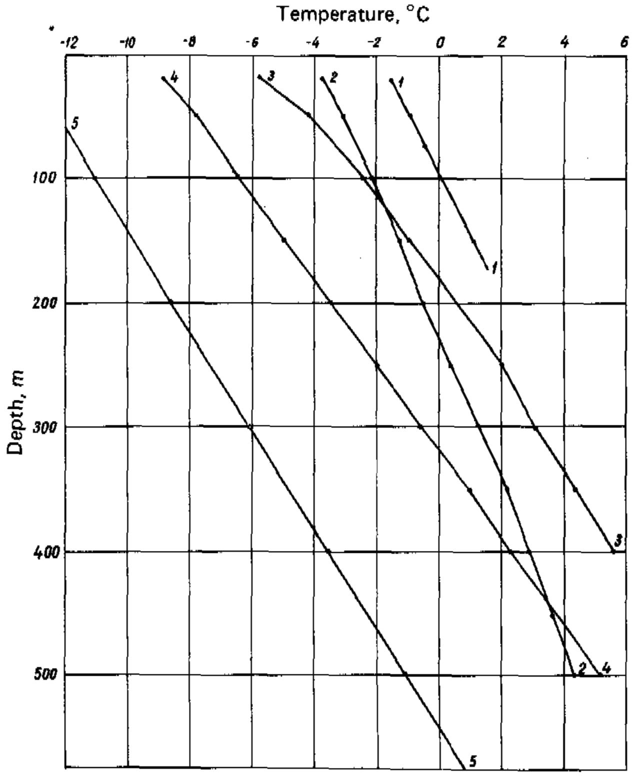

FIGURE 1 Vertical profiles of the steady-state temperature field of the ground in various areas. 1--Ardag (Northern Mongolia); 2--Amga; 3--Ulakhan-Egelyakh (Yano-Adychskoe highland); 4--Deputatskii (Polousnyi Ridge); 5--Nordvik.

in the Amga area (curve 2) than in Ulakhan-Egelyakh, despite the greater surface temperature, confirms the role of internal heat flow in the formation of the permafrost. The heat flow in the area of the Amga is equal to 1.2 μcal/cm^2•s, which also led to the formation of a less-thick frozen series in a colder area. In a uniform environment in the absence of the relief effect, the thickness of the frozen series is determined by the dependence

$$\xi = \frac{(T_0 - T_s)\lambda_F}{q} \qquad (2)$$

where T_0 is the freezing temperature of water, λ_F is the thermal conductivity of frozen material, T_s is the temperature of the earth's surface. Steady-state series of frozen rocks are found everywhere in areas of development of compact slightly fissured rocks of Jurassic and post-Jurassic age. These rocks, on the one hand, have a comparatively high thermal conductivity and, on the other hand, a comparatively low water content. The processes of heat transfer and freezing-thawing take place in them relatively quickly, in time to follow secular changes in heat exchange conditions on the surface. At the same time, the wide development of steady-state series of frozen rocks shows that paleoclimatic changes in these areas took place rather slowly.

The second group of geothermal data (Figure 2) gives a number of interesting and unexpected results. It characterizes the temperature field of the western part of the East Siberian platform, extending from the Anabar massif in the north to the latitudinal section of the course of the Lena River in the south. All of this area is composed mainly of ancient Cambrian and Ordovician carbonate rocks containing strongly mineralized waters and brines. The greatest thickness on Earth of freezing and cooling of the Earth's crust to negative temperatures is observed here. In the Markhinskaya wells (Figure 2, curves 1 and 2), the zero isotherm is situated at a depth of 1,500 m. The distributions of temperatures with depth shown are the results of repeated observations

TABLE 1 The Effect of Relief on the Heat Flow in the Ground (μcal/cm^2•s)

Depth, m	Heat Flow Crest	Slope	Foot	Slope	Crest	Slope
70	1.41	1.75	3.00	1.98	1.50	2.19
150	1.48	1.80	2.50	1.98	1.69	1.95
300	1.85	1.85	2.10	1.95	1.91	1.91

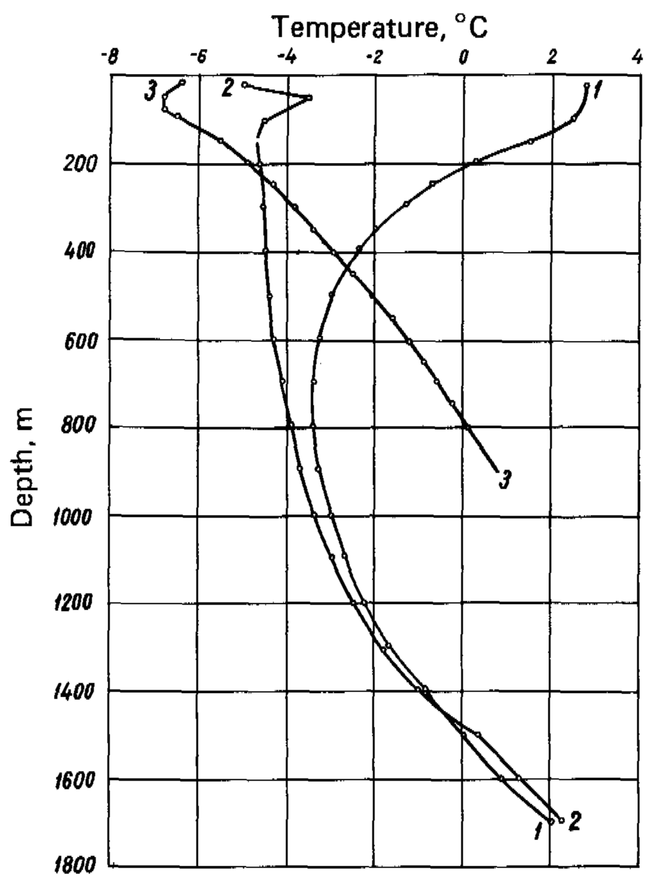

FIGURE 2 Vertical profiles of the temperature field of rocks west of the East Siberian platform. 1--Markhinskaya well 1; 2--Markhinskaya well 2; 3--well of the "Udachnaya" pipe.

during an 8-yr period from the moment drilling stopped. The very unusual appearance of the curves is due to the fact that the heat flow constantly increases with depth, while in the area of well 1 it even changes sign because of the development of a local surface talik, the genesis of which is obscure. All of the findings given in Figure 2 relate to the latitude of the Arctic Circle. If temperature curve 2 is analyzed, then its form and features of change of heat flows can show, in the first place, the non-steady-state of the heat regime of the ground caused by increase in surface temperature, or secondly, the existence of negative sources (or flows) of heat distributed in the rocks, or both at the same time.

The picture obtained of the temperature field is difficult to explain in terms of the non-steady-state of the heat regime. At a depth of 1,500-1,700 m, the heat flow is equal to 0.6 μcal/cm·s. This value is approximately half the normal value for platform conditions of the same type. This means that unsteadiness of the temperature field must also be observed deeper, as far as 2,500-3,000 m. Such a field can be a consequence of very prolonged and wide-ranging changes in the temperature on the surface. Neither contemporary conditions on the surface, nor the temperature field of rocks of adjacent areas, indicate this magnitude of change, although the thermal processes in the area of distribution of frozen rocks develop precisely in this direction. In addition, limestones and dolomites, predominantly distributed in the section, are characterized by high thermal conductivity (λ = 2.1-2.9 kcal/m·h degree), which must promote rapid stabilization of the heat regime. Curve 3 (Figure 2), obtained 100 km to the west of the "Udachnaya" pipe, does not show any tendencies to unsteadiness of the temperature field, although the heat flow there is also anomalously low and is equal to 0.4-0.5 μcal/cm^2·s at a depth of 600-800 m.

Absorption of the heat flow in its movement to the surface as the result of physicochemical processes in earth materials must be assumed. Subterranean waters in these areas are brines with a salt content of 250-350 g/l. A number of dissolved substances can be found in saturated condition, and consequently, in these conditions, processes of crystallization, exchange, and replacement with absorption of heat can take place.

More severe conditions existed in the permafrost zone in the period preceding the present time. There are many proofs of this, including geothermic proofs, which will be discussed below. The zone of cooling and freezing was lower than the existing zone. In the ensuing period of warming up it contracted from the surface. In the areas under study, this contraction in the thickness of the zone of ancient cooling was slowed down and is being slowed down by the physicochemical processes accompanying the absorption of heat. If rocks and the solutions in them are considered as a balanced thermodynamic system, then, when such a system is heated from without, only those processes can occur in it that impede warming up or that are accompanied by absorption of heat (Le Chatelier-Brown principle). These could be processes of crystallization from solution of gypsum, limestones, and dolomites; dissolving of rock salt; and dehydration reaction.

As the result of these processes, the diffusion of heat from the surface is inhibited, and rocks at great depths remain cooled and even continue to get colder. The diffusion of heat flows by depth shows that in the area of Shelagontsy (Markhinskaya wells) processes involving absorption of heat take place in the zone of negative temperatures, as well as considerably below it. In the area of the Udachnaya pipe, these processes are observed only in a zone with positive temperatures, since above this the heat flow does not change with depth, remaining, however, very low.

Thus, in the zone of development of ancient carbonate rocks that contain strongly mineralized waters we find a complex temperature field with internal sources of absorption of heat. It is marked by a thick zone of negative temperatures within which rocks contain ice only in the upper (100-300 m) near-surface part.

If such a field is steady, then its change with depth can be described by the formula

$$T = T_s + \frac{q}{\lambda} z + \frac{P}{2\lambda} z(2H-z), \qquad (3)$$

when the heat sources P are evenly distributed in a uniform rock massif, starting from the surface, and

$$T_1 = T_s + \frac{z}{\lambda}(q + PH - Ph), \quad z < h, \qquad (4)$$

$$T_2 = T_s + \frac{q}{\lambda} z + \frac{P}{2\lambda}(2zH - z^2 - h^2),$$

$$z > h, \qquad (5)$$

when the heat sources are found only below depth h. The heat flow begins at depth H, below which there are no sources.

It can be shown that with an appropriate selection of a P value Equation (3) corresponds rather well to curve 2, while Equation (4) corresponds to curve 3 (see Figure 2). But because of the ambiguity of the solution of the inverse problem of geothermy, this does not signify steadiness of the temperature field. In all probability, it is nonsteady in the area studied and is described by more complex relations.

We cited factual data for a comparatively small area, but similar conclusions with some nuances hold for all of the region indicated. In similar conditions, but 450 km to the south, in the area of Mirnyi, with a ground surface temperature of -3°C, the zone of negative temperatures has thickness 560-600 m, despite the fact that the boundary of continuous permafrost passes 200 km to the south (in the area of Lensk the ground surface temperature is positive).

Unfortunately, to accurately map this specific area is impossible, since there are few factual data, but the anomalously low temperature gradients at many points indicate extensive size.

The third group of geothermic data (Figure 3) characterizes a clearly transient temperature field accompanied by processes of thawing of the frozen series.

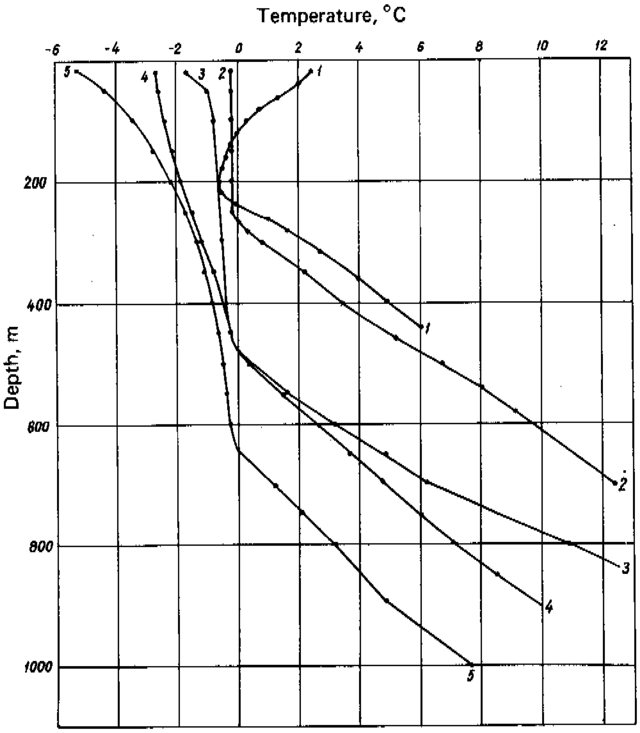

FIGURE 3 Vertical profiles of a transient temperature field of rocks in different areas. 1--Lower Obl; 2--Kostrovskii; 3--Middle Vilyui; 4--Namtsy; 5--Bakhynai (Zhigansk).

All curves of distribution of temperature with depth show that $q_T > q_F$ (Table 2). This corresponds to thawing of frozen rocks under the influence of the heat flown from the thawed zone.

The rate of thawing at the present time can be calculated by the formula:

$$-\frac{d\xi}{d\tau} = \frac{q^{\tau} - qF}{Q}. \qquad (6)$$

Its calculated value is equal to 1-2 cm/yr. It is obvious that direct measurements cannot detect movement of the phase boundary.

Non-steady-state frozen rocks are confined to areas of development of rather thick series of young, slightly cemented sediments of Quaternary, Tertiary, and Cretaceous age. They are found in the north of western Siberia as far as the Yenisei,[3] in the central part of Yakutsk along the middle and lower courses of the Lena and Vilyui rivers, and in the lower reaches of the Aldan River. Their existence in the basin of the Khatanga and the lower reaches of the Anabar and Olenek rivers, in the Anadyr, Markovo, Penzhino, and Main lowlands, and also in the intermontane depressions of the Verkhoyansk-Kolyma and Chukotsk mountain regions is beyound doubt.

Thawing of frozen rocks takes place in these territories at the present time mainly from below with the preservation of a negative temperature on the surface. This shows the existence of a more severe climate in the past. Figure 3 shows that the contemporary series of perennially frozen rocks could have been formed when the temperature on the surface was less than -13° to -15° in the Zhigansk area (Bakhynai well), -10° to -12° in central Yakutsk (Namtsy and Middle Vilyui well) and -6° to -8° in the north of western Siberia and in the lower reaches of the Yenisei. It is now 7°-8° warmer in these areas. If it is taken into consideration that thawing of frozen series has continued now for 20-25 thousand years, then still more severe conditions must correspond to its maximum thickness.

An approximate evaluation of the dynamics of the thickness of the frozen series during thawing from below and linear increase in surface temperature

$$T^{s}(\tau) = T_0 + \beta\tau$$

can be made by means of the following formulas:

$$\text{with } \beta < \frac{q\tau^2}{4\lambda_F Q}$$

$$= \ln\frac{T_0^{\,2}(Q\beta\xi^2 + q_T\xi T_s + \lambda_F T_s^{\,2})}{T_s^{\,2}(Q\beta\xi_0^{\,2} + q_T\xi_0 T_0 + \lambda_F T_0^{\,2})} - \frac{q\tau}{\sqrt{-\Delta}}\left[\ln\frac{(q\tau - \sqrt{-\Delta})T_s + 2Q\beta\xi}{(q\tau + \sqrt{-\Delta})T_s + 2Q\beta\xi} - \ln\frac{(q\tau - \sqrt{-\Delta})T_0 + 2Q\beta\xi_0}{(q\tau + \sqrt{-\Delta})T_0 + 2Q\beta\xi_0}\right] \qquad (7)$$

$$\text{with } \beta = \frac{q\tau^2}{4\lambda_F Q}$$

$$2\ln\frac{T_0}{T_s} = \ln\frac{T_0^{\,2}(Q\beta\xi^2 + q\tau\xi T_s + \lambda_F T_s^{\,2})}{T_s^{\,2}(Q\beta\xi^2 + q\tau\xi_0 T_0 + \lambda_F T_0^{\,2})} + \frac{2q\tau T_s}{q\tau T_s + 2Q\beta\xi} - \frac{2q\tau T_0}{q\tau T_0 + 2Q\beta\xi_0}, \qquad (8)$$

$$\text{with } \beta > \frac{q\tau^2}{4\lambda_F Q}$$

$$2\ln\frac{T_0}{T_s} = \ln\frac{T_0^{\,2}(Q\beta\xi^2 + q\tau\xi T_s + \lambda_F T_s^{\,2})}{T_s^{\,2}(Q\beta\xi_0^{\,2} + q\tau\xi_0 T_0 + \lambda_F T_0^{\,2})} - \frac{2q\tau}{\sqrt{\Delta}}\left[\arctan\frac{q\tau T_s + 2Q\beta\xi}{T_s\sqrt{\Delta}} - \arctan\frac{q\tau T_0 + 2Q\beta\xi_0}{T_0\sqrt{\Delta}}\right], \qquad (9)$$

TABLE 2 Heat Flows in Thawed and Frozen Zones ($\mu cal/cm^2 \cdot s$)

Region	q_T	q_F
Namtsy	1.1	0.38
Middle Vilyui	1.3	0.28
Bakhynai	1.1	0.20
Turukhan	1.2	0.00

where $\Delta = 4\lambda_F Q\beta - q_T^2$, ξ_0 is the initial thickness of frozen rocks at time $\tau = 0$.

Despite the copious data, there are still none showing contemporary increase in the thickness of the frozen strata in any area, not counting local areas of new formations of frozen rocks in the flood plains of rivers, in the bottoms of drying-up lakes, and in a zone of their wide development. The dynamics of frozen ground, basically of small thickness, is very complex here, and is frequently associated with nonclimatic factors. This indicates the unity of the process of warming of frozen ground in the entire north of the USSR.

Geothermic methods of research make it possible to give the quantitative characteristics of this process at the present time and in the past. However, drilling of deep special wells in different areas, together with a wide program of research, is necessary for this.[4]

During thawing, frozen ground is in an unbalanced thermodynamic state. In these conditions, heat transfer cannot take place in isolation, but is accompanied by a number of secondary processes which lead to changes in all the basic thermodynamic parameters and properties, together with the enclosed gases and solutions. The energy of phase transitions of water is so great that the intensity of these processes was very considerable, and the changes in earth materials resulting from them were noticeable. There already exist factual data that indicate intensive fissuring on the phase boundary, the presence of a powerful compression-suction mechanism of movement of underground waters, which changes their composition and the characteristics of gas in rocks.

Thermodynamic phenomena in series of frozen and thawed rocks and soils during thawing and freezing have still not been sufficiently clarified. Their study is one of the main tasks for the immediate future.

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THERMAL CONDITIONS IN THE TOPSOILS IN THE PERMAFROST REGION AS A FACTOR OF CRYOGENESIS AND PEDOCRYOGENESIS

V. N. DIMO V. V. Dokuchaev Soil Institute

The temperature field in the layer of annual temperature fluctuations in permafrost depends first of all on the thermal balance of the active surface. The thermophysical properties of topsoils, soils, and rocks are of great importance here.

The mean annual temperature of the topsoil is a parameter that determines during the change of seasons the heat exchange in the topsoil itself and in the system: air layer at the surface-vegetation-topsoil-soil-rock. The mean annual temperature may remain the same, although it annual amplitudes may vary. At the same time, the greater the role of soil heating in the annual cycle, the higher the mean annual temperature, and the greater the role of cooling, the lower the mean annual temperature. The sign of the latter corresponds to the sign of the difference between the sums of above- and below-zero temperatures.

Therefore, in the study of contemporary cryogenesis, the mean annual temperature of the topsoil is a much more reliable indicator than the air temperature at the ground surface.

Maps of the latest meteorological data[4] show that the zero isotherm of the mean annual temperature of the topsoil crosses the territory of the USSR along a diagonal from the Cheshskaya Guba in the northwest to Kyakhta on the Mongolian border in the southeast (Figure 1). Generally speaking, the region of negative mean annual temperatures of the topsoil coincides in the USSR with the permafrost region. At the same time, the geographical boundary of the permafrost region does not coincide with the zero isotherm. For example, in western Siberia the latter is displaced to the north. The southern part of the Krasnoyarsk Krai (south of the Angara River), Khakasskaya and Gorno-Altaiskaya Oblasts, Tuva ASSR, and Irkutsk Oblast, as well as the southern part of Chita Oblast, the southeastern part of Amur Oblast, the southern part of Khabarovsk Oblast, and most of Kamchatka, have positive mean annual temperatures of the topsoil although all of them lie within the permafrost region. Displacement of the zero isotherm occurs in the direction of the cold pole from the south, west, and east.

From the point of view of the heat balance, the temperature of the upper layer of topsoil must correspond to the temperature of the lower surface of the layer of annual fluctuations, where the annual amplitudes become zero. The thickness of the layer of annual temperature fluctuations depends mainly on the thermophysical properties of the system: topsoil-soil-rock. Three-dimensional distribution of the depth of the lower boundary of annual temperature fluctuations depends on the supply of radiant energy of the sun, as well as on the changes in the effective thermal conductivity of this layer.

On the basis of the thermophysical factor, we may assume that the permafrost region, which is characterized by positive mean annual temperatures of the topsoil, should contain relict formations only, while the boundary of permafrost untouched by degradation should be displaced at present in the directions of the cold poles.

We analyzed the soil temperature data gathered by the Hydrometeorological Service (*Handbooks on the climate of the USSR*) and on this basis carried out a comparative study of thermal conditions in topsoils that rest on permafrost.[3]

The factor chosen to distinguish between different topsoils is their heat absorption, which is estimated from the numerical value of the ratio of the sum of active soil temperatures at a depth of 0.2 m to the sum of active temperatures of the air layer at the surface. Figure 2 shows the changes in the heat absorption as a function of soil and bioclimatic conditions in the plains.

The heat-absorption factors are below unity in Eurasian polar (A), east Siberian permafrost-taiga (D), and eastern brown-soil-forest (H) soil-bioclimatic regions that form part of the permafrost region. The heat absorption factors are above unity in western meadow-forest (B), central taiga-forest (C), western brown-soil-forest (F), far eastern taiga-meadow-forest (E), central forest-steppe and steppe (G), and desert-steppe and desert (I) soil-bioclimatic regions, as well as in subtropical, moderately warm regions: wet-forest (J), xerophyte-forest (K), and desert-steppe and desert (L).

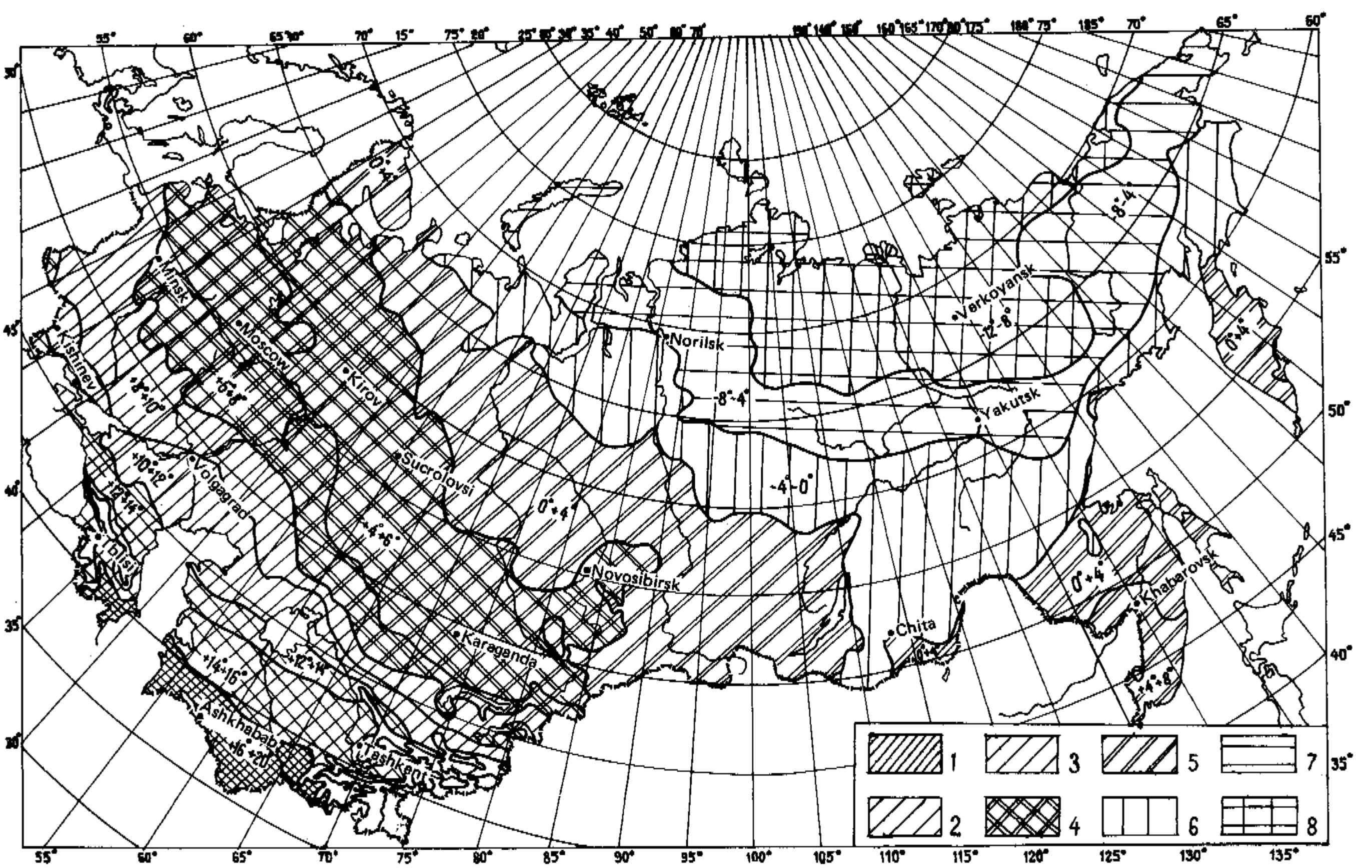

FIGURE 1 Mean annual temperature of soils (°C) at a depth of 0.2 m. 1--from +16 to +20, 2--from +12 to +16, 3--from +8 to +12, 4--from +4 to +8, 5--from 0 to +4, 6--from -4 to 0, 7--from -8 to -4, 8--from -12 to -8.

Since the physical meaning of the heat absorption factor excludes the effect of radiation, the differences in its numerical values can be explained by differences in the thermal state of the topsoil.

The sums of active soil temperatures and the heat absorption factor will differ, even if the sums of active air temperatures at the surface, which indirectly characterize the radiation conditions,[1,2] are the same or almost the same.

	$\Sigma t_A > 10°$	$\Sigma t_S > 10°$	Index of Heat Absorption
Vychegda province of the middle-taiga subzone of podzolized soils	1,350	1,410	1.04
Central Yakut province of the middle-taiga subzone of frozen-taiga and pale-yellow soils	1,370	1,280	0.94

NOTE: t_A = air temperature;
t_S = surface temperature.

Consequently, the "cold reserves" in permafrost have a direct effect on the thermal state of soils during the period of active temperatures and hence on the thermal state of the thawing layer.

It has been established that permafrost is not in a state of physical and chemical rest. The processes of energy and mass transfer are constantly active in permafrost owing to heat and moisture gradients. The thermal gradients of the system topsoil-soil have been determined for the USSR.[4]

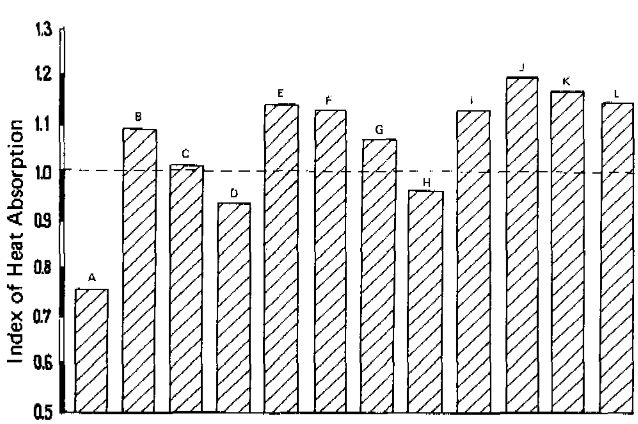

FIGURE 2 Heat absorption of top soil in different soil-bioclimatic regions: A--Eurasian polar; B--western meadow-forest; C--central taiga-forest; D--East-Siberian permafrost-taiga; E--far eastern taiga-meadow-forest; F--western brown-soil-forest; G--central taiga-steppe and steppe; H--eastern brown-soil-forest; I--desert-steppe and desert; J--subtropical, moderately warm, wet-forest; K--subtropical, moderately warm, zerophyte-forest; L--subtropical, moderately warm, desert-steppe and desert.

TABLE 1 Thermal Gradients of the System: Topsoil-Soil and Continental Characteristics of the Climate (Plains)

Soil-Bioclimatic Region	Annual Temperature Amplitudes, °C: Air	Annual Temperature Amplitudes, °C: Soil at a Depth of 0.2 m	Thermal Gradients, deg/m: Warmest Month	Thermal Gradients, deg/m: Coldest Month
A	31.3	17.0	-4.36	2.98
B	19.8	7.5	No Data	
C	31.8	20.2	-5.29	3.07
D	49.0	32.7	-9.15	8.43
E	33.4	19.4	-4.79	3.86
F	24.0	21.0	No Data	
G	43.7	29.8	-7.58	7.36
H	35.3	26.7	-6.93	5.71
I	34.7	31.3	-6.15	5.86
J	17.8	20.1	-3.43	2.43
K	21.0	24.8	-5.58	6.07
L	28.6	29.1	-5.80	6.15

The thermal gradient determines the intensity of heat fluxes, while its sign indicates the direction of the latter* and the type of heat exchange (insolation, radiation).

Comparison of thermal gradients of the system topsoil-soil (0.2-1.6 m) in different soil-bioclimatic regions shows that the gradients decrease with continental characteristics of the climate (Table 1). The distribution of thermal gradients in soil-bioclimatic regions in the coldest month represents a mirror reflection of their distribution in the warmest month.

It must be assumed that in accordance with the laws of heat-and-mass transfer, the heat transfer is especially pronounced in the topsoils of the most continental parts of the permafrost region, and this applies to the underlying soils as well. The intensity of heat-and-mass transfer determines to a great extent the nature of the cryogenesis of topsoils (pedocryogenesis) as well as soils (geocryogenesis).

Since pedocryogenesis and geocryogenesis are interdependent and in the final analysis represent a single process of cryogenesis, a classification of the temperature regime of topsoils, in which the taxonomic determination of types is based on the intensity of freezing, may be of interest to geocryologists as well (Table 2). Such a classification has been developed on the basis of the topsoil temperature data provided by the Hydrometeorological Service.[2]

Topsoils have been divided into two groups:

*In our understanding, a positive sign of the gradient indicates a heat flux toward the surface of the soil, while a negative sign indicates a heat flux from the surface. In both cases the heat is flowing toward a heat sink.

1. Freezing soils. The temperature in the coldest month at a depth of 0.2 m is below zero.
2. Nonfreezing soils. The temperature in the coldest month at a depth of 0.2 m is above zero.

The topsoils in the permafrost region belong to the first group. They are characterized by two types of the temperature regime:

1. Perennial freezing conditions. The cooling process predominates and is accompanied by freezing of soil moisture down to the permafrost table. Permafrost is of a continuous type. Permafrost islands and pereletoks may occur. The mean annual soil temperature is always below zero.
2. Long-term seasonal freezing conditions. The warming process predominates and is accompanied by slow thawing of soil moisture. Permafrost is of the island type. There are pereletoks. Permafrost may be absent altogether. The mean annual temperature is mostly above zero.

If the topsoil is dry, there is naturally no freezing (within the range of possible negative temperatures). Soil scientists often use the term "dry frost," meaning the absence of freezing moisture. It is much better to use the term "frostiness" suggested by P. F. Shvetsov.[5]

Dry freezing (frostiness) cannot have the same effect in the process of topsoil formation as normal freezing of moist soils, which is accompanied not only by a redistribution of soil moisture and substances dissolved in the latter, but also by a change in the soil structure. Therefore, we feel that it would be better not to include a dry-freezing regime in the classification, since the phenomenon of frostiness occurs mainly with sandy soils in regions where there is no snow and no precipitation in the fall.

The subtypes of the temperature regime of soils were established on the basis of the mean annual temperatures of the warmest and the coldest months, the sums of positive temperatures and the depths of their penetration (the main characteristics of the warm period), the sums of negative temperatures and the depths of their penetration (the main characteristics of the cold period), and the continental characteristics of the topsoil climate as indicated by the annual temperature amplitudes.[2,4]

The perennial freezing temperature regime is characterized as follows.

1. On the basis of the annual cycle:
   (a) very cold subtype (the mean annual topsoil temperature at a depth of 0.2 m ranges from -12°C to 0°C).
2. On the basis of conditions prevailing during the warm period, two subtypes:

   (a) very cold (the soil temperature at a depth of 0.2 m in the warmest month ranges from 0°C to +12°C, the sum of positive temperatures ranges from 0°C to +500°C, and the depth of their penetration is less than 1 m);
   (b) cold (the soil temperature at a depth of 0.2 m in the warmest month ranges from 8°C to 20°C, the sum of positive temperatures from 500°C to 1,500°C, the depth of their penetration may reach 2 m).
3. On the basis of conditions prevailing in the cold period, two subtypes:
   (a) very cold (the soil temperature at a depth of 0.2 m in the coldest month varies from -20°C to -16°C, the sum of negative temperatures ranges from 2,500°C to 2,000°C, the depth of their penetration is less than 3 m);
   (b) cold (the soil temperature at a depth of 0.2 m in the coldest month ranges from -16°C to -4°C, the sum of negative temperatures ranges from 2,000°C to 1,000°C, the depth of their penetration is 2-3 m).
4. On the basis of continental characteristics of the soil climate indicated by the annual temperature amplitudes at a depth of 0.2 m, there are the following subtypes:
   (a) moderately continental (from 16°C to 24°C);
   (b) continental (from 24°C to 32°C);
   (c) distinctly continental (from 32°C to 40°C);
   (d) extracontinental (above 40°C).

The temperature regime of the permafrost type occurs in several provinces on the plains of the Eurasian polar and the east Siberian permafrost-taiga soil-bioclimatic regions. In the first of these the warm period is characterized by a very cold subtype, while the cold period is characterized by a cold subtype. The climate is moderately continental.

In the east Siberian region we find the cold subtype of the warm period and the very cold subtype of the cold period. The climate varies from continental to distinctly continental. In the mountainous provinces (Verkhoyansk and Kolyma) the climate is, in places, of the extracontinental subtype.

The temperature regime of the seasonally frozen type is characterized as follows.

1. On the basis of the annual cycle by two subtypes:

   (a) very cold (the mean annual temperature of soil at a depth of 0.2 m ranges from -12°C to 0°C);
   (b) cold (the mean annual soil temperature at a depth of 0.2 m ranges from 0°C to +8°C).
2. On the basis of conditions prevailing in the warm period, by three subtypes:
   (a) very cold (the soil temperature at a depth of 0.2 m in the warmest month ranges from 0°C to +12°C, the sum of positive temperatures ranges from 0°C to 500°C, the depth of their penetration is less than 1 m);
   (b) cold (the soil temperature at a depth of 0.2 m in the warmest month ranges from 8°C to 20°C, the sum of positive temperatures ranges from 500°C to +1,500°C, the depth of their penetration may reach 2 m);
   (c) moderately cold (the soil temperature at a depth of 0.2 m in the warmest month ranges from 16°C to 24°C, the sum of positive temperatures ranges from 1,500°C to 2,500°C, the depth of their penetration may reach 3 m).

TABLE 2 Classification of the Temperature Regime of Topsoils of the USSR

Groups	Types	Subtypes with Respect to Temperature Conditions			Subtypes with Respect to Continental Character of Climate
		Year	Warm Period	Cold Period	
	Permafrost	Very cold	Very cold Cold	Very cold Cold	Moderately continental Continental Very continental Extracontinental
Freezing topsoils	Long-term seasonal freezing	Very cold Cold	Very cold Cold Moderately cold	Cold Moderately cold	Mild (oceanic) Moderately continental Continental Very continental
	Seasonal freezing	Cold Moderately cold Moderately warm	Very cold Cold Moderately cold Moderately warm Very warm	Moderately cold	Mild (oceanic) Moderately continental Continental Very continental
Nonfreezing topsoils	Nonfreezing	Moderately cold Moderately warm Warm	Moderately cold Moderately warm Warm Very warm Hot	Moderately warm	Mild (oceanic) Moderately continental Continental

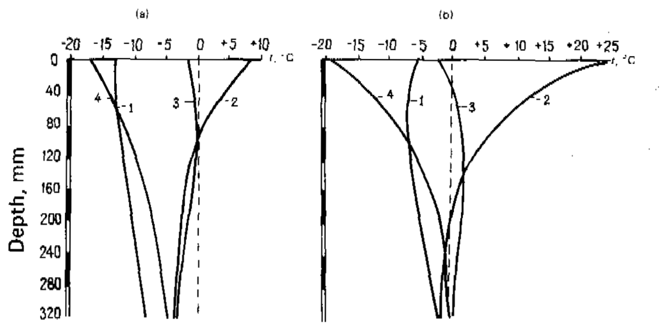

FIGURE 3 Seasonal temperature distribution in topsoils with a temperature regime of the permafrost type. (a) Subarctic tundra soils of the Chukotka-Anadyr province, (b) Frozen-taiga and pale-yellow soils of the central Yakut province, 1--April; 2--July (August); 3--October; 4--January (February).

3. On the basis of conditions prevailing in the cold period, two subtypes:

(a) cold (the soil temperature at a depth of 0.2 m in the coldest month ranges from -16°C to -4°C, the sum of negative temperatures ranges from 2,000°C to 1,000°C, the depth of their penetration is 2-3 m);

(b) moderately cold (the soil temperature at a depth of 0.2 m in the coldest month ranges from -12°C to 0°C, the sum of negative temperatures ranges from 1,000°C to 0°C, the depth of their penetration does not exceed 2 m).

4. On the basis of the continental characteristics of the soil climate as indicated by the annual temperature amplitudes at a depth of 0.2 m, there are the following subtypes:

(a) mild oceanic (the annual amplitude below 16°C);

(b) moderately continental (16°C-24°C);

(c) continental (24°C-32°C);

(d) distinctly continental (32°C-40°C).

We can see that together with subtypes which are common for the temperature regimes of both permafrost and seasonally frozen types, there are subtypes which in the latter case characterize warmer conditions in both winter and summer as well as a trend towards less continental climate and warmer mean annual temperatures. Furthermore, within the same subtypes there may be higher temperatures in the warm period and lower temperatures in the cold period. Figure 3 shows the temperature distribution throughout the soil cross-section in different seasons.

The temperature distribution throughout the cross section of subarctic tundra soils in the Chukotka-Anadyr soil province (Figure 3a) is characteristic of the permafrost type of soil temperatures with a very cold subtype of the warm period and a cold subtype of the cold period. The same type but with an inverse distribution of subtypes is shown in Figure 3b, which illustrates the changes in the seasonal temperature throughout the cross section of the permafrost-taiga and the pale-yellow soils in the central Yakut province of the middle taiga subzone of the east Siberian permafrost-taiga region.

Figure 4 shows the temperature distribution characteristic of soils with a long-term seasonal freezing type of the temperature regime. The subarctic tundra soils in the Kanin-Pechora province are characterized by a very cold subtype of the temperature regime of the warm period and a moderately cold subtype of the cold period (Figure 4a). The frozen soddy-taiga podzolized soils of the Upper Zeya province are characterized by a different distribution of subtypes of the temperature regime. The warm period is of the moderately cold subtype, while the cold period is of the cold subtype (Figure 4b).

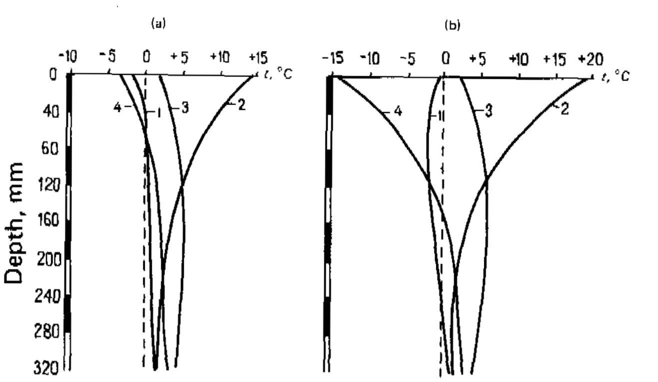

FIGURE 4 Seasonal temperature distribution in topsoils with a temperature regime of the long-term seasonal freezing type. (a) Subarctic soils of the Kanin-Pechora province, (b) Soddy-frozen-taiga and podzolized soils of the upper Zeya province. Symbols same as in Figure 3.

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RADIATION-HEAT BALANCE OF THE SOIL SURFACE AS A FACTOR IN THE FORMATION AND DYNAMICS OF SEASONALLY FROZEN SOILS AND PERMAFROST

V. A. KUDRYAVTSEV, B. N. DOSTOVALOV, AND L. S. GARAGULYA
Moscow Lomonosov State University

The formation and dynamics of seasonally frozen soils and permafrost depend on the structure of the radiation-heat balance of the surface. It is essential to examine the effect of all components of the radiation-heat balance on the temperature of the surface and the underlying soils, their freezing and their thawing within shorter periods of time in the annual cycle.

The change of seasons in middle and high latitudes is accompanied by abrupt changes in the amount of direct (Q) and scattered (q) radiation, the nature of the surface and its albedo (α), and, in the final analysis, in the amount of absorbed radiation. Considerable changes occur in the effective radiation (J) due to shifts in the surface temperature. The same is true of evaporation (LE). All these components change quantitatively throughout the year, but their signs remain the same. Not so with the turbulent (air convection) heat exchange (P) and heat storage in the soil (B). They undergo both quantitative changes and a change of sign. The latter is, as a rule, connected with the change of season.

Direct and scattered radiations reach a maximum in the summer and are reduced to a minimum in winter. At the same time, the snow cover and its albedo, which is close to 70-80 percent, reduce absorbed radiation almost to zero in winter. In the summer, effective radiation is invariably lower than absorbed radiation, since the radiation balance is positive and reaches such large values that its sum per year remains positive also anywhere on the Earth. The bulk of this energy is spent on evaporation, turbulent heat exchange, and heat exchanges in the soil. In winter, when absorbed radiation is abruptly reduced and the surface temperature passes through zero, evaporation and turbulent heat exchange become almost zero, and the radiation-heat balance can be described as follows:

$$(Q + q)\ (1 - \alpha) = J + B. \qquad (1)$$

It follows that such components as the heat spent on evaporation and the turbulent heat exchange are very important in the formation of positive temperatures of the soil surface in the spring and summer, when the supply of solar radiation is at a maximum. Formation of negative temperatures depends mostly on the effective radiation.

Freezing of soils is related to an abrupt reduction in the amount of solar radiation and an excess of effective radiation over absorbed radiation, i.e., to a negative radiation balance, as a result of which the temperature of 0°C is established on the surface of the soil. It is easily shown that at this time the negative radiation balance consists essentially of a negative annual heat exchange in the soil and the underlysing rocks. Therefore, the negative radiation balance depends on the structure of the annual heat storage in the soil.

At the temperature of the soil surface equal to 0°C, the main factor in the heat storage is the latent of fusion. This process maintains the zero temperature on the soil surface or close to it for a certain period of time (the zero curtain). Subsequently, during freezing of soil in winter, the temperature of the surface under the snow remains slightly higher. At the same time, absorbed radiation drops to a minimum, and, as a result of this, the difference between the latter and the effective radiation rapidly increases. The abrupt increase in this difference (the negative radiation balance) is due to heat released from soil as a result of the phase transition of water during freezing. Because of this, the negative radiation balance is a function of the latitudinal zonality.

Freezing of soil is greatly affected by the duration of the period during which a negative radiation balance exists on the radiating surface. It is this ratio of the duration of the periods with positive and negative radiation balances that determines the sign of the mean annual temperature of the soil surface, in spite of the fact that, as a rule, the sum of the positive radiation balance greatly exceeds the sum of the negative balance.

Geological and geographical factors play an important role in the formation of the structure of the radiation-heat balance. They include various kinds of natural cover (snow, vegetation, water), surface relief and the exposure of slopes, the composition and the water content of the surface deposits, and the hydrogeological conditions. These factors determine the insolation of the surface, the albedo, and all thermal processes. This is the reason why permafrost can exist in latitudinal zones with a large supply of radiation, while numerous taliks may be present in the zones where the amount of incoming radiation is small.

The relationship of the above factors to the radiation-heat balance of the surface and its temperature regime can be represented as follows:

$$T_{av} = \frac{1}{2\sqrt[4]{\sigma s}}$$

$$\left(\sqrt[4]{\frac{QG_{max} - LE_s - P_s - B_{av} - U + V}{(0{,}4 - 0{,}06\ e_s)(1 - cn_s^2) + 4\ \frac{TG \cdot s}{T_{max}} - 1}} + \right.$$

$$+\sqrt[4]{\frac{QG_{min} - LE_{w} - P_{w} + B_{av} - U}{(0,4 - 0,06\sqrt{e_{w}})(1 - cn^{2}_{w}) + 4\left(\frac{Tg.w}{T_{min}} - 1\right)}}\Bigg), (2)$$

where σ is the Stefan-Boltzmann constant = 8.14 × 10^{-11}; s is the radiation of the surface relative to a black body expressed as a fraction (0.85 - 1.0); $Q_{G\ max}$ and $Q_{G\ min}$ are the extremal decade sums of absorbed radiation in the summer and winter half-periods, respectively; the indices "s" and "w" mean that the decade sums of the components LE and P apply to summer and winter extremal values of Q_G; B_{av} is the decade value of the heat storage equal to the semiannual heat storage divided by 18 (the number of decades per year); e is the absolute moisture content of the air; n is the cloudiness expressed in fractions; c is the change of cloudiness with latitude; U is the amount of heat lost by soil in a decade in accordance with the gradient of the mean annual temperature in the layer of seasonal freezing or thawing; V is the amount of heat received by soil from precipitation in a decade; T_{av} is the mean annual temperature of the soil surface (°K); $T_{G.s}$ and $T_{G.w}$ are the mean decade temperatures of the soil surface at a time of maximum and minimum solar radiation, respectively (°K); T_{max} and T_{min} are the mean decade maximum and minimum temperatures of the air.

Depending on geological and geographical conditions, each region has its own specific aspects of formation of the radiation-heat balance. The structure of the latter may vary with different landscapes within one region. Permafrost surveys must account for these differences and indicate the importance of each element of the environment in the formation of the radiation-heat balance. This type of investigation makes it possible to forecast changes in permafrost resulting from construction. In this way it becomes possible to investigate the possibility of changing the permafrost conditions at will to ensure optimum performances of buildings and structures.

In permafrost studies, the study of the radiation-heat balance makes it possible to estimate the dynamics of the upper boundary conditions and the general nature of permafrost conditions in the region as a whole. Deviations from typical structure of the radiation-heat balance on individual sections can be estimated also. In this way the geographical distribution and the mode of occurrence of permafrost, as well as the conditions leading to formation of taliks, will be determined.

Of great importance is the fact that the structure of the radiation-heat balance of the surface is largely responsible for the mode of formation of seasonally frozen soils and permafrost. And this, in conjunction with the geological structure and the composition of soils, determines the cryogenous textures.

Thus, the study of formation of seasonally frozen soils and permafrost as a function of the radiation-heat balance determines not only the thermophysical characteristics (the temperature regime, the depth of seasonal freezing and thawing, the distribution and occurrence of permafrost and taliks), but also the cryogenous structure of permafrost.

Evidently, there is an equally close relationship between the radiation-heat balance and the nature of cryogenous processes and phenomena. Thermokarst is in essence a vivid expression of the effect of changes in the structure of the radiation-heat balance on the changes in the permafrost conditions. Thus, the increase in absorbed radiation resulting from the disturbance of the veaatation cover increases the temperature amplitude on the soil surface and the depth of thaw and hence leads to formation of thermokarst in the presence of ice. The same applies to solifluction, heaving, fracturing, and other processes.

The study of the radiation-heat balance and its structure is very important for the understanding of altitudinal and latitudinal zonalities in permafrost studies. It is evident that direct, scattered, and absorbed radiation, as well as effective radiation, evaporation, turbulent heat exchanges, and heat economy in the soil, depend on the altitudinal and latitudinal zonality and have different values in different climatic regions and geobotanical zones. Because of this, the study of the radiation-heat balance is evidently the only possible scientific basis for the compilation of small-scale permafrost survey maps.

Human activity changes the structure of the radiation-heat balance, and each of its components may be affected in a different way. Therefore, it is expedient to examine how a change in each component affects the temperature regime of soils, the depth of seasonal freezing and thawing, and the distribution, the mode of occurrence, and the composition of permafrost.

The amount of absorbed radiation $Q_G = (Q + g)(1 - \alpha)$ depends on the supply of the direct solar radiation Q, which within each region depends on the steepness and the exposure of terrain. Hence the leveling of terrain may considerably change Q. For example, at the latitudes of 60°N, the leveling of natural 30° slopes with northern exposure until a horizontal surface is obtained increases Q by 20-30 percent. In the case of similar slopes of southern exposure, there will be a decrease in Q, as compared with undisturbed conditions.

The relative change in Q_g depends also on the scattered radiation q, which remains constant in each region and does not depend on the changes in the steepness and the exposure of slopes and other surface parameters. The larger q, the smaller the effect of changes in Q on the changes of the radiation-heat balance.

The change in the albedo (α) of the ground surface resulting from human activity plays a more important role in the change of the radiation-heat balance. Removal of vegetation cover, afforestation, ploughing, and sowing of grass and shrubs change the albedo of the surface by 7-10 to 25 percent. One and the same change in α results in different changes in the amounts of absorbed radiation. Under conditions of severe continental climate, where $Q + q$ assumes large values, even the smallest change in α may result in considerable changes in the temperature of the soil. Under conditions of marine climate, these changes will be negligible.

In the Far North, a change in the amount of absorbed radiation is often related to a change in α of the snow surface in winter. Thus, the blackening of the snow in the Vorkuta area owing to a deposition of coal dust results in an earlier disappearance of snow, increases the annual amount of absorbed solar radiation, and increases the mean annual temperatures of the soil.

Changes in absorbed radiation lead to changes not only in the mean annual temperatures, but also in the annual temperature amplitudes on the ground surface as well. For example, an increase in the absorbed radiation in the summer will increase the amplitude of the temperature fluctuations throughout the year. Calculations and actual observations show that changes in the annual amplitude resulting from changes in the steepness and exposure of slopes do not exceed 2°-3°. Much greater changes result from the removal of vegetation, as well as from changes in the albedo of the surface. In this case the temperature amplitudes can change by 4°-5°, or even more.

The most significant changes in the temperature regime of the surface occur if an artificial cover is installed on the ground. For example, the amount of absorbed radiation resulting from a change in the albedo can change the mean annual temperatures and amplitudes by as much as 2°-3° in the case of concrete cover, and by 3°-4° in the case of asphalt.

Changes in the mean annual temperatures and annual amplitudes on the ground surface resulting from an artifical cover will change the depths of seasonal freezing and thawing. Since an increase in absorbed radiation leads to an increase in the soil temperature and in the annual amplitudes on the surface, there will be a slight decrease in the depth of seasonal freezing and a sharp increase in the depth of seasonal thaw.

The structure of the radiation-heat balance of the surface is greatly dependent on evaporation. In each region where the atmospheric conditions (the temperature and the moisture content of air, the wind rose) remain constant, the evaporation on each individual section depends on the moisture content of soil and the nature of vegetation. Such measures as drainage on territories under development result in an abrupt decrease in evaporation and in the final analysis lead to an increase in the mean annual temperatures of the soil. The same applies to the removal of vegetation, which dries the soil and reduces evaporation.

It should be noted that a change in absorbed radiation has a strong effect on evaporation, since the individual components of the radiation-heat balance are closely interrelated. In particular, an increase in Q_g resulting from changes in the steepness and exposure of slopes or in the albedo of the surface, with all other conditions remaining the same, will lead to an increase in the mean annual temperatures of the soil and in evaporation.

Changes in evaporation affect B_{av} and V on the surface of the ground. Because of this, there will be abrupt changes first of all in the temperature regime of soils in the warm season (summer). An increase in evaporation will reduce the maximum temperatures and hence the annual amplitudes, since the temperature regime on the ground surface in winter is practically independent of evaporation. The amplitudes are above normal on dry sections, where evaporation is negligible. Consequently there will be only slight changes in the depth of seasonal freezing, on sections with strong evaporation, as compared with dry sections. In the permafrost region, changes in evaporation result in abrupt changes in the depth of seasonal thaw.

These phenomena will occur in the course of economic development of a region on draining the soils and regulating the surface runoff, as well as in the case of changes in the moisture regime of soils due to the flow of groundwater or an unregulated runoff.

Similar phenomena may occur as a result of removal of natural vegetation or in the case of afforestation and similar measures, since this would have a profound effect on evaporation.

Installation of cover (asphalt, concrete) greatly reduces evaporation by preventing the groundwater from reaching the surface.

The structure of the radiation-heat balance is also dependent on the turbulent heat exchange (P), which is the function of the temperature of the ground surface (t), and the coefficient of the heat exchange at the boundary between the topsoil and the atmosphere (K). At higher temperatures of the topsoil, with all other conditions remaining the same, the turbulent heat exchange will be greater than at lower temperatures. Similar relationships occur in the case of different values of K.

There is no direct relationship between changes in the temperature regime of soils and the turbulent heat exchange. This relationship is of a more complex nature and can be detected only on examining the radiation-heat balance as a whole, with allowances for the interaction of all its components. We shall merely mention that with the mean annual temperature of the air remaining constant, the mean annual soil temperature always increases the turbulent exchange. Because of this an increase in the turbulent exchange is related to a decrease in the depth of seasonal freezing and an increase in the depth of seasonal thaw. In a more detailed study, with allowances for all components of the radiation-heat balance, it may be found that this relationship is in fact opposite to what it was supposed to be. It is evident that there may be considerable changes in the turbulent heat exchange, for example, on removing the vegetation cover, or in the case of afforestation, paving, etc.

Changes in the heat exchanges change the soil temperature by a certain value depending on a number of conditions and on the structure of the radiation-heat balance. Calculations show that measures such as drainage will reduce the heat exchanges in the soil and alter the structure of the radiation-heat balance of the surface in such a way that the mean annual temperature of the soil will rise. Similar changes will occur on leveling the ground, excavating, and placing fill. These measures may affect the composition of the soil. its density and its moisture regime. All this will have a direct bearing on the structure of the radiation-heat balance and the temperature of the soil.

It is also essential to consider the nature and the depth of seasonal freezing and thawing and the changes in these factors resulting from human activity. It is an important problem, since most heat exchanges depend on phase transitions of water on freezing and thawing of the soil.

Of great importance is also the temperature shift resulting from human activity, which may be as high as 1°-2°C. The temperature change may seem to be insignificant, and yet close to the southern boundary of permafrost it may lead to abrupt changes in the depths of seasonal freezing and thawing and in annual heat exchanges. The latter had to change the structure of the radiation-heat balance of the soil.

Therefore the study of the structure of the radiation-heat balance within each region provides the only basis for determining the qualitative and quantitative aspects of formation of seasonally frozen soils and permafrost. The most important factor is the study of the relationship between the structure of the radiation-heat balance and the geological and geographical conditions.

ON THE FORMATION OF THE CRYOLITHOZONE IN THE MOUNTAINS OF EASTERN SIBERIA

I. A. NEKRASOV Permafrost Institute, Siberian Branch, Academy of Sciences of the USSR

The beginning of the Quaternary period coincided with a worldwide decrease in the temperature of the atmosphere, the hydrosphere, and the lithosphere. In Europe and the Atlantic Ocean, the temperatures during the glacial ages were at least 8°C lower than at present.[14,18] In the central part of the Scandinavian ice shield, the mean annual temperature of the air dropped to -45°C, and in its marginal parts to -5°C in the west and -25°C in the east.[12] According to K. K. Markov,[7] the mean annual temperature on the glaciers was -50°C.

In eastern Siberia the air temperature at that time dropped to -35°C or -40°C, and even now it may be as low as -17°C or lower (in Artyk, Oimyakon, and Delyankir, for example).

This cooling trend, which lasted at least 600,000 yr, led to a considerable expansion of the earth's cryosphere. If in the Paleogene and even Neogene the cryosphere was entirely within the atmosphere, in the Quaternary it acquired its present form and incorporated both the ionosphere and the cryolithozone, i.e., the soil layers whose temperature is constantly below 0°C.

While considering the formation of the cryolithozone, a distinction is usually made between the distribution of the latter in space, relating it to the present state of the cryolithozone, and the distribution in time, i.e., the development of the cryolithozone is examined from the point of view of paleoclimatology. Since the cryolithozone incorporates soils and rocks of vastly different genetic origin, ages, and lithological composition, and since it is singled out as a separate entity merely on the basis of its temperature, which is always below 0°C, all morphological characteristics of the cryolithozone and its changes in space and time are those of the temperature field of the upper horizons of the lithosphere.

The distribution of the cryolithozone is characterized by the following morphological features: aerial distribution (continuity), vertical thickness, and structure. There is still no generally accepted classification of the cryolithozone, and, therefore, a more detailed differentiation is not yet possible.

Bearing in mind the position of the cryolithozone with respect to the sun, the main source of heat that forms the temperature field of the heliothermozone, we would like to suggest three main types of the cryolithozone: subaerial, subaqueous, and interstitial.

On the basis of its areal development, the cryolithozone of the subaerial type is subdivided into continuous, discontinuous, and sporadic.[8] Recent studies have shown that the main parameters of the subaerial cryolithozone (its areal development, thickness, and temperature at the base of the layer of annual fluctuations) obey the laws of latitudinal and altitudinal supply and redistribution of heat from solar radiation.

The latitudinal zonality of the temperature of the cryolithozone is expressed by the fact that the temperature rises progressively in the southerly direction (for the same elevations). Thus at the latitude of 60°N, the ground surface at sea level receives in the present epoch a considerable amount of heat, so that perennial freezing of soils is no longer possible.

As far as mountain regions are concerned, we see that, as we proceed southward, the cryolithozone at first retreats from the valley bottoms to the slopes and then rises gradually by about 150 m for each degree of latitude.

As is the case in all high, and medium-high, mountain regions of eastern Siberia, the main effect of altitude is a continuous lowering of soil temperatures, with the hypsothermal gradient ranging from 0.8° to 3.5° for each 100 m of altitude. Detailed actinometric observations and measurements of the gradient carried out in the southern part of the Verkhoyansk Range[4] and on the Stanovoi upland[9] have shown that this temperature drop results from an insignificant radiation balance at high altitudes. The radiation balance there is almost 4 times lower than on the lowlands, due to strong radiation in winter resulting from the predominance of anticyclonic weather and a relatively small absorption of heat in the summer owing to the reflection and scattering of radiation by snow. Furthermore, in spite of low air temperatures, evaporation in high mountains is 2 to 3 times as high as at the foot of their slopes, which is due to abundant precipitation and strong winds.

Investigations conducted in low mountain regions[9,16] revealed a somewhat different picture. Here, on slopes of southern exposure at elevations of up to 1,200 m, the soil temperature gradually increases with altitude, so that the cryolithozone is often interrupted by taliks on slopes and even water divides. In some cases the negative hypsothermal gradient reaches 4° per 100 m of altitude. The rise of soil temperature with altitude is due to an uneven distribution of snow, as well as to oversaturation of the valley bottoms in the summer, which leads to considerable cooling as a result of a more intense evaporation. Of great importance is also the heat supplied by atmospheric precipitates seeping through the soil on water divides.

Moreover, some investigators relate the distribution of the cryolithozone on low mountains to the orographic inversion of the air temperature in winter, which sometimes play an important role in the formation of the temperature regime of the cryolithozone in all mountain regions. For example, I. Ya. Baranov[1] states that this inversion is responsible for a "two-story" system of cooling in the mountains, consisting of a lower cooling layer, a separating inversion layer, and an upper layer of cooling. Conditions in the lower layer are favorable for cooling the soils on the valley bottoms, in depressions, and on the lower parts of slopes. A slow rise in temperature with altitude is characteristic of the inversion layer. In the upper cooling layer the temperature drops with altitude in a linear fashion. No inversion is formed in regions within the sphere of oceanic influence. Similar arguments were postulated by P. N. Lugovoi,[6] who substituted "the two-story cooling system" by a "three-layer cooling structure."

However, the data obtained by the author in the course of 20 yr of work in Siberia and by other investigators[2,3,5,13] show that nowhere in eastern Siberia, from the shores of the Arctic Ocean to the Mongolia steppes, are there any "two-story systems" or "three-layer structures." There is only a "lower" (on southern slopes of low mountains) or an "upper" (on medium and high mountains) story of these systems, and in a number of regions where there is a deep inversion of the air temperature, "the upper story" starts at sea level.

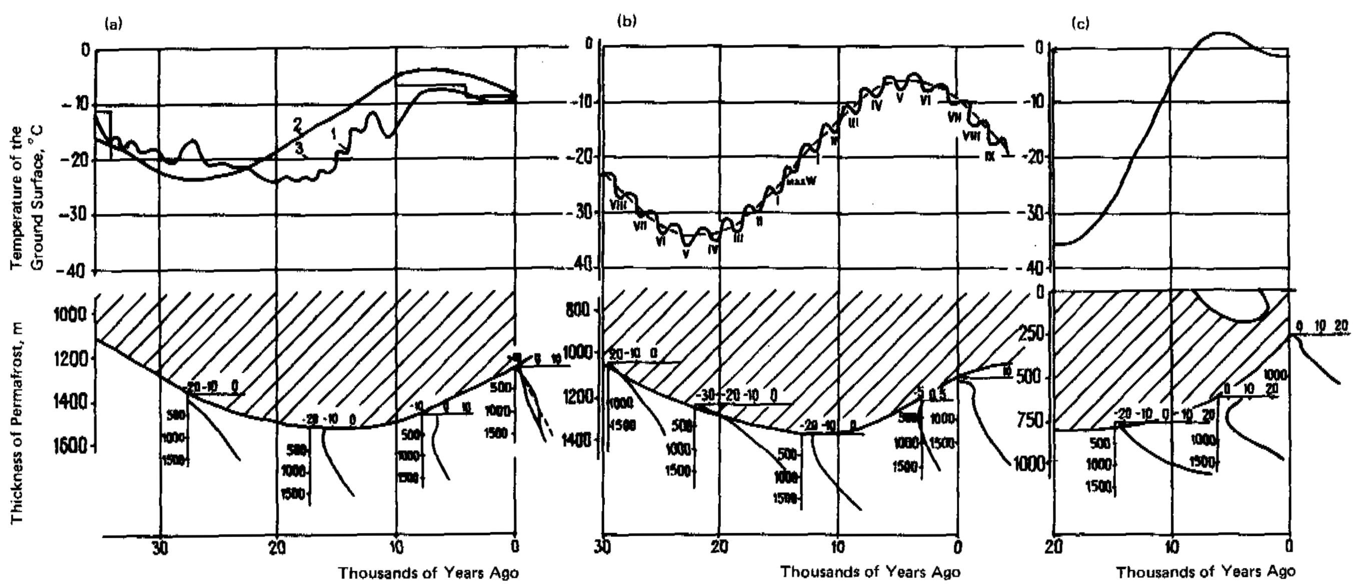

FIGURE 1 Soil temperature on the surface and changes in the thickness of the cryolithozone and the temperature of soils throughout their depth at different times in the upper Pleistocene and the Holocene (calculated data and results of model studies). (a) After Yu. G. Skastkevich;[15] (b) After I. A. Nekrasov *et al.*;[9] (c) After I. A. Nekrasov and G. E. Li.[10] 1, 2 and 3--"reference" marks; I, II, III, etc.--the stages of development and retreat of the last glaciation; max W--the stage of maximum development of glaciers.

The present morphology of the cryolithozone has been largely predetermined by the history of its development, since deep freezing of the upper horizons of the lithosphere has been going on for hundreds and thousands of years. Formation of the cryolithozone in eastern Siberia evidently started in the early Pleistocene, as soon as the mean annual temperature on the soil surface dropped to -3°C or -4°C. In the north this process started earlier than in the south, and on the mountains earlier than on the lowlands.

During the Pleistocene and the Holocene, the development of the cryolithozone in the mountains of eastern Siberia was complicated by repeated glaciations interrupted by relatively warm interglacial periods.

Several attempts were made recently to reconstruct the paleogeocryological picture using analytical methods and models. In all cases the mean annual temperature of the soil surface was reconstructed first, followed by determinations of the temperature field of the upper horizons of the lithosphere.

It is easiest to reconstruct the temperature field of water divides, where the heat exchange is simple and depends on the radiation balance only, since the supply of heat from ground and internal water is at a minimum and no heat is used up on evaporation and phase transformations of water within the soil.

Such approximation of perennial freezing was carried out, for example, for the water divides on the Stanovoi upland.[15] The boundary conditions on the surface are given by the mean annual temperature curve developed by P. Woldstedt[19] for central Europe and adapted by Yu. G. Shastkevich to conditions prevailing on the Stanovoi upland. The present temperature on the water divides is -9°C, while the minimum temperature (during the Sartan glaciation) was -25°C (Figure 1a, curve 1).

Since this curve cannot be described by any equation, the problem was solved by several methods. A good result was obtained by substituting a sinusoidal curve with a period of 40,000 yr and an amplitude of 10° for the Woldsted curve. The initial phase was selected in such a way that the last maximum coincided with the Holocene climatic optimum (Figure 1a, curve 2). Good results were obtained also on assuming that there were abrupt changes in the surface temperature (Figure 1a, curve 3).

Determinations were made of the temperature throughout the depth of the soils at different moments of time and of the thickness of the cryolithozone (Figure 1). It was found that there were considerable changes in the temperature field in different periods. In some epochs the temperature curves at depths of up to 500 m had practically no gradient or even a slight negative gradient. At other times the gradient was greater than the present gradient.

The greatest thickness of the cryolithozone equal to 1,530 m (at given parameters) occurred approximately 12,000-13,000 yr ago, while the last temperature minimum on the surface occurred 20,000 yr ago. This was followed by a gradual decrease in the thickness of the cryolithozone. Hence the maximum thickness occurred some 7,000 or 8,000 yr after the minimum temperature.

Having determined the surface temperature in the upper Pleistocene-Holocene from the sinusoidal curve with a period of 40,000 yr, we repeated the calculations,[9] but this time we incorporated the mesofluctuations with a period of 1,850 yr reflecting the long-term variation of climate.[17] The temperatures used were: the minimum temperature during the Sartan glaciation of -35°C, the optimum temperature of -5°C, and the present-day temperature of -10°C (Figure 1b). The parameters of the cryolithozone calculated in this way were close to those obtained earlier. The surface temperature fluctuations with a period of 1,850 yr and an amplitude of 3°-4° attenuate at depths of 100-200 m.

These calculations refer only to water divides that were represented by nunataks in the glaciation periods. Development of the cryolithozone in the valleys assumed an entirely different course. Here the drop in the soil temperature on water divides was accompanied by a rise in the temperature of the soils buried underneath the ice. In the interglacial periods, ice melting gave rise to streams flowing rapidly along the valleys, and this completely degraded the cryolithozone on the valley bottoms. Therefore, in the valley-troughs, there was, as a rule, a rapid freezing of sediments that had melted as a result of melting of the Sartan glaciers.

It is very difficult to determine the mode of formation of the cryolithozone in the troughs, intermontane basins, and depressions, since the conditions of the heat exchange there were continuously changing (for example, during the entire Quaternary period), due both to downwarping of the bottoms of depressions accompanied by an accumulation of unconsolidated water permeable sediments and to changes in the water content of soils resulting from changes in glaciation on the surrounding mountains. Hence, in contrast to the water divides, formation or degradation of the cryolithozone in such areas took place at a time of considerable environmental changes and at a much slower rate, since freezing and thawing of soils in depressions were accompanied by the aggregate transformations of water.

The results of our model studies of the sandy soils in the Tunkinskaya depression near Lake Baikal[10] clearly show how difficult it is to reconstruct the history of development of the cryolithozone in depressions.

We assumed in our calculations that in the last 20,000 yr the soil temperature in the center of the depression at a depth of 2,000 m remained unchanged at 53°C and on the surface ranged from -35°C to -2.5°C. At the time of the climatic optimum, when oak and elm were growing in depressions, the temperature was 5°C (Figure 1c). We assumed that the problem was one-dimensional and that the soil was of uniform sandy composition with a constant water content of 5 percent. The results showed that the thickness of the cryolithozone within the sandy soils was continuously decreasing and that at the time of the climatic optimum it also retreated from the surface to a depth of over 100 m. Later, during the period of cooling in the late Holocene, this thawed layer became completely frozen once again.

However, on comparing our results with direct observations in the depression it was found that permafrost is now completely absent in the sandy massifs in the Tunkinskaya and similar depressions of the Baikal type. We explain this by the high permeability of sand,[11] for which no allowances were made in our model studies.

All this indicates that the cryolithozone in the mountains of eastern Siberia is still one of the most dynamic natural factors. The cryolithozone and its parameters remain relatively stable under natural conditions if the heat and mass exchange within the system: Atmosphere-soil surface-lithosphere is not interfered with in any way.

However, on developing an area, rapid and substantial changes in the parameters of the cryolithozone may occur, making it necessary to modify development plans. These are the changes that, if not accounted for, will result in numerous deformations or even total destruction of engineering structures in eastern Siberia.

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HEAT EXCHANGE IN THE ACTIVE LAYER

A. V. PAVLOV Institute of Permafrost Studies, Siberian Branch of the Academy of Sciences of the USSR

The thermal regime of the Earth's mantle is studied in the USSR in geological, geographical, hydrometeorological, and geophysical institutes. In particular, the physicogeographical characteristics of external components of the heat balance on a natural surface have been studied in detail by the hydrometeorological services. Investigations of the upper soil layer several meters in thickness, with its annual temperature fluctuations, phase transformations of water, and heat cycle, are of great interest to permafrost scientists.

Such investigations were already in progress in the early stages of Soviet permafrost research. In particular, N. I. Bykov in Skovorodino and P. I. Mel'nikov in Igarka initiated in the thirties long-term studies of formation of the active soil layer.* More detailed observations of the heat cycle in the soil were started in the last few years and have been extended to the surface cover and the atmospheric layer at the surface.

Such studies on sites with different types of cover were first started by the V. A. Obruchev Permafrost Institute of the USSR Academy of Sciences in 1956 in Zagorsk (a region of seasonal freezing), in Vorkuta and Mirnyi (the permafrost region), and a year later in the Suntar-Khayata area. At present such investigations are carried out by the Permafrost Institute, of the Siberian Branch of the USSR Academy of Sciences, at stations in Yakutsk (since 1969) and Igarka (since 1971).

The following experimental sites have been set up in Yakutsk:

I. Natural grass and an undisturbed snow cover (Figure 1),
II. Forest (Figure 2),
III. Stripped (grass is removed in the summer and snow in winter),
IV. Paved,
V, VI, VIII. With [plastic] film (Figure 3) and heat insulating covers,
VII. With board covers,
IX. With controlled water content in the soil.

In Igarka there are experimental sites of types I, II, III, and V. The year-round observations carried out in both localities are summarized in Table 1.

The results of observations will be used in general and applied permafrost studies, since the relationships between the climate, the surface cover, and the heat and water regimes of the soil derived from the data make it possible to compile permafrost maps using the steadily increasing volume of hydrometeorological information and to improve the methods of forecasting and controlling the thermal regime of the active soil layer.

*M. I. Sumgin used the term "active layer" to describe a layer of soil or rock that freezes in winter and thaws in summer.

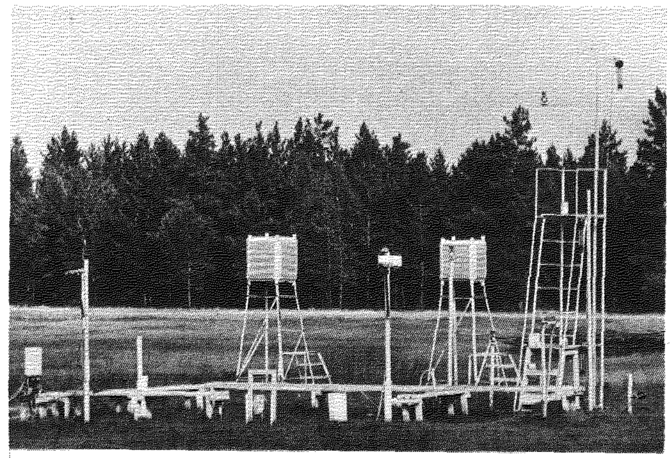

FIGURE 1 Natural site near Yakutsk, August 1972.

These investigations help to determine the annual characteristics of external and internal elements of the heat cycle in the soil in relation to climatic and permafrost conditions. Let us cite some examples based on observations in Yakutsk (2-yr cycle from May 1970 to April 1972), Vorkuta (3-yr cycle from September 1958 to August 1961), and Zagorsk (2-yr cycle from September 1957 to August 1959).

The heat balance of the active layer in Yakutsk is as follows (Table 2). During seasonal thawing from May 3 to October 3, the radiation balance B comprises 47 percent of the total radiation Q. The heat used up on evaporation, LE (E is the intensity and L is the heat of evaporation) and the turbulent [convective] heat exchange, P are the main outgoing components of B. P exceeds LE by 30 percent. The heat, q, used up on heating and thawing of soil, comprises 8.3 percent of B (4.0 percent of Q). In general, during the period of seasonal thaw, P and LE exceed their values in the winter season by 1 order of magnitude. In winter all components of the heat balance are of the same order. The turbulent heat exchange is directed toward the snow surface and is one of the energy sources for evaporation and melting of snow. In Vorkuta, which is at a higher latitude, the ratio q/B increases to 14.3 percent.

The heat balance in the active layer in the region of seasonal freezing (Zagorsk) is charac-

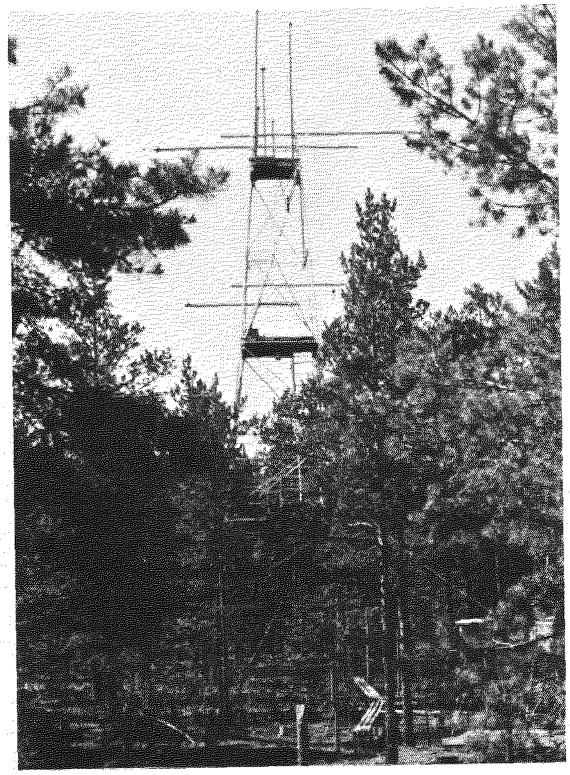

FIGURE 2 Site in a pine forest near Yakutsk, June 1972.

terized first of all by larger amounts of heat used up on evaporation (76 percent of *B* in the summer) and smaller accumulations of heat in the soil (q/B = 3.9 percent). While the albedo of the natural surface in the summer is close to that in Yakutsk (19-21 percent), the ratio of the effective radiation to the total radiation is considerably smaller, since the surface temperature is lower (Table 3).

If the hydrometeorological conditions are close to average long-term values* (Yakutsk, Zagorsk), the difference between the incoming and the outgoing heat in the soil does not exceed 20 percent of the heat exchange. However, if there is a considerable change in hydrometeorological conditions, this difference may be as great as the heat exchange itself. For example, increased snow precipitation during the observation period led to formation of a talik (Table 3). The extreme values of the heat flux through the surface precede the corresponding temperatures by about 1½ months. Due to the delay in the arrival of the heat wave, the fluxes in some months are less at greater depths than at shallower depths (Table 4).

Of all natural covers, snow and vegetation, and especially forest, have everywhere the greatest effect on the thermal regime of soil. The data in Table 3 serve as quantitative illustrations of the reduction in the sum of degree-days of negative temperature on the soil surface as compared with the air temperature in the presence of a snow cover. For example, a 36-cm-thick snow cover in Yakutsk reduced the absolute sum by more than 3,000 degree-days.

Forest cover, which is the main natural landscape in Siberia, considerably changes the proportion of components of the heat balance, as compared with open sites where the meteorological stations are located. Forest growth increases the absorption of radiation throughout the year. This abruptly increases the radiation balance (by several times in the spring and by 50-60 percent in the summer in central Yakutiya) and the external parameters of the heat exchange (*P* and *LE*). The heat is transferred by advection from the surface of the crowns to the open areas. This process is especially pronounced in the spring, which accelerates the melting of snow in the fields. According to observations in Yakutsk, on clear days the heat used up on the evaporation of snow on a clearing surrounded by a pine forest may exceed the radiation balance or even the total radiation. As a result of shading, the heat exchange in the forest soils is reduced by 25-50 percent, while the mean annual temperature is 1.0°C to 2.5°C lower. Therefore, afforestation results in a lowering of the temperature of permafrost.

In engineering geocryology, the heat exchange in the soil is controlled for two reasons: (a) to lower the temperature of permafrost and reduce the depth of seasonal thawing, for example, in the case of hydro or other types of construction, and (b) to protect the soil from freezing and increase its thawing, for example, on developing placer deposits. Detailed studies of the entire combination of different physical processes of heat and mass exchange--which take place in the

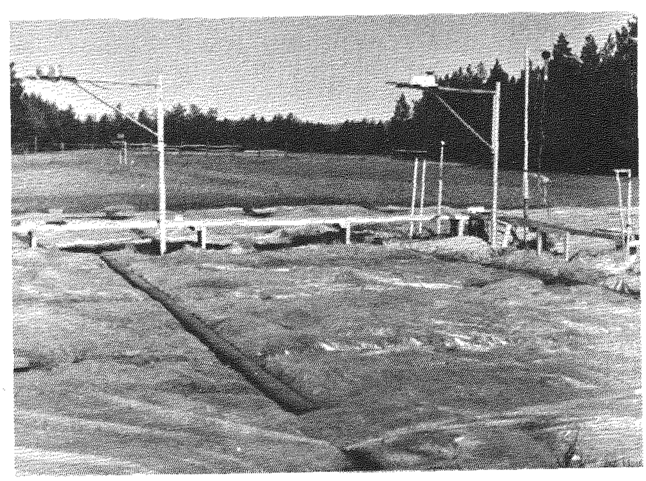

FIGURE 3 Site with a polyethylene film near Yakutsk, June 1972.

*During the periods of observation, the air temperature at all points was close to the average long-term value. The annual precipitation was 224 mm in Yakutsk, 594 mm in Vorkuta, and 682 mm in Zagorsk (11, 31, and 6 percent above normal, respectively). The depth of snow was 36 cm in Yakutsk, 78 cm in Zagorsk, and 77 cm in Vorkuta (20, 13, and 64 percent above normal, respectively). Supes in Yakutsk, suglinok in Vorkuta and Zagorsk.

TABLE 1 Main Observations on Experimental Sites

Observations	Type of site: Yakutsk	Igarka	Equipment
Actinometry			
Total radiation	I,II,III,V,VIII	I,II	Yanishevskii and Kozyrev pyranometers
Reflected radiation	I-IX	I,II,III,V	Yanishevskii and Kozyrev pyranometers
Radiation balance	I-IX	I,II,III,V	Yanishevskii and Kozyrev balancers
Micrometeorology			
Air temperature and humidity	I,II	I,II	Meteorological observation point
Liquid and solid precipitation	I,II,III	I,II,III	Precipitation gauge
Gradients of temperature, air humidity, and wind velocity	I,II	I,II	Psychrometer with remote control, MS-13 anemometer
Evaporation from soil	I,II,III	I,II,III	GR-26 evaporator
Heat and Mass Exchange in the Surface Cover			
Depth, structure, stratigraphy, and density of snow cover	I,II	I,II	Metering rod, lens, densimeter
Evaporation from snow	I,II	I	Plastic evaporator
Thermophysical properties of snow	I	I	Flat probe
Radiation penetration into snow	I	I	Kozyrev pyranometer and pyranometer with a plastic window
Temperature of snow	I,II	I	Thermal set, infrared radiometer
Radiation penetration in a forest	II	II	Yanishevskii pyranometer
Temperature regime of tree stands	II	II	Thermistor
Gradients of temperature, air humidity, and wind velocity in a forest	II	II	Psychrometer with remote control, MS-13 anemometer
Surface temperature of film and insulating covers	V,VIII	V	Thermocouple assembly
Hydrothermal regime of soil			
Soil temperature	I-IX	I,II,III,V	Thermocouple assembly, IR-radiometer, thermal set, thermistors
Water content of soil	I-IX	I,II,III,V	Drilling
Freezing-thawing	I-IX	I,II,III,V	Danilin permafrost gauge
Heat flows	I-IX	I,II,III,V	Plastic thermometer
Thermophysical properties	I,IX	I	Flat probe

TABLE 2 Annual Record of Components of the Heat Balance of Soil in an Open Area, kcal/cm^2

	Months												
Components	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Year
Yakutsk													
Q	0.6	2.7	7.3	10.8	14.5	15.2	15.9	10.3	6.5	3.0	1.2	0.4	88.4
S	0.5	2.1	5.4	6.5	3.8	2.6	2.7	1.9	1.0	1.6	0.9	0.3	29.3
B	-0.5	-0.5	-0.3	2.1	6.5	8.0	7.7	4.8	2.4	-0.0	-0.7	-0.4	29.1
LE	-0.0	0.0	0.2	1.6	1.6	2.8	1.7	2.0	1.5	0.2	0.0	-0.0	11.6
q	-0.4	-0.3	-0.2	0.1	0.9	0.6	0.6	0.3	0.0	-0.5	-0.4	-0.5	0.2
Vorkuta													
Q	0.1	0.8	0.5	8.8	10.4	11.1	14.5	8.9	3.7	1.5	0.2	0.0	64.7
S	0.1	0.5	3.0	5.5	3.0	1.6	2.5	1.4	0.6	0.8	0.1	0.0	19.1
B	-0.7	-0.9	-0.2	0.8	5.1	7.1	9.0	4.6	1.7	-0.6	-0.8	-0.6	24.5
LE	--	--	--	--	--	1.7	2.6	1.7	1.1	--	--	--	--
q	-0.2	-0.2	-0.1	-0.1	0.0	1.6	1.3	0.5	-0.2	-0.3	-0.8	-0.3	1.2
Zagorsk													
Q	0.9	2.6	8.0	10.6	13.2	14.7	15.6	10.6	5.8	2.5	1.2	0.6	86.3
S	0.7	2.0	5.0	3.6	2.5	3.2	3.5	2.3	1.3	0.5	0.4	0.4	25.4
B	-0.8	-0.4	0.2	3.6	6.5	7.2	8.4	5.5	2.5	0.6	-0.8	-0.8	31.7
LE	0.1	0.2	0.6	1.6	4.8	5.3	6.3	4.0	2.2	1.0	0.1	0.1	26.3
q	-0.2	-0.1	-0.2	0.4	0.6	0.6	0.6	0.2	-0.3	-0.4	-0.6	-0.4	0.2

REMARKS: S = reflected radiation; for other symbols, see text.

TABLE 3 Annual Record of Air Temperature (t_A) and Soil Temperature (t_S) at a Depth of 3 m ($t_{3.0}$), °C

Temperature	Months I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Σ(+t) Degree Days	Σ(−t) Degree Days	t_{av}
Yakutsk (t_0 = −2.0°)															
t_A	-45.2	-37.6	-21.4	-5.7	6.5	15.0	21.0	14.0	7.0	-8.5	-28.2	-40.3	1,957	-5,697	-10.2
t_S	-18.3	-18.8	-15.6	-6.8	8.2	18.3	23.9	16.1	6.4	-3.2	-8.7	-13.5	2,239	-2,574	-0.9
$t_{3.0}$	-0.8	-1.6	-3.0	-3.8	-3.9	-3.0	-2.0	-1.3	-1.0	-0.9	-0.8	-0.8	--	-698	-1.9
Vorkuta (t_0 = −0.5°)															
t_A	-18.5	-19.4	-19.5	-11.2	-2.4	6.4	13.2	10.0	3.2	-5.2	-14.3	-18.5	1,007	-3,294	-6.4
t_S	-3.7	-4.0	-3.5	-2.4	-0.2	6.8	14.1	10.8	3.3	-2.8	-3.4	-3.0	1,075	-705	1.0
$t_{3.0}$	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.3	0.2	0.0	0.0	18	0	0.0
Zagorsk (t_0 = 6.7°)															
t_A	-7.1	-6.2	-5.6	4.4	11.5	14.1	17.4	15.1	8.6	3.5	-3.4	-6.7	2,286	-880	3.8
t_S	-0.0	-0.1	0.0	3.5	11.9	15.4	18.5	17.0	9.6	4.6	1.3	-0.2	2,507	-10	6.8
$t_{3.0}$	6.3	5.4	5.0	4.4	4.7	6.0	7.4	9.0	9.8	9.4	8.7	6.8	2,518	0	6.9

REMARKS: t_0--the mean annual temperature of soil at the depth of zero annual heat exchanges.

TABLE 4 Annual Record of Heat Fluxes in the Soil, kcal/cm^2

Depth, m	Months I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Year In-coming	Year Out-going	Year Dif-ference	Year Ampli-tude
Yakutsk																
0.0	-445	-324	-166	147	917	643	644	252	2	-547	-442	-547	2,605	-2,471	134	5,076
0.2	-405	-286	-186	110	515	620	590	282	65	-164	-333	-390	2,182	-1,764	418	3,946
0.5	-265	-204	-150	15	232	371	386	209	82	-54	-178	-236	1,295	-1,087	208	2,382
Vorkuta																
0.0	-205	-198	-168	-125	28	1618	1294	533	-238	-310	-751	-265	3,473	-2,333	1,270	5,806
Zagorsk																
0.0	-230	-179	-174	338	592	545	562	94	-378	-492	-533	-290	2,131	-2,276	-145	4,407
0.1	-234	-180	-174	280	575	523	567	116	-349	-468	-496	-330	2,061	-2,231	-170	4,292
0.5	-232	-217	-208	86	610	528	676	291	-218	-282	-474	-383	2,191	-2,114	77	4,305
1.0	-239	-183	-173	-45	340	386	482	323	-18	-120	-305	-308	1,531	-1,391	140	2,922
2.0	-158	-130	-130	-100	72	182	261	262	115	23	-107	-171	915	-796	119	1,711
3.5	-10	-27	-43	-45	-24	24	59	56	49	26	-16	-35	214	-200	14	411

air layer at the ground surface, the surface cover, and the active layer throughout the year--make it possible to determine the optimum way of controlling the incoming and the outgoing components of the heat exchange in the soil.

The studies have shown that the two simplest and most effective methods of cooling the soil in winter without using artificial sources of cold are periodic removal of snow and construction of decks at a height exceeding that of snow. From the point of view of thermophysics, both of these methods are about equally efficient. They result in mean winter temperatures of the surface some 1°C to 3°C higher than the air temperature and increase the outgoing component of the heat exchange by a factor of 1.5 to 2.5. The same decks are not sufficiently effective in preventing the soil from thawing. For example, the depth of seasonal thawing on site VII in Yakutsk was reduced by merely 25-35 percent, as compared with site I. The investigations have shown that the most effective method of protecting the soil from thawing was the application of foamed plastic covers, which were first used by us at an experimental station in Yakutsk. In the fall of 1972, the depth of seasonal thawing below two layers of PS-4 foamed plastic with an overall thickness of 20 cm was as little as 70 cm, which was almost 3 times less than on a natural site, where it reached 202 cm. Foamed plastic has good radiating and reflecting properties (its albedo is 80-85 percent) and a low coefficient of thermal conductivity. Therefore at night, when the air temperature was above zero, the temperature on the surface of foamed plastic was often -6°C to -8°C. In the daytime, the difference between the temperature of the plastic surface and that on a natural site reached 30°C.

Increased influx of heat into the soil in the summer is achieved by controlled changes in the net radiant heat exchange and by reducing the outgoing components of the heat balance, i.e., the turbulent heat exchange and the heat used up on evaporation of porewater and melting of snow. In this connection studies were made of transparent and opaque synthetic films. It was found that the use of polyethylene films in Yakutsk almost doubled the albedo of the surface and reduced the radiation balance by 5-15 percent. The accumulation of heat in the early summer increased 1.5-2.6 times and then decreased abruptly. The discharge of precipitation accumulated on the film into the soil increased the heat flow still further. A still greater increase in the rate of thaw under a film cover was noted in wet soils.

The investigations described here include a wide range of problems related to the study of the heat exchange in the soil under different natural conditions, the methods of forecasting the thermal regime, and the methods of controlling the latter. The most important problems for future investigation are as follows:

1. Continuation of studies of the heat exchange in the active layer under natural conditions not included in the observations at the meteorological stations, i.e., on the large regions covered by forests (especially earth forests) and in the areas with dissected relief.
2. Determination of relationships between the forest cover, the active layer, and the permafrost; forecasting of changes in geographical and geological conditions (evaporation, water discharge, seasonal freezing, and thawing), which may occur in the case of depletion of forests in Siberia.
3. Studies of the heat balance on slopes of different steepness and aspect.
4. Determination of latitudinal and meridional zonalities of the heat exchange in the soil and subsequent mapping on the territory of the USSR.
5. Improvement of methods of evaluating the elements of the thermal regime of soil for different natural landscapes and for thermal improvements on the surface.

THE TEMPERATURE FIELD IN THE GROUND OF THE NORTHERN PART OF WESTERN SIBERIA AND ITS RELATION TO THE HEAT BALANCE AT THE EARTH'S SURFACE

N. A. SHPOLYANSKAYA   Moscow Lomonosov State University

Apart from general patterns of behavior of the temperature field in the ground of western Siberia, consisting of strictly zonal changes in temperature and thickness of permafrost from north to south, which have been described in the literature,[2,10,11] there are other patterns that manifest themselves on a smaller scale. These may be called local patterns, which concern wide spatial changes in the temperature of permafrost from one type of terrain to another within a single physicogeographical region. The regions may be quite small. The aim of this paper is to investigate these local patterns.

The investigation is based on the analysis of the heat exchange between the ground and the atmosphere. The region under study is bounded by the Kara Sea in the north, the Ural Mountains in the west, the Yenisei River in the east, and the east-west section of the Ob'River in the south.

Our observations and a large amount of factual data obtained from published and unpublished sources show that in the given region the temperature field in the ground varies within a wide range. The patterns of these variations are not the same in different parts of the region and they change from north to south.

For example, according to the exploration data gathered by the "Fundamentproekt" Institute in 1969-1970 in the northeastern part of western Siberia (the area around the Messoyakha-Noril'sk gas pipeline), the distribution of the mean annual temperature of the ground is a function of the topography. The lowest temperatures, ranging from -5°C to -7°C, are characteristic of soils on open watersheds covered with spotted tundra and peat bogs. The soils on the sections of watersheds covered with moss and lichen usually have a slightly higher temperature (-4°C to -6°C). The temperatures on low elements of relief are higher: On bogs covered with sedge and grass, they range from -3°C to -4°C, while in depressions and on the sections of floodplains with shrubs and thin larch forests they vary between 0°C and -2°C.

According to the exploration data collected by the Geological Department of Moscow State University in 1969-1970, the mean annual ground temperature in the northernmost part of the Yamal ranges from -4°C to -8°C and is distributed in space as follows. The lowest temperature, ranging from -6°C to -8°C, occurs on flat surfaces irrespective of altitude, lithological composition, and contours of various plant associations. In depressions, along runoff strips and in the valleys of small creeks, the temperature of permafrost varies between -4°C and -6°C, depending on how well these elements are expressed in the relief.

According to our observations carried out in the lower reaches of the Taz River in 1968, there are marked temperature differences between the open watersheds (-3°C, -4°C) and the bottoms of small valleys and ravines (-1.5°C).

All this indicates that in the northern part of western Siberia the greatest differences in the ground temperatures occur between open elevated sections and protected low-lying sections. Other differences in the physicogeographical elements, including differences in lithology, do not cause temperature changes.

In the more southern regions of western Siberia, the nature of the spatial distribution of the mean annual ground temperature is somewhat different.

According to A. I. Popov,[10] the ground temperature in the region of the Polui River varies not only in the direction from elevated sections to low-lying areas, but also from flat sections to slopes, which is not the case in the northern regions. The aspect of the slope plays an important role here. For example, the ground temperature on forest-covered slopes of northern aspect is lower than the generally uniform zero temperature of flat, forest-covered watersheds.

A similar temperature field was observed in 1970 in the Nydin-Nadym interfluve slightly north of the Pravaya Khetta River (the right tributary of the Nadym). The ground temperature on the watershed covered by a thin larch forest is above zero to a depth of 15 m, while the soils even on gentle northern slopes also covered with forests have a negative temperature.

Here the change of soil temperature with a decrease in elevation is often directly opposite to that observed in the north. In the Polui River basin (according to A. I. Popov), the shallow depressions on plakors* covered with spruce and mixed forests are composed of permafrost, while dry elevated sections with mixed birch and larch forests are free of permafrost.

In the Nydin-Nadym interfluve, permafrost is often found in shallow wet depressions. In deep and narrow gullies and runoff strips, the ground temperature is usually higher than on elevated surfaces.

In contrast to the northern part of western Siberia, there is a distinct relationship between the temperature and the lithology of the soils. The temperature of bog peats is as a rule 1°-2°C lower than that of suglinoks and sands.

Still farther south, in the basins of the Kazym, Vakh, Turukhan, and other rivers, the most noticeable changes occur first of all in the soils of different compositions (as noted by A. I. Popov). The temperature of soils covered with peat is almost always negative, irrespective of their position in the relief. The temperature of purely mineral soils is usually positive.

The water content of soils plays an important role in this part of western Siberia. According to A. I. Popov, the soil temperature of some sections of plakors with a sufficiently wet surface covered with spruce and larch forests and composed of clays and suglinoks is zero. Sandy sections, which are usually dry and are covered with birch forests and lichen, consist of thawed soils with a temperature of 1°C.

This clearly demonstrates the fact that there are considerable differences in the temperature field and hence in the distribution of permafrost in different parts of western Siberia (we have in mind the contemporary permafrost region).

Since the temperature field results from the heat exchange between the ground and the atmosphere, the reasons for this should be sought among the characteristics of the heat exchange at the surface.

CALCULATION OF THE HEAT BALANCE

It follows from the equation suggested by M. I. Budyko[3] that

$$R = LE + P + B. \quad (1)$$

Hence for the cold half-year,

$$B = R - LE - P, \quad (2)$$

and for the warm half-year,

$$B = R - LE - LW - P, \quad (3)$$

*Russian term for well-drained, flat, or slightly undulating interfluves, where soils and vegetation are typical of the region.

TABLE 1 Heat Balance and Its Components, kcal/cm^2·year

		Duration of Period		Warm Period					Cold Period				ΔB
°N	Met. Station	Warm	Cold	R	LE	P	LW	B_W	R	LE	P	B_C	
1	2	3	4	5	6	7	8	9	10	11	12	13	14
73.5	Dikson	V-IX	X-IV	21.3	-10.5	-3.6	-1.7	5.5	-9.1	-1.4	2.2	-8.3	-2.8
72.5	Drovyanoi	Same	Same	19.3	-10.5	-3.4	-1.1	4.3	-9.0	-1.4	1.8	-8.6	-4.4
72.5	Cape Leskin	Same	Same	19.3	-10.5	-1.8	-1.1	5.9	-9.0	-1.4	1.8	-9.2	-3.4
71.5	Kharasovoi	Same	Same	19.4	-10.5	-4.2	-1.1	3.6	-8.7	-1.4	1.4	-8.6	-5.1
71.5	Tambei	Same	Same	19.4	-10.5	-2.6	-1.1	5.2	-8.8	-1.4	0.7	-9.5	-4.4
69.5	Marre-Sale	Same	Same	19.4	-10.5	-4.2	-1.1	3.6	-8.2	-1.4	1.1	-8.5	-5.0
70.0	Se-Yakha	Same	Same	19.4	-10.5	-4.2	-1.1	3.6	-8.8	-1.4	0.4	-9.8	-6.2
70.0	Tadibe-Yakha	Same	Same	19.5	-10.5	-3.6	-1.1	4.3	-8.3	-1.4	0.4	-9.3	-5.1
68.5	Cape Kamennyi	Same	Same	22.9	-13.2	-3.1	-1.1	5.5	-7.7	-1.8	0.7	-8.8	-3.4
67.5	Novyi Port	Same	Same	23.4	-13.2	-4.2	-1.1	4.9	-7.7	-1.8	0.4	-9.1	-4.3
67.5	Tazovskoe	Same	Same	27.5	-15.9	-2.6	-1.3	7.6	-6.7	-1.9	0.4	-8.2	-0.6
67.0	Yambura	IV-IX	X-III	27.1	-16.7	-3.4	-1.0	6.0	-4.9	-1.3	1.2	-5.0	+1.0
67.0	Yar-Sale	IV-IX	X-III	26.7	-16.7	-4.0	-1.0	5.0	-4.9	-1.3	1.2	-5.0	0.0
66.5	Nyda	V-IX	X-IV	26.8	-15.9	-2.6	-1.0	7.2	-4.9	-2.2	0.7	-6.4	+0.6
66.5	Sidorovsky	Same	Same	27.4	-15.9	-2.1	-1.7	7.6	-7.5	-2.2	2.5	-7.2	+0.4
67.5	Igarka	Same	Same	29.2	-15.8	-3.9	-1.7	7.8	-6.3	-2.2	3.3	-5.2	+2.6
66.5	Salekhard	IV-IX	X-III	30.9	-16.7	-5.3	-1.0	7.9	-4.9	-1.3	0.9	-5.3	2.6
65.5	Pitlyar	Same	Same	27.6	-17.8	-3.4	-1.0	5.4	-6.3	-1.4	3.5	-4.2	1.2
65.0	Polui	Same	Same	27.6	-17.8	-1.2	-1.0	7.5	-6.3	-1.4	2.8	-4.9	2.0
65.5	Urengoi	V-IX	X-IV	30.0	-16.9	-3.4	-1.7	8.0	-6.7	-2.3	3.9	-5.1	2.0
65.5	Krasnosel'kup	Same	Same	30.4	-17.1	-3.1	-1.7	8.5	-6.8	-2.1	1.8	-7.1	1.4
64.5	Yanov Stan	Same	Same	30.4	-18.6	-2.6	-1.7	7.5	-6.8	-2.3	4.7	-4.4	3.1
66.0	Turukhansk	Same	Same	29.1	-15.8	-3.6	-1.7	8.0	-6.1	-2.0	3.9	-4.2	3.8
65.5	Muzhi	IV-IX	X-III	27.6	-18.7	-1.6	-1.0	6.3	-6.3	-1.2	1.3	-6.2	0.1
65.5	Nadym	V-IX	X-IV	30.0	-17.1	-3.4	-1.3	6.3	-6.7	-2.1	4.7	-4.1	2.2
65.0	Tarko-Sale	V-IX	X-IV	30.0	-17.6	-4.2	-1.5	6.7	-6.7	-2.2	3.6	-5.3	1.4
64.0	Tol'ka	IV-IX	X-III	30.8	-20.9	-2.2	-1.7	6.0	-5.5	-1.2	5.6	-1.1	4.2
64.0	Saranpaul	Same	Same	30.8	-19.7	-1.9	-1.5	7.7	-5.5	-1.3	5.0	-1.8	5.9
64.0	Sos'vinskaya	Same	Same	30.8	-19.7	-3.5	-1.5	6.1	-5.5	-1.3	4.7	-2.1	4.0
63.5	Khalesovoi	Same	Same	30.8	-20.9	-1.6	-1.4	6.2	-5.5	-1.3	3.5	-3.3	3.8
63.0	Igrim	Same	Same	31.2	-20.9	-4.1	-1.2	5.0	-4.5	-1.3	4.4	-1.4	3.7
63.5	Berezovo	Same	Same	31.2	-20.9	-2.8	-1.2	6.3	-4.5	-1.3	2.8	-3.0	3.4
63.5	Kazym	Same	Same	31.2	-20.9	-3.5	-1.2	5.6	-4.5	-1.3	4.7	-1.1	4.6
63.0	Numto	Same	Same	30.8	-20.9	-3.8	-1.2	4.9	-5.5	-1.3	1.9	-4.9	0.0
62.5	Kellog	Same	Same	30.8	-20.9	-0.9	-1.4	7.6	-5.5	-1.3	4.7	-2.1	5.7
62.5	Nyaksimvol	Same	Same	31.2	-21.2	-4.1	-1.4	4.5	-4.5	-1.6	4.4	-1.7	3.0
62.5	Oktyabr'skoe	Same	Same	31.5	-21.2	-3.1	-1.4	5.8	-4.2	-1.6	2.5	-3.2	2.8
62.0	Cape Sosnovyi	Same	Same	31.5	-21.7	-1.9	-1.4	6.5	-4.2	-1.6	4.4	-1.3	5.4
61.5	Khantymansiisk	Same	Same	31.5	-21.7	-4.1	-1.2	4.5	-4.2	-1.6	2.2	-3.6	1.0
62.0	Gorshkovo	Same	Same	31.5	-21.7	-3.5	-1.2	5.1	-4.2	-1.6	4.1	-1.7	3.5
61.5	Sytomino	Same	Same	31.5	-22.3	-5.0	-1.2	3.0	-4.2	-1.7	3.8	-2.1	1.0
61.5	Surgut	Same	Same	31.5	-22.3	-3.1	-1.2	4.9	-4.2	-1.7	2.5	-3.4	1.6
62.0	Ermakovo	Same	Same	31.5	-22.3	-2.5	-1.2	5.5	-4.2	-1.7	4.4	-1.5	4.1
62.0	Var-Egan	Same	Same	31.5	-22.3	-3.5	-1.2	4.5	-4.2	-1.7	3.5	-2.4	2.2
61.5	Lar'yak	Same	Same	31.5	-22.3	-4.4	-1.2	3.6	-4.2	-1.7	4.4	-1.5	2.2
60.5	Yartsevo	Same	Same	31.5	-22.3	-4.4	-1.2	3.6	-4.2	-1.7	4.1	-1.8	1.9

where R is the radiation balance at the active surface, LE is the heat loss due to evaporation, LW is the heat loss due to the melting of snow, P is the turbulent heat exchange with the atmosphere, and B is the heat exchange with the soil. We have calculated the heat balance on the surface at 46 points that are fairly uniformly distributed over western Siberia. Our aim was to determine B for two half-yearly periods: the warm period (the heat flux is directed into the soil), and the cold period (the heat flows from the soil into the atmosphere).* The average long-term values of components of Equations (2) and (3) were taken from the literature,[1,3,5,6,7,9] or were calculated by methods used at the Main Geophysical Observatory. B was determined as a remainder term of the equation. It was assumed that the patterns of the distribution of B in time and space would explain the main relation-

*The warm period includes the months for which $R > 0$; the cold period, the months with $R < 0$.

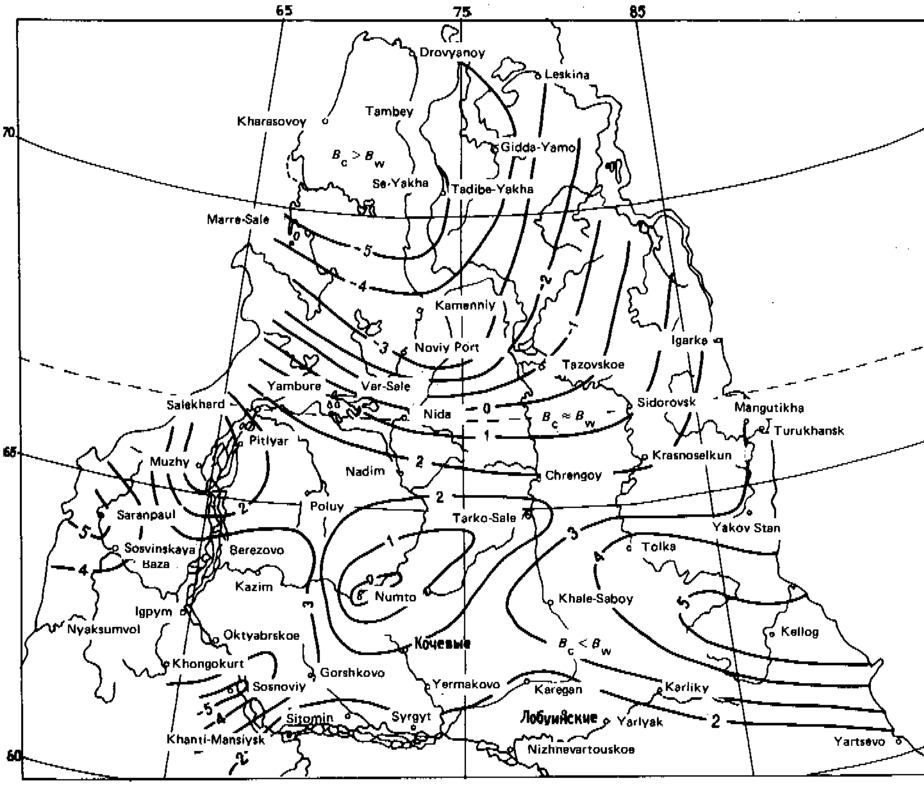

FIGURE 1 Distribution of ΔB in western Siberia.

ships of the temperature field of soils in western Siberia.

The determined components of the heat balance are given in Table 1.

ANALYSIS OF RESULTS

As indicated by Table 1, the common assumption that the inflow of heat into the soil in the warm period (B_W) is practically equal to the outflow of heat in the cold period (B_C), which was probably based on observations in the temperate zone, cannot be applied in the region under investigation.

The fact that $\Delta B \neq 0$ was noted previously by A. V. Pavlov[8] in Vorkuta, N. S. Ivanov[4] in Yakutsk, and at Houghton [Michigan] in the Northern Hemisphere. We would like to stress that, at this stage of the investigations, the qualitative side of the problem is the most important. As far as the quantitative aspects are concerned, we cannot claim that our values of ΔB are accurate, bearing in mind the inaccuracies in the method of calculation of various components of the heat balance and especially P. However, we must not ignore the distinct qualitative patterns in the spatial distribution of ΔB (there is not a single deviation from these regular patterns in the values of ΔB). To analyze these patterns, we used the computed ratio of B_C to B_W.

The map of the ΔB isolines (see Figure 1) shows that three regions can be singled out in western Siberia: the northern region to about 67°N in the west and 68°-69°N in the east, where the heat exchange in winter is considerably higher than in the summer ($B_C > B_W$); the central region, which forms a narrow strip along the Arctic Circle and where we may assume that in the first approximation $B_C \approx B_W$; and the southern region, which encompasses the remaining part of the permafrost region and where $B_C < B_W$.

The isolines of the heat flux in the northern part of western Siberia stretch in a sublatitudinal direction and roughly from north to south in the eastern part (near the Yenisei River). This zonality is disturbed south of the Arctic Circle, where the isolines form closed loops, and these are found in the entire southern part of the given region. The zonality is evidently restored in the extreme south, in the area around the east-west section of the Ob.

The spatial distribution of B explains many factors related to the temperature field in the ground of western Siberia, as well as a number of general physicogeographical relationships.

First of all, the ratio of B_C to B_W described earlier shows that formation of the temperature field in different parts of western Siberia is a function of various factors.

In the north, formation of the temperature field is affected by the thermal regime in winter, and, therefore, the spatial distribution of the ground temperature is related to the winter factors of the heat exchange. The snow cover with its heat-insulating properties is one of the important factors that vary greatly from place to place. The spatial distribution of the ground temperature in the northern parts of western Siberia is a function of the spatial distribution of the snow cover. Elevated open sections have the lowest ground temperature, while low-lying protected sections have the highest. The specific composition of vegetation is of importance only if it affects the density and the height of the vegetation cover. In this case, it controls the accumulation of snow and, indirectly, the ground temperature.

In the given region, the summer factors of the heat exchange have little effect on the ground temperature changes from place to place. The most important among them, i.e., direct radiation and heat losses due to evaporation, are small: The first because of an abundance of clouds; the second because of little evaporation due to low radiation. Therefore, they cannot cause any temperature changes from slope to slope and from wet sections to drier areas. There are no such differences in the region. Since the differences in the water content of soils are not sufficient to change their temperature, the soil lithology, which greatly affects the water content, has practically no effect on the temperature changes either. This is precisely the reason why, with all other conditions being equal, the mineral soils and the peat bogs in the northern part of western Siberia have the same temperature.

In the central part of the region, where the heat fluxes into the ground in winter and summer are approximately the same, the ground temperature depends on both the winter and the summer factors of the heat exchange.

As before, snow is the most important winter factor and, therefore, the soils in deep ravines and tree-covered areas have the highest temperature.

Moreover, the more important role of the heat balance in the summer frequently disturbs the relationship between the temperature field and the spatial distribution of snow characteristics of the northern part of the region.

Fewer clouds than in the north and fairly extensive radiation in the warm period result in noticeable differences in the amount of heat supplied to slopes of different aspect and steepness, which under conditions of dissected relief leads to changes in the ground temperature, depending on the slope and orientation of the surface. Therefore, most slopes of northern exposition consist of permafrost, which is not the case on the southern slopes and horizontal sections where the ground temperature is higher or is even positive.

Relatively large amounts of summer heat bring about stronger evaporation, and, as a result of this, differences in the water content of soils lead to different heat losses. This is the reason why wet sections (wet forests and water-logged shallow depressions) often have a lower ground temperature than drier places and sometimes rest on permafrost.

This also increases the importance of the lithology of soils, which is closely related to their water content. The sands, which on plakors have the least water content, have a higher temperature than clays and suglinoks, which in most cases have a high water content. Peaty soils, which have the highest water content, have the lowest temperatures (providing there is no constant water reflector on their surface).

In the southern part of the given region, the temperature field is affected mainly by the heat regime in the summer. Therefore, the main differences in the ground temperature depend on the position of a given section in the relief (the radiation input is sufficiently large and the differences due to the slope and orientation of the surface are considerable) and the water content of the surface layers (and hence the lithology of soils).

The above observations have confirmed these relationships, especially with respect to the latter factor. In this part of western Siberia, peaty soils and often clays and suglinoks with sufficiently high water content are frozen both in depressions and on plains. The sand composing dry elevated sections of watersheds are unfrozen and have the highest temperature.

The spatial distribution of ΔB shown in Figure 1 permits us to assume that accumulation of permafrost is taking place in the norther part, that it is evidently relatively stable in the central part, and that it is undergoing degradation in the south.

The general physicogeographical relationships which govern the B_C/B_W ratio are as follows.

The above zonality in the distribution of B is the cause of the natural zonality in western Siberia, including the permafrost zonality. Many natural boundaries, including the ground isotherms, coincide with the ΔB isolines.

The boundaries of vegetation zones evidently depend on the spatial ratio of B_C to B_W. Since vegetation depends mainly on the thermal regime in the summer, the changes in vegetation zones occur in places with a considerable increase in B_W as compared with B_C. For example, the boundary between the tundra and the forest-tundra coincides with the boundary between the areas with $B_C > B_W$ and $B_C \approx B_W$. The boundary between the forest-tundra and the taiga coincides with that between the areas with $B_C \approx B_W$ and $B_C < B_W$. From this it follows that the latitudinal extent of the forest tundra zone is just as small as that of the area with $B_C \approx B_W$.

The B_C/B_W ratio determines yet another pattern in the distribution of vegetation. The most abundant vegetation in the northern parts of western Siberia is found predominantly in ravines and in the valleys of small rivers (which, as a rule, are covered with shrubs or trees). As we proceed southward, the forests disappear in the valleys and appear on watersheds and their slopes. This is evidently due to the fact that at $B_C > B_W$ in the north the microclimate is warmest in valleys and ravines--the sites of snow accumulation. In the more southern part of the region, where $B_C < B_W$, the higher water content of soils on the bottoms of valleys and ravines, and hence considerable heat losses due to evaporation, lead to a more severe microclimate than on watersheds.

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CLIMATIC (HEAT BALANCE) FACTORS GOVERNING FORMATION OF PERMAFROST IN THE MONGOLIAN PEOPLE'S REPUBLIC

M. K. GAVRILOVA Permafrost Institute, Siberian Division, Academy of Sciences of the USSR. Joint Soviet-Mongolian Geological Expedition

Climate is one of the main factors governing formation of permafrost. Although persistent freezing of the Earth's crust has been developing over thousands of years, its present development is largely dependent on the contemporary climate. The effect of contemporary climatic conditions is especially noticeable along the southern boundary of the permafrost region, where the thickness of permafrost is not very great.

The formation of permafrost depends on the ratio of the climatic factors of the cold and warm seasons, and particularly their heat exchange characteristics, or the ratio of winter freezing to summer thawing of soils.

The contemporary climate of Mongolia is the result of its geographical position. The relatively southern latitudes (43°-52°N) ensure a fairly large supply of heat from solar radiation, especially in summer. The fact that Mongolia is far removed from the oceans results in abrupt temperature changes and a considerable dryness of the air, which is especially pronounced in winter. The mountainous relief in the northern part of the country and in some southern regions creates a number of mesoclimatic and a variety of microclimatic conditions.

Table 1 shows some data from a number of meteorological stations in Mongolia. The main climatic and heat balance characteristics of freezing and thawing seasons are given in Tables 2 and 3. The data are based on meteorological measurements and calculations (the heat-balance characteristics).

The freezing season in Mongolia lasts from 5 months in the south to 7 months in northern and mountainous regions. At high altitudes the soils begin to freeze approximately in early October, and at stations below 1,500 m in northern and central regions, at the end of October. Freezing stops in southern Mongolia in early April, in central and northern regions in the middle of the month, and in high mountainous areas at the end of April.

The surface freezing index is: 1,500°-2,500°C in the south, 2,500°-3,500°C in central and northern regions, and up to 3,700°C in the northwestern regions and at high altitudes. Total radiation: from 30 kcal/cm^2 per season in the north to 40 kcal/cm^2 in the south. Radiation balance: from -1 to -4 kcal/cm^2 per season in western, northern, and eastern regions, and from 0 to +3 kcal/cm^2 per season in the southern parts of central regions and in the south. The heat flux to the soil is negative everywhere and ranges from -3.0 to -4.5 kcal/cm^2 per season.

Intensive cooling of the soils occurs throughout Mongolia in winter when temperatures are low and there is little snow on the ground. In February the average soil temperature at a depth of 1 m ranges from -11°C to -7°C in the central part and from -6.0°C to -3.5°C in the south. The soil freezes everywhere, including the Gobi Desert. The maximum depth of freezing ranges from 5 m in the north to 1.5-2.0 m in the south.

The thawing season starts in early April in central and northern regions and in late April at high altitudes. Thawing stops in early or late October. Therefore, the length of the thawing season is 5 months in northern regions, 6 months in central parts, and 7 months in the south.

The surface thawing index comes to 2,000°C at high altitudes, 2,500°-3,200°C in northern and central regions, and 3,500°-3,800°C in the south. The total radiation ranges from 78 kcal/cm^2 per season in the north to 104 kcal/cm^2 in the south. The radiation balance is 30 kcal/cm^2 per season in the north, 35 kcal/cm^2 in central regions and 40 kcal/cm^2 in the south. The heat flux to the soil is positive everywhere: 3.0-3.5 kcal/cm^2 in western, northern and central regions; 3.5-4.0 kcal/cm^2 in the east, and 4.5-5.0 kcal/cm^2 in the south.

On receiving such amounts of heat (as indicated by meteorological data), the soils thaw everywhere in the summer. However, it should be remembered that the meteorological stations in Mongolia are located mainly on the plains or in relatively open river valleys. The climatic characteristics of these locations depend on macroclimatic factors.

Permafrost in Mongolia is usually found in mountainous regions, i.e., at high altitudes and under conditions of complex relief. Retention of soils in a frozen state in these areas is due not so much to climatic conditions in winter (it is warmer in the mountains in winter and there is less snow), than to conditions prevailing in the summer (general decrease in temperature with altitude and the effect of shade). For example, the summer radiation balance in a narrow tree-covered gorge having a northern exposure decreases 2 to 4 times as compared with a wide valley, and the heat flux to the soil is about 1 kcal/cm^2 per season. This is obviously not enough to heat the frozen soils.

In the formation of a climate it is essential to distinguish between macrofactors, mesofactors, and microfactors. The macrofactors are the latitude (which determines the total heat supply from solar radiation) and the location with respect to the oceans (which determines the continental nature of the climate related to the general atmospheric circulation). The mesofactors are the altitude above sea level and large relief forms. The microfactors are small, low hills; convex and concave forms of local relief; vegetation; soil cover; mineral composition of soil; amount of moisture; human activity; and so forth. Such differentiations in explanations of the formation of permafrost in relation to climate are important in permafrost studies in general and especially so in Mongolia.

The macroclimatic factors of the winter season in Mongolia result in freezing of soils to great depths everywhere in the country. However, the same factors are responsible for complete thawing of frozen soils in the summer. The presence of permafrost in Mongolia is related to the effects of mesoclimatic and microclimatic factors.

TABLE 1 Some Data on the Meteorological Stations in Mongolia

Station	Absolute Altitude, m	Soils
Ulangom	936	Clay
Kobdo	1,500	Detrital sand
Muren	1,281	Supes
Bulgan	1,210	Suglinok to 40 cm, then detrital sand
Ulan-Bator	1,266	Detrital suglinok
Choibalsan	750	Detrital sand to 40 cm, then clay
Altai	2,170	Detrital sand
Galut	2,160	Detrital supes
Choir	1,425	Detrital supes
Sain-Shand	952	Detrital supes
Zamyn-Ude	961	Sand
Dalan-Dzadgad	1,470	Sand

TABLE 2 Average Characteristics of the Freezing Season

Meteorological Station	Start of Freezing (October)	End of Freezing	Duration of Freezing Season, Days	Average Surface Temperature in the Coldest Month, °C	Surface Freezing Index, °C	Average Thickness of Snow Cover (One Winter), cm	Seasonal Totals: Total Radiation, kcal/cm^2	Seasonal Totals: Radiation Balance, kcal/cm^2	Seasonal Totals: Heat Flux to the Soil, kcal/cm^2	Mean Soil Temperature at a Depth of 1 m in the Coldest Month, °C	Depth of Seasonal Freezing, cm
Ulangom	18	16/IV	181	34	3,690	10	37.1	2.6	4.6	--	--
Kobdo	19	3/IV	167	27	2,587	1	35.1	1.1	4.0	--	291
Muren	16	11/IV	178	26	2,812	2	32.1	1.7	3.7	--	>320
Bulgan	14	10/IV	179	24	2,792	4	38.9	3.9	4.2	6.9	>320
Ulan-Bator	11	13/IV	185	29	3,203	3	40.5	2.7	3.8	8.6	>320
Choibalsan	22	6/IV	167	22	2,414	1	36.5	2.3	4.2	7.9	278
Altai	10	16/IV	189	23	2,548	--	44.1	2.6	3.2	--	--
Galut	2	20/IV	200	29	3,635	--	50.6	2.5	4.0	--	--
Choir	22	8/IV	169	22	2,350	--	36.8	0.6	3.8	8.6	>320
Sain-Shand	29	26/III	149	20	1,969	1	36.0	1.8	4.7	6.3	237
Zamyn-Ude	29	26/III	149	20	1,999	--	36.1	0.3	4.8	7.0	264
Dalan-Dzadgad	29	25/III	148	18	1,791	3	34.1	0.8	4.4	3.5	177

TABLE 3 Average Characteristics of the Thawing Season

Meteorological Station	Start of Thawing	End of Thawing (October)	Duration of Thawing Season, Days	Average Surface Temperature in the Warmest Month, °C	Seasonal Totals: Surface Thawing Index, °C	Seasonal Totals: Total Radiation, kcal/cm^2	Seasonal Totals: Radiation Balance, kcal/cm^2	Seasonal Totals: Heat Flux to Soil, kcal/cm^2	Mean Soil Temperature at a Depth of 1 m in the Warmest Month, °C	Depth of Seasonal Thawing
Ulangom	16/IV	18	184	24	3,107	78.8	29.8	4.3	--	--
Kobdo	3/IV	19	198	25	3,203	89.0	31.3	3.6	15.7	Complete
Muren	11/IV	16	187	23	2,920	83.2	34.4	3.4	--	Complete
Bulgan	10/IV	14	186	20	2,451	87.7	35.0	3.1	10.6	Complete
Ulan-Bator	13/IV	11	180	21	2,625	78.1	29.3	3.7	9.6	Complete
Choibalsan	6/IV	22	198	26	3,241	87.8	37.8	3.7	14.9	Complete
Altai	16/IV	10	176	19	2,230	78.1	26.8	2.5	--	--
Galut	20/IV	2	165	17	2,043	85.5	35.9	3.1	--	--
Choir	8/IV	22	196	22	2,891	89.3	36.8	3.2	15.0	Complete
Sain-Shand	26/III	29	216	28	3,810	103.6	38.6	4.9	18.9	Complete
Zamyn-Ude	26/III	29	216	27	3,788	98.8	39.9	4.9	19.1	Complete
Dalan-Dzadgad	25/III	29	217	25	3,510	99.2	38.6	4.3	18.8	Complete

TRANSVERSE DISSIPATION OF HEAT IN SATURATED PERMEABLE SOILS

V. E. KAPRANOV AND G. Z. PERL'SHTEIN Eastern Research Institute of Gold and Rare Minerals, Magadan

In the course of the movement of liquids and gases in fine-grained media, the elementary flows, on moving around the grains, deviate from the general direction of the main flow. If the temperature or the concentration of the liquid vary along the normal to the vector of the seepage rate, this will result in a transverse dissipation of heat (matter). This phenomenon is usually regarded as a conductive process by considering the role of convection with the help of the coefficient of effective transverse thermal conductivity λ_e.

Several authors[1-3] represented the dependence of λ_e on the seepage rate (V_f) as follows:

$$\lambda_e = \lambda_0 + DV_f \tag{1}$$

where λ_0 is the coefficient of thermal conductivity of the medium at $V_f = 0$; D is the coefficient of transverse dissipation of heat, which is directly proportional to the size of the pores.

On explaining the reasons for the increase in λ_e and deducing Equation (1), no allowances were made for the thermal conductivity of the liquid and the heat exchange between the flows moving in opposite directions in the adjacent transverse channels. This resulted in increased values of λ_e and D, especially with small values of V_f.

These shortcomings can be avoided by making the following assumptions:

1. The coefficient of thermal conductivity of a medium is expressed as a sum of the thermal conductivities of its components:

$$\lambda_e = n\lambda_K + (1 - n)\lambda_s, \tag{2}$$

where n is the porosity, λ_K is the effective thermal conductivity of the liquid in transverse channels,

$$\lambda_K = \frac{q_l}{T_0 - T_l},$$

q is the resulting heat flux in the channels, λ_s is the thermal conductivity of that part of the medium where there are no transverse flows, and l is the distance between the T_0 and T_1 isotherms equal to the length of the transverse flows.

2. The real trajectories of liquid particles are replaced by two components--along and across the flow (Figure 1).

3. The temperature fields of the elementary transverse flows are one-dimensional.

4. The heat flux between the adjacent channels 1 and 2 at any point x depends only on the difference between the temperatures $T_1(x)$ and $T_2(x)$ and the thermal resistance of the layer h.

In the case of a steady-state regime, the water temperature in adjacent channels is described by the following system of equations:

$$\left.\begin{aligned} \lambda_w S \frac{\partial^2 T_1}{\partial x^2} - c_w VS \frac{\partial T_1}{\partial x} - \frac{\lambda\tau}{h}(T_1 - T_2) = 0 \\ \lambda_w S \frac{\partial^2 T_2}{\partial x^2} + c_w VS \frac{\partial T_2}{\partial x} + \frac{\lambda\tau}{h}(T_1 - T_2) = 0 \end{aligned}\right\} \tag{3}$$

with the boundary conditions

$$T_1(0) = T_2(0) = T_0, \quad T_1(l) = T_2(l) = T_l, \tag{3a}$$

where λ_w and c_w are the thermal conductivity and the [specific] heat capacity of water respectively, λ_T is the thermal conductivity of solid components of the medium, $2 \cdot S$ is the width of the flows, and h is the equivalent distance between adjacent flows.

Substituting into Equation (2) the value of q corresponding to the solution of Equation (3), we obtain

$$\lambda_e = (1 - n)\lambda_s + n\lambda_w \frac{\lambda\tau\lambda_w \tfrac{1}{2} c_w^2 V^2 hS}{\lambda\tau\lambda_w + c_w^2 V^2 hS \frac{\mathrm{th}\,\frac{1}{2}pl}{pl}}, \tag{4}$$

where

$$p = \sqrt{\frac{2\lambda\tau\lambda_w + c_w^2 V^2 hS}{\lambda_w^2 hS}}; \qquad V = \frac{V_f}{2n}. \tag{4a}$$

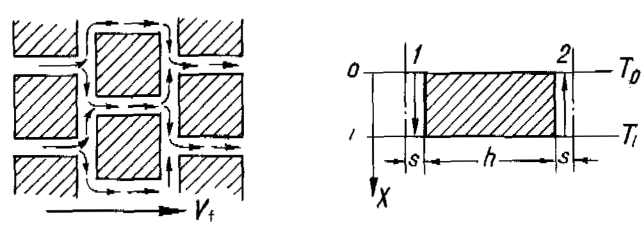

FIGURE 1 Explanations for Equation (4).

Most experiments[1,4] were carried out on materials used in the chemical industry and in turbulent regimes of movement of liquids and gases.

The coefficients of effective thermal conductivity of permeable soils were first determined by V. G. Gol'dtman, who obtained $D = 3.08 \times 10^4$ J/m^2·K. S. D. Chistopol'skii investigated two types of soils and obtained D equal to 1.01×10^4 and 0.45×10^4 J/m^2·K.

Studies of the transverse dissipation of heat were started again in the permafrost-hydrogeological laboratory of VNII-I in 1969. We studied the heat transfer from a seepage flow cooled from the side of the section under investigation. The first series of experiments was carried out in a rectangular trough. A 2- to 4-cm-thick crust of ice and soil was maintained on the sides of the specimens to cool the flow. In the second series, conducted in a cylindrical trough, cooling was achieved by rapid circulation of cold water (2°-3°C) in the external jacket of the cylinder. Altogether we carried out over 100 experiments.

In both series a steady-state temperature regime was established. Hence, ignoring the molecular transfer of heat in the direction of seepage, the problem of heat exchange could be reduced to the well-known problem of cooling a plate or a cylinder of infinite length.

Realization of boundary conditions in devices where no ice-soil crust was formed was investigated both experimentally and theoretically. It was shown that the temperature of the soil surface may be taken as equal to that of the cooling liquid.

It was observed in many experiments that there were abrupt temperature changes in the direction away from the cooling surface and towards the axial part of the specimen. Because of this, the calculations were based on the total heat transfer of the flow.

We determined λ_e for homogeneous gravel soils with particle diameters of 2-3, 3-5, and 5-7 mm and seepage velocities ranging from 0 to 5 m/h (Re < 10). There was good agreement between the results of both experimental series and only a slight scatter of experimental points.

There was no clearly defined relationship between λ_e and the particle diameter (Figure 2). The increase in λ_e with the seepage rate was found to be negligible. At seepage rates of 4 to 5 m/h, λ_e ranged from 1.63 to 2.09 W/m·K (Figure 2), while λ_0 ranged from 1.57 to 1.65 W/m·K.

We also investigated the effective thermal conductivity of detrital soil (particle diameter 15-35 mm), the pores of which were filled with fine (2-3mm) gravel. In the absence of the filler, λ_e ranged from 3.49 (V_f = 0.6 m/h) to 7.84 W/m·K (V_f = 3.0 m/h). In the case of complete filling of pores, λ_e was practically equal to that of homogeneous gravelly soils.

There are considerable discrepancies between these results and the data of V. G. Gol'dtman. In our opinion his value of D is too high because of (a) inadequate size of the model for the coarse-grained soil used, which should produce a hydraulic wall effect and increase the heat transfer of the flow, and (b) calculation of λ_e from temperature measurements at individual points, which led to a very considerable scattering of factual data.

We processed the experimental data of S. D. Chistopol'skii with allowances for the boundary conditions in the apparatus used. The "new" treatment produced values of λ_e, which were close to our data and were practically independent of V_f. Furthermore, the scattering of experimental points was greatly reduced (Figure 2).

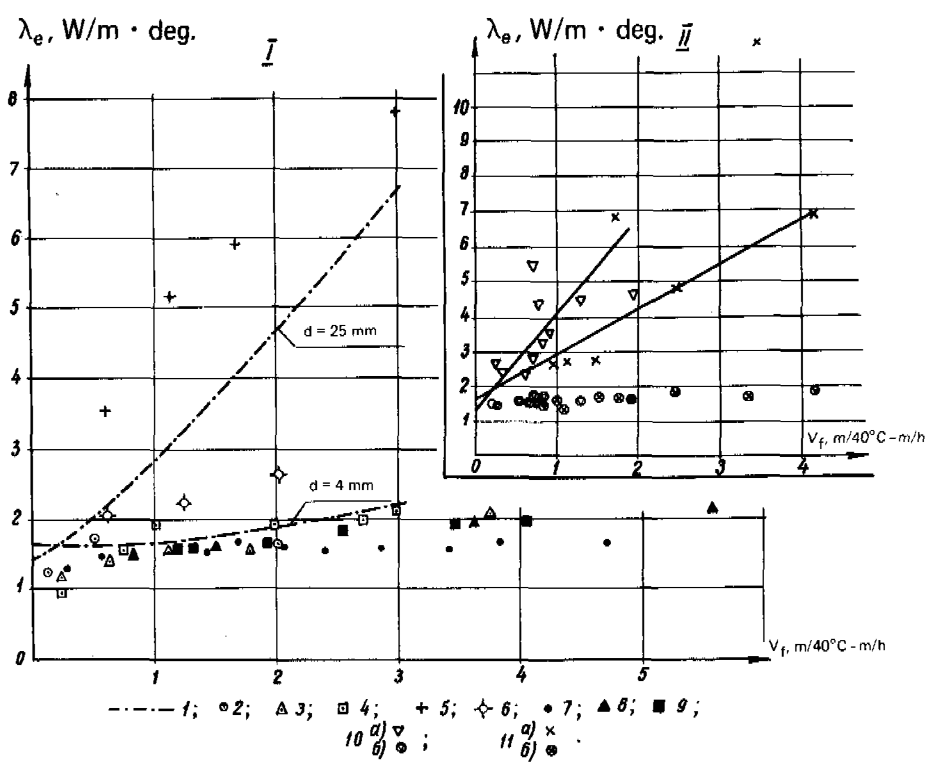

FIGURE 2 λ_e versus V_f. I. After Kapranov and Perl'shtein: 1--Equation (4), I series of experiments; 2--gravel 2-3 mm; 3--same 3-5 mm; 4--same 5-7 mm; 5--detritus and gravel 15-35 mm; 6--mixture (15-35 mm and 2-3 mm - 50 percent). II. series of experiments: 7--gravel 2-3 mm; 8--same 3-5 mm, 9--same 5-7 mm. II. Results of Chistopol'skii: (a) processed by the author; (b) processed by Kapranov and Perl'shtein; 10 a,b = "soil 1"; 11 a,b = "soil 2."

The slight increase in λ_e of gravelly soils is in complete agreement with the results of Aerov and Umnik for a laminar flow in materials with a corresponding particle size.

There is a reasonable agreement between the experimental data and Equation (4).

The completed part of the investigation shows that a noticeable increase in λ_e is unlikely in the case of a laminar flow in most soils, the large pores of which are, as a rule, filled with sandy gravel or silty clay.

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WATER MIGRATION IN SOIL DURING FROST HEAVING

PROFESSOR SEIITI KINOSITA Institute of Low Temperature Science, Hokkaido University, Sappore, Japan

INTRODUCTION

When a soil containing water freezes, the expansion observed is greater than the expansion resulting from the freezing of the water contained in the soil. The water in the unfrozen part of the soil migrates towards the freezing interface and there segregates in ice layers. It is presumed that the water migration is governed by the diffusion equation,

$$V = k \left(\frac{\partial w}{\partial z}\right) z = + D,$$

where V, k, w, z, and D represent respectively migration rate, diffusion coefficient, water content, vertical distance, and depth of the frost penetration from original soil surface, while

$$\left(\frac{\partial w}{\partial z}\right) z = + D$$

indicates the vertical gradient of the water content just below the freezing interface.[1]

A waterproof basin (area: 3 m × 3 m, depth: 192 cm) was constructed at Tomakomai, Hokkaido, for verifying the above Equation.[2] The basin was filled with a frost-susceptible soil (consisting of 28 percent sand, 31 percent silt, and 41 percent clay with an average specific surface area of 57 m^2/g). The water level in the basin was measured by reading the water level in the pipe placed close by and connected to the basin. There was no supply of water from outside the basin and water was neither supplied from above after the soil began to freeze.

RESULTS OF FIELD RESEARCHES

In the winter of 1970-1971, the freezing index calculated from daily mean air temperatures through the winter reached 600°C-day (Figure 1). Heave amounts, soil temperatures at several depths, freezing depths, and water levels were measured on nearly every tenth day (Figure 2). As seen from the figure, the soil began to freeze around November 30 and completely melted in early May. Since the intensive growth of ice needles at the beginning of freezing made heave measurements difficult, the surface was tamped down to the original level on December 17, 1970. After December 17, the soil surface continued to rise and reached a maximum (20 cm) at the middle of February. Both the freezing front and the water table continued to sink and became deepest (39 cm and 172 cm, respectively) at middle of March. The distance between the freezing front and the water table increased almost linearly from 25 cm to 100 cm and then became stationary. From these relations it becomes evident that the heaving of the soil surface was due to the moisture supply from the groundwater to the frozen soil.

Pits were dug and samples of soil taken at various depths on November 30, January 6, and February 27. For the frozen sample, its mass in

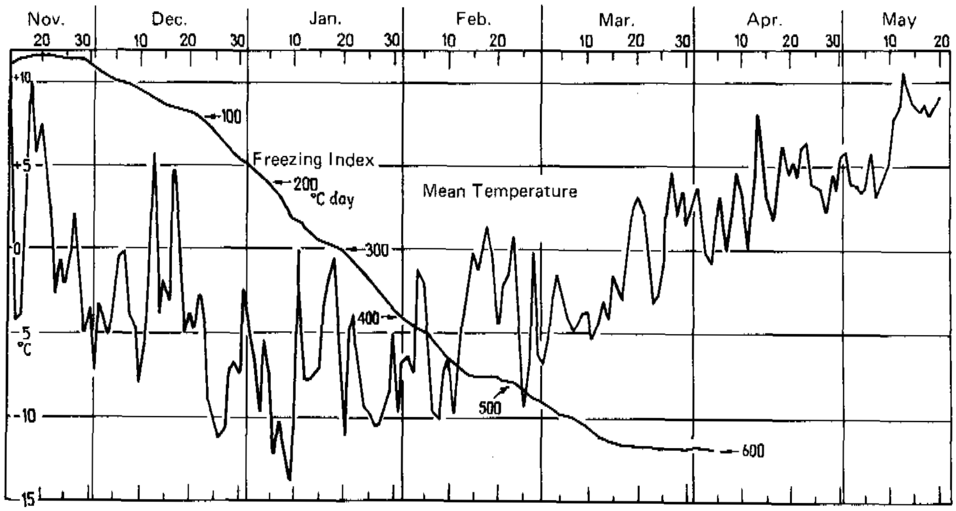

FIGURE 1 Daily mean air temperature and freezing index in the winter of 1970-1971 (Tomakomai, Hokkaido, Japan).

air, M_a, and that in kerosine, M_k, of known density ρ_k, were measured immediately after sampling, while its dried mass, M_d, was measured later in a laboratory. The following quantities were calculated from them:

Water content of the sample:

$$w = 100(M_a - M_d)/M_d;$$

Density of the sample:

$$\rho = \rho_k M_a/(M_a - M_k);$$

Mass of soil particles in a unit volume of the sample:

$$m_s = \rho_k M_d/(M_a - M_k).$$

Mass of moisture in a unit volume of the sample:

$$m_m = \rho_k (M_a - M_d)/(M_a - M_k).$$

Volumes of soil particles, V_s, moisture, V_m, and air, V_a, in a unit volume of the sample are given by $V_s = m_s/\rho_s$, $V_m = k'/\rho_w + (1 - k')/\rho_i \times m_m$, and $V_a = 1 - V_s - V_m$, where ρ_s is the density of the soil particles, ρ_w that of water, ρ_i that of ice and k' the ratio of unfrozen water to total moisture. Values of ρ_s were determined as 2.54 ~ 2.65. For the unfrozen sample ρ was calculated by measuring the weight of a rectangular prism (5 × 5 × 3 cm). The vertical distributions of W, ρ, and the relations of V_s, V_m, and V_a for $k' = 0$ are shown in Figure 3.

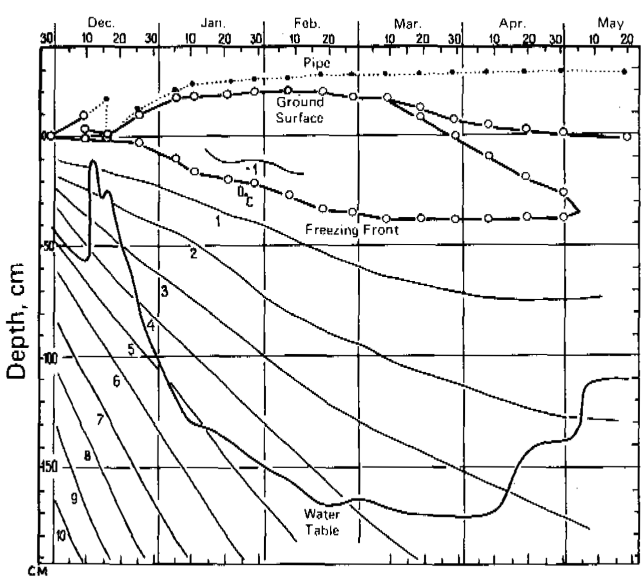

FIGURE 2 Heave-time relationships of the pipe and the ground surface, and depth-time relationships of the freezing front and the water table.

DISCUSSIONS

The freezing period between December 17 and February 27 was divided into five intervals (Table 1) and the change of water content $(M_2 - M_1)$ in each frozen layer was computed with the aid of relations, $M_1 = dD \cdot W_1$, and $M_2 = (dD + dh) \cdot W_2$, where dD is the thickness of the layer before freezing, dh is the heave amount and W_1 and W_2 are the volumetric water contents of the layer before and after freezing, which were computed by the above mentioned methods from the data on the nearest available days. The quantity, dh might be equal to the surface heave amount during the interval. But the soil surface may melt and sink in the daytime so as to make the heave amount less than dh. Therefore, dh was measured by the heave of a light plastic pipe which had been inserted deep into the soil (the dotted line of Figure 2). The changes of the water content divided by the length of the interval are the rates, V, of the water migration from the unfrozen parts toward the freezing front. The moisture gradient,

$$\left(\frac{\partial w}{\partial z}\right) \quad z = + D$$

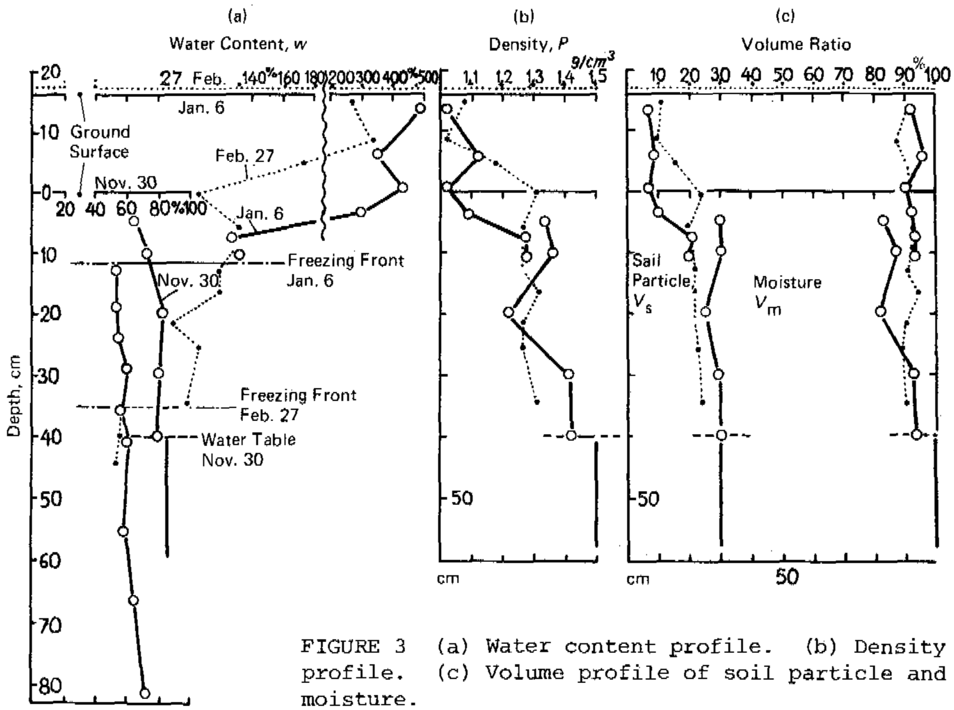

FIGURE 3 (a) Water content profile. (b) Density profile. (c) Volume profile of soil particle and moisture.

was obtained approximately from $W_d/(L - D)$, where L is the depth of the water table and W_d is the difference in the water contents of the soil at the water table and unfrozen layer just below the freezing interface.

The relation between V and

$$\left(\frac{\partial w}{\partial z}\right) z = + D$$

is expressed fairly well by the following equation:

$$V = 700 \quad \left\{\left(\frac{\partial w}{\partial z}\right) \quad z = + D^{-0.002}\right\} \tag{1}$$

This empirical equation provides an interpretation that the water migration occurs only when the gradient of the water content just below the freezing interface exceeds a certain value, 0.002, in the equation.

It is well known that there remains some unfrozen water in the frozen soil. If it is assumed that α percent of the original water remains unfrozen, the following relations hold:

$$dh = dh_1 + dh_2, \tag{2}$$

$$dh_1 = 1.09(M_2 - M_1) + 0.09M_1 \left(1 - \frac{\alpha}{100}\right), \tag{3}$$

$$dh_2 = \delta(dD + dh) - \gamma dD. \tag{4}$$

The heave amount dh may be divided into two parts: the one dh_1 is due to the water supply from below, while the other dh_2 is a correction term due to the change of air content in the soil freezing. γ and δ are volumetric air contents before and after freezing. Also, the following thermal equilibrium relationships exist at the freezing surface:

$$k_1 \left(\frac{\partial T_1}{\partial z}\right)_{z = + D} - k_2 \left(\frac{\partial T_2}{\partial z}\right)_{z = - D} = LV - L\,\frac{d(w - \alpha)D}{dz}. \tag{5}$$

Where k_1 and k_2 are the thermal conductivities of frozen and unfrozen soil, respectively, T_1 and T_2 are the temperatures of frozen and unfrozen parts, respectively, and L is the latent heat of fusion. The best fit value of α in this case was searched for and found to be euqal to 20 percent. The comparison of the three columns, dh, dh_1, dh_2, of

TABLE 1 Experimental Data

Layer	Freezing Period	dD, cm	dh, cm	M_2, g/cm^2	M_1, g/cm^2	V, g/cm^{2}day	$\left(\frac{\partial w}{\partial z}\right)$ z=+D, g/cm^4	γ, %	dh_1, cm	dh_2, cm
I	Dec. 17 to 26	3.0	11.7	12.8	1.6	1.24	0.0036	17	12.2	-0.2
II	Dec. 26 to Jan. 6	5.1	8.5	10.5	2.2	0.75	0.0031	27	9.2	-0.7
III	Jan. 6 to Jan. 21	10.5	4.5	10.5	4.5	0.40	0.0024	27	6.7	-2.1
IV	Jan. 21 to Feb. 8	6.3	1.4	5.1	2.7	0.13	0.0022	27	2.7	-1.3
V	Feb. 8 to Feb. 27	7.4	0.9	5.5	3.3	0.12	0.0021	27	2.5	-1.5

Table 1 were the values of dh_1 and dh_2 were computed from the Equations (3) and (4) with $\alpha = 20$ percent, showed that the Equation (2) was fairly well satisfied.

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CHANGE IN THE STRUCTURE OF THE RADIATION-HEAT BALANCE OF VARIOUS LANDSCAPES DURING ECONOMIC DEVELOPMENT OF A REGION

A. YA. KURENNAYA PNIIIS, Salekhard Geocryological Laboratory

Owing to the economic development of the Far North, the study of changes in geocryological conditions resulting from these activities is steadily increasing in importance. The present communication deals with changes in the structure of the radiation-heat balance of various natural landscapes of the Uyandinskaya depression and the southern spurs of the Polousnyi Ridge during construction and operation of industrial and residential complexes. The equation of the radiation-heat balance for various landscapes is as follows:

$$R = LE + \Delta W_s + P + A - Lr, \qquad (1)$$

where R is the radiation balance of the surface, LE is the heat used up on the evaporation of water in the active layer, ΔW_s is the heat used up on the evaporation and melting of snow, Lr is the heat released during the condensation of water vapor in the layer of seasonal thawing, P is the turbulent heat exchange between the soil and the atmosphere, and A is the heat flow in the topsoil. All parameters are in kcal/cm^2 for the warm season (March-August).

To determine the mean annual temperature of the ground, use was made of slightly modified calculations suggested by V. A. Kudryavtsev and V. V. Smirnov.[1]

The initial stage of forecasting presupposes a study of the region in its natural state. The analysis of its natural history reveals a clear differentiation into a number of landscapes and microlandscapes, each of which has its own specific heat and mass exchanges with the atmosphere. Among the large variety of structures of the radiation-heat balance of various landscapes, we shall examine the limits of variations of its components for the valleys and the mountainous parts only (Table 1).

The analysis of geocryological conditions revealed that a number of cryogenous phenomena occur in association with sections having a definite structure of the radiation-heat balance. E.g., solifluction is most characteristic of sections

TABLE 1 Components of the Radiation-Heat Balance of Natural Surfaces (in kcal/cm^2 per Warm Season)

Absolute Altitude, m	Slope Angle, deg	Aspect	Q	R	LE	P	A	Lr
River valleys, gently sloping foothills								
200-400	0-12		67.0-78.7	16.8-25.2	11.8-15.0	0.5- 4.0	3.1-4.8	---
Mountainous part of the region								
	12-20	N	62.2-67.7	13.4-17.1	4.5- 8.7	0.9- 4.5	4.0-4.7	0.0-1.8
		Mountainous 8.3	72.5	17.9-20.7	5.0- 9.5	2.6- 8.7	4.3-4.9	0.7-1.9
400-800		S	78.7-82.2	21.2-26.3	5.5-10.7	4.8-12.9	4.5-5.3	1.2-2.0
	30-40	N	51.2-56.0	7.6-10.5	1.7- 3.1	0.8- 2.1	3.1-4.5	0.0-1.6
		Mountainous 8.3	71.0-72.1	17.7-18.9	1.8- 3.5	7.7-12.2	3.4-4.9	0.4-1.8
		S	84.5-87.0	22.6-27.5	1.9- 3.9	12.8-18.5	3.7-5.2	0.7-1.9

TABLE 2 Change of Components of the Radiation-Heat Balance Resulting from a Disturbance of the Natural Cover over Extensive Areas in River Valleys and on Gently Sloping Foothills (in kcal/cm^2 per Warm Season). Absolute Altitudes: 200-400 m; Slopes: 0°-12°.

Type of Disturbance	ΔR	$\Delta LE + \Delta W_S$	ΔP	ΔA
Snow removal: $h = 0.5$-0.7 m, $\rho = 0.17$-0.19 g/cm^3	+(1.6-3.2)	+(1.5-1.6)	+(0.1-1.2)	+(0 -1.1)
Compaction accompanied by contamination of snow: $h = 0.10$-0.20 m, $\rho = 0.34$-0.42 g/cm^3	+(1.4-2.9)	+(1.2-1.4)	+(0.3-0.7)	+(0 -0.9)
Increase in snow depth: $h = 0.3$-0.5 m	-(0.4-1.4)	+(0.1-0.8)	+(0.4-1.3)	-(0.1-0.9)
Removal of trees and brush	+(0.9-3.0)	-(1.2-2.3)	+(0.7-1.9)	+(1.4-3.0)
Removal of vegetation	+(1.9-4.3)	-(2.2-3.2)	+(0.9-2.0)	+(1.6-4.5)
Snow removal	+(2.5-6.0)	-(0.5-1.2)	+(1.5-4.3)	+(1.3-3.1)

NOTE: (+) means increase in the values of parameters; (-) means decrease; h = the snow depth, ρ = the snow density.

where $0.20 < R/Q < 0.25$, $0.35 < LE/R < 0.5$, $0.07 < P/R < 0.15$, and $Lr/R = 0$. This holds for gentle (up to 20°) windward slopes facing north-east, east, or north (Q is the total solar radiation in kcal/cm^2 in the warm season). Solifluction lobes are found on gentle (up to 10°) slopes facing in any direction, or at the foot of such slopes, where the ratios of the components are:

$$0.23 < \frac{R}{Q} < 0.27;\ 0.5 < \frac{LE}{R} < 0.6;$$

$$0.03 < \frac{P}{R} < 0.09;\ \frac{Lr}{R} = 0.$$

Sections with numerous frost fractures are characterized by the ratios

$$0.8 < \frac{LE}{R} < 0.9;\ 0.04 < \frac{P}{R} < 0.1;\ \frac{Lr}{R} = 0.$$

The sections where $0.08 < LE/R < 0.35$, $0.25 < P/R < 0.67$, and $0.03 < Lr/R < 0.11$ are characterized by the occurrence of frost sorting of rocky material, physical weathering, and nivation.

Industrial development of a region changes the conditions of heat and water exchange between the soils and the atmosphere, which results in changes in the temperature regime of the ground and the depth of thaw. Tables 2 and 3 show the most typical limits of changes of components of the radiation-heat balance, which result from the most common types of disturbance of the natural cover. Landscapes with a changed structure of the radiation-heat balance are usually characterized by transformations of cryogenous phenomena or by variations in the intensity of the latter. For example, removal of snow in a mountainous area (reduction of LE/R, increase in P/R and Lr/R) may activate physical weathering, and in a valley (increase in R/Q, P/R and LE/R) may lead to more intensive frost fracturing, disappearance of syngenetic segregation ice, formation of hummocky relief, and appearance of seasonal ice wedges. On the other hand, an increase in the thickness of the snow cover in the mountains may lead to formation of firn, more intensive nivation, and on gentle slopes, to solifluction.

Partial or complete removal of the vegetation cover in the valleys increases R/Q and P/R and considerably reduces LE/R (total heat used up on the soil evaporation), which in the case of a high groundwater table may lead to the formation of water-filled depressions resulting in gradual melting of ice and formation of thermokarst and thermal erosion structures. As an example of this, we can cite the construction of the Kuiga-Deputatskii highway.[2]

TABLE 3 Change of Components of the Radiation-Heat Balance Resulting from a Disturbance of Natural Covers in Mountainous Areas (in kcal/cm^2 per Warm Season)

Type of Disturbance	ΔR	$\Delta LE + \Delta W_S$	ΔP	ΔLr	ΔA
Snow removal: $h = 0.2$-0.3 m, $\rho = 0.25$-0.33 g/cm^3	+(1.8-4.0)	-(0.2-0.6)	+(1.1-2.5)	+(0.3-0.6)	+(0.2-1.5)
Snow compaction accompanied by contamination: $h = 0.1$-0.15 m, $\rho = 0.34$-0.42 g/cm^3	+(0.5-3.0)	±(0.1-0.3)	+(0.7-2.3)	+(0.5-0.6)	+(0.6-1.0)
Increase in snow depth: $h = 0.3$-0.6 m	-(1.2-2.8)	±(0.3-0.6)	-(0.2-0.9)	±(0.2-0.7)	-(0.6-3.5)
Removal of trees and brush	+(0.3-2.1)	-(0.4-1.1)	+(0.5-2.0)	+(0.1-0.3)	+(0 -1.7)
Removal of vegetation	+(0.7-4.1)	-(0.3-1.5)	+(0.6-3.5)	+(0.3-0.6)	+(0.7-3.0)
Snow removal	+(1.9-6.2)	-(0.4-1.9)	+(2.0-6.7)	+(0.3-0.9)	+(0.6-2.3)

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PHASE BOUNDARY CONFIGURATION IN STRUCTURALLY NONUNIFORM EARTH'S CRUST FROZEN TO GREAT DEPTH

V. V. LOVCHUK PNIIS, Salekhard Geocryological Laboratory

Attempts to determine the causes of floods in mines in different parts of the Magadan Oblast, Yakutiya, and Chukotka brought out the need for very accurate determinations of phase boundary configurations in complex geological structures, because thawed and frozen zones are mined by vastly different methods. Calculations of the configuration of the lower permafrost boundary from the available data on the nature and the properties of a given geological structure would considerably reduce the volume of costly drilling operations and yet provide sufficiently accurate information on the area where mining would be safe.

Most investigators base their analysis of the effect of structural nonuniformities in the Earth's crust on the existence of a steady-state distribution of heat. In the case of unsteady temperature distributions, use is made of numerical solutions of the Stefan problem suggested mainly for horizontally layered media.

We shall analyze a two-dimensional model of freezing, since most geological structures normally represent systems of folds strongly elongated in one direction, inclined beds, and ore bodies with distinct boundaries between different lithological units. Hence they can be adequately described within the framework of a two-dimensional model. The surfaces that separate layers with different physical properties are given by certain implicit functions with two variables. No allowances are made for the effect of possible sources of heat and the processes of mass transfer (both factors could be considered by modifying the program).

Cooling of the upper surface of a structurally nonuniform medium results in formation of a frozen layer of variable thickness $y = \xi(x,\tau)$, i.e., a mobile boundary (phase transition front) is formed, which is the site of transitions from one aggregate state to another. The latent heat of fusion required for this is δ = 80 kcal/kg. The amount of water contained in the rocks $\rho_i W_i$, which is transformed directly to ice, is fixed in each single volume of the given region. It is assumed that within each layer $\rho_i W_i$ is a function of y. It is required to determine the configuration of the phase transition front and the temperature field in frozen and thawed zones at different moments of time.

The mathematical formulation of the problem is as follows:

$$\rho_i{}^{M}(x,y)\, c_i{}^{M}(x,y)\, \frac{\partial T_M}{\partial \tau} = \frac{\partial}{\partial x}\left[\lambda_i{}^{M}(x,y)\frac{\partial T_M}{\partial x}\right] + \frac{\partial}{\partial y}\left[\lambda_i{}^{M}(x,y)\frac{\partial T_M}{\partial y}\right], \qquad (1)$$

$$\rho_i{}^{T}(x,y)\, c_i{}^{T}(x,y)\frac{\partial T_T}{\partial \tau} = \frac{\partial}{\partial x}\left[\lambda_i{}^{T}(x,y)\frac{\partial T_T}{\partial x}\right] + \frac{\partial}{\partial y}\left[\lambda_i{}^{T}(x,y)\frac{\partial T_T}{\partial y}\right], \qquad (2)$$

$$\lambda_i{}^{M}[x,\xi(x,\tau)]\frac{\partial T_M}{\partial n} - \lambda_i{}^{T}\left[x,\xi(x,\tau)\right]\frac{\partial T_T}{\partial n} = \delta\rho_i{}^{T}W_i\left[\xi(x,\tau)\right]\frac{\partial \xi}{\partial \tau}\cdot\frac{1}{/\mathrm{grad}F/}, \qquad (3)$$

where $F = y - \xi(x,\tau)$; $\partial/\partial n$ is the derivative along the normal to the phase boundary $y = \xi(x,\tau)$

$$T_M[x,\xi(x,\tau)] = T_T[x,\xi(x,\tau)] = T_F = 0°\mathrm{C}, \qquad (4)$$

$$\left.\begin{aligned} \frac{\partial T_M(o,y,\tau)}{dx} &= \frac{\partial T_T(o,y,\tau)}{dx} = 0 \\ \frac{\partial T_M(L/_2,y,\tau)}{dx} &= \frac{\partial T_T(L/_2,y,\tau)}{dx} = 0 \end{aligned}\right\}, \qquad (5)$$

$$d\xi/\partial x = 0, \quad x = 0, \quad x = L/_2, \qquad (6)$$

$$T(x,y,o) = f(x,y), \qquad (7)$$

$$T(x,o,\tau) = T_0 + A\sin\omega\tau, \qquad (8)$$

$$\frac{\partial T_T(x,\psi,\tau)}{\partial y} = q_0 = \text{const}, \qquad (9)$$

$$\xi(x,o) = b = \text{const}, \qquad (10)$$

where T is temperature, °C; T_F is the temperature of the phase transition water-ice, °C; c is the specific heat, kcal/kg · deg; ρ is the density, kg/m^3; λ is the coefficient of thermal conductivity, kcal/m·h·deg; ω is the frequency of temperature fluctuations on the surface, h^{-1}; τ is the time, h; q_0 is the heat flux from the interior of the Earth, kcal/m^2·h; x,y are the space coordinates; m and T are the indices of frozen and thawed zones, respectively; i is the number of the layer; w is the water content, percent.

Condition (5) realizes the periodicity of the structure. If the structure is not periodic, the heat flux at the vertical boundaries $x = 0$ and $x = L/_2$ should be substituted for Condition (5).

Problems (1) to (10) are solved in a BESM-6 computer by numerical finite-difference methods using our program[1] for a system of recurring synclinal folds. As an example, we shall analyze the configuration of the phase boundary and its dynamics during freezing of a "five-layer" model with a combination of layers SC, SC, and S, where S is sandstone and C is coal. The thermophysical properties of the S layer are: $\lambda_T = 1.6 \times 10^4$ kcal/m yr deg; $\lambda_M = 1.7 \times 10^4$ kcal/m yr deg; $\rho_T = \rho_M = 2{,}300$ kg/m^3; $c_T = c_M = 0.19$ kcal/kg deg; $W_S(y) = 0.1\text{exp}(-0.023y)$; $\delta\rho_T \sim 184{,}000$ kcal/m^3; layer C: $\lambda_T = 0.22 \times 10^4$ kcal/m yr deg; $\lambda_M = 0.35 \times 10^4$ kcal/m yr deg; $\rho_T = \rho_M = 1{,}200$ kg/m^3; $c_M = c_T = 0.31$ kcal/kg deg; $W(y) = 0.36\text{exp}(-0.047\ y)$; $\delta\rho_T = 96{,}000$ kcal/m^3.

$T_0 = -0.1$°C; $A = 6$°C; $q_0 = 480$ kcal/m^2 yr; $p = 2\pi/\omega = 10^3$ yr. Both C layers are semicircular (Figures 1 and 2). The radius of the first (measured from the center) boundary SC is 15 m and that of CS is 20 m. The radius of the second SC is 130 m. $L/_2 = 300$ m; $\Psi = 1{,}200$ m.

The results of calculations indicate that the phase boundary configuration undergoes complex changes during freezing of a structurally nonuniform medium. For example, within the coal layer C (the first layer from the center of the fold), the minimum thicknesses of the frozen zone at the time $\tau = (0.05\ p;\ 0.1\ p)$ are observed near the outcrop of the layer C on the surface (Figure 1). As the boundary $y = \xi(x,\tau)$ moves towards the interior, the region of minimum thickness shifts towards the center of the fold. Between the first and the second coal layers, the phase boundary moves subparallel to the cooling surface. Near the outcrop of the second layer, the minimum thicknesses are observed at $\tau = (0\text{-}0.2\ p)$ (Figure 1). At $\tau = 0.2\ p$ the thickness is 15.9 m for an average structure of 40.0 m. Outside the structure, the boundary $y = \xi(x,\tau)$ descends rapidly and then remains parallel to the cooling surface.

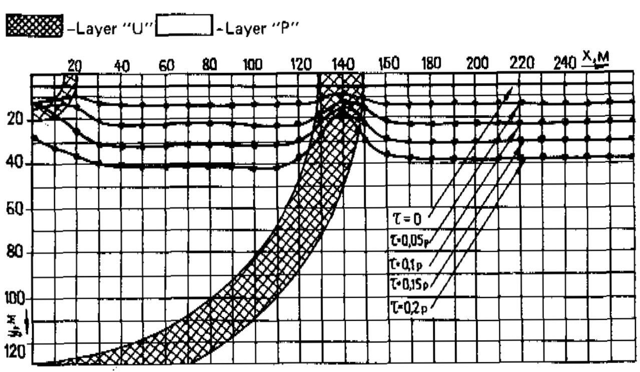

FIGURE 1 Phase boundary configuration at various times in a five-layer synclinal fold. (a) = C [=u] layer; (b) = S [=p] layer

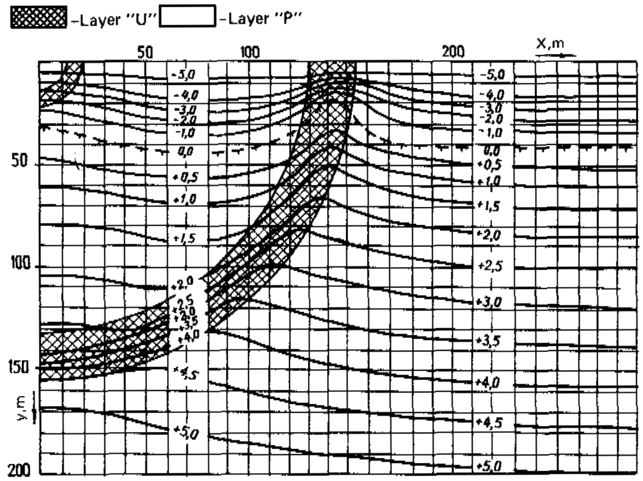

FIGURE 2 Temperature field in a five-layer synclinal fold at $\tau = 0.2\ p$ (symbols the same as in Figure 1)

Structural nonuniformities in the upper horizons of the Earth's crust lead to a redistribution of the heat flux from the interior of the Earth. For example, on reaching the C layer with its high thermal resistance, some of the heat flows around it. From 150 to 290 kcal/m^2· yr, 35 to 80 percent of the heat flux (65 percent on the average), pass through the coal layer. The heat flux below the C layer is greater (370-430 kcal/m^2 yr) than that within the layer. This means that in the course of freezing of the fold, the amount of heat used up from above is 1.5 to 2.0 times greater than the heat flux passing through the layer from below. As a result of this, the rocks in the interior of the fold become considerably colder. This is incicated by the fact that the maximum drop of the isotherms occurs in the interior of the fold (Figure 2). On the other hand, the isotherms rise below the structure (Figure 2), which indicates that the underlying S horizon receives more heat.

The horizontal heat fluxes in the S layer, which accommodates the structure, are directed away from the latter and reach 100-200 kcal/m^2 yr, while at the outcrop of the C layer they reach 900-1,300 kcal/m^2 yr.

Therefore, the analysis of two-dimensional freezing indicates a considerable horizontal redistribution of heat. Comparison of the phase boundary configuration obtained by solving the two-dimensional problem [(1)-(10)] with that obtained by means of a system of thermally insulated "flow lines" showed that if the average thickness of the frozen zone is less than the vertical dimensions of the structure (synclinal fold), the error in the determination by the second method of the thickness of the frozen zone near the steep boundaries between the layers may reach 20-30 percent or more. In the case where the average thickness of the frozen zone is greater than the vertical dimensions of the structure, the phase boundary configuration below the structure may be distorted altogether, if determined by a system of thermally insulated "current lines."

The phase boundary configuration obtained by the computer method outlined here reflects very well the configuration of the lower permafrost boundary within the synclinal structures of the Arkagalinskaya depression in the northeastern part of the USSR.

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INVESTIGATION WITH THE HELP OF AN ELECTRONIC DIGITAL COMPUTER OF CRYOGENOUS STRUCTURE AND HEAVE OF FREEZING FINE-GRAINED SOILS

V. G. MELAMED AND A. V. MEDVEDEV Moscow Lomonosov State University

Quantitative investigation of the freezing process in fine-grained soils is one of the important tasks of permafrost studies. Such investigations are difficult because freezing of fine-grained soils is accompanied by conductive heat exchange, as well as water migration toward the freezing front, both of which must be considered. This is possible only by solving a system of equations describing heat and mass exchange in freezing and thawed soils. These equations are nonlinear, which is because the coefficients describing heat and water conductivities of frozen and thawed soils are strongly dependent on the total water content, which is unknown. Furthermore, most phase transitions in fine-grained soils occur within the temperature range corresponding to the curve of unfrozen water. Another important point is the possibility of heaving and formation of ice layers, as a result of which the changes in the dimensions of frozen, and thawed, zones obey different laws. Hence freezing of fine-grained soils presents a difficult problem. We shall assume that water migration occurs in the liquid phase only, that deformation of the soil matrix and migration in the frozen zone are negligible, and that the soil is homogeneous with the water content remaining constant.

Interrelationship between heat and water exchanges is expressed in the boundary conditions that, at the freezing front, depend on the dynamics of the process. For example, in the case of formation of a massive or a thin-layered streaky structure, the positive average velocity of the mobile phase boundary $y(t)$ is $y'(t) > 0$, where t is time. The water content at the front (facing the thawed zone) is at a minimum and is practically constant (close to the plastic limit, W_0). The criterion of such freezing at a given moment of time is the condition where the difference between the heat fluxes at the front $A(t)$ exceeds the heat required to freeze the migrating water $B\{t,W[y(t)]\}$, with the water content W on the phase boundary equal to W_0. If, however, $A(t) < B(t,W_0)$ at $A(t) > 0$ (i.e., during freezing), an ice layer of finite thickness will be formed. The dimensions of the thawed zone remain unchanged ($y'(t) = 0$), while the frozen zone is increasing owing to the freezing of migrating water. The water content on the moving boundary is not known beforehand (it varies from W_0 to saturation, W_S), while water transfer and heave depend on $A(t)$.

The given problem is reduced to a system of quasi-linear equations of the parabolic type describing the temperature field in both zones (with allowances for the translation of the frozen zone due to heaving), as well as the moisture field in the thawed zone.

The mathematical formulation of a single-front problem in the case of freezing is as follows. It is required to determine the temperature $\bar{u}(x,t)$, $u(x,t)$ and the unit total water content $\bar{W}(x,t)$, $W(x,t)$ and the phase boundary

$y(t)$ at $t \varepsilon (0,T)$ from the following conditions:

$$\bar{c}(\bar{u},\ \bar{w})\left[\frac{\partial \bar{u}}{\partial t} - h'(t)\frac{\partial \bar{u}}{\partial x}\right] = \frac{\partial}{\partial x}\left[\bar{\lambda}(\bar{u},\ \bar{W})\frac{\partial \bar{u}}{\partial x}\right],\ -h(t) < x < y(t),$$

$$\frac{\partial u}{\partial t} = a^2{}_0\frac{\partial^2 u}{\partial x^2},\ \frac{\partial w}{\partial t} = \frac{\partial}{\partial x}\left[k(w)\frac{\partial w}{\partial x}\right],\ y(t) < x < l,$$

$$u[y(t),\ t] = \bar{u}[y(t),\ t] \equiv u_0, \bar{u}[-h(t),\ t] = \varphi(t),\ h(0) = 0,$$

$$y(0) = y_0 \varepsilon (0,\ l),\ \lambda_0 u_{x/x} = 1 = \varphi_1(t),\ k(w)W_{x/x} = 1 = \varphi_2[w(l,t),\ t],\ \bar{u}(x,0) = f_1,\ (x),\ \bar{W}(x,0) = f_2(x),\ x\varepsilon(0,\ y_0),$$

$$u(x,\ 0) = f_3(x),\ w(x,\ 0) = f_4(x),\ \ x\varepsilon(y_0,\ l),$$

$$h'(t) = \tfrac{1}{2}\{\nu B[t,\ w(y(t,)t]\} + (m - w_n\ y'(t) + |\nu B\{t,\ w[y(t),\ t]\} + (m - w_n)y'(t)|\},$$

$$\bar{w}[y(t),\ t] = \frac{\nu B\{t, w[y(t),\ t]\} + my'(t)}{h'(t) + y'(t)}.$$

On the phase boundary at $A(t) > 0$ -, $A(t) - B[t,\ w(y(t),t)] = Qy'(t)$, $w[y(t),\ t] = W_0$, if $A(t) \geq B(t,\ w_0)$. If $A(t) \leq B(t,\ w_0)$, $A(t) = B[t,\ w(y(t),\ t)]$, $y'(t) = 0$.

Here $A(t) = \lambda(\bar{u}_0,\bar{W})\ u_x|x = y(t) - \lambda_0 u_x|y(t)$, $B\{t,w[y(t),t]\} \equiv \mu\ [k(w)w_x]_{y(t)}$; λ, $\bar{c}$ are the heat conductivity and the specific heat of frozen soil; λ_0, $a_0{}^2$ are the heat conductivity and thermal diffusivity of thawed soils; γ is the coefficient of volumetric expansion, water-ice; μ is the latent heat of fusion of 1 m^3 of water; $Q = \mu[w_v - w_H(u_0)]$, where $w_H(\bar{u})$ is unfrozen water; $k(w)$ is the coefficient of potential conductivity; $m = [w_0 - w_H(u_0)]\nu + w_S(u_0)$, $w_H(u_0) < w_0$; u_0 is the temperature at the start of freezing, $\Phi(t) \leq u_0$, $\psi \geq 0$, $t\varepsilon[0,T]$, $f_1(x) \leq u_0$, $f_2(x)\varepsilon[w_0,\ 1]$, $f_3(x) \geq u_0$, $f_4(x)\varepsilon[w_0,\ w_S]$.

To solve this problem at $t\varepsilon(O,\ T)$, it is essential that $A(0) > 0$. The problem cannot be solved at $t < T$, if a third phase is formed or thawing occurs from below [$A(t) < 0$], or $y(t) = l$. The total dimensions of the frozen zone are given by $\xi(t) \equiv y(t) + h(t)$. An algorithm of numerical integration of this problem has been developed on the basis of the straight-line method (the Ruttë scheme), and used with slight modification in an M-20 electronic digital computer. The solution makes it possible to determine the rates of freezing and heaving in time, as well as the distribution of temperature and water content (averaged) throughout the depth for an arbitrary moment by time.

On solving the problem of freezing with water migration, it is possible to determine quantitatively the processes of freezing, ice accumulation, and heaving in both open and closed systems in relation to heat and water exchanges in the boundary region, as well as the natural water content, the coefficients of heat and water conductivity, etc. Figure 1 shows the calculated dynamics of long-term freezing at a surface temperature $\phi(t)$ with a period of 300,000 yr, an amplitude of 4°C, and an average amplitude of 1°C ($n = 1$). In the second case the period considered was 7 times shorter ($n = 7$). In the latter case, with the same freezing index, $\phi(t)$ warms up by $n - 1$ during "winter." The calculations were performed for soil with a high potential conductivity (using Gardner's data). The other parameters were: the geothermal gradient 0.01, $W_0 = 0.2$; $W_S = 0.4$; $f_4(x) \equiv 0.28$; $u_0 = 0$. The freezing process at $n = 1$ and $n = 7$ was the same, while heaving and especially the cryogenic structure were different. For example, at $n = 1$, massive texture was formed everywhere (except the base of the permafrost bed), and the total water content gradually increased with depth. At $n = 7$, however, three sections with increased ice contents corresponding to the first three warm periods were formed within the permafrost. The highest ice accumulation occurred in a 10-m-thick section at a depth of 130 m, corresponding to the first warm period and containing isolated ice layers with a

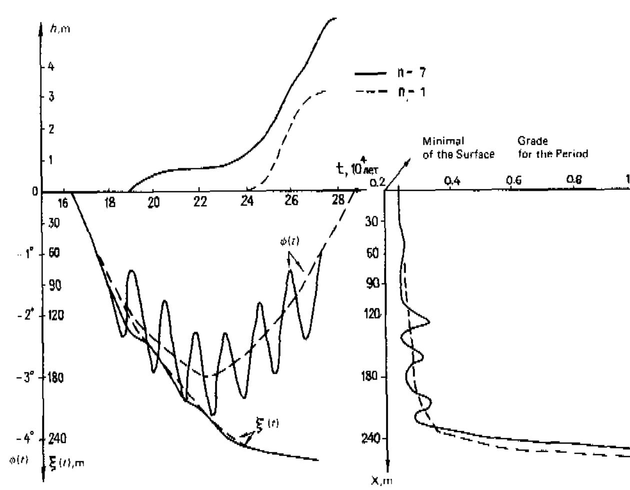

FIGURE 1 Dynamics of freezing $\xi(t)$ and heaving $h(t)$ in time and distribution of the ice content (w) throughout the depth (x) at different surface temperatures $\phi(t)$.

total thickness of 50 cm (in the other two sections the total thickness of ice layers did not exceed 20 cm). At the same time, at $n = 7$, ice formation at the base of the permafrost layer occurred more rapidly also. Therefore, calculations of freezing with allowances for the water migration make it possible to determine quantitatively the relationship between the cryogenous structure of permafrost and the conditions of its formation.

Considering the difficulties involved in solving the problem in a general case, as well as the large number of factors affecting the process as a whole, individual parts of the process can be determined by applying model methods, and in this electronic digital computers can be of considerable help. However, it is essential to remember that in model methods where the problem is solved in a semilimited region with boundary conditions remaining constant, any rough approximations in the postulation of the problem are extremely dangerous. For example, the use of a model solution without allowances for heaving in the cases where intense ice accumulation is taking place produces meaningless results. Lately an algorithm has been developed for a model solution of the problem with any given accuracy. The specific characteristic of the problem in the model is the constancy of the ice content throughout the cross-section, so that formation of an ice layer is possible on the surface only and is not limited in time. In contrast to a model solution without allowances for heaving, which can be applied only to soils with water content not exceeding the heaving limit, this solution provides an effective method of carrying out serial calculations of freezing, heaving, and ice accumulation processes using any input data. In particular, it helps to determine for any natural soils $W(x,0) \in (W_0, W_S)$ the surface temperature $\phi^* \equiv \text{const} < \mu_0$ such that at all $\phi \in (\phi^*, \mu_0)$ an ice layer is formed on the surface, while any $\phi < \phi^*$ brings about freezing of soil with formation of massive or streaky structures. Let us note that at $\phi = \phi^*$ ice formation on the surface is especially rapid. Figure 2 shows the calculated ϕ^* and corresponding ice layers as functions of the initial water contents of three types of soils with different physical and aqueous properties. For soils 1 to 3, the coefficients of potential conductivity are correspondingly reduced: $W_0 = 0.2$, 0.36, and 0.4; $W_S = 0.4$, 0.54, and 0.5.

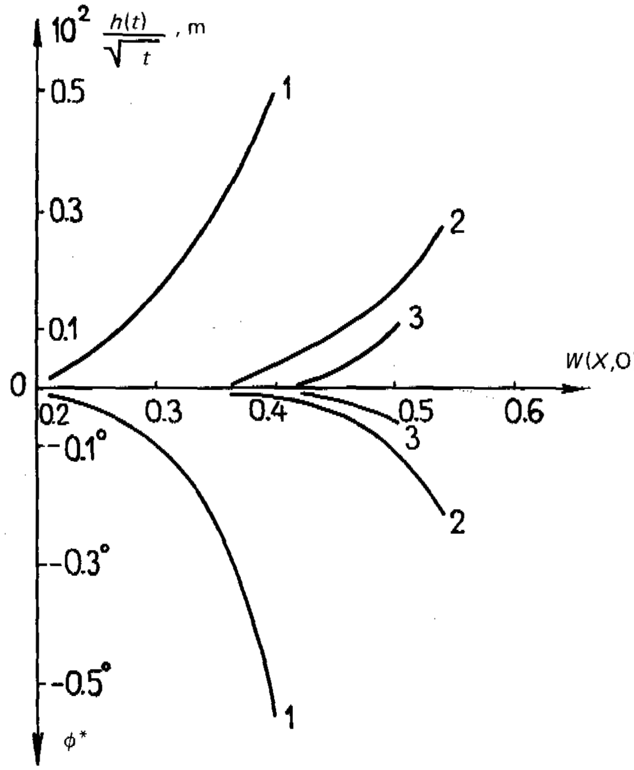

FIGURE 2 Optimum (from the point of view of ice formation on the surface) surface temperatures ϕ^* and corresponding ice layers as functions of natural water content under model conditions. Soils with different hydrophysical properties (the numbers designate different soils).

Hence, the study of heat and water transfer on freezing makes it possible to solve the most complex problems of permafrost forecasting in an actual geological-geographical environment, as well as to investigate the problem of determing the freezing conditions, which would be optimum from the point of view of ice accumulation and heaving.

MOISTURE TRANSFER IN THAWING CLAY SOILS

F. YA. NOVIKOV NIIOSP, Research Institute of Foundations and Subsurface Structures

Thawing of frozen clay soils is invariably accompanied by evaporation or condensation of moisture on their surface. The moisture content of soil decreases on evaporation and increases in the case of condensation, if the moisture exchange between the thawing layer and the surrounding medium occurs via its surface only and the layer does not receive moisture from any other sources (e.g., groundwater).

Changes in the moisture regime of thawing soils resulting from evaporation or condensation on the surface are so insignificant that they can be ignored, while the study of freezing and thawing can be based on the theory of heat conductivity. However, there are cases where the mass transfer in thawing soils cannot be ignored.[1]

Moisture transfer and the resulting changes in the thermophysical properties of soils are, as a rule, not considered in the existing methods of quantitative determinations of the thawing process in frozen clay soils. In calculations of the bearing strength of these soils, changes in the mechanical properties resulting from changes in their moisture content are not considered either. At the same time, allowances for the moisture distribution in the thawing layer make it possible to determine more accurately the thawing process and the condition of the limiting equilibrium of soils in the walls of underground workings and trenches, in the slopes of pits and embankments, on lake shores, etc.

We shall examine the moisture transfer in clay soils under natural conditions, in which thawing and freezing result from natural changes in the temperature and the moisture content of the surrounding medium. In most cases the moisture in clay soils is transferred in a capillary-film state. The temperature gradients are relatively low and therefore their effect on the moisture transfer can be ignored. Moisture transfer in the frozen zone can be ignored also.

Moisture is transferred in thawing clay soils due mainly to the moisture gradient or the moisture transfer potential.

It is known that the mass-transfer characteristics depend on the moisture content. The effect of changes in the moisture content of soils on their coefficient of potential conductivity, the moisture-transfer potential, and the specific mass capacity was investigated by experimental methods developed by A. V. Lykov.[2] Use was made of various suglinoks from Vorkuta (see Table 1).

The coefficient of potential conductivity was determined by the steady-state mass-exchange method. It was found from the curves of the moisture distribution in the samples and the intensity of the moisture flow using the following formula:

$$K = \frac{i}{\gamma_d (\partial u/\partial x)},$$

where K is the coefficient of potential conductivity, i is the amount of moisture evaporated in unit time, γ_d is the dry unit weight of the soil, and $\partial u/\partial x$ is the moisture gradient in the direction of the axis of the cylinder.

The curves showing the moisture content of various suglinoks consist of broken lines. The moisture-content curve is close to a slightly inclined straight line, which indicates that K is little dependent on the moisture content. Calculations show that suglinoks with a maximum molecular moisture capacity of 11.2 percent have $K = 1.42 \times 10^{-4}$ m^2/h at moisture contents ranging from 15 to 21 percent. In suglinoks with a maximum molecular moisture capacity of 12.1 percent, $K = 1.36 \times 10^{-4}$ m^2/h at moisture contents of 14 to 20 percent.

Under natural conditions, changes in the moisture content of clay soils during thawing and freezing accompanied by evaporation or condensation on the surface remain within the limits at which K can be taken as constant.

The moisture-transfer potential and the specific mass capacity of suglinoks were determined by the thermodynamic equilibrium method.

The experimental results are given in Figure 1. The moisture content of the reference material and the potential scale are shown along the abscis-

TABLE 1 Average Physical Characteristics of the Vorkuta Suglinoks (after G. P. Mazurov)

	Fractions, %			Plasticity Limits					
Lithological Type	Sandy Gravel, > 0.05	Silt Particles, 0.05-0.005	Clay, > 0.005	Liquid Limit	Plastic Limit	Plasticity Index	Grain Weight, g/cm^3	Unit Weight, g/cm^3	Natural Moisture Content, %
Surface	23.2	59.1	17.1	37.9	20.2	17.7	2.69	1.91	25.8
Glacier Ice	45.3	39.4	13.3	34.5	18.2	16.3	2.70	1.99	21.4
Sea Ice	51.2	31.6	17.2	30.7	16.0	14.7	2.69	2.02	16.8

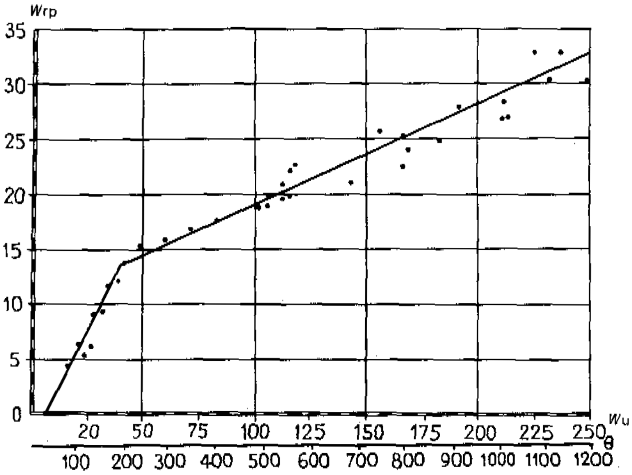

FIGURE 1 Specific mass content of suglinoks in relation to that of reference material (filter paper) in a state of thermodynamic equilibrium.

sa. The maximum hygroscopic moisture content of the reference material (filter paper) was taken as 100 potential units. The moisture content of suglinoks is shown along the ordinate. The points corresponding to a state of equilibrium lie approximately along a straight line. Hence the specific mass capacity of the Vorkuta suglinoks is constant within a wide range of moisture content and can be calculated from the tangent of the inclination of the straight line towards the abscissa at a point corresponding to the chosen potential, i.e.,

$$\varepsilon = \frac{\partial u}{\partial \Theta},$$

where ε is the specific mass capacity of suglinoks, u is the potential of thawing soil, and Θ is the moisture transfer potential.

At a moisture content ranging from 12 to 32 percent, the Vorkuta suglinoks with a unit weight of 1,600 kg/m^3 have $\varepsilon = 2.0 \times 10^{-4}$ kg/pot. units or $\varepsilon_v \simeq 0.32$ kg/m^3 pot. units (ε_v is the unit mass capacity). The suglinoks from other areas, e.g., from the temperate region (according to M. G. Murashko), have approximately the same mass-transfer characteristics.

These experimental results can probably be extended to other regions with seasonal thawing and freezing. Within a certain range of moisture content, the mass-transfer characteristics will be independent of the latter, as long as thawing is not accompanied by intensive drying of the soil at large temperature gradients. In this case the moisture field in thawing soil is described by the following equation:

$$\frac{\partial \Theta}{\partial \tau} \nabla (k \nabla \Theta), \tag{1}$$

where τ is the time.

The moisture field forms while the thawing boundary is moving in accordance with the heat exchange. This process consists, so to say, of two components: the heat exchange on thawing and the mass transfer. However, both phenomena occur within the same region and have a common mobile boundary, which should be allowed for in solving the boundary problems in both cases.

The problem of the moisture distribution in a thawing layer must be solved with allowances for the changes in the moisture-transfer potential in the thawing soil in the immediate vicintiy of the thawing plane. These changes depend not only on the general drop in the moisture-transfer potentials in the thawing zone, but also on the ratio of the rate of movement of the thawing plane to the rate of leveling of the moisture content within the thawing zone. In the case of a relatively high thawing rate and a low coefficient of potential conductivity, the moisture-transfer potential at the thawing plane will be close to the initial potential, Θ_a.

Let us examine a one-dimensional problem--the thawing of a flat wall (Figure 2). At time τ, thawing amounts to h and the distribution of the moisture-transfer potential in the thawing zone is indicated by curve 1. In time $d\tau$, thawing amounts to dh and the potential distribution is described by curve 2. During the same time, water in the amount of $\gamma_d \varepsilon(\Theta_a - \Theta_h - d\Theta)dh$ leaves the layer dh due to the gradient of the moisture-transfer potential (Θ_h is the moisture-transfer potential at the thawing plane). In other words, the moisture content of a layer, which had thawed in an infinitely short interval of time, would decrease (increase) by the amount equal to the amount of moisture lost to (received from) the thawed zone due to the gradient of the moisture-transfer potential at the thawing plane. Consequently, ignoring infinitesimally small quantities of the second order, the following equation holds for the thawing plane:

$$\kappa \left.\frac{\partial \Theta(x,\tau)}{\partial x}\right/_{x=h} d\tau = \gamma_0 \varepsilon (\Theta_a - \Theta_h) dh, \tag{2}$$

where κ is the mass transfer coefficient and $\Theta(x,\tau)$ is the transfer potential at any point in the thawed zone.

This specific characteristic of the boundary condition on the mobile boundary gives rise to a separate group of thermophysical problems, the theoretical aspects of which have not been ade-

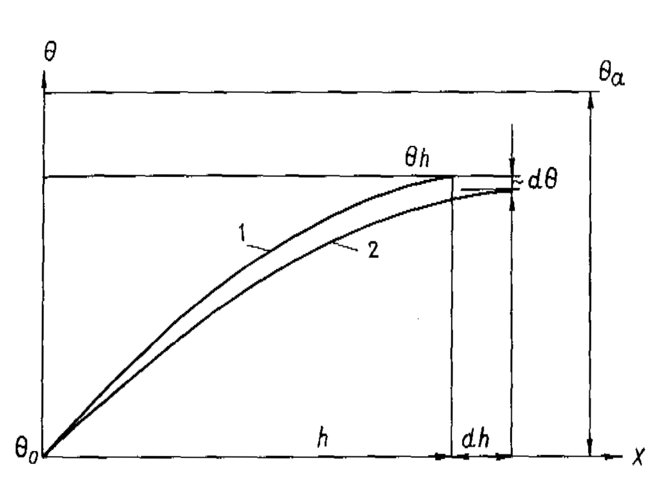

FIGURE 2 Boundary condition at a moving boundary.

quately studied. This group belongs to thermal problems with a mobile boundary, where it is required to determine the law of moisture distribution in porous bodies in a confined area.

The general method of solving such problems in the case of an arbitrary movement of the boundary is very complex, and it is practically impossible to apply it to special problems. However, in a number of cases, which are important in practice (one-dimensional plane problem), thawing is characterized by the movement of the phase boundary described by $\beta\sqrt{\tau}$, where β is a constant coefficient, which depends on the thermophysical properties of the soil and the conditions on the steady-state boundary.

As an example, we shall examine the solution of a one-dimensional problem of moisture transfer. For the sake of simplicity, we shall assume that: (1) the moisture is uniformly distributed throughout the frozen soil, the transfer potential is Θ_a; (2) the moisture transfer in the thawing zone is due to the moisture gradient only; (3) the mass-transfer characteristics during the moisture transfer are assumed to be constant and equal to the average values; and (4) the moisture transfer potential at the stationary boundary during the moisture transfer is taken as constant (Θ_0).

In this case the moisture field in thawing clay soils, where the movement of the thawing plane is described by $h = \beta\sqrt{\tau}$, can be determined as follows:

The moisture field in thawing soils is described by the following equation:

$$\frac{\partial\Theta(x,\tau)}{\partial\tau} = k\frac{\partial^2\Theta(x,\tau)}{\partial x^2};\ \tau > 0;\ 0 < x \leq \beta\sqrt{\tau}. \qquad (3)$$

The identity condition at the stationary boundary is:

$$\Theta(0,\tau) = \Theta_0. \qquad (4)$$

The initial conditons are:

$$\Theta(x,0) = \Theta_a. \qquad (5)$$

We shall first find a solution for the case of evaporation on the ground surface.

If $h = \beta\sqrt{\tau}$, then for a confined area ($0 < x < \beta\sqrt{\tau}$), we shall solve Equations (2)-(5) as follows:

$$\Theta(x,\tau) = C_1 + C_2\ erf\frac{x}{2\sqrt{\kappa\tau}}. \qquad (6)$$

The integration constants are determined from the identity condition. Substituting Equation (4) into Equation (6) at $x = 0$, we obtain:

$$C_1 = \Theta_0,\ \text{then}$$

$$\Theta(x,\tau) = \Theta_0 + C_2\ erf\frac{x}{2\sqrt{\kappa\tau}}. \qquad (7)$$

The moisture-transfer potential at the thawing plane in the thawed zone is described by:

$$\Theta_h = \Theta_0 + C_2\ erf\frac{\beta}{2\sqrt{\kappa\tau}}. \qquad (8)$$

Since $dh/d\tau = \beta/2\sqrt{\tau}$, Equation (2) will assume the following form:

$$\frac{\partial\Theta(x,\tau)}{\partial x}\Big|_{x=h} = \frac{\beta}{2\sqrt{\kappa\tau}}(\Theta_a - \Theta_h). \qquad (9)$$

The gradient of the moisture-transfer potential is found from Equation (7):

$$\frac{\partial\Theta(x,\tau)}{\partial x}\Big|_{x=h} = C_2\frac{e^{-\frac{h^2}{4\kappa\tau}}}{2\sqrt{\kappa\pi\tau}}. \qquad (10)$$

Equating the right halves of Equations (9) and (10), and at the same time denoting $\mu = \beta/2\sqrt{\kappa}$ and replacing $h = \beta(\sqrt{\tau})$, while $\Theta_h = \Theta_0 + C_2\ erf\ \mu$, we obtain:

$$C_2 = \frac{\mu(\Theta_a - \Theta_0)}{\mu erf\ \mu + \frac{1}{\sqrt{\pi}}e^{-\mu^2}}. \qquad (10a)$$

Substituting the values of C_1 and C_2 into (6), we obtain the equation describing the moisture field in thawing soil at any moment of time:

$$\Theta(x,\tau) = \Theta_0 + \frac{\sqrt{\pi}\mu e^{\mu^2}(\Theta_a - \Theta_0)}{1 + \sqrt{\pi}\mu e^{\mu^2}\ erf\ \mu}\ erf\ \frac{x}{2\sqrt{\kappa\tau}}. \qquad (11)$$

If condensation ($\Theta_0 > \Theta_a$) rather than evaporation is taking place on the stationary boundary, i.e., the initial distribution of the moisture-transfer potential in frozen soil is lower than on the soil surface (on the banks of water bodies, condensation in ventilated mine shafts, etc.), the moisture field in thawing soil is described as follows:

$$\Theta(x,\tau) = C_1 - C_2\ erf\frac{x}{2\sqrt{\kappa\tau}}. \qquad (12)$$

Applying the same arguments as before, we obtain the solution of this problem in the following form:

$$\Theta(x,\tau) = \Theta_0 - \frac{\sqrt{\pi}\mu e^{\mu^2}(\Theta_0 - \Theta_a)}{1 + \sqrt{\pi}\mu e^{\mu^2}\ erf\ \mu}\ erf\ \frac{x}{2\sqrt{\kappa\tau}}. \qquad (13)$$

Evidently such a problem can be solved also if the conditions on the stationary boundary differ from those above, but irrespective of the lay of movement of this boundary, Equation (2) must hold.

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CHANGES IN THE COMPONENTS OF THE EXTERNAL HEAT AND MASS EXCHANGE BETWEEN THE EARTH'S SURFACE AND THE ATMOSPHERE AND IN THE METEOROLOGICAL REGIME OF BUILT-UP AREAS IN THE FAR NORTH

L. N. KHRUSTALEV AND V. M. GORBACHEVA Northern Branch of the Research Institute of Foundations and Subsurface Structures, Gosstroi, USSR

Construction of settlements and industrial enterprises changes the natural heat exchange between the earth's surface and the atmosphere. In the Far North, quantitative estimates of these changes make it possible to improve the planning of construction, as well as to increase the accuracy of forecasts of possible changes in the frozen ground within the built-up area and hence increase the reliability of structures erected on permafrost.

One of the main indicators of the dynamics of permafrost within a built-up area is the temperature of the ground surface outside the buildings. It is determined from well-known formulas, which include the following parameters: the radiation balance, the heat used up on evaporation, the turbulent heat exchange (the heat-transfer coefficient), the air temperature, and the height and density of the snow cover.

In a series of special investigations the authors have determined the possible ranges of changes in these parameters in built-up areas in the Far North. The selected area was Vorkuta, where the above changes are especially obvious due to high density of construction, severe pollution of the atmosphere, and large volumes of coal and soil dust on the ground, as well as local climatic characteristics: frequent snow storms that redistribute the snow within the city limits, low air temperatures in winter, and long temperature inversions in the air layers at the surface, which reduce the vertical exchange. The investigations consisted of simultaneous meteorological and balance observations on the most characteristic sections within and beyond the city limits. These observations were supplemented with thermometric, anemometric, and snow survey data. The results were as follows.

Radiation balance. The city affects the components of the radiation balance by weakening the intensity of the total radiation, reducing the duration of sunlight, increasing the absorption of the short-wave radiation, and reducing the long-wave radiation of the Earth. Changes in the components of the radiant heat exchange are due mainly to two causes: the pollution of the city atmosphere with industrial wastes and the changes in the properties of the underlying surface.

Increased aerosol content in the city atmosphere reduces the total radiation by 18-20 percent on the one hand, and increases the long-wave atmospheric radiation on the other, which reduces the effective radiation of the Earth.

In the summer, the reduction in the incoming part of the radiation balance is compensated by the reduction in its outgoing part, and, therefore, the radiation balance as a whole does not undergo any substantial changes. In the fall and winter, when insolation is practically absent, the radiative cooling of the active surface within the city is 15 to 20 percent higher than outside its limits.

In the spring, due to a lower albedo of the snow cover and earlier removal of the latter, the absorption of the solar energy is 1½ to 2 times higher within the built-up area than outside the city limits. This results in an annual increase in the radiation balance within the city of 8 to 10 percent.

Heat losses due to evaporation. From 1 m^2 of the surface within the city and in the tundra may be assumed to be the same. However, the total heat loss due to evaporation is somewhat lower within the city, since a part of the built-up area is covered artificially (asphalt, concrete), on which there is practically no evaporation.

In the summer the evaporation within the city of Vorkuta amounts to 80 percent of that from an equal area in the tundra. There is condensation in the Vorkuta area in winter, which is about the same within the city and in the tundra.

The turbulent heat exchange above the evaporating surfaces is 30 to 40 percent lower in the city than in the tundra. This is due to a greater stability of the air layers at the surface resulting from an overall reduction in the wind velocity within the city and a heat exchange with the walls of buildings.

Modern city planning makes it possible to reduce the wind velocity within a built-up area by 35-40 percent. The wind regime in the city is extremely variable. Within the so-called wind shadow zone next to the buildings, the wind velocity is reduced by a factor of 3 or more. It increases outside this zone, but remains below the wind velocity in an open area. An increase in the wind velocity on an average of 20-30 percent occurs only in the streets whose axes coincide with the direction of the wind.

As a result of the heat exchange with the walls of buildings, the differences in the air temperature within the 0.5-2.0-m layer and above the evaporating surfaces are 1½ to 2 times smaller in the city than in the tundra. Above nonevaporating surfaces (asphalt and concrete), however, they become considerably higher. This is due to overheating of such surfaces, since there are no heat losses on evaporation.

The heat transfer coefficients above asphalt and concrete exceed their values under natural conditions by a factor of 2 or more.

The specific characteristics of the radiation and heat regimes of the active surfaces, as well as the heat released by buildings and other structures, increase the mean annual air temperature within the city limits. Under conditions prevailing in Vorkuta, this is due mainly to the winter maxima of the temperature differences. For example, on clear quiet days in winter the overheating of air within the city may reach 5-6°C (see Figure 1).

To obtain the long-term differences in the air temperatures, we compared the results of periodic observations in the city and in the tundra with the long-term data from the meteorological station at the Vorkuta airport. This enabled us to correlate the air temperatures in the city ($t_{a.c.}$) and in the tundra ($t_{a.t.}$) for different temperature stratifications of the atmosphere:

for superequilibrium,

$$t_{a.c.} = 0.94\, t_{a.t.} + 1.1; \qquad (1)$$

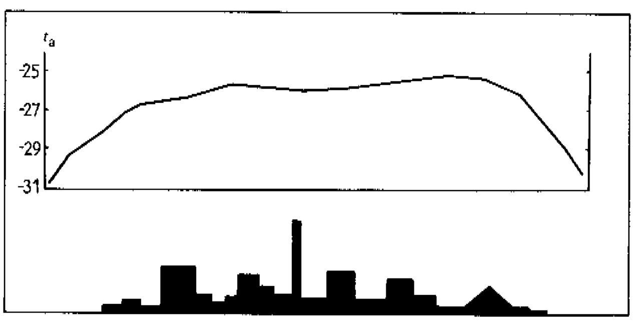

FIGURE 1 Air temperature in Vorkuta and the surrounding area on January 22, 1971 (automobile survey data). t_a = air temperature (°C) at a height of 1.5 m above ground

for inversions

$$t_{a.c.} = t_{a.t.} + 0.065\,(t_{a.t.}) + 0.65. \qquad (2)$$

Equations (1) and (2) were used to calculate the average long-term air temperature in the city, which was -4.4°C. Outside the city limits this temperature was -5.7°C. Hence the mean annual air temperature in Vorkuta is 1.3°C higher than in the tundra. This result is in good agreement with the known examples of the thermal effects of considerably larger cities in the temperate zone. This is mainly due to long heating seasons and large heat losses of buildings in the North.

The snow regime is one of the main factors determining the thermal state of permafrost. Most precipitation in the Vorkuta area occurs in winter, and if the wind velocity exceeds 5 m/s the snow begins to drift. Since the buildings represent artificial barriers, much snow accumulates within the built-up area. Snow surveys indicate that there is 40 to 50 percent more snow within the city limits than on an equal area in the open tundra. Snow accumulations increase towards the center of the city. For example, in the winter 1958-1959, the deduced thickness of the water layer in the snow was as follows: in the center of the city, 243 mm; on the outskirts, 195 mm; and in the tundra, 152 mm. The snow is not uniformly distributed within the built-up area: in the yards and boulevards the snow accumulations are higher, and in the streets and squares, lower, than in the tundra.

All these changes in the components of the external heat and mass exchange and in the meteorological conditions increase the total heat flux to the ground in Vorkuta on the average by 0.16 kcal/cm^2·yr. Together with heat escaping from buildings and other structures, this leads to a rapid degradation of permafrost within the city limits. In the last 25 yr the "cold reserves" of permafrost in Vorkuta have been reduced by 20-25 percent.

FORMULAS FOR THE DETERMINATION OF THE MINIMUM TIMES OF FORMATION AND DEGRADATION OF THE CRYOLITHOZONE

A. A. SHARBATYAN Institute of Water Problems, Academy of Sciences of the USSR

The studies of the origin of the cryolithozone, its history, and the rate of its development must include a comprehensive thermophysical analysis of freezing and thawing of the Earth's crust. Of great interest are the quantitative correlations that make it possible to relate the development of permafrost to all determining parameters. It is especially important to have simple formulas for the calculation of the limiting times of formation and degradation of the cryolithozone, since they provide quantitative answers to the problems of paleoclimatic reconstructions, long-term forecasts, and other aspects of historical cryology.

With this in mind we shall examine the classical problem that has been studied by many authors starting with Krylov.[1] Let there be a semi-infinite body ($z > 0$) with the initial temperature $T(z,0) = T^0 + g_T z$ which is increasing linearly in accordance with the geothermal gradient g_T, and the surface temperature $T(0,t) = T^0 + (T_0 - T^0)\eta(t)$, which changes nonuniformly. Here T^0 is the freezing temperature of soil and $T_0 = const < T^0\eta(t)$ is the Heaviside step function. It is required to determine the position of the freezing boundary at any moment of time from the Stefan equation, and the temperature distribution in frozen and thawed parts of the body from the Fourier equation.

The assumption that the temperature field in both zones is quasi-stationary, which was first made by M. M. Krylov, reduces the problem to the solution of its right half only. The required relationship between the depth of freezing ξ and the time t in a nondimensional form is as follows:

$$\hat{t} = \hat{\xi} - \ln(1 - \hat{\xi}). \tag{1}$$

The symbols used are: $\hat{\xi} = \xi/\xi_0$, where $\xi = \lambda_M (T^0 - T_0)/\bar{q}$ is the limiting depth of freezing for the given temperature difference $T^0 - T_0$, the heat conductivity of frozen soil λ_M, and the heat flux from the interior q; $t_0 = \xi_0/V_0$, where $V_0 = q/\sigma\gamma_d w$ is the limiting rate of thawing of the cryolithozone from below, σ is the latent heat of fusion, $\gamma_d w$ is the moisture content of soil per unit volume, and $\hat{t} = t/t_0$.

Equation (1) is the integral of the Stefan equation $\sigma\gamma_d \frac{d\xi}{dt} = \lambda_M \partial T/\partial z - q$. With allowances for the assumption made, it can be transformed to $d\hat{\xi}/d\hat{t} = 1/\hat{\xi} - 1$, since $\partial T/\partial z = (T^0 - T_0)/\xi$; $\frac{\lambda_M}{q} = \frac{\xi_0}{T^0 - T_0}$. The initial condition: $\hat{\xi} = 0$ at $t = 0$.

Comparing Equation (1) with the Saalschuetz-Stefan formula $\hat{t} = \hat{\xi}^2/_2$, which follows from (1) at low $\hat{t}$, it is possible to estimate the duration of the period P_0 of the maximum rate of increase in the depth of freezing in the case of a nonuniform decrease in the surface temperature to T_0, and also the fraction of the total thickness of the stationary cryolithozone P_0, which is formed in times P_0, $P_0 = 0.05\ t_0$, and $P_0 = 0.3$. Therefore, during P_0, the movement of the freezing front occurs in accordance with equation $\hat{t} = \hat{\xi}^2/_2$. Although this initial period is very short, almost one-third of the total thickness of the frozen zone ξ_0 is formed within this time.

Formation of the remaining two-thirds of ξ_0 requires much more time. Indeed, the $\hat{\xi} = 1$ straight line in the $(\hat{\xi},\hat{t})$ plane is an asymptote of Equation (1). Therefore, to calculate the minimum time $\hat{t}$ of formation of the cryolithozone, it is sufficient to take the admissible error in the determination of its thickness $\varepsilon = \delta\xi_0$ and substitute $\hat{\xi} = 1 - \varepsilon$ into Formula (1), which yields $\hat{t}_m = \varepsilon - 1 - \ln \varepsilon$. In particular, if we take $\varepsilon = 0.01$, then $\hat{t} = 3.61$. It is important to consider that the widely used method of determining the thickness, h, of the layer of annual temperature fluctuations is based on a completely similar principle, but here the factor which limits the desired value of h is the admissible error in the temperature measurement.

The given model makes it also possible to calculate the minimum time of thawing of the entire cryolithozone from below. For this it is sufficient to modify the initial and the boundary conditions as follows:

$$T(z,0)\begin{cases} = T_0 + g_M z, & 0<x<\xi_0, \\ T^0 + g_M(z - \xi), & z>\xi_0, \end{cases}$$

$$T(0,t) = T_0 + (T^0 - T_0)\eta(t),$$

i.e., to assume that at $t = 0$ the surface temperature of the stable cryolithozone with the thickness ξ_0 increased abruptly from T_0 to T^0.

In this case the quasi-steady-state assumption cannot be applied but the problem is easily solved by combining the temperature fields at the freezing front. The deviation of the temperature gradient at $z = \xi_0 - 0$ from the quilibrium value $g_M = (T^0 - T_0)/\xi_0$ is expressed as follows:

$$\Delta g = (2/\xi_0)(T^0 - T_0)\sum_{n=1}^{\infty}(-1)^{n-1}(-\kappa_M n^2\pi^2 t/\xi_0{}^2),$$

where $\tau_0 = \kappa_M t_0/\xi_0{}^2 = \sigma w/c_M(T^0 - T_0)$ is a nondimensional parameter, the numerical value of which at the specific heat $c_M \leq 3$ cal/g°·C, the moisture content $w \geq 4$ percent and the temperature difference $T^0 - T_0 \leq 10.6°C$ (i.e., almost always) is greater than unity; κ_M is the thermal diffusivity of the frozen soils.

Integration of the last equation if $\hat{t} = 0$, $\xi = 1$, yields the desired formula for degradation of the cryolithozone:

$$\hat{\xi} = 1 - \hat{t} + (1/\tau_0) - (2/\pi^2\tau_0) \sum_{n=1}^{\infty} \frac{(-1)^{n-1}}{n^2} (-n^2\pi^2\tau_0\hat{t}), \qquad (2)$$

from which it follows that if $\hat{t} = 1.6\ \tau_0$, the thickness of the frozen zone decreases by $(2/\pi^2\tau_0)\exp(-\pi^2/6)$, while the temperature distribution in it becomes almost nongradiental. It becomes nongradiental at $\hat{t} = 1/3\tau_0$ and remains like that until the cryolithozone has completely thawed from below, since at $\hat{t} = 1/3\tau_0$ for any possible τ_0 the last term in Formula (2) containing the exponential expression becomes negligible.

Therefore, during the degradation of the cryolithozone resulting from an abrupt increase in the surface terperature to the melting point T^0, it is also possible to single out two characteristic periods. These, however, are qualitatively different from the previous periods. During the first period $P_1 = t_0/3\tau_0$, the duration of which is close to P_0, the temperature in the entire cryolithozone levels off to the constant value T^0, while the ratio of thawing from below gradually increases from zero to the limiting value $V_0 = q/\sigma\gamma_d w$. During this time the total thickness of the frozen zone decreases by $P_1 = 1/6\tau_0$, which is usually not more than a few percent, but the "temperature inertia of the cryolithozone" is completely overcome. During the second period, the freezing boundary rises gradually until it reaches the surface of a given massif.

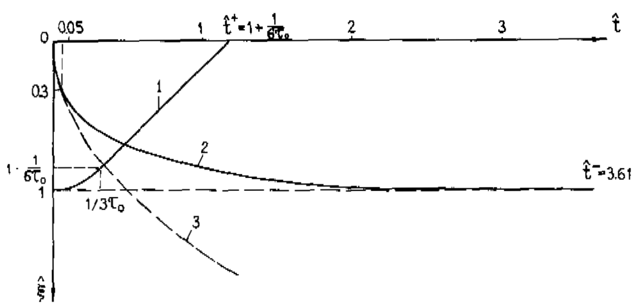

FIGURE 1 Nondimensional relationship between the depth of freezing and time: 1 = during aggradation, 2 = during degradation, 3 = as given by the Saalschuetz-Stefan equation, t = time, and ξ = depth of freezing (Refer to text for meanings of the symbols $\hat{\xi}$, $\hat{t}^+$, $\hat{t}^-$, and τ_0.)

The total duration of the degradation period in such a model is calculated from $\hat{t}^+ = 1 + 1/6\tau_0$, i.e., $\hat{t}^+$ is almost three times lower than $\hat{t}^-$. Hence the minimum times of formation of the cryolithozone are much greater than the minimum times of its complete thawing from below. Figure 1 shows the secular pattern of freezing and thawing of the Earth's crust caused by abrupt changes in the surface temperature--the same with respect to magnitude but with different signs. The hatched line is the $\hat{\xi} = \sqrt{2\hat{t}}$ curve--a nondimensional representation of the Saalschuetz-Stefan formula, which was used in the past to calculate the approximate age of the cryolithozone. The validity of this formula has been questioned repeatedly. Nevertheless one can still find statements in the literature to the effect that the thickness of the cryolithozone is proportional to the square root of time from the start of its formation. For example, the unusually large thickness of the cryolithozone in Yakutia is often explained by its Tertiary age, i.e., by the fact that the process of freezing of the Earth's crust extended over a long period of time.

Figure 1 shows that in this case the theory is far removed from the truth. It is important to note that not only does the Saalschuetz-Stefan formula greatly distort the quantitative results, but that it also leads to qualitatively wrong conclusions, since at a sufficiently large value of t it can yield any preassigned depth of freezing ξ. In actual fact, however, the depth of freezing for any possible values of $T^0 - T_0$, λ_M and q is strictly limited. The same applies to the period of time required to reach this limiting depth. The thermophysical model adopted here and the quantitative relationships that follow from it make it possible to estimate this time with required accuracy.

The suggested formulas are sufficiently simple and reliable and can be used to determine the main limiting characteristics of a nonstationary cryolithozone.

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PART II

Regional Geocryology

N. A. GRAVE, *Editor*

THE REWORKING OF SHORELINES IN THE PERMAFROST ZONE

F. E. ARE   Permafrost Institute, Siberian Division Academy of Sciences of the USSR

The processes leading to the reworking of shorelines in the cryolithozone are characterized by a number of features that are associated with the specific properties of permafrost. This applies particularly to shorelines consisting of ice-rich, unconsolidated Quaternary deposits. Since the reworking of rocky shorelines depends but little on the state of the rock, such shores are not discussed in the present report.

The main motive force in the development of shores consisting of thawed strata is the mechanical energy of water. In the case of frozen materials, an important role is played by heat. In the complex of interrelated processes under whose influence the reworking of shorelines consisting of permafrost takes place, it is necessary to distinguish three fundamental and independent processes:

1. *Thermal abrasion.* The disintegration of the shore zone under the influence of the mechanical and thermal energy of moving water.
2. *Thermal erosion.* The disintegration of the shoreline under the influence of the thermal energy of the air and solar radiation.
3. *Thermokarst.* The thawing of the bottom of a body of water under the influence of the thermal energy of the water, leading to subsidence of the bottom surface.

The thermal abrasion process is accompanied by erosion of the underwater offshore slope and wave-cut bench, by caving of the offshore benches and erosion of the caved blocks, and by removal of the disintegration products by currents of water.[1,11] The mechanical and thermal effects of the moving water on the frozen materials making up the shorelines complement one another. When thawing occurs, this reduces the strength of the materials and thus facilitates their mechanical disintegration. The erosion lays bare the frozen materials, thereby contributing to their thawing.[2]

Generally speaking, thermal abrasion leads to retreat of the shoreline. The development of thermal abrasion in its purest form can be observed on shores consisting of unconsolidated, perennially frozen deposits that have not undergone subsidence and are not subject to thermokarst processes.

In the case of riverbanks, in which mechanical erosion takes place mainly under the influence of a current of water, the process being discussed is appropriately termed thermal erosion. There are no differences in principle, however, between thermal erosion and thermal abrasion, since under river conditions we still have to take into consideration the effect of the waves on the banks, while at sea an important role is played by the migration of disintegration products along the shore.

During the reworking of shorelines, thermal erosion develops on offshore terraces, the stability of which is disturbed by thermal abrasion or thermokarst and is a consequence of the latter. In the course of its development, the seasonally thawed layer disintegrates and outcrops of permafrost are revealed, which undergo intensive thawing. In migrating downward along the slope, the disintegration products advance toward the foot of the offshore terrace and on to the underwater offshore slope. Following the cessation of thermal abrasion and thermokarst, thermal erosion dies down, forming an offshore terrace. On the whole, it leads to flattening out of the offshore terraces and does not give rise to retreat of the shore.

Thermokarst deepens a body of water and thereby contributes to the development of thermal abrasion, but under certain conditions it can give rise to retreat of a shore without the involvement of thermal abrasion. In the pure form, thermokarst (thermal) reworking of shorelines is observed in small bodies of water and also in most of the water bodies of the taiga zone of eastern Siberia, which is characterized by light winds.

In 1965, M. N. Boitsov[8] first drew attention to the possibility of a theoretically infinite thermal reworking of lakeshores consisting of permafrost. V. K. Ryabchun[13] discussed this process as a function of a coefficient proposed by him, which he called the degree of nonuniformity of the conditions of development of a water body.

In the absence of permafrost, the retreat of the shores under the influence of abrasion continues until the profile of the underwater offshore slope[12] is no longer being formed. This situation also holds true in the case of shorelines situated in a permafrost zone, with the sole difference that there the formation of the ultimate profile is complicated by the thermokarst subsidences of the bottom and can only be completed after the thermokarst process has fully died down. Furthermore, the character of the reworking of the shorelines depends radically on the ice content of the materials constituting them. Apparently, when the mean annual water temperature is above freezing, a shoreline consisting of pure ice will retreat indefinitely. On the other hand, the retreat of a shoreline of materials containing no ice is a finite process. Accordingly, for each shoreline situated in the permafrost region there exists a certain critical value of the total ice content of the materials, and, when this value is exceeded, the shoreline is nonstabilizing, and vice versa.

The critical ice content corresponds to the content of the ice in the materials at which their thawing gives rise to subsidence of the surface to an elevation below the level of the water in the water body by a value approximately equal to the minimum depth at which a positive mean annual temperature of the surface of the bottom of the water body is assured. If the level of subsidence is above this elevation, the retreat of the shoreline can take place only when abrasion is involved. If it is below it, thawing of the permafrost beneath the bottom of the water body will inevitably lead to its outcropping at the base of the offshore terrace. Hence, the shoreline is nonstabilizing and will even retreat without the involvement of abrasion.

Obviously, the notion of the critical ice content only makes sense when there is a given level of water in the water body. These two characteristics cannot be considered independently of one another. Sometimes, instead of the critical ice content, it is more convenient to use the notion of the critical water level, understanding this to mean the particular level, above which the shore becomes nonstabilizing, and vice versa.

The possibility of a thermal reworking of a shoreline without the participation of abrasion is determined by the relation

$$H_c - hK - h_w \lessgtr H_w,$$

where H_c is the surface elevation of the land in the zone of the expected reworking of the shore, h is the thickness of the subsiding sediments containing ice, K is the relative subsidence of the sediments containing ice, h_w is the minimum depth of water at which a positive mean annual temperature of the surface of the bottom of the water body is assured, and H_w is the height of the water level in the water body.

If the right-hand side of the inequality is greater than the left-hand side, the shoreline is nonstabilizing, and vice versa.

The rate of development of thermal abrasion, thermal erosion, and thermokarst is evidently determined by climatic factors. Therefore, the processes that lead to reworking of shorelines are characterized by regional characteristics of climatic origin. Permafrost occurs mainly in taiga and tundra zones. The climate of the tundra is distinguished by strong winds and a cold summer. Since the water temperature in the bodies of water is low, in the arctic and subarctic coastal lowlands, for example, a major role in the reworking of shores is played by thermal abrasion. The rates of retreat of the shorelines attain high values. The thawing of the bottom of the water bodies occurs slowly. Closed taliks are widespread, even beneath the bottom of large lakes.[17] The climate of the taiga zone of eastern Siberia is characterized by gentle winds, a warm summer, and pronounced insolation. There, the role of thermal abrasion in the reworking of shorelines is of little consequence; the shorelines retreat mainly under the influence of thermokarst and thermal erosion, the process being much slower than in the coastal lowlands.

Only a very limited amount of data is available on the rate of reworking of the shorelines in the permafrost zone. It is known that the shorelines of the small islands in the open sea, which consist of deposits containing ice, can retreat at a rate of 100 m per year.[10] Observations of the shorelines of the Laptev Sea made by N. F. Grigor'ev[11] and the author indicate that in recent years the average rate of retreat has not exceeded 4-6 m annually. The rate of retreat of the lakeshores in the subarctic coastal lowlands, where thermal abrasion has been actively developing, is as high as 10 m annually.[16] Observations made by the author and N. P. Bosikov of several actively developing thermokarst lakes of central Yakutia have revealed that there, even in areas of very intensive disintegration, the rate of retreat of the shorelines does not exceed 3-4 m annually.

An approximate indicator of the rates of retreat of a shoreline is the shape of the profile of the offshore terrace. Four main types of profile are distinguished: inclined, inclined with a perpendicular lower part, perpendicular, and stepped. Although the last of these gives the greatest opportunity for estimating the rate of retreat of a shore, it is only very rarely encountered.[3]

If the offshore terraces are perpendicular and have wave-cut notches, or if in the vicinity of their foot there are remains of caved blocks and no large accumulations of disintegration products resulting from thermal erosion, this means that the shoreline is rapidly retreating and that the rate of retreat exceeds the rate at which the offshore terrace is disintegrating under the influence of thermal erosion. The latter may be termed the rate of thermal erosion and may mean the migration of the brow of the offshore terrace along the horizontal per unit of time.

If the offshore terraces are inclined or inclined with a lower perpendicular part, there are outcrops of permafrost at their surface, and there are no substantial accumulations of disintegration products resulting from thermal erosion near their foot, then the shoreline is retreating at a rate approximately equal to the rate of the thermal erosion.

Thus, to obtain an approximate estimate of the rate of retreat of shorelines from the shape of the profile of the offshore terraces it is necessary to know the rate of thermal erosion, which can be obtained relatively simply only in the case of shorelines consisting of ice-rich Quaternary deposits that have a high content of regenerating ice wedges.

P. A. Solov'ev[14] has termed such a deposit an "ice complex." Ice complexes are widespread in the lowlands of eastern Siberia and the Soviet Northeast and are characterized by a high content of ice, being of the order of 75 percent by volume, and not infrequently up to 90 percent in the upper layers. When such materials become completely thawed, their thickness decreases by 2-4 times. When reservoirs are being built in the zone of occurrence of an ice complex, their banks may prove to be nonstabilizing, and, in the course of their use, there may be appreciable changes in their capacity. Accordingly, the study of the mechanism whereby shorelines consisting of an ice complex develop, and also the working out of a method of predicting their reworking, are of special interest.

The rate of thermal erosion of an ice complex is in general determined by the rate of thawing of the ice wedges, although during the thermal erosion process the thawing of the outcropping country rocks may somewhat outstrip or lag behind the thawing of the ice.[3,4] The rate of thawing of the ice can easily be calculated from the climatic characteristics. To this end, in particular, it is possible to use the formulas worked out by V. T. Balobaev[7] or F. E. Are,[4] which were derived for the purpose of calculating the rate of melting of naleds. Regime observations on Mostakh Island in the Laptev Sea reveal that on the mainland coast of Yakutia and the coastal islands the rate of thermal erosion of offshore terraces consisting of an ice complex is 4.5 m annually, regardless of their exposure and steepness.[6] Similar observations on the shorelines of three actively developing lakes in central Yakutia, carried out by the author in 1971, showed that in this region the rate of thawing of perpendicular outcrops of regenerating ice wedges in the offshore terraces is 4.0-5.5 mm for each degree above freezing of the mean diurnal air temperature, depending on the exposure of the outcrop, which corresponds to 7.0-9.5 m annually.

Quantitative data on the rate of reworking of shores consisting of permafrost can be derived not only thorugh a study of the morphology of the shorelines or long-term observations of their retreat, but also by using the geothermal method. In the course of the reworking of the shorelines, the permafrost is located beneath the bottom of the water body or river and begins to thaw out. The depth to which it thaws depends on the length of time that the area in question has been underwater. After determining the thickness of the talik at a particular point by means of drilling or through the use of geophysical methods, it is possible to make a rough calculation of the length of time that would be required for the thawing of the permafrost to this depth under subaqueous conditions. On dividing the derived result by the distance of the selected point from the shore, we obtain the rate of retreat of the shore during the period of thawing.

Determinations of this type were made by S. V. Tomirdiaro[15] for certain lakes of the lower Anadyr' lowland, and also by the author and N. P. Bosikov in central Yakutia. In 1971, some holes were drilled close to the shore on the bottom of Lakes Oner and Syrdakh. These holes reached the permafrost. In the case of Lake Oner, the hole was drilled at a point located 18 m from the shoreline, where the depth of the water was 1.1 m and the height of the offshore terrace was 30 m. The thickness of the thawed materials proved to be 8.3 m. On the Lake Syrdakh the hole was drilled at a point 5 m from the shoreline where the depth of the water was 0.3 m and the height of the offshore terrace was 16 m. Permafrost was encountered at a depth of 8.2 m from the bottom. An approximate calculation, which took into consideration the subsidence of the bottom and the changes in the water temperature with time, revealed that the talik took 32 yr to form in Lake Oner and 40 years in Lake Syrdakh. Here, the mean annual rates of retreat of the shoreline were 0.6 and 0.1 m respectively.[5]

The determination of the rate of retreat of the shoreline by the geothermal method is an example of the additional opportunities that are offered by geocryology for the study of shoreline processes.

Methods of predicting the reworking of shorelines consisting of permafrost have not as yet been worked out. Studies along these lines are only just beginning.[9,16]

In this connection, mention can be made of certain possibilities of predicting the reworking of shores of small bodies of water, in whose disintegration thermal abrasion does not play an important role. Currently, there are no reliable data for determining the limiting sizes of such water bodies. In the case of the northern coastal lowlands, it is possible to cite an approximate figure of 1 km. Observations of the reworking of the shores of the actively developing lakes in central Yakutia revealed that thermal abrasion has only a minor role in the case of water bodies in which the "fetch" is at least as great as 3.5 km.

For water bodies such as these, on the basis of F. E. Are[1] and the results of engineering-geological surveys, it is possible to find the limiting value of the retreat of a shoreline, keeping in mind that under the conditions being considered it is only a nonstabilizing shore that can retreat. To this end, a tracing is made of the profile of the shoreline zone, on to which is plotted the outline of the surface which is expected after the ice-containing deposits have completely thawed.

The point of intersection of this surface with the line of the water level in the body of water, depressed by the value h_w, gives the limiting position of the water's edge. As a rule, the rate of thermal erosion is higher than the rate of thermal reworking of the shore. Thus, in the subarctic coastal lowlands the annual rate of retreat of the shorelines of small bodies of water does not exceed ~5 m, and in central Yakutia, 10 m. In order to define these figures more precisely, in each specific instance it is possible to make use of the results of observations of the rate of retreat of lakes under similar natural conditions. This procedure was used by the Permafrost Institute of the Siberian Division, USSR Academy of Sciences, to compile a forecast of the reworking of the shorelines of one of the reservoirs to be included in the projected system of transferring water from the Amga River to the Tattu River in central Yakutia.

In conclusion, we would note that in order to work out reliable methods of predicting the reworking of shores consisting of permafrost, it is necessary to arrange for comprehensive regime observations on the shores of various water bodies and rivers. Here, the principal research tasks are as follows: Ascertaining the specific features and mechanisms involved in the formation of the limiting profile of an underwater offshore slope in conjunction with the development of thermokarst subsidences of the bottom; determining the erodibility coefficients of various types of permafrost; studying the heat exchange between water masses and outcrops of permafrost; and determining the possibility of using the specific features and properties of the frozen materials making up the shorelines for the purpose of working out methods of predicting their reworking.

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THE MAIN FEATURES OF THE STRUCTURE AND DEVELOPMENT OF THE PERMAFROST OF THE WEST SIBERIAN PLAIN

V. V. BAULIN, G. I. DUBIKOV, AND YU. T. UVARKIN Operations and Research Institute for Engineering Site Investigations

The western Siberian Plain is a unique region, in which the latitudinal and zonal variations of the permafrost are exhibited to the maximum extent. Based on an analysis of the geological and paleoclimatic events, it is possible to reconstruct the process that let to the perennial freezing of the deposits in Quaternary times. On the other hand, owing to its high degree of thermal inertia, the permafrost has preserved in its structure traces of large-scale climatic changes in the past and thus serves as a valuable aid in gaining an understanding of the complex history of the region's development. Important assistance in this regard is also rendered by the traces of the existence of permafrost, which have been preserved in the Quaternary deposits. After specific corrections have been applied, the established mechanisms that resulted in the development of the frozen ground in western Siberia can be extended to other regions.

The modifications that have occurred in time and in space with respect to various characteristics of permafrost are the result of both zonal (e.g., climatic, paleoclimatic, and geobotanical zoning) and regional (e.g., geological and geomorphological structure, the composition of the ground, and groundwater) factors. Frozen ground occupies the greater part of the western Siberian Plain, trending in the north-south direction over a distance of more than 1,000 km.

DISTRIBUTION

In the tundra and forest-tundra zones situated in the northernmost part of the lowland, frozen ground underlies almost all of the land surface in an almost continuous distribution. It is even to be found in the shallow regions (up to 1.5-2.0 m in depth) of the Ob' Gulf.[6] The existence of permafrost is also possible in the shallow regions of the Kara Sea shelf, where under normal water salinity conditions (35 percent) the temperature of the benthic sediments may be as low as -1.5°C to -1.8°C.

Within this zone, open taliks are only possible beneath the beds of major rivers and large lakes and in regions of tectonic faults. Closed taliks measuring up to 30-100 m in thickness exist beneath most of the thermokarst lakes and rivers.[9] At latitude 66°-67°N., occasional taliks whose number and dimensions increase towards the south are to be found among the frozen massifs, while southward of the 63d and 64th parallels it is preeminently peatlands that constitute the permafrost. The thawed islands are most often formed in areas consisting of sandy soil, which is due to the seepage of warm atmospheric precipitation.

The southern boundary delimiting the occurrence of permafrost outlines the peatlands, among which are to be found isolated mineral hummocks that have long-existing frozen cores. According to recent data,[3] this boundary runs along the lower course of the right-hand tributaries of the Ob' (B. Yugan, B. Salym, *et al.*).

TEMPERATURE OF THE PERMAFROST

The mean annual temperature of the ground (at a depth of 5-15 m) ranges from -9°C or -10°C to 0°C in the north-south direction. South of the Arctic Circle, in the floodplains of the rivers, there are areas overgrown with thickets, which consist for the most part of ground with above-freezing temperatures. Within the forest-tundra and taiga zones, in the dry sandy terraces, it is usual for the temperature of the ground situated at the surface to be above freezing. The temperature and occurrence of the permafrost also depends on the geobotanical characteristics of the territory and on the microtopography, which affect the distribution of the snow cover. On the north of the plain, in the valleys of the shallow creeks and in the deep gulleys where the snow collects, the temperature of the ground increases by 2°-3°. Between latitudes 65°-67°N, almost all of the treeless areas consist of ground with temperatures as low as -3°C-4°C, while the forested areas are made up of thawed or high-temperature frozen ground.

In these latitudes depending on the mean annual soil temperature, the different areas can be divided into forested (pine, larch) sandy areas, consisting of thawed soil; forested (spruce, larch) clayey-silt areas, consisting of frozen soil with a temperature of 0°C to -1°C; and treeless (peatland) areas, consisting of frozen soil with a temperature as low as -2°C.

The variation in the climatic conditions in the western and eastern parts of the lowland, in particular, and the increase in the thickness of the snow cover from west to east, leads to a rise in the mean annual soil temperature in this direction.

A specific feature of the temperature distribution in the frozen soils of western Siberia is the existence of a broad zone (300-400 km) in which the temperature of the soil ranges from 0°C to -1°C or -2°C. The appearance of this zone is due to the increase in the mean annual air temperature that has occurred since the end of the last century and to the "zero curtain effect."

There is very little information on the temperature of the ground situated below the depth of zero annual amplitude. An analysis of geothermal observations of the permafrost of western Siberia makes it possible to distinguish four main types of temperature distribution curves:

1. The temperature of the ground gradually increases until the base of the permafrost is reached (curves with a positive gradient). Curves such as these are obtained for the Ust' Port region and Tazovskii settlement. Curves of this type are typical of regions situated north of the 66th and 67th parallels, but can even be found in the treeless areas of the more southerly regions.

2. There is almost no gradient in the temperature distribution of the ground lying between the surface and the base. In this case it is usual for the temperature values to be close to zero. They are highly typical of the areas lying to the south of the Arctic Circle. This type of temperature distribution is indicative of a decrease in the "reserves of cold" in the permafrost.

3. The temperature distribution in the frozen ground has a negative gradient (the vicinity of Mys Kamennyi and the town of Salekhard). These values can be found throughout almost the entire northern part of the lowland, although they are most typical of the polar regions, where temperature variations can occur within the negative range of values. These curves indicate that the climate has begun to ameliorate.

4. The fourth type of temperature curves consists of curves with a variable gradient. They are found in areas where the frozen ground has a two-layered structure and in which only relics of the permafrost have been preserved. The gradient changes from being positive in the upper frozen layer to zero in the intrafrozen talik and in the relic permafrost. Such a complex temperature curve was formed as a result of the influence of climatic variations that continued for different lengths of time. The temperature of the frozen ground in the upper layer ranges from 0°C to -2°C or -3°C; in the subjacent thawed strata, it is about zero; and in the relic frozen layer, it ranges from zero to -0.5°C or -0.6°C.

ICE CONTENT OF THE PERMAFROST

The ice content of the permafrost and its cryogenous structure are closely linked with the contemporary climatic conditions of the Plain and its paleoclimate. The main part of it (covering approximately 80 percent of the region) was typically epigenetic in its formation. The late Pleistocene and Holocene continental and coastal marine sediments froze syngenetically. Their thickness is as much as 10-15 m, and the distribution is confined to the tundra zone of the lowland.

The structure of the thick epigenetic permafrost is complex. It varies in space and is chiefly determined by the lithological composition of the constituent rocks, the degree of diagenesis attained by the sediments by the time they became frozen, and also by the presence in the section of aquifers. The freezing of the homogeneous suglinoks that had passed through the stage of diagenesis in the course of their accumulation was accompanied by appreciable ice segregation only in the upper 10-15-m layer of the permafrost and by a rapid decrease in the ice content of the ground with depth. The structure of the suglinok deposits that froze syngenetically under the conditions presented by the constricted lagoons is specifically cryogenous. In this case, the process that led to the freezing of the deposits fulfils the role of a major lithogenetic process. Frozen ground with reticulate or ataxitic texture and an ice content of up to 40-60 percent is formed from fluid or cryptofluid sediments.[5] The total moisture content of the frozen lagoon sediments exceeds that of the liquid limit of the soil. On rare occasions, the explanation for the high moisture content of the mineral parts (25-40 percent) of the deposits lies in the specific features of the deposits themselves (a high content of organic remains that retain water) and in the time relationship between their accumulation and freezing. The ice-rich suglinok deposits of the lagoons, with thicknesses of up to 10-20 m, occur in the northern part of the Yamal and Gydan peninsulas within the upper Pleistocene marine terraces. The cryogenous structure of the frozen deposits of nonuniform composition (the interstratification of suglinok and sandy varieties) is complex in form. Their freezing was accompanied by the formation of several ice-containing horizons of great thickness at various depths (up to 150-200 m), which is attributable to the presence of the aquifers in the sections.

The ice-containing syngenetic frozen deposits of the northern part of western Siberia were formed between the end of the Kazantsevian and the present time from alluvial, alluvial-lacustrine, lake-swamp, and coastal marine sediments. The climatic changes that occurred in the Holocene greatly decreased the area of occurrence of syngenetic and epigenetic permafrost and altered their relationship in favor of the epigenetic type of permafrost.

Syngenetic frozen deposits of Pleistocene age occur to the north of the region between the 68th and 69th parallels, where their thickness attains 10-15 m. To the south of this latitude, there are shallow, syngenetic horizons consisting of late Holocene floodplain and lake-swamp deposits.[2] In some areas of the coastal part of the lowland, where there is an unstable tectonic regime in the upper Pleistocene, multistage polygenetic permafrost was formed (the valleys of the Yuribei and Gyda rivers), in which there are two horizons of syngenetic frozen remains.

In the polar regions of the lowland, underground ice (regenerating ice wedges and sheet formations) is widespread. According to their type of growth, syngenetic and epigenetic regenerating ice wedges are distinguished. The southern boundary of occurrence of the syngenetic ice wedges runs to the north of latitude 68°. This ice is concentrated in the deposits of terraces situated above the tidal flats, and in the lake-swamp and coastal marine sediments of the Kazantsev Plain. The height of the ridges can be as much as 8-12 m; the width, 2.0-2.5 m. The epigenetic ice wedges are present in mineral soils (to the north of latitude 67°-68°) and in peatland formations (to the north of 65°N) in all of the topographic elements. The height of the epigenetic wedges is 2-5 m; the width, 1.0-1.5 m. Sheet-ice formations have a major role in the structure of the permafrost of western Siberia. These are mainly concentrated on the Yamal and Gydan peninsulas. It has been found from studies carried out in these regions that the thick (up to 10-20 m) sheets and lenses of ice in marine Pleistocene deposits lie at depths ranging from several to 200 m and have a horizontal extent of up to 300 m. They form the largest masses of ice in western Siberia. The sheet-ice formations are associated with the shallow coastal marine clays, suglinoks, and sands, which are nonpersistent and frequently alternate in section. The partial thawing of the rocks during the optimum period and their subsequent freezing has had a decisive effect on the development of the cryogenetic structures.

As is well known, for the frost-*cracking* of soils and the formation of regenerating ice wedges, a combination of certain necessary conditions must be met; the soil must be sufficiently monolithic and there must be sufficiently high temperature gradients in that part of the sediments lying near the surface.[4] The first of these conditions is most likely to be met in the areas consisting of ice-rich mineral soils and in peatlands; the second, in the areas where there are low mean annual soil temperatures (about -4°C to -5°C in western Siberia). Depending on the composition of the ground and the magnitude of the temperature gradients, the frost cracks form polygons measuring from 10 m or 25 m to 100 m and greater. In accordance with their origin, the regenerating ice-wedge formations are divided into two types: syngenetic and epigenetic. The visible height of the syngenetic wedges can be as much as 8-10 m; the maximum width, 2-4 m. The dimensions of the epigenetic wedges are somewhat smaller. Their height ranges from 1-2 m to 5-6 m, and they can be up to 1-2 m in width.

An analysis of the mechanisms that gave rise to the distribution of the regenerating ice-wedge structures and of the alteration in the recent climatic and paleoclimatic conditions that accompanied their formation makes it possible to subdivide the permafrost region into four zones. In the first (northernmost) zone, extending to latitude 70°-71°N, ice of both types if formed in the mineral soils and peatlands. In the second zone, extending to 68°-69°, the regenerating ice wedges continue to form mainly in the peatlands. Although this type of ice has been preserved in the mineral soils, it develops in exceptional cases only. Further to the south (to latitude 65°-66°N), only isolated regenerating ice wedges are known to have occurred in mineral soils. While spreading ice formations can be encountered in the peatlands to the north of the zone, in the south, the formation of ice wedges has ceased. In this zone, pseudomorphs after regenerating wedges are to be seen, which frequently occur in the more southerly regions of the plain in deposits of the various stratigraphic horizons of the Pleistocene.

In Nature there are two main causes leading to the *frost heaving of soils*: First, the migration of water toward the freezing front and the formation of ice layers; second, the infilling of large quantities of water or saturated soil during freezing of the closed spaces in the ground and the formation of injection-type ice masses or ice-and-soil masses.

Migration heaving can take place in soils in which moisture is capable of being drawn toward the freezing front, given the presence of an adequate reserve of intrasoil moisture or the possibility of its being drawn in from without. Such conditions are most frequently observed during the freezing of argillaceous soils situated near bodies of water or above aquiferous horizons. The origination of closed taliks and their rapid freezing, both of which are necessary for injection expansion, are most probable when the mean annual temperatures of the ground in the floodplains of the rivers are low and during freezing of the sublacustrine taliks. These are the conditions that have predetermined the main zonal mechanisms involved in the distribution of the heaving processes at the present, their development in the past, and the morphological peculiarities of heaved structures.

Strictly speaking, all permafrost has been subject in varying degrees to migration heaving. The nonuniform composition of the soils, however, the uneven distribution of moisture in them, and the varying conditions under which the ground became frozen have led to considerable differentiation between the heaving processes. It is the peaty and clayey soils that are most subject to heaving. Not infrequently, they are clearly outlined in the topography totally as a result of heaving (the so-called areas of heaving). They can be as much as several meters high. The thermokarst erosion processes lead to the disintegration of the heaved mass and the formation of flat-hummocky and very hummocky peatlands. A well-defined zonation has long been discerned in their distribution:[7,8] the flat-hummocky peatlands are confined to the south of the tundra zone and to the forest-tundra; the very hummocky peatlands, to the northern taiga subzone.

The majority of the migration heaving hummocks occur in regions of epigenetic soil freezing (to the south of latitude 68°-69°N), i.e., where the freezing of the ground occurred after the climatic optimum was reached. Quite frequently, however, they also occur in the more northern regions. Favorable conditions for the contemporaneous development of heaving hummocks exist where ground freezing is taking place in the floodplains of rivers, etc. The contemporary heaving hummocks are most frequent in areas where the mean annual temperature ranges between -1°C to +2°C and -3°C to +5°C. The height of the hummocks varies between 3-5 m and 10-15 m and, in general, increases in a south-north direction. The diameter of the base of a hummock can be as much as 100-200 m or more; in plain view, the hummocks are sometimes elliptical.

As a result of the formation of injection-ice bodies, open-system pingos form close to the surface. Outwardly, these are difficult to distinguish from large heaving hummocks. The main difference between them lies in the structure of the hummocks. While open-system pingos contain an ice core, migration-heaving hummocks consist of ice-rich soils containing ice layers. The height of the open-system pingos can be as much as 20-25 m or more. In the upper reaches of the Yarudei River, there are hummocks up to 35 m in height. As a rule, the contemporary open-system pingos are confined to areas with a fairly low mean annual temperature (below -3°C to -4°C). The open-system pingos that were formed during the colder ages in the past occur further south (to latitude 66°-65°N), in the zone of discontinuous permafrost.

Mention must be made of the fact that in many regions a concentration of heaving hummocks is noted within the confines of tectonic uplift. This is due, first, to fine-grained sediments, which are highly susceptible to frost heaving being present as outcrops or situated close to the surface, and, secondly, to the highly saturated condition of these sediments.

THERMOKARST PROCESSES

Thermokarst occurs almost everywhere in the permafrost zone of western Siberia. The most favorable potential conditions for the formation of thermokarst depressions exist in the regions where the soils have a high ice content, i.e., for the most part in regions where there has been syngenetic freezing of the deposits and in peatlands. The territories where the thermokarst processes are most widely developed are thus the regions of accumulation of upper Pleistocene and Holocene sediments, situated north of latitude 67°-68°N. Here also, however, the thermokarst processes are confined to definite geomorphological levels (chiefly to the third and fourth terraces), which, because of the colder paleoclimatic conditions of sediment accumulation, have resulted in a higher ice content of the upper group of sediments. If the distribution of the contemporary thermokarst structures is traced from north to south, it becomes possible to discern yet another general rule. Between the southern boundary of the permafrost and latitude 68°-70°N, the number of thermokarst lakes gradually increases, whereas in the Far North of the plain, where the climate is very severe, they decrease in number.[9]

Depressions that were formed as a result of the melting of segregation ice are widespread in western Siberia. They have a rounded or elongated shape and a length (or diameter) ranging from 10-20 m to 1,000-2,000 m. As a rule, their depth does not exceed 1.5-3.0 m. During melting of the large concentrations of sheet ice, subsidence-basin lakes of up to 10-15 m in depth and 100-200 m in diameter form. The warming of the climate that accompanied the Holocene climatic optimum led to a marked regional acceleration of thermokarst processes and to an aggregate subsidence of the levels consisting of ice-rich soils.

Since between latitude 63°-64°N and 68°-69°N the sediments of the alluvial terraces evidently froze syngenetically and their ice content rose to 40-60 percent, it can be presumed that, as a result of the thawing that occurred during the optimum period, the levels may have subsided several meters.[1]

THE THICKNESS AND STRUCTURE OF THE PERMAFROST

The depth of the base of the permafrost is determined by the heat-exchange conditions existing at the Earth's surface (both currently and in the past) and also by the heat flux from the Earth's interior to the base of the permafrost. Other things being equal, the most favorable conditions for the plutonic freezing of the rocks occur in the interfluves confined to the tectonic depressions.

In conformity with the wide variation in the conditions of heat exchange at the Earth's surface, the thickness of the permafrost decreases from 400-500 m to 50-100 m and less in the north-south direction. Drillholes sunk in the vicinity of Ust' Yeniseisk port and in the lower course of the Taz and Nida rivers have revealed the presence of permafrost measuring about 500 m in thickness. Permafrost thicker than 300 m has been found to be widespread on the Salekhard and Kazantsev uplands of the Yamal Peninsula (according to the findings of an expedition of the Faculty of Geology, Moscow State University).

As a rule, the thickness of the permafrost decreases with movement away from the interfluves and toward the low-lying terraces that border the major rivers (such as the Ob' and Yenisei); on the floodplains at latitudes 66°-68°N, it is as little as 100-150 m.

The analysis of the permafrost distribution patterns reveals a pronounced increase in its thickness from west to east, which agrees completely with the decrease in the heat flux density in that direction. Whereas in the Urals section of the lowlands at latitudes 63°-65°N the depth of perennial freezing does not exceed 200-300 m (according to the regional geological administrations), in the Yenisei section it is as much as 400-500 m. It is in the eastern part of the lowlands that the permafrost is seen to extend furthest towards the south (as far as latitude 59°-58°N, according to A. A. Zemtsov), whereas in the western part it only reaches 61°-62°N.

The changes that occur in the structure of the permafrost with depth are a result of the paleogeographic features of the lowland during Quaternary times. The warming of the climate that took place during the optimum period, the partial thawing of the ground that occurred south of latitude 68°-69°N, and the subsequent freezing of the deposits had a decisive effect on the basic features of the frozen ground. In those regions (lying chiefly north of 66°-67°N) where there was only a moderate degree of thawing during the optimum period, the new freezing restored the monolithic nature of the permafrost with depth. Further south, the thawed ground did not freeze completely in all places, so that along with the monolithic permafrost a two-layered structure of the permafrost is also widespread. The so-called relic frozen layer was formed, which attests to the major climatic changes that occurred in the past. It has remained to this day, even in those regions in which permafrost is absent at the surface (between latitudes 62°-63°N and 59°-60°N).

The depth to the permafrost table of the relic layer increases from 50-80 m at latitude 66°-67° to 150-200 m at 59°-60°N, and the thickness ranges from 100-150 m to 200 - 300 m. According to available data, the temperature of rocks from this layer is not less than -1°C. Also affecting the existence of the relic frozen layer, in addition to the climatic and paleoclimatic conditions, is the composition of the ground and the vegetation. The relic permafrost is best preserved in deposits with a low permeability.

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BASIC FEATURES IN THE CRYOGENESIS OF THE EUROPEAN PLAINS IN THE UPPER PLEISTOCENE

A. A. VELICHKO The Institute of Geography, Academy of Sciences of the USSR

For the study of general questions relating to the dynamics and origin of permafrost, the European area is of special interest. At present, the European area, with the exception of some very small zones located in the north of Scaninavia and the European part of the USSR, and also in the high Alps, is free from the influence of processes associated with permafrost. Nevertheless, at certain times in the past, beginning with the middle Pleistocene, and especially in the upper (late) Pleistocene, not only the mountains but also the vast European plains were in a permafrozen state. During this period, Europe and the northern half of Asia constituted a single cryogenous region, which seemed to emphasize the unity of some of the climatic and physical-geographic features of the European mainland. The subsequent existence of two extreme states--the total predominance throughout all of northern Eurasia of a region of permafrost and the subsequent falling away from this region of the entire western half of the mainland (Europe)--is a phenomenon that continues to await a fully substantiated explanation. It may be that the search for the causes of such striking changes in the structure of the cryogenous region will being to light properties of the process of cryogenesis as a whole.

A fairly extensive body of data has now been assembled with respect to the fossil and relic ancient permafrost formations that have been discovered and described in various regions of Europe.

The so-called periglacial phenomena in Poland have been described in papers by Dylik,[16,17] Yahn,[19] Moiskii,[9] and Ol'khovik-Kolyasinskaya.[14]

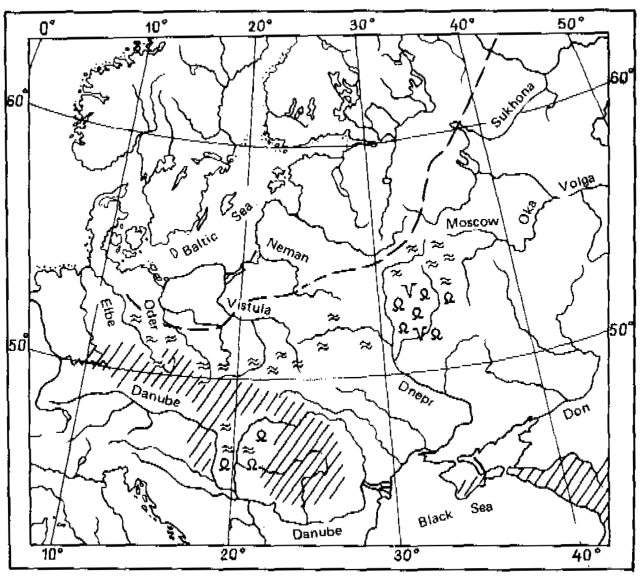
In Hungary they have been described by Pecsi. There are also data relating to the area comprising East Germany (Lembke *et al.*[8]) and Czechoslovakia (Demek,[15] Sekyra[23]). The cryogenous structures of France have been reported in papers by Tricart and Cailleux,[24,25] and also by Guillien,[18] while those in Belgium have been described by Paepe.[21] Some observations on cryogenous structures in East Germany, Czechoslovakia, Poland, Hungary, Austria, Belgium, and France have been obtained by the author during geological field trips in those countries

There are also extensive publications by various authors that relate to the European part of the USSR.

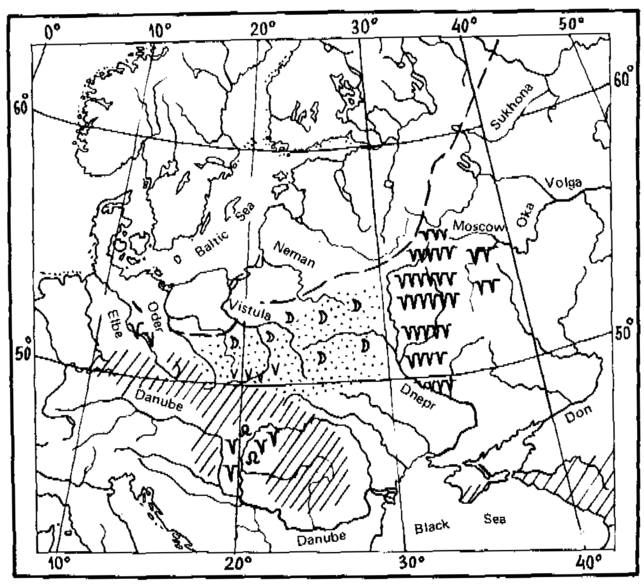
First and foremost among these, mention must be made of the papers by A. I. Moskvitin,[10] Yu. M. Vasil'ev,[2] Yu. V. Krylkov,[7] D. L. Nazarenko,[11] M. N. Grishchenko,[6] V. V. Berdnikov,[1] and also those of the author.

There also exists certain cartographic information reflecting the actual distribution of cryogenous formations. These include maps of ancient cryogenous phenomena with respect to Poland[16] and Hungary;[22] an atlas of the periglacial phenomena of France;[24] an outline map of locations of ancient cryogenous structures;[13] and a drawing illustrating the distribution of a relic cryogenous morphosculpture on the Russian plain.[3] Moreover, together with Berdnikov,[5] the author has made small-scale drawing illustrating the occurrence of cryogenous formations in the Valdai (Würm) epoch in the area comprising central and eastern Europe (Figure 1).

Currently, an attempt is being made to use the existing data as a basis for making a small-scale schematic map that would cover a more extensive area, embracing the principal European plains regions, the cryogenesis of which, as distinct from the mountainous regions, is more closely and di-



(a)



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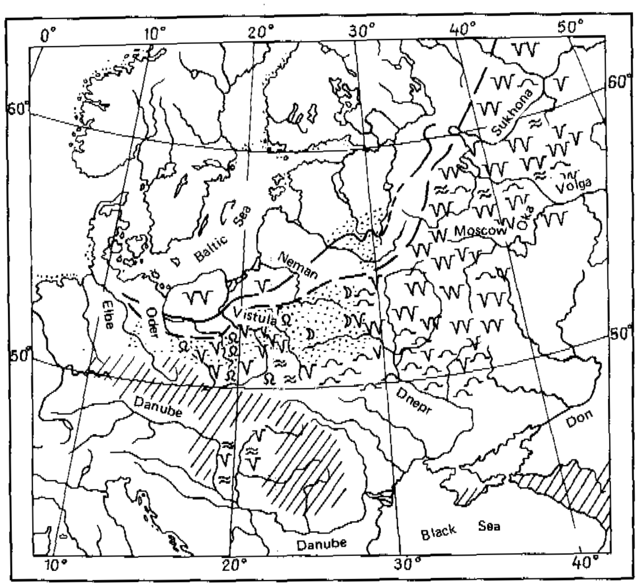

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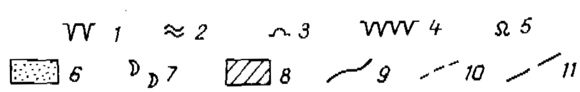

FIGURE 1 (a) Distribution of cryogenous structures dating from the early Pleistocene, (b) late Bryansk, and (c) late Pleistocene. 1--ice wedges and polygons bounded by ice wedges; 2--solifluction structures; 3--cryoturbations and other structural disturbances; 4--deformations of the Bryansk fossil topsoil; 5--thermokarst structures; 6--regions containing sand; 7--dunes; 8--mountainous regions containing a complex of periglacial-permafrost structures; 9--limit of the Valdai glacier in the Bologov stage; 10--southern limit of the early Valdai permafrost zone; 11--limit of maximum extent of the Valdai glacier.

rectly linked with the general climatic changes. Despite their inherent shortcomings and simplifications, which indeed are characteristic of any first attempt when large areas of land are involved, these schematic maps will nevertheless make it possible to reach certain general conclusions, a critical analysis of which, it may be hoped, will facilitate the compiling of more complete maps and charts through the efforts of scientists of various countries.

Since when ascertaining the main cryogenous features of the European plains, a major role is attached to their relationship with the general climatic changes, it is important to establish the chronological-stratigraphic order of succession in which the cryogenous processes occurred and their correlation with the main natural events in the late Pleistocene.

In solving the chronological-stratigraphic problems, a more positive role was played by the information derived from studying the structure of the upper Pleistocene loess and fossil soil beds of Europe, since it is within these beds and their facies-related deposits that the numerous and diverse ancient cryogenous structures are localized, by virtue of which they have obtained a secure stratigraphic and age position.[20]

An analysis of the cryogenous processes at particular chronological and stratigraphic levels has made it possible to distinguish three main cryogenous stages in respect of the upper (late) Pleistocene. Furthermore, the great horizontal extent of these beds provided an opportunity of discerning certain spatial variations in the manifestations of the cryogenesis within the individual cryogenous stages.

THE FIRST (EARLY VALDAI) STAGE OF CRYOGENESIS

In the loess section, the characteristics of this stage were recorded in a soil profile belonging to the Mikulin (Riss-Würm) glaciation and the beginning of the Valdai (Würm) glaciation (the Shtillfried-A profile in Austrai, the Naumberg profile in East Germany, the RK-III + RK-II profile in Czechoslovakia, and the Mezino profile in the European part of the USSR). An early (loess) phase and a late (steppe) phase are distinguished in its structure. The extent to which the entire upper part of the soil profile is disturbed by deformations led us to the view that the early Valdai cryogenous stage originated immediately after the formation of the second--steppe--soil phase had ended. On the basis of the relationship between the structures belonging to this stage, the cultivated layers at the halting places in southeastern France, and the existing dates as to absolute age, the approximate limits of the early Valdai cryogenous stage can be set at between 65,000 and 60,000 yr ago.

On the plains of eastern Europe, frozen structures dating from this period extend southward to latitude 53°N, while in central and western Europe, they occur in East Germany, Poland, Czeckoslovakia, Hungary, Belgium, Austria, and France, reaching latitude 47°-48°N. This initial upper Pleistocene wave of cryogenesis was characterized by its own main type of permafrost formation: Throughout the entire area, the formations consisted chiefly of solifluction and cryoturbation deformations, combined into a single zone. Within this zone it is possible to trace at least two provinces: (1) a western province, extending from France to the western regions of East Germany and to the Czechoslovakian SSR (whereas here, the above-mentioned structureless deformations completely predominated; in the area comprising France, it is only the solifluction phenomena that are distinctly recorded); (2) an eastern (central-eastern European), in which together with the processes of plastic deformations there is evidence of a process involving the formation of small, elementary ice wedges measuing 15-20 cm wide in the upper part and 1.0-1.2 m high. It is typical that in the European part of the USSR, north of latitude 55°-56°N, this combination of cryogenous phenomena is accompanied by evidence of soil movement along the slope, not only in the spring, but also in the autumn, when blocks of the frozen upper parts of the seasonally thawed layer were subject to creep.

In general, the early Valdai cryogenous stage was characterized by the preeminent development of structureless deformations, by the absence of pronounced geomorphological manifestations in the form of specific topographic features within a gently undulating framework, by the high moisture content of the layer of seasonal freezing and thawing, by the moderate (0.5-0.8 m) thickness of the seasonal layer, and by the small temperature gradients.

THE SECOND STAGE OF CRYOGENESIS

This is the middle Valdai (middle Würm) stage. The onset of cryogenesis is seen to have occurred at the end of the bryansk (Shtillfried-B) period of soil formation and is well recorded in the loess sections of many European countries. The conclusion of the stage occurred at the beginning of the second post-Bryansk phase of loess accumulation. The period in which the cryogenous processes developed was about 25,000-22,000 yr ago.

In comparison with the early Valdai stage, the climatic conditions under which this stage of cryogenesis took place were more severe, and the area of the permafrost region extends to latitude 49°N. The following are the specific spatial peculiarities of the cryogenesis. In the latitudinal direction, two provinces are distinguished. During this period, as in the first stage, beginning with the area comprising Volyn-Podolia in the European part of the USSR, Poland, Czechoslovakia, East Germany, and farther west, solifluction and, in part, cryoturbation processes predominated. These, however, were much more active and dynamic than those in the early Valdai stage. Of the other structures, only rudimentary cracks 0.2-0.3 m high and 5-8 cm wide were sporadically noted there (for example, in Belgium).

The eastern European province differed significantly and qualitatively from the western

European province during this stage. It was characterized by the extensive occurrence of fissure-forming processes, although these were on a small scale (the length of the sides of the polygons ranged from ~2.0-2.5 m, the heights of the wedges was ~1.5 m and they exhibited a two-layer structure). This polygonal fissuring was accompanied by processes of plastic soil flow, which arose as a result of the dynamic stresses in the layer of seasonal freezing and led to the formation of clay boil type structures that were quite clearly defined in the surface microtopography.

Another typical feature of the middle Valdai cryogenous stage consists of the very weak meridional variability of the cryomorphic processes throughout the entire eastern European province. The morphotype, i.e., the clay boils, predominated between the central Ukraine and the region lying to the north of Moscow (the Vladimir Oblast'). Its distinctiveness, however, varied in the north-south direction, there being a worsening toward the north, since in this direction the thickness of the seasonally thawed layer decreased (from 1.0 m to 0.5-0.3 m), along with an accompanying increase in the dynamic processes in the dynamic processes in the soil near the surface.

In general, the moisture content of the soil in this stage remained high, especially in the western province. Toward the east it diminished and there was a rise in the temperature gradient in this direction, which reached a maximum when systematic fissure formation began, even though this was on a very small scale.

THE THIRD STAGE OF CRYOGENESIS

This is the late Valdai (late Würm) stage. The upper parts of the cryogenous structures are in contact with the top of the loess strata. An approximate estimate of the time segments in which this stage existed is between 15,000 and 10,000 yr ago. The third, most recent, stage was characterized by very complete and active development of cryogenesis. It was during this period that the cryogenous region reached its southernmost limit of 45°-46°N, and the climatic conditions leading to cryogenesis became most severe and attained their maximum continental extent. All the European plains came to be characterized by polygonal crack formation processes in which the "normal" size of the polygons was 20-40 m and ice wedges were formed. Under the gently undulating conditions that obtained (with the exception of areas with a high rate of sedimentation, for example, floodplains), these latter were preeminently of the epigenetic type.

Judging from the map of this stage, certain features were noted in the structure of the late Valdai cryogenous zone that connect it with the contemporary Siberian region of permafrost. Thus, in eastern Europe, in the north-south direction, there were three broad belts whose features resemble the northern, central, and southern zones of the permafrost region in Siberia.

In the west-east direction, the plains of Europe were subdivided into the following four provinces, in which cryogenesis underwent modificiations, depending on the increasing continentality in this direction: (1) a western European, (2) a central European, (3) a western Russian, (4) an eastern Russian.

There is reason to suppose that in the late Pleistocene the overall structure of the European cryogenous zone was characterized by considerable originality. First, it is probable that in western and central Europe the mountainous and flatland regions of permafrost were interlocking, thereby forming a single area. In eastern Europe, on the other hand, there was a hiatus between the permafrost regions on the plain and in the mountains (the Caucasus and in the Crimea). There, these regions were most likely divided by the area situated in the lower regions of the Dnieper and Volga and characterized by processes of deep-seated seasonal freezing (the third latitudinal belt in the western Russian province). It is as if the belt of seasonal freezing in eastern Europe had become wedged into the zone of Pleistocene permafrost. On the whole, in comparison with the preceding stages, the third and latest European stage of cryogenesis was characterized by the lowest moisture content of the soil under the prevailing gently undulating conditions (especially in the eastern regions) and by the highest temperature gradients, which coincided with the greatest thickness of the seasonally thawed layer (2-3 m).

The available information pertaining to the development in Europe of permafrost processes in the upper Pleistocene affords a basis for making a comparative analysis of the various stages of cryogenesis and leads on to certain general concepts.

In the upper Pleistocene in the European plains, cryogenesis was dissimilar both in time and in space. At the present level of understanding, it remains difficult to arrive at a fully substantiated classification of the types of cryogenesis that existed in the late Pleistocene in Europe, the more so if geophysical and quantitative criteria are to be used. There are even grounds for assuming that today, complete analogue types that existed in antiquity are lacking. In particular, in the case of the ancient Valdai stage, there are no analogues such as high soil moisture, small gradients, and structureless deformations. This type can be provisionally called the Atlantic type of cryogenesis. As its opposite, the eastern Siberian type can be distinguished (abruptly continental conditions and large temperature gradients). Between these extreme types, it is possible to distinguish an intermediate-transitional type.

In that case one can speak of the fact that in the ancient Valdai stage, a greater degree of uniformity of the cryogenous conditions was typical of the European plains, although at the level of the gentlest, Atlantic type of cryogenesis (Figure 2). The differences between the west and the east did not overstep the bounds of the same type, although some increase in continental extent is to be seen toward the east.

The European cryogenous region was characterized by a different type of structure during the middle Valdai stage of cryogenesis. Here, fairly

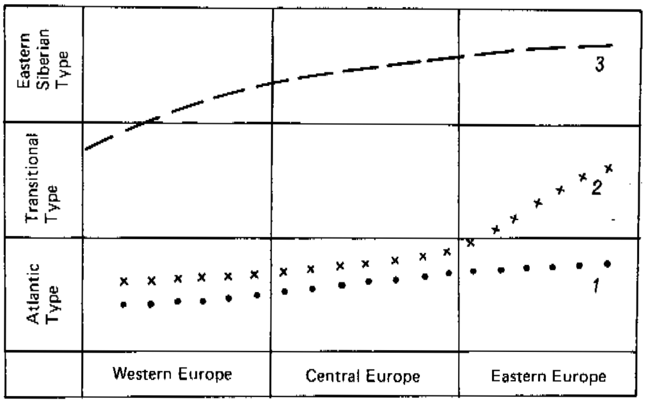

FIGURE 2 Variation in the structure of the permafrost region of the European plains in the (1) early, (2) middle, and (3) late Valdai stages.

marked differences between the western and eastern regions are to be seen. Whereas the western regions have remained at the level of the Atlantic type (even though the processes in the seasonally thawed layer are characterized by increased activity), the eastern regions fall within the more severe type of cryogenous conditions, which can be termed transitional. The conditions of this type are such that the processes leading to plastic deformations of the soil are accompanied by polygonal cracking, although this is rudimentary.

Another typical feature of this stage, which has yet to be explained, is the absence of a succession of cryogenous morphotypes in the meridional direction, in other words, the simplified morphoclimatic structure of the cryogenous region as a whole.

Finally, in the third, late Valdai stage, some equalization of the cryogenous conditions again appears, although here it is at a qualitatively different level than in the first, early Valdai stage. The development of cryogenosis in the late Valdai in the European plains is patterned on the eastern Siberian type, although within the cryogenous region of Europe there also existed differences associated with the increase in the degree of continentality from west to east. It should be noted that the use of the term "eastern Siberian type" must not be taken to mean a complete analogy, since undoubtedly, the moisture content of the ground in eastern Europe was lower than in the coastal lowlands of eastern Siberia today.

The analysis of the development of cryogenesis in the late Pleistocene of Europe is suggestive of a discreteness (discontinuity) of this process with time. Evidence in support of this notion is to be seen in the presence of three well-defined, independent cryogenous horizons (stages). Did permafrost exist in the intervals between them, during the main periods of ice accumulation? If this was the case, then the dynamic processes accompanying it were in an extremely depressed state. This depression could have been caused by the high aridity of the climate and the very low ice content of the soil, rather than by unfavorable temperature conditions.

On the other hand, the analysis of the cryogenesis being discussed reveals the direction they have taken, both in time and in space. Between the early and late Valdai, the area of cryogenesis became greater, and at the end of the Valdai it had advanced farthest towards the south. Throughout the Valdai stage there was also an increase intensity and severity of the cryogenesis, so that at the end of the Valdai there arose, not only territorially but also structurally, a single cryolithozone. This was the vast cryogenous region of the whole of northern Eurasia, the area of which, equalling approximately 22 million km^2, was almost double the area of the contemporary permafrost on that continent (approximately 12 million km^2), while its limit, in contrast to the contemporary submeridional limit, occupied a latitudinal position.

We shall now turn to one of the most important moments in the history of the cryogenesis of Europe and Eurasia as a whole. During the changeover from the Pleistocene to the Holocene, the vast cryogenous region, which had only reached the peak of its development, both spatially and in content, right at the end of the Pleistocene, began to undergo very rapid degradation (Figure 3). According to palynological data, accompanied by radiocarbon datings,[12] for the most part the degradation of the permafrost in Europe could have embraced a period of 1,000-1,500 yr. This rapidity of the degradation process was close to being catastrophic, while the rearrangement that occurred throughout the entire cryogenous region by the falling away from it of Europe was of a radical nature. Naturally, the rapidity of this degradation and its areal dimensions gradually diminished from west to east with increasing proximity to the region of contemporary permafrost, which can be regarded as the remaining, preserved portion of the vast Eurasian cryogenous region of the late Pleistocene.

The question that naturally arises is: What is at the root of the unique process whereby the late Pleistocene region of Eurasian permafrost underwent progressive development and degradation? We have already drawn attention to the fact that

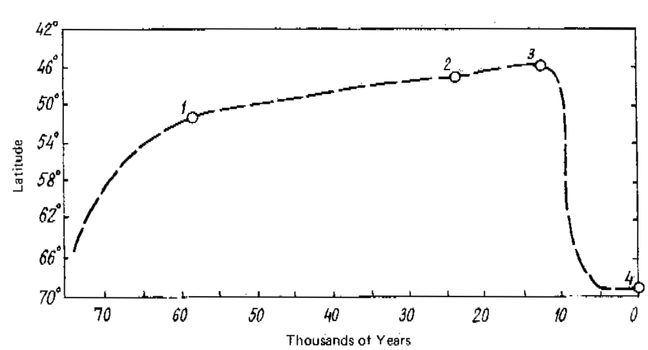

FIGURE 3 Changes in the position of the southern limit of permafrost in eastern Europe in the upper Pleistocene period. 1--early Valdai stage, 2--middle Valdai, 3--late Valdai, 4--contemporary stage.

this process cannot be directly linked with the history of the continental ice sheet and that the explanation for it must be sought primarily in the overall climatic changes of the late Pleistocene.[4]

It was in the late Pleistocene that the onset occurred of the specifically continental, cold conditions throughout the entire extratropical expanse of the Northern Hemisphere (the indigenous stage of cryogenesis). This led to a very marked expansion of the area occupied by sea ice, to its southward advance and to rearrangement of the surface structure of the hemispheric mantle, and also of the climate, since the ocean, being covered by ice for a long period, could not be the source of solid precipitation in Europe. As is well known, in the late Pleistocene the overall climatic changes were superimposed by a major regression (northern 100 m) of the sea, which was likewise conducive to an increasingly continental type of climate. Presumably the vast dry expanses of the present-day shelf were also an arena for the development of permafrost.

The overall climatic warming, with which is associated the beginning of the holocene, led primarily to the degradation of the thin sea ice, which reacted very quickly everywhere, even to minor variations in the air temperature (this is a well-known phenomenon in the history of the Holocene). Together with the retreat of the sea ice and the general warming trend there was also a change in the structure of the Earth's mantle: the European continent began to receive war Atlantic air masses carrying precipitation, with the result that rapid degradation of the permafrost began.

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COMPREHENSIVE REGIONAL MAPPING OF PERMAFROST

Exemplified by A. V. VOSTOKOVA   Moscow Lomonosov State University

The basis of comprehensive mapping consists of the development of indices reflecting physico-geographical natural phenomena as parts of a single entity and revealing the relationships among these phenomena. Comprehensive mapping ensures that a complete and many-sided description is given of the nature of the individual regions. During the past decade, comprehensive regional atlases of natural conditions and natural resources have been compiled.

In the mapping of northern and northeastern regions, permafrost is an important element of the natural framework. The most suitable way to generalize permafrost research is to present it in the form of a series of geographic maps, which can give a complete picture of this phenomenon and serve as a reference aid when resolving regional planning problems, developing engineering and economic construction designs, and estimating future development prospects. A series of permafrost maps will make it possible to convey the fullest amount of information on the properties of the permafrost and the variability of its qualitative and quantitative characteristics in relation to the sum total of the geological and geographic conditions. The inclusion of permafrost maps in a comprehensive atlas incorporating a series of multipurpose maps of the natural framework (such as geological, geomorphological and vegetation maps) opens the possibility of visually tracing the two-way effect and relationships between permafrost and other natural phenomena.

In the mapping of the ground it is essential to specify its structure and properties and special features that are linked with the thawing and freezing processes. Where permafrost is concerned, it is preferable to provide comprehensive descriptions to reveal the diversity of the zonal and regional variations. In small-scale maps it is possible to distinguish only the main characteristics, the sum total of which gives some idea of the specific features of the permafrost conditions in the region being mapped. The most important of these are the following: the pattern of distribution of the permafrost; the temperature state, thickness, cryogenous structure, and permafrost relief; and also the characteristics of seasonal freezing and thawing.

The compilation of a series of interrelated permafrost maps gives rise to major difficulties, which stem firstly from the insufficiently high quality of the source information and its consequent low reliability; and secondly, from the fact that little progress has as yet been made in working out the problems associated with the small-scale mapping of permafrost.

As a rule, the author's contribution to deciding upon the content of permafrost maps is made by specialist geocryologists who are well acquainted with the region being mapped. When preparing the maps for publication a whole series of cartographic problems arise which call for timely and effective solutions. These problems are as follows: evaluating the source information being used to ensure the correct positioning of the phenomena; working out the content and preparing the geographical bases, which are specially drawn up for each map in relation to its specific subject area; determining the amount of information to be conveyed by the maps and the procedures to be followed in plotting the legends; making recommendations relating to the choice of the methods of cartographic representation, and also their integration on the same map; developing the map compilation techniques to increase their reliability and ensure that the permafrost maps are in agreement with one another and also with maps portraying other kinds of natural phenomena; making recommendations relating to the coloring of the map series.

A number of problems relating to the small-scale mapping of permafrost were worked out in the course of integrated studies of the Tyumen' Oblast' by specialists from Moscow University.[2] During the comprehensive mapping of the Tyumen' Oblast' the system of interrelated small-scale maps of the natural setting was arranged to include a series of maps providing a detailed representation of the features of the permafrost in this Oblast'.*

The area constituting the Tyumen' Oblast' is characterized by a wide diversity of permafrost conditions, ranging from the northern regions, where there is a predominance of continuous permafrost with low temperatures and of great thickness, to the southern regions, where freezing is absent from the upper layers of the ground and occurs at great depths from the surface. To provide for the mapping of the main permafrost characteristics a system of qualitative and quantitative indices has been worked out, as well as recommendations pertaining to the most logical modes and means of cartographic representation (see Table).

The characteristics of the permafrost in the Tyumen' Oblast' are depicted in nine coordinated maps. The following are the most important of these:

* The maps were compiled by Moscow University's Faculty of Geography and are printed in polychrome. The authors are A. I. Popov and N. A. Shpolyanskaya; the editor, A. V. Vostokova.

TABLE 1 Characteristics and Method of Depicting Permafrost in Comprehensive Atlases

Cartographic Element	Qualitative Indices	Quantitative Indices	Mode of Representation	Means of Representation
Permafrost distribution	1. Degree of continuity or discontinuity (a) in area; (b) with depth	3. Distance (in m) of the permafrost table from the ground surface	Qualitative background	Range of half-blue color shades
	2. Southern limit of permafrost		Linear marks	Numerical indices
Thickness of the permafrost		1. Thickness (in m)-- (a) latitudinal variation in thickness; (b) differences in thickness depending on regional natural factors	Isolines (of scale of gradation in thickness)	Shading of varying intensity and design between the isolines
Temperature of the permafrost		1. Temperature (in °C) at the depth of zero annual amplitude: (a) latitudinal variation in °C; (b) differences in °C depending on regional natural factors	Isolines (of scale of gradation in °C)	Coloring of varying intensity between the isolines or a variable design of the isolines
Cryogenous structure of the ground	1. Genetic types of underground ice 2. Types of cryogenous textures 3. Age of cryogenous rocks		Qualitative background	Range of half-blue color shades Color shading Indexes

	4. Mode of formation of cryogenous rocks 5. Lithological composition of cryogenous materials		Qualitative background	Explanatory notes in the legend Shading design
Seasonal freezing and thawing	1. Region of seasonal freezing and seasonal thawing 5. Lithological composition of the layer of seasonal freezing and thawing of the ground	2. Depth of freezing and thawing (in m) 3. Times of initial and maximum freezing and thawing 4. Range of temperatures during period of freezing and thawing at various depths	Qualitative background Isolines (of scale of depth gradations) Data localized by stations Localized graphs by stations Qualitative background	Range of half-blue color shades Tinting of background Shading of varying intensity between isolines Variously colored numerals Graph of temperature curves Shading of variable design
Permafrost relief	1. Distribution of permafrost relief forms (and combinations of them) by reference to the genetic principle 2. Dynamics of development of the permafrost forms		Areas	Conventional symbols for the various relief forms Variable coloring of the symbols
Dynamics of the permafrost	1. Development and state of the permafrost during the various geological time stages: active, stable, degradational		Qualitative background	Range of half-blue color shades

1. permafrost (distribution, temperature and thickness);
2. cryogenous structure;
3. seasonal freezing and thawing;
4. permafrost relief.

Of the remaining maps we would mention especially those depicting the development of permafrost during the different geological time stages, as they emphasize the specific nature of the development of the permafrost process in the area constituting western Siberia.

The content of the maps, their completeness and reliability, and the degree of detail with which the permafrost characteristics are represented, are to a large extent determined by the existing level of knowledge concerning the area and by the availability and quality of the sources of information. As to the completeness of content of a map, and the principles to be followed in the distinguishing of various features on it, as well as the relationships between them, all this can usually be ascertained from the legend.

The development of the legend is the first step in compiling a map. In the case of new, original maps for which there is no single system of symbols, the plotting of the legend is a complex process, especially when compiling a series of maps in which it is important to allow for their comparison and comparability.

A clear idea of the content of the map and the degree of correlation between the indices can only be obtained when the legend has been successfully plotted. This plot provides for a suitable arrangement of the symbols and their explanatory superscripts in a definite system, grouping and sequence, which makes it possible to separate the main content of the map from its secondary content. If necessary, the legend is supplemented by explanations of the terms and by diagrammatic aids for use when working with the map. The classification of the various indices in the legend and the reflection of their degree of coordination makes it possible to distinguish the main divisions of the legend and its lower categories. The plotting principles and graphic form of the legends of permafrost maps of dissimilar subject areas are largely dependent on the specific nature of the permafrost conditions within the area in question. The content of the permafrost maps of a proposed subject area includes a combination of qualitative and quantitative indices. The content and graphic form of the legends are determined by the distribution characteristics of the indices within the region.

The diversity of the cryological conditions within the Tyumen' Oblast' (distribution of the permafrost by area and with depth, cryogenous structure, temperature distribution and thickness, etc.), is such that in the legends of the different types of maps, the specific nature and range of the individual properties of the permafrost can be shown to the full, as can their interdependence and dependence on regional natural factors, all of which is indicated graphically.

The representation of the distribution, temperature, and thickness of the permafrost in the Tyumen' Oblast' is combined on the 1:4,000,000-scale map (Figure 1). The plotting of the map legend is based on distinguishing three major regions, which differ in the character of occurrence of contemporary and relic permafrost in the vertical section: continuous (continuous distribution with depth), discontinuous (separated by thawed layers), and deep-lying (freezing absent at the surface and occurring at great depths). The presence of these regions, successively alternating in the north-south direction, is a common feature of the permafrost distribution in western Siberia.[1,4]

Each of these regions is characterized by the spatial relationship between the frozen and thawed areas, by the temperature distribution and by the thickness of the strata. The zone of continuous permafrost is characterized chiefly by a continuous areal distribution, which becomes discontinuous only in the southern part, but with a marked predominance of permafrost. The discontinuous region is characterized by the sporadic distribution of permafrost, the relationship between the frozen and thawed ground changing towards the south of the region, where permafrost is observed only in the form of scattered islands, for the most part in organic soils. Within the region, deep-lying relic permafrost is found only in the form of isolated patches and as pereletoks. The areal permafrost distribution pattern is indicated on the map by a colored background in the case of contours drawn to the map scale, whereas for regions containing isolated patches of permafrost and pereletoks it is denoted by symbols not drawn to scale.

To provide for representation of the temperature regime of the permafrost within the above-named regions, the following geothermal zones have been distinguished, each of which is characterized by specific temperature indices: a northern and a southern arctic zone, a northern and a southern subarctic zone, and a northern and a southern boreal zone. Also taken into consideration were certain natural factors that have a special effect on the temperature regime of the ground. Despite the flat nature of the Oblast', the ground temperatures within the zone change appreciably, depending on the degree of geomorphological confinement of the terrain: in the valleys the temperature is found to be higher, while in interfluves it is lower. Besides being caused by the relief, in certain zones the variation in the temperature regime is due to lithology. In the watershed areas in the subarctic zone, for example, the temperature is 1° to 2° lower in organic soils than in mineral soils. Therefore, on the map and in the legend the influence of the topography is reflected by distinguishing a watershed and valley type of permafrost within each temperature zone. The mountain type of permafrost with its characteristic ground temperatures is specially indicated. Because the map is crowded, the relationship between ground temperatures and lithology is shown by explanatory notes in the map legend for each geothermal type. The lithological composition of the earth materials is subdivided into organic-sedimentary, mineral-sedimentary, crystalline, and metamorphic rocks.

In order to maintain unified principles of

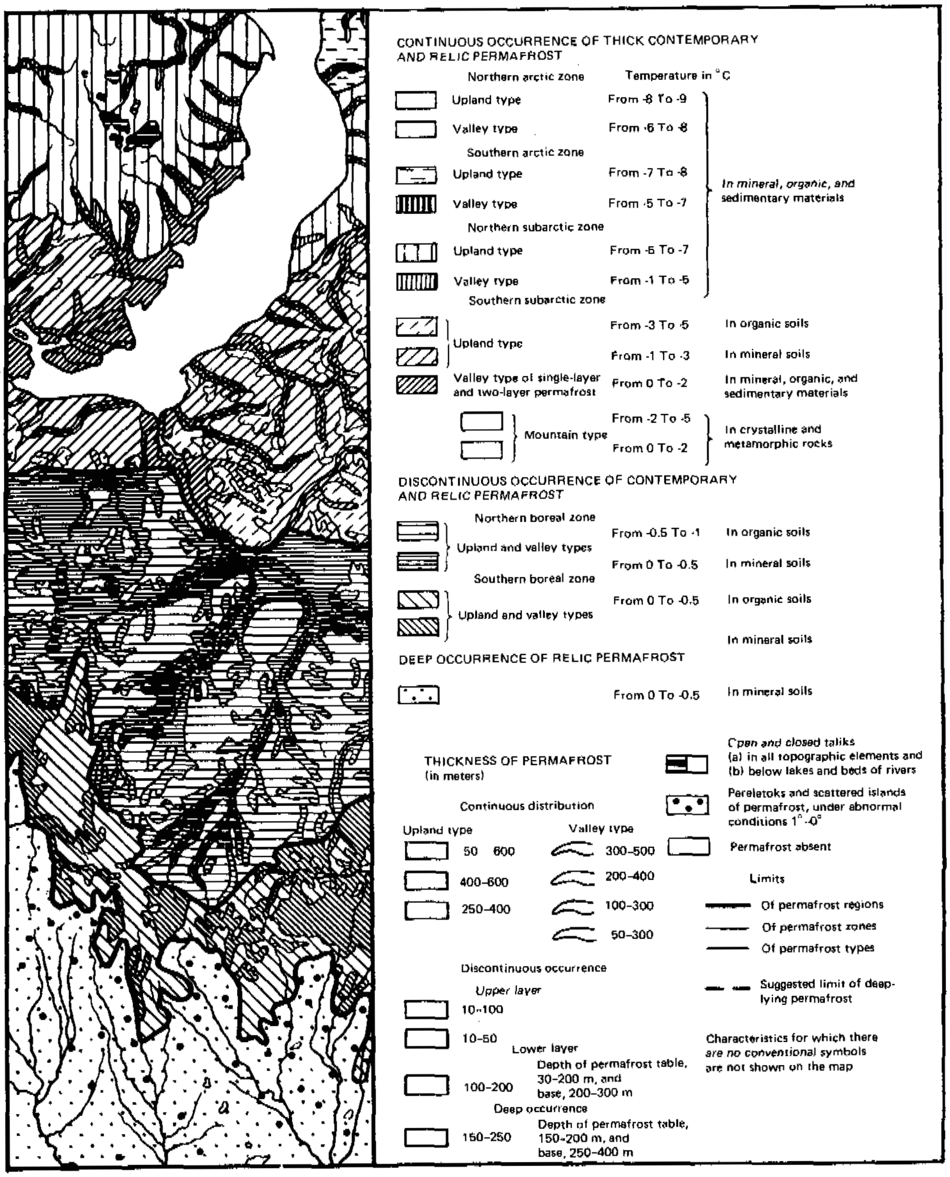

FIGURE 1 Permafrost (distribution, temperature, thickness).

representing the permafrost characteristics on the same map, the thickness categories are shown in the legend. This is done by plotting the thickness scales separately for each of the three previously distinguished regions of continuous, discontinuous, and deep-lying contemporary and relic permafrost. Furthermore, in the continuous permafrost zone, differences in the thickness values, which depend on the relief, are quite clearly traced within a zone. This fact gave rise to a requirement for the plotting of separate thickness scales for the upland and valley types of permafrost as well. In view of the two-layered nature of the permafrost in the discontinuous zone, different thickness scales were also compiled that characterize the upper (contemporary) and lower (relic) layers of permafrost. Unfortunately, the sparseness of the available information makes it possible to give only a very approximate picture of the thickness of the lower layer. The thickness indices of the lower layer are denoted in the legend by numerical values indicating the distances of the permafrost table and base from the ground surface.

The above described method of mapping the distribution, temperature and thickness of the permafrost can also be used and adopted for the compilation of similar maps for areas where the permafrost conditions display well defined zoning features. Here, the legends of such maps must reflect quite distinctly the specific permafrost features in the region being mapped, besides which, the factors giving rise to the formation of the permafrost as a whole and of its various characteristics must stand out sharply. In the case of other regions, where as a result of the limited areal extent or special conditions of permafrost development there is only moderate evidence of zoning, the use of simplified legends is possible, for example, those with a single scale of thickness.

The principal features of the cryogenous structure of the ground in the Tyumen' Oblast' are shown on a separate map, with a scale of 1:6,000,000. The specific nature of the cryogenous structure is determined by the distinctiveness of the ground ice distribution.[3] The cryogenous earth materials are characterized on the map by a number of indices: by the spatial arrangement of the genetic types of ground ice and cryogenous character of the ground; by the mode of freezing of the ground (epigenetic or syngenetic); and by the age of the ice formations and lithological composition of the enclosing materials. The plotting of the legend (Figure 2) is based on the subdivision of the cryogenous materials into two groups: cryoliths (monomineral rocks), and cryolithites (ice-containing polymineral rocks in which the ice shows up as one of the mineral components). But cryoliths and cryolithites are not always well-defined formations in the ground. They sometimes occur in combination. Therefore, a third group is distinguished in the legend: cryoliths in combination with cryolithites.

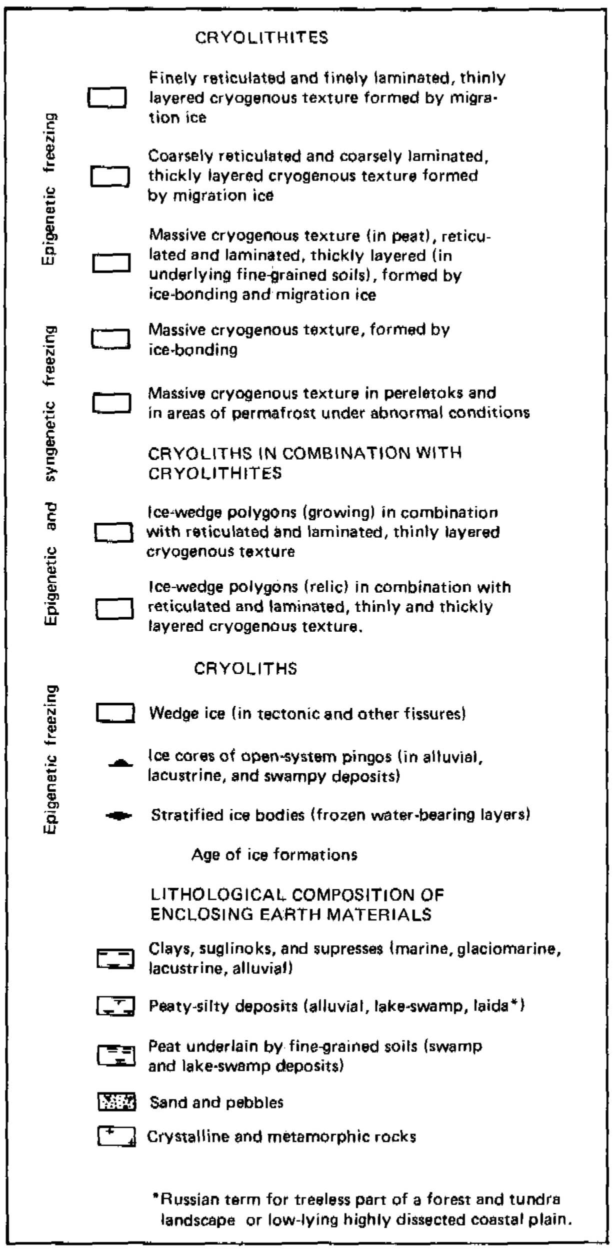

FIGURE 2 Cryogenous earth materials.

All ground ice formations are divided into a series of major genetic types. Cryoliths are made up of ice-wedge polygons, forming chiefly in fissures of crystalline and metamorphic rocks, ice cores of open-system pingos, and the stratified ice bodies of frozen water-bearing layers. With cryolithites, the migration type of ground ice and ice bonding are typical. In the cartographic representation of ice-containing polymineral soils (cryolithites), a decisive role is played by the characteristics of the ground ice. Three main types are distinguised--layered, reticulated, and massive--although most often it is conbinations of these that are observed within the same area. The cryogenous characteristics are concretely defined by the dimensions of the ice lenses. For example, finely laminated (or finely reticulated), thinly lensed, and coarsely laminated (or coarsely reticulated), thickly stratified earth materials are indicated separately on the map. The cryogenous nature of the ground and the genetic types of ground ice are regarded as integral groupings, reflecting the distinctive features of the cryogenous structure of the ground of the region in question. They are, therefore, indicated without prior division.

In the 1:4,000,000 map a description is given of seasonal ground freezing and thawing. The plotting of the legend (Figure 3) is based on the subdivision of the area into regions of seasonal thawing (the northern part), seasonal freezing (the southern part), and a region in which there is both seasonal thawing and seasonal freezing (the central part). The main index, characterizing the layer of seasonal freezing and thawing, is its depth, measured in meters. The depth of freezing (thawing) is extremely variable from place to place and depends on a diversity of geological and geographic conditions.[4] Therefore, the description of variations in the depth of seasonal freezing and thawing is especially important, having regard to the influence of the main physical factors. In the Tyumen' Oblast' a special influence is exerted on the variation in depth of seasonal freezing and thawing by lithology and by variations in relief (uplands and valleys).

Accordingly, the depths of freezing and thawing are shown in the main lithological groups, i.e., in the mineral and organic soils, and in the uplands and valleys. The description of the layer of seasonal freezing and thawing is augmented by data on the times when freezing and thawing begin and attain a maximum. For this purpse the 10-day period of the month in question is indicated. The range of temperatures at various depths during the period of freezing and thawing is indicated by localized graphs, which are plotted from the results of long-term observations at meteorological, and permafrost, stations.

The permafrost relief is indicated on a separate 1:6,000,000 map: The content of the legend is based on a description of the types of permafrost relief. These are distinguished by their genetic characteristics, namely, polygonal relief, heaving hummocks, and relief caused by solifluction and nivation (Figure 4). The typical relief forms in each genetic type are distinguished by conventional symbols. Besides describing the distribution of these forms of relief, and the dynamics of their development are indicated.

The permafrost relief, which is formed when the ground becomes frozen and ice accumulation occurs (the process of aggradation), undergoes important modifications. The alternation of the heat-exchange conditions results in the thawing

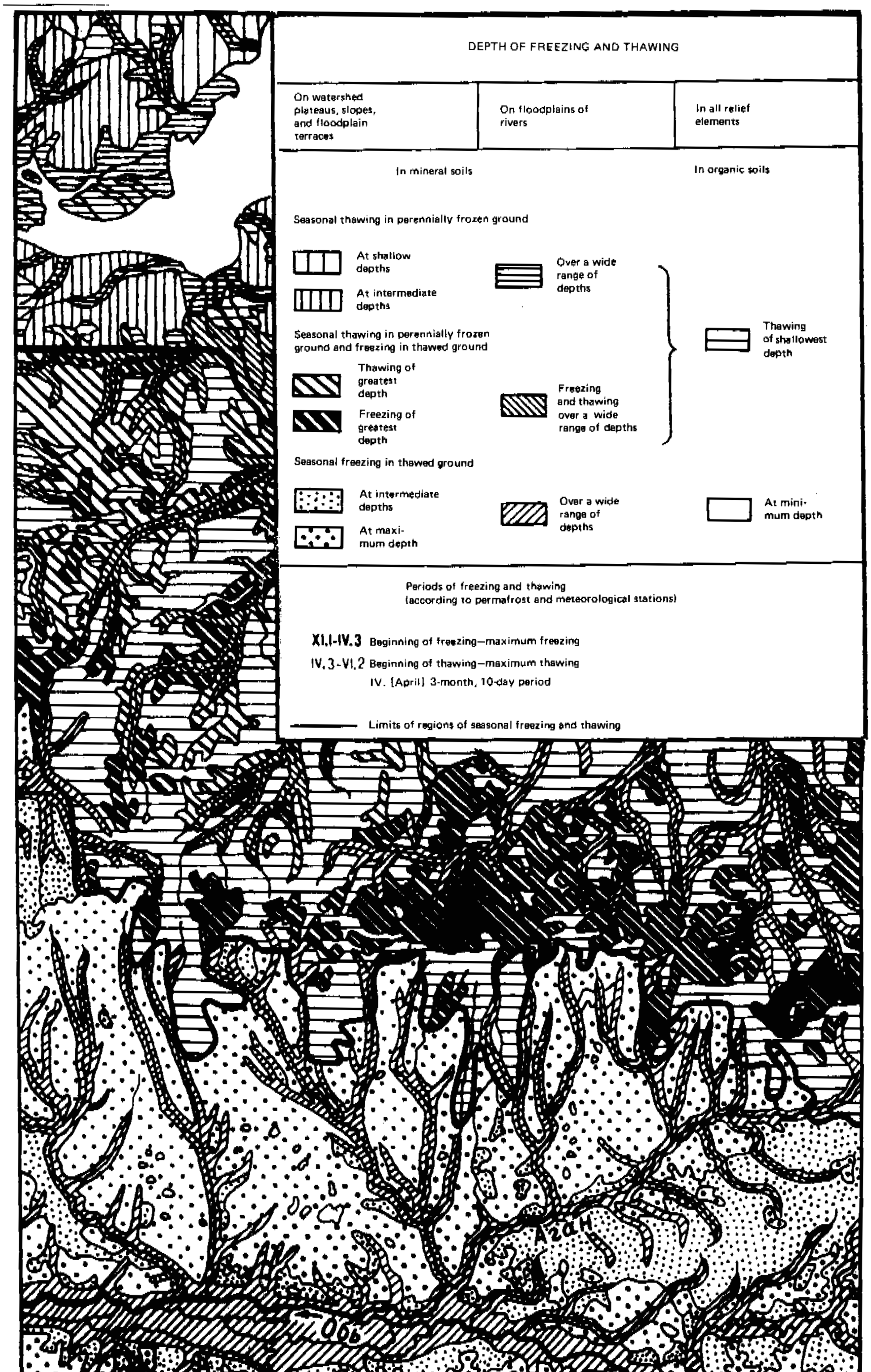

FIGURE 3 Seasonal freezing and thawing.

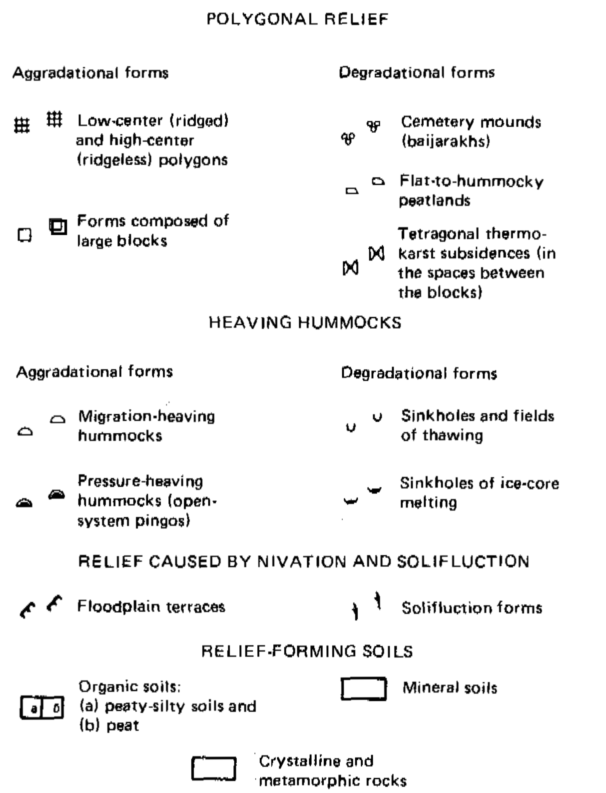

FIGURE 4 Permafrost relief.

of frozen soil and melting of ground ice. The thawing of ice-rich soils and large ice accumulations (the process of degradation), causes subsidence and settlement of noncohesive soil, leads to break-up and modification of the relief which was formed when the ground became frozen. The development dynamics of the permafrost relief is described by distinguishing within each group the forms resulting from aggradation and degradation which are indicated on the map by variously colored symbols. For example, in the polygonal relief group, included among the aggradation forms are the low center (ridged) and high center (ridgeless) polygons and also forms consisting of large blocks, all of which are typical in the region being mapped. The degradation forms include baijarakhs (cemetery mounds), flat-to-hummocky peatlands, and tetragonal thermokarst subsidences. Indicated as a supplement to the basic content of the map are the main lithological groupings (mineral, organic, crystalline), in which the various permafrost forms originate.

In addition to describing the basic characteristics of the perennially and seasonally frozen ground, in the case of western Siberia it is essential to provide a graphic representation of the history of development of the permafrost, associated with the geological history of the Quaternary period. This is of great value in helping to reveal the specific features of the contemporary state of the permafrost. In line with the prevailing scientific concepts regarding the development of the permafrost of the western Siberian Plain, provision has been made for the distinguishing, on 1:16,000,000-scale maps, of four stages. In particular, on some of the maps a description is given of the character of development of the permafrost during the periods of maximum glaciation (middle Pleistocene), the Zyryanka glaciation (upper Pleistocene), and the thermal maximum (Holocene), as well as during the contemporary period. Used as the basis for the content of the maps is the state of the permafrost during the different time stages: the regions of active freezing and initial degradation of relic permafrost are distinguished. Also taken into consideration are the general patterns observed in the distribution of the temperature and thickness of the permafrost during these time stages and also the spatial relationship between the areas affected by marine submergence, glaciation, and the permafrost.

The production of the maps of the Tyumen' Oblast' made it possible to use a special technique for their compilation, with a resulting high degree of precision in the positioning of the permafrost phenomena and consequent reliability of the maps. A single reproducible geographic base was used for all of the maps. During the author's work on the compilation, however, special bases were produced for the various permafrost maps, which include a detailed description of the topography and a number of other components (contours of the forests, peat bogs, and lithological variations, etc.), depending on the peculiarities of content of the map being compiled. The use of these contours, which serve as indicators, makes it possible to refine the boundaries of a number of permafrost characteristics. Also contributing to the increase in reliability are other, previously compiled specialized maps (lithological, geomorphological, hypsometric, and the like).

A series of permafrost maps makes it possible to:

1. Describe the general patterns of distribution and development of the permafrost and emphasize graphically its role in the evolution of the natural landscape,
2. Indiate (in conformity with the existing understanding of the region) the particular characteristics of the permafrost conditions, their specific regional nature, and variations in some of the characteristics as determined by local natural factors,
3. Estimate the influence of the sum total of the permafrost conditions on the various aspects of economic activity.

The series of permafrost maps for an individual region assists in the working out of theoretical principles and the compilation of maps for the assessment of permafrost conditions and also of small-scale prediction maps.

Undoubtedly, the further expansion and deepening of permafrost research and the resulting increase in our understanding of perennially and seasonally frozen ground will call for improvements in the methods used to describe them (the addition of new indices and an increase in reliability and cartographic precision).

The foregoing principles of representing the permafrost characteristics of the Tyumen' Oblast' on small-scale maps in connection with the comprehensive mapping of the area are also applicable to other regions. But the compiling of the maps, the subject areas they cover, and the choice of the qualitative and quantitative indices for describing perennially and seasonally frozen ground will depend on the specific nature of the permafrost conditions in the area being mapped.

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THE GEOCRYOLOGICAL CHARACTERISTICS OF THE MONGOLIAN PEOPLE'S REPUBLIC AND SOME CHARACTERISTICS OF PERMAFORST DEVELOPMENT IN THE PAST

G. F. GRAVIS, S. I. ZABOLOTNIK, A. M. LISUN, AND V. L. SUKHODROVSKII

Research conducted by co-workers of the geocryological detachment of the Joint Soviet-Mongolian Geological Expedition during the periods 1967-1968 and 1970-1971 resulted in a marked expansion of the earlier concepts of the development of seasonally and perennially frozen ground in the Mongolian People's Republic.[4-7,9]

Yachevskii's idea concerning an "alpine" type of permafrost in Mongolia[9] was corroborated, and the close connection between this phenomenon and the mountainous topography of the country was elucidated, as was its relation to altitude. At the lowest absolute elevations there are islands of permafrost that are confined to spring sinkholes. As the terrain becomes higher, the permafrost occupies a progressively greater area, extending initially to the bottoms of valleys and tectonic depressions, then successively to slopes with a northernly exposure and to scree trains and their bases, and, finally, to slopes with a southerly exposure. In each of these topographic features, the lower limit of permafrost is in contact with facies consisting of permanently moist, salt-free suglinoks.

This relationship between permafrost distribution and relief, which has been revealed and repeatedly checked under field conditions and confirmed by computations, was used for distinguishing altitudinal zones of permafrost for geocryological mapping purposes (see Figure 1 and Table 1). The area occupied by permafrost as shown on the schematic map is 985,950 km^2, which is 63 percent of the Republic.

Also reflected in Table 1 are the geographical patterns of variation in the depths of seasonal thawing and freezing for the altitudinal zones, as revealed by computer methods based on V. T. Balobaev's formula[1] and utilizing data collected by meteorological stations and during field studies. The maximum depths of thawing and freezing were calculated separately for suglinoks and sands and for the most probable ranges of variation in moisture content, obtained after statistical processing of the findings.

In conformity with the tectonic structure of the country and the distinctive features of its topography, the altitudinal zones of permafrost are arc-shaped.

The south and far east of the country are occupied by a zone of seasonally frozen ground. Within this zone, according to the calculations, the potential depths of seasonal thawing are 0.4-1.3 m greater than the actual depths of seasonal freezing. In most cases, therefore, permafrost does not form even in the total absence of a snow cover, the thickness of which is normally 1-2 cm.

The altitudinal zone of sporadic permafrost is bordered on the south by all of the principal mountain formations of the Republic and by an arcuate southerly extension of the Siberian platform. Its lower limit is at the same time the lower (southern) limit of occurrence of permafrost in the Republic. Further south, there are only 12 small isolated areas of sporadic permafrost, which are confined to the ranges of the Gobi Altai, the Matad upland region, and the Dariganga volcanic plateau.

As the potential depths of seasonal thawing within this zone are 0.2-0.5 m greater than the

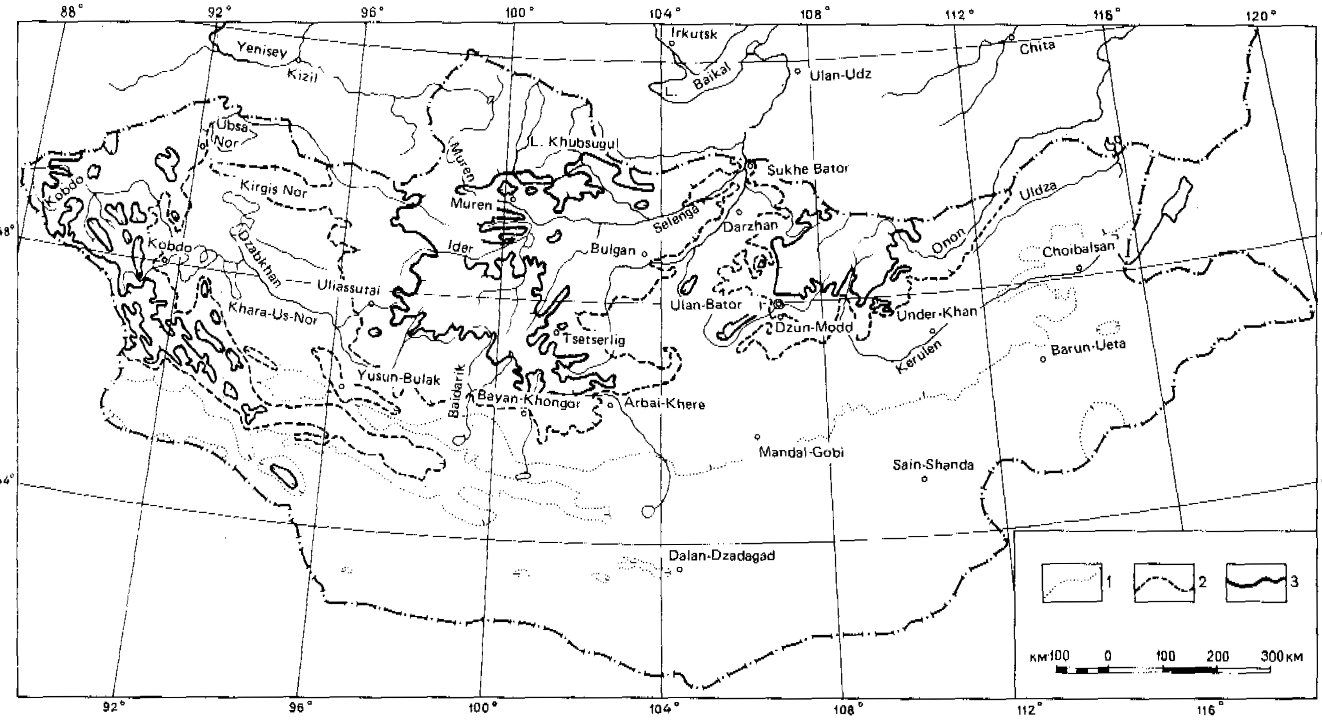

FIGURE 1 Schematic map of distribution of permafrost in the Mongolian People's Republic. Lower limits of the altitudinal zones of: (1) sporadic permafrost, (2) isolated patches and scattered islands of permafrost, (3) discontinuous and continuous permafrost.

actual depths of seasonal freezing, generally speaking the conditions within the zone are wholly unfavorable for the formation of permafrost. In the salt-free suglinok deposits topped by a dense turf, however, which are confined to centers of permanent moisture (to spring sinkholes and, less frequently, to closed depressions in areas consisting of blister basalts, the dry valleys of streams originating high in the mountains, or the valleys of rivers crossing the area), the depth of seasonal thawing decreases rapidly, whereas there is little change in the depth of seasonal freezing. In such places there are small islands of permafrost. The areal extent of the islands is not more than several thousand square kilometers, and they are dozens of kilometers apart, with the result that permafrost accounts for much less than 1 percent of the zone.

Removal of the snow cover, which is usually from 2 cm to 5 cm thick, or local artificial irrigation of the suglinok deposits can lead to the formation of pereletoks and even lenses of permafrost.

A belt of isolated islands and discontinuous permafrost surrounds each of the three great mountain systems of the Mongolian Republic: Khentei, Khangai, and the Mongolian Altai. The belt of continuous permafrost is confined to the higher elevations of these systems, especially in the Khangai and Prikhubsugul'ya Mountains.

In the zones of isolated patches of permafrost the calculated depths of seasonal thawing and freezing are almost identical. This means that even small changes in the heat-exchange components can lead to the formation of permafrost or to its thawing. Most often, the islands of permafrost are formed in the tectonic depressions along the periphery of Khentei, Khangai, and the Mongolian Altai. The mountain-fed rivers constitute a continuous source of irrigation of these depressions. Moreover, islands of permafrost are also found in the river valleys themselves and in spring sinkholes. They range from several to tens of kilometers in size, and the individual islands are often dozens of kilometers apart. Accordingly, not more than 5 percent of this zone is occupied by permafrost. Destruction of the snow cover can lead almost everywhere to the formation of pereletoks and permafrost.

It is evident from the above that within the zones of sporadic and isolated patches of permafrost the moisture conditions that are necessary for perennial freezing of deposits are set up as a result of the local redistribution of moisture by surface or underground drainage.

Within the overlying zones of continuous, discontinuous, and island permafrost, their formation is directly linked with the specific hydrothermal regime, which, under the prevailing climatic conditions, is to a certain extent typical of each shaded slope. Where permafrost is present in the form of islands, the potential depths of seasonal freezing in the moist deposits exceed by 0.3 m to 0.4 m the depths of thawing; it is only in the dry deposits that the possible depths of seasonal thawing exceed the depths of freezing by 0.2 m to 0.3 m. In the more elevated zones the potential depths of seasonal freezing exceed those of thawing over the greater part of the area. Within these three zones, therefore, permafrost occurs regularly and occupies from 5 to 40 percent of the area containing permafrost islands, 40 to 80 percent of the discontinuous permafrost zone, and more than 80 percent of the zone of continuous permafrost.

All the limits of the altitudinal zones become lower towards the north, although not uniformly so, i.e., each limit exhibits its own pattern of change. Thus, in the northern part of the Republic, for every 100 km the lower limit of the zone of sporadic permafrost decreases in altitude by an average of 130 m and in the southern part, by 400 m, the corresponding figures for the lower limits of the zones containing isolated patches and scattered islands of permafrost being 210 m and 71-130 m, respectively.

The abrupt changes in the position of the altitudinal zones of permafrost are due, in part, to the situation of the terrain in relation to the mountain ranges, which serve as barriers to the prevailing moist currents of air. On the windward slopes all of the limits are lower, and there is an increase in the areas occupied by permafrost, while on the lee slopes, the converse is the case. An exception is Khangai; since its western slopes are still in the wind shadow of the Altai, the altitudinal zones are higher there than on the eastern slopes. The maximum difference in the altitude of the zones as between the windward and lee slopes on the same parallel of latitude can be as much as 600-800 m.

Over the greater part of the Republic, the thickness of the permafrost ranges from several to a few dozen meters. The maximum thickness--140 m--was found in 1968 in one of the hollows to the north of the Bolnai Range at an altitude of 1,850 m by extrapolating data on temperature variations in a drillhole 65.4 m deep. There is every reason to assume that at altitudes of 3,500 m to 4,000 m (Khangai and the Mongolian Altai) the thickness of the permafrost increases to 1,000 m.

At a depth of 20 m the mean annual ground temperature ranges from 10°C in some regions of the trans-Altai Gobi to -5°C at altitudes of 3,000 m 4,000 m in Khangai and the Mongolian Altai (see Table 1). The geothermal gradient, depending on the composition of the ground, ranges from 0.014 deg/m to 1.00 deg/m, although its most likely values are 0.02-0.03 deg/m.

In Mongolia, the permafrost was in the main formed epigenetically, through a downward accretion of the frozen strata. They are characterized by the presence of three layers of ground-ice formation.

The layer of pressureless migration of moisture towards the plane of freezing is the uppermost of these. Depending on the hydrogeological conditions of freezing, it varies in thickness from several centimeters to 3-5 m. This layer contains thin lenses of segregation ice from 1 mm to 3 mm thick and up to 4 cm to 5 cm long. Because of seasonal temperature variations, giving rise to acceleration and retardation of freezing, the ice lenses at first become compressed into indistinct, thin belts and then become rarified. As a result, the frozen suglinoks have an indistinctly zonal lenslike cryogenous texture of the segregation type[2] and their moisture content falls from 40-50 percent to 20 percent. In coarse-grained deposits only ice bonding is seen, and their moisture content is 10-20 percent.

The layer of pressure-migration of moisture occurs lower down, toward the plane of freezing. This layer is several dozen meters thick. During its freezing, hydrogeological factors are of decisive importance in the ice-formation process: the degree of saturation of the deposits, the presence of hydrodynamic or hydrostatic pressure, the magnitude of this and also other characteristics, which do not remain constant but alter in the course of the freezing. The pressureless (pellicular) migration of moisture toward the freezing front is combined with processes of shuttlelike pressure migration of free water from the more stressed to the less stressed areas of the freezing front. In suglinok deposits the distance over which water migrates under pressure in the subhorizontal direction is measured in centimeters and decimeters; in sandy deposits, in kilometers. The ice forms as a result of two closely interrelated processes: segregation of the growing ice crystals and of mineral soil particles, which becomes possible only because of hydraulic pressure, and freezing of injected water.

The fluctuation of the hydraulic stresses, which is to a large extent due to the ice-formation processes themselves, and the resulting shuttlelike pressure migration of water, give rise to a rhythmic cryogenous structure of the permafrost. The rhythms are formed by ice lenses or series of these and consist of two parts: an initial (upper) part, containing less ice and a terminal (lower) part, which contains more ice.

The shape of the ice intrusions is mainly determined by the properties of the freezing ground. In banded (varved) sediments, a rhythmically bedded reticulated cryogenous texture is formed (a combination of horizontal and vertical ice wedges); in indistinctly bedded strate, a rhythmically bedded irregularly reticulated texture (a combination of horizontal and inclined ice lenses); and in unstratified deposits, a rhythmically lenslike cryogenous texture (vertical or inclined ice veins are not typical). All three forms of cryogenous textures belong to the injection type.[2]

The rhythms form a cycle embracing the entire layer in which moisture is migrating under pressure towards the freezing front. Three stages are distinguished in the cycle: an initial, a culminative, and a final. The initial stage is characterized by rhythmical accretion of ice, the culminative stage by the maximum ice content in the stratum, and the final stage by a rhythmical decrease in ice.

The formation of the culmination stage of the cycle is associated with the maximum hydraulic stresses. When there is a favorable combination of geological-geomorphological and hydrogeological conditions, the ice-formation processes in the culminative cycle become especially pronounced, and, in a number of cases, deposits of ice measuring up to several tens of meters thick are formed. Such areas can be called zones of intensive ice formation, by analogy with ore-formation zones. Six types of such zones have been iden-

TABLE 1 Altitudinal Zones of Permafrost in the Mongolian People's Republic and a Brief Description of Them

Altitudinal Zone	Lower Limit of Altitudinal Zone	Absolute Elevations of Lower Limits of Altitudinal Zones, m: Khentei	Khangai and Prikhubsugul'e	Mongolian Altai	Area of Altitudinal Zones: KM²	Percentage of Total Area of Republic (1,565,000 km²)	Percentage of Area of Republic Occupied by Permafrost
Of seasonally frozen ground					579,050	37.0	--
Of sporadic permafrost distribution	Lower limit of permafrost in spring sinkholes	700-1,200	1,400-1,600	1,200-1,800	460,110	29.4	46.6
Of isolated patches of permafrost	Lower limit of permafrost in valleys and depressions	700-1,600	1,200-2,400	800-2,400	190,930	12.2	19.4
Of scattered islands of permafrost	Lower limit of permafrost on slopes with northern exposure	1,200-1,800	1,200-1,300	1,600-3,000	159,630	10.2	16.2
Of discontinuous permafrost	Lower limit of permafrost on scree trains near the base of slopes with northern exposure	1,200-1,800	1,400-1,600	2,200-2,800	28,170	1.8	2.9
Of continuous permafrost	Lower limit of permafrost on slopes with southern exposure	1,400-2,000	1,500-3,000	2,800-3,000	147,110	9.4	14.9

tified: spring sinkholes, the marginal parts of alluvial debris cones, the central parts of Baikal-type depressions (according to Florensov's classification[8]), the marginal and central parts of depressions of the Gobi type in northern Mongolia, and the central part of the depressions of this type in southern Mongolia. In most cases the zones of intensive ice formation are distinguished in the topography as perennial hummocks or areas of heaving. By virtue of the foregoing ice-formation characteristics the moisture content of the frozen ground of the layer in question ranges between wide limits.

Occurring lower down is a layer of ice bonding. Ice can form in it only in the hollows and cracks that were present in the ground prior to its freezing. The moisture content of the deposits ranges from 10 to 30 percent.

During the cold ages of the Pleistocene, permafrost was much more widespread than it is today. Traces of its existence have even remained in the topography and in Quaternary deposits in the form of various geomorphological and geological postcryogenous formations. They can serve as direct or indirect paleocryogenic indicators.

Under the conditions prevailing in Mongolia, the following geomorphological formations serve as direct cryogenous indicators: relic perennial heaving hummocks, thermokarst sinkholes, and solifluction lobes on dry slopes impervious to great depth, as well as the following geological formations: postcryogenous textures (injection and composite segregation textures); stationary mushroom-type and cryoturbations, corresponding to cemetery mounds (baijakakhs); large undulating deformations of the layers, associated with the initial stages of development of perennial heaving hummocks; pseudomorphs after ice wedges; and solifluction deformations in the deposits on dry slopes impervious to great depth. The direct paleocryogenetic indicators can be used with confidence as a basis for inferring the existence of permafrost in the past, and in some cases they can even be used for determining the type of freezing --epigenetic of syngenetic.

Under the conditions prevailing in Mongolia, the following geomorphological formations serve as indirect paleocryogenous indicators: relic raised terraces and solifluction terraces on dry slopes characterized by the presence of impervious strata close to the surface, as well as the following geological formations: elementary postcryogenous textures of the segregation type; nonstationary cryoturbations corresponding to thufurs (hummocks), sand wedges, and solifluction deformations in the deposits on dry slopes characterized by the pres-

Probable Minimum and Maximum Calculated Depths of Seasonal Thawing (Numerator) and Freezing (Denominator)				Maximum Thickness of Permafrost, m		Temperature, °C		
Suglinoks		Of Sands		Directly Ascertained or Calculated	Presumed	At Ground Surface	At a Depth of 20 m	The Most Typical Features of the Cryogenous Relief
-25%	-10%	-15%	-3%					
2.1/1.7	3.0/2.1	2.5/2.0	5.5/4.2	--	--	+1.6+8.0	+3.0+10.0	Thufurs
1.9/1.4	2.7/2.4	2.2/2.0	5.0/4.5	6.5	10	-1.4+5.0	+1.0+5.0	Thufurs, cemetery mounds, and perennial-heaving hummocks
1.7/1.6	2.4/2.4	2.0/1.9	4.8/4.8	50	--	-2.8+4.2	-1	
1.7/2.0	2.5/2.3	2.0/2.4	4.7/4.4	50	100	-4.2+2.4	+5.0	Thufurs, cemetery mounds, perennial-heaving hummocks, and solifluction formations
				160	1000	-1.6+2.3	-5.0+1.0	The same, together with solifluction terraces, frost-fracture polygons, raised terraces, and structural formations
1.6/2.0	2.0/2.3	1.9/2.4	4.0/4.6					

ence of impervious strate close to the surface. Indirect paleocryogenous indicators attest to the more severe climatic conditions in the past, or at least to the seasonal freezing of the deposits, but they do not prove the existence of permafrost.

The study of direct and indirect paleocryogenetic indicators assures us that there was a marked two-stage expansion in the area occupied by permafrost in Pleistoncene times. During the course of the first expansion, the entire area comprising the Mongolian People's Republic became frozen, including its southern Gobi regions. The results of spore-pollen analysis and geological-geomorphological findings make it possible to correlate this period with that of the Samaranian-Tazovian glaciation of Siberia.

During the second period, the limit of the zone containing isolated islands of permafrost* was approximately 150-200 km south of its present position. The southern limit of the zone is clearly recorded by a belt of relic thermokarst sinkholes, which can be seen to extend between depressions located in the eastern part of the Republic and a trough of lakes in the central part of the country. The results of spore-pollen analysis and their correlation with the various paleocryogenetic indicators and the character of the deposits in closing them afford a basis for confidently dating this period to that of the Zyryansk-Sartanian glaciation of Siberia, or, more precisely, to the second half of the Zyryansk period.

The analysis of the existing information has made it possible to conclude that the development of the glaciation and permafrost in Zyryansk-Sartanian times and, according to preliminary data, in the Samaranian-Tazovian period in Mongolia was asynchronous. Against the background of the decrease in mean annual air temperatures, the increase in the humidity of the climate that occurred during the first half of the ice ages give rise to the formation and advance of the glaciers in the mountains. Simultaneously, the lakes began to transgress[3] and the thickness of the snow cover became greater, which hindered substantial aggradation of permafrost. The drying up of the climate during the second half of the ice ages caused the mountain glaciers to retreat. The lakes began to regress, and the thickness of the snow cover decreased. The outcome of the two latter phenomena was an increase in the areal extent of the permafrost. The end of the ice ages

*The limit of the altitudinal zone of sporadic permafrost has not been determined by paleocryogenetic indicators.

evidently coincided with degradation of the permafrost.

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GEOCRYOLOGICAL CONDITIONS OF THE SOUTHERN PART OF THE YANA-INDIGIRKA PLAIN

K. A. KONDRAT'EVA, S. F. KRUTSKII AND
A. B. CHIZHOV *Lomonosov State University*

The results of a cryological and hydrogeological survey, conducted in 1969-1971 by the Yakutsk Expedition of Moscow University's Faculty of Geocryology, have made it possible to ascertain the basic mechanisms involved in the development of permafrost under the distinctive conditions occurring at the junction between the mountainous structures and the Yana-Indigirka plain. This special morphostructural zone is confined to the structural contacts between the major components of the most recent tectonic formations: the Kolyma-Indigirka platform and the Verkhoyansk-Chukotka region of orogenesis.

Included in the geological structure of the young Kilyma-Indigirka platform are the deposits of the Verkhoyansk complex, which form a folded basement, and a platform mantle of loose Cenozoic deposits. The Verkhoyansk complex corresponds to the geosynclinal stage of Mesozoic development and is a typically miogeosynclinal formation. It consists of a thick (up to 8-10 km) interstratification of sandstones, siltstones, and clay shales. Within this area the Cenozoic deposits consist exclusively of continental formations. Occurring at the base of the section are sands, pebbles, and clays interstratified with lignite. They are succeeded by Pleistocene sands, supesses, suglinoks, and silts containing thick regeneration ice wedges. The contemporary deposits consist of moderately thin, residual and slope formations; lake-swamp deposits; and alluvium from floodplain terraces. The total thickness of the Cenozoic deposits is 50-100 m, but in the tectonic depressions of the basement it can be as much as 200 m or more.

The topography of the southern part of the Yana-Indigirka plain is an intricate combination of aggradation and denudation plains with the latter exceeding the former by 20-50 m, occasionally by as much as 100 m. In its contemporary structure, the zone of denudation-aggradation plains can be defined as a keyboard-block formation, caused by activation of the feathered system of dislocations characteristic of structural contacts.

The entire area is characterized by an abruptly continental type of climate and low (-13° and -14°) mean annual air temperatures. The mean annual precipitation is of the order of 300 mm. The distribution of the snow cover and its thickness are greatly affected by shifting winds. On the average, the snow cover ranges from 15-20 cm to 40-50 cm in thickness. At sites of drifting, snow patches of up to 2 m or more form. On the average, the snow cover remains for 230-240 days; the snow patches, until late July or early August.

A large part of the region is occupied by typical arctic tundra. The areas of northern taiga tend to be localized in the foothills and are wedged far to the north only along the valleys of the major rivers, while isolated stands of trees are also found on the lee slopes of the lake-filled basins. The influence of the vegetation on the temperature regime of the ground is chiefly expressed by the thickness of the snow, as a rule, this is 20-25 cm greater in tree-covered areas.

Under conditions of flatland topography, in which the surface radiation regime is fairly uniform, the relationship between areas of uplift and subsidence, i.e., denudation and sediment accumulation, emerges as a powerful factor in the differentiation between permafrost conditions. Denudation plains are widely developed in the immediate proximity of the Polousnyi and Ulakhan-Sis ranges. On the average, their absolute elevations are 100-200 m. The variously leveled denudation plains form a belt of foothills, which border the mountain ranges on the north. To all appearances, the lower level of these plains consists of river terraces and pediments, which merge to form a single surface.

In the submeridional direction the piedmont zone is cut through by narrow tectonic depressions (fault-block valleys) within which the surface and subsurface drainage is concentrated. It is here that in Neogene-Quaternary times the rhythmically built sandy-pebbly groups of strata measuring up to 100 m in thickness, which serve as traps for groundwater, were accumulated. In their geomorphology they are alluvial plains.

Further north, the piedment zone and the middle-upper Quaternary lacustrine alluvial plain, together with the younger, extensive alas depressions enclosed by the latter are seen to be intricately linked. Frequently, the boundaries between the denudation and lacustrine alluvial plains extend along very recent faults and are defined by distinct scarps. Within the relatively sunken blocks delimited by these faults, in middle and upper Quaternary times the accumulation occurred of fine-grained silty sediments of lacustrine alluvial origin that overlie more ancient Cenozoic formations--pebbles, sands, and clays containing lignitic inclusions. The overall thickness of the Cenozoic strata is as much as 100 m or more.

Thus, by the time that the ground first showed signs of becoming perennially frozen, which was in the early Pleistocene period,[2] the geological setting and its attendant diversity of conditions of permafrost development was characterized by the following features:

1. The differentiated character of the sedimentation, which determined the regular alteration of the facies composition of the Cenozoic mantle, ranging from detrital and clayey-silty residual formations on the interfluvial denudation plains, through gravelly-pebbly-sandy strata of the riverbed facies in the piedmont depressions, to silty sandy-silt, clayey-silt alluvial, and lake-swamp deposits in the zones of regional aggradation.

2. The concentration of the surface and subsurface drainage in grabenlike tectonic structures, where, owing to the accumulation of sandy-pebbly strata, conditions were set up for the generation of powerful currents of groundwater.

3. Activation of the feathered system of dislocations, which, besides being responsible for the development of the blocklike structure of the area, gave rise to variability of the geothermal conditions due to the convective heat transfer by the groundwater moving through the fractured zones.

Within the confines of the denudation plains in which lateral planation processes predominated, the freezing penetrated the rocks more or less uniformly. Here, the thickness of the permafrost zone was chiefly determined by long-term temperature fluctuations at the soil surface, caused by general climatic variations. Because the rocks of denudation plains are least subject to tectonic fracturing, fluctuations in the thickness of the permafrost occasioned by local variations in the geothermal conditions near the plane of freezing were manifested there to a much lesser degree than in the regions of the tectonic depressions. At the present time the permafrost thicknesses within the watershed denudation plains rarely deviate from the mean values of 350-450 m. The permafrost consists of the epigenetically frozen terrestrial rocks of the Verkhoyansk folded complex. The cryogenous textures of the latter are, in the main, of the inherited fracture, fracture wedge, and stratified fracture types. The relatively thin (to 5 m) loose formations of residual, detrital, and solifluctional origin overlying the rocks of the Mesozoic mantle are characterized by the development of structural soils and a polygonal relief containing soil wedges and thin regeneration ice wedges.

Within the alluvial plains confined to the fault-block valleys, the development of the permafrost was characterized by a more complicated history. Here, the heat exchange between the ground and the atmosphere was accomplished under the highly dynamic conditions of alluvial sediment accumulation, which, in virtue of the tectonic conditions, took place during the formative phases. The negative rhythmic movements of the Earth's crust were such that the tectonic depressions formed by the river network became heavily filled with clastic material; it was as if constantly migrating river channels were crossing over to continually higher levels in relation to the alluvial strata.

When frozen alluvial deposits were present, this process was accompanied by migration of the talik zones situated beneath the beds and encompassing a large part of the alluvial strata. This circumstance facilitated the complete reduction of the floodplain and oxbow lake facies of the alluvium, the most typical feature of which, from the beginning of the permafrost, was the very high ice content, due to the development of regeneration wedge ice. As a result, the alluvial strata in the depressions (with the exception of the relief-forming deposits of upper Pleistocene age) consist, for the most part, of riverbed facies. Furthermore, the migration of the river beds and bed taliks gave rise to a repeated cycle of freezing and thawing of the alluvial deposits.

In view of these sediment accumulation dynamics, the Cenozoic strata within the aggrading piedmont plains are characterized by the development of two layers of sediments, which differ sharply in their cryogenous structure. The lower layer consists preeminently of polycyclically frozen pre-upper-Plaistocene, relatively ice-poor pebbles, sands and finely dispersed sandy silts, and clayey silts.

Depending on the lithological characteristics, the cryogenous textures range from being crustal and massive in the pebbles and sands to microlensed (up to 1 mm), finely lensed (1-10 mm), and, less frequently, medium lensed (1-3 cm). As the thicknesses of the lenses increase, the distance between them becomes greater, amounting to 5-8 cm on the average. The moisture content by weight of the deposits is fairly constant through the section and amounts to 30-40 percent.

The upper layer consists preeminently of syngenetically frozen alluvial deposits (floodplain and oxbow lake facies), less frequently of lake-swamp facies with epigenetic freezing of spotted tundra deposits, and also of syngenetically frozen alas deposits. All of the syngenetically frozen deposits consist of finely dispersed sandy and clayey silts with fine inclusions of detritus and lenses and interstratifications of peat. Deposits in the form of pillars of soils are embedded in a 6-12-m-thick stratum of regeneration ice wedges. The most typical cryogenous textures of the deposits are thin lenses (1-3 cm) and thick lenses (up to 5-7 cm) being 5-8 cm apart. The cryogenous texture between the ice zones and thick lenses of ice is finely lensed, stratified, and reticulate. The moisture content of the deposits is 120-240 percent by weight and 70-95 percent by volume.

Whereas in the lacustrine alluvial plains, situated far from the regions of ablation, it is preeminently sandy-silty bed alluvium that was deposited, which hardly differs from the other alluvial facies, in the piedmont zone the bed facies of the fault-block valleys, consisting of pebbly strata, had a pronounced effect on the character of the freezing. This found expression in the favorable conditions of percolation of the surface and valleyside suprapermafrost waters into the alluvial stratum, which in turn dictated the existence within these valleys of thick (up to 50-100 m) infrabed and floodplain taliks. At the present time such taliks are developed beneath a number of rivers (the Berelekh, Tenkel', Nuchcha, Dodomo, Igan'e, and others). In some cases, (e.g., in the middle course of the Nuchcha River), open taliks form in the valley areas that coincide with the zones of fracture in the folded basement of the plain.

The temperature regime within the zones of the taliks and their subjacent frozen strata is unstable. Investigations by expeditions working in the valley of the Tenkel' have revealed the existence of neoformations of permafrost 10-20 m thick, associated with displacement of the riverbed. In this case the frozen stratum in the marginal region of the talkik is in the form of a peak, which tapers off in the direction of its axial part.

The maximum width of the thawed zone increases in the downstream direction from a few hundred meters to 1 km. With transition from the piedmont zone to the lacustrine alluvial plain, the spacious floodplain taliks taper off, beyond which it is possible to trace only closed, intermittent taliks situated below the riverbeds and confined to the deepest sectors of them. This is due to the lessening of the longitudinal gradient of the valley and to alterations in the percolation properties of the alluvial deposits. Furthermore, within the lacustrine alluvial plain there is a pronounced decrease in valleyside drainage, which plays an important part in the influx of heat to the taliks. Naleds usually form near the zone in which the floodplain taliks taper off.

Geocryologically, the region of occurrence of the ancient lacustrine alluvial plains differs appreciably from the above-described piedmont zone. Here, the so-called ice complex of deposits as thick as 15-30 m and incorporating large beds of syngenetically frozen regeneration ice wedges is extensively developed. Its formation occurred during the upper and, in part, middle Quaternary times in the course of the freezing of preeminently floodplain facies of alluvial sediments.

The stratification and thickness of the perennially frozen lacustrine alluvial plains were greatly affected by thermokarst processes, which have been especially pronounced during the contemporary period. These processes have led to the formation of numerous lakes averaging between 2-3 and 8-10 m in depth and from several hundred meters to upwards of 1.5 to 2.0 km in plan. In summer, the temperature of the upper layers of the benthic deposits can reach 10°C to 13°C.

The minimum depth of water that is necessary for the formation of a sublacustrine talik, according to direct observations and calculations,[3] is 1.5 m. The minimum dimensions at which a talik can be open are estimated at 600-800 m, and the period necessary for complete thawing of the permafrost, depending on the depth of occurrence of the folded basement, ranges from 1,500-2,000 yr to 15,000-20,000 yr.

The situation at present is that the sublacustrine taliks of the Yana-Indigirka plain have been insufficiently studied. Special attention should therefore be paid to the discovery by the Yakutsk expedition that an open talik exists beneath Lake Baka. This thermokarst lake measuring 4.2 × 2.5 km and with a depth of 3-4 m is situated in a small tectonic depression cutting through the Polousnyi Range. From the south it opens in the direction of the Uyanda intermontane depression; on the north it abuts upon the piedmont zone of the Yana-Indigirka plain. The basin is filled with Cenozoic deposits of not more than 40-50 m in thickness and consisting of gravelly sands and clayey silts. Along the shores of the lake, the thickness of the permafrost is 300-400 m. The formation of the open talik here is primarily due to the prolonged warming influence of the water body and the favorable geological conditions.

In the piedmont part of the area, there are mostly small lakes with closed taliks below them. Thus, a talik with thickness 25 m was discovered below one of the lakes measuring 310 × 180 m. The thickness of the talik, as calculated from a diagram of the stationary thermal field, came to 35 m. This attests to the instability of the geothermal field, which is due to the dynamics of development of the lake itself and to its relatively young age.

The influence of the lakes is such as to cause an elevation of permafrost base. Electrical logging, conducted from the offshore shelf region of a large lake in the Yana-Omoloi interfluve, revealed that the base of the permafrost is located at depths not exceeding 150-200 m. The thickness of the permafrost at the low alas level and in the connecting areas between the lakes is between 200 m and 300 m; at the high alas levels it increases to 400 m. At the same time, in the residual outcrops of the ancient lacustrine alluvial plain it is as much as 500-550 m. The subpermafrost waters of this area are weakly mineralized, and that there is no significant layer of cryopegs* at the base of the permafrost.

The decrease in the thickness of the permafrost in the alasses is a consequence of their existence at a place where large bodies of water were present in the recent past. Their drying out was due to the drainage of the lake waters into the river system. This process was greatly facilitated by the change in the sign of the tectonic movements at the boundary between the Pleistocene and Holocene, which caused activation of the erosion processes and an increase in the downcutting and sizes of the river valleys.

Rough calculations indicate that the age of the large alas depressions containing regenerative thermokarst lakes must be at least 3,000-4,000 yr. Thus, the formation of the alas depressions can be dated to the period of the thermal maximum or to even earlier stages of the Holocene. On the question of the possible migration of the lakes, it might be well to note that at present the annual rates of migration in the area concerned are scarcely in excess of 0.3 m to 0.5 m. The direction of migration of the lakes is the outcome of a very large number of factors. Therefore, in order to study the processes involved in the development of thermokarst lakes it is advisable to use statistical methods and the theory of probability. It should also be noted that the local decrease in the thickness of the permafrost to 300-200 m is associated with zones of increased fracturing in the basement. Such zones are usually confined to the valleys of rivers and creeks, aligned along very recent faults. The influence of fracture tectonics is especially pronounced in the areas where the thickness of the Cenozoic deposits does not exceed 80-100 m.

In conclusion we shall deal with some of the mechanisms involved in the development of the temperature conditions of the upper layers of the rocks and arising from the contemporary heat-exchange conditions in the rocks-atmosphere system.

In the development of the thermal radiation regime at the surface of the rocks in the zone of the piedmont denudation-aggradation plains, a very important role is played by adiabatic processes in the air masses. The superpositioning of the paths of movement of cyclones and anticyclones (the Nizhnekolymsk, Novosibirsk, and other trajectories) along the latitudinally oriented Polousnyi and Ulakhan-Sis ranges causes foehns to have a major influence on the development of the temperature conditions in the layer of air near the ground. From the results of observations during 1969-1970, the mean annual air temperature under these conditions proved to be -12.8°, i.e., the highest value for the entire Yana-Indigirka plain. It is evidently this that accounts for the relatively favorable conditions that resulted in the growth of thin stands of deciduous trees, which, in the basins of the Khroma, Chondon, and Berelekh rivers, extend as far north as 71°.

The existence of mean annual air temperatures that are relatively high for subarctic lowlands, together with the widely prevalent thin stands of deciduous trees and dense undergrowths of willow and alder, the snow cover within which can be as thick as 0.6 m to 0.7 m, have resulted in the development within the piedmont zone of highly typical permafrost rock temperatures, falling between -5° and -7°.[1] Rock temperatures such as these are characteristic of the gentle, wooded slopes, low tree-covered watersheds, terraces and residual outcrops of the ancient plain, and also of shallow lake basins. The temperature variations in this range are due to such factors as changes in the soil composition, the character of the plant associations and conditions of snow accumulation, and the surface moisture conditions. Specifically, the highest temperatures are characteristic of the Earth materials of denudation and aggradation plains, whose surface deposits consist of detrital-sandy or pebbly-sandy varieties.

The lowest permafrost temperatures occur in the tundra zone of the lacustrine alluvial plain. Here, as a result of blowing and compaction of the snow, at the high alas levels that extend for several kilometers, the peat-containing, lake-swamp, clayey-silts develop mean annual temperatures of between -9° and -11°. The elevated areas between them, consisting of residual outcrops of the ancient lacustrine alluvial plain, are characterized by temperatures in the -8° to -10° range. A rise in the mean annual temperatures to -7° and -6° is noted on lee slopes with predominantly westerly and northwesterly exposures.

The mean annual temperatures of the rocks of the thick floodplain and bed taliks are usually between 1°C and 2°C. Such temperatures are typical of most of the shallow, sublacustrine taliks and it is only in the open talik beneath Lake Baka that the mean annual temperature reaches 4°.

To sum up, it must be emphasized that piedmont denudation-aggradation plains are typical of most of the transitional zones between the major morphostructural components of the Soviet Northeast (the uplifts and depressions). Being situated for the most part within structural contacts, these plains probably have much in common as regards the history of development of the topography, the Cenozoic sedimentation, and the formation of the permafrost conditions.

*Bodies of mineralized unfrozen water at a negative temperature.

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THE EFFECT OF THERMAL ABRASION ON THE TEMPERATURE OF THE PERMAFROST IN THE COASTAL ZONE OF THE LAPTEV SEA

E. N. MOLOCHUSHKIN Permafrost Institute of the Siberian Department of the Academy of Sciences of the USSR

Permafrost first appeared in northern Yakutia during the early period of the Pleistocene, and from that moment on it has existed continuously, degrading to some extent in the interglacial epochs.[6]

During the last transgression, which terminated about 5,000-6,000 yr ago,[2] the sea flooded extensive areas of coastal lowlands comprising permafrozen strata on the shelf of the Laptev Sea. The coastline at that time was located 50-100 km north of its present location. Further ingression of the sea is due to thermoabrasive erosion of the permafrost forming the shores of the coastal lowlands.[7] On certain sectors of the Laptev seacoast, the rate of coastline retreat today amounts to 15-20 m annually.[1]

The flooding of coastal lowlands brings about a substantial change in the thickness and temperature of permafrost. This fact has been established by the author's observations conducted in the coastal zone of Muostakh Island, located in Buor-Khaya Bay on the Laptev Sea.

A number of boreholes were drilled on the island and in the coastal waters for observing temperatures. Borehole 1 is located on the island 100 m away from the coastline. Boreholes 2 and 4 are situated on the northeastern side of the island in the area of intensive coastal erosion. Borehole 2 is at the very foot of a cliff, and boreholes 3 and 4 at a distance of 50 m and 150 m respectively. Borehole 3, established on the 1-m isobath, penetrated 10 m into the frozen rocks, while borehole 4, having passed through a 2.3-m layer of water, penetrated 15 m into the frozen rocks. Boreholes 15-20 m deep, established on the 0-, 1-, 2-, 3-, and 4-m isobaths, including borehole 5 on the 3-m isobath, were drilled on the western side of the island in the area of sediment accumulation. The boreholes were established at the respective distances of 0, 60, 150, 600, and 900 m away from the coastline. Permafrost was not encountered during drilling operations.

According to routine observations carried out during the year, the mean annual water temperature in the coastal zone of the island is 0.2°-0.3°C, while the salinity has a well-defined dependence on time. The lowest salinity is observed at the end of July and beginning of August when the water becomes almost fresh. From October to April the salinity does not alter greatly; its mean value is 20 percent.[4]

During 4 yr of observations, the shores of the island retreated at the average rate of 5 m/yr in the area of boreholes 2-4. Judging by this, it may be assumed that the locations of boreholes 3 and 4 have been under water for 10 yr and 30 yr, respectively.

Naturally, the temperatures of permafrost that ends up at the bottom of the sea as a result of coastal erosion at first do not differ from the ground temperatures inshore (Figure 1).

As time passes, very significant temperature changes take place in the upper layer of the submerged permafrost. This is clearly seen in Figure 1, which also shows the ground temperatures taken in boreholes 3-5. According to these data, the temperature of permafrost at a depth of 10 m went up by 4° during the first 10 yr under water, which means that the rate of increase averaged ~0.4° annually during this period and 0.25 deg/yr during the first 30 yr. Undoubtedly, other rates of above-water and underwater coastal erosion will produce other rates of increase in the temperature of permafrost.

As seen in Figure 1, when the cooling of the seabed at small depths takes place directly through the ice sheet during a period of 6-7 months a year, the general nature of seasonal fluctuations in ground temperature remains almost the same as on land. However, the thickness of the layer has somewhat decreased with the annual fluctuations of temperature, as has the range of temperature change at the various depths.

An entirely different picture is observed in the area of borehole 4. Figure 1 shows that the appearance of a water layer that does not freeze through in winter has resulted in significant

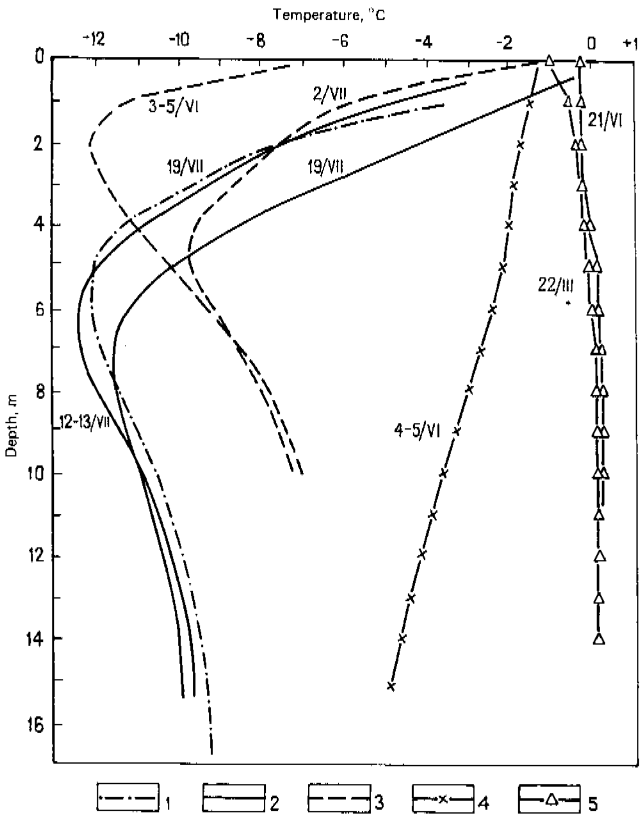

FIGURE 1 The temperature of continental permafrost and the different stages of temperature change under the thermal effect of the sea. 1--temperature of continental permafrost; 2--ground temperature in the initial stage; 3,4--ground temperature in the intermediate stages; 5--ground temperature in the final stage. [*date--22nd of March]

rearrangement of the temperature field of the permafrozen strata and a specific linear distribution of temperature, with a very gradual decrease corresponding to the depth of the layer. During the past 30 yr, the ground temperature in the upper boundary layer went up to approximately -1°C, and, at the depth of 15 m, to 5°C. However, as pointed out earlier, the ground down the entire section of the borehole was frozen.

It is not difficult to estimate that during the 30 yr the temperature of permafrost at a depth of 15 m increased by an average 0.17 deg/yr. In the future, the rates of increase in temperature will also continue to fall, and the temperature at which the phase transition of ice begins will be attained during the first 100 yr of submergence. Naturally, this assertion will be correct if the rate of coastal erosion remains constant. It follows that permafrost with a relatively high temperature can be found in an area of intensive thermal abrasion, even in the sectors with a positive mean annual water temperature at a distance of up to 1.0-1.5 km from shore. They may be covered over with a layer of thawed or frozen deposits of 10-m thickness or greater. Apparently, this type of permafrost may also be found in the open sea at quite a distance from shore, where islands have been washed away by the sea during the past century.

The rates of increase in ground temperature will in effect be equal to zero when the ice begins to thaw at a given depth. Indeed, simple calculations show that approximately 8-10 times less heat is spent on raising the temperature of frozen rocks by 10° (i.e., from the initial value to the beginning of phase transition) than is spent on melting ice if the ice content of the rocks is equal to 30 percent. Therefore, on the basis of even the most approximate estimates, it may be assumed that the temperature field of underwater permafrost at a depth of 15-20 m in the area under observation attains a practically stationary state not sooner than 800-1,000 yr after submergence. This, apparently, will take place when the temperature curve of borehole 4 is superposed on the curve of borehole 5 (see Figure 1). Judging by the results of five measurements taken monthly, the temperature of seabed deposits at a depth ranging from 9 m to 15 m in the accumulation zone is constant and equal to the mean annual water temperature. Thus, Figure 1 shows the initial, intermediate, and final stages of change in the temperature field of the upper 15-m layer of permafrost, which is now under water as a result of the thermoabrasive action of the sea and a positive mean annual temperature of the water.

The temperature dynamics of underwater permafrost in the shelf areas with a mean annual water temperature of 0°C and lower can roughly be judged on the basis of model tests. The modeled rock mass consists of two layers. The upper one is made up of iced-over Quaternary deposits with a thickness of ~50 m, and the bottom layer is composed of bedrock.

Since the mean annual water temperature fluctuates from 0.5°C to 1.3°C on an extensive area of the Laptev shelf, constant temperatures ranging from 0°C to -0.8°C and -1.3°C were maintained on the upper boundary of the model. A model test at 0°C was carried out with the help of a Lukyanov hydrointegrator, taking into account the latent heat of fusion of the ice contained in the Quaternary deposits, while the model test at negative temperatures was performed on the YCM-1 computer. A geothermal heat flow of 0.045 kcal/m^2·h was applied at the lower boundary of the model located at a depth of 700 m. Initial temperature distribution in the rock mass was the same as in a deep borehole on the shores of Tiksi Bay on the Laptev Sea.[3]

Descriptions of coastal outcrops and boreholes located at different points of the coastal zone of Yakutia, as well as data on the laboratory findings concerning the thermal properties of permafrost in the Yana River delta,[1] were used to establish the composition and thermophysical characteristics of the Quaternary deposits. The geological section of the subjacent layer was the same as in the Tiksi borehole.

The modeling results show that the temperature of submerged permafrost increases by 5°-6° during the first 100 yr if the mean annual water temperature at a depth of 50 m is 0°C. By the end of

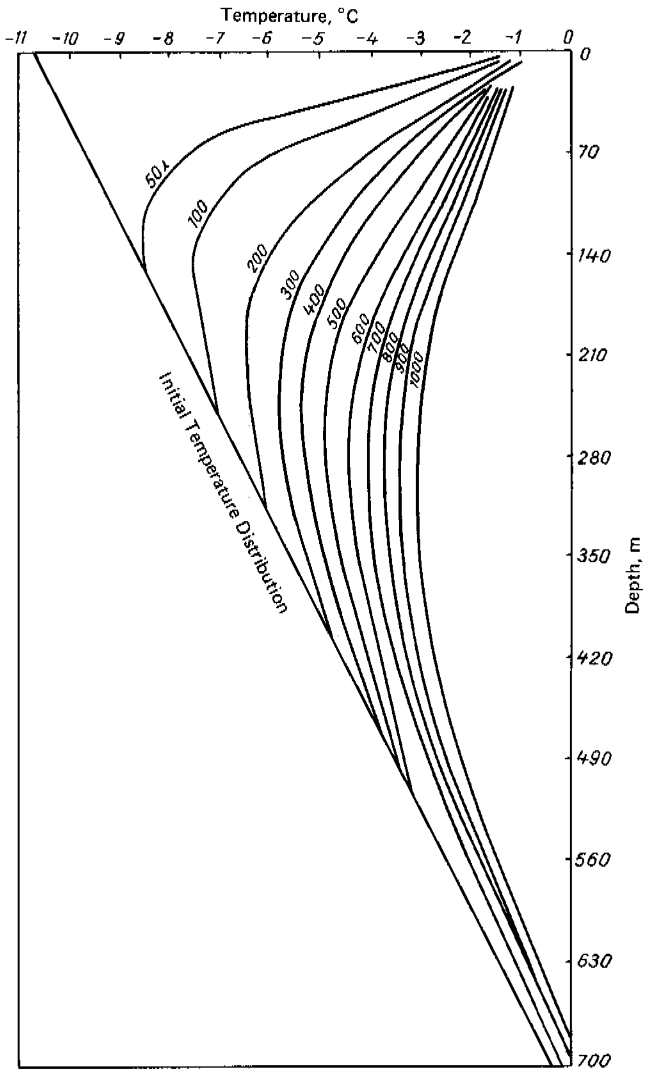

FIGURE 2 Changes in temperature of continental permafrost under the thermal influence of water with a mean annual temperature of -0.8°C.

a 400-yr period, the rate of temperature increase for 100 yr diminishes to 2° and by the end of the first 1,000-yr period to 1°. The temperature of permafrost in this case reaches approximately -1°C.

As seen in Figure 2, the temperature disturbance caused by an abrupt change in surface conditions penetrated to a depth of over 100 m with the mean annual water temperature at -0.8°C during the first 50 yr after submergence. The ground temperature at a depth of 50 m went up by approximately 4° during this time. After that, the rates of temperature increase suddenly diminished, amounting to ~1.5°C in the following 50 yr. By the end of the first 1,000-yr period, the general increase in the temperature of underwater permafrost amounted to about 9° at a depth of 50 m and 0.5° at a depth of 700 m.

With the mean annual water temperature at -1.3°C, the nature of the change in the temperature field of continental permafrost submerged in the sea is basically the same as in the case described above.

The results of actual observations and model tests show that the nature of temperature distribution in the ground with set conditions and in recently submerged permafrost differs.

In a zone with set conditions, a thicker rock mass, whose temperature corresponds to the mean annual water temperature, serves as a basement for thin layers with yearly fluctuations in temperature. The temperature gradient within this rock mass is in effect equal to zero. Below that, it will be positive in value. During the first several hundred years, the temperature gradient in the series of underwater permafrost of continental origin in the zone of active thermal abrasion will remain negative. Therefore, the presence of a negative temperature gradient at a depth of several dozens of meters below the surface of the seabed can be an indication of a relatively recent transition of permafrost to subaqueous conditions.

Table 1 shows the salinity of deposits found in boreholes 4 and 5. It indicates that the salinity of certain layers constituting the seabed is 2-11 times less in the zone of intensive coastal erosion than in the accumulation zone. According to the available classification, the beds in the first zone are considered nonsaline, while the deposits of the second are of average salinity.

The nonsalinity in the zone of thermoabrasive erosion shows that the permafrost is impervious to seawater of the indicated salinity. Apparently, the depth of salt penetration into the basement is restricted by the depth of summer thawing. The thawed layer is periodically removed by the waves, thus exposing the surface of the permafrost. The rate at which the underwater coastal slope is washed away is determined by the rate at which the permafrost of the seabed thaws. In such areas, permafrost can be found directly under the water at quite a distance from shore.

TABLE 1 Salinity of Deposits Found in Boreholes (in g/kg)

	Distance from the Seabed Surface,m														
Borehole	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
4	2.5	1.1	--	--	--	--	2.5	--	1.5	1.0	--	--	--	--	2.3
5	6.5	12.0	6.9	--	6.4	7.0	4.4	5.9	6.8	5.4	4.9	5.0	6.0	--	--

The results of cryopedological and geological investigations, carried out by the author in 1970, are proof of the extensive occurrence of nonsaline deposits on the southeastern outskirt of the Laptev shelf. With the help of a VPGT-59 vibro sampler, some 200 cases of samples were collected on different sectors of the shelf where water depth ranged from 4 m to 25 m. The length of the cores varied from 2-3 cm to 350 cm, depending on the composition and state of the material. At certain points of the shelf, the sampler could not penetrate the basement at all. Judging by the separate pieces extracted by the sampler, the beds were frozen.

Most of the selected samples underwent different types of laboratory analysis. The chemical composition of the aqueous extract was determined in particular. Approximately 40 of the 130 samples subjected to chemical analysis were nonsaline.

A number of sectors containing nonsaline material were found in Ebelyakhsky Bay in the Dmitri Laptev Strait (Figure 3).

Sectors of the seabed, composed of nonsaline deposits, may be "outcrops" of relict continental permafrost submerged at the beginning of the Holocene. However, it may well be that some of these "outcrops" are newer formations that were recently submerged as a result of thermoabrasive erosion of the seacoast.

It should also be pointed out that in certain cases the temperatures of phase transition of the ice contained in underwater permafrost of continental origin and in recent saline deposits may differ by 4°-5°. As was mentioned previously, the seabed in the area of Muostakh Island in the zone of thermal abrasion was frozen at a temperature of -1.0°C. At the same time, recent frozen deposits with temperatures up to -6°C[5] were discovered on the opposite side of the island in the accumulation zone. It follows that permafrost may sometimes be covered with frozen deposits of the same or even much lower temperature in some sectors of the sea.

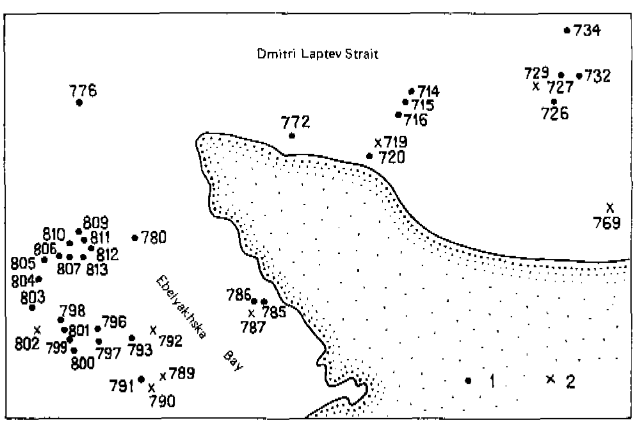

FIGURE 3 Sectors of the Laptev shelf where nonsaline deposits are found. 1--points at which the nonsalinity was established by means of chemical analysis; 2--points at which frozen pieces of material were extracted; the figures denote the core numbers.

The data given in this paper allow us to assert that with time a layer of new permafrost of considerable thickness may accumulate on permafrost submerged during the last transgression or as a result of thermal abrasion, in sectors of the Laptev Sea where negative mean annual water temperatures are observed. Especially thick layers of permafrost accumulate in semienclosed bays and lagoons to which the deposits are transported by currents and waves. The places of accumulation of such deposits are the eastern out outskirt of Vankin Bay and the vicinity of Muostakh Island, where their thickness exceeds 20 m. On the contrary, permafrost on the sectors of seabed subjected to erosion have no protective cover formed by recent deposits, are not saline, and are directly washed by seawater.

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LAKE THERMOKARST ON THE LOWER ANADYR' LOWLAND

S. V. TOMIRDIARO AND V. K. RYABCHUN Northeastern Integrated Research Institute, Far Eastern Scientific Centre, Academy of Sciences of the USSR

The studies of V. A. Kudryavtsev[5] were the first to provide convincing evidence that the formation of thermokarst lakes is not an indication of general degradation of the permafrost and a warming of the climate and that such lakes are also capable of actively developing at the present time, even in the very cold arctic tundras. For this to happen there must be widespread ground ice, an increase in the depth of seasonal thawing, and conditions that result in the protection of the emerging lake from drying out. These basic theoretical tenets opened the way for field and experimental studies of the thermokarst processes and their consequences. Based on the results of their own 9-yr program of field studies pertaining specifically to the lower Anadyr' lowland and also on the contributions of other investigators, the authors will discuss here the principal laws governing the development of lake thermokarst on plains of diverse origin.

The mechanisms involved in the development of lake thermokarst can be subdivided into general and regional. The former are characteristic of all northern lowlands; they are the general mechanisms involved in the build-up of heat in primary and growing bodies of water, development of sublacustrine taliks, and the thermal disintegration of shorelines. The latter pertain only to certain regions and are indicative of the overall course of the lake thermokarst process, as dictated by the previous history of development of the region. They are highly dissimilar on plains of marine and alluvial accumulation and on marshlands and deposits of the glacial complex. This subdivision is, to a certain extent, arbitrary, since the operation of the general mechanism is also expressed by way of the actual regional situation.

If it were not for the fact that new ice bodies have recently formed close to the ground surface, the thermokarst would only have been able to develop by reason of the upper Pleistocene fossil ice of the alluvial and marshland complexes. But, as is well known, in spite of the cessation of the blanket sediment accumulation process on the plains of the Northeast in the Holocene, even in our times multiple ice wedges are continuously forming when the conditions are suitable[4] in all of the arctic and subarctic lowlands.

There is no denying that this is ice of epigenetic origin. It develops most actively in the swamp-filled and breached lacustrine alas depressions, and also on young marine river terraces. As the multiple ice wedges grow, the soil and ice of the wedge polygon block undergoes bending in the direction of the vertical axis of the latter.

Ryabchun's studies indicate that it is this that results in the gradual disappearance of the mineral material above the ice wedges, which come to be protected by only a thin covering of moss.

Tomirdiaro's studies indicate that the simultaneous development of the low-center polygon type of microrelief with its characteristic inter-ridge grooves above the caps of the ice wedges leads to the origination of a primary sinkhole containing water, to thawing out of the ice wedges, and to development of lake thermokarst.

When the conditions are such that there is an excess of moisture and soils containing ice are prevalent, the development of the thermokarst lakes can continue indefinitely. Sooner or later, therefore, all of them will converge upon a river erosion system and become drained. Wedge ice will again begin to grow on the dried-up lake bottoms. The study of these alternating processes made it possible for Tomirdiaro to ascertain the general laws governing the polycyclical development of epigenetic frozen strata in arctic and subarctic lowlands; namely, the accretion of epigenetic multiple ice wedges inevitably leads to lake thermokarst, its disappearance inevitably leads to the accumulation of a new epigenetic regeneration ice wedge system, and so on.[8]

The thermokarst lakes swiftly extend their water area and merge to form larger bodies of water, thus continuously changing the cryological-geological and geomorphological conditions.

A most important characteristic of the thermal abrasion of shorelines is the trend towards continuous reconstruction of the vertical offshore scarp. This leads to unlimited widening of the lake, which is interrupted only by its breaching and drainage.[8,6,12]

Four types of thermal abrasion lake shorelines distinguished and studied by Tomirdiaro;[8,9] floating, trench-block shearing, notch-block undercutting, and erosion-solifluction flattening. Ryabchun distinguished and studied a fifth type--the erosion-solifluction nonflattening shoreline.

All five of these main types of shorelines are defined by various combinations of the following parameters: h_D--the extent to which the brow of the offshore scarp exceeds the level of the bottom of the water body; h--the possible settlement of the shore deposits in the course of their melting to an elevation which is below the lake; H_0--the depth of the oldest part of the lake that originated without the involvement of thermal abrasion; h_L--the extent to which the shore exceeds the level of the water; h_S--the maximum depth of seasonal thawing; h_w--the height of the waves (measured from crest to trough); S_0--the area of the oldest part of the lake that originated without the involvement of thermal abrasion; S--the ultimate area of the water body; $h_{av.}$--the weighted mean depth of the lake; h_{CT}--the critical depth of the lake at which freezing, and heaving, of the bed occurs (see Table 1).

The inhomogeneity of the conditions of thermal abrasion in the different directions frequently leads to the formation of oriented lakes.[7]

After the lakes have undergone degradation, an alas topography develops. The bottoms of the alasses become overgrown with meadow-steppe mixed herbage. With the development of the new system of multiple ice wedges and the establishment of swamp conditions, mosses and sedges replace the herbage. When sublacustrine taliks of coarse-grained deposits are present among the herbs, bulgunnyakhs (open-system pingos) develop. The original deposits are everywhere overlain by a layer of lake silts. This even made it possible for N. A. Shilo[13] to distinguish a special type of continentallithogenesis with large thermokarst lakes. The basic features of the deposits in lakes are: They are generally organic; they are less mineralized (i.e., the underlying soil); and they are finer grained from the bottom upward.

Within each region it is possible to isolate some zones of thermokarst lakes according to differing geologic-geomorphologic regions. In this connection in the lower Anadyr subdivision of the shore zone, there are alluvial, fluvioglacial, and glacier accumulations.[10-11] In the valleys of large tributaries in the various Anadyr zones, there are ellian accumulations (Figure 1).[9,11,13]

In the thermokarst lakes of the shore zone, there are accumulations of elongated discontinuous bands along the shores of the Anadyr, Kamchalan, and Onemen estuaries. Morphologically, they appear to be Holocene terraces. The zones of alluvial accumulations are discontinuous along the larger rivers. The thermokarst lakes in the zone have been studied and have been described by the theory of polycyclic growth of epigenetic layers.[8,9]

In the low-lying shore accumulations, it is possible to observe the mutual enclosure of dry lakes in alassial depressions, but an area seen to be imbedded in one another, the area between the alases amounting to as much as several percent. This means that in practically any sector of the marine plain, thermokarst lakes came into being and were destroyed several times in the recent past. These lakes are characterized for for the most part by shorelines of the first three types. The temperature conditions of the taliks are unstable.[10]

The greater part of the lower Anadyr' lowland is occupied by plains of glacial and fluvioglacial accumulation. In the northern part of the lowland, four belts of terminal moraine formations of the first late Pleistocene (Q^2_{III}) glaciations are observed. Three such belts[1-3] exist in the southern part of the lowland. The spaces between the terminal morainic ramparts are occupied by a sheet of fluvioglacial deposits.

The fluvioglacial lake-thermokarst zone is subdivided into subzones of outwash belts and bow-shaped depressions between the belts of the zones the development of the thermokarst lakes is patterned on polygenetic and epigenetic regeneration ice wedges, and also on texture-forming and injection ice. Typical of the subzone of outwash belts, occuring in the form of a 3-4-km strip on the outer slopes of the terminal moraine ridges, are some small lakes (diameters ranging from tens of meters to a few hundred meters), whose depths are between 1 and 4 m. The bottoms of these lakes have been leveled to a smaller extent than the bottoms of the thermokarst lakes on the marine plains, owing to the lack of uniformity in the surface conditions that led to the development of the lakes and because the summer winds in the interior parts of the lowland are not as strong or as frequent as they are on the coast. In the case of the thermokarst lakes of the outwash belts, trench-block shearing shorelines are most typical; notch-block undercutting shorelines occur quite frequently; floating shorelines are almost completely absent; and erosion-solifluction shorelines are relatively rare. The lake basins are frequently asymmetrical, with higher upland and lower piedmont shorelines. The total area occupied by the lakes and alases within the outwash belts is relatively small and it is apparently no more than a few percent. The good drainage of these regions serves to hinder the widespread development of lake thermokarst.

Here also, however, well-developed alasses are frequently encountered, which attests to a repeated succession of ice and lake formation in some sectors of the outwash belts. The ice- and lake-formation processes in the bow-shaped de-

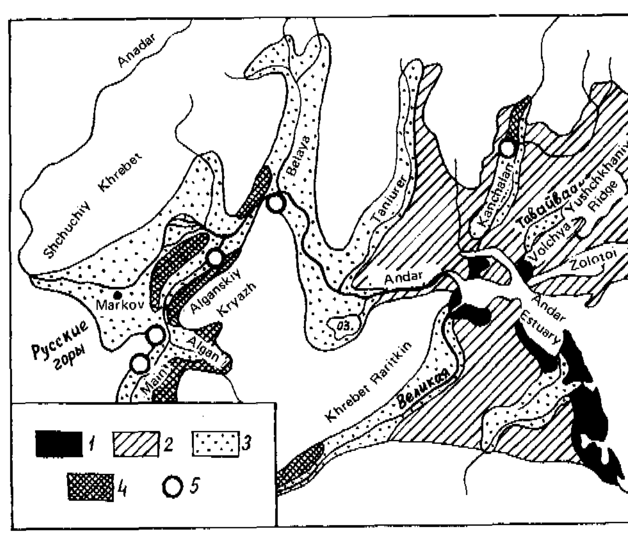
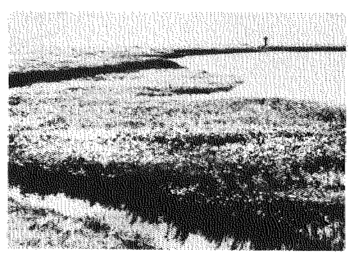
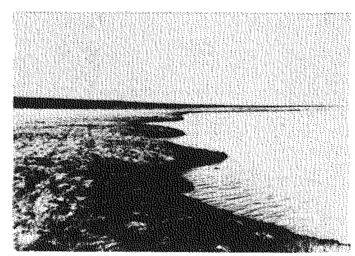
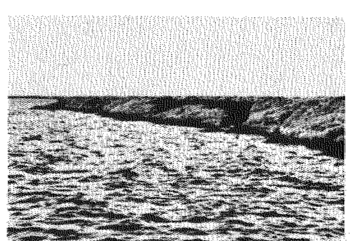
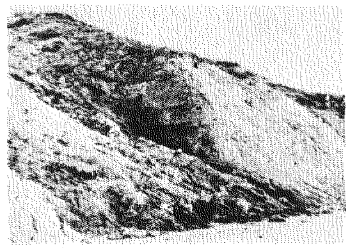
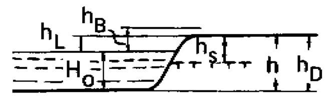
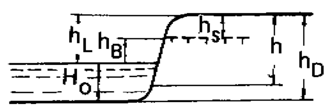
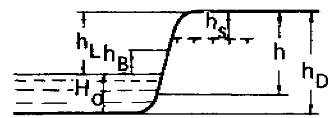
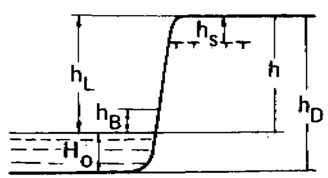
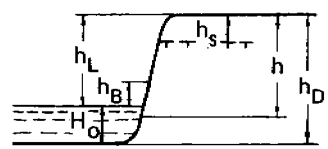


FIGURE 1 Schematic map of lake-thermokarst zones of the Anadyr' lowland (after Tomirdiaro). 1--zone of flat-basin lakes and alasses on the Holocene marine terrace; 2--zone of glacial-thermokarst lakes and the lakes on fluvioglacial plains; 3--flat-basin lakes and allases on alluvial plains; 4--loess-glacial plain with cave-in-basin lakes; 5--sites of investigated loess-glacial scarps.

TABLE 1 Main Types of Thermal Abrasion Shorelines (after Tomirdiaro)[a] and Schematic Drawings of Their Development (after Ryabchun)

Types of Shorelines	Examples of Typical Shorelines	Relationships between the Basic Parameters of the Shorelines	Conditions Determining the Type of Shorelines	Rate of Retreat of the Shoreline, m/yr	Remarks
1. Floating			$h_L \leqslant h_s$; $\frac{h_D - h}{H_o} = 0$	1; $t = 2$ yr	No accumulative shoal
2. Trench-block shearing			$h_L = h_s + h_B$; $\frac{h_D - h}{H_o} > 0$	0.5; $t = 5$ yr	No accumulative shoal

Types of Shorelines	Examples of Typical Shorelines	Relationships between the Basic Parameters of the Shorelines	Conditions Determining the Type of Shorelines	Rate of Retreat of the Shore-line, m/yr	Remarks
3. Notch-block undercutting			$h_L > h_s + h_B$; $\frac{h_D - h}{H_o} > 0$; $h_D - h \ll H_o$	5–10, $t = 4$ yr	Intermittently emerging accumulative shoal
4. Erosion-solifluction nonflattening			$h_L \gg h_s + h_B$; $\frac{h_D - h}{H_o} \gg 0$; $h_D - h < H_o$; $\frac{H_oS_o + [H_o-(h_D-h)]}{S_o + S} = h_{ar} < h_{cr}$	2.5; $t = 8$ yr	Developing accumulative shoal
5. Erosion-solifluction flattening			$h_L \gg h_s + h_B$; $\frac{h_D - h}{H_o} \gg 0$; $h_D - h < H_o$; $\frac{H_oS_o + [H_o-(h_D-h)]S}{S_o + S} = h_{av} > h_{cr}$	0; $t = 6$ yr	Stable accumulative shoal

[a]Type 4 was distinguished and studied by Ryabchun.

pressions between the terminal morainic ramparts are more intensive. These smooth depressions, measuring up to 10 km in width, cut across the interfluvial region of the major rivers and have a noticeable gradient only in the approach to their valleys. According to rough calculations, the total area of the alases and lakes ranges in some of the sectors between 10 and 30 percent. Not infrequently, lakes with diameters of 1-3 km are observed here. The depth of the lakes varies between 1 m and 4 m; for the most part it is 2.0-2.5 m. The bottoms of the large lakes are smooth and flat. The smaller lakes have a differentiated bottom, whose relief is due to the lack of uniformity in the conditions that governed the development of the lake and to the modest degree of wave leveling. Trench-block shearing shorelines are highly typical of the thermokarst lakes of the bow-shaped depressions, and quite frequently, shorelines of the other types as well. Pingos are a distinguishing feature of the fluvioglacial lake thermokarst zone. Ryabchun noted 14 pingos between the second and third belts of the terminal morainic ramparts in the northern part of the lower Anadyr' lowland. They are situated in alases. Their heights vary from 2 m to 15 m and the diameters of their bases from 20 m to 100 m. Within the fluvioglacial plains of the lower Anadyr' lowland, the pingos form a regular pattern, since it is only here that there are large volumes of coarse grained outwash deposits in which reserves of free water become concentrated with the initiation of the thermokarst lakes and the development of sublacustrine pseudotaliks.

The development of the thermokarst lakes in the glacial accumulation zone is patterned on the buried remains of glacier ice. The basins of these lakes are oval and funnel-shaped, the the lakes being 5-20 m deep. The margins of the basins rise above the water surface by as much as 20-25 m. Occasionally, the surface dimensions of the lakes reach 200-300 m. They are arranged in the form of cascading chains extending from the crests of the morainic hills to their feet. There are from two to five lakes in one chain. The lakes of each chain are fed by a single stream. Older lakes with relatively stable shorelines are distinguished, the depths of which rarely attain 10 m. Comprising a large group are lakes with erosion-solifluction shorelines that are actively disintegrating at the present time. They are from 10 m to 20 m deep, the center line being displaced in the direction of the least stable shoreline.

Serving as proof of the burying of large masses of glacier ice during the retreat of the glaciers are first and foremost the glacial thermokarst lakes themselves. Ordinarily, small continental bodies of water rapidly develop the shoreline profile until a state of dynamic equilibrium is attained. It is only the thermokarstic lakes and major reservoirs during the first 10 yr of their operation that form an exception to this rule. The shorelines of the lakes being discussed are currently actively disintegrating, which bears witness to their thermokarstic origin. Sometimes, parallel retreat of the offshore scarps is observed, which is possible only when there is submarine lateral melting of the ice lying beneath the steep offshore slope. The great depth of the lakes is due to the melting of thick layers of ice, which under the prevailing geological and geomorphological conditions (the enclosing soils are rubbly suglinoks; the relief component morainic ramparts) could have formed only through the burying of glacial remains. The prolific nature of the thermokart lakes (they number in the hundreds) attests to the considerable areal extent of the buried ice.

The mineralization of the ice classified by us as buried glacier ice is 10-20 mg/l and is equivalent to the mineralization of snow. Nevertheless, it is approximately an order less than the mineralization of the other ground ice formations of the lower Anadyr' lowland.

Thus, during the retreat of the upper Pleistocene glacier, large masses of "dead ice" were preserved, owing to the freezing of the morainic blanket and the cooling of the buried glacial remnants. Subsequently, with denudation of the hilly morainic relief, melting of the ice and the formation of glacial thermokarst lakes occurred. This process is continuing to take place at the present time. The drainage of the glacial-thermokarst lakes and the freezing of the sublacustrine taliks is followed by the formation of multiple ice wedges and texture-forming ice, with the subsequent origination of normal thermokarst lakes. These widen the original basins, even though to a less degree. The areas occupied by basins of this type, however, are very small.

It has been demonstrated that, irrespective of the sizes of the lakes, the thickness of the sublacustrine taliks on a marine and alluvial plain does not exceed a few dozen meters.[10] This is due to the short-lived existence of the lakes under conditions characterized by the complete emergence of a polycyclical alternation of ice and lake formation. On fluvioglacial and glacial accumulation plains, where the mobility of thermokarst lakes is restricted by the lack of uniformity in the conditions governing the development of the lakes in plan, and consequently, the probability of their rapid breaching and drainage is sharply reduced, the thickness of the sublacustrine taliks should be greater. As regards glacial-thermokarst lakes and those that have been blocked up, beneath many of them there are open taliks.[10] Young lakes with active thermal abrasion shorelines form an exception to this rule.

Occupying a special place is the lake thermokarst within the Anadyr' marshland, which has been most fully studied by Tomirdiaro[9,11] in the valley of the Main River. The thickness of the dusty loesslike supesses and suglinoks in this marshland is 15-20 m. They are penetrated from top to bottom by thick syngenetic ice wedges, contain osseous remains of upper Paleolithic mammals and are also penetrated by rootlets of herbs. In all of their features, these deposits are analogous to those of the marshland complex on the eastern Siberian lowland. Until very recently both kinds were considered to be deposits of the upper Pleistocene alluvial terraces.[6] Shilo[13] was the first to point to the possibility that the deposits of the marshland complex are not of alluvial origin. More recently, on the basis of paleogeographic, paleoclimatic, paleontological, and cryolithological analyses, Tomirdiaro[9,11] brought forward a number of arguments in favour of the eolian-cryogenetic origin of the deposits of the marshland complex. As is well known, these deposits were formed in the Zyrianian and Sartanian glacial phases, which were characterized by a high degree of continentality and aridity of the climate and also by the very active development of subaerial processes and the accumulation of eolian loess blankets. This assumption, which is known to hold true for Europe, and also for those regions of western Siberia and North America that are contiguous with the Soviety Northeast, is now being extended to include northern Yakutia and the Chukchee Peninsula, following studies by Gerasimov.[9] The very fact of the discovery of loess-glacial marshlands in Chukotka and the Anadyr' basin[11] has forced us to reconsider the existing concepts with regard to the exclusiveness of this region from the standpoint of the absence there of the marshland mantle and corresponding thermokarst landscapes that characterize the eastern Siberian lowland.[2]

The development of lake thermokarst through the ice-rich marshland deposits is especially rapid and gives rise to the formation of an alas topography containing craters and deep sinkholes. In the alasses conditions are established that favor the complete emergence of polycyclical ice- and lake-forming processes. By virtue of the great thickness of the marshland deposits, which are 70-90 percent ice-saturated, the thermokarst lakes there have very deep basins. Depending on the height of the water level, shorelines that are of the floating, notch, or solifluction type form there. In this event, total disintegration of the marshland occurs and typical Holocene lake-alas plains of the maritime lowland type form (Figure 2). Only isolated loess-ice hillocks remain on such a plain among the flat-basin lakes and swamps, forming the so-called hillocky-marshy lake landscape.

In many of their features--the saturation with macroremains of trees, the extensive polygonal network of epigenetic wedge ice, and the absolute dating of individual specimens--all of the deposits of the thermokarst lakes and alases that have been studied within the Main-Anadyr' upper Pleistocene marshland are undoubtedly of Holocene age. This supports the well-known thesis that contemporary lake thermokarst, as a general phenomenon, did not originate until the conclusion of the last glacial episode and reached the peak of its development at the time of the Holocene climatic optimum.[9]

On the lower Anadyr' lowland, both of the principal types (in terms of their overall course of development) of lake thermokarst are manifested: the nonreversible type--on the plains of glacial and eolian-cryogenetic accumulation--and the polycyclical type--on all of the other plains and on those of eolian and glacial accumulation that have been reworked by nonreversible thermokarst. Furthermore, the role of the thermokarst lakes in the polycyclical transformation of the natural setting depends on the genesis of the deposits in which the lakes have originated, and gradually diminishes in a number of plains of marine, eolian, alluvial, fluvioglacial, and glacial accumulation, although nowhere does it wholly disappear, even when the historically formed conditions are highly unfavorable.

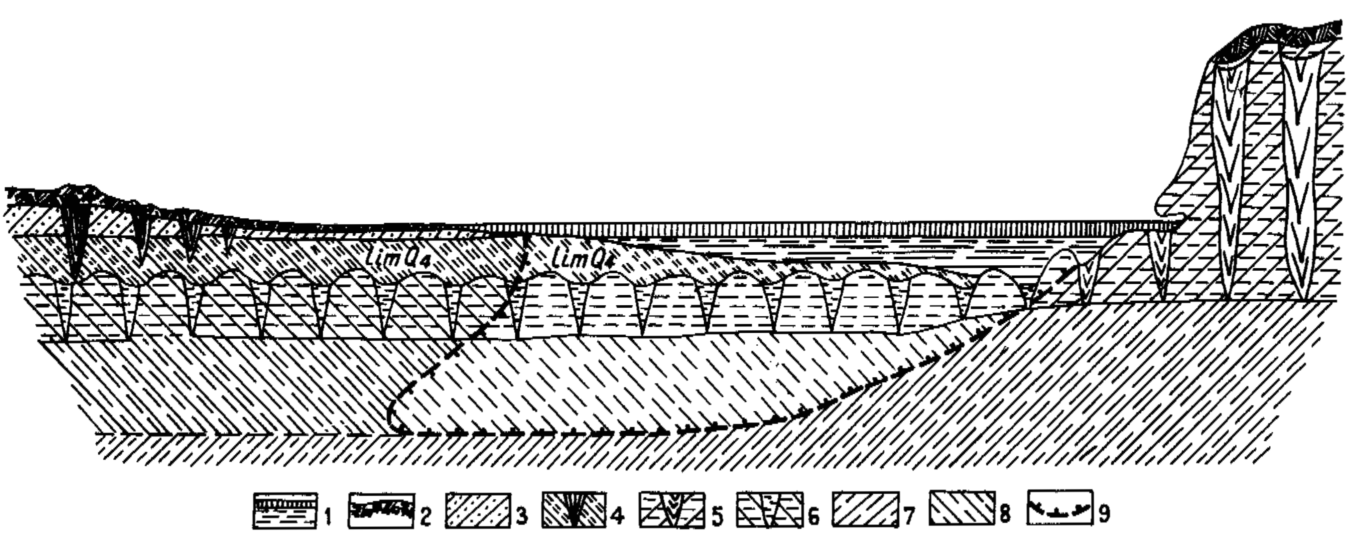


FIGURE 2 Diagram of the principles of general disintegration of loess-glacial plains and the formation of young lacustrine-alluvial lowlands (after Tomirdiaro). 1--thermokarst lake, advancing towards loess-glacial plain; 2--layer of young peatlands; 3--layer of peat-containing alas deposits with extruded veins; 4--contemporary ice wedges (epigenetic type); 5--steeply plunging fossil ice wedges dating from glacial times (syngenetic type); 6--soil wedges (cavities from melted veins filled with lake silt); 7--loesslike deposits containing ice, dating from glacial times; 8--deposits that have become thawed out and compacted in the sublacustrine talik; 9--boundary between lake and moving sublacustrine talik.

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PERMAFROST AND VEGETATION

A. P. TYRTIKOV Moscow Lomonosov State University

The existence of a close correlation between permafrost and the vegetation has been established in the writings of many investigators.[1,4-8] This correlation is due to the interaction between the vegetation and low-temperature processes.

Vegetation as a distinct transitional zone changes the conditions of heat and moisture exchange between ground and atmosphere. The effects of a vegetative cover on permafrost are:

1. It intercepts and reflects most of the solar radiation (often more than 90 percent). In summer, the diurnal radiation balance beneath the forest canopy is at least 10 to 20 times less than it is above the forest.[9]
2. It lessens the turbulent exchange of heat between the ground and the atmosphere by decreasing the wind speed.
3. It hinders the radiation of heat by the ground.
4. It hampers the exchange of heat between the ground and the atmosphere on account of its low thermal conductivity and as a result of the decrease in the thermal conductivity of the soil. In this respect the roles of the mossy cover and peaty layers of soil are especially important. By reason of their drying out during the warm season of the year, the moss and upper layer of peat become such poor heat conductors that almost all of the heat being received from the sun is returned to the air in its entirety. In the permafrost region the moss and peat normally freeze in the wet state, with the result that their thermal conductivity is dozens of times greater than when they are in the dry state. Consequently, while doing little to hinder the outflow of heat from the ground in winter, the moss and peat strongly impede influx to the ground in summer, thereby contributing to the predominance of freezing over thawing. It is these properties of the mossy cover and peat that serve to explain the confinement of islands of permafrost situated near the southern limit of permafrost to the moss-covered forests and peatlands.

Another reason for the decrease in the thermal conductivity of the soil beneath the vegetative cover is that voids form following the decay of plants. Organic residues dry out as moisture is absorbed by the roots (this also leads to a reduction in the thermal capacity of the soil).

5. Beneath the cover there is a 2 to 10 times reduction in evaporation from the surface of the soil and in the expenditures of heat on this process, as compared with unvegetated zones.

6. When the vegetation holds a large quantity of precipitation, the amount of heat taken in by the ground along with the precipitation (or the loss of heat from the ground) is reduced.

7. Vegetation facilitates the intake of heat along with precipitation by the ground (or its loss from the ground) by restricting surface drainage and favoring the percolation of moisture into the soil.

8. The vegetation contributes to a decrease in the loss of heat by the ground in winter by favoring the accumulation of snow and its deposition in the loose state.

9. In dense cedar, spruce, silver fir, and pine forests, up to one-third of the snow is held in the crowns of the trees, with the result that conditions are set up for more pronounced cooling and freezing of the ground than is the case in deciduous forests.

The effect of the vegetation on the heat and moisture exchange between the ground and the atmosphere varies the seasonal and long-term dynamics of the vegetation and depends on the composition and structure of the vegetative cover and on the conditions in which it develops.

By exerting an appreciable influence on the heat and moisture exchange between the ground and the atmosphere, vegetation plays a major role in the temperature conditions, freezing and thawing of the ground, the permafrost dynamics, and other low-temperature processes.

Beneath the vegetation there is a reduction of the ground temperature ranges. The ground temperatures are lower in summer and higher in winter than they are in the exposed areas. The differences between the July and January mean monthly ground temperatures of the forested and cultivated areas reach 15°-17°.[2]

The effect of the vegetation on the mean annual ground temperature depends on various conditions and, particularly, on the snow cover. Beneath the trees and tall shrubs in the forest tundra of western Siberia, where about 100 cm of loose snow accumulates, the ground temperatures are 3°-5° higher than in the surrounding tundras.

The depths of thawing and freezing are lower beneath a vegetative cover than in areas where there is no vegetation. In the permafrost region the thawing of the ground is influenced mainly by the mossy covering and peaty layers of soil. The depth of thawing beneath these is only 25-50 percent of that in the exposed areas.[8]

The freezing of the ground is influenced mainly by the woody vegetation and shrubbery, since these contribute to the build-up of loose snow. In regions where there is little snow, the freezing of the ground is greatly reduced by the poorer heat-conducting mossy covering and peaty layers of the soil.

The vegetation decreases frost-cracking of the ground, preserves the permafrost, and hinders the development of thermokarst, erosion, solifluction, mudflows, and landslips.

THE EFFECT OF PERMAFROST AND LOW-TEMPERATURE PROCESSES ON THE VEGETATION

The effect of the permafrost on the vegetation is mainly expressed by changes in the soil. Permafrost causes the following soil variations:

1. *Soil swampiness*--a consequence of the imperviousness of the frozen ground and also of the condensation of water vapor from the air.

2. *A lessening of the aeration of the soil*--a consequence of its swampiness. The suprapermafrost layers of the soil are so saturated with moisture that the air is completely squeezed out of them, besides which it is impossible for oxygen to be brought in by the flow of water. A consequence of the worsening aeration of the soil is its gleying.

3. *A lowering of the soil temperature.*

4. *The soil becomes depleted of the mineral elements taken up by the plants* and of mineralizing organic remains, a consequence of the weakening of the vital activity of the organisms under conditions of low temperature and poor soil aeration.

5. *The accumulation of the organic remains* in the soil and on its surface (the formation of the peaty layer), by reason of their poor mineralization, also leads to the soil becoming depleted of the mineral elements taken up by the plants, since a large part of these elements is retained in the peaty layer in a state of inaccessibility to the roots and is thus excluded from the circulation.

These changes in the soil give rise to suppression of the growth and development of flowering plants, besides which they set up conditions favorable for the establishment of marshy vegetation and swampiness among the woods, thickets, and tundras.

The extensive development of rootless mosses and lichens in the permafrost zone is in part attributable to their being less dependent on the soil conditions.

The condensation of airborne water vapor in the soils of the permafrost region is an important source of the plants' water supply.

The retention of moisture in the soils due to its inability to percolate downwards, its influx into the root inhabiting layer in the course of the gradual thawing of the soil during the summer, and also the condensation of water vapor in the soils give rise to favorable conditions in the arid regions (for example, central Yakutia) for the springing up of trees and for agriculture without irrigation.

The following are the low-temperature processes that have the most noticeable effect on the vegetation:

1. *Frost-cracking of the soil*, besides being responsible for direct damage to the vegetative cover (the breaking of roots and stems, etc.) provides an impetus for the initiation of many freezing processes that disturb or destroy the vegetation, namely, mudflows, landslips, erosion, thermokarst, and solifluction.

2. *Thermokarst* (the formation of subsidence relief features as a result of the melting of the ice in the soil) often develops with such rapidity that it destroys the vegetation over a large area. The thermokarst establishes new habitats (exposed slopes and water bodies), which again become overgrown.

3. *Erosion* is more intensive in regions characterized by the occurrence of noncohesive soils containing ice than it is outside the permafrost zone. The reasons for this are as follows:

(a) Subsequent to thawing out, the frozen soils become unstable and are easily eroded.

(b) Above the frozen ground, water accumulates, which, after draining along the gradient of the frozen surface and setting up favorable conditions for soil creep (mudflows and landslips), flows out at the surface of the slopes and erodes them.

(c) The erosion is everywhere accompanied by melting of the ground ice. The water that forms during this process both adds to and intensifies the erosion.

(d) The erosion is usually accompanied by subsidences of the ground, which increases the gradient of the bed of the temporary watercourse and adds to the intensity of the erosion.

(e) The erosion continues unceasingly throughout the warm season of the year.[3] This results in the destruction of the vegetation and the establishment of new habitats.

4. *Mudflows and landslips* destroy the vegetation on the slopes of hills, mountains, terraces, etc., and by exposing the soils containing ice set up conditions for thermokarst and erosion.

5. *Solifluction* (a slow-moving flow of thawed soil under the influence of gravity), by causing local disturbances in the vegetative cover and weakening the soil consolidation, also frequently gives rise to conditions of thermokarst and erosion.

6. *Soil heaving* also appreciably alters the vegetation. The slow heaving that takes place during the formation of hummocky tundras, coarsely hummocky peatlands, and open-system pingos, although it gives rise to no marked disturbance of the vegetative cover, results in the establishment of new habitats and favors a more variegated and heterogeneous vegetation. Rapid heaving (with the formation of seasonal and icing hummocks) frequently results in death of the vegetation.

7. *Naleds* give rise to habitats that differ from the surrounding ones in having a shorter growing period, a lower temperature, and higher soil humidity and are characterized by a specific type of vegetation (usually swamp-tundra).

These low-temperature processes give rise to highly diverse habitats and contribute to the unusual variegation and heterogeneity of the vegetation in the permafrost region.

A close correlation between the vegetative cover and the low-temperature processes is also observed in the dynamics of the vegetation.

The vegetation is constantly changing, thereby giving rise to corresponding variations in the freezing and thawing of the ground and in the permafrost. These variations in turn affect the vegetation.

Of the different kinds of changes in the vegetative cover, the greatest influence on the freezing and thawing of the ground and on the permafrost is exerted by gradual changes in the development of the vegetation and also by catastrophic changes in it.

The extent of the relationship between the low-temperature processes and changes in the development of vegetation increases in the north-south direction. This situation is well illustrated in the case of western Siberia.

In the subzone of lichenous, mossy tundras of western Siberia the changes in the vegetation that occur when the water-bodies and exposed areas of dry land are being overgrown do not cause major alterations in the permafrost. They are accompanied by a decrease in the depth of thawing of the ground and by a reduction in the low-temperature and other physical processes in it. Beneath a vegetative cover in the concluding phases of development, the depth of thawing of the ground is only a quarter to a half what it is in areas where there is no vegetation. Consequently, the freezing, thawing, heaving, settlement, ice-segregation, and solifluction take place in a layer that is 2-4 times less thick and give rise to correspondingly shallower variations in the ground. The dense, mossy-lichenous covering and peaty layer of soil measuring 5-15 cm thick, which are typical of the concluding phases of development of the vegetation, completely eliminate erosion and thermokarst. Since a relatively thin (5-20 cm) vegetative cover does little to retain the snow, there is only a slight abatement of cooling and frost-cracking of the soil in the main poorly heat conducting peaty-mossy layer.

In the subzone consisting of scrub tundras and in the northern forest tundra the changes in the vegetation appreciably affect the permafrost, mainly on account of the shrubs and mosses. Establishment of the shrubs gives rise to conditions favoring the accumulation of loose snow and thereby contributes to a decrease in the cooling, freezing, and frost-cracking of the ground. The development of the mosses and eventual formation of a peaty layer as a result of their dying off are accompanied by a progressive decrease in heating and thawing and by an abatement of the physical processes in the ground.

In areas where shrubs are developed, three stages are distinguished in the dynamics of the permafrost. These correspond to definite phases in the development of the vegetation.

Stage I--the degradation of the permafrost--corresponds to the development of shrubs with a herbaceous cover or of thin forests with an underbrush of these shrubs. During this stage it is only the upper regions of the soil that freeze, and the soil's seasonally frozen layer annually thaws during part of the year, following which the subsoil thaws, forming open taliks, the soil temperatures being $3°-6°$ higher than they are in the tundra areas.

Stage II--the restoration of the permafrost--corresponds to the development of a mossy cover in the shrubbery and of tundras replacing the latter. During this stage, as the mossy cover in the shrubbery develops there is a lessening of the heating and thawing of the ground, and the taliks become frozen. Also, the soil deteriorates, the shrubs are succeeded by tundras, and the ground temperatures are $3°-6°$ lower than in the areas containing the shrubs and herbaceous cover.

Stage III--the stable condition of the permafrost--corresponds to the tundra phases in the development of the vegetation.

In the subzone consisting of thin stands of trees and in the southern forest tundra, the relationship between vegetation and permafrost is most pronounced. The vegetation is the leading factor determining alterations in the freezing and thawing of the ground and in the permafrost. This role of the vegetative cover is due in the main to the development of mosses, trees, and shrubs. As the vegetation changes, the relationship between the above-mentioned components, which to a large extent determine the heat exchange between the ground and the atmosphere, is substantially altered. The changes are therefore accompanied by deep-seated alterations in the permafrost. The latter develops, undergoes degradation, or is in a more or less stable state in accordance with the dynamics of the vegetation.

Three stages are distinguished in the development of the permafrost. These correspond to definite phases in the development of the vegetation.

Stage I--the degradation of the permafrost--corresponds to the development of birch groves or shrubs in which there is no continuous mossy cover and of swamps containing surface water. During this stage it is usual for open taliks to form and for the ground temperatures to be from $2°-5°$ higher than in the tundra-covered areas. The development of the mosses and the accumulation of peat in the forests, thickets, and swamps inevitably lead to a situation favoring the formation of permafrost, which signals the onset of the next stage.

Stage II--the formation of the permafrost--corresponds to the development of moss-covered forests and thickets, their replacement by swamp-ridden forests, thin forests and thickets, and then by tundras. The stage is characterized by a lowering of the ground temperature, an increase in the thickness and ice content of the permafrost, and a decrease in the depth of thawing with the buildup of the mosses, the accumulation of the peat, and the thinning and dying out of the tree growth and thickets. Having formed under the influence of the development of a specific vegetative cover, the permafrost subsequently alters in accordance with the dynamics of the vegetation. It is, however, the permafrost that to a great extent determines the dynamics of the vegetation. By contributing to swamp formation, a worsening of soil aeration, its depletion of the mineral elements taken up by the plants, and a lowering of the temperature and depth of thawing of the ground, the permafrost is conducive to the development of swamp vegetation and to the replacement of the forests and thickets by swamps, and of the latter by tundras.

Stage III--the stable condition of the permafrost--corresponds to the tundra phases in the development of the vegetation.

The development of the vegetation in ground subject to heaving establishes conditions favoring the formation of hummocky tundra, coarsely hummocky peatlands, and other topographic features.[8]

Catastrophic changes in the vegetation (its destruction) abruptly alter the freezing and thawing of the ground, the low-temperature processes, and the development of the permafrost. As the permafrost region is opened up, there is an increase in the area in which the vegetation is destroyed or disturbed. Following the destruction of the vegetation, there is a doubling or quadrupling of the depth of thawing of the ground, as compared with the undisturbed zones. In areas where there are fine-granular soils containing ice, the destruction of the vegetation is followed by the intensive development of thermokarst and erosion, which leads to disintegration of positive topographic features and of structures situated on them. Frequently, erosion and thermokarst are set in motion after a small local disturbance of the vegetation. Following the passage of tractors and cross-country vehicles, which interfere with the integrity of the vegetation along the route, after a few years deep gullies form, which gradually become wider and destroy the vegetation in areas far removed from the original disturbed region.

Solifluction also becomes more intensive after the vegetation has been disturbed. The reasons for this are as follows:

1. There is a decrease in the reinforcement of the soil by the vegetation.
2. Frost-cracking of the ground becomes more intensive on account of the reduction in snowfall, and the gaps in the vegetation become greater.
3. There is an increase in the depth of thawing and, consequently, in the weight of the thawed layer of the ground.

The destruction of the vegetation is frequently followed by the development of mudflows and landslips, as well as by increases in the frost-cracking and heaving of the ground. It is in this way that favorable conditions are established for thermokarst and erosion.

The preservation of the vegetation in the permafrost region is the most important of the measures that must be taken to protect positive relief features consisting of noncohesive soil from disintegration by erosion and thermokarst.

When engaged in any course of action that results in disturbance or destruction of the vegetation on soils containing ice, steps must be taken to guard against the development of catastrophic low-temperature processes (thermokarst and erosion), especially on slopes.

The close correlation between the vegetation and the low-temperature processes serves as a basis for using the vegetation as an indicator of the depths of freezing and thawing of the soils and of their temperature and state (frozen or thawed).

The vegetation serves as an index of the direction of development of the permafrost (degradation or formation), and as an indicator of the age and rate of formation of the permafrost and low-temperature relief features. The vegetative cover can be used as a basis for distinguishing the stages in the development of the permafrost and for predicting its dynamics, both during the course of development of the vegetation and following its disturbance or destruction, in connection with the development of the area.

The patterns revealing the correlation between the vegetation and the low-temperature processes serve as a basis for working out rational measures aimed at controlling the freezing and thawing of the ground and the development of permafrost; protecting the vegetation; averting the catastrophic consequences of the disturbance or destruction of the vegetation; the thermal improvement of the soils and ground for purposes of construction, agriculture, sylviculture, and afforestation of the tundra; and changing the microclimate of individual regions.

These measures are apt to differ in the various regions, and their working out necessitates making a careful study of the specific nature of the regional patterns characterizing the correlation between the vegetation dynamics, the low-temperature processes, and the development of permafrost.

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ZONAL AND REGIONAL CHARACTERISTICS OF THE PERMAFROST IN CENTRAL SIBERIA

S. M. FOTIEV, N. S. DANILOVA AND N. S. SHEVELEVA Operations and Research Institute for Engineering Site Investigations (PNIIIS)

Central Siberia is a unique geocryological region, with well-defined geological-tectonic and physical-geographic boundaries. It differs essentially from the neighboring regions: western Siberia and the mountainous areas of southern and northeastern Siberia.

The geocryological conditions of central Siberia have been studied very erratically. The most detailed study was that of the Yakut Province.[1,2,5-7,11,17,18] In the Anabar, Olenek, and Vilyui provinces, geocryological studies were conducted for the most part in small areas.[2,8,11,15,19] In the Putoran Province, the most detailed studies were those of the Igarka and Noril'sk districts.[9,12,13,22] The Taimyr and North Siberian provinces were studied mainly along the Arctic coast.[5,12,14,16] In the Yenisei Province it was chiefly the lower geocryological zone that was studied.[9] In the Angara-Lena Province the valley of the Angara has been studied relatively well.[3,10] The least studied areas continue to be Tunguss Province, the southern and eastern parts of Putoran Province, and the western part of Vilyui Province.

The steady uplift of the region that occurred during the Neogene-Quaternary stage of tectogenesis with amplitudes of up to 300-500 m resulted in the formation of a plateau-shaped relief with absolute elevations of 400-800 m, dissected by a network of river valleys, incised to the extent of 200-250 m of more. In the areas that had undergone uplift to 1,000 m, platform-type mountains rising to 1,500-1,700 m and with erosional downcutting to 500-800 m, were formed. As a result, over a large part of the area, the thickness of the Quaternary deposits is no more than a few meters. It is only in the regions of relative subsidence that lowlands were formed, within which the thickness of the Quaternary deposits exceeds 100 m.

The climate of the region is markedly continental. The mean annual air temperature on the Taimyr Peninsula is -16°; on the Angara Plateau it is -2°. The sum of the mean monthly temperatures during the cold season varies between 200°C and above in the North and 80°C in the South. The annual amount of precipitation ranges from 150-200 mm in the North to 350-500 mm in the South. This appreciable variation in the main climatic indices, which is due to the enormous north-south extent of the region (of the order of 3,000 km), even resulted in the formation of latitude-zone inhomogenieties of the geocryological conditions.

The uniqueness of the region is determined by its confinement to the ancient Siberian Platform, the basement of which was formed in the early pre-Cambrian period.

Over the greater part of the region the crystalline rocks of the basement are overlain by a mantle of sedimentary and volcanic formations. Repeated changes in the tectonic regime predetermined the variable structure, composition, and thickness of the rocks forming the mantle and gave rise to substantial differences in the hydrogeological conditions in certain of the region's structures.

The buildup of salt water and brines in the lower Paleozoic rocks, the accumulation of fresh and brackish waters in the Mesozoic rocks, and also that of limited sources of fresh water in the joints of the mantle of waste consisting of crystalline rocks--specific features of certain of the region's geological-tectonic structures--played a major role in the evolution of the thickness and structure of the permafrost.

Beginning with the lower Quaternary period, i.e., over the full extent of the period in which the permafrost was formed, repeated changed occurred in the latitudinal conditions of heat exchange. For the region as a whole, the changes were not well defined. An analysis of the latitudinal heat-exchange characteristics throughout the Quaternary period has made it possible to distinguish two geocryological zones: a northern and a southern.

In the northern zone the harsh climatic conditions persisted throughout the entire period of cooling, which resulted in almost continuous cooling and freezing of the ground. Occasional rises in temperature did not result in substantial changes in the perennial freezing process. It is for this reason the permafrost in the northern zone is the oldest (Q_1, Q_2^1) and is characterized by continuity, great thickness, and low temperature.

In the southern zone, repeated and substantial changes in the climatic conditions occurred throughout the freezing period, which resulted in a succession of perennial freezing and thawing processes. During Samarovo, Taz-Sanchugovka, Zyrianian, and Sartan times (periods of perennial freezing), the southern limit of permafrost moved southwards for a considerable distance and passed outside the region. In Tobol, Messo-Shirta, and Karginskii times (periods of perennial thawing), the southern limit of permafrost climbed northwards to latitude 60°-64°. Permafrost degradation from the surface of the southern zone was greatest during the period of the Holocene thermal maximum. During this period, in the west of the region, the southern limit of permafrost passed near latitude 66°, in the center of the region--near 62°--and in the east--to the south of 60°. In the southern provinces of the zone, the permafrost evidently thawed completely, whereas in the northern provinces, especially in the Podkamennaya Tunguska-Nizhnyaya Tunguska interfluve, the permafrost thawed only to a depth of 150-220 m, below which relic permafrost was preserved. In the southern zone the contemporary permafrost was formed after the thermal maximum. The distinctive features of the region's paleogeocryological conditions pointed to the need for distinguishing two geocryological zones in all of the permafrost regions. The frontier between them, which reflects an important paleogeographic boundary line, separates areas with substantially different contemporary geocryological conditions.

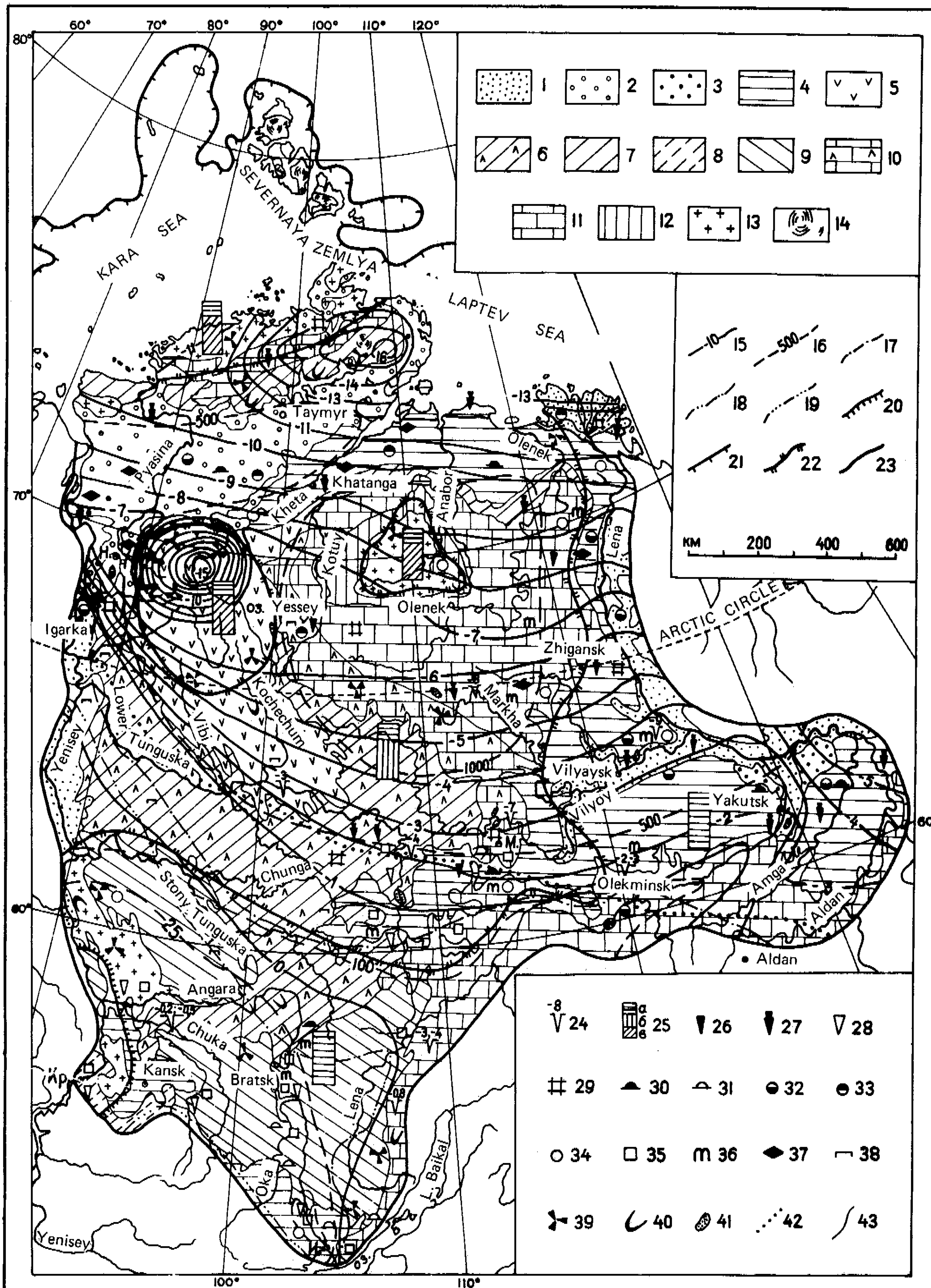
An analysis of published and unpublished findings and maps, the authors' own studies, and also a detailed evaluation of the natural setting have made it possible to identify the following zonal and regional characteristics of the permafrost in central Siberia and to show them on a schematic geocryological map (see Figure 1).

The Distribution of the Permafrost

In the northern geocryological zone the permafrost is continuous. Here, in local regions, situated for the most part in the bottoms of valleys, there are only taliks of the hydrogenous class, although they occupy an extremely small area (up to 5 percent) in comparison with that of the permafrost. In the southern geocryological zone the discontinuity of the permafrost increases sharply toward the south. The substantial increase in the area comprising the taliks is primarily a product of conditions favoring the formation of precipitation seepage taliks and those resulting from radiation factors at the boundaries between the watershed plateaus. In the northern part of the zone, the permafrost is discontinuous; in the southern part, it is sporadic. The islands of permafrost are confined exclusively to the shaded areas in the bottoms of valleys consisting of clayey-silty, and clayey, deposits.[20]

The Temperature of the Ground

The present day climatic conditions obtaining in the latitude zones have predetermined the increase


60°
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70°
60°
100°
110°
KARA SEA
SEVERNAYA ZEMLYA
LAPTEV SEA
ARCTIC CIRCLE
Taymyr
Pyasina
Khatanga
Kheta
Anabor
Olenek
Kotuy
Yessey
Igarka
Zhigansk
Lena
Lower Tunguska
Yenisey
Vibi
Kochechum
Markha
Vilyaysk
Vilyoy
Yakutsk
Chunga
Stony Tunguska
Olekminsk
Amga
Aldan
Angara
Chuka
Kansk
Bratsk
Oka
L. Baikal
Yenisey
KM 200 400 600

FIGURE 1 Schematic geocryological map of central Siberia (combiled by S. N. Fotiev, N. S. Danilova, and N. S. Sheveleva). 1--suglinoks, supesses, and sands of Quaternary age (glacial-marine, marine and glacial, alluvial); 2--suglinoks, supesses, and sands (alluvial and lacustrine) of Quaternary age; 3--sands, sandstones, and silts (sometimes with intercalations of pebbles) of Neogene age; 4--sands, sandstones, silts, clays, and argillites with intercalations of bituminous and brown coals of Jurassic and Cretaceous age; 5--volcanic formations (basalts, tuffs, tuffites, and tuffogenic sandstones) of Triassic age; 6--sandstones, silts, and argillites containing lenses and seams of coal; tuffs and tuffogenic sandstones of Carboniferous, Permian, and Triassic age, intruded by traps; 7--sandstones, silts, and argillites of Permian age; 8--marls and argillites containing intercalations and bands of limestones, dolomites, silts, and sandstones of Devonian age; 9--sandstones, silts, and argillites containing intercalations and benches of limestones, marls, and less frequently, gypsums of Silurian and Ordovician age; 10--limestones, dolomites, and marls of Silurian and Ordovician age, much intruded by traps; 11--limestones, dolomites, and marls (in depressions containing lenses and intercalations of gypsum and rock salt) of Silurian and Ordovician, Cambrian, and Sinian age; 12--sandstones, quartzitic sandstones, dolomties, limestones, marls, and calcareous clay shales of Sinian age; 13--gneisses, plagiogneisses, and granites of the Archean complex and silicified sandstones and sheared silts of Proterozoic age; 14--glaciers; 15--isolines of the mean annual ground temperatures for typical watershed conditions; 16--isolines of frozen soil thicknesses (in m); 17--southern limit of mainly continuous permafrost; 18--southern limit of discontinuous permafrost; 19--southern limit of sporadic permafrost; 20--southern limit of permafrost region; 21--northern limit of possible permafrost beneath the sea bottom; 22--limit of distribution of various types of permafrost; 23--boundaries of region; 24--temperature of earth material in the bottoms of the valleys; 25--structure of permafrost: (a) earth material, the joints and pores of which are filled with ice (stage of freezing), (b) earth material, the joints and pores of which are filled with water and brines (the stage of cooling), and (c) earth material, the joints and pores of which are filled with air, or alternatively solid earth material (the stage of cooling); 26--epigenetic multiple ice wedges; 27--syngenetic multiple ice wedges; 28--pseudomorphs after multiple ice wedges; 29--polygonal fractured microrelief; 30--perennial heaving hummocks; 31--seasonal heaving hummocks; 32--thermokarst depressions after segregation and multiple ice wedges; 33--thermokarst depressions after ice segregation--multiple ice wedges--and stratal ice; 34--thermokarst depressions after segregation ice; 35--hummocky-sinkhole relief; 36--finely hummocky microrelief; 37--stratal ice; 38--hummocky peatlands (relic); 39--placers, stony rivers, circles, etc; 40--solifluction features; 41--icings; 42--southern limit of northern geocryological zone; and 43--limit of distribution of soils of varied composition.

in ground temperatures from -14° in the north to +3°C in the south. The regional temperature deviations from the strictly zonal values can be summarized as follows: In the eastern part of the region, the geoisotherms swing southward. This is most pronounced in the central provinces of the region. Here, as a result of the substantial increase in the continentality of the climate and the decrease in the thickness of the snow cover toward the east, the -1° isotherm is deflected southward by almost 600 km. Constituting an equally importnat factor, which greatly interferes with the zonal distribution of the ground temperatures, are the altitude belt variations in the climatic conditions. Predominating over most of the region is a continental type zonation with well-defined lower and transitional belts. It is for this reason that the increase in ground temperatures between the bottoms of the valleys and watersheds normally has a gradient of 2°/100 m. It is only within the Angara Plateau, the Yenisei Ridge, and the eastern part of the Putoran Plateau, which rise above the inversion ceiling, that the temperature of the ground in the upper geocryological zone again decreases. An altitudinal zonation of the oceanic type appears only on the western slope of the Putoran Plateau, in the Byrranga Mountains and the Chekanovskii and Pronchishchev ranges. Here, the decrease in ground temperatures, with the increase in elevation, has a gradient of 0.5°-0.7°/100 m. It is owing to this that at the highest points of the Putoran Plateau the ground temperatures are 6°-8° lower than the zonal temperature, and in the Byrranga Mountains they are 2°-3° lower. The appreciable (up to 3°-4°) local deviations in ground temperatures from the zonal values shown on the map are due to the exposure of slopes (especially in the central and southern provinces), the composition of the Quaternary deposits, and the thickness of the vegetation and snow cover.

The Thickness and Constitution of the Permafrost

The decrease in the thickness of the permafrost from 1,800-1,500 m in the north to 3-5 m in the south was predetermined by latitude zone changes in the climatic conditions that occurred throughout the Quaternary. The disparate age of the permafrost in the northern and southern geocryological zones is responsible for the abrupt increase in thickness at the boundary between these zones.[4,15] In the southern zone the increase in the thickness of the permafrost to 200 m corresponds to the post-Holocene latitude zone variations in the heat-exchange conditions, whereas in the northern geocryological zone, the latitude zone increase in the thickness of the permafrost is a result of the heat-exchange conditions in the Pleistocene.

Among the regional factors that have the most noticeable effect on the variation in the zonal thickness of the permafrost it is necessary to distinguish the following: the heat flux originating at great depths; surface water and groundwater; and the altitude-belt characteristics of the heat exchange.

The magnitude of the region's geothermal gradient ranges from 1°/100 m and less within the Anabar Shield to 4°/100 m in regions of Mesozoic troughs and in local, tectonically mobile areas. The low geothermal gradient was conducive to the formation of permafrost of great thickness and did not make for any appreciable degradation of it from below during the period of climatic amelioration. The greatest degradation of the permafrost from below occurred and continues to occur in the Mesozoic troughs. The deficit of overburden pressure recorded in the Yakut and Khatanga basins serves to corroborate this. Evidently, only an appreciable (up to 500 m) degradation of the permafrost from below could explain the difference in the thickness of the permafrost at the western margin and in the central part of the Vilyui Trough, which is as much as 400 m and more.

The role of groundwater in the redistribution of plutonic heat and the influence of this factor on the abrupt intrazonal changes in the thickness of the permafrost has been demonstrated by P. I. Mel'nikov. In the interconnecting second- and third-order positive and negative structures in the Yakut Basin, the difference in the thickness of the permafrost is as much as 200-400 m. The effect of the altitude-belt conditions of heat exchange on the variation in the zonal thickness of the permafrost if most pronounced within the Putoran Plateau and the Byrranga Mountains. Here, it has been found that for every 100 m in altitude, the thickness of the permafrost increases by 40-50 m, attaining maximum values in the high-altitude zones.

As the ground became cooler, permafrost that differed in its constitution formed in the various structures, depending on the characteristics of their hydrochemical profile. Three types of permafrost are most typical of the region:

1. *Single-stage permafrost of the first type* consists of only a freezing stage made up of perennially frozen ground widespread in the eastern and southern parts of the region, where the thickness of the zone of free water exchange exceeds the depth at which there is cooling of the ground to negative temperatures.
2. *Two-stage permafrost of the second type* consists of the freezing stage of pereniallly frozen frozen ground and a water-filled stage of cooling, in which the joints and pores of the ground contain saltwater that has a negative temperature. Permaforst of this type occupies the central and northern parts of the region, within which the zone of free water exchange was found to be much shallower than the depth of cooling of the ground to a negative temperature.
3. *Two-stage permafrost of the third type* consists of the freezing stage of the perennially frozen ground, below which there is a stage of air-dried frozen ground. Highly favorable conditions for the formation of permafrost of the third type evidently existed within the Anabar and Putoran massifs.[21]

The Cryogenous Structure of the Perennially Frozen Soils Stage

These soils consist of the polygenetic and epigenetic types.

The polygenetic type predominates in the northern zone. Its upper layer, which become frozen syngenetically (early Pleistocene-Holocene), consists mainly of slope and alluvial deposits and less frequently of lake-swamp, coastal marine, and other deposits. Owing to the predominance of plateau relief in central Siberia and removal of the products of weathering, the thickness of this layer is usually not more than 3-5 m. It is only in the major valleys that it increases to 10-20 m, although in the lowland areas it can be as much as 50 m or more. Usually, the syngenetic layer is characterized by a high ice-saturation value, amounting to 20-40 percent and in some of the facies varieties of alluvium reaching 80-90 percent, compositely bedded or concavely bedded cryogenous textures and thick syngenetic regeneration ice wedges. In the contemporary syngenetic layer, a zonal decrease in ice content is traced from north to south.

Over the greater part of the area, the lower layer, which became frozen epigenetically and constitutes the bulk of the polygenetic permafrost, consists of bedrock and to a lesser extent of alluvial, glacial, eolian, and marine deposits. The cryogenous structure of the bedrock is determined by the extent of its disaggregation and infilling prior to becoming frozen. Fractured and fracture-bedded cryogenetic textures predominate, and in the residual rock--bedded, reticulate, and basal textures. The highest ice content (up to 10 percent or more) is typical of the Paleozoic soils of the northeastern part of the Siberian Platform. Although the ice content of the bedrock decreases with depth, ice inclusions are typical of the entire perennially frozen soils stage and have been traced in drill holes to depths of 300-350 m.

The epigenetic permafrost is comparatively widespread in regions of marine transgressions and glaciations. It is distinguished by a low ice content; only in the upper layers (5-12 m) is it capable of reaching 20-40 percent. A reticulate-horizontal cryogenous texture is typical of these deposits. Not infrequently, glacial marine deposits contain stratal sills of ice of 10-20 m or more in thickness.

In the southern zone, polygenetic permafrost in which there is a thin syngenetic layer occurs only along the northern margin. The bulk of the permafrost is of the epigenetic type and consists of fine-grained alluvial, slope, and lake-swamp deposits. The coarse-grained deposits and bedrock became frozen only in local areas: in the bottoms of valleys, on slopes with a northern exposure, and where there is a thick peaty-mossy covering. The ice content of the frozen soils does not normally exceed 10-15 percent. The highest ice content is confined to the upper 5-10-m-thick layer in which bedded and lens-shaped cryogenous textures predominate. In the event of water migrating from subjacent water-bearing layers, there can be a substantial increase in ice content.

The Cryogenous Formations

Contemporary and ancient cryogenetic processes are of prime importance in the modeling of the surface, in that they determine many of the features of its meso- and microtopography.

The region's exceptionally severe cryological conditions have resulted in the extensive occurrence of cryogenous features associated with weathering, frost-cracking, and hillside processes.

Cryogenous weathering gives rise to the formation of coarsely lumpy debris and various types of structural soils. They are most prevalent to the north of the Arctic Circle, where they frequently cover up to 50 percent of the area. Blanketing suglinoks, which are regarded as the end product of cryogenous weathering, cover the entire area from the far north to the south.

Frost-cracking is very widely developed. The distribution of multiple ice wedges exhibits a latitudinal zoning. In the northern zone it exists almost everywhere and is developed not only in the valleys, but also on the watersheds and slopes and in soils of varying composition, including rubbly residuum. Thick, ancient, multiple ice wedges are widely prevalent.

In the southern zone, multiple ice wedges develop only when there are favorable conditions. Small ice veins are encountered almost to the southern boundary of the region. In contrast to western Siberia, they exist not only in peatlands, but also in mineral soils. High-center fractured polygons are typical. Traces of ancient frost-cracking (pseudomorphs after multiple ice wedges, etc.) are widespread in the southern zone and attest to the activity of this process in the Pleistocene.

Solifluction formations are one of the predominant types of cryogenous formations in the tundra zone. In the taiga zone, dells are widely prevalent.

Heaved hummocky formations have a local distribution. In the northern zone there are some small perennial heaving hummocks. Pingos form in the larger lake basins, becoming overgrown with vegetation. In the southern zone, hummock formation plays a relatively minor role in comparison with the other cryogenous processes. There are both perennial and seasonal-heaving hummocks. Some hummocky peatlands are present in the western half of the zone; they have been studied in detail in western Siberia.

The moderate thickness of the ice-saturated Quaternary deposits, which is typical of much of the region, serves to limit the development of thermokarst processes. For this reason, thermokarst formations of limited dimensions and depth (to 2 m) predominate. Large thermokarst basins are developed for the most part within the North Siberian, central Yakut lowlands and in some of the major river valleys, where there are thick deposits containing ice and multiple ice wedges. Relic thermokarst features (sinkhole-hummocky topography and lake basins) are particularly widespread in the southern zone.

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THE THICKNESS OF THE PERMAFROST OF THE WEST SIBERIAN PLAIN

V. V. BAULIN AND A. L. CHEKHOVSKII Operations and Research Institute for Engineering Site Investigations (PNIIIS)

The permafrost of western Siberia formed under the influence of two groups of factors: exogenetic (determining the upper boundary conditions of freezing of the soil materials) and endogenetic (determining the lower boundary conditions of freezing). The exogenetic factors are divided into zonal (long-term fluctuations of the climate) and regional (geological structure, transgressions, alluvial flows, and the like). The first of these subdivisions mold the zonal features of the permafrost and constitute the background against which the regional factors operate. The emergence of the regional factors is specific in each frozen zone and can lead to a decrease in the thickness of the permafrost or to complete thawing of the soil.

EXOGENETIC FACTORS

The main zonal features of the structure and distribution of the contemporary thick permafrost of western Siberia developed during the second half of the upper Pleistocene and Holocene, i.e., subsequent to the boreal transgressions of the Polar Basin. The very cold conditions that prevailed during Zyrianian times gave rise to intensive freezing of the soils far beyond the limits of their contemporary distribution. Attesting to this are traces of perennial freezing in the Syrianian sediments. The very pronounced warming of the climate in the Holocene led to partial thawing of the soils to latitude 68°-69°. Later, the soils again became frozen, although to the south of latitude 66°-67° the frozen layers did not interlock with one another, while to the south of latitude 61°-62° permafrost did not form at all.

As a result of these climatic fluctuations, in the area making up the west Siberian plain, three zones are distinguished in which the structure of the permafrost is dissimilar (Figure 1). In the northern zone its structure remains continuous as the depth increases. In the central zone, a two-layered structure of the permafrost is seen to exist in parallel with its continuous structure. The thickness of the upper layer, which formed after the climatic optimum, decreases from 40-80 m to 10-20 m. The lower layer of frozen ground (relic permafrost) has been preserved since the pre-Holocene periods of cold. Southward, the depth to permafrost gradually increases from 50-100 m to 100-150 m. The thickness of the relic frozen layer ranges from 50-100 m to 200-300 m. In the southern zone of freezing, only the relic frozen layer has been preserved. Its table lies at a depth of up to 200 m or more, its base at a depth of 300-400 m.

Regional exogenetic factors determine the intrazonal variations in the conditions of perennial freezing and thawing of the soils. Under the beds of the major rivers (the Ob', Yenisei and others), the soils completely thaw out (or do not develop). Permafrost can occur only in those parts of the bed that are situated near a bank being undercut by the water. Consequently, freezing of the soils in the terraces of such rivers and lakes begins after the area has been fully drained (when the climatic conditions are favorable). The thickness of the permafrost in the major river valleys and lakes will therefore increase with movement away from the younger terraces towards the older ones. This is why there is no relic permafrost in the floodplains of the rivers that formed in the Holocene.

Thermokarst lakes, the duration of whose development is measured in tens and hundreds of

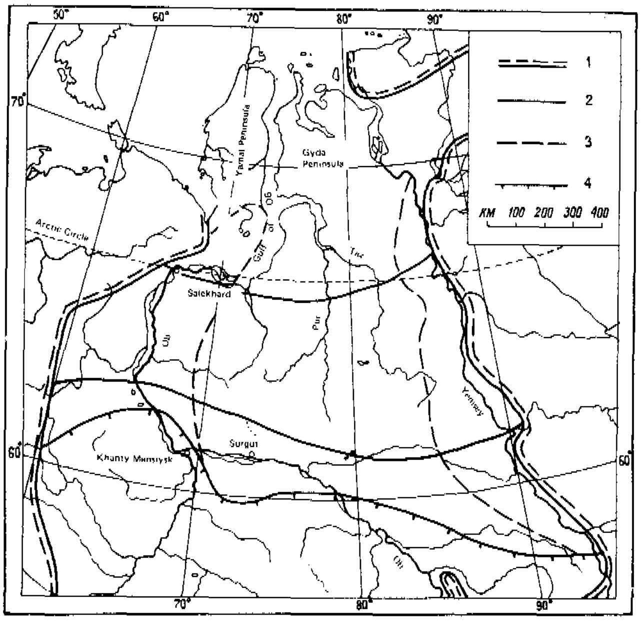

FIGURE 1 Diagram of the zoning of the west Siberian plain. 1--boundary of the west Siberian plain; 2--boundaries of the permafrost zone; 3--boundaries of the permafrost provinces; 4--southern limit of relic permafrost.

years, also interfere with the thermal regime of the frozen soils, although it is not everywhere (especially in the north) that they lead to total thawing of the latter. Most often, closed taliks form beneath thermokarst lakes or the temperature of the soils rises and conditions are established that favor their thawing from below as a result of the heat flux from the Earth's interior.

Relic permafrost is most often preserved in local areas where argillaceous deposits have developed that interfere with the seepage of warm atmospheric precipitation.

ENDOGENETIC FACTORS

In western Siberia the relationship between the distribution of the permafrost and the tectonic structure of the plain is clearly manifested. An increase in the depth of the basement (metamorphosed rocks of Paleozoic age) leads to an increase in the thickness of the permafrost.[5] In some cases, this relationship can be quantitatively expressed.

The decrease in the thickness of the permafrost is seen to be most pronounced beneath third-order structures. The difference between the depth of occurrence of the base of the permafrost on the limbs and in the arch portion of an uplift is as much as 150-200 m.[1,6] The cause of this phenomenon was seen to lie in the increase in the heat flux from the Earth's interior, which is governed by the thermal anisotropy of the rocks[3] and by processes of convective heat transfer.[4]

In the event, however, that gas, constituting a thermal screen, exists in the arch portion of a fold, the depth of the base of the permafrost may be even greater. Above the gas, the process leading to cooling of the rocks can be intensified on account of adiabatic expansion of the gas.[2]

The size of the heat flux depends essentially on the age of the basement (on the period of its consolidation)[7] and increases by roughly 1.5 times with movement away from the regions of ancient folding towards those of young folding. As compared with the western regions, the overall increase in the thickness of the permafrost in the part of the plain lying adjacently to the Yenisei fits the more ancient basement. Most clearly manifested is the relationship between the depths of the base of the permafrost and the geothermal gradient, which, under the conditions that obtained on the west Siberian plain is proportional to the heat flux. In the west-east direction the steepness of the geothermal gradient decreases from 5°-6° to 1.5°-2.0°, and the thickness of the permafrost increases. These variations tions in the steepness of the gradient show up best of all in the southern part of the northern permafrost zone and also in the central and southern zones. The deep southward penetration of the layer of relic permafrost in the central and eastern regions of the plain coincides with the regions in which there is a decrease in the gradient. Taking the endogenetic factors into consideration has made it possible to draw anew, and in a more valid way, the southern limit of relic permafrost distribution.

On the basis of the lower boundary conditions of freezing of the rocks, the area making up the plain is divided into three provinces: Ob', central, and Yenisei. The zonation that is apparent in the depths of occurrence of the base of the permafrost must be considered separately for each province.

Overall, the Ob' province is characterized by very steep gradients of 3°-4°C and by a permafrost thickness ranging from 300-400 m in the north to 200 m in the south. Against this background occasional zones in which the gradient increases to 5° and even to 6° are to be seen. The permafrost with the least thickness is undoubtedly confined to these zones.

The central province covers the greater part of the plain. It is characterized by a smooth change in the gradient from 3°-4°. The relatively small heat flux ensures the development of permafrost of up to 400 m in thickness in the central zone of freezing and of up to 300 m in the southern zone.

In the Yenisei province the gradient decreases to 2° and even to 1.5°, which leads to an abrupt increase in the depth of the permafrost base. At the 62d parallel in the Yenisei province, permafrost has been traced to a depth of approximately 500 m, and at the 60th parallel to 400 m.

The intersection of the permafrost zones and provinces yields nine permafrost regions that have specific features as regards the thickness of the permafrost and its sectional structure.

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FEATURES OF THE CONTEMPORARY DEVELOPMENT OF PERMAFROST IN WESTERN SIBERIA

E. B. BELOPUKHOVA Operations and Research Institute for Engineering Site Investigations (PNIIIS)

The development of permafrost is governed primarily by climatic variations and the associated evolution of the vegetation. The effect of climatic fluctuations on the permafrost of western Siberia has been the subject of studies by A. I. Popov,[5] V. V. Baulin,[1] and other investigators. A. P. Tyrtikov[6] has devoted considerable attention to the study of the relationship between the permafrost and the dynamics of the vegetation. In this paper, an attempt is made to analyze the combined effect of recent climatic changes and the vegetative, primarily that which covers the soil, on the permafrost conditions.

Since the beginning of the second half of this century, a tendency has been noted in western Siberia, as also in the other regions of the Northern Hemisphere, for the air temperature to decrease. For example, during the period 1930-1950, not once did the average air temperature in the town of Salekhard fall below -5.9°C, whereas in subsequent years it was constantly below -6.2°C, and during the period 1959-1971 it ranged from -7.0°C to -7.3°C. The cooling trend has been accompanied by an increase in the amount of precipitation, both during the warm season and cold season. The maximum quantity of summer precipitation for Salekhard weather station was observed during the period 1948-1959.

The increase in the amount of precipitation and the decrease in the air temperature, especially during the summer, led to a rise in the surface moisture content and to an intensification of swamp formation processes, which were accompanied by increased accumulation of moss and peat.[2,4] As will be seen from published sources and from our observations, the rate of accumulation of sphagnum moss in the taiga zone of western Siberia is as much as 1 cm or more annually.

Undoubtedly, both the cooling of the climate and the activation of the swamp formation were reflected in the development of permafrost. Rough estimates based on the method developed by V. A. Kudryavtsev[3] indicate that, as a result of the cooling of the climate, notwithstanding an observed slight increase in the snow cover, at latitude 65° the surface temperature had dropped by approximately 1°C. A no less appreciable effect on the permafrost conditions was produced by the rapidly increasing moss cover. According to estimates based on G. M. Fel'dman's method,[7] a 10-cm-thick layer of moss and peat with a moisture content of 200-500 percent decreases the mean annual temperature of the soils by 0.5°-0.6°C.

When there is intensive swamp formation, the temperature of the soils over a 10-yr period may be further decreased by approximately 0.5°C solely on account of the surface becoming overgrown by moss.

Serving to corroborate these calculations is a large volume of information accumulated in recent years, which affords evidence of the regeneration of permafrost over vast areas of the west Siberian plain. Newly formed layers of permafrost, measuring up to 2-3 m in thickness, have been discovered at all of the geomorphological levels, beginning at latitude 67° and continuing to latitude 61°, i.e., not only within the region shown on the cryological maps to contain permafrost, the southern limit of which extends roughly along the 62d parallel, but even to the south of this.* In the northern part of this area, where taliks have a limited distribution, the regeneration of permafrost is occurring on a limited scale. At latitude 64°-65°, perennial freezing of the deposits was recorded over an area occupying up to 25 percent of the surface. Regeneration of the permafrost is observed under various landscape conditions, but it is most typical of swamp-ridden forests containing thin stands of trees. In western Siberia large areas are occupied by sparse forests that have a coarsely hummocky microtopography. The hummocks are moss-covered. The thickness of the mossy cushions of these hummocks (together with the dead part) can be as much as 0.2-0.5 m. Everywhere, beneath the mossy hummocks there are nuclei of permafrost measuring up to 1.5 m in thickness. Apparently, intensive perennial freezing of the deposits is being noted for the most part in swamp-ridden areas, where it occurs in accordance with moss overgrowth.

Furthermore, in some regions of western Siberia it has proved possible to notice a decrease in the mean annual temperature of the ground. Thus, based on numerous temperature measurements conducted in drillholes in 1948 in the vicinity of Cape Kamennyi, a mean annual temperature of about -6°C was established for the permafrost of the treeless expanses. In not a single drillhole of this region was a lower temperature discovered. On the other hand, in 1969 an expedition of PNIIIS working in approximately the same region succeeded in recording ground temperatures of below -7°C in eight cases out of twelve, while a temperature of somewhat less than -6°C was recorded in 1966 in drillholes in the vicinity of Novyi Port, i.e., 75 km to the south of Cape Kamennyi.

Also attesting to a decrease in the temperature of the ground is a comparison of temperatures recorded in the lower reaches of the Ob' during the periods 1954-1956 and 1969-1971.

The cooling of the climate and the increased accumulation of moss also contributes to a reduction in the thickness of the seasonally thawed layer.

In most cases the transition of the lower part of the seasonally thawed layer into the perennially frozen state is accomplished as a result of freezing from below. With this mode of freezing, ice-rich deposits form. These ice-containing deposits occur very close to the surface and have become widespread in the northern part of western Siberia. Accordingly, in spite of their moderate thickness, special attention must be devoted to them when studying the cryogenous makeup of surface deposits.

Evidently, the progressive development of permafrost under present conditions is due to the unidirectional effect of the decrease in air temperature and the increased accumulation of moss on the heat exchange between the soil and the atmosphere. Nevertheless, the effect of changes in the air temperature and the development of vegetation situated on the permafrost changes its direction with the passage of time. Along with the climatic fluctuations, variations in the rates of accumulation of moss and peat constantly occur. Furthermore, peat is also capable of accumulating against a background of a rise in the air temperature. When studying the dynamics of the permafrost, an analysis must therefore be made of the combined effect of the climate and vegetative coverings.

*Evidently a need has arisen for a southward revision of the present versions of the southern limit of the permafrost in western Siberia.

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AN ANALYSIS OF LARGE-SCALE MAPS OF RELIC CRYOGENOUS MORPHOSCULPTURE (AS EXEMPLIFIED BY THE BASIN OF THE UPPER VOLGA)

V. V. BERDNIKOV Institute of Geography, USSR, Academy of Sciences

The study of the main types of relic permafrost microrelief and its large-scale mapping was first undertaken by us within the confines of a relatively small region (the basin of the upper Volga). These investigations were a further development of the basic tenets formulated when relic cryogenous morphosculpture was distinguished in the European part of the USSR.[5,6]

The basin of the upper Volga is situated in the central part of the northern zone of distribution of relic cryogenous morphosculpture. As a result, the dynamic limit of the ancient zone of permafrost extended almost 10° further south, and, in the case of the area being studied, the stage was set for a highly prolonged period of permafrost and for relatively stable conditions.

The studies comprised two lines of investigation. The first--a study of the geological structure of the forms of relic cryogenous morphosculpture and their geomorphological degree of expression, as well as a morphogenetic analysis of the principal features of these formations. The second--the large-scale mapping of relic topographic forms based on their lithological and geomorphological degree of expression at the contemporary surface. A large-scale composite hydrogeological survey of the area served as the baseline in the mapping of these forms. The mapping of the permafrost relics revealed that the area is replete with residual permafrost microrelief, which may occur in three states: it can be buried in such a way that it does not show up in a topographic survey to the scale of 1:1000-1:100; the second state pertains to microrelief of those areas, in which it is partially buried and only isolated elements of it show up in a topographic survey; the third state pertains to microrelief of those areas in which it is clearly expressed at the contemporary surface and where all of the characteristic features and elements show up in a topographic survey.

The results of the mapping were plotted against the legend of a map showing the distribution of the relic permafrost microrelief[6] and incorporating 10 principal types of microrelief distinguished through the interpreting of photographs and a complex of field studies. These types were subdivided by us into five groups, at least three of which (polygonal, block, and hummocky and hummocky-sinkhole microrelief) had polygonal systems as their foundation. The widespread occurrence of relic microrelief forms belonging to these groups is indicative of the marked development during the permafrost epoch of polygonally fractured forms containing wedge ice and rocks with a high ice content. The information obtained is applicable to various lithological provinces and affords a basis for making comparisons between closely similar polygonal forms.

Central to the microrelief of the three groups are polygonal systems that in profile are seen to consist of wedge-shaped structures of variable shape and degree of expression. We studied a series of wedge-shaped structures encountered in morainic suglinoks, loesslike suglinoks, and supesses. An analysis of these formations based on their morphology, the structural character of the wedge-shaped forms, and the ratio between the enclosing and infilling material must be approached with caution and not worked out conclusively. In particular, not all of the investigators believe that the criteria that have been proposed for classifying wedge-shaped forms as permafrost relics or for determining the category of the formations are applicable in practice. The existing criteria, which are considered to be definitive, are not always able to indicate the true origin of polygonal forms as primevally glacial or land formations. There are at present no diagenetic or syngenetic criteria that are considered to be single-valued when it comes to determining the origin of enclosing rocks and wedge-shaped forms.[8,9]

In our view, which is based on previous investigations and data derived during field studies, a description and analysis of relic permafrost structures must include the following points: the geomorphological position of the profile and its orientation according to the cardinal points of the compass, the composition and texture of the material constituting the enclosing and infilling deposits, the shape (cuneiformity) and detailed description of the morphology of the structures, the elements of two-storying, the asymmetry of the boundaries of the structures and character of their infilling, and the relationship between this asymmetry and the orientation of the polygonal wedges.

In a stratigraphic plan the emergence of a system of wedges is one of the conditions that are necessary for the distinguishing of a stratigraphic discontinuity.

We shall not dwell in detail on each of the foregoing points, but will consider only those of them that have been further clarified or otherwise dealt with in the author's papers. According to our information, in the area comprising the European part of the USSR, the dimensions and configuration of wedge-shaped forms can vary within wide limits: from 1.0-1.5 m to 4-5 m in height and from 0.3-0.5 m to 4 m in width, the observed sizes of the polygons ranging from 2-3 m to 25-30 m. They are grouped into at least three systems of fractured formations with correspondingly different parameters. One of the determining factors with respect to these differences is the lithology of the deposits and their ice content and conditions of existence in the relic state.

In recent years a two-storying arrangement of wedge-shaped structures has been regarded as an index of the depth of seasonal thawing in the concluding phase of existence of the permafrost, and it is altogether possible that it is further governed by the lithology of the enclosing earth materials, having regard to the factor of the asymmetry of the formations.

Asymmetry of the boundaries of wedge-shaped forms is characteristic of polygonal wedges that are clsoe to latitudinal in their orientation. Meridionally oriented wedges have a shape that is close to symmetrical. This is due to the variable effect of the solar radiation during the period of the melting of the ice and permafrost degradation. We made a detailed investigation of this situation, as exemplified by wedge-shaped structures discovered in an area of polygonal microrelief and located in close-grained morainic suglinoks (Figure 1).[4] It was found that the asymmetry characteristic is not only one of the points of the "study," but even makes it possible to categorize the relic wedge-shaped forms as structures that have passed through the ice-wedges stage of their development.

In most cases the material and character of infilling of wedge-shaped structures is regarded as a manifestation of random factors. It seems to us, however, that the character of the infilling is also subject to definite laws, which become apparent when an overall analysis is made of the material comprising the infilling that takes into account the asymmetry characteristic, the stability of the ground (lithology), and the factor of the gravitational instability of the masses of the enclosing and infilling materials. Ascertaining the character of the infilling of wedge-shaped structures is thus an important corrective in the diagnosing of formations.

The sum total of the above-described characteristics of relic wedge-shaped forms is one of the components in the conducting of a paleocryogenous analysis of large-scale maps of relic cryogenous morphosculpture. The essence of such an analysis consists of (1) ascertaining the basic parameters of the ancient permafrost that existed in the area in question, the distribution patterns of the ancient permafrost relief forms, and the relationship between these characteristics and the geological-geomorphological and facies conditions of the past; and (2) determining the leading factors and conditions that were manifested during the period of permafrost degradation and transition of the systems into the relic state.

A paleocryogenous analysis of large-scale maps also includes the division of the area into regions, as was done in the case of the upper Volga basin. Forming the basis of this operation is the distinguishing of combinations. We define these as being the sum total of the types of relic microrelief occurring within the confines of a major geomorphological unit and on lithologically similar soils. Four paleocryogenous regions are distinguished on the appended map, each of which is characterized by a specific "collection" of forms of relic cryogenous morphosculpture (Figure 2).

The greater part of the right bank of the area in question is occupied by a region of morainic,

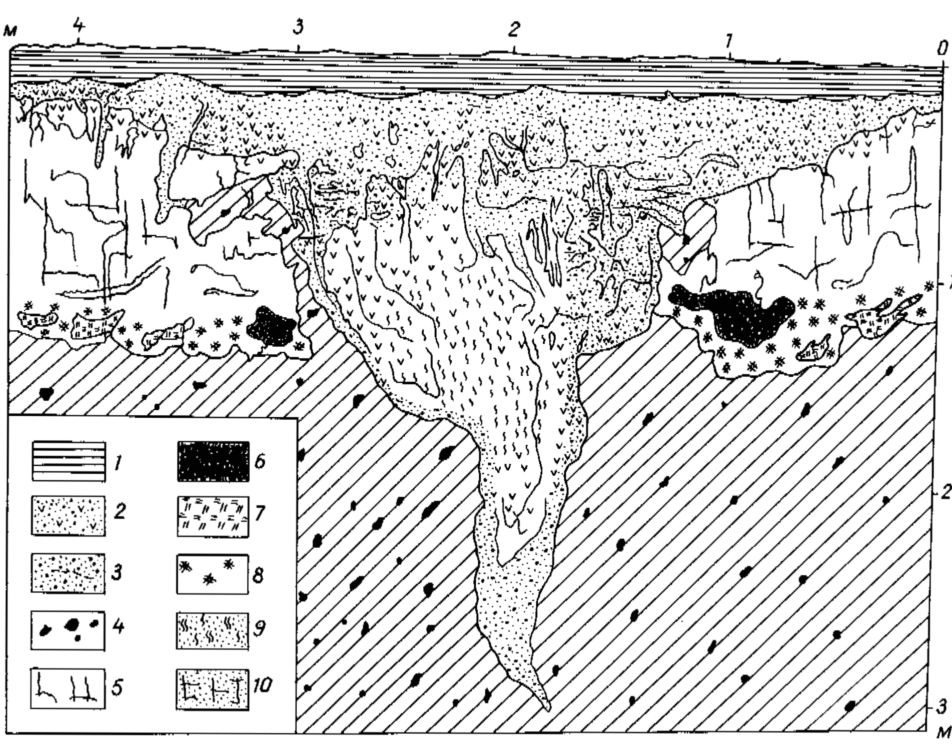

FIGURE 1 Sketch of a wedge-shaped structure in a trench near Kir'yanovo village. 1--layer of contemporary topsoil; 2--medium- and fine-grained sand with patches of iron accumulation; 3--boundary of the podsol layer; 4--red-brown suglinok admixed with morainic gravel; 5--light brownish-chestnut suglinoks; 6--dark brownish suglinok in which humus has formed; 7--gray, dove-colored, gleyed suglinok; 8--brown and orange-colored suglinok with granular structure; 9--light brown suglinok admixed with sand; 10--interstratifications consisting of red and brown iron accumulations.

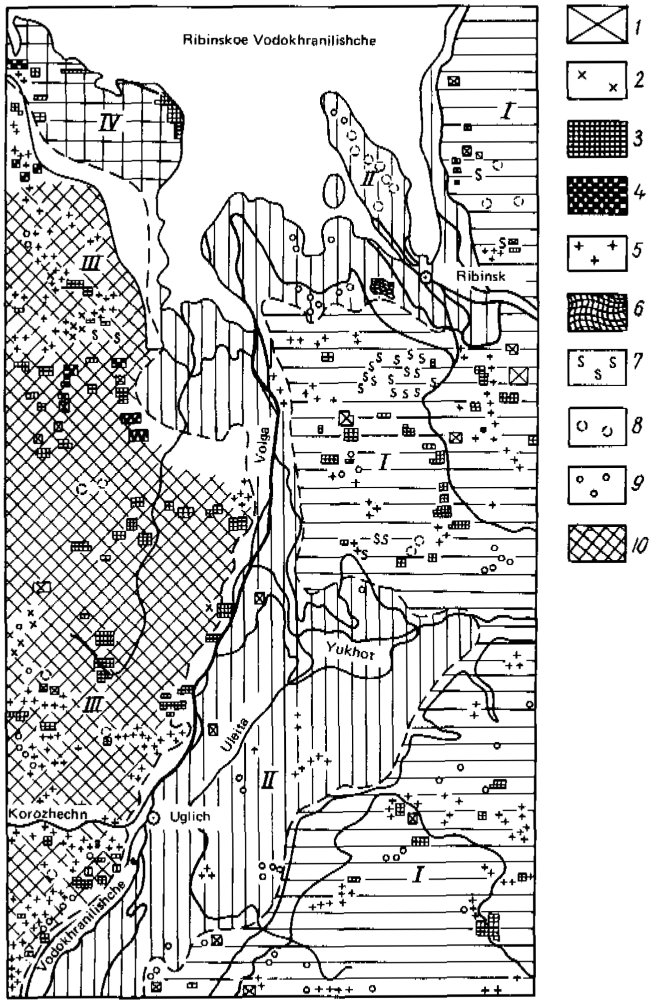

FIGURE 2 Map showing distribution of relic cryogenous microrelief in the basin of the upper Volga.

Forms of relic microrelief: 1--fine- and medium-polygonal microrelief with a dark mesh; 2--fine- and medium-polygonal microrelief with a light mesh; 3--polygonal-block microrelief; 4--hummocky-sinkhole and block-sinkhole microrelief; 5--massive forms of block and hummocky microrelief; 6--vestigial forms; 7--relief consisting of "microunevennesses" (translucent spottiness); 8--circular structures; 9--sinkhole formations; 10--large-block type of relief.

Paleocryogenous regions: I--region of morainic, fluvioglacial, and lacustrine glacial plains with widespread polygonal, polygonal-block, and massive forms; II--region consisting of a lacustrine-glacial plain with sparse occurrences of massive forms and circular structures; III--region consisting of a lacustrine-glacial plain in which polygonal, polygonal-block, and massive forms combined with large-block relief are prevalent throughout; IV--region of ancient lacustrine terraces characterized by a hummocky-sinkhole type of microrelief combined with block and hummocky forms.

----- Boundaries of the regions.
——— Boundaries of the geomorphological levels.

fluvioglacial, and lacustrine-glacial plains containing widespread polygonal, polygonal-block, and massive forms. Situated to the west of Region I on the right bank of the Volga is Region II, which falls away abruptly on a level with the low-lying terraces of the Volga. This region is a lacustrine-glacial plain in which there are sparsely occurring massive forms and circular structures.

Region III occupies the main area of the left bank of the Volga and is a lacustrine-glacial plain in which polygonal, polygonal-block, and massive forms in combination with large-block relief are prevalent throughout. Region IV occupies small areas in the northern part of the left bank of the Volga on a level with the ancient lacustrine-glacial terraces in which a specific hummocky-sinkhole type of microrelief is prevalent.

Ancient polygonal cracking was a leading morphosculptural process, and the presence of relic forms of polygonal groups on a level with such geomorphological features as high terraces and plateaus undoubtedly attests to the severity of the freezing conditions during the period when the ancient permafrost existed and to the high ice content of the frozen ground.

In the upper Volga basin the depths of seasonal thawing during the period when permafrost existed ranged from 1.4-1.8 m; most of the wedge-shaped structures studied have the outlines of epigenetic formations. The widespread occurrence of fracture-polygon types of microrelief is also indicative of the drastic changes that occurred in the amplitudes of the annual temperatures, there being very low absolute values and a weak development of the snow cover. In the course of correlating the map showing the distribution of the relic permafrost microrelief with a geological map of the Quaternary deposits of the area, it was found that the lithology of the ground is a factor that governs the origination, degree of expression, and preservation of the relic permafrost relief forms to a greater extent than does the geomorphological localization of small areas. Thus, polygonal systems are not encountered in sands, the large-block type of relief is prevalent only in a region where there are supramorainic suglinoks, and wedge-shaped structures are much more clearly defined and better preserved in dense morainic suglinoks than in supesses.[2,3]

It is important to note that this was the first time that in the mapping of permafrost relics the same methods were employed as are used in the mapping of relief forms in the contemporary permafrost zone.[1,7] Also, this was the first time that large-scale mapping of relic microrelief was accomplished for the purpose of solving specific problems of geological mapping. Individual relief forms and systems that had previously been classed as manifestations of glacial activity were identified as cryological phenomena.

A paleocryogenous analysis of large-scale maps of the relic microrelief can be augmented and extended through the obtaining of new characteristics and parameters. In the case of the area making up the upper Volga basin, the results of the analysis are fully in agreement with other data, in particular spore-pollen data, which characterized the area during the concluding period in which permafrost existed. Thus, the information adduced attests to the existence in the past of a vast zone of permafrost that had all the typical manifestations and processes.

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ON SOILS SLOWLY SHIFTING UNDER THE INFLUENCE OF CRYOGENOUS REDISTRIBUTION OF THE MATERIAL IN THE SURFACE LAYER

A. M. GORELIK AND M. K. DRUZHININ All-Union Research Institute of Transport Construction

In the mountainous and foothills districts of the region where permafrost and harsh climatic conditions exist, large expanses of hillside ranging from 2° to 30° in slope are covered by soils that have been profoundly influenced by low temperatures. In the course of cryogenous redistribution of the material in the overlying mantle, these soils have shifted slowly downhill under the constant influence of gravitation. The main factors giving rise to these displacements are alterations in the volume of the particles forming the deposits; the freezing and thawing of moisture both within and at the foot of the deposits at the time of abrupt variations in the temperature; and the accompanying uplift (along the perpendicular surface of the hillside), subsidence (along the vertical), and fragmentation and aggradation of the particles. As was established by our observations in the trans-Baikal and Kuznetskii Alatau, also of great importance in the steady downhill displacement of coarsely fragmented material is the washing out of fine particles, extending all the way to gravel and rubble, by melt water, rainwater, interstitial water, and condensation water.

A surface layer of this kind has been given a special name--"deserptium" (B. V. Ryzhov, 1966; G. S. Zolotarev, 1969). The term "deserptium" thus refers to variously composed overlying hillside deposits, the entire mass of which moves slowly downhill under the influence of cryogenous redistribution of the material in the surface layer, and of gravitation.

In the permafrost region deserptium is a zonal formation, as are solifluction [areas] or mari. It covers extensive hillside areas, since it caps most of the soils in this region. It forms on soils of any origin (Figure 1).

Usually, the thickness of deserptium is 3-4 m. This value is greater, however, in lumpy deserptium (rock trains and rock flows) forming in residuum (bald placers) or in the colluvium of landslides and screes (No. 4 in Figure 2), which are readily pervious to such heat carriers as air and water.

At the present time the entire circumpolar zone of permafrost (in the USSR, Canada, Alaska, and other countries) is being progressively developed. Construction activity (such as road building, mining and industrial construction, installation of hydroelectric power-generating and transmission systems, and civil construction) is expanding there. Since deserptium occupies large areas in the mountainous districts of this zone, and as its thickness is considerable, in many cases it will serve as a material source, a fill for buildings or installations, material for earthworks, subbasement fill, and so on.

It was only a few years ago that deserptium was first conceived of as being a zonal type of

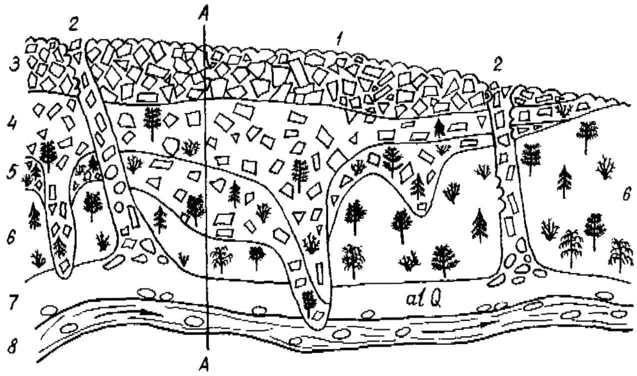

FIGURE 1 View of hillside on which deserptium is present. 1--rock-strewn brow of hillside; 2--rocks and boulders in channels on the hillside (stone streams); 3--rocks from landslides and screes (colluvium); 4--treeless rock trains in various parts of the hillside; 5--tree-covered rock trains in the central part of the hillside; 6--close-grained mantle of deserptium with inclusions of coarsely fragmented material; 7--terrace of river valley; 8--river; 9--permafrost table; 10--rock-strewn bed of surface deposits.

deposit. In documents held in archives that pertain to the surveying, planning, construction, and utilization of different types of buildings and installations designed for a variety of purposes, these deposits are not even mentioned. At the All-Union Research Institute of Transport Construction (TSNIIS), a study was made of information pertaining to hillside mantle deposits that had undergone cryogenous reworking. The aim of the study was to augment information on the composition, texture, consistency, and properties of deserptium, and to supplement experience gained in building on it.

Furthermore, specialists at TSNIIS determined the engineering-geological properties and conducted observations of the shifting of deserptium deposits. Such a study of deserptium is necessary in order to establish its distinctive properties when used as a foundation for buildings and structures under a variety of conditions.

The composition of deserptium is such that coarsely fragmented material is almost always present. The content, coarseness, and distribution of the fragments in the overlying mantle may therefore be highly variable, depending on the

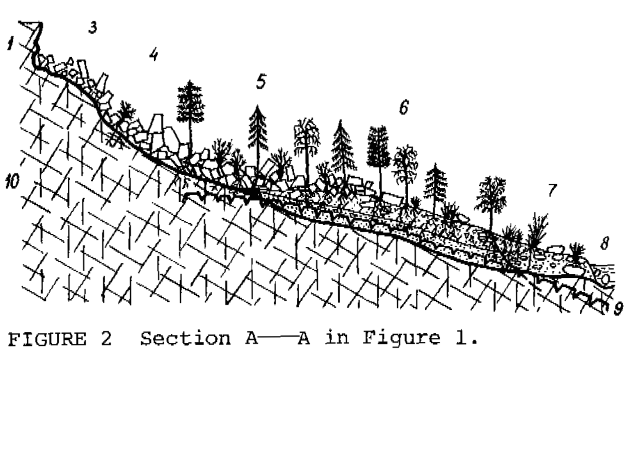

FIGURE 2 Section A—A in Figure 1.

sector and the steepness of the slope, the characteristics of the substrata, and the permafrost and hydrogeological conditions (see Figure 2). To a large extent they are inherited from the parent rock (residuum, colluvium, talus, solifluction, etc.).

The upper parts of the hillsides are usually overlain by colluvium (No. 3 in Figures 1, 2), but with lessening of the rockslide and scree processes the latter is superseded by lumpy deserptium (rock trains). In the central parts of the slopes, the lumps of the rock trains are often loosely packed (the porosity can be as much as 50 percent) (see No. 4 in Figure 2), since the fine particles, gravel, and rubble are washed down onto the lower-lying sectors of the hillsides. Conversely, the voids between the lumps of the rock trains situated in the lower parts of the hillside are more often filled with rubble, shingle, gravel, sand, supesses, and suglinoks, which are also present in the lower part of the mantle (see No. 5 in Figure 2). Fine-grained deserptium (see No. 6 in Figure 2), bordered by alluvium (see No. 7 in Figure 2) or other sediments predominates lower down the slope.

By reason of the variegation in composition and the presence of coarsely fragmented inclusions. it is difficult to estimate the density and shear strength of deserptium. The selecting of representative samples in order to determine the unit weight and conduct laboratory tests for shear strength is also a complicated procedure. Prisms and cylinders of deserptium cut from borehole samples for undisturbed testing for shearing, crushing, fracture, and bulging can only be provisionally regarded as representative of a given sector of the hillside and layer of the surface. All this makes it essential to explore beneath each of the corners and each section of the supporting walls and structures within the area to be occupied by the buildings and installations being planned. The aim, therefore, must be to assemble a large volume of data with a view to statistical processing or the use of approximate values of the density and strength of deserptium when buildings are founded on it.

On the whole, the strength of deserptium consisting of coarse fragments, rubble, gravel, and sand is considerable. This is due to the presence in such deserptium of poorly rounded, coarse particles. Thus, friction is set up between granitic lumps in contact with one another and with a rock-strewn bed; likewise, the internal friction of rubble, shingle, gravel, and sand is substantial, and the corresponding [friction] angles are 35°-40° and sometimes even greater in the case of deserptium occurring in rocky residuum, the friction is greater on account of mineral crystals (such as quartz and feldspar) that have an irregular shape, acute angles, and ridges, becoming trapped between the fragments. If the deserptial and cohesive soils are not swollen, which is often found to be the case in regions with sever winters but lying outside the permafrost zone (such as the Urals), these also have an increased strength due to the reinforcement of the deserptium by coarse inclusions.

In the case of hillsides with a steepness of between 2° and 30°, a deserptium mantle shifts slowly downhill at a rate of millimeters to centimeters annually (although sometimes somewhat faster). Colluvium normally overlies the steeper slopes. An annual shift rate of up to 1 cm is not of any consequence in the case of roads that have beds that can be widened in the uphill direction to provide for periodic changes in the positioning of their center lines throughout the total estimated lifetime of the route. But even such a modest rate of displacement is dangerous and gives rise to distortions that cannot be tolerated when they affect the long-term operation of pipelines, power transmission towers, and the foundations of many types of buildings and installations.

On slopes that are steeper than 20°, irrespective of the rate of shifting of coarsely fragmented deserptium (rock trains), and on gentler slopes at shift rates greater than 1 cm annually, in order that the subgrade of railway lines can be maintained in the planned position, a need arises for a system of retaining walls, sunk into the rocky base of the surface layer. Generally speaking, the supports of artificial structures (such as piers, bridges, and culverts) must be securely fastened to it.

The sinking of foundations into the foot of the deserptium gives rise to additional expense, especially as it often overlies rock or is underlain by permafrost. By reason of its tendency to shift downhill, a layer of deserptium, particularly of the lumpy type, exerts pressure on a foundation cut through it. While there are currently no reliable data on the extent of this presure under actual conditions, an approximate estimate of it is possible.

In excavations, trenches, coal mines, the walls and faces of quarries, and sections and reserves sunk in deserptium, we observed a thinning out of suglinoks and supesses and, during the cold season, icing, formed as a result of the influx of water from the intercalations and lenses of sand and the coarsely fragmented material.

The rock fragments, rubble, shingle, gravel, and sand in deserptium are satisfactory materials for the construction of earthworks. Nevertheless, we managed to observe the formation of water pockets and lenses of waterlogged soil in inclusions of sand and coarsely fragmented material situated within embankments built from cohesive deserptium soils and along the periphery of these inclusions. Accordingly, for building embankments, dams, and weirs, and as subbasement fill for the foundations of structures, the use of overlying deserptial supesses and suglinoks, enriched with an admixture of coarsely fragmented material, gravel, and sand, is undesirable.

It is almost impossible to work rock trains and fine soils containing inclusions of rocks and boulders by excavation machinery unless the fragments within them have first been crushed by small-scale explosions. In many cases, this increases the cost of working deserptial soils.

The development of construction in regions of permafrost and a severe alpine type of climate is contributing to the accumulation of information on the properties of deserptial deposits. In its turn, the study of the characteristics of deserptium is establishing a basis for the working out of rational methods of surveying, planning, constructing, and utilizing installations in the regions in which it occurs.

DISTRIBUTION AND CONDITIONS OF FORMATION OF THE LOESS DEPOSITS OF CENTRAL SIBERIA

N. S. DANILOVA Operations and Research Institute for Engineering Site Investigations (PNIIIS)

At the present time the view is being expressed with increasing frequency that prolonged, systematic seasonal freezing and thawing is having a pronounced effect on the earth materials and is resulting in the formation of certain loessial characteristics in loose deposits of diverse origin.[5,7,10,12] According to this hypothesis, central Siberia, which has a markedly continental climate and rigorous geocryological conditions, is a region where favorable conditions exist for the cryogenous weathering and loessification of surface deposits. Moreover, the harsh climate and continental regime of sediment accumulation throughout the greater part of the Quaternary period presuppose the existence of more favorable conditions for the loessification of deposits in this region than in western Siberia and the European part of the USSR, where loess deposits are fairly widespread. The fact is that published and unpublished literature contains a great deal of information on the exitence of deposits with loessial characteristics in various regions of central Siberia.

This paper discusses the distribution of such deposits, certain of their regional peculiarities, and the conditions of their formation. Of the deposits that have loessial characteristics, blanket suglinoks have come to hold pride of place. Those of central Siberia have many features in common with the western varieties. These are the siltiness, porosity, pale yellow color, carbonate content, capacity for forming vertical walls, and so on. In many cases they occur in pseudomorphs substituting for regeneration wedge ice.

Blanket suglinoks have been studied in the Far North, on the Taimyr Peninsula[7] and in central Yakutia.[2,6] Furthermore, we described the formations situated in the Podkamennaya Tunguska basin, in the Lena-Vilyui interfluve, in the vicinity of the town of Lensk, and in the Lena delta. In the southern regions of central Siberia, blanket suglinoks may pass into typical loesses. Blanket suglinoks are found not only in interfluves and on gentle slopes, but also in valleys. They are absent only on the floodplains and first terraces. This also holds true in western Siberia. Blanket suglinoks form on deposits of highly diverse origin, although lake-swamp deposits and shifting eolian sands constitute an exception. It is evidently on this basis that several varieties of blanket suglinoks, possessing dissimilar morphological features and a differing degree of expression of the loessial characteristics, can be distinguished.

In central Siberia, in conformity with the region's geological and geomorphological structure, blanket suglinoks are very widely developed on bedrock. In contrast to the blanket suglinoks on the fine-grained deposits, those overlying bedrock are less thick. As a rule, the thickness does not exceed 0.5-0.8 m and frequently falls to 20-30 cm. In the lower layers of the bedrock, there are large quantities of gravel and bedrock detritus, into which the blanket suglinok gradually passes. As with many of its other characteristics, the thickness of the suglinok blanket depends in large measure on such regional conditions as the relief, geological structure, and tectonic regime. Thus, highly favorable conditions for its formation exist on the ancient surfaces of planation. In areas that have been subject to tectonic uplift, or on steep hillsides, blanket suglinoks are superseded by coarsely lumpy placers. All these characteristics bear witness to the long duration of the process whereby compact bedrocks were subjected to weathering.

Due to the latitudinal variations in the natural conditions of freezing and thawing, substantial changes occur in many of the characteristics of blanket suglinoks. Thus, whereas the thickness of the blanket suglinok developed on the fine-grained deposits in central Yakutia and in the Podkamennaya Tunguska basin averages 1.2-3.0 m, in the extreme east it does not exceed 0.8-1.0 m. Inasmuch as southward of the line between the mouth of the Podkamennaya Tunguska and the town of Kirensk on the Lena River the blanket suglinoks are in the thawed state, and northwards of this line the lower part of the layer of these deposits is, as a rule, in the perennially frozen state, in the northern half of central Siberia they are distinguished by the presence of a high content of segregation stratal and regeneration wedge ice. Furthermore, in the north there is a higher content of organic matter and alterations in the texture and other properties. All of the existing data for central Siberia confirm the correspondence between blanket suglinoks and seasonally frozen and thawed layers. The thickness of the blanket suglinok on coarse-grained materials, however, is usually less than that of the seasonally frozen and seasonally thawed layers, while on loose, fine-grained deposits it may be one-and-a-half times to twice as thick. This circumstance is indicative of the changes that occurred in the thickness of the seasonally thawed and seasonally frozen layers as a result of fluctuations in the climate.

Certain characteristics of loessial soils, primarily the siltiness and carbonate content, pale yellow color, and capacity for forming vertical walls in the event of prior thawing, are also noted in loose deposits of great thickness. These are much thicker than the layer of seasonal thawing and measure 10-20 m or more. The thick loess deposits, including typical loesses, have been studied to best advantage in the southern part of central Siberia. These regions are characterized by relatively moderate geocryological conditions, and the loess deposits are, as a rule, in the thawed state. The alluvial and hillside deposits are loessial. In the more northerly regions of central Siberia, however, where all of the rocks are in the perenially frozen state, loose deposits that have certain loessial properties, and also typical loesses, are very often reported. These properties are most frequently reported in the alluvial deposits of terraces situated above floodplains in the eastern regions of central Siberia[3] and have been best studied in the valley of the Aldan River.[9,11] They are also known to exist, however, in the valley of the Vilyui and the tributaries of the lower Lena, in the basin of the Olenek, and elsewhere. While alluvial deposits are a complex combination of a number of facies, loessian properties are characteristic of the suglinok deposits of the high floodplain facies proper.

In these regions, the suglinok deposits of gently sloping hillsides,[1] at whose bases sediments of great thickness can accumulate, also have a number of loessial properties.

Being silty in composition, the loess deposits or those coeval with them are, as a rule, characterized by a higher ice content and show signs of syngenetic freezing: bedded or concavely bedded cryogenous textures and syngenetic regeneration wedge ice. The analogous deposits in the southern regions, which are currently in the thawed state, bear uniform traces of cryogenous formations: pseudomorphs substituting for regeneration wedge ice and various cryogenous disturbances in the stratification.[4,8] Based on the cryogenous structure of these deposits or traces of the same, it is possible to visualize the natural historic conditions that resulted in their formation and loessification. As is well known, syngenetically freezing sediments with bedded cryogenous textures and syngenetic regeneration wedge ice are restricted to fairly severe geocryological conditions. In central Siberia, judging from the sum total of the paleogeographic data, these conditions may approximate to the conditions of low-center polygon tundra landscapes.

Thus, the existing data indicate that granular deposits differing in their origin and possessing loessian properties are widely prevalent throughout central Siberia as a whole. Loessification occurs under severe climatic and geocryological conditions and is associated with seasonal freezing and thawing.

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MIGRATIONAL FROST MOUNDS

V. P. EVSEEV Moscow Lomonosov State University

Migrational frost mounds are widespread in the north of the European part of the USSR and western Siberia, both in the forest tundra zone and in the northern part of the taiga.[1] They show certain disparities with respect to their occurrence with genetic landforms, which are in turn caused by disparities in the lithological structure of the upper layers of the Quaternary deposits of these regions.

The lithological composition of the rocks in the north of western Siberia indicates that the conditions there were more favorable than in the European part of the USSR for the areawide development of migrational frost mounds.

When discussing the distribution patterns of frost mounds within genetic landforms, it should be noted that they are sometimes confined to areas in which well-sorted, fine-grained soils are replaced by sands. This is attributable to the fact that it is precisely here that lateral migration processes, which have an important role in the formation of frost mounds[1] take place most intensively, since clay has a greater particle surface energy than has sand.

The lithological structure of the areas in which migrational frost mounds occur influences the height of the mounds in a definite way. A vertical section through such mounds shows peat, a layer of sands or supesses (sometimes an interstratification of these), and, lower down, well-sorted suglinoks and clays. Since these latter heaved the most, it is the thickness ratio of the above-mentioned lithological varieties within the layer of active cryolithogenesis that governs the height of the mounds. Thus, when peat is absent and there is a predominance of coarse-grained varieties and a decreased thickness of fine-grained soils, mounts 1-2 m in height form. It is within this group that the hummocky-sinkhole landforms, which figure predominantly in the river valleys of the northern part of western Siberia, should probably be placed. If peat and homogeneous clays and suglinoks occur in a section in which there are only moderate thicknesses of sands and supesses, then the formation of entire areas of heave with individual mounds of up to 8-10 m in height is possible.

The analysis of the lithological composition of soil material in frost mounds reveals that the height of the mounds is greatest when the thickness ratio of this material is as follows: a 1.0-1.5-m layer of peat, a layer of sand or supesses measuring 2-3 m, and a 5-6-m layer of well-sorted clays or suglinoks.

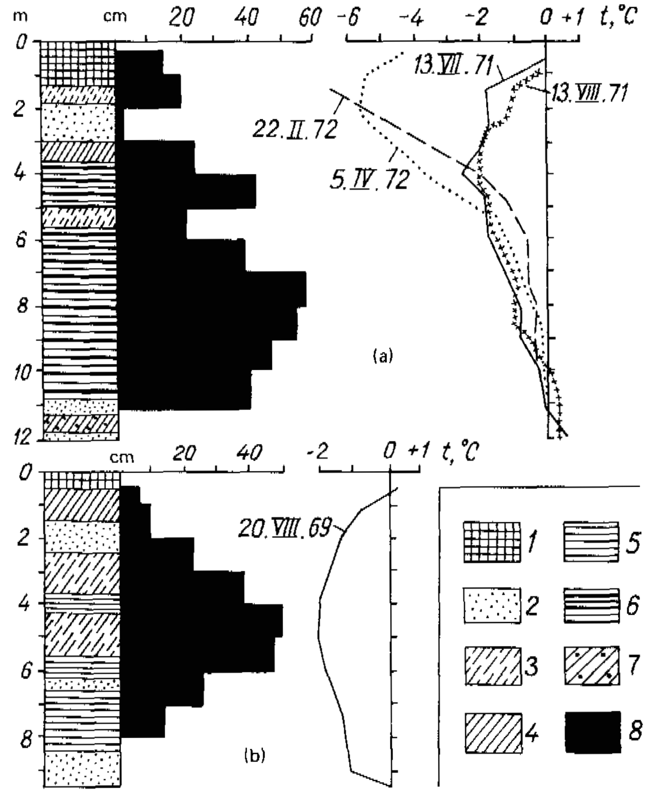

FIGURE 1 Distribution of the overall thickness of the ice inclusions and temperatures in a frost mound. (a) In the north of the European part of the USSR (village of Abez'); (b) In the north of western Siberia (basin of the Pravaya Khetta River). 1--peat; 2--sand; 3--supes; 4--suglinok; 5--clay; 6--varved clay; 7--boulder suglinok; 8--overall thickness of the ice inclusions.

In the region of sporadic permafrost, areas in which frost mounds have developed appear as permafrost islands. In outline, a single mound forms a frozen island, equal in area to the base of the mound. Concentrations of mounds or areas of heaving form islands of permafrost measuring several square kilometers, the area of these as a rule exceeding that of the frost mounds.

The thickness of the permafrost beneath frost mounds in a zone of sporadic permafrost is variable and depends directly on the height of the mounds and the area of heaving. Beneath single mounds averaging 2-4 m in height, the thickness of the permafrost is 10-12 m. At heights of 1-2 m, its thickness is 6-8 m. In the case of high mounds (8-10 m), which are only to be seen in areas of heaving, the permafrost is more than 15-20 m thick. In such instances, in the northern part of western Siberia the permafrost beneath the mounds may merge with a second layer of permafrost.

The thickness of the perennially frozen material beneath frost mounds is also seen to depend on its region of occurrence. Thus, in the north of the European part of the USSR, at the southern limit of migrational frost mounds (village of Abez'), the thickness of the permafrost beneath a mound measuring 3.5 m in height is 11 m (Figure 1a). At the same time, at the northern limit of their occurrence in this region (village of Seida), where perennially frozen soils predominate over thawed soils, the thickness of the permafrost beneath a mound of the same height is 16 m, and in a number of cases it merges with the permafrost of the zone in question.

The temperature field of a frost mound is highly distinctive. The ground of the intramound areas has a temperature that is either above-freezing or close to 0°. The ground making up these mounds has a fairly low temperature. A close relationship is observed between the height of the mound and its temperature: the higher the mound, the lower its temperature.

We shall consider the temperature regime of the ground situated within frost mounds underlain by thawed material (Figure 1).

Illustrated in Figure 1a are temperature measurements conducted in July, August, February, and April in a frost mound situated in the European part of the USSR (village of Abez'). It will be seen that in summer the lowest temperatures (-2.4°) were noted at a depth of 4-5 m. Roughly similar temperatures were noted at this depth in February and April. Further down there is a gradual rise in temperatures to 0° at a depth of 10-11 m. Figure 1b pertains to a single measurement conducted in August in a frost mound in western Siberia.

Here, the lowest temperatures (-2°) were observed at a depth of 4-5 m. It should be noted that the heights of these mounds are approximately the same and amount to 3.5 m and 3.0 m, respectively.

It can therefore be stated that the effect of the surface conditions (such as fluctuations in the air temperature and the thickness of the snow cover) largely disappears at a depth of 4-5 m. In the zone being discussed, this depth is identical in the northern regions of the European part of the USSR and western Siberia. The winter temperature gradients in this layer range from 2.4 to 1.4 deg/m (see Figure 1). Beginning at a depth of 4-5 m, there is a slow increase in the ground temperature. The temperature gradient in this layer is approximately the same in winter and summer and amounts to 0.2-0.4 deg/m.

The analysis of the temperature field of the frost mounds shows that the thickness of the layer of active cryolithogenesis in which the temperature gradient exceeds 1 deg/m^2 is 4-5 m. Situated below this layer is a layer of passive cryolithogenesis with temperature gradients of 0.2-0.4 deg/m.

A. I. Popov[2] shows that frost mounds are formed as a result of intensive migration of moisture in the layer of active cryolithogenesis and that the height of the mounds corresponds to the total number of ice inclusions in this layer.

Our investigations in the northern regions of western Siberia[3] and also of Europe corroborated Popov's viewpoint that the height of the mounds corresponds to the overall thickness of the ice inclusions. As is indicated in Figure 1, however, at the present time the thickness of the ice

TABLE 1 Determination of the Thickness of the Layer of Active Cryolithogenesis

Region of Investigation	Height of Frost Mounds, m	Contemporary Thickness of Frozen Ground beneath Frozen Mounds, m	Total Thickness of Ice Inter-layers, m	Thickness of Layer of Active Cryolithogenesis at Moment of Freezing (without Allowing for the Ice), m
North of western Siberia	1.0	8.0	1.0	7.0
	2.0	10.0	1.7	8.3
	3.0	9.5	2.2	7.3
North of the European part of the USSR	3.5	12.0	3.5	8.5

interlayers determining the height of a mound falls mainly within the layer of passive cryolithogenesis. It can therefore be assumed that at the time of formation of the mound, the thickness of the layer of active lithogenesis was greater. On the other hand, with the formation of the cryogenous textures, a part of the layer of active cryolithogenesis shifted its position to that of a layer of passive cryolithogenesis.

It will be seen from Table 1 that if a deduction is made in respect of the amount of ice contained in a layer for which the total thickness of the ice inclusions has been calculated, the thickness of the layer will be 7-9 m. This value will also fit the layer of active cryolithogenesis at the moment when the frost mounds began to form. In view of the fact that at the present time the layer of active cryolithogenesis with a thickness of this magnitude is situated to the north of the zone containing contemporary migrational frost mounds, it can be assumed that the conditions that coincided with the formation of the frost mounds were more severe.

The thickness of the layer of active cryolithogenesis and the height of the mounds also determine the character of their cryogenous structure.

Thus, in frost mounds with heights of 1-3 m, with increases in depth, a regular replacement of reticulated layered textures is to be seen, as well as an overall rarefaction of the ice network, the cryogenous structure being directly dependent on the lithology of the soils. In frost mounds with heights of 5-10 m, it is preeminently basal textures that are noted, which are also typical of sand and peat. Here the rarefaction of the ice network begins at a depth of 15-18 m.

This disparity in the distribution of the cryogenous textures can be explained in the following way. Since the overall thickness of the ice interlayers (or the height of the individual mound) is determined solely by the intensity of the migration processes in the layer of active cryolithogenesis, the distribution of a disparate number of ice inclusions in a layer of the same thickness will also determine the type of the cryogenous textures. In the event that the overall thickness of the ice inclusions does not exceed the thickness of the layer of active cryolithogenesis at the moment of freezing, frost mounts measuring 1-3 m in height with netted and layered textures will form. In the event that the thickness of the ice interlayers is equal to or greater than the thickness of the layer of active cryolithogenesis, high frost mounds (8-10 m) that have typical basal textures will form.

The following conclusions can be drawn from the foregoing description of migrational frost mounds:

1. Migrational frost mounds have a local distribution in the north of the European part of the USSR and an areawide distribution in the north of western Siberia, which is caused by the lithological composition of the soil material in the upper layers.
2. At the present time the layer of active cryolithogenesis in frost mounds equals 4-5 m in thickness, whereas at the moment of their formation it amounted to 7-9 m, which is indicative of more severe climatic conditions in the initial stages of formation of the mounds.
3. Their cryogenous structure is determined by the height of the frost mounds and the thickness of the layer of active cryolithogenesis.

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A COMPARATIVE DESCRIPTION OF THE COMPOSITION AND PROPERTIES OF THE LACUSTRINE ALLUVIAL DEPOSITS OF NORTHWESTERN SIBERIA AND THE YANA-INDIGIRKA LOWLAND, CONSIDERED IN THE LIGHT OF CRYOLITHOGENESIS

L. A. ZHIGAREV AND I. S. BOCHAROVA Operations and Research Institute for Engineering Site Investigations (PNIIIS)

Today, the influence of the epigenetic and syngenetic types of freezing on the cryogenous structure, saturation, and formation of large accumulations of ice is widely known.[1,5,7,9] It is essentially on the basis of these characteristics that layers of permafrost are subdivided into epigenetic and syngenetic. It has also been established that the different types of freezing are of great importance in the cryodiagenetic transformation of sediments, and, in particular, in their lithification.[2,10] Almost nothing has been done, however, to elucidate the formative role of the particular type of freezing in the physical constitution and engineering geological properties of deposits. It is with the object of helping to fill this gap that the present paper has been written.

In certain regions situated in the northen part of western Siberia (the Yarudei and Polui rivers) and the Yana-Indigirka lowland (the Burguat and Ulukhan-Yurege rivers and Lake Spirka), there are silty-clayey lacustrine alluvial deposits of upper Pleistocene age that contain ice. Qualitatively, despite their regional discreteness, these deposits have many common features that are the result of cryolithogenesis. They are characterized by a high silt content, a predominance of hydromica with low colloidal activity in the clayey fraction, a very low content of electrolytes, and the presence of iron and aluminum hydroxides and of organic matter of a definite type.

Nevertheless, the content of these components in the lacustrine alluvial deposits of northwestern Siberia and the Yana-Indigirka lowland lacks uniformity (Table 1), which is due to the conditions of formation of the deposits and preeminently, it would seem, to the characteristics of their cryodiagenetic transformations.

It will be seen from Table 1 that the lacustrine alluvial deposits of northwestern Siberia are distinguished by a lack of uniformity in their constituent material, a predominance of sandy-silty fractions, and a low content of organic matter. The deposits of the Yana-Indigirka lowland are characterized by uniformity of particle-size distribution, an absolute predominance of the silty fractions, a high content of organic matter, and a very low absorptive capacity. This latter characteristic is evidently associated with the screening capability of organic ferruginous compounds.

The formation of the lacustrine alluvial deposits in the northwestern Siberia was, in the main, accomplished before they began to freeze, when they had entered upon the stage of diagenesis and had been little affected by cryogenous transformation. This has been reflected in the composition of the deposits. In the particle-size analysis, the yield of the fractions consisting of clayey particles is higher and that of the silty fractions lower than is the case in the grain-size analysis. This probably indicates that the increase in the silty fractions occurred as a result of aggregation of the clayey particles. The low content of organic matter can evidently be explained as follows: Being already in the stage of sedimentation and early diagenesis, it was carried away by surface and interstitial water and the freezing of the deposits fixed the remaining amount of it. This is corroborated by the fact that there is a much higher organic matter content in the solid runoff from the rivers

TABLE 1 Composition of Lacustrine-Alluvial Deposits

Region	Particle Size Distribution (Numerator) and Microaggregate Composition (Denominator), %: Sand, 0.5–0.05 mm	Silt, 0.05–0.005 mm	Clay, 0.005 mm	Organic Matter Content, %	Content of Electrolytes, %	Absorptive Capacity, mg/equiv per 100 g of Absolutely Dry Soil
Northwestern Siberia	20-37 / 19-31	45-55 / 55-62	17-26 / 9-21	0.11-0.38	0.035-0.60	10.28-13.58
Yana-Indigirka lowland	7-20 / 12-25	55-73 / 54-72	20-25 / 15-20	1.81-3.61	0.050-0.078	4.00-4.63

and in the recent sediments accumulating on the floodplain.

The formation of the lacustrine alluvial deposits of the Yana-Indigirka lowland occurred simultaneously with freezing. Naturally, by the time that they became frozen, these deposits, in bypassing the stage of ordinary diagenesis, had undergone substantial cryogenous transformation, which has likewise been reflected in their composition. To all appearances, the identical content of silty fractions in the particle-size and microaggregate analyses attests to their having originated in a fine-grained state, inasmuch as the yield of the silty particles was not affected by the differences in the preparation of the soil for analyses. The fragmenting of the sandy particles to the sizes of silty particles depends on the frequency of the phase transformations and the extreme temperature values of the deposits.

With the epigenetic mode of freezing, except for the upper layer, the deposits undergo a smaller number of phase transformations and their temperature range is narrower than is the case with syngenetic freezing. They therefore contain a smaller number of silty particles. The substantial content of organic matter in the lacustrine alluvial deposits of the Yana-Indigirka lowland is due to the permafrost preservative factor.[8]

With freezing of the deposits from below, when the sediments are in the process of accumulating, the organic matter together with the subsiding vadose water is concentrated at the surface of the water-confining stratum, which is frozen soil. The entire series of syngenetically frozen deposits is therefore more or less uniformly saturated with organic matter. At a depth of 14 m, the organic matter content is 2.2 percent.[6] At the surface of the water-confining stratum, many water-soluble compounds can accumulate, in particular, ferrous forms of iron. It is therefore probable that the syngenetically frozen deposits contain a somewhat larger quantity of water-soluble salts than the deposits that froze epigenetically.

The degree of fineness and chemical and mineral composition of the deposits predetermined the specific structural characteristics of the lacustrine alluvial deposits and the main differences in their properties subsequent to freezing. The character of the structural bondings is determined from the aggregation coefficients for particles smaller than 0.001 mm and 0.005 mm.[4] In the lacustrine alluvial deposits of northwestern Siberia, subsequent to thawing, massive structures form that have interparticle bonds of the stabilization type. The aggregation coefficient ranges from 1.66 to 1.78 for particles < 0.001 mm and from 1.22 to 1.24 for particles < 0.005 mm. In the deposits of the Yana-Indigirka lowland, subsequent to thawing, massive structures form that have interparticle bonds of the plasticized-coagulation type. The aggregation coefficient of these deposits ranges from 2.23 to 2.64 for particles < 0.001 mm and from 1.23 to 1.28 for particles < 0.005 mm.

The physical and mechanical properties of the epigenetically frozen (northwestern Siberia) and syngenetically frozen (Yana-Indigirka lowland) deposits subsequent to thawing are given in Table 2.

It will be seen from Table 2 that epigenetically frozen deposits with a low content of organic matter are characterized by a higher density and compactness, although they have lower porosity values, plasticity, strength, and hydrophily. Deposits with such characteristics are fairly resistant to static and dynamic influences. With the increase in moisture content, however, they become markedly less strong, less compact, and more liquescent.[3] These deposits are almost nonthixotropic on account of the lowered capability for the formation of thixotropic structure by the fine-grained constituents.

The syngenetically frozen lacustrine-alluvial deposits of the Yana-Indigirka lowland differ from the deposits of northwestern Siberia in having a lower density and unit weight and a higher porosity and plasticity, even though the strength is greater, which is evidently because of the higher content of organic matter. In these deposits the thixotropic properties are more pronounced than in the deposits of northwestern Siberia.

The dissimilar properties of the lacustrine alluvial deposits are giving rise to differences in the development of the natural processes in the regions being discussed. In northwestern Siberia there is widespread gullying, ground slumping is prevalent on the hillsides, and the

TABLE 2 Properties of Lacustrine Alluvial Deposits

Properties of Deposits	Northwestern Siberia	Yana-Indigirka Lowland
Grain density, g/cm^3	2.67-2.70	2.62-2.67
Dry unit weight, g/cm^3	1.55-1.79	1.24-1.70
Void ratio	0.50-0.66	0.57-1.11
Degree of saturation, %	0.84-0.95	0.81-1.06
Liquid limit, %	24.1-24.1	26.2-38.9
Plastic limit, %	17.1-19.9	19.4-27.8
Plasticity index	4.22-6.48	6.80-11.1
Liquidity index	0.24-0.43	0.28-0.57
Hydrophily index	0.99-1.59	1.06-1.98
Resistance to penetration, kg/cm^2	0.72-4.65	2.80-5.20

thawing out of polygonal wedge ice is resulting in the formation of a hummocky-sinkhole type of relief. Within the Yana-Indigirka lowland, gullying is due almost entirely to the thawing out of polygonal wedge ice; slumps occur only on slopes that are in the process of being undercut and when offshore scarps are eroded and baijerakhs (cemetery mounds) form when the ice melts.

Thus, the differences in the composition and properties of these deposits, which are caused by diagenesis, make it possible to distinguish two principal engineering geological types of lacustrine alluvial deposits in the northern part of Siberia: epigenetic and syngenetic.

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CRYOGENOUS DEPOSITS OF THE UPPER REACHES OF THE TATTA RIVER (CENTRAL YAKUTIA)

M. S. IVANOV AND E. M. KATASONOV Permafrost Institute, Siberian Section, Academy of Sciences of the USSR

We use the term cryolithogenous to describe deposits that form under permafrost conditions. These deposits are the result of cryolithogenesis, that is, the aggregate of geological and cryological processes originating in a layer of seasonal thawing or in taliks and leading to the formation of ground that is characterized by a distinctive structure. They freeze either syngenetically if the requisite cryogenous textures form in them as a result of weak compaction,[4] or parasyngenetically, when the sediments while still in the thawed stage undergo the initial stages of diagenesis and are subject to lithogenetic fracturing, with the result that cracked cryogenous textures form. We will apply the method of permafrost facies analysis to cryolithogenous deposits.[6]

During the period 1969-1970, geocryological studies were conducted in the upper reaches of the Tatta River, in central Yakutia. A total of 30 holes, ranging in depth from 20 m to 60 m were drilled over a comparatively small area, and data were obtained on the temperature regime of the ground and the distribution pattern of the ground ice.

In the upper reaches of the Tatta, four basic relief forms are distinguished: outliers consisting of Jurassic sandstones, an aggradational surface of Pleistocene age, alaslike hollows dissecting the aggradational surface, and the contemporary valley of the Tatta.

The Pleistocene deposits have a thickness of up to 55 m and consist in the main of silts, their mineralogical composition being arkosic and subarkosic with a high degree of aggregation of the clayey particles. Two cryolithogenic complexes are distinguished in the drillhole sections: below, a subaqueous; above, a subaerial (see Figure 1).

The subaqueous complex is made up of muddy bluish-gray silts with inclusions of plant remains and fragments of wood (driftwood). A. I. Popova has concluded that the pollen and spore content makes it possible to date this complex to the Ice Age. It does not contain distinctive

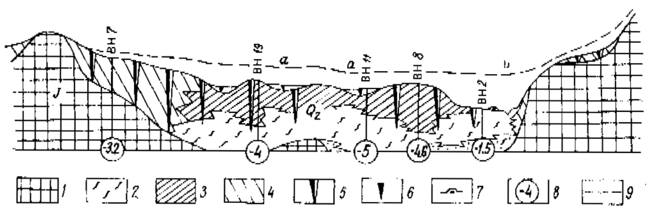

FIGURE 1 Schematic representation of the structure and disintegration of the Quarternary deposits as a result of thermal denudation. 1--bedrock; 2--syngenetically frozen deposits of shallow bodies of water--the ice complex; 3--deposits of the swampy depressions and alluvial fans; 4--valleyside deposits--the ice wedges in the deposits; 5--of Pleistocene age; 6--of Holocene age; 7--cores of open pingos; 8--ground temperature at depths of 15-20 cm; 9--surface of the ice complex by the time that it began to disintegrate: (a) thermal denudation hollows; (b) alas valley. The seasonally thawing deposits are not shown.

sills of ground ice. At the top, only the ends of the wedges intrude into it from the subaerial complex. The ice inclusions consist of broken lenslike streaks. The latter have a slanting orientation and cut across the sedimentary bedding, which attests to the existence of a perennially frozen substrate: an orienting factor in the syngenetic freezing of the bottom sediments. Some of the drillholes revealed subaqueous deposits that had frozen parasyngenetically. Their cracked cryogenous textures indicate that these sediments had originated some distance away from a frozen mass (they formed a large talik) and, therefore, until their transition into the perennially frozen state they were subject to the initial stages of diagenesis. In general, the complex of subaqueous deposits forms a single layer measuring 15-20 m thick. Its ice content is not great, being 10 to 20 percent by volume, and the icy streaks become thinner with depth.

The subaerial (ice) complex, measuring up to 25-30 m thick, consists of loesslike, peat-laden, less frequently gleyed suglinoks and supesses with typical hairlike rootlets of grasses and bedded lenslike or reticulate cryogenous textures and contains ice wedges of up to 6 m wide in the upper parts. The total ice content of these deposits is 60-80 percent by volume. Their buildup and gradual (syngenetic) freezing, judging from the bedded cryogenous textures, occurred when there was a surplus of moisture in a layer of seasonal thawing, the thickness of which did not exceed 30-100 cm.

These complexes are differentiated by the coloring of the ground, the content of ice, and the cryogenous structure. The boundary between them--an uneven one, with projections and large depressions--is difficult to make out. The waterborne and terrestrial deposits exhibit facies transitions one into the other (see Figure 1). It is typical that in some cases the subaerial deposits enclosing the ice wedges pass gradually downwards into the waterborne sediments, while in others they lie directly on the Jurassic sandstones. By filling the irregularities in the topography, they ascend along the original slopes to considerable heights. These facts afford evidence that the glacial complex is genetically heterogeneous and consists not only of the floodplain accumulations of the Greater Tatta, but also of deposits received from the slopes and alluvial fans of its tributaries, ravines, and drainage hollows. G. F. Gravis'[2] conclusion as regards the hillside or mixed hillside-floodplain origin of the loesslike suglinoks filling the shallow valleys is borne out by the example of central Yakutia.

The formation of an ice complex amid conditions of dissected relief is the main reason for the absence of closed alasses in the region being considered. The uneven, steeply sloping surface precluded the existence there of water bodies--the main factor in the development of thermokarst basins. Thermal erosion and thermal denudation processes predominated. The formation of the indistinct alaslike hollows is associated with them. In its upper reaches, the Tatta Valley itself is a typical flat-bottomed alas valley with sharply delineated widenings, constrictions, and blindly terminating branches.

It is relatively easy to correlate and compare the Pleistocene deposits of the upper reaches of the Tatta with those of the key sections of central Yakutia. Subaqueous sediments similar to those described above are revealed in the "Mount Mamontov" and "Mount Chuiskaya" outcrops and at other points along the banks of the Aldan, as well as in sections of the 50-60-m terrace (the third terrace according to G. F. Lungersgausen;[5] the Abalakskaya terrace according to P. A. Solov'ev.[7] Their thickness can be as much as 40 m. Upwards along the section they are replaced by the deposits of the ice complex, the boundary with which is very uneven. These sediments, forming a kind or marker horizon, are revealed by the drillholes to the south of Borogonets (at the latitude of Yakutsk). Pebbles and boulders appear in them increasing proximity to the Verkhoyansk Mountains; they gradually pass into moraine.[3]

The age of the terrace consisting of the lacustrine deposits is determined as middle Quaternary.[1,5] Their geological and cryogenetic features are indicative of the existence at that time of "permafrost." A heavily lake-studded plain stretched southwards from the glaciated Verkhoyansk Mountains. The spacious, shallow bodies of water were separated by relatively narrow tracts of land containing polygons and accreting ice wedges. As sediment accumulation proceeded, these tracts both widened and tapered, with the result that in the sections we now observe the masses and blocks of the ice complex, which pass like roots into the lake sediments.

In the late middle and early upper Pleistocene, the lake-studded plain was superseded by a polygonal plain. Judging from the sections of the younger terraces of central Yakutia (in the upper reaches of the Tatta they do not show up), the humid permafrost-landscape situation, which had evolved against a background of mountain glaciation, was gradually replaced by a semiarid permafrost situation. Here, the contemporary cryolithogenous deposits, as distinct from those of upper and middle Pleistocene age, consist mainly of ice-poor soil material containing primordial ground wedges. Swamp facies, of which ice wedges are typical, are only rarely encountered.[4]

The investigations in the upper reaches of the Tatta are an example of the genetic approach to the study of permafrost. Their results support one of the basic tenets of cryolithology: the connection that ground ice has with genesis, that is, with the conditions of accumulation and freezing of the enclosing deposits. In the final analysis, the distribution and position in the section of cryolithogenic complexes, which vary in their ice content, is predetermined by the evolution of the paleogeographic and permafrost-landscape conditions.

The history of formation of the cryolithogenous deposits and the relief forms associated with them also explains, in our view, certain features of the thermal regime of permafrost. In the upper reaches of the Tatta, the highest mean annual temperatures (-1.5°C to -2.7°C) are reported for the holes drilled at the bottom of the Tatta Velley (see Figure 1, Drillhole 2); the lowest (-4.0°C to -5.4°C) for the deposits of the ice complex and thermal erosion hollows (see Figure 1, Drillholes 8, 11 and 19). We attribute these differences to the fact that the ice complex consists of low-temperature facies containing heavily peat-laden deposits, the present-day analogues of which in central Yakutia currently have temperatures of -6°C to -7°C. The breakup of the ice complex, which occurred as a result of thermal denudation, did not essentially alter the moisture content and thickness of the layer of seasonal thawing and, consequently, the character of the heat exchange. The alas valley formed as a result of the repeated widening and drying out of the water bodies. At the present time, the deep or open taliks, which have existed at certain of them, are continuing to freeze or have only just become frozen, evidence of which is to be seen in the development there of bulgunnyakhi (open-system pingos).

In the case we have been discussing, namely lacustrine bodies of water, the importance of morphogenetic factors in the shaping of the temperature regime of the permafrost becomes intelligible if the effect of these factors is considered in a historical context. In the past, just as is the case today, the thermokarst depressions were periodically inundated at intervals ranging from tens to hundreds of years, and this could not have had an effect on the temperature conditions of the ground that constitutes their bases.

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GEOCRYOLOGICAL CONDITIONS OF THE BASIN OF THE MIDDLE OLEKMA

T. N. KAPLINA AND O. P. PAVLOVA   Operations and Research Institute for Engineering Site Investigations (PNIIIS)

Until very recently the geocryological conditions of the area making up the basin of the middle Olekma and its main tributary, the Nyukzha River, had not been studied at all. Field investigations conducted in this region by the authors during the period 1969-1970 have made it possible to work out the basic concepts with respect to the peculiarities of distribution and fluctuations in the thicknesses of the permafrost, its temperature regime, and the cryogenous structure.

The basin of the middle Olekma encompasses a vast mountainous region, the northwestern part of which is a sharply dissected area of medium-high mountain relief in which the absolute elevations of the watersheds and valleys range from 900-1,500 m and 300-400 m, respectively. The southeastern, low-mountain relief part has smoothed watersheds of not more than 1,000 m high and wide valleys with absolute elevations of 400-600 m. The tectonic structure of the area is governed by its confinement to the juncture of the greatly uplifted folded zone of the Baikal Rift (the northwestern part) with a lagging block of the northern limb of the Stanovoi Anticlinorium (the southeastern part). The predominant formations within the basin of the middle Olekma as a whole are massive crystalline rocks of pre-Cambrian age. Quaternary formations ranging from 1.5 m to 30 m in thickness consist of residuum and the alluvial and swamp deposits of the slope sequence.

Low mean annual air temperatures (from -6.5°C to -8°C), in conjunction with the high absolute elevations and mountainous nature of the relief have predetermined what is for the most part a harsh geocryological environment. The basic and the sole reason for the existence of open taliks is the groundwater, localized within a widely ranging network of very recent and rejuvenated tectonic dislocations.

It is possible to distinguish in the region three main topographic levels, within each of which the degree of continuity in the distribution and the thickness of the permafrost will be different.

The first level includes the large and medium-sized river valleys of the Olekma; Khana; Nyukzha; Lopcha; upper, middle, and lower Larb, and others. The deeply incised swamp-covered valley bottoms with a thin covering of snow are subject to pronounced cooling as a result of a temperature inversion and are characterized for the most part by continuous permafrost.

The mean annual temperatures of the ground in the valleys range from above-freezing to -5°C and display a distinct relationship with the surface conditions and lithology. Furthermore, they are somewhat higher along a line extending from the northwest to the southeast on account of the changes in the climatic conditions (the rise in the mean annual air temperatures and the increase in the thickness of the snow). The mean annual temperatures for the most typical conditions existing in the valleys are presented in Table 1.*

Above-freezing temperatures are typical of the closed taliks with a thickness of up to 5-10 m that have developed in the lake-studded tracts of the high floodplain and within the confines of the sandy crests of the latter.

The results of deep drilling indicate that the thickness of the permafrost in the valleys ranges from 100-200 m. Close to the thawed flooded zones of the tectonic dislocations, the thickness of the permafrost may decrease to between 10 m and 50 m. In almost every valley, open water-bearing taliks are observed, as evidenced by heavy icings with volumes ranging from several hundred to between 3 million and 4 million cubic meters of ice. The area occupied by the taliks is sharply circumscribed by the dimensions of the water-removing tectonic dislocation and is usually of moderate proportions.

The second level incorporates watersheds with absolute elevations ranging from 800 m to 1,200 m (extensively developed in the southeastern low-mountain relief part of the area). As a result of the inversion, the mean annual air temperatures here have remained approximately the same as those in the valleys. At the same time, the increase in the thickness of the snow cover and the warming effect of the summer precipitation are conducive to a softening of the geocryological setting on the flat, smooth watersheds. The permafrost thicknesses here are ~100 m. According to the calculations, the mean annual temperatures of the ground range from -0.5° to -2.0°, which in conjunction with the presence of highly pervious rocks (fissured zones) establish the conditions that are necessary for the formation of atmospheric seepage taliks.

In the areas with a highly dissected relief (the left bank of the Olekma and the basin of the lower Nyukzha), the warming effect of the atmospheric precipitation does not show up within this level and, as a result the geocryological situation in the ridged watersheds, does not differ appreciably from that in the valleys.

Pertaining to *the third level* are watersheds with absolute elevations of more than 1,200 m (developed for the most part in the northeastern, medium-high mountain relief part of the area). At such elevations and in the presence of a highly dissected relief, the heat-exchange level is so low that the continuity of the permafrost is everywhere undisturbed, and its thicknesses range from 300 m to 500 m, although they may be higher than this. It can be assumed that the temperatures of the ground here fall below -5.0°C.

The study of the cryogenous structure of the perennially frozen bedrocks has been limited to that of the unconsolidated mantle.

The residuum of the low, smoothed watersheds, consisting of fine-grained material (supesses with rubble and gravel), is characterized for the most part by massive cryotextures containing occasional thin streaks and crusts of ice among the stones.

The residuum of the mountain summits and slopes of the medium-high region of mountain relief often consists of rocky and rubbly placers. In the absence of fine soil, ice serves as a filler in them below the layer of seasonal thawing. Locally, the ground takes on a basal cryotexture (fragments suspended in the ice). Such is the structure of the stony placers and rock streams situated on the slopes. On the steep and well-drained slopes, where coarsely clastic colluvium with a sandy or sandy-silty filler predominates, massive, or, less frequently, bedded-netted cryo-

*The temperatures were obtained as a result of comparing calculated values and data derived from *in situ* measurements. The calculations were performed by V. P. Chernyad'ev and I. P. Kuznetsova on an IG-1 hydrointegrator [hydraulic-analog computer].

textures predominate, sometimes in combination with ice crusts among the stones.

The accumulations of the gentle slopes (preeminently diluvial and diluvial-solifluctional) occupy large areas in the low-mountain relief region. They consist in the main of supesses containing rubble and gravel, and sometimes suglinoks, and are characterized by a higher ice content. The cryotextures are crustal, thinly bedded, and thickly bedded; irregularly shaped ice pockets are encountered. The ice content of the deposits on the gentle slopes ranges from 25 to 50 percent by volume.

The alluvial deposits can be numbered among those of Holocene age (the high floodplain and first terrace above the floodplain), as well as among those that are even older but have not been precisely dated (the terrace levels ranging from 15 m to 60 m).

In the main valleys the riverbed alluvium of the low terraces ranges from rubbly-pebbly (in the river valleys of the medium-high region of mountain relief) to finely pebbly and even sandy-gravelly accumulations (in the valleys of the low-mountain relief region). The deposits making up the shoals near the riverbeds consist of sands that vary in their degree of coarseness and at times even become silty. The thickness of the riverbed alluvium in the main valleys ranges from 1 m to 2 m and from 20 m to 30 m. Massive cryotextures (often a basal variety of them) predominate in the sandy-gravelly and sandy filler, in conjunction with ice crusts among the stones. It would appear that the riverbed alluvium is swollen in many of the areas, evidence of which is also to be seen in its high electrical resistivities (up to 180,000 ohms [sic]).

The floodplain alluvium consists of fine and silty sands and supesses, often peat-laden, while in the low-mountain relief region it also consists of suglinoks and has a thickness of between 2 m and 7-8 m. The sections of the low terraces, especially in those parts of them situated adjacent to the slopes, are often overlain by peat measuring 2-5 m in thickness.

Over the greater part of the area comprising the low terraces, the floodplain deposits are distinguished by a very high ice content, which is on account of their syngenetic freezing. Along with the basal cryotextures (in the sands), lensed, bedded, and reticulate textures are widely prevalent in the silty sands and especially in the supesses and suglinoks. An increase in ice content is to be seen from below upward along the section on account of the replacement of the sands by supesses and, higher up, by peat-laden supesses or peat. This is because of the emergence of thick (2-3 cm) ice streaks or "zonules," situated at intervals of 3-5 cm. Being superimposed on the thin bedding or reticulation, they form composite cryotextures. The ice content by volume is as follows: 25-50 percent for the sands with massive (including basal) textures, 30-60 percent for the supesses with multiple thinly lensed textures, and 50-70 percent for the peat-laden supesses and peat with composite textures.

The sands belonging to the riverbed ridges and floodplain crests, which are characterized by massive cryotextures and an ice content by volume of 20-30 percent, are less icy.

Over roughly 40 percent of the area, the alluvium of the lower terraces contain multiwedge ice. The development of the ice wedges is greatest in the peatlands, although they are also found in the supesses and silty sands. The wedges vary in their dimensions: the width ranges from 0.3 m to 1.5 m at the top; the vertical extent is from 1 m to 4 m. In the upper parts the wedges show signs of syngenesis, and in areas with mean annual ground temperatures of below -3°C they are actively growing. The dimensions of a network of ice wedges range from 5 m (in peatlands) to 20 m (in sands). The macro-ice content of the upper layer of alluvium, which can be ascribed to the multiwedge ice, ranges from 3 percent to 15-17 percent by volume, depending on the dimensions of the ice wedges and their network.

The alluvium of the high terraces is particularly extensive in the southeastern, low-mountain relief part of the area, although it is also locally developed in the valleys of the medium-high region of mountain relief (the lower reaches of the Nyukzha and Olekma) and consists in the main of sand, or, less frequently, conglom-

TABLE 1 Mean Annual Temperatures of the Deposits in the Valleys

Surface Conditions and Composition of the Deposit	Mean Annual Temperature of the Deposits, °C: Northwestern Region of Medium-High Mountain Relief	Southeastern Region of Low-Mountain Relief
Mossy and hillocky-mossy treeless mari, supesses containing peaty formations, silty sands	-3.5-5.0	-2.5-4.0
Tree-covered drained tracts with a woody floor; fine and silty sands	-2.0-3.0	-1.5-2.5

erate. The sands are characterized for the most part by massive cryotextures with an ice content of 20-30 percent by volume. Thus, an increase in the ice content of the alluvium is found between the high and low terraces.

The foregoing information affords evidence that the northwestern medium-high region of mountain relief and the southeastern low-mountain relief part of the area making up the basin of the middle Olekma vary appreciably in their tectonic structure, relief, and climate.

THE BAIKAL AND TRANS-BAIKAL TYPES OF PERMAFROST IN DEPRESSIONS

R. YA. KOLDYSHEVA All-Union Research Institute of Hydrogeology and Engineering Geology

The folded mountains of the Siberian Platform--Baikal and Trans-Baikal--exhibit differences not only in the relief, climate, and geological structure, but also in the geocryological conditions. The Baikal is characterized by high-mountain type of relief; the Trans-Baikal mainly by medium-high and low-mountain relief. In Baikal, where there is a harsh continental climate, the precipitation is heavier, while in the Trans-Baikal the winter temperature inversion is more abruptly manifested. A result of the disparate development of these areas in Mesocenozoic times was the formation of tectonic depressions of the Baikalian and Trans-Baikalian types (after N. A. Florensov).

Many investigators have noted that in the high-mountain region of Baikal and in some high-mountain regions of the Trans-Baikal the maximum thicknesses of the permafrost are confined to the mountain ranges, the minimum thicknesses to the river valleys and depressions; while overall, a reverse relationship is observed in the case of the medium-high mountain regions and especially the low-mountain regions of the Trans-Baikal. This is because Baikal is located in the upper geocryological belt.[1] According to the heat-balance calculations of V. P. Alekseev and A. T. Naprasnikov, in northern Baikal this belt begins at latitudes of 550-880 m; in the Trans-Baikal, the upper geocryological belt, being located above the 1,500-m level,[4] is a local phenomenon, whereas up to elevations of 1,000-1,100 m there is a decrease rather than an increase in the temperature. In the tectonic depressions, even more pronounced differences in the geocryological conditions are observed. These have given rise to numerous debates and attempts to distinguish them as to type (I. Ya. Baranov, V. P. Solomenko, I. A. Nekrasov *et al.*).

In the depressions of the Baikal and Trans-Baikal types, which extend over great distances and have a submeridional trend, the specific features in the distribution of permafrost are subordinate to a latitudinal geocryological zoning (the Barguzin, Chikoi-Ingoda, and other depressions).

The geocryological conditions are also dependent on the particular structural-tectonic and hydrogeological characteristics of the depressions that are typical of each of them.[2,3] Thus, the highest values of the heat fluxes and geothermal gradients, especially when the conditions are such that there is a discharge of groundwater and also perhaps of thermal water, obtain in anticlinal structures, which is conducive to a decrease in the thickness (or at times even to the total disappearance) of the permafrost. Minimum heat flux density values favor the formation of an anomalously great thickness of the permafrost in synclinal structures.

The smaller depressions (those of the order of 500-300 km^2 and less) that have a thinner mantle (200-600 m) and are situated in the upper geocryological belt in which the distribution of the permafrost is continuous, are completely frozen. The deeper depressions are frozen only in the upper part. In the large depressions of the Baikal type, however, under the influence of groundwaters and surface waters, taliks may occupy up to 20 percent or more of the area, thereby giving rise to a greater degree of discontinuity of the permafrost than is the case in the Trans-Baikal depressions (up to 12 percent).

The depressions of the Baikal and Trans-Baikal types, being located in the middle and lower geocryological belts, where the permafrost distribution is discontinuous and sporadic, have more heterogeneous geocryological conditions, which in turn depend on the distinctive geological-structural, neotectonic, climatic, and hydrogeological conditions. Based on the special features of the distribution (occurrence and degree of discontinuity) and formation of the permafrost, the Baikal and Trans-Baikal types of permafrost are distinguished in the tectonic depressions of Baikal and the Trans-Baikal.

The permafrost of the Baikal type is confined to the tectonic depressions of Baikal, which have large dimensions (the "mature" depressions of the Baikal type, after Solonenko), are filled with sedimentary Neogene-Quaternary deposits in which the Quaternary sediments are of great thickness (up to 500 m) and Pleistocene outwash sands are present, are highly seismic, are bounded by deep seismically active faults, abound in water, and contin in the lower part of the section highly thermal stratal waters.

It is chiefly the Pleistocene-Holocene deposits, occurring everywhere and at great depths that are in the frozen state, although in rare instances (the marginal portions of the depressions) the Neogene sediments are frozen.

The permafrost that has the greatest thickness (100-300 m), the lowest temperature, and the most continuous distribution is found in areas where the bottoms of the depressions have undergone recent subsidence[3] and that contain silty-peaty and both sandy-silty and clayey-silty soils. At the present time, the rapid accumulation of sediments derived from the excessively wet soils in the tectonically sunken blocks is contributing to the formation of the permafrost and to its anamalously great thickness.

According to Florensov, the high degree of discontinuity in the vertical section of the permafrost in the areas of discontinuous and sporadic distribution, and also the marked variations in its thickness, are a consequence of both the synchronous formation of the river valleys and Baikal depressions and the lack of uniformity in the trend of the neotectonic movements, given the nonuniform seepage properties of the freezing Quaternary soils.

The deep-seated (up to 100-150 m) permafrost table, as determined from the drillholes in the Tunkin and Barguzin depressions, is a consequence of the intensive activity of groundwater near the beds of the major rivers, and also of degradation of the permafrost, which, in the sandy Pleistocene massifs, is characterized by a low ice content.

The permafrost base is to a large extent subject to degradation, which is because of the active influence of geothermal heat in the Baikal depressions, given their unusual degree of seismic activity, steep geothermal gradients, and substantial resources of thermal groundwater.

The permafrost that is confined to the Quaternary deposits is of syngenetic origin. Pointing to this are the sediment accumulation characteristics (intensive downwarping of the depressions) coupled with the harsh climate that existed in Pleistocene-Holocene times and the high ice-saturation levels of the soils containing ice lenses and wedge ice in some of the areas.

The marked variability in the conditions of occurrence of the permafrost in each depression (the great thickness amplitudes, discontinuous distribution in the vertical sections, deeply lying permafrost table, and high mobility of its base), and also the syngenetic mode of its formation, are typical features of Baikal-type permafrost. Some of the leading factors in the formation of this type of permafrost are the active neotectonics of the depressions and the intensive interaction between the frozen soils and the groundwaters and surface waters, with the processes of insolation and heat absorption tending to play a secondary role in the evaporation of moisture from the surface of the saturated soils.

Permafrost of the Trans-Baikal type is prevalent in the tectonic depressions of the Trans-Baikal. In comparison with the depressions of Baikal, it is for the most part more fine-grained, consists of effusive-sedimentary Mesozoic-Cenozoic formations in which the Quaternary sediments are not as thick (up to 50-100 m), is less seismic and watery, and with rare exceptions stores up cold water.

Present in the frozen state in the depressions are variously aged (Mesozoic-Cenozoic) formations that have a varied cryogenous structure. The permafrost reaches a maximum thickness of 100-350 m at temperatures ranging from -0.1°C to -3°C. The permafrost table lies at a depth of up to 5 m, although in the south it can be as deep as 10 m or sometimes deeper.

The existence of permafrost in the intermontane depressions of the Trans-Baikal is undoubtedly influenced by the pronounced winter temperature inversion, which has a marked cooling effect on the bottoms of the valleys and depressions. It has also been repeatedly noted that the deposits of latitudinally trending basins are more deeply frozen than those with a submeridional trend. Here, the orientation of the valleys may determine not only the thickness, but also the specific arrangement[4] of the frozen soil masses in relation to the valley bed. This is because of the varying conditions of insolation in the individual depressions and their parts.

Given the relatively uniform heat-exchange conditions at the surface of depressions in denudation terraces, when the steepness of the slopes ranges from 4° to 15°, the thickness and positioning of the islands of frozen soil depend to a greater degree on their exposure than on the moisture content and thermal conductivity of the soil.[2]

In high alluvial terraces consisting of thick (up to 60 m) sands containing a deep-lying groundwater table, the formation of permafrost is wholly determined by the composition of the deposits and is observed in areas containing clayey and clayey-silty soils. In the event of the alluvial sands being overlain by diluvial argillaceous formations, according to V. P. Portnova the permafrost will be more extensively developed.

On the floodplain and first terraces above the floodplain, where the thickness of the alluvial deposits is not great (up to 6 m), there is a concentrated discharge of groundwater and a hydraulic linkup between the groundwater and surface waters, when the conditions of groundwater circulation are similar. The greatest thickness of the permafrost is confined to the areas with the wetted clayey-silty, clayey, and silty-peaty varieties (this was also noted by A. V. L'vov, P. F. Shvetsov, and N. A. Shpolyanskaya *et al.*) or those containing sandy deposits underlain by water resistant Mesozoic soils.[4]

The permafrost of the depressions of Trans-Baikalia is for the most part of epigenetic origin (in soils predating the Pleistocene), whereas in small depressions and troughs filled with Pleistocene-Holocene sediments, it originated syngenetically.

Thus, the formation of the permafrost of the

Trans-Baikal type is a result of the climatic peculiarities (the temperature inversion), the conditions of insolation in the individual depressions and their constituent areas, and the enormous losses of heat due to evaporation of excessively wet soils. On the whole, permafrost of the Trans-Baikal type is characterized by a clear-cut dependence on the altitudinal and latitudinal positioning of the area in question, with a predominance of soil materials of epigenetic origin.

In conclusion, it can be noted that given the presence of relatively uniform "external" factors in the formation of permafrost of the Baikal and Trans-Baikal types, the influence of "internal" factors on the formation of Baikal type permafrost is so great that they substantially modify the effect of the "external" factors.

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PERMAFROST AND LANDSCAPE

G. S. KONSTANTINOVA Moscow Lomonosov State University

Permafrost is one of the most important properties of landscapes in the region where it occurs. The frozen soils develop in conjunction with the overall complex of natural conditions in the area. The presence of permafrost close to the surface and the cryogenetic processes caused by it--the formation of cracks and the growth of venous ground ice, thermokarst, solifluction, and heaving--have a decisive effect on the formation and shaping of the new meso- and microrelief. The change in the relief entails a change in the conditions of drainage and snow drifting and thereby creates new habitats for the plants. The new pattern of development of the vegetative cover leads to changes in the soil and upper layer of the ground (the formation and accumulation of peat, etc.), the microclimate, the temperature of the upper layers of the permafrost, and the depths at which they become thawed.[8]

Thus, by influencing primarily the formation of the relief, permafrost has an effect on the development of the landscape as a whole and on the isolation within it of smaller, natural territorial units: the terrain, natural landmarks, and facies.[6]

A basic principle in the distinguishing of landscapes and their morphological components must be the cryogenetic approach. In the identification of a landscape, the following must be taken into account:

1. the genetic type of permafrost;
2. the type of ground ice;
3. the prevailing cryogenetic processes and relief; and
4. the latitudinal and geographic zoning determining the character and intensity of contemporary processes, as expressed by the heat-exchange level and the vegetative and snow cover.

In the course of small-scale mapping of tundra plains,[4] we distinguished 12 types of landscapes. Several of the most typical and widely ranging types are:

1. lake-swamp, thermally leveled with recent thin ice-wedge polygons and polygonal relief, low and flat;
2. marine, with buried thin ice-wedge polygons and polygonal relief, low and very slightly dissected;
3. ancient marine, reworked by thermokarst and thermal erosion, with buried injection and epigenetic ice-wedge polygons; and
4. ancient alluvial, thermokarst, and lake-alas, with thick buried ice-wedge polygons deeply dissected.

Correlating permafrost and the landscape is a very important task, since it opens the way to the use of a very economical and reliable method,

TABLE 1 Interrelationship between Some Natural-Territorial Complexes and the Permafrost in the Northern Taiga of the Northern Yenisei[a]

Landscape	Terrain	Natural Features	Permafrost: Content (% by volume of soil) and Character of the Ice in the Upper Layer of the Ground	Permafrost: Depth of Occurrence, m	Permafrost: Temperature (°C) at Depth of 8-10 m	Permafrost: Thickness, m	Permafrost Process
Lacustrine alluvial plain, sandy-silty to clayey-silty, thermokarst with migration, and injection type ground ice	Hummocky peatlands on the interfluves and terraces	Peaty hummocks and flat peatlands covered by sparse stands, shrubs, mosses, and lichens in peaty soils. Snow not more than 50-60 cm thick	60-90%; thick (up to 20 cm) frequent interlayers; occasionally continuous ice cores	0.4-0.6	-1.0 -1.5	25-40	Heaving on the summits, frost-fissuring, thermokarst
		Flat marshes in depressions between hummocks, snow from 0.5 m to 2-3 m thick	Open and, at times, also closed taliks				Seasonal freezing and heaving, at times neogenesis of permafrost
		Lakes; snow up to 2-3 m thick	Open taliks				Thermokarst at times along the shores
	Gently crested interfluves	Gently sloping hillocks 10-15 m high, clayey-silty, and covered by sparse stands of trees growing in peaty-latent podsol and gleyed soils; snow 1.0-1.5 m thick	Up to 20-30 %; interlayers from 2-5 cm to 10 cm thick, thin veinlets and lenses	From 0.8-1.0 to 3-5 (in the lower parts of the lee slopes)	-0.1 -0.5	From 15-20 to 30	Thermokarst, solifluction
		Marshes in depressions and consisting of sedges, trefoil, and sphagnum; snow up to 2-3 m thick	Open, less frequently closed taliks				Seasonal freezing, less frequently seasonal heaving
		Variously sized thermokarst lakes; snow 1-3 cm thick	Open, more frequently closed taliks				Thermokarst

[a]Not all of the terrain and natural features noted by us are included here, but only some of the more typical formations. The facies are not given at all.

namely that of landscape, in the study of permafrost. The landscape method of studying permafrost was proposed and first used by V. F. Tumel'[7] in a large-scale permafrost survey. He attached very great importance to the use of this method, since he maintained that it should form the basis for carrying out studies in the field of engineering geology.

The method was subsequently used by a number of geocryologists when studying the permafrost of various regions, as a result of which the interrelationships between certain types of permafrost and landscapes were revealed. The further elaboration of this method, however, was not deemed worthy of special attention.

Today, the landscape method is finding increasingly extensive applications in geology and engineering geology. Attempts are being made to use the landscape method in an engineering-geological survey of the northern regions, in which permafrost is one of the phenomena being studied.[5] For the time being, however, this attempt must be acknowledged to have been unsuccessful. In the first place, the specific nature of landscapes in the permafrost region has not been taken into account at all, added to which the relief and vegetation formed the basis for the distinguishing of the landscapes. Secondly, the relationships between the landscape indicators and the properties of the soils have not been ascertained or taken into account, and the indicators themselves are taken as the landscape. Thirdly, the experience of geocryologists and landscape specialists has not been taken into consideration.

Obviously, the landscape method must rest on a solid scientific basis. This basis must consist, first, of the skilled identification and comprehensive study of permafrost landscapes and their constituent parts, both on the ground and on the map. It is in this connection that the experience of landscape specialists should be considered.[6] Second, it must consist of a skilled study of the properties of the frozen soils that form the lithogenous basis of these landscapes. It is here that the experience of the geocryologists is pertinent. Third, there is the establishing of the interrelationship between the type of permafrost and the type of landscape or between their morphological components and the patterns of development of the various landscapes.

For example, the scientific approach will make possible the following:

1. Exposing the relationship between the permafrost and the general properties of the landscape and using the latter as indicators (such interrelationships were established in a number of areas of the permafrost region). In this connection, Table 1 shows the interrelationship with respect to the northern taiga landscape of the northern Yenisei.

2. Determining the role of permafrost in the shaping and evolution of landscapes, not only in the past, which helps in the reconstruction of the history of the region's evolution, but also in the present and future.

3. Establishing the relationship between the dynamics of the permafrost and the dynamics of the landscapes (and conversely) both in the natural setting and in areas undergoing economic development. A knowledge of these relationships will offer strong possibilities of predicting the pattern of evolution of an area in the course of its economic development.

4. By reference to the permafrost relief forms and permafrost phenomena, and in the light of their relationship with the character and properties of the permafrost and landscapes as a whole, it is possible to interpret them on aerial photographs. We have identified and published specific interpretation indicators (direct and indirect) of certain permaforst phenomena.[1-3]

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THE EFFECT OF PERMAFROST ON THE NORTHERN TREE LINE

V. V. KRYUCHKOV Moscow Lomonosov State University

Permafrost is one of the main factors determining the positioning of the northern tree line. In the Asiatic part of the USSR, a definite parallelism has been found to exist between the treeline and the summer thawing depths of 40-60 cm.

Natural terrain complexes consisting of suglinoks, heavy supesses, and clays are usually overgrown by a layer of moss. The subsequent growth of the mosses and the dying off of their lower part leads to the formation of a mossy-peaty layer, with the result that there is a decrease in summer thawing and a lowering of the soil temperature, an increase in its moisture content, and a worsening of aeration. This hampers the growth and development of trees and large shrubs and not infrequently causes them to die. We will consider these reactions with reference to four major regions.

The first region includes the extreme northeastern part of Asia, where the mountain pine *(Pinus pumila)* forms the limit of trees. In accordance with the Geobotanical Map of the USSR,[1] the stands of mountain pine in this region are classed with the forests, although the question of the particular type of vegetation to which they belong has not yet been conclusively resolved.[2] We shall not touch upon this matter here. The second region extends between the Pyasina River and the lower course of the Kolyma River, that is, the northern part of central and eastern Siberia. Here, the larch (*Larix dahurica*) forms the northern tree line. The third region--extending from the polar Urals and the lower course of the Ob' River to the Pyasina River--comprises the northern part of western Siberia. The Siberian larch *(Larix sibirica)* forms the northern tree line. The fourth region is northeastern Europe, where the Siberian spruce *(Picea obovata)* forms the tree line.

The harshness of the conditions near the tree line has resulted in there being no dense woody undergrowths. This is quite natural, of course, since dense undergrowth would have accumulated a considerable amount of snow in winter, the prolonged thawing of which would have shortened the growing season. In summer the shading of the soil is likewise not conducive to its warming.

In northeastern Asia (the first region), with increasing proximity to the northern limit of trees, the effect of the drop in the temperature and the decrease in the thickness of the layer of seasonal thawing is such that the stands of mountain pine become increasingly more sparse, the distances between them increase, and the trees themselves become smaller. At the tree line the mountain pine appears as isolated clumps. In the central, more shaded part of a clump, there is usually considerable moss, and it is in this part that the fallen needles are largely concentrated. The accumulating layer of organic matter hinders the warming and thawing out of the soil.

In the central part of the clump the thickness of the layer of seasonal thawing decreases to 20-25 cm, and there is a sharp drop in the soil temperature.[2] It becomes impossible for the roots of the mountain pine to exist here. The branches of the tree that are capable of sending out accessory roots grow in the direction of the warm area to which heat is being supplied. In the central part of the clump, the roots and the lower regions of the plants die off eventually, while the branches that have taken root and assumed the role of providing for their own root nourishment become independent plants, no longer connected with one another.[2] It is in this way that a clump undergoes degradation, by disintegrating into separate branches. The central part of the former clump, having been deprived of its plants, that is of a definite thermal resistance, is more receptive to heat transmission, whereupon the thickness of the layer of seasonal thawing becomes greater. In time, therefore, a clump of mountain pine will again establish itself.

Thus, the insular distribution of the mountain pine at its northern limit is dictated by the presence of permafrost and the relative shallowness of the summer melting of the soil (40-80 cm). On the other hand, these factors in the interaction with the vegetation are resulting in a constant shifting of the mountain pine near its northern limits. The role of permafrost in this process is usually underestimated.

In central and eastern Siberia (the second region) only *Larix dahurica* is present at the northern limits of arboreal growth. The accumulation of a mossy-peaty layer and the accompanying decrease in the thickness of summer melting, as well as the worsening of its aeration, promotes the formation of accessory roots stemming from the narrow region of the root. This takes place after the lower part of the trunk has become covered by a layer of moss, the lower region of the root being held solidly in position by the ascending frozen layer. When the buildup of moss is slow, the accessory roots will penetrate the overlapping mossy-peaty layer, thus forming a surface rooting system. Such a rooting system effectively fulfills a key role. When the buildup of moss and the ascent of the frozen layer is rapid, there will be no opportunity for a surface rooting and branching system to develop. In this event, a radishlike thick root will form, as well as numerous small lateral roots that will be unable to play a key role. At this time windfalls will occur. A fallen tree normally remains connected with the soil by a part of its roots. When this is the case, some of the outwardly directed branches will begin to grow vertically and assume the form of a young tree. In the accumulating mossy-peaty layer, the fallen trunk will die and decompose, while the branches that have become transformed into saplings will develop their own rooting system. The rapid buildup of the mossy-peaty layer, along with the described phenomena, will likewise lead to the lower branches lying on the surface of the ground taking root and to their conversion to independent saplings of vegetative origin (propagated by cuttings).

Thus, in both cases larch clumps will form as a result of vegetative propagation. Similar processes leading to the formation of clumps (from fallen trunks and by cuttings) are also observed when there is slow thermokarst subsidence in the soil surface. The "normal" northern limit of trees in this region, that is, the tree line that has not undergone pronounced anthropogenic modification, is usually clumplike and insular in character.

The disparate distribution of *Larix dahurica* at its northern limit in this region attests to the anthropogene influence. The onset of the formation of most clumps coincides with periods in which there is a somewhat heightened moisture content and cooling effect, leading to a rise in the permafrost table and to the growth of mosses, and also with periods of slow thermokarst subsidences of the surface.

In western Siberia (the third region), the processes leading to the buildup of the mossy-peaty layer are more intensive than in the two preceding regions. This is because the area in question has a more level topography (here, weakly dissected marine plains of Pleistocene age predominate) and the climate is less continental. On the other hand, the Siberian larch is much less capable of sending out accessory roots than is *Larix dahurica*. Accordingly, the variants described for the second region are possible here. Most often, it is another variant that is peculiar to interfluvial, flat watersheds and the upper parts of flat-crested hills that is observed. As a rule, the buildup of the mossy-peaty layer leads to a worsening of the hydrothermal and aeration properties of the soil. This hampers the growth and development of the Siberian larch and scrub alder. The seeds of these plants do not germinate when they meet up with such natural-terrain complexes. At a maximum melting depth not exceeding 20-30 cm, the trees and shrubs will begin to die off. In time, the dwarf birch will also die in these natural-terrain complexes. The snow held by the trees and shrubs will begin to be blown away in these areas, which will lead to the death of the mossy cover and degradation of the entire mossy-peaty layer and to the emergence of mottled ground devoid of plants.

Thus, in areas covered by thin stands of larch, a mottled tundra will form. Mottled tundras are characterized by a thicker layer of seasonal thawing, a higher soil temperature, and improved aeration. Here, seedlings of trees and shrubs will begin to appear. In time the area will again become covered by thin stands of larch--at the northern limit of trees a unique cyclical process involving the alternation of thin stands of larch with mottled tundras will be observed (see Figure 1). The main role in this process belongs to the interaction between the permafrost and the vegetation. These phenomena do not take place synchronously: In one natural territorial complex, it is possible for a thin stand of larch to be in the stage of dying off, even though another (situated within the same region) is becoming overgrown by mottled tundra consisting of larch and alder. A lack of understanding of this phenomenon on the part of certain geographers and botanists is leading to an abundance of contradictory and incompativle hypotheses, either that there is a northward circumpolar advance of the forests or, conversely, that there is, in general, a retreat toward the south.

The tundralike natural-terrain complexes that have been stripped as a result of successive changes in arboreal vegetation only become overgrown when larch seeds are present. Actually, these areas become overgrown mainly in the proximity of reforested river valleys and fruit-bearing islets of trees, i.e., not more than a few hundred meters away from the sources of the seeds. Over most of the northern part of western Siberia, the forested islets situated on the upper parts of smoothed watersheds and flat-crested hills (the outliers of marine terraces) are many hundreds of meters or even kilometers distant from the islets and sparse stands of larch. Accordingly, in the northern part of western Siberia, the trees that have died as a result of the plant successions (and human economic activity) are not regenerated, leaving behind them a collection of semirotted stumps and trunks. Consequently, in the unchanging climate of western Siberia, the tree line is actually retreating, and there is a decrease in the area occupied by the Siberian larch. The role of permafrost in this process is usually underestimated.

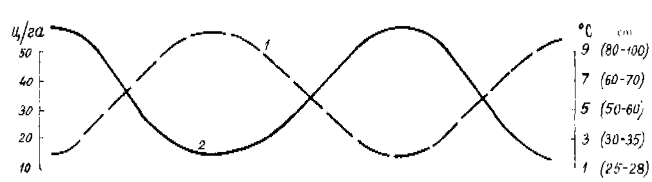

FIGURE 1 Schematic representation of the interrelationship between the phytomass and the soil temperature in some of the biogeocoenoses of the western Siberian tundra and northern forest-tundra. 1--phytomass (air-dried state) in which the dominant role belongs to the mosses, cm/ha; 2--soil temperature in July at a depth of 20 cm. It is directly dependent on the depth of summer thawing. The bracketed figures indicate this depth, cm; 3--time span. The cycle of development of the biogeoceonoses (the period between the dashes) is completed during the life of one or two generations of trees under conditions of flat topography and will evidently embrace a longer time span with rolling topography.

In northeastern Europe (the fourth region), as a result of the degradation of the permafrost (the increase in season melting, the rise in ground temperature, the appearance of taliks, and the increase in their number, (etc.), the arboreal vegetation is advancing slowly towards the north. Other conditions being equal, the pronounced summer melting of the permafrost and its deeper positioning relative to the surface of the soil is favoring an increase in the heating of the upper layers. Furthermore, the fact that the frozen ground is more deeply situated in summer means that there is at the same time a decrease in the level of the water-resisting ground and, consequently, somewhat better drainage of the upper layers of the soil. This frees a part of the heat, previously expended on the evaporation of surplus soil moisture, for heating of the soil. Thus, the degradation of the permafrost gives rise to a warming of the upper layers of the soil in the southern tundra and accordingly results in a northward shift of the arboreal vegetation. After becoming established at the new location, the woody vegetation holds the winter snow and protects the ground from severe freezing, which in turn, leads to an increase in the intensity of the permafrost degradation process.

This process, whereby ground that is subject to temperature variations interacts with the arboreal vegetation, is of interest in view of the general law noted by P. F. Shvetsov.[3] Essentially, what this amounts to is that in the course of warming trends the gain in the soil temperature and, consequently, the density of the heat flux in its upper layer increases from south to north during the same time period. Or, when there is an identical rise in the air temperature and the temperature of the soil surface, the increase in the thickness of the layer of seasonal thawing will be greater in the north than in the south. Such temperature increases result in the capture of new areas by the arboreal vegetation. In the course of subsequent slight decreases in temperature, the trees in the newly captured areas will not die off but will merely alter the pattern of growth (clumps, mountain pine). By retaining the snow, under the high temperature permafrost conditions obtaining in northeastern Europe, even if the arboreal-shrubby plants do not contribute to the degradation of the permafrost during a period of cooling, they will in any case hinder its aggradation right up to the time when a new warming trend sets in.

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SPECIAL FEATURES IN THE EMERGENCE OF CRYOLITHOGENESIS IN THE LACUSTRINE-ALLUVIAL DEPOSITS OF THE VALLEYS OF THE SMALL RIVERS IN CENTRAL YAKUTIA

T. P. KUZNETSOVA Moscow Lomonosov State University

If we examine a cryolithological section of the lacustrine-alluvial deposits of the valleys of small rivers as a whole, it will be seen to present a complex picture, caused by the variegated nature of the landscape conditions, which is typical of river valleys. Inasmuch as the first terraces above the floodplains of the rivers Khanchala and Kenkeme are very young relief forms (some areas of these terraces are either in the formative stage or in the stage where floodplain conditions predominate), they will include many floodplain relief forms--oxbow lakes, oxbow troughs, ridges, and interridge depressions that are still of recent origin. Many areas of the terraces (such as the "near-estuarine" areas of the drainage belts) are subject to swamp-formation.

Widely prevalent in the valleys of the small rivers are thermokarst lakes that develop at the site of oxbow lakes (these are inevitably transformed into thermokarst lakes, which often attain considerable dimensions, surpassing those of oxbow troughs). In expanding, these lakes gradually "eat up" areas containing relic ice-wedge polygons. As a result, oxbow and lake-swamp facies consisting of muddy-peaty, ice-rich, clayey-silty deposits occupy a fairly prominent place in the predominantly sandy composition of the deposits of the small rivers. The conditions that are either near bodies of water in the process of becoming overgrown or at the site of already overgrown oxbow lakes have always been favorable to the growth of ice-wedge polygons.

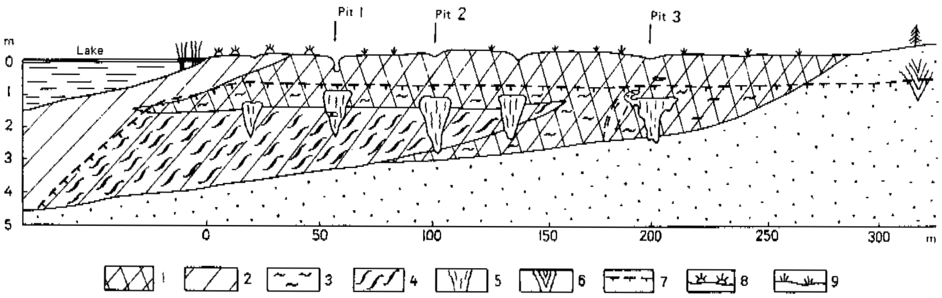

FIGURE 1 Cryolithological section of the lacustrine deposits. 1--brownish suglinoks; 2--greenish and dove-colored suglinoks; 3--thin "lamellae" of ice; 4--obliquely oriented large ice streaks; 5--ice wedges (exaggerated horizontal scale); 6--soil wedge; 7--permafrost table; 8--sedgy-hummocky meadow; 9--mixed herbage-grassy meadow.

We investigated the near-shore area of a thermokarst lake in the valley of the Khanchala River. The area is covered by a dry meadow consisting of mixed herbage and grasses. Polygonal relief, the shape of which approximates that of a flower bed, is quite clearly manifested on it. This is a former oxbow trough, a part of which is ccupied by the thermokarst lake. Today, there are no longer any morphological features of the oxbow trough at the surface. Figure 1 illustrates a cryolithological section traced from sections of several prospecting pits that were dug in the area in question. At the top and extending to a depth of 1.3-1.5 m there is a layer of brownish suglinoks, at times heavily peat-laden (often with inclusions of slightly decomposed peat). This layer is exceptionally dry, there being almost no ice inclusions visible in it. Only beginning at a depth of 0.9-1.0 m, that is, below the active layer, are some slender, sparse, and barely visible "lamellae" of ice to be seen; the moisture content is 15 percent. Occurring below 1.3-1.5 m is a layer of greenish suglinoks, which gradually become bluish-gray. Here, beginning at a depth of 1.5-1.6 m (the overlying part of the layer of greenish suglinoks is almost devoid of ice inclusions), the dove-colored and greenish suglinoks are seen to be exceedingly ice-rich. Usually they occur as irregular, broken, and very closely spaced ice lenses with thickness 0.2-0.5 cm, less frequently 1-2 cm. A well-defined oblique orientation of the ice lenses is to be seen; the moisture content is 50-80 percent.

The dissimilar coloration and varying pattern of the ice content in the suglinoks in the shore area of the lake attests to the presence of two facies varieties in the section: a subaqueous (lacustrine) variety, consisting of greenish and dark-colored, ice-rich suglinoks, with the obliquely oriented, broken lenses of ice that are typical of lacustrine (oxbow) deposits, and a subaerial (floodplain) variety, consisting of dry, brownish, peat-laden suglinoks that were formed under conditions closely similar to a floodplain regime of sediment accumulation (every year the areas adjacent to the shore are flooded with lake water in the spring).

In terms of its cryolithology, this area of lake shore is of great interest, and an analysis of its cryogenous textures makes it possible to describe in outline the conditions under which its constituent sediments became frozen. Evidently, the ice-rich lower layer of dove-colored suglinoks containing large, broken lenses of ice consists of the subaqueous sediments of an oxbow lake. These sediments had become frozen from the sides and from below. As the sediments accumulated and the lake regressed, this subaqueous syngenesis resulted in the dove-colored suglinoks becoming ice-rich and also in a predominantly slanting orientation of the ice streaks.* The upper, ice-poor layer of brownish suglinoks was formed above the subaqueous sediments under subaerial conditions. The accumulation regime of these sediments closely resembles floodplain conditions: Every year in the spring the shore areas of the lake are flooded with lake water. The accumulation of the deposits and their synchronous freezing can be regarded as a subaerial syngenesis. The deposits are characterized by a very low ice content, which is associated with the constant drying out of the active layer when the sediments are accumulating and becoming frozen. At the time when these deposits were formed, the active layer (as is the case today) was characterized by great dryness. The gradual transition of the lower part of the active layer into the perennially frozen state, which occurred as a result of the annual accretion of the sediments, led to a general dryness of the entire layer of subaerial deposits (floodplain facies). The drying out of the active layer was aided by an abrupt change in the landscape (surface) conditions: The wet sedgy-hummocky marsh, which is typical of the initial stage in lakes that are becoming overgrown, is eventually succeeded by a dry meadow consisting of mixed herbage and grasses.

Thus, the lake-alluvial (oxbow-thermokarst) deposits that are widely prevalent in the valleys of small rivers are characterized by a well-defined two-membered arrangement of the cryogenous structure of the section. The cause of this is the disparate emergence of cryolithogenesis.

The lithological composition of the deposits and the surface conditions of the area being discussed were conducive to intensive frost-cracking throughout the entire period in which the deposits were being formed, beginning with their drying out. Ice-wedge polygons were widely developed among the deposits (see Figure 1). In

*E. M. Katasonov. The cryogenous textures of perennially frozen alluvial deposits. Tr. Sev-Vost. Otd. Inst. Merzlotoved., No. 2, Yakutsk, 1960.

the area in question, three fairly thick ice wedges were uncovered by the prospecting pits. The thinnest wedge is located in Pit No. 1, which is situated in the hummocky-sedgy meadow, close to the lake (60 m). Its upper and lateral contacts are distinct, its lateral contacts are irregular, and the greater (lower) part of the wedge is situated in ice-rich dove-colored suglinoks. The ice is contaminated by mineral impurities, and there are inclusions of soil material in the form of fairly thick horizontal lenses.

The second wedge (Pit No. 2) is located at a distance of approximately 100 m from the lake (on a prolongation of the same frost crack) in the area of dry mixed herbage and grassland, where the polygonal relief takes on the appearance of a flower bed. The wedge is revealed at a depth of 1.15 m and has a smooth upper surface, its transverse thickness being 1.4 m. As in the preceding case, the larger lower part of the wedge is situated in ice-rich, dove-colored suglinoks.

The third ice wedge (Pit No. 3) is revealed in the same polygonal area (a unified system of ice wedges), though it stands apart from the wedges already described, being 200 m from the lake. The wedge is more than a meter wide and measures about 2 m along the vertical. In its morphological appearance, it differs markedly from the preceding wedges. The character of the enclosing deposits is also different, a most interesting feature of which is the almost total absence of ice inclusions.

The thickness of the wedge and its morphological features indicate that it developed quite actively and did not "lag behind" the ice wedges that have developed in a more favorable medium. This is evidently attributable to the fact that the wedge in question was situated in a unified polygonal system containing wedges whose conditions of development were more favorable and that it had maintained this developmental trend.

Apparently, the lower part of the ice wedges, being situated in subaqueous sediments, was formed epigenetically. Their subsequent development was patterned on the syngenetic type and occurred synchronously with the accumulation of the sediments.

One of the interesting characteristics of the type of section being discussed is the identical pattern of development of the ice-wedge polygons throughout the full extent of their formation, notwithstanding the markedly different ways in which cryolithogenesis affected the enclosing deposits. As noted earlier, the type of cryolithological section observed in this area is distinguished by an almost total absence of ice in the enclosing deposits and by the presence of relatively thick polygonal wedge ice. The unusual dryness of the soil materials in comparison with those in the remainder of the area, where the ice content is as much as 80 percent, is attributable to the fact that this area, being located relatively far from the lake, had not "survived" the subaqueous stage of development and comprised the shore zone of the lake. Lacking here is the sharply defined, two-membered arrangement of the cryogenous structure, which is typical of lacustrine-alluvial (oxbow-thermokarst) deposits that are prevalent in the valleys of small rivers and is caused by the disparate emergence of cryolithogenesis.

The general dryness of the enclosing soil materials did not interfere with the intensive growth of the ice-wedge polygons, probably because the area in question was at the stage when it was subject to floodplain conditions.

It should be noted that the active process of development of ice-wedge polygons in ice-poor (dry) soils is an unusual phenomenon, since, on the whole, confinement to definite lithological-facies conditions, distinguished primarily by a high content of segregated ice, is normal in the case of ice-wedge polygons.

MULTIWEDGE ICE IN THE UPPER REACHES OF THE TATTA RIVER

E. I. LEONKIN AND Z. I. TKACHENKO *Lengidroprokt*

The multiwedge ice in the upper reaches of the Tatta River was investigated by "Lengidroprokt" in connection with engineering geological studies pertaining to the contract design entitled "Systems for transferring the waters of the Amga River to the Tatta River," the object of which is to provide for the irrigation of the agricultural districts of central Yakutia.

In terms of its geomorphology, the area being studied is situated within an ancient lacustrine-alluvial plain, corresponding to the fifth terrace above the floodplain of the Lena River--the Abalakhskaya terrace--and characterized by alas topography. Distinguished in the area being discussed are alaslike depressions, outliers of the interalas region and the valley of the Tatta, the latter having reformed the alases inherited by it.

The geological structure of the area is characterized by the extensive development of a thick series of lacustrine-alluvial, preeminently clayey-silty deposits (middle and upper Quaternary), resting on an eroded surface of Middle Jurassic sandstones. In the alaslike depressions and the alas valley of the Tatta River, they are overlain by a relatively thin series of upper Quaternary and contemporary alas deposits.

The area is situated in the zone of continuous permafrost and has a disparate cryogenous structure. The permafrost, engineering, and geological conditions of the area are complicated by the fact that the texture-forming ice that is present in the Quaternary deposits is augmented by thick, variously aged multiwedge ice, developed in the form of polygonal nets of various sizes.

The most important areas in which it was necessary to provide a quantitative and qualitative estimate of the multiwedge ice were the floodplain of the Tatta River (the base of a dam) and its left bank, which is confined to the interalas outlier (the area adjoining the dam). The study of the ice was hampered by the haphazard appearance of its polygonal net at the surface, by the presence of soil wedges with parameters similar to those of the polygonal net, and by the absence of natural ice outcrops.

Of all of the prospecting holes that were dug (in order to substantiate the permafrost-geological survey), 60 percent of them revealed ice in the outliers of the interalas region, while, in the floodplain of the Tatta River, only a few of the holes revealed ice. This indicates that the ice in the Tatta floodplain is less thick and occurs in the form of a more sparse polygonal net than is the case in the outliers of the interalas region. The dimensions of the polygonal net of multivein ice in the Tatta floodplain were ascertained by means of electrical prospecting operations, while in the interalas outlier this was done in a test excavation.

In view of the fact that there is no single method that is suitable for solving this problem, the electrical prospecting operations were carried out in two stages. Initially, tests were conducted in order to select the procedure; these were followed by studies at preselected sites.

In the first stage a study was made of a geoelectrical profile and its variability, both horizontally and with depth. This was done by means of multiazimuthal and triple-electrode vertical soundings using several beams. As a result of these studies, the following information on the geoelectrical profile was obtained:

1. It was found that the region is characterized by a geoelectrical profile of type $K(\rho_1 < \rho_2 > \rho_3)$, in which the first layer correlates with the thawed surface suglinoks, the second corresponds to frozen loose sediments, and the third to bedrock.

2. The thickness and resistivity of the layer of seasonal thawing were determined (the thickness is 1-2 m, the resistivity is in the 80-100 Ω-m range).

3. The range of variations in the specific electrical resistivity of the series of frozen suglinoks was determined (2,500-5,000 ohm-m). The relative persistence of the ρ_2 parameter over a great distance (the resistivity of the second layer) and its comparatively small variation indicate that the ice veins have only a slight effect on this parameter, that is, that the ice veins are of limited thickness and width (presumably 0.5-1.5 m) in this area.

4. The circle diagrams indicated that no preferred direction of the vertical inhomogeneities was registered in the area. This confirmed that the ice veins are polygonal in structure.

The data derived from the study of the geoelectrical profile made possible a well-grounded approach to the question of the choice of the method to be used in the fieldwork and the processing of the resulting information.

Of all of the methods tested the following were found to be the most promising:

1. The median gradient method, which is characterized by a high degree of efficiency. The ρ_K curves of the gradient fix the surface position of a body of high resistivity more accurately than do other methods of electrical profiling. A disadvantage of the method is its sensitivity to surface inhomogeneities, which limits its use in areas with dissected relief.

2. In such areas the test results are given by combined profiling, as this is a more interference-free and efficient method. A disadvantage is the rather indistinct surface delineation of high conductivity anomalies on the ρ_K curves.

The experimental studies revealed that the AB separations are 65-130 m and that the magnitude of the MN separations is 2 m.

Also included among the procedural problems that were solved by the experimental studies is the choice of the network of investigations pertaining to the spacing of the observations, and the direction of the cross sections in relation to the objects being sought.

The choice of the spacing of the investigations was based on the results of the study of the geoelectrical profile noted above. The spacing adopted equals 1 m.

In view of the fact that the polygons have sides of varying length, often not exceeding 5 m, the distances between the profiles running across them should not exceed 2-3 m.

An important factor influencing the size of an anomaly is the angle of approach of the prospecting profile (and instrument separations) to the linear-shaped object being sought. It is known that the "permissible departure of the profiles from the direction running transversely to the orientation of a vertical wedge should not exceed 30°." The results indicated that only a 4-azimuthal system of observations satisfies the requirement that at all times ones of the prospecting profiles should pass at an angle of 90° ± 25° to veins of arbitrary direction.

In the second stage, electrical profiling studies were conducted at preselected sites, again in conformity with the method worked out. As a result, charts were drawn of the axes of the anomalous ρ_K (max), axes corresponding to the polygonal net of multiwedge ice. In compiling these charts initially, the strongest and most clearly defined anomalies, the orientation of whose axes was close to the perpendicular in relation to the prospecting line, were delineated and correlated on the charts of the ρ_K curves. These axes were entered on the area plan of the studies. Later, confirmation of the axis delineated by less distinct anomalies, identified by profiles intersecting the body at an angle of 90° ± 25°, was sought from the profiles intersecting the area in other azimuths. In this way, the position of the electrical axis on the plan was ascertained. It was found that the polygons are quadrangular to multiangular in shape; the diameters of the polygons range from 3-5 m to 10-12 m; the width of the ice wedges in the upper part is between 0.9 m and 1.2 m; and the volume of ice in proportion to the volume of soil material is 15-20 percent, in some areas up to 25 percent. The electrical profiling data were corroborated by drilling and excavation.

The dimensions of the polygonal net of ice in the interalas outlier were ascertained by us in a "test excavation," where a bulldozer was used to remove the soil material until the upper parts of the ice wedges were revealed. At places where the polygons were not clearly visible, holes were drilled in order that the cross section of the ice wedge could be precisely defined. As a result of these operations, a map of the multiwedge ice polygons was compiled. The shape of the polygons is mainly quadrangular with smoothed angles, the diameter of the polygons ranging from 2-3 m to 5-6 m; the width of the ice wedges in the upper part is from 2-3 m to 6 m (at their points of intersection). The volume of ice is about 60 percent of the volume of soil material.

It is important to note that the operations in the "test excavation" are time-consuming and that their execution is complicated by the thawing of the ice and soil. They therefore extend over a long period (a year) and are costly in comparison with the electrical prospecting studies. These latter, together with the experimental and verificatory drilling and excavation, are much cheaper. Furthermore, by means of electrical profiling it is possible to cover larger areas without disturbing the natural permafrost conditions. When used in conjunction with other methods, in particular seismic prospecting and drilling and excavation, it is capable of solving the three-dimensional ice-saturation problem.

The exposing of a polygonal net of multiwedge ice in plan made it possible to carry out to better effect the drilling and excavation by means of which the vertical thickness and shape of the ice wedges and their physical properties were ascertained.

The vertical height of the ice wedges in the deposits of the interalas outliers is up to 14 m; often, ice was found to be present in two stages, and the lower limit of its occurrence was found to be at a depth of 20-25 m from the surface. Its upper limit is between 1.1 m and 6 m, with depths of 1.1-1.5 m predominating. The ice is not distinctly wedge-shaped in appearance.

The vertical height of the ice wedges in the floodplain deposits of the Tatta River does not exceed 4 m and as a rule is 2-3 m. The depth of occurrence from the surface is 0.7-1.1 m. The ice is clearly wedge-shaped, the width of the wedge in the upper part being 0.9-1.2 m and in the lower part 0.1-0.2 m. Occasionally, a wedge formed of several elemental veinlets is clearly visible.

As a rule, the ice is crystalline, dense, both clear and cloudy, and less frequently snowlike. Ice of this latter variety is often found in the floodplain of the Tatta River and throughout the entire thickness of the vein, whereas in the outliers it occurs only in the form of lenses in the upper part of the vein. Furthermore, the latter is characterized by a cracked appearance, the walls of the cracks being tinted by ferric oxide, which is indicative of their greater age. The ice exhibits a vertically layered texture, caused by the thin interlayers of suglinoks and also air bubbles, frequently arranged in the form of vertical chainlets. It should be noted that the air bubbles in the ice in the Tatta floodplain are more elongated and tapering in shape, while those of the interalas outliers are more oval, or even rounded, which likewise attests to their having existed for a longer period.

The foregoing data pertaining to the parameters of a polygonal net of multiwedge ice, its shape, and physical properties, have made it possible to conclude that the ice in the floodplain of the Tatta River and the interalas outliers is noncoeval. This was one of the substantiating indicators of the noncoevality of the deposits and has made it possible to provide a more complete and accurate assessment of the permafrost engineering and geological conditions at the base of the dam and in the area immediately adjacent to it.

DETERMINATION OF THE THICKNESS OF PERMAFROST ON THE ARCTIC COAST

N. G. OBERMAN AND B. B. KAKUNOV The Vorkuta Integrated Expedition of Geological Exploration

The considerable thickness of the permafrost on the arctic coast frequently makes it necessary to extrapolate the depth of its lower boundary. Its assumed thickness depends (all other conditions being equal) on the value of the estimated geothermal gradient, which, as is well known,[6,18] varies with depth. This paper is concerned with the selection of depth intervals whose geothermal gradient would make it possible to determine reliably the lower boundary of the cryolithic zone.

The region in question has the following characteristics: The water has receded from the coastal belt relatively recently and that is why here there are widely distributed subterranean aquifers, which are waters at subzero temperature, that strongly affect the deep cooling of the upper zone of the earth's crust.[9,11,17] The second characteristic is that the region is washed by the Arctic Ocean, whose aquifers have a temperature of up to -1°C and an average thickness of 3 km,[10] which must lower the temperature of the rocks and the values of the geothermal gradients.

When generalizing geothermal data pertaining to large areas with dissimilar geological structures, it is useful to compare the gradients for equal intervals without reference to the geological structures, it is useful to compare the gradients for equal intervals without reference to the geological characteristic of the sections of the boreholes.[6] This is all the more justified for this region, where the effect of geological structure on the formation of geothermal gradients is less significant owing to the presence of frozen strata.

To compare the geothermal gradients of the cryolithic zone (see Table 1), we selected in each region of the arctic coast (see Figure 1) 1-2 boreholes from which the most reliable data were obtained, and we also selected mineworkings with the oldest shafts before taking their thermograms with delayed mercury thermometers. For comparative purposes Table 1 also shows the geothermal data for several continental regions containing permafrost rocks.

We shall divide the cryolithic zone of this region into three intervals according to the formation of the geothermal gradients: 0-30, 30-250, and 250-1,000 m. The 0-30-m interval (subsequently) this interval will not be discussed) was accepted on the basis of the ultimate thickness of the layer of annual thermal cycles in the region. The temperature field of the rocks to a depth of 250 m is due mainly to the influence of the radiation balance on the surface of the earth.[16] Allowing for this and for the depth at which the geothermal gradient changes from negative to positive (at the coast this depth is usually from 100 m to 250 m), the lower boundary of the second interval is assumed to be at a depth of 250 m. The boundary of the third interval is selected on the basis of the ultimate thickness of the cryolithic zone at the coast.

In Table 1 the gradients are also determined for comparative purposes by the conventional method[6,13,15], that is to say, for the entire cryolithic zone (the interval is 30-1,000 m).

An analysis of the geothermal gradients calculated for the selected intervals reveals some distributional relationships in the first vertical gradients, depending on the temperature of the rocks at the base of the layer of annual thermal cycles and on the mineralization of the subterranean waters.

The upper interval (30-250 m) has a negative, as well as a positive, geothermal gradient. The negative gradient is characteristic only of this

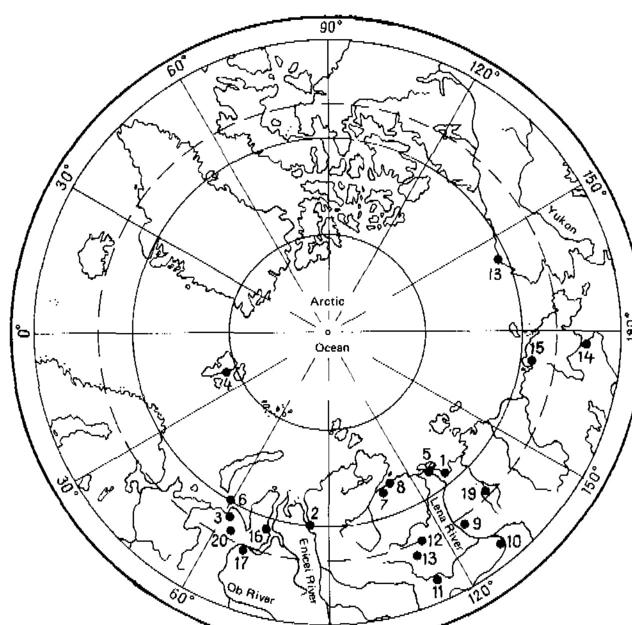

FIGURE 1 1--Tiksi Bay; 2--Ust' Port; 3--Bol'shezemel'skaya tundra; 4--Spitzbergen; 5--mouth of the Lena River, Chai-Tumus; 6--Amderma; 7--Kozhevnikov Bay; 8--Nordvik; 9--Bakhynai; 10--settlement Namtsy; 11--city of Mirny; 12--Markha River; 13--Alaska, Point Barrow; 14--Anadyr; 15--Pevek; 16--Gulf of Ob', Cape Kamennyi; 17--Gulf of Ob', city of Salekhard; 18--Yakutia, Udachnaya; 19--Verkhoyansk; 20--Vorkuta.

TABLE 1 Geothermal Gradients of the Cryolithic Zone of the Arctic Coast and of Continental Areas

Places of Observations	References	Thickness of Cryolithic Zone Based on Original Source	Depth (m) at Which Temperatures Were Taken	Depth Intervals (m): 30 -250		250-1,000	30-1,000	Rock Temperature at Depth of 30 m	Borehole Group	Mineralization of Aquifers, g/l
				Geothermal Gradients of the Cryolithic Zone (°C/100 m): Negative	Positive					
Rock Temperature at Base of Annual Thermal Conductivity and Geothermal Gradients										
Coast										
Nordvik (hole 8)	13	600(800?)	320	--	2.40	1.31	2.10	-11.7		~300
Kozhevnikov Bay (hole 6)	13	~600	503	--	4.09	0.40	2.13	-11.5		165
(hole 5)	13	~600	380	--	4.00	0.38	2.58	-11.0		~165
Tiksi Bay	3	640	500	--	1.50	1.88	1.71	-11.1	I	2
Mouth of the Lena River, Chai-Tumus (hole 5)	3	540-560	330	--	2.45	0.63	1.97	-10.8		No data
Alaska, Point Barrow (hole 3)	8	390	179	--	2.49	No data	2.49	- 9.0		Several g/l
Anadyr, mine	15	>150	77	--	5.41	Same	5.41	- 5.8		27-70
Pevek	16	230	348	--	2.86	--	2.86	- 5.2		<1
Average	13			--	3.2	0.9	--	- 9.5		
Amderma (hole 75)	13	400	274	-0.67	0.87	0.42	0.82	- 4.7		63-133
Ust' Port (hole 32-bis)	14	>400	400	--	1.00	0.40	0.76	- 3.2		45
Spitzbergen (hole 248)	12	~400	310	-0.94	1.20	0.75	1.04	- 3.0	II	34-44
(hole 13)	13	180	180	-1.00	2.00	--	2.00	- 2.4		34-44
Gulf of Ob' (hole K-1)	1	120	120	-1.50	1.25	No data	1.25	- 1.0	II	Brackish waters
Anderma (hole AD-2)	authors	900	773	-0.62	--	0.44	0.44	- 1.7		60-92
Ust' Port (hole 2GBG)	14	310	850	-1.08	0.75	1.17	0.85	- 0.6		9
Bol'shezemel'skaya tundra (hole 9)	authors	350	1060	-0.43	--	1.02	1.02	~-0.5		~10-20
Gulf of Ob' (hole 59)	1	250	245	-0.12	--	No data	--	- 0.3		1-18
Average				-0.8	1.2	0.7		- 2.0		

Continental areas										
Yakutia "tube" Udachnaya (hole 32)	9	~500-550	150	--	1.42	No data	1.42	- 7.7	I	15-150
Verkhoyansk	17	200-250	>250	--	4.0-5.0	--	4.0-5.0	- 7.0		<1
Yakutia, "tube" Udachnaya (hole 47)	9	~500-550	170	-1.00	0.62	No data	0.62	- 5.0		15-150
Bakhynai (hole 1-R)	7, 16	650	2780	-2.00	0.24	0.65	0.55	- 0.9	II	1-5
Vorkuta (hole 35-23)	authors	50	100	--	1.50	--	1.50	- 0.3		<1

Mineralization of Groundwaters and Geothermal Gradients

Coast										
Tiksi Bay	3	640	500	--	1.50	1.88	1.71	-11.1		2
Ust' Port (hole 2GBG)	14	310	850	-1.08	0.75	1.17	0.85	- 0.6	III	9
Bol'shezemel'skaya tundra (hole 9)	authors	350	1060	-0.43	--	1.02	1.02	- 0.5		~10-20
(hole 16)	authors	500	1000	--	0.85	1.16	1.05	No data		~10-20
Average				-0.8	1.0	1.3		- 4.1		up to 20
Spitzbergen (hole 248)	12	~400	310	-0.94	1.20	0.75	1.04	- 3.0		34-44
Mouth of the Lena River, Chai-Tumus	3	540-560	330	--	2.45	0.63	1.97	-10.8	IV	No data
Ust' Port (hole 32-bis)	14	>400	400	--	1.00	0.40	0.76	- 3.2		45
Amderma (hole AD-2) and	authors	900	773	-0.62	--	0.44	0.44	- 1.7		60-92
(hole 75)	13	400	274	-0.67	0.87	0.42	0.82	- 4.7		63-133
Kozhevnikov Bay (hole 6) and	13	~600	503	--	4.09	0.40	2.13	-11.5		165
(hole 5)	13	~600	380	--	4.00	0.38	2.58	-11.0		~165
Nordvik (hole 8)	13	600(800?)	320	--	2.40	1.31	2.10	-11.6		~300
Average				-0.7	2.3	0.6		- 7.2		>30
Continental areas										
Bakhynai (hole 1-P)	7, 16	650	2780	-2.00	0.24	0.65	0.55	- 0.9	III	1-5
Settlement Namtsy	4, 7	475	3000	--	0.47	0.65	0.53	- 2.5		1-5
Mirny (hole 3, MGU)	5	~600	520	--	0.41	0.20	0.29	- 1.8	IV	29
Markha River (hole 2)	9	1450	2010	0.00	--	0.14	0.14	No data		311

interval and of points with a temperature of -5°C at a depth of 30 m and above (second group of boreholes, see Table 1). The first group of boreholes is situated in areas to the east, and the second group in areas to the west of the Taymyr peninsula. The fixed negative geothermal gradient in the second group may be connected with the warming influence of the Gulf Stream or with a warming trend in the climate.

The value of the positive geothermal gradient in the upper interval is also determined by the temperature of the rocks. In the first group of boreholes, this gradient varies from 1.5-2.0 to 4-5 deg/100 m and more (average 3.2 deg/100 m, see Table 1); in the second group of boreholes, it varies from 0.0 to 1.5-2.0 deg/100 m (average 1.2 deg/100 m). One of the possible reasons for the drastic difference in gradient values is that the gradients in the cryolithic zone noticeably diminish with depth (below 15-70 mm).[18] Hence in the holes of the second group, the positive gradient (usually noted only in the lower part of the 30-250 m interval) has lower values than in the holes of the first group, where it remains fixed for the entire length of the interval. It is significant that in the 250-1,000-m interval, which usually does not have a negative gradient, the average values of the positive gradients of both groups of holes are very similar (see Table 1).

The relationship of the geothermal gradients to the temperature of the layer of annual thermal cycles has been noted in the upper interval of the cryolithic zone and in a number of continental areas, including areas in the United States and Canada (see Table 1).

The 250-1,000-m interval which is related to the diminishing effect in these depths of climatic zonation, does not reveal the above-mentioned distributional relationships. In this interval, compared to the preceding one, the geothermal gradient increases if the rock strata contain fresh and brackish subterranean waters, and it diminishes (2-4 times and more) if the strata contain very salty (above 30 g/liter) and briny aquifers. The gradient values, diminishing with the increasing mineralization of subterranean waters, amount to 1.0-2.0 deg/100 m in the first case and to 0.2-0.8 deg/100 m in the second (see Table 1); also in the latter case the cryolithic zone has the greatest thickness.

In some continental areas the lower section of the cryolithic zone reveals a similar relationship between the mineralization of the aquifers and the geothermal gradient, although the latter's absolute value differs from that at the coast (see Table 1).

The relationship between the geothermal gradient in the cryolithic zone and the mineralization of the subterranean waters is directly or indirectly recognized (without indicating the interval of the section) by many investigators;[6,9,17] this link may be due to the fact that at depths of 250-1,000 meters the thermal zonation of the rocks is increasingly, and in this region to a determining degree, affected by the regional hydrogeological characteristics and by the paleoclimatic zonation. The average ratio of the geothermal gradients of the third and fourth groups of boreholes of the 250-1,000 m interval amounts to 2.2 (see Table 1), and the gradient ratio of the anticlines and limbs of 57 American structures for the 30-900-m interval is 1.1 (data by Van Ostrand[6]). To ascribe this difference in the gradients to the influence of the tectonic effect is unjustified.

A comparison of the geothermal gradients of the 30-250, 250-1,000, and 30-1,000 meter intervals indicates that emphasis on the first two intervals gives a truer picture (the dependence of the gradient on the temperature of the layer of annual thermal cycles and on the mineralization of the subterranean waters is traced only in the first two cases).

In determining the thickness of the cryolithic zone, the use of the gradient of the 30-1,000-m interval is not justified, since it frequently gives distorted results owing to its understated value in cases involving fresh groundwaters, or, conversely, exaggerated results in cases involving mineralized (above 20-30 g/liter) groundwaters. The thickness of the cryolithic zone on the arctic coast can be determined most reliably with reference to the geothermal gradient of the 250-1,000-m interval; hence it should be utilized when extrapolating the position of the zero isotherm.

The foregoing suggests that the thicknesses of the cryolithic zone calculated by other investigators are too low for the following areas: Kozhevnikov Bay (borehole 6), Amderma (borehole 75), Chai-Tumus (borehole 5), Ust' Port (borehole 32-bis, 6, 10, and others). It is more probable that the zero isotherm at the above locations is found at depths of 900-1,000, 900, 1,000, and 500 m, respectively. It is significant that near Ust' Port the thickness of permafrost was recently established to a depth of 500 m.[4]

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MORPHOTECTONICS AND THE STUDY OF THE GEOTHERMAL CONDITIONS OF THE PERMAFROST ZONE

M. V. PIOTROVSKII Moscow Lomonosov State University

The term "morphostructure" refers to a tectonic structure that has been partly modeled by exogenetic processes but is wholly expressed in the relief. Both active and passive morphostructures are distinguished. The morphostructure of an area is its true structure. By its morphotectonics we mean the systems of morphostructures in their regular relationships, and their genesis and history. Morphotectonics is also the science of the relationships between tectonics and relief, also in the broad sense. The relief of the permafrost zones of Eurasia and North America is determined by the active morphotectonics of the Mesocenozoic cycle of activation. Mesocenozoic morphotectonics are intimately linked with the long-lived structural plan of the consolidated basement. Such a role is evidently played by the plan resulting from late Archean and early Proterozoic tectogenesis, which had reworked the early Archean plans but had incorporated a number of their structures: blocks, faults, and magma centers. They continued to be relatively independent and active, sometimes right up to the neotectonic stage. The morphotectonics of the permafrost zone are of the anticline-block type, that is, they comprise systems of outwardly smooth and more rugged steplike deformations. They occur as a result of vertical movements and distortions of basement blocks that are separated by faults and discontinuous dislocations of varying orders of magnitude (horizontal movements will not be considered here).

The processes envisaged are as follows: Recurrent planetary deformations give rise to the genesis and maintenance of relatively regular grids of dislocations and to movements of the blocks, which to a large extent are small and microamplitudinal ("amplitude free") and vibratory in character. The combination of grids of dislocations belonging to different systems and also of local dislocations results in the formation of a highly complex mosaic of the block-fracture tectonics. Further, the processes in the upper mantle and changes in the volumes of the abyssal masses cause vertical "high-amplitude" movements of the crustal mantle and crustal blocks occurring in complex combinations, away from the areas making up the faulted grid. Faults and discontinuous dislocations of lower orders of magnitude are delineated by a valley system. In mountainous relief and in outland areas, the direct participation of block dislocations in the evolution of the valleys is ascertained from the gradients of the slopes. The valleys also pass into distinct grabens, sometimes containing a thick layer of alluvium. Such a valley system points to a widespread reduction of faults, which can be ascribed to folding of the morphostructures.

The mechanisms leading to the formation of depressions and their energetics are in many respects not well understood. An analysis of the problem leads to the distinguishing of the two main types of activation, which are provisionally referred to here as rift and nonrift. Rift activation is the most intensive type. It is characterized by opposite movements of the various blocks, which entails the formation of steeply rising elevations and fully descending depressions, often filled with sediments that are hundreds or thousands of meters thick (the Baikal-Patoma region). This type is presumably associated with pronounced freeing of energy and local fusion of the upper mantle, resulting in the formation of an asthenospheric layer. In the nonrift type the depressions participate in uplifts that are only surpassed in elevation by mountains (the Aldan region). Both types are capable of replacing one another in the various stages of their development. Thus, although the Cis-Stanovoi downwarp of the Aldan region evolved in Jurassic times as a system of rifts, in the most recent stage its depressions have been participating in uplifts. The nonrift type is presumably associated with an escape of energy and a partial congealing of the asthenosphere in a previous stage of development.

The permafrost and hydrogeological conditions are linked with morphotectonics in the following way. Differentiation of the morphostructures, in reflecting the abyssal processes, should also reflect the differentiation of the abyssal heat flux. This problem has hardly been studied at all. The distinguishing and mapping of the various types of morphostructures can guide the geothermal observations and their analysis. It appears that regions of orogenesis are not characterized by a general rise in the abyssal heat flux, but by its differentiation: by an increase in uplifts and by a lessening in depressions.[3,4] Apparently, these are preeminently nonrift depressions. Those of the rift type, on the other hand, are characterized by a sharply increased heat flux. Nonrift depressions, however, are capable of retaining the heat flux differentiation features that are associated with weakened faults and are typical of the rift type.

Thus, in the Chul'man depression of the Cis-Stanovoi downwarp, the metamorphism of the coal-bearing Jurassic deposits reflects the differentiation of a heat flux that increased sharply during early Cretaceous tectogenesis (I. S. Bredikhin). Today, however, the areas of more pronounced metamorphism are characterized by a weakened development of permafrost. The junctions between the major blocks may be characterized by an increased heat flux. Such a junction exists between the Baikal-Patoma and Aldan regions along the Olekma River, where a seismic focus and hot springs (+50°) are located. Anticlinal-block morphotectonics and the resulting patterns of weakened dislocations in the surface layer of the crust are of the greatest importance in the circulation of fracture waters and the formation of a perennially frozen layer, taliks, and icings. Acting in conjunction with cryogenous processes, the tectonic jointing and its very recent weakening are giving rise to intensive denudation of the slopes, etc.

The phenomena described show up most distinctly in activated shields, where the relief is clearly linked with the block-fracture tectonics of the basement. On stratal plains and plateaus with a sedimentary mantle, discontinuous dislocations are much rarer and the deformations are greater. Here also, however, and even on low aggradational plains (for example, on the west Siberian plain) a relationship has been found to exist between the relief and loose mantle and the block relief of the buried basement. Here also the valley system often follows the faults, evidently because of the gently sloping depressions that have formed along them and the disintegration of the loose deposits.[1]

Morphotectonic mapping and analysis would seem to be of fundamental importance in permafrost-geological studies, especially in area surveys. The method that has been outlined rests on the experience gained from Soviet geomorphological and morphotectonic mapping.[5] A logical step is the compiling of systems of maps drawn to various scales: These range from detailed maps in the case of key areas to small-scale regional maps and interregional survey systems. The unity of the maps must be assured by adhering to a "rigid" style of depiction when representing the grids of faults, systems of blocks, and morphostructural zones.[2]

On such maps, many elements associated with the asthenosphere (the 120-180-km interval), the Mohorovicic Discontinuity (40 km), and centers of granitization (5-15 km) are clearly traced from the intervals between the grids, as well as the dimensions and types of the morphostructures. The analysis of the morphostructural profiles enables a rough estimate to be made of the overall volume of fracture tensions, while the maps indicate the sites where these tensions are localized and make it possible to select areas and positions in order that guided field studies can be conducted.

The patterns envisaged are common to the permafrost zones of Eurasia and North America. In particular, there is a close similarity between most of the morphotectonic features of the Baikal-Patoma and Aldan regions and the Canadian Shield. The conditions under which the morphtectonics are studied in these regions of Siberia are more favorable than on the Canadian Shield, where they are partly obscured by the effect of the thick glaciation and marine ingression. The patterns studied in Siberia may therefore be found helpful in clarifying the morphotectonics of the Canadian Shield.

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ZONALLY CAUSED GEOMORPHOLOGICAL CHARACTERISTICS OF THE CONTEMPORARY RIVER VALLEYS AND THE FLATLAND RIVERS OF THE EASTERN SUBARCTIC

G. E. ROZENBAUM Moscow Lomonosov State University

The presence of permafrost in the floodplains of rivers situated in the flatland areas of the eastern subarctic determines the specific (zonal) characteristics of their geomorphological structure.

The close interrelationship between the character of sediment accumulation, the cryogenous structure of the alluvium, and the cryogenous relief of the floodplain is dictated by the syngenetic freezing of the floodplain alluvium, together with the attendant syngenetic growth of ice-wedge polygons. The most strking geomorphological expression of this interrelationship is to be seen in the polygonal relief. The presence of polygonal patterns is peculiar to all of the relief components of the floodplain, although the character of the polygonal relief is exceedingly varied. This is associated with such factors as the facies conditions pertaining to the growth of the ice-wedge polygons in the different areas of the floodplain, the age differences between the latter, and whether the pattern of surface development of the individual areas of the floodplain is predominantly aggradational or degradational. The frost-cracking process itself predetermines the positioning of the polygonal relief. Frost-cracking takes place on the surfaces of riverbed shoals that are devoid of vegetation and are periodically drained of water, despite the presence in them of cracks resulting from dessication. But traces of frost-cracking on the denuded surface of riverbed shoals are rarely seen and consist merely of shallow grooves running perpendicular to the bank and corresponding to the direction of the surface drainage.

Widely developed in the young segments situated in the riparian zone of the floodplain are flat polygons that are gradually replaced by low-center polygons as the distance from the riverbed becomes greater. Against the background of this general trend, a marked diversity of the polygonal relief is to be seen in the various larger elements of the relief in the riparian part of the floodplain, on the crests of the floodplain, and in the depressions between the crests. The surface of the crests is broken into flat block-polygons, separated by interblock trenches, while in the swampy intercrestal depressions it is usual to encounter flat polygons with weakly developed ridges.

The low-center polygon relief attains its maximum development in the inner zone of the floodplain. Here, one sees evidence of the transition from a developing to a degradational type of polygonal relief. The degradation of the low-center polygon relief occurs in two ways. On the one hand, there is an infilling process, preeminently on account of the accumulation of peat in the intrapolygonal depressions, and also a planation of the surface of the polygons. Here, the low-center polygonal relief is transformed into the cryptopolygonal relief that is typical of the most elevated areas of the inner zone of the floodplain. On the other hand, in the relatively low-lying areas we find the development of lake thermokarst, by which we mean the expansion of the intrapolygonal bodies of water, together with their coalescence and the formation of secondary thermokarst bodies of water. The activation of the thermokarst process in the inner zone of the floodplain is associated with the marked decrease in its rates of sediment accumulation, which is due to its remoteness from the riverbed that constitutes the source of the material.

The development of thermokarst on a floodplain is determined, in the first place, by the thickness of the icy layer of floodplain alluvium containing the widest upper parts of the ice wedges, and, secondly, by the rate of sediment accumulation, which in turn is dependent on the height of the individual areas of the floodplain and on their remoteness from the active riverbed. The development of the thermokarst on the surface of the floodplain is in a state of dynamic equilibrium with the rate of sediment accumulation. A decrease in the latter, attendant upon a vertical growth of the floodplain and the floodplain massif becoming far removed from the riverbed, is accompanied by activation of thermokarst at the surface. Thermokarst, for its part, contributes to a lowering of the surface, and consequently, to the activation of sediment accumulation, the effect of which is to weaken and suspend the development of the thermokarst until such time as the regenerative growth of the floodplain leads to a repetition of the entire cycle.

Oxbow lakes are also subject to thermokarst reworking. When this happens, the extent to which the oxbow lake is transformed can be so great that it becomes completely impossible to guess what the original erosive shape was from the outlines of the oxbow-thermokarst lake. In contrast to shallow secondary thermokarst lakes, the floodplains of the oxbow-thermokarst bodies of water have shallow depths only in the near-shore regions, while at their centers there are areas with substantial depths that record the original erosive shape. The thermal erosional effect on the floodplain of the flowing surface waters is very great. This shows up in the transformation of the low-center polygonal and cryptopolygonal relief into the polygonal block type of relief in the more elevated and well-drained areas of the floodplain. This type of polygonal relief is expressed particularly clearly near floodplain ravines (the steep banks of the active riverbeds and oxbow lakes). Here, deep trenches separating the polygon-blocks form above the ice wedges. Areas of the floodplain that have been reworked in this way and are situated close to steep scarps that are in the process of being undermined are ripe for disintegration. In contradistinction to the strictly thermal erosive effects of the river stream on the riverbed, the thermal effect of the flowing waters of the surface of the floodplain has come to be known as the thermal denudation process.

Both of these processes, that is to way, thermal erosion and thermal denudation, participate in the deformation of the banks of rivers and oxbow lakes. Many investigators have noted the exceptional vigor that characterizes the effect produced by the thermal erosion of a fluvial stream on the surface migrations of riverbeds. It seems to us that single (annual) regressions of a bank under the influence of the thermal erosion process cannot be used when determining the perennial intensity of surface migrations of the bank at a particular site. Our views in this matter are based on *in situ* observations. Very often the base of a crumbling shore is armored by blocks of frozen ground that have already caved in. Thawing takes place more slowly than the process whereby they become drifted over with fresh sediment and a semisubmerged shoal is formed, which, following its expansion over a number of years, causes the deep channel to be displaced in the direction of the opposite shore and consequently, leads to temporary stabilization of a previously disintegrating shore.

It is typical of the floodplains of the rivers situated in the flatland regions of the eastern subarctic that the riparian zone of a floodplain is not greater than its inner zone. To a degree, this is caused by the weak development of the channel banks in the riparian zone of the floodplain. But the main cause of this phenomenon is that the lessening of aggradation in the inner zone of the floodplain, as compared with that in the riparian zone, is compensated by an intensive peat and ice buildup in an environment characterized by stagnant moisture content conditions, by the development throughout of a wet covering, and by a fine-grained mechanical composition of the sediments in the inner zone of the floodplain.

Another distinctively zonal peculiarity in the geomorphological structure of floodplains is the presence of expanses of superimposed floodplain. These form when the bottoms of alasses that had originated on higher and more ancient relief elements and have descended to the level of the floodplain merge with the surface of the floodplain. As a result of this merger, the rear junction of the floodplain takes on meandering outlines from which the shape of the original alas basins can be guessed and towards whose surface the waters of the freshet begin to percolate. This process is accompanied by the overlapping of the alas deposits by the floodplain alluvium.

PRINCIPAL LAWS GOVERNING THE DEVELOPMENT OF THERMOKARST IN WESTERN SIBERIA

YU. T. UVARKIN AND I. I. SHAMANOVA   Operations and Research Institute for Engineering Site Investigations (PNIIIS)

In the northern part of the west Siberian lowland, thermokarst is one of the main exogenetic processes. Although relief forms generated by it are present in abundance throughout the area, the prevalence of particular morphogenetic varieties of thermokarst forms, and the trend and stage of their development, are subject to definite zonal laws.

In terms of the conditions governing the development of thermokarst, the area comprising the lowland is subdivided into three zones: a southern (south of latitude 63°), a central (between 63° and 66°30'), and a northern (north of 66°30').

These zones are associated with the geocryological zoning of western Siberia[1] and are to a considerable degree determined by the latter. The use of geocryological and also general geographical (natural) zoning when making a study of the general laws governing the development of ancient and contemporary thermokarst is necessary in order that a qualitative evaluation can be made of the series of thermophysical factors determining the emergence of thermokarst processes (such as the level of insolation, the distribution of the snow cover, the modification of the plant associations, etc.).

In the course of field investigations conducted by the authors in the different geocryological zones of western Siberia during the period 1967-1971,[2-4] information was obtained which makes it possible to describe the manifestly zonal nature and potential for development of thermokarst within the confines of the main zones.

The northern zone is a zone of contemporary and ancient thermokarst formations situated for the most part in mineral deposits. This zone is characterized by the presence of continuous permafrost, containing an abundance of all of the known varieties of ground ice with which the development os subsidence processes is associated. Here, the emergence of thermokarst is determined by the conditions of heat exchange between the upper layers of the permafrost and the external environment. Despite the presence of the main condition--a high ice content of the ground--at the boundaries of this zone contemporary thermokarst is very unevenly developed.

The southern part of the zone, where the thermophysical prerequisites for the emergence of thermokarst are relatively favorable, is characterized by very great morphogenetic diversity of the subsidence relief forms and by an exceptionally intensive development of contemporary thermokarst of the insolation and hydrothermal types. Ancient formations are rarely encountered, as, following the thawing of the permafrost during the period of the middle Holocene thermal maximum, the ground again became frozen. This was accompanied by a reworking of the thermokarst forms by various exogenetic processes (including cryogenous heaving and secondary thermokarst).

In the northern part of the zone, the existing conditions of heat exchange at the surface are highly unfavorable for the emergence of contemporary thermokarst, because of the meager insolation, the relatively low mean annual temperature of the permafrost (to -9°C), and the small amount of precipitation, which falls preeminently during the frost-free season. In rare instances, thermokarst of the hydrothermal type is seen to be present to a limited degree. The principal landscape-shaping role is played by the ancient thermokarst relief forms. The larges of these originated during the period of the thermal maximum, the smallest during more recent episodes of climatic warming.

Within the northern zone, the potential for the initiation of very recent thermokarst processes becomes less in the south-north direction, because of the above-mentioned worsening of the thermophysical conditions for the development of thermokarst.

The central zone is a zone of contemporary, ancient and relic thermokarst formations situated in peatlands and mineral deposits. In contrast to the northern zone, here the thermophysical preconditions for the emergence of thermokarst processes are uniformly favorable: the high level of insolation coupled with the presence of a thick snow cover provide a capability for the intensive development of both the insolation and hydrothermal types of thermokarst.

The specific features that characterize the development of thermokarst in this zone, and its confinement to particular geomorphological levels, are determined exclusively by the conditions of distribution and occurrence of the permafrost and by its cryogenous structure and iciness.

The southern part of the central zone is preeminently characterized by relic (middle Holocene) and, to a lesser degree, contemporary subsidence relief forms. The latter are found within the peatlands in which the masses of thin permafrost are mainly localized.

Typical of the northern part of the zone, in which the area occupied by permafrost is much larger, are contemporary and ancient thermokarst forms situated in mineral deposits and peatlands. Relic subsidence formations are only rarely encountered. The age of the ancient forms is associated with the middle Holocene thermal maximum and also with the later episodes of climatic warming. Here, the morphogenetic diversity of the contemporary thermokarst relief forms is much greater than in the southern part of the zone, which is because of the highly diverse nature of the thawing ground ice.

Within the central zone the potential for the development of very recent thermokarst decreases from north to south, which arises from the fact that in this direction there is a marked contraction of the areas occupied by shallowly occurring permafrost and also from the decrease in their overall ice content.

The southern zone is a zone of relic thermokarst formations situated for the most part in peatlands. Here, the subsidence relief forms consist exclusively of relic formations of the lacustrine kettle hole and dry sinkhole types. Their origination was a consequence of the thawing through of the frozen peatlands and perhaps also of the melting of wedge ice during the course of the thermal maximum, processes that were not followed by restoration of the permafrost. Within the southern zone, there is every reason to suppose that seasonal subsidence processes could show up as a result of thawing out of pereletoks. The probability that these have originated cannot be ruled out.

The southern limit of this zone has not as yet been precisely established. Evidently, it may run close to latitude 58°-60°, where pseudomorphs replacing wedge ice have been noted.

Thus, in the south-north direction and extending to latitude 68°-69°, the intensiveness of development and morphogenetic diversity of the contemporary thermokarst relief forms occurring within the western Siberian lowland increase in accordance with regular laws. These characteristics are most prominent in the southern part of the northern zone, where all of the morphological and genetic varieties of the thermokarst formations known to exist in western Siberia are found. This increase can be ascribed to the existence of a south-north amelioration in the conditions that are necessary for the contemporary emergence of thermokarst: an enlargement of the area occupied by permafrost and an increase in its iciness. Simultaneously, a natural deterioration of the thermophysical preconditions for the perennial thawing of the permafrost is noted in this direction, which is a consequence of the worsening of the conditions of insolation and the gradual swing of the surface heat balance into the negative range of values.

The cryogenous structure of permafrost and its ice content prove to be the dominant factors right up to latitudes 68°-69°, where the thermophysical conditions are relatively unfavorable and the development of contemporary thermokarst is taking place, particularly intensively because of the widespread distribution of shallowly located ground ice.

The above-described zonal characteristics in the development of thermokarst are complicated by a number of regional factors, of which the following are deserving of special mention.

Geomorphological Structure

Within each zone a regular variation is observed in the conditions of distribution, the morphological characteristic, and the intensity of development of the thermokarst formation at the different geomorphological levels. This is associated with the age of the latter, the history of their development, and the geological structure. On the whole, the number of morphogenetic varieties and the age of the subsidence relief forms increase between the younger and the older geomorphological levels, reaching a maximum at the boundaries of the poorly drained surface of the lacustrine alluvial plain (the third marine terrace). Within the higher levels (the fourth and fifth terraces), the development of thermokarst decreases noticeably in consequence of their more pronounced ruggedness. Associted with the degree of surface drainage is a second-order natural law, in accordance with which a substantial lessening in the development of thermokarst processes takes place, when approaching the valleys of the major stream flows. It is this that accounts for the fact that the multilake type of landscape formed by the thermokarst processes that is so typical of the northern part of western Siberia is usually confined to vast, flat, swampy interfluves.

Most Recent Tectonic Movements

In recent years a number of investigators have noted that the intensiveness of development of thermokarst processes is dependent on the orientation and degree of differentiation of the most recent tectonics. It proved possible to determine this dependence by analyzing the results of a comparison between large-scale maps showing the extent to which the area is lake-studded and maps of the most recent tectonic structures. The analysis revealed that there is a regular increase in the extent to which the surface is lake-studded and that this occurs at the expense of the thermokarst bodies of water in a direction leading from the arched portions of positive structures toward the tectonic depressions and downwarps. In regions of submergence, conditions are set up that favor active swamping, with the result that drainage of the subsided water bodies is impeded. The swamping process is accompanied by a substantial change in the coefficient of thermal conductivity of the ground. As a result, perennial thawing of the ground, which is normally not typical of well-drained tracts, is frequently to be seen within vast, swampy areas.

On the whole, the foregoing azonal factors do not disturb the latitudinal zoning that characterizes the emergence of thermokarst processes in the area comprising the west Siberian lowland.

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PART III

Origin, Composition, and Structure of Permafrost and Ground Ice

A. I. POPOV, *Editor*

THE AGE OF THE PERMAFROST IN THE FAR NORTHEAST OF THE USSR

A. A. ARKHANGELOV AND A. V. SHER   Moscow Lomonosov State University

The evolution of the permafrost in the vast expanses of Eurasia is considered by most investigators to be associated with the middle Pleistocene, which is customarily regarded as a period of maximum glaciation (cooling) in which an extensive periglacial zone was formed. In the history of the animal kingdom, this stage is notable for the wide dispersion of mammoth ("periglacial") fauna, which included, along with the widely ranging boreal species of the open spaces (mammoths, rhinoceroses, horses, and bison), a number of subarctic species (lemmings, musk oxen, caribou, and arctic foxes), and also certain of the steppe animals (such as antelopes and gophers).

Many Soviet and foreign investigators acknowledge the existence of more ancient--early Pleistocene and even late Pleistocene--continental glaciations. They consequently accept the development of the periglacial zone and permafrost along the periphery of the ice caps, citing, in particular, traces of cryogenous processes in pre-Rissian deposits. Other scientists consider this viewpoint to be unsubstantiated. Without going into the details of this endless debate, we will merely note that during the cold phases of the early Pleistocene (the "Mindel"), the first representatives of a subarctic fauna (*Rangifer, Ovibos,* and possibly *Lemmus*) made their appearance in the more temperate latitudes of Eurasia. At present, however, there are no grounds for speaking of the existence at that time of a periglacial faunal complex.

The wide range of the cold-loving animals in the middle Pleistocene clearly indicates that their development occurred no later than the early Pleistocene. Hence, it inevitably follows that typical subarctic landscapes containing permafrost must have existed, in the high latitudes of Eurasia at least, by the beginning of the Pleistocene. Only in such a setting would the development have been possible of the indigenous representatives of the animal kingdom of the sub-Arctic, which are characterized by a unique complex of specific adaptations to the conditions of life imposed by the harsh climate of the high latitudes.

Recently, some indirect evidence has been obtained in support of this hypothesis. To begin with, there are indications that the cold arctic zones of the Earth were formed long before the beginning of the Quaternary. Thus, the advent of the antarctic ice sheets and the emergence of the first ice caps in the Arctic are dated to the late Miocene--early Pliocene. Authentic information is available on the Pliocene glaciations of Alaska and Iceland. It has been established that the ocean waters underwent pronounced cooling during the Pliocene. It is typical that all of these very recent studies are as a rule accompanied by findings relating to the absolute age of the deposits and to an analysis of the paleomagnetic, or the paleotemperature, regime. During the past few years, so many papers have been published on this theme that there is hardly any point in mentioning them here. References to the main lines of investigation in this research and to many specific studies are to be found in brief reviews published in the Soviet press.[1,13,17] The overall trend of these studies, conducted by various methods and in different countries, clearly indicates that the climatic cooling process in the high latitudes was in large measure accomplished in the Pliocene.

A second group of facts, forcing us to take a new look at the late Cenozoic history of the climate of the northeastern part of the USSR, is linked with the more precise definition of the age of the key floral assemblages, which until very recently were placed within the "Eopleistocene." Today, the age of these relatively heat-loving floras has been narrowed down to the Miocene-early Pliocene. Until very recently, even the youngest of the floras belonging to this group--the Enemten flora in western Kamchatka--was thought to be of early Pleistocene age. Today, it is placed within the early Pliocene.[16] The latest findings indicate that an analog of this flora--the Klamgalech flora of Alaska--is more than 5,700,000 years old.[19] At that time, western Alaska was the site of taiga-type birch-spruce forests admixed with pine and hemlock, and also certain species of plants that are known to exist today in southern Alaska and British Columbia. The northern limit of occurrence of these comparatively rich coniferous forests lay close to the Arctic Circle. Until quite recently, this was precisely the way that the early Pleistocene vegetation of the Soviet Northeast was envisaged. Today, however, pursuant to the updating of floras of this type to the beginning of the Pliocene, a hiatus has developed in the history of the landscapes of this region, the magnitude of which is several times greater than the universally accepted duration of the Pleistocene. There are grounds for assuming that this hiatus will gradually be filled up by much more cold-loving floras than were previously believed to have existed.

It was in this way that the objective preconditions arose that have made it possible to postulate that the permafrost in the high latitudes of Eurasia emerged as early as pre-Quaternary times. From the standpoint of obtaining information of this kind, the coastal plains of the Soviet Northeast, which since the Neogene have existed under continental climatic conditions and did not undergo major marine transgressions in the late Cenozoic, are considered highly promising. In particular, one might expect that some fairly old traces of the permafrost have been preserved in the thick sedimentary mantle of the Primorskaya (Maritime) lowland.

It is, however, impossible to find convincing evidence in the literature of the existence of either Pliocene or early Pleistocene permafrost in the Soviet Northeast. It is only in the middle Pleistocene that the development of the permafrost has been satisfactorily proven. Some authors even consider this to have been the era when persistent permafrost first appeared,[2,4] and several investigators have "rejuvenated" the permafrost right down to the late Pleistocene.[8]

A sizeable group of scientists, basing themselves both on theoretical considerations[11,14] and on firm geological data,[5,9,10,15] have even gone so far as to maintain that permafrost existed in the early Pleistocene. One of those who reached this conclusion is Yu. A. Lavrushin,[12] who uncovered buried ice wedges in the deposits of the Shanga Formation on the Indigirka River, which he dated to the early Pleistocene. O. A. Ivanov[7] postulates that permafrost first made its appearance in the early Pleistocene, based on the presence of pseudomorphs substituting for ice wedges in presumably early Pleistocene deposits of the Verkhneserkinskaya subformation of the Primorskaya lowland.

The data adduced by these authors serve as a basis for assuming that the permafrost in the Soviet Northeast is of early Pleistocene age. To obtain proof of this hypothesis, however, it is not sufficient to identify traces of the syngenetic development of permafrost phenomena in a relatively ancient group of strata. It is also necessary to have paleontological evidence of the age of the deposits, in the absence of which their dating to the early Pleistocene would not be sufficiently conclusive. Unfortunately, the named authors are not in possession of such evidence. As a rule, the ages of the strata were established either from their position in the section and their relation to other groups of strata, which were likewise not in all cases conclusively dated, or from spore-pollen assemblages.

Presented below is information on cryogenous disturbances noted in a group of strata, the early Pleistocene age of which is not to be doubted. This is the Olerskaya Formation, type sections of which are found in the Chukoch'ya River in the eastern part of the Kolyma lowland and, judging from recent data, are widely prevalent in coastal plains. It was in the deposits of the Olerskaya Formation that the initial discovery was made of the existence of a rich early Pleistocene mammalian fauna in the Soviet Northeast.[18] More than 1,000 skeletal remains of large mammals are known to have been recovered from these deposits, included among which are such typical early Pleistocene forms as the bullhead moose, giant horses predating the caballoid type, beavers, ancient musk oxen of a genus predating *Ovibos*, and a unique early Pleistocene ungulate, zorgelium. There are also numerous rodent remains belonging to ancient species. *None of these animal forms is known to have existed in Eurasia and North America later than the early Pleistocene.* The uniform state of preservation and identical evolutionary level of the Olerskaya mammals, coupled with the wholesale character of their burying and the finds of almost entire skeletons of early Pleistocene ungulates, indisputably attests to the synchronous origin of the fauna and the sediments.

The Olerskaya Formation, a lacustrine-alluvial group of strata with thickness of up to 15-20 m, consists of silts and fine-grained silty sands containing plant detritus, lenses, and intercalations of wave-built and autochthonous peat. The uniform lithological composition of the deposits makes facies analysis of the strata very difficult. Nevertheless, in a number of cases it has proved possible to establish the involvement in the structure of the strata of a number of facies, a set of which is typical of the river valleys of the subarctic lowlands. Most often, deposits of riverbed shoals are seen to be interstratified with floodplain sediments. The former consist of dingy grey, sometimes light gray, close-grained sands, at times silt-laden, with a horizontal lens-shaped bedding or with a fine cross-bedding manifested by a rippling course, which is sometimes registered by thin lenticules of plant detritus. Not infrequently, fine gravel consisting of consolidated silt is to be seen. The floodplain facies consist of silts with an indistinct horizontal bedding, intercalations and lenses of detritus, and lenses of peat, sometimes containing a large quantity of driftwood. The riverbed sediments proper consist of clearly defined bands of silt and sand with lensing and cross-bedding and of lenses of gravel, the grains of which are composed of silt that has been cemented by iron oxides.

More readily distinguished are the deposits of the facies of small lakes of the oxbow type. They consist of whitish aleuropelites with the typical fine horizontal bedding of the lake type. The bands of lacustrine sediments are normally crowned by lenses of autochthonous peat or by intercalations of peat-laden silt.

The only other facies of the deposits of the Olerskaya Formation--gleyed, fragmented silts containing a large number of ferruginized intercalations--is considered by us to be lake thermokarst deposits, the formation of which took place at the time when the oxbow lakes were transformed into thermokarst lakes. The shorelines of these lakes are often delineated by dipping intercalations of peat that are analogous to the peatlands found along the shores of contemporary thermokarst lakes.

All the above-listed facies are intricately combined in the sections of the Olerskaya Formation. Sometimes, the riverbed facies are seen to be rhythmically succeeded by floodplain sediments and then by oxbow lake sediments; in a number of cases, these cycles are repeated. Frequently, because of the exceptional homogeneity of the material, it is impossible to detect the transitions between the facies. Taken together, the modest thickness of the separate facies intercalations, their close succession along the horizontal and the vertical, and the rhythmical nature of this succession all indicate that the Olerskaya strata accumulated in the valley of a relatively small flatland river in which there was intensive lateral migration of the river bed.

Along the entire section of the Olerskaya Formation numerous wedge-shaped structures were observed, a study of which has revealed that they are pseudomorphs after ice wedges and furnish clear evidence that the sediments of the formation evolved under permafrost conditions. A number of characteristics of the wedge-shaped structures in the Olerskaya strata are suggestive of traces of melting ice wedges. Such diagnostic characteristics include: narrow, ferruginized "tails"; subsidence of the beds of enclosing earth materials and the existence of faults at the contacts; at the base of the structures the presence of lenses of material displaced from the walls; the infilling of some of the structures by concave lenses of peat, which demarcate the stages of the phase-by-phase filling of the thermokarst cavity; the presence of investing laminations; and the overlapping of the structures by lenses of autochthonous peat, registering an interruption in the sediment-accumulation process. These characteristics rule out the possibility of categorizing the formations in question as conventional structures or primary soil wedges.

Typical pseudomorphs substituting for ice wedges were observed in almost all of the outcrops of the Olerskaya Formation. A detailed description of some of them has already been published.[3] As the pseudomorphs are situated at various levels in the formation and are composed of many stages, the peaty intercalations overlapping one stage of the pseudomorphs are often broken up by the fissures of the level immediately above.

The pseudomorphs of one stage form a distinct polygonal network, which, in a number of cases, could be traced for several hundred meters. It is important to note, however, that no regional layers of pseudomorphs are distinguished in the Olerskaya Formation.

With the meandering of the riverbed along the floodplain, the floodplain sediments and ice wedges that had become deposited and frozen were overlain by riverbed or oxbow sediments, which evidently led to wedges melting and to formation of pseudomorphs. The latter could have occurred under both subaqueous and subaerial conditions, since pseudomorphs of both types are observed at one and the same level of the section of the Olerskaya Formation.

The dimensions of the ice wedges were evidently limited by the thickness of the floodplain sediments, which is in the 0.5 m to 3-4 m range. Under such conditions, both epigenetic and syngenetic ice wedges could have formed, with the latter, judging from some of the pseudomorphs, being of only moderate width. With closer interbedding of the floodplain and riverbed deposits, the ice wedges that were formed would have been predominantly of the epigenetic type and there would have been a superimposing of the network of one stage on another. The rhythmical structure of the Olerskaya alluvium reflects definite cycles in the embedding of the original, relatively thin, polygonal network and its subsequent thawing. On the whole, the process that led to the formation of the ice wedges took place uninterruptedly and synchronously with the deposition of the alluvium of the Olerskaya Formation. This is evident from the confinement of the pseudomorphs to definite facies cycles, regularly succeeding one another along the section. Inasmuch as numerous remains of large and small mammals have been found along the entire section, both in the deposits enclosing the pseudomorphs and in the sediments of the next cycle overlying them, there is no reason to doubt that an Olerskaya early Pleistocene fauna existed under conditions characterized by widespread permafrost.

The cryogenous texture of the deposits of the Olerskaya Formation being observed today is of secondary origin. At times, however, traces of a primary texture are seen to have been preserved in the strata in the form of netted or fragmented minute soil forms and also of the typical occurrence of ferruginized intercalations, repeating the outlines of the more ancient intrapolygonal depressions. These facts, taken in conjunction with the presence of pseudomorphs, likewise confirm the syngenetic freezing of the Olerskaya deposits.

Thus, we arrive at the conclusion that climatic conditions favoring the growth of ground ice existed without interruption throughout the period in which the deposits of the Olerskaya Formation were accumulating and subsequently thawed in consequence of changes in the environmental conditions. That the climate did not become appreciably warmer during the period in which the formation evolved is confirmed on the one hand by the absence of regional layers of pseudomorphs and on the other by the results of the detailed paleobotanical probing of the strata, which indicate that a harsh, subarctic climate predominated throughout this period.

There is reason to suppose that as the Olerskaya Formation accumulated, even thicker syngenetic ice wedges were formed. This could have occurred in both parts of the valley where conditions favoring the accretion of ice existed over a longer period by virtue of the higher stability of the facies conditions, for example, in the areas of the floodplain far removed from the riverbed. Owing to the deposits being more ice-rich, with their subsequent thawing out these wedges would not have formed such well-defined pseudomorphs as would the relatively thin wedges of the riverbed areas. Apparently, the massive, highly ferruginized silts, the roofs of which are several meters lower than those of the areas consisting of ice-poor riverbed facies, are a product of thermokarst reworking of the deposits of the intrafloodplain areas containing ice wedges.

The combination of traces of primary syngenetic and secondary epigenetic freezing makes it possible to state that the Olerskaya strata are characterized by permafrost that had originated polygenetically. The period in which the primary texture underwent destruction remains unclear. In all probability, it could have been linked with the episode of regional interruption in sediment accumulation, which is recorded by the water-eroded roof of the Olerskaya Formation.

The most important conclusion to be drawn from the foregoing discussion is that stable permafrost existed on the Kolyma lowland in the early Pleistocene. While this conclusion, as was indicated earlier, stems from a number of general considerations, the findings adduced are nevertheless of great importance since the Olerskaya Formation is, in essence, the first group of strata with clearly defined traces of cryogenesis, the early Pleistocene age of which is so conclusively proven by rich paleontological material. Also of great interest is our ability to visualize the harsh climatic conditions that existed at the time when the Olerskaya Formation was being built up, not only on the basis of cryogenous dislocations, but also on the overall aggregate of information contained in the sediments. The Olerskaya invertebrate fauna differs markedly from the coeval faunas of the more southerly regions in that there is an absence of southern forms and the leading role is played by such subarctic animals as lemmings (Common and Subarctic), caribou, and musk oxen. The widely ranging boreal animals of the open (treeless) spaces--horses, bisons, etc.--became a major component of this fauna. The presence of these forms, and also of certain steppe species, indicates that the climatic conditions under which the Olerskaya fauna existed were not only cold, but also fairly dry and markedly continental. This is in excellent agreement with the voluminous paleobotanical data that sketch a preeminently treeless vegetation, consisting of tundra and xerophytic grassy associations.[6] All these features bring together the Olerskaya biocoenosis and the periglacial biocoenoses of the second half of the Pleistocene, notwithstanding the fact that as a result of its much greater geological age it differs from them in the composition of the fauna and in certain other characteristics. Serving as additional evidence that the sediments were formed under conditions of permafrost is the excellent state of preservation of the organic matter, the specific features of the mineralogical composition of the deposits, and a number of other factors.[18] The sum total of the information on the Olerskaya Formation suggests that the conditions accompanying its formation were in the nature of a prototype of the periglacial situation that characterized the middle-late Pleistocene.

The Olerskaya Formation familiarizes us with the oldest of the cold-resisting biocoenoses. Its typical subarctic appearance, however, is so striking that we are fully justified in assuming that what is imprinted here is far from being the initial stage in the existence of a harsh climate in the high latitudes of Eurasia. For the development of the distinctive features of the cold-loving animals and plants, likewise of the biocoenosis as a whole, there must have been a sufficiently long period in which subarctic conditions predominated. The evolution of the subarctic forms to the stage at which we meet them in the early Pleistocene must have taken place in the Pliocene.

This hypothesis can rightfully be extended to the period in which the permafrost was formed. Indirect confirmation of this is to be seen in the presence of cryogenous dislocations in the sediments of the Begunovo Formation on the Kolyma, which is somewhat older than the Olerskaya Formation,[18] and also in the sediments enclosing the even more ancient fauna in Alaska.[20]

The foregoing data on the Neogene age of the cold arctic zones of the Earth are fully in agreement with this hypothesis, while the presence in the stratigraphic schemes of the Maritime lowlands of a hiatus embracing, in essence, the greater part of the Pliocene, permits us to hope that considerably older traces of the existence of permafrost will be found in the Pliocene deposits of this region.

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PATTERNS OF DISTRIBUTION AND A QUANTITATIVE ESTIMATE OF THE GROUND ICE IN THE USSR

B. I. VTYURIN Pacific Institute of Geography, Far Eastern Scientific Centre Academy of Sciences of the USSR

The geographical patterns of distribution of the various types of ground ice are a consequence of the characteristics of their formation. Primary interstitial texture-forming ice (segregation and, to a lesser extent, injection ice in loose soils, wedge ice in rocky soils) is widely distributed throughout the USSR (Figure 1), although it takes second place to the cementing ice of the granular soils. The latter is found in granular sediments, irrespective of their origin and composition, throughout the entire mass of the permafrost.

Segregation ice, constituting the bulk of the so-called visible ground ice[4] and forming a class of streaky cryogenous textures, is confined to fine-grained deposits of differing origin and occurs mostly in the upper 10-15-m layer of epigenetic permafrost and throughout the entire mass of syngenetic permafrost. In the USSR the latter does not normally form strata that are thicker than 30-40 m. Accordingly, only the uppermost layer of permafrost averaging 10-40 m in thickness has been taken into consideration in the discussion of the patterns of distribution and the quantitative estimate of the segregational texture-forming ice.

Despite the vagueness and highly sporadic nature of our understanding of permafrost, an analysis of the information on ground ice that has been amassed to date make it possible to distinguish regions of widely occurring segregation texture-forming ice. In the Eruopean part of the USSR, to a depth of 8-10 m on the average, the peatlands and subjacent silty and fine-grained lacustrine, lake-swamp, and glacial suglinoks of upper Quaternary and Holocene age are characterized by ice with a very high degree of streakiness. In the "barrens" region (the islands of Novaya Zemlya and Franz-Josef Land), segregation ice is found only in small areas containing fine-grained marine, glacial marine, and glacial sediments.

In the Urals, segregation texture-forming ice is confined to the valleyside and, in particular,

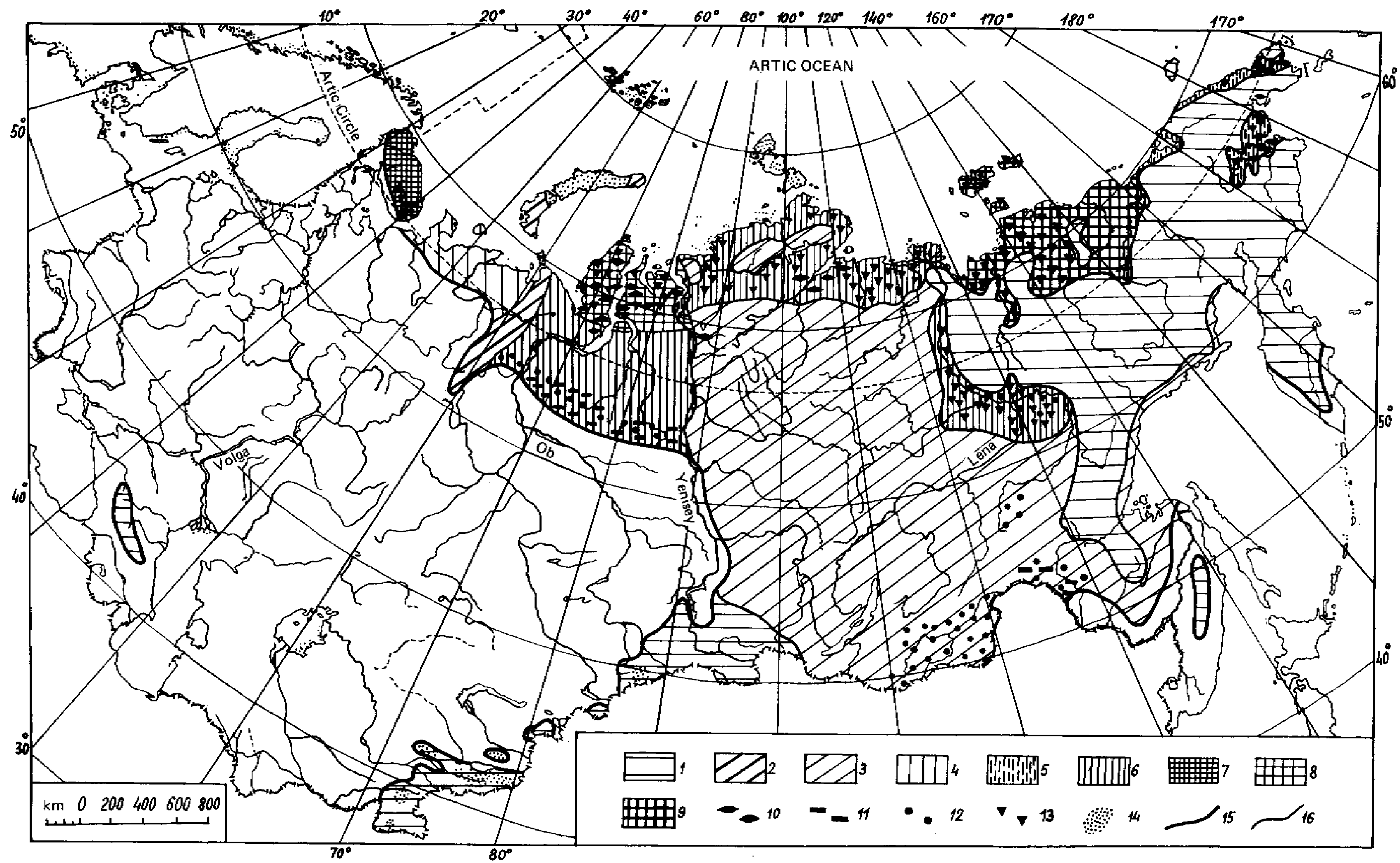

FIGURE 1 Sketch map of the total visible ice content of the permafrost in the USSR. 1--3 percent to a depth of 50 m; 2--10-20 percent to a depth of 5 m; 3--the same to a depth of 10 m; 4--20-30 percent to a depth of 10 m; 5--20-30 percent to a depth of 20 m; 6--30-40 percent to a depth of 10m; 7--40-50 percent to a depth of 10 m; 8--40-50 percent to a depth of 20 m; 9--40-50 percent to a depth of 30 m; 10--regions of thick (>3 m), deep (>3 m) stratal sills of primary interstitial ice; 11--regions of relatively thin (<3 m), shallow (<3 m) stratal sills of primary interstitial ice; 12--regions where frost mounds containing ice cores occur; 13--regions where thick sills of multiwedge ice occur; 14--contemporary glaciers; 15--limit of permafrost; 16--limits of areas with variable ice content.

to the lake-swamp deposits and peatlands of the flat swamp-covered watersheds and bottoms of depressions.

Western Siberia is the region where segregation texture-forming ice is most widespread. Here, fine-grained Quaternary deposits of marine, alluvial, lake-swamp, and colluvial-soliflunctional origin occur almost everywhere at the surface. Test borings have revealed the presence of segregation ice in a series of granular deposits extending to a depth of 130 m or more. In the main, however, streaky cryotextures are traced to a depth of 10 m in the southern and central parts of the lowland and to 20 m in the northern part.[3]

In central Siberia, the regions where segregation ice is most widespread are the Lena-Vilyui and north Siberian lowlands. These are regions where the thickness of the permafrost strata containing streaky cryotextures ranges within wide limits; on the average it is 10-20 m. The highly diversified complex of streaky cryotextures has been described by many investigators from the alluvial-lacustrine and valleyside deposits of the Lena-Vilyui lowland. These cryotextures are less widespread in the granular dsposits of the mountainous regions of central Siberia and were not noted at all in the islands of Severnaya Zemlya.

Segregation ice occurs only infrequently in the southern mountainous regions: in the Caucasus, the Pamirs, the Tien Shan, and the Sayansk Mountains. In the trans-Baikal it is confined to the floodplain and oxbow facies of alluvium, to lacustrine and valleyside fine-grained deposits, and to the peatlands in the flat and swampy areas of the river valleys and intermontane depressions.

In the Soviet Southeast, segregation ice is most widespread in the fine-grained Quaternary deposits of the Amur-Okhotsk region and particularly in areas where *maris* exist. There is virtually no information on the occurrence of ice of this type in Sikhote-Alin, Sakhalin, and Kamchatka.

In the Soviet Northeast, the highest content of segregation ice is in the alluvial, lacustrine, swamp, valleyside, and other deposits of the northern coastal plains and the islands of the Novosibirsk Archipelago. On the plains of the Chukchee Peninsula, moreover, segregation texture-forming ice is observed almost everywhere in marine and glacial marine deposits and less frequently in glacial deposits.

Wedge ice, forming streaky-cementing cryogenous textures, is noted in rocky, cracked soils and is also very widespread (see Figure 1). It is confined to the mountainous regions, where rocky soils outcrop at the surface or are covered by a thin layer of valleyside or granular residual formations. On the average, the zone characterized by intensive cracking and the formation of wedge ice is 30-70-m thick, although it ranges from several meters to 200-300 m. Nevertheless, as has been shown by the findings of a number of investigators (N. V. Gubkin, N. F. Brakhin, A. B. Chizhov *et al.*), it is only in the upper several tens of meters (50 m on the average) of the cracked rocky soils that there is an appreciable quantity of wedge ice.

Of the sill-forming primary interstitial ice, the most widely distributed varieties are segregation and injection ice, which give rise to the formation of large lens-shaped (mainly cores of frost mounds) or stratal-shaped ice bodies. A map indicating the distribution of frost mounds containing ice cores was compiled for the first time by the author and P. A. Shumskii in 1955. Later[7] it was augmented by information on ancient, deep (>3 m), thick (>3 m) stratal sills. Recently derived data on the distribution of stratal and lens-shaped sills of primary interstitial ice make it possible to discuss the following patterns of their distribution in the USSR.

In the arctic zone or barrens region, frost mounds are an exceedingly rare phenomenon. In the northern zone, such mounds are usually found in freezing taliks, while in the southern zone the mounds form when the ground first becomes frozen. Also widely distributed in this zone are frost mounds at the outlets of springs. The zone evidently attracts thin, shallowly situated strasills of which there is no surface indication or which appear as areas of heaving. There is no difference in principle between the mechanism of their formation and that of frost mounds (predominantly segregation ice).

Deeper and thicker stratal sills are currently known to exist only in the northern part of the permafrost zone. Their genesis is associated with the composition and hydrogeological characteristics of the epigenetically frozen deposits. They are confined to marine sediments in western Siberia, to sediments of marine, glacial, or other origin in the northern part of central Siberia, and to marine and glacial-marine deposits in Chukotka.

Secondary interstitial ice is for the most part a sill-forming variety. It is very widespread and forms thicker sills of multiwedge ice. The general patterns of distribution of the multiwedge ice in the USSR are indicated in a map which we compiled jointly with Shumskii in 1955[7] and in maps produced by Baranov[1] and Popov.[6] During the past 15 yr there have been many contributions to our knowledge of the distribution of multiwedge ice. These enable us to define more precisely the limits of its geographic range and also to arrive at an approximate estimate of the reserves of it in the USSR (~100 km^3).[4] The latter became possible because the multiwedge ice, owing to the regular pattern of the polygonal network formed by it and the clear manifestation of the latter at the surface (wedge-polygon relief), enabled us to work out a procedure for determing the three-dimensional and two-dimensional macro ice content of the permafrost.[5] The distribution of the permafrost in the USSR, together with approximate indices of the three-dimensional macro ice content as derived from the multiwedge ice, is illustrated in Figure 2.

Of the other types of secondary interstitial ice, mention can be made of cave ice. This is azonal in its distribution and is found within the permafrost zone and outside it (e.g., the Carpathians, the Crimea, the Caucasus, and the Urals).

Primary surficial buried ice (that of the snow, glacier, marine, lacustrine, river, icing, and interstitial types) is limited in its distribution

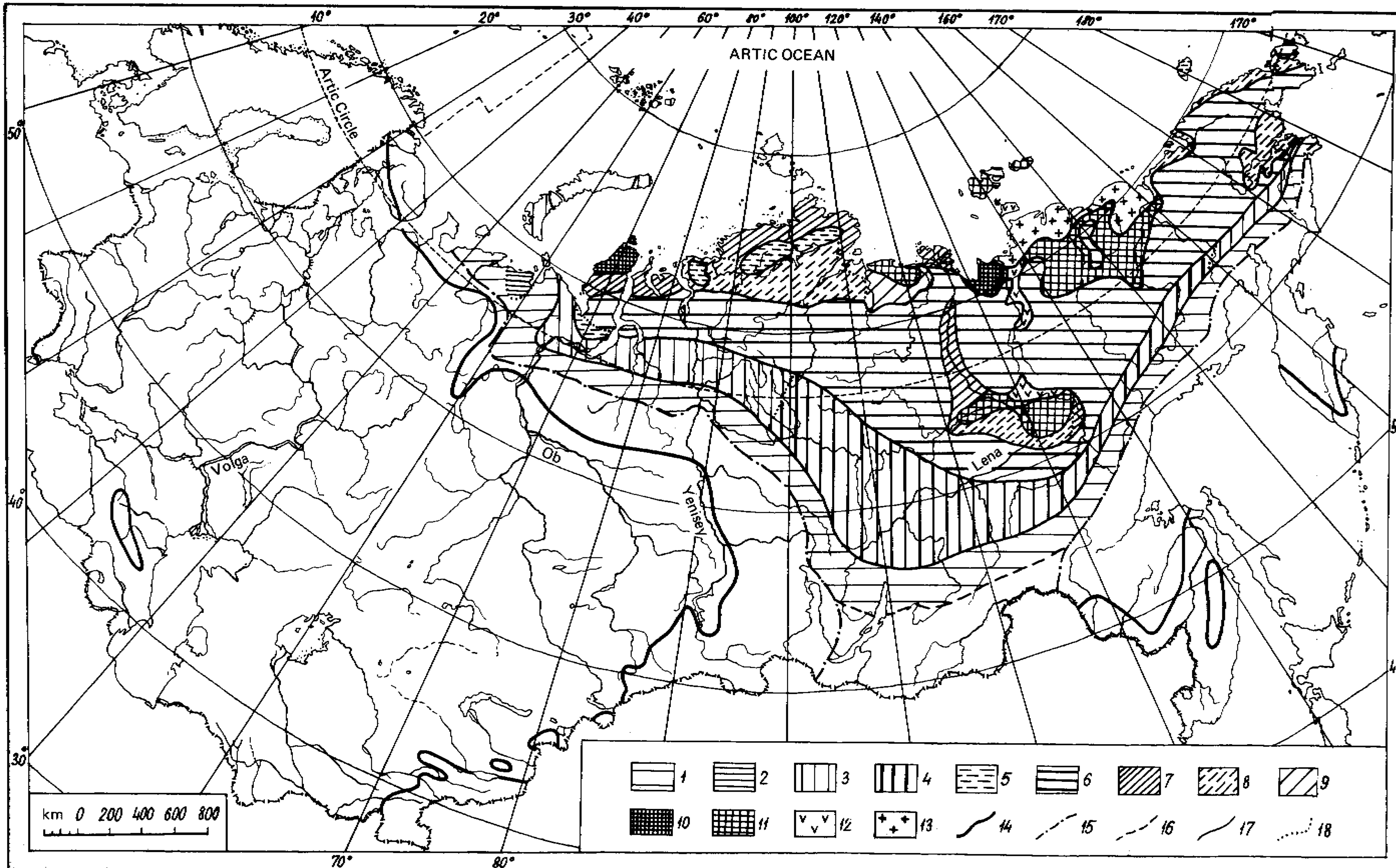

FIGURE 2 Sketch map of three-dimensional macro ice content of the permafrost, as derived from the multiwedge ice in the USSR. 1--up to 3 percent to a depth of 1.5 m; 2--3 percent to a depth of 2.5 m; 3--3.5 percent to a depth of 1.5-2 m; 4--3.5 percent to a depth of 2.5-5 m; 5--5-10 percent to a depth of 3 m; 6--5-10 percent to a depth of 3-5 m; 7--10-15 percent to a depth of 5 m; 8--10-15 percent to a depth of 5-10 m; 9--10-15 percent to a depth of 10-15 m; 10--15-20 percent to a depth of 10 m; 11--15-20 percent to a depth of 10-20 m; 12--20-30 percent to a depth of 10-20 m; 13--20-30 percent to a depth of 20-30 m or more; 14--limit of permafrost; 15--southern limit of multiwedge ice; 16--the same, as found in the trans-Baikal; 17--boundary line of regions with differing macro ice content; 18--boundary line of areas containing ice wedges that penetrate to different depths within one region.

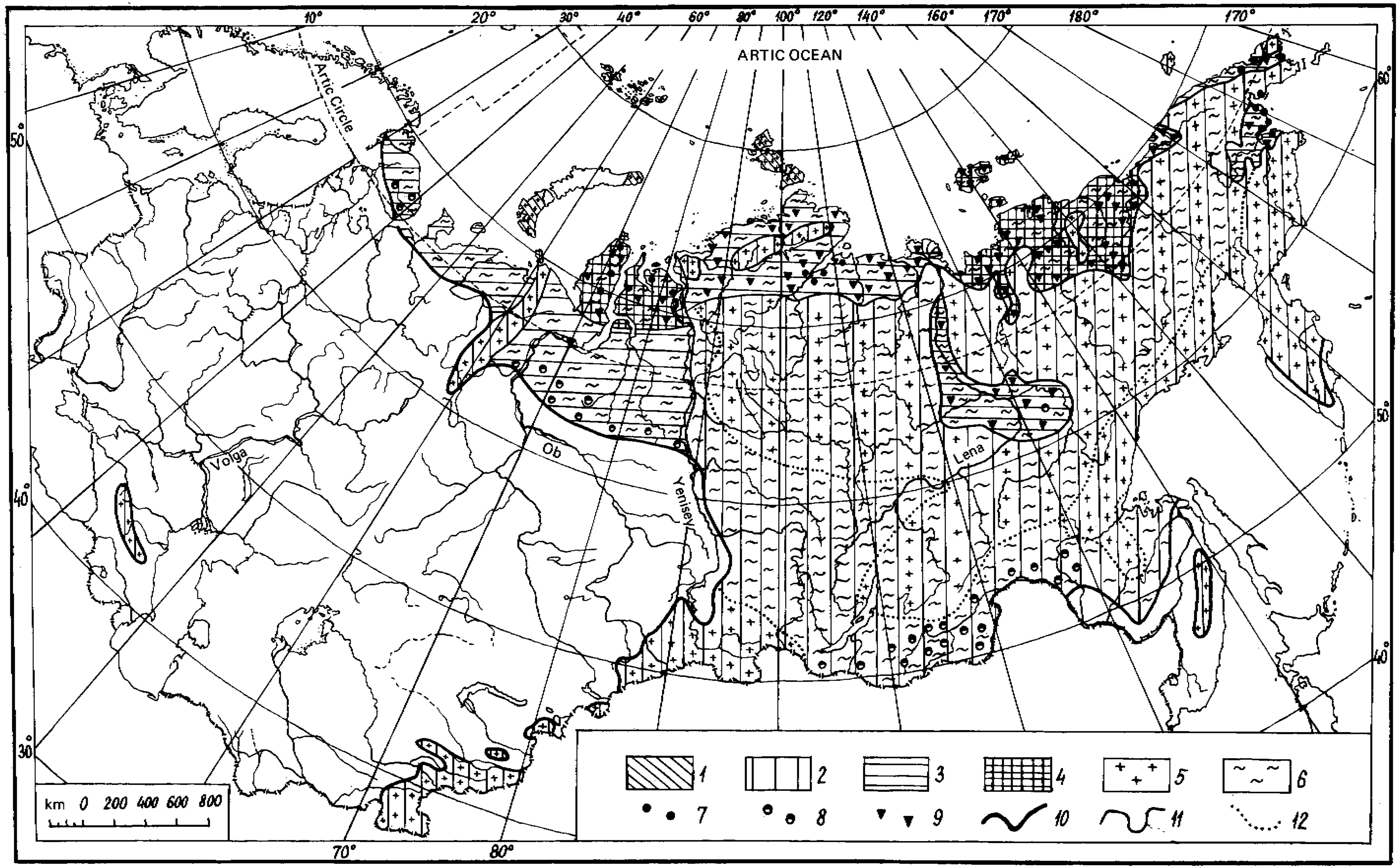

FIGURE 3 Sketch map indicating total amounts of visible ice in the upper layers of the permafrost (5-30 m in unconsolidated material and 50 m in rocky material) in the USSR. 1--<1,000,000 m^3/km^2; 2--1,000,000-2,000,000 m^3/km^2; 3--2,000,000-5,000,000 m^3/km^2; 4--5,000,000-10,000,000 m^3/km^2 or more; 5--mainly accounted for by texture-forming wedge ice in rocky soils; 6--mainly accounted for by texture-forming segregation ice in loose soils; 7--in which large quantities of deep stratal sills of primary interstitial ice of great thickness are present; 8--in which large quantities of thin, shallow stratal sills of primary interstitial ice are present; 9--in which large quantities of sills and multiwedge ice are present; 10--limit of permafrost; 11--boundary line of region with varying reserves of ground ice; 12--boundary line of areas containing various types of ground ice.

TABLE 1 A Quantitative Estimate of the Visible Ground Ice Contained in the Shallow Permafrost of the USSR

Region	Area Occupied by Permafrost, km^2	Quantity of Ice, km^3	Quantity of Ice accounted for by 1 km^2 of Permafrost, millions of m^3
European part and the Urals	190,000	370	1.9
Western Siberia	538,000	3,200	5.9
Central Siberia	2,342,000	5,500	2.3
Southern mountains	317,000	480	1.5
Trans-Baikal and southeast	1,035,000	1,450	1.5
Northeast	2,537,000	8,000	3.0
USSR as a whole	6,959,000	19,000	2.7

and confined to regions where syngenetic permafrost has accumulated. It is not of interest from the standpoint of determining the reserves of ground ice. The largest accumulations form buried glacier ice near the termini of the present-day glaciers.

Based on the information available, we made an attempt at estimating the total reserves of ground ice occurring in the USSR and giving rise to the so-called visible ice content of the permafrost (see Table 1). As the main ice reserves are concentrated in the upper layers of the permafrost, we used the following permafrost thicknesses for the calculations: rocky material, 50 m; loose, epigenetic material, 10-20 m; syngenetic material, on the average 10 m but in some areas of the northern zone, 20-30 m. For the calculation, only the permafrost occurring at the surface, immediately below the layer of seasonal thawing, was taken into account. According to our estimates, in the USSR the area occupied by this type of permafrost does not exceed 7,000,000 km^2, whereas, according to Baranov,[2] the total area occupied by permafrost is 11,115,000 km^2. Recent data, however, on the relic permafrost of western Siberia, indicate that this figure should be 11,451,000 km^2.

Overall, the total reserves of ground ice in the USSR, given the above-mentioned conditions, are estimated as approximately 19,000 km^3, of which about 1,000 km^3 are accounted for by multi-wedge ice. All the remaining types of sill-forming ground ice presumably amount to a further 1,000 km^3, approximately.

The data presented in the Table 1 denote only approximate, highly averaged amounts of ground ice by regions. Allowance must be made for the fact that most of the regions extend over a great distance in the north-south direction and include all of the geocryological zones, together with the distinctive features of their ice formation and their reserves of ground ice. In practice, therefore, the amounts of ice accounted for by 1 km^2 of permafrost range within wide limits in these regions. Thus, in central Siberia, with highly diversified geocryological conditions, they range from 50,000 m^3 in the islands of Severnaya Zemlya to 3,500,000 m^3 in the Lena-Vilyui lowland. In the northeastern part of the USSR, they range from 1,500,000 m^3 in the mountainous regions to 12,000,000 m^3 in the areas of coastal lowland. The figure for western Siberia is from 4,000,000 m^3 in the southern and temperate zones to 9,000,000 m^3 in the northern zone (Figure 3). Overall, however, the derived figures evidently do reflect the regional characteristics of the distribution of ground ice in the USSR.

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THE ROLE OF CRYOGENOUS FACTORS IN PROCESSES LEADING TO THE FORMATION OF PLACERS IN THE COASTAL ZONE OF THE SHELF OF THE LAPTEV SEA

N. F. GRIGOR'EV Moscow Lomonosov State University

At the present time the discovery and development of mineral resources in the coastal zone of the shelf of the arctic seas that border the USSR is an economic task of very great importance. In recent years prospecting and exploration activities relating to placers of valuable minerals have been vigorously conducted in a number of areas of the Laptev Sea littoral.

As is well known, various natural factors are involved in placer formation processes in the coastal zone of the arctic shelf. These include geological-geomorphological, hydrological-hydrodynamic, and cryogenous factors. The role of the latter has been studied least of all, although in certain situations they give rise to favorable conditions for the formation of placers, especially present-day placers.

The entire mainland part of the the Laptev coast is in a zone of continuous thick permafrost, and it is only beneath the beds of major rivers that this is cut through by narrow open taliks. At some points on the mainland coast (Nordvik and Tiksi), the total thickness of these strata is as much as 500-550 m. Within the transitional zone lying between the continental shoreline and the continental slope in the Laptev Sea, three main types of permafrost are traced: the abraded continental, submerged continental, and shelf types.

The abraded continental type occurs beneath the sea bottom opposite steep coastal bluffs consisting of Quaternary deposits containing copious amounts of ground ice (up to 50-70 percent of the total soil mass). As a result of the abrasive and thermal abrasive action of the sea and the direct influence of thermokarst processes, coasts consisting of ice-rich soils rapidly disintegrate and the shoreline not infrequently retreats at the rate of 5-6 m/yr. As the offshore ledges constituting the upper layers of thick permafrost were destroyed by the action of the sea, their unabraded part became covered by a layer of seawater. Thus, in the littoral areas of the shelf of the Laptev Sea, where active retreat of the shore has been in progress for many hundreds of years, beneath the sea bottom a surface-abraded relic layer of permafrost was found, which, in the form of a thick northwardly tapering submarine peak, underlies inshore areas of the bottom that were once dry land. In the immediate proximity of the shore, this submarine layer of permafrost has preserved its great thickness and relatively low temperature, while in its upper part, as indicated by drilling of boreholes in the vicinity of Muostakh Island, what may be the lower ends of thick wedges of fossil ice have been preserved.

The submerged continental type is distinguished from the abraded continental type by the transformations that occur in the frozen strata after the latter have become covered by a layer of seawater. As a result of the boreal transgression, the northern margins of the vast maritime lowlands, which by that time had become frozen to great depths, were flooded by cold seawater. As a result, there was a rise in the temperature of the submerged continental strata, and in conformity with the prevailing temperature of the benthic layer of water (between -0.8°C and -1.2°C) and the determining value of the geothermal gradient, its thickness decreased abruptly (evidently to 40-50 m).

The present-day shelf type of permafrost is widely distributed in the shallow areas of the bottom of the Laptev Sea. It is subdivided into the coastal delta, coastal marine, and marine subtypes. The first of these, consisting chiefly of river sediments, is mainly confined to the coastal areas of the sea bottom at sites where bay mouth bars and their analogous accumulative forms of bottom relief have evolved (the Yana and Indigirka bars, for example, are between 10 km and 20 km distant from the sea margin of their deltas). The second subtype occurs mainly in the shallow areas of the bottom of gulfs (such as the Siellyakh and Khrom gulfs). The third is found chiefly on marine banks and dried-out areas, as well as on the wave-built islands situated in the open part of the Laptev Sea, where as a result of intensive accumulation of sea drift, the depths are at most 1.0-1.2 m.

The main determining factor giving rise to the formation of the present-day shelf type of permafrost is cooling of the benthic sediments through a layer of sea ice measuring up to 1 m in thickness, with the result that in winter they freeze to a greater depth than the depth to which they succeed in thawing out in summer.

In the Soviet Northeast, the principal placer deposits of gold, tin, and other valuable metals were formed in the Paleocene-early Oligocene and Neogene periods, when thick crusts of weathering were generated and valuable minerals were released from the bedrock and redeposited by currents of water. It was during this period that the majority of the buried river placers were formed, both those located in the contemporary land area and those situated in tracts of mainland that had sunk below sea level at the time of the transgression. The processes leading to the formation of placers persisted on a vastly smaller scale throughout the whole of the Cenozoic period and are continuing to take place today.

According to present-day concepts, the formation of the permafrost in the Soviet Northeast dates from the first half of the Pleistocene and has continued without interruption right up to the present time. Evidently, in this area the placer formation process was most active during the period leading up to the emergence of permafrost there; the majority of the placer-containing Quaternary deposits froze epigenetically. Simultaneously, these deposits became saturated with fossil ice of various types and were cemented by it.

Despite the fact that placer formation was most active during the period leading up to the development of the cryogenous processes, their study is of independent interest, particularly since it enables us to obtain a better understanding of the distinctive features that characterize the formation of present-day placers under arctic conditions. When discussing the interrelationship that exists between the local cryogenous conditions and the placer formation processes in the coastal zone of the Laptev Sea, mention must be made of the role of the temperature regime of the frozen soils, which in many ways determines their strength, and also of the cryogenous structure of these soils, since it is this that is primarily responsible for the development of cryogenous phenomena and coastal processes.

On the coast of the Laptev Sea there are outcrops of bedrock containing pockets of valuable minerals. The abrupt temperature fluctuations in the surface layer of these rocks lead to an intensification of the physical weathering processes, the grinding down of rock masses, the development of physicochemical processes, and the more rapid release of the minerals from the rocks enclosing them. For the concentration of heavy minerals and origination of placers, it is of vital importance that there should be firm areas of bedrock on whose surface a placer deposit often forms. The hard upper surface of fine-grained perennially frozen deposits occurring both on land and under water at various depths below sea level constitutes a unique cryogenous bed on whose surface it is possible for minerals that have a high specific gravity to become concentrated. In the immediate beach zone, where present-day placer formation processes are occurring, such a bed is furnished by the base of the layer of seasonal thawing, which in winter joins up with the permafrost layer underlying it, normally at a depth of between 0.6 m and 1.5 m. While the cryogenous bed consisting of a submarine abraded-continental layer of permafrost frequently lies at depths of between 1 m and 15 m, it is known that the hard surface of the submerged continental layer can plunge to a depth of 86 m below the sea bottom. Thus, in the spring of 1971, when exploratory drillhole No. 1,000 was being sunk through sea ice at a preselected point in Van'kinaya Bay of the Laptev Sea, a hard surface of fossil ice containing a large quantity of inclusions in the form of finely fragmented plant remains was found at a distance of 10 km from the shore (depth ~4 m) beneath a layer of sandy and muddy silts. The possibility cannot be ruled out that this type of cryogenous bed forms the upper surface of a submerged continental layer of permafrost. In this case, the presence of a thick layer of ground ice at depths exceeding 86 m below the bottom may be associated with pronounced tectonic subsidence of this area.

The presence of a shallow cryogenous foundation in the bay-mouth bars of the major rivers, which together with the copious solid runoff are responsible for the transportation of the fine particles of minerals, may also give rise to conditions favoring the formation of delta-bar placer deposits that are concentrated in the upper layers of the contemporary shelf-type permafrost. Whereas the low temperatures of the weakly mineralized layers of permafrost are responsible for an increase in the strength of fine-grained sediments and the origination of a cryogenous bed, the rapid freezing of the surface layer of benthic and coastal deposits leads to consolidation of the fine-grained material containing the minerals and, at the same time, prevents the placer-containing detritus from being washed away and redeposited by the sea. As a rule, the ice-rich, upper and middle Pleistocene deposits that make up the bluffs of the coastal lowlands contain a small quantity of disseminated minerals. As the blocks of frozen soil undergo thermal abrasion and caving in the sea, they lose their content of ground ice through thawing. The thawed portion of soil material gets onto the beach and eventually onto the continental shelf. Presumably, natural separation of the soil takes place under the influence of wave cutting and offshore currents. The fine particles of heavy minerals will naturally collect on the surface of the cryogenous bed--the upper surface of the abraded continental strata of permafrost. Thus, as a result of repeated erosion and redeposition by the sea of the material arising from the shore, unique reservoirs consisting of layers of soil enriched with finely particulate minerals are able to form.

Where a primary mineral deposit is located in the immediate proximity of the shore, metalliferous detrital material will get onto the beach as a result of the development of slope and, in particular, solifluction processes. As a rule, colluvial-solifluction deposits contain pockets ice, mostly in the form of interlayers and lenses of varying thickness. The thawing of this ice results in the slope deposits becoming inundated, and this hastens their movement downhill onto the beach where they are then reworked by the sea and a coastal mineral placer is formed.

When engaged in paleogeographic reconstruction of placer-containing series of granular deposits, in some cases identifying the distinctive features of their cryogenous structure may be of great importance, since these can serve as genetic and stratigraphic indicators. For example, based on what has been made known to date about the peculiarities of cryogenous structure, and the possibility arises of determining whether a series of granular sediments froze epigenetically or syngenetically. This, in turn, may help us to ascertain under what conditions--subaqueous or subaerial--the placer-containing layers of granular sediments took shape during the period in which the permafrost series were being formed.

For example, the presence of thick syngenetic ice wedges as components of the permafrost not only attests to the markedly continental conditions that existed at the boundary between the middle and upper Pleistocene, but also indicates the pattern of sediment accumulation that prevailed when the thick series of lacustrine, fluvial, and slope sediments were being formed.

The cryogenous structure of the present day placer-containing coastal marine deposits have hardly been studied at all. The development of well-defined cryogenous diagnostic criteria will provide us with a firm foundation for the possible distinguishing of a Holocene complex of coastal marine sediments. Ascertaining the interrelations between the individual stages of placer formation (which were accompanied by cryogenous sedimentation of the coastal deposits), the dynamics of the coastal processes and the neotectonic activity appears promising as a means of predicting the contemporary placer formation processes in the coastal zone of the shelf of the Laptev Sea.

THE POSITION OF CRYOLITHOGENESIS IN THE OVERALL SCHEME OF POLAR LITHOGENESIS

I. D. DANILOV   Moscow Lomonosov State University

In N. M. Strakhov's[12] basic work on the theory of lithogenesis, four main types are distinguished: icy, humid, arid, and extrusive-sedimentary. The icy type of lithogenesis, according to Strakhov, encompasses only the areas in which ice sheets of varying thickness existed for a long period of geological time. The areas of marine, lacustrine, and fluvial sediment accumulation situated in the polar regions immediately adjacent to the ice sheets belong to the humid type of lithogenesis. He notes that even the "deposits of melted glacial water belong to the other, humid type of sedimentary process" (Volume 12, p. 137).

The specific nature of the lithogenetic processes in the arctic zones, which is a consequence of the severe climate, has led some authors[8] to the conclusion that it is desirable to distinguish a special, polar type of lithogenesis, encompassing areas of both continental and marine sediment accumulation. The specific features of polar lithogenesis that distinguish it from lithogenesis in the temperate climate of the humid zone are considered to be negative mean annual air temperatures, precipitation mainly in the solid state, the presence of sheet ice on the land and on the seas, the low temperature of the waters in the basins of sedimentation, the predominance of the mechanical type of erosion, the involvement of floating ice and icebergs in the dispersion of the sedimentary material, the formation of mainly terrestrial deposits admixed with small quantities of material of chemical and organic origin, and the retarded character and primitiveness of the diagenetic processes.[8,9]

B. A. Zubakov[6] has put forward the notion of a cryogenous structure consisting of two main varieties: an icy and a glacial. Among the icy structures the author distinguishes sea-ice and ground-ice structures, and among the glacial structures, land-glacial and sea-glacial. The formation of these structures takes place under the conditions of the cryogenous type of lithogenesis, which is characterized by the involvement of ice (as distinct from "normal" lithogenesis in which only water in the liquid state is involved). Zubakov places great emphasis on the fundamental importance of the differentiation of cryolithogenesis into the glacial and icy subtypes, this differentiation being the result of the dissymmetrical division of the Earth's crust into continental and oceanic blocks and, accordingly, the dissymmetrical division of the climate into continental and oceanic types. Lithogenesis of the ground-ice formation zone (cryological lithogenesis) is lumped together with sea-ice lithogenesis into a single, icy type of lithogenesis.

In emphasizing the specific nature of the lithogenetic processes in the permafrost zone, A. I. Popov[10] defines cryolithogenesis as a special type of lithogenesis occurring at below-freezing and low positive temperatures. It is characterized by phenomena resulting from the formation of ice in the Earth's crust and the accompanying typical geological and geomorphological effects. Popov[10] suggests that a distinction should be made between cryogenous diagenesis (*cryodiagenesis*) and cryogenous weathering (*cryohypergenesis*). Cryodiagenesis shows up in sustained irreversible ice formations, both in new sediments and in loose, previously lithified soils. Cryohypergenesis is caused by the regular alternation of freezing and thawing, that is, by short-lived reversible ice segregation, and is completed by the formation of a fine-grained cryoresidual product: cryopelite.

Sh. Sh. Gasanov[1] agrees on the necessity of distinguishing cryolithogenesis as a special type of lithogenesis confined to areas in which there is terrestrial, ground- and sea-ice formation. N. A. Shilo and Yu. V. Shumilov[13] use the specific features of the inland lithogenesis that occurred on the plains of the Soviet Northeast throughout the whole of the Quaternary period as grounds for the distinguishing of a periglacial type of lithogenesis.

An analysis of published sources indicates that essential differences underlie our knowledge of the extent, most typical features, and attributes of the icy, polar, cryogenous, and periglacial types of lithogenesis and also of the properties of the end products of lithogenesis that are determined by them, i.e., sedimentary rocks.

Information derived mainly as a result of a study of the Pleistocene deposits in the plains of the northern part of Eurasia affords a basis for stating that all the types of deposits that form in the various facies arrangements under arctic and subarctic climatic conditions possess specific features that distinguish them from the homologous deposits of the temperate zone, where the climate is humid. Evidently, in the zones where there is stable cooling of the Earth, it is necessary to distinguish a whole set of processes leading to the formation of special types of sediments and, in consequence, of sedimentary rocks under inland and marine conditions.

The main factors determining the specific nature of the sedimentary process in the polar latitudes are the low temperature and water in the solid state, i.e., ice. Consequently, *polar lithogenesis* can be taken to mean lithogenesis in the zones where there is stable cooling of the Earth and the active involvement of ice (the terrestrial, ground, and surface ice of impounded bodies of water).

It is the low temperatures in the polar zones that are mainly responsible for the terrestrial composition of the products of ablation and the accumulating sediments, as well as the high migratory capacity of certain of the elements (iron, manganese, and possibly others), which enrich the benthic soils of the resulting incompletely impounded bodies of water.

Terrestrial glaciers and the surface ice of water bodies only influence lithogenesis during the stage in which sedimentary material is being transported and sediments are accumulating (i.e., that of sedimentogenesis). It is accordingly possible to distinguish an ice and a glacial type of sedimentogenesis (but not of lithogenesis as a whole), and among these, an *ice- and-inland-glacial* type and an *ice-and-sea-glacial* type. The formation of the marine deposits involves the participation of surface ice, ice shelves, and icebergs, while that of the deposits of the closed basins and the alluvial deposits takes place only under the influence of the surface ice of bodies of water. Inland moraines are deposited solely by terrestrial glaciers.

The effect of the ice-glacier factor on the marine sediment accumulation process gives rise to such specific features of the deposits of the arctic basins as weak grading of the material, its predominantly clayey-silty composition, and the presence of pockets of rudaceous matter. In closed bodies of water, by leveling the surface irregulatities the continuous and permanent sheet ice favors the emergence in the sediments that are in the least altered form of rhythmical patterns in the inflow of terrestrial material. As a result, thick strata of rhythmically bedded (and in particular, ribbon-bedded) rocks are built up. Thus, the character of the sedimentation basin determines the lithological effect of the involvement of the ice-glacier factor in the sedimentary process. River ice has an appreciable effect on the alluvial sediment-accumulation process.

Ground-ice formation influences the sedimentation process in all of its stages: in the stage of hypergenesis, transportation of the material (solifluction and creep), sediment accumulation, and conversion of the sediments into rock. In the main, however, it is manifested only under continental conditions.

It is probably advisable that the set of cryogenous processes associated with the origination of ground ice and its thawing should be regarded as *cryolithogenesis*,[11] by which is meant a part of polar lithogenesis as a whole. Depending on when the sediments become frozen and the ice separates out, cryolithogenesis shows up in various stages of total lithogenesis, usually in completing it.

The stage in which the sedimentary material is formed, i.e., that of weathering, and which is accomplished through the active influence of seasonal ice formation, Popov[11] calls *cryohypergenesis*. The stage in which the simultaneous accumulation of the sediments and their freezing occurs can probably be called *cryosyngenesis*. If freezing and ice segregation occurs under benthic conditions, i.e., during the course of diagenesis of the sediments, the latter pass through a stage of *cryodiagenesis*. The stage in which the already-formed sedimentary rocks become frozen after their emergence from beneath the level of the sedimentation basin and the completion of the complex of physicochemical diagenetic transformations can be called the stage of *cryoepigenesis*. The concept of the interrelationship between polar lithogenesis as a whole and the constituent parts of cryolithogenesis is illustrated in Figure 1.

The lithogenesis of the marine deposits, and also that of the deep freshwater and brackish (lacustrine, lagoonal, estuarine) basin deposits is characterized by a maximum degree of diversity in the stages of weathering and the depths to which it is manifested. The accumulation of these deposits and their subsequent transformation oc-

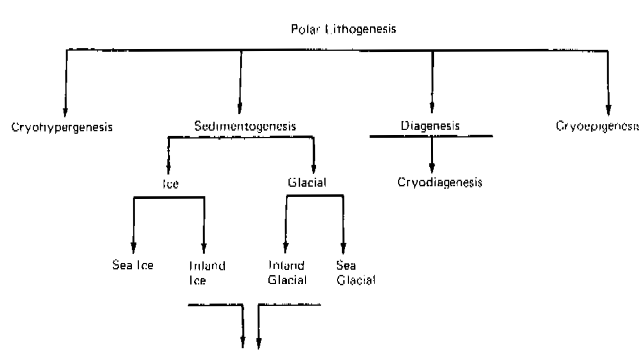

FIGURE 1 Diagrammatic representation of the constituent parts of polar lithogenesis.

curs in the thawed state. The sediments pass through all of the usual stages of lithogenesis as a whole. The diagenetic stage of their transformation is typified by complex and diversified physicochemical transformations that are fully comparable with those that take place in the terrigenous sediments of the temperate, humid climatic zone. As diagenesis proceeds, the relatively deep marine, freshwater, and brackish basin deposits become dehydrated and contract. Diagenetic cracks and subsidences form. Under certain conditions variously shaped plastic and discontinuous injection structures originate. The diagenesis of the marine sediments is accompanied by the transformation of buried muddy waters, which in some cases are converted into waters of the continental type. Under marine conditions an exchange of absorbed bases takes place quite actively in argillaceous sediments, with the result that calcium is replaced by sodium and in part by magnesium. Evidently, partial diagenetic transformation of the argillaceous minerals is accomplished in those areas of the cold sea basins that are deepest and furthest removed from the influence of land.

As diagenesis of the marine argillaceous sediments proceeds, there is a reduction of the hydroxides of iron and manganese and also of sulphates sulphides are generated that contract in the areas where organic matter is localized and form pyritic concretions.[5] In certain paleogeographic situations, lithogenesis of the marine and deep freshwater basin deposits is accompanied by a relative enrichment of them by dispersed carbonates. This occurs when there is an unusually high loss by erosion of soluble carbonates from the land, which end up in the ultimate incompletely impounded water bodies and become concentrated in the bottom sediments where the conditions are such that a harsh continental climate of the periglacial type prevails. As diagenesis proceeds, redistribution of the dispersed carbonates buried in the bottom sediments leads to the formation of concretions containing carbonaceous cement.[4] In marine sediments, calcium with a heavy admixture of carbonates of manganese, iron, and sometimes magnesium forms part of the cement contained in the concretions. In freshwater sediments, there is a marked predominance in the concretions of calcium carbonate admixed with a small quantity of manganese carbonate.

The specific nature of the authigenic mineral-formation processes, which is a consequence of the low temperatures, showed up as follows: Under marine conditions a variously composed group of carbonates was generated, whereas when the conditions were those of the freshwater (lacustrine and lagoonal) basins the concretions that formed were almost exclusively calcitic.

Thus, until they become frozen the marine and abyssal facies of the freshwater and brackish basin deposits undergo profound physicochemical diagenetic transformations. The sediments freeze subsequent to their emergence from beneath the level of the basins, by which time the complex of basic lithogenetic processes in them is almost complete. As the compacted and dehydrated soils are largely in the frozen state, on the whole their cryogenous textures are ice-poor (netted or laminated-netted). Stratal and lens-shaped ice accumulations are limited to lenses that are sandy and silty in composition. In this case, with respect to lithogenesis, freezing and ice segregation are on the whole postdiagenetic, i.e., cryoepigenetic.

In the shallow littoral zones of the seas and in the shallow closed and semiclosed basins, freezing and ice segregation can occur under benthic conditions.[2,3,7] The highly saturated soils become frozen. Freezing and ice segregation are leading factors in the dehydration and compaction of the sediments, i.e., they are cryodiagenetic processes. In this case the ice can be regarded as a unique low-temperature authigenic mineral.

Nevertheless, the main group of geochemical and diagenetic transformations is accomplished under benthic conditions in the moisture-saturated sediments in advance of their freezing. A group of dispersed authigenic minerals is generated and concretions form. In sandy, well-aerated varieties of soils, weakly individualized concretions and nodules of iron and manganese hydroxides develop. In those places where the carbonates deposited are of organic origin, clastocarbonaceous concretions form. Nodules of vivianite and microconcretions of pyrite are typical of argillaceous-aleuritic varieties of soils enriched with organic matter.

As the moisture-saturated sediments are frozen through, the soils have a high ice content. The inflow of currents of cold from below and from the sides leads to the formation of distinctive slanting cryogenous textures in the peripheral parts of the basins and of laminated-reticular textures in the central parts.[3,7]

The cryogenous process, i.e., cryolithogenesis, becomes a leading factor in the lithogenesis of alluvial, tidal marsh, laida, and shallow lake sediments. There is a marked decrease in the number of stages involved in the lithogenetic processes and also in the profundity of their physicochemical diagenetic transformation.

The deposits of the shoal areas, which are subject to periodical flooding or drying out (those of the tidal marsh, laida, and lacustrine alluvial types), and also the alluvial floodplain deposits become frozen while their sediments are in the process of accumulating, i.e., in the stage of cryosyngenesis. Ice-rich cryogenous textures develop, as do ice-wedge polygons and buried stratal ice. The usual physicochemical diagenetic transformation of the sediments is almost nonexistent. They remain in the sediment accumulation stage. There is no authigenic mineralogenesis.

In a number of cases the cryogenous transformation of the deposits that are subject to periodical drying out or flooding is complex in character. Partial thawing of the sediments that had previously become frozen and subsidence of the permafrost table may occur when there are ephemeral rises in the level of the basins or when fetch phenomena are present due to action of the wind on the water. Consequently, the soils again become frozen, although their ice content is appreciably lower. The subsidence at the perma-

TABLE 1 Stages of Cryolithogenesis

Cryosyngensis	Cryodiagenesis	Cryoepigenesis
Glacial deposits		
Alluvial deposits of floodplain facies--freezing simultaneously with sediment accumulation	Alluvial deposits of riverbed facies--freezing, partial-thawing, and refreezing	
Tidal marsh, laida, and lake-alluvial deposits--freezing simultaneously with sediment accumulation	Tidal marsh, laida, and lake-alluvial deposits--freezing, partial-thawing, and refreezing.	
	Lacustrine, lagoonal, and estuarine deposits--freezing under benthic conditions	Lacustrine, lagoonal, and estuarine deposits--freezing under subaerial conditions
	Coastal deposits--freezing under benthic conditions	Coastal deposits--freezing under subaerial conditions
		Marine shelf deposits--freezing under subaerial conditions

frost table is accompanied by a simultaneous thawing of the ice-wedge polygons and their substitution by subaqueous pseudomorphs, which are one of the distinguishing features of the cryogenous transformation of the soils of aqueous origin.

In proportion to their build up, the alluvial deposits of the riverbed facies are also subject to freezing and thawing, following which they again become frozen as a result of the riverbeds migrating when the alluvial series are being formed.

The cryogenous transformation of the soils, which is associated with freezing and subsequent partial thawing and is accompanied by a change in the ice content and the development of psudomorphs, can be regarded as a unique cryodiagenetic process.

Thus, the involvement of processes leading to the development of ground ice and subsequent thawing (cryolithogenesis) in the formation of sedimentary soils shows up in the overall scheme of lithogenesis and in its various stages, depending on the facies conditions under which the sediments accumulate. This dependence is presented in Table 1.

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PERMAFROST-FACIES ANALYSIS AS THE MAIN METHOD OF CRYOLITHOLOGY

E. M. KATASONOV Institute of Permafrost Studies, Siberian Branch of the USSR Academy of Sciences

Cryolithology is a branch of science that came into being on the borderline between permafrost studies and lithology. According to our definition,[5] cryolithology concerns itself with the study of frozen, predominantly Quaternary, sedimentary strata--their composition and structure in relation to the conditions of accumulation and freezing. The main objects of cryolithogical investigations are deposits formed under permafrost conditions. We call them cryolithogenic deposits. The corresponding special type of soil formation is cryolithogenesis.

Cryolithogenesis (it is also known as polar or periglacial lithogenesis) and individual problems of crylithology have been studied from various points of view by V. A. Zubakov,[4] A. I. Popov,[13] N. A. Shilo,[16] I. D. Danilov,[3] Sh. Sh. Gasanov,[2] and N. N. Lapina *et al.*[11] These authors made an important contribution to the development of the new science. However, their studies also contain some, and in our opinion important, basic shortcomings.

First, while providing different interpretations of cryolithogenesis and cryolithology, these authors say nothing at all about the methods of the new science. Secondly, they ignore the most important rule of sedimentology, i.e., the rule concerning the relation between the composition and structure of deposits, as well as inclusions in them on the one hand and their environment and conditions of formation on the other. In particular, there is no clear answer to the question: Does the cryogenous structure of perennially frozen deposits depend on their genesis? This dependence is rejected in some studies dealing with problems of cryolithology. If this is so, if cryogenous textures, soil and ice veins, sheet ice, and other phenomena that constitute the main features of permafrost are not related to the genesis of the latter, then there are no reasons for singling out a special lithogenesis and use the term cryolithology. Lithology and lithogenesis without the genesis of deposits are meaningless.

We believe that cryolithology studies the composition and structure of perennially frozen deposits in relation to their genesis. Cryolithogenesis is interpreted as a combination of related, alternating or parallel (although not coinciding in time) geological and cryogenous processes, as a *superposition* of cryogenesis on lithogenesis.

According to the definition provided by the terminological commission of the Institute of Permafrost Studies,[12] cryogenesis is "a combination of processes of physical, chemical and mineralogical *changes* and *transformations* (underlined by this author) of soils and rocks in the crust of weathering and the hydrosphere which occur at below-zero temperatures." Hence we shall interpret cryogenesis as changes in the structure and composition of sediments, rocks, and corresponding deposits resulting from processes that occur at negative temperatures.

The main processes of cryogenesis are various kinds of water crystallization. All kinds of migration of water (gravitational, capillary, and film water) occur mainly in unfrozen soils, and therefore they are regarded as conventional physical processes. As far as cryophilic crystallohydrates ($NaCl \cdot 2H_2O$, $MgCl_2 \cdot 12H_2O$, etc.) are considered, which exist at negative temperatures only and whose role in the structure of permafrost is still unknown, their crystallization is not an independent process, since salts and ice incorporated in cryohydrates are formed simultaneously. Frost fissuring, a variety of a more general process of thermal fissuring, plays an important role in cryogenous transformation of deposits.

The most noticeable changes in the structure of earth materials are caused by ice formation. The latter does not merely affect the mineral skeleton (for example, during freezing of monolithic rocks its effect is zero), but develops in close relationship with specific geological (hypergenic, sedimentation, and postsedimentation) processes. Ice formation may follow these processes or may become superimposed on them. The nature and scale of cryogenous transformations (cryogenesis) of deposits vary, depending on the place occupied by the aforementioned processes in the general scheme of lithogenesis.

In the sedimentation stage (after[14]), ice formation and frost fissuring, which repeatedly occur in the seasonally thawing layer, are superimposed on other processes leading to the disintegration of rocks. Frost weathering is of considerable importance, leading to formation of cryogenous residual material.[6,13] Denudation on slopes and sedimentation in drainage basins become more complex. Cryodeserption* and cryosolifluction here play an important role.

The diagenetic stage is the period of interaction of cryogenous and postsedimentation processes. As is well known, the subaqueous sediments are at first loose and unconsolidated. They become more compact with depth and are transformed to consolidated rocks in which lithogenetic fissures appear with time. Each stage of their lithification is distinguished by specific activities of oxidation and reduction reactions; rates of redistribution of salts; and rates of capillary, film, or pressure migration of water, i.e., the processes that facilitate or prevent ice formation and hence predetermine the changes in earth materials during freezing. In the final stages of diagenesis, the redistribution of water and hence ice formation abruptly slow down or cease altogether. Strongly compacted materials are practically immune to the effects of frost. As a rule, cryogenesis in them is reduced to freezing of water in previously formed fissures that sometimes remain empty.

Cryogenesis of deposits on flood plains, in deltas, and on slopes occurs under different conditions. The subaerial diagenesis that takes place there is characterized by the fact that transformation of sediments in the process of burial occurs in a relatively open system and is affected by the atmosphere, ground vegetation, active water exchange, and weathering agents: free oxygen, carbon dioxide and humic acids.[15] Materials whose diagenesis occurs in the seasonally thawing layer are especially affected by the environment. The direction and intensity of diagenetic and cryogenous processes taking place there are fully dependent on the landscape and the facies where sedimentation occurs. This makes it possible and desirable to study the ice content and the cryogenous structure of perennially frozen Quaternary deposits by methods based on lithological principles.

The interaction of cryogenous and geological processes, either in the seasonally thawing layer or in the taliks, is constantly affected (directly or indirectly) by the perennially frozen substrate. Incidently, this effect is often ignored by authors who talk about a "migration of water to great depths" and an ice lattice that thins out with depth.

*Formation of detritus by frost action.

The occurrence of the perennially frozen substrate at shallow depths, particularly below lakes and shallow sea bays, results in a gradual transformation of deposits to a perennially frozen state. This transition is accompanied by an abrupt cessation of weathering and diagenesis.

In the light of all this, the suggestions to regard freezing as a kind of cryometamorphism[1] or cryodiagenesis[13] cannot be accepted without criticism. Diagenesis and metamorphism are *irreversible* processes of regeneration of sediments and other materials. Perennially frozen deposits had been subjected to these processes in the unfrozen state. Freezing, however, is freezing. It is usually related to repeated ice formation, which, as we know, leads to frost weathering. Hence a contradiction arises: freezing, a factor of frost weathering, is cryodiagenesis.

In fact, as was already mentioned, freezing stops both diagenesis and weathering and preserves in the earth materials any changes which had taken place earlier. Therefore it would be more appropriate to talk about "cryoconservation." It would make sense to refer to various changes in the structure and composition of deposits by a good but hitherto little-used term--cryogenesis.

In all stages of earth-material formation, from the mobilization of substances to the transition to a perennially frozen state, cryogenesis and lithogenesis emerge as two sides of a single process--cryolithogenesis. The latter occurs over a wide area (the permafrost region), which includes both humid and arid areas, i.e., different climatic zones distinguished by the amount and the predominant type of meteoric precipitation. Therefore, cryolithogenesis cannot be regarded as a new or supplementary type, or as a variety of one of the types of lithogenesis proposed by N. M. Strakhov.[14]

Cryolithogenesis is related to the presence of permafrost and different permafrost-climatic conditions. On the basis of these conditions, we single out the following types of cryolithogenesis: subaerial-glacial, humid, semihumid, semiarid, and arid, as well as subaqueous and glacial-subaqueous.*

In our opinion, cryolithogenesis is an independent *series* of earth-material formation types, all of which exhibit common *permafrost-climatic* conditions. This series is comparable to the conventional lithogenesis series, whose types (ice, humid, arid) are combined in groups on the basis of a common climate. Further development and improvement of the general scheme of sedimentary-material formation, as outlined by N. M. Strakhov, would make it possible to single out other series as well. For example, we feel that the effusive-sedimentary type, as it is described in "Funda-

*These types form the basis of the classification of cryolithogenic formations that will be shown on the model of the geocryological map of the USSR prepared by the Institute of Permafrost Studies for the II International Permafrost Conference.

mentals of the Theory of Lithogenesis," should be elevated in importance and made into a series containing the types (humid, arid, etc.) of effusive lithogenesis. The improved scheme, in which subaerial deposits must be clearly separated from subaqueous sediments, will accommodate all types and forms of cryolithogenesis.

We are compelled to digress into the field of general lithology to underline its unbreakable bond with cryolithology and determine the place of cryolithogenesis in the general scheme of formation of sedimentary material. This problem is of fundamental importance in the choice of direction the new science will take.

The result of cryolithogenesis is formation of cryolithogenic deposits. The latter are distinguished by the fact that their "complete" cycle of postsedimentation changes from the moment of deposition to final freezing lasts only a few hundred years in some cases, and thousands of years in other instances.

The shortest cycle of changes occured in deposits that were freezing syngenetically, i.e., the deposits that make up shallow taliks and the seasonally thawing layer. At the time of their transition to a perennially frozen state, they were in an early or even the "zero" stage of diagenesis, as indicated by undecomposed moss, soft animal tissues, etc. They contain characteristic ice inclusions: layers and bands with smooth outlines in subaerial deposits, vertical and inclined ice wedges in subaqueous sediments.[7]

As we have stated on numerous occasions, the ice layers, bands, and lenses repeat the outlines of the lower surface of the seasonally thawing layer and the outlines of taliks. Their formation is directly influenced by the perennially frozen substrate and they, therefore, provide abundant information on the conditions of accumulation and freezing of sediments and help to divide frozen Quaternary deposits into cryolithological complexes with different structures and ice contents. Therefore, we refer to their cryogenous textures (layered and banded textures, inclined and vertical latticed textures[9]) as *marker* textures by analogy with the marker species of fauna and flora. We regard formation of these textures in little-changed earth materials and sediments as the main criterion of cryostngenesis. We have in mind the simultaneity (in the geological sense of the word) of accumulation and freezing of Quaternary deposits.

Syngenetically frozen deposits are very common in Yakutia and other regions of Siberia. As an example of their subaerial variety, we can take the "ice complex," i.e., peaty and gleyed supeses and suglinoks with banded or layered textures and thick ice veins. The syngenetically frozen subaqueous deposits have been described in the lower reaches of the Yana, Tumara, and Aldan rivers, including the well-known Mamontovaya exposures,[9] and in the upper reaches of the Tatta River (Ivanov and Katasonov, 1972).*

Syngenetically frozen deposts are vastly different from deposits in which accumulation and diagenetic changes were not influenced by cryogenesis prior to formation of permafrost. Such deposits may be called epigenetically frozen. They contain remnants of relatively heat-living plants (lower Quaternary-Pleistocene) and are as a rule strongly compacted. Ice inclusions formed due to migration of film, capillary, or pressure water are not characteristic of such deposits. Ice occurs only in fissures, which are narrow and sometimes half empty.† Lower Quaternary-Neogene deposits with a low content of fissure ice are exposed in the Mamontovaya Mountains in the lower reaches of the Omoloi River.

Apart from epigenetic and syngenetic permafrost, as defined by us, there are deposits that occupy an intermediate place with respect to their structure and the duration of postsedimentation changes. These deposits, which are mainly lacustrine and coastal-marine sediments, are formed under permafrost conditions but remain within large taliks away from the surrounding permafrost for a long time. Therefore, they have sufficient time to pass through the early and the middle stages of diagenesis before final freezing. On the basis of their relationship to the perennially frozen substrate, we call them parasyngenetic deposits.

The parasyngenetically frozen deposits are fairly compact and contain fissures, the walls of which reveal no traces of iron. As a rule, the fissures are completely filled with ice, which gives rise to fissured cryogenous textures. There are also inherited ice layers--the result of ice formation along the bedding planes. However, the marker cryogenous textures and other features bearing witness to a direct influence of the perennially frozen substrate are lacking in these deposits. The existence of the frozen substrate at the time of their accumulation is indicated indirectly by the presence of injection ice and by gradual transitions from parasyngenetically to syngenetically frozen deposits.

Parasyngenetic permafrost occurs over wide areas. Monolithic subaqueous parasyngenetic deposits with fissured cryogenous textures fill the depressed sections of trough valleys (e.g., near Allakh-Yun they are up to 50 m in thickness) and occur on individual sections of coastal plains and on the bottom of sea gulfs.[10] They were encountered in some boreholes in central Yakutiya (the upper reaches of the Tatta River), where they are on the whole not common due to the predominance of syngenetic high water-glacial deposits of middle Quaternary age (Ivanov and Katasonov).* Our investigations in the lower

*See Proc. II Int. Permafrost Conf., "Regional Geocryology."

†We have in mind the deposits that were still undergoing consolidation at the time of freezing, i. e., were not affected by weathering, or eluviogenesis as it is interpreted by E. V. Shantser.[15] Weathered and especially saturated rocks and soils in the weathered layer are characterized by a variety of cryogenous textures, which are not examined here.

*See Proc. II Int. Permafrost Conf., "Regional Geocryology."

reaches of the Enisei and the analysis of published data permit us to conclude that the greater part of frozen glacial-marine sediments in western Siberia is of parasyngenetic origin.

Therefore, perennially forzen deposts are subdivided into cryolithogenic, formed under permafrost conditions, and epigenetic. Cryolithogenic deposits are represented by syngenetically and parasyngenetically frozen formations. It is difficult to determine the precise boundaries between these subdivisions. The important thing is that there are distinct characteristics, i.e., marker and fissured cryogenous textures, which make it possible to distinguish these deposits in cross sections. Of course, there are also some "barren" deposits that do not contain any cryogenous features. But then, after all, the task of science is to correlate difficult (with respect to cryogenesis) layers with reliably established syngenetic, parasyngenetic, or epigenetic layers.

Cryolithology, as an independent science that concerns itself with the composition and structure of perennially frozen deposits in relation to conditions of their accumulation and freezing has its own specific methods. One of them is the method of permafrost-facies analysis.[5,6]

The essential points of this method are as follows. A field study is made of both sedimentary and low-temeprature features of earth materials: their composition, plant remains, cryogenous textures, ice and soil veins, etc. The analysis of these features in a cross section helps to single out facies, i.e., deposits, which were accumulating and freezing under specific landscape and low-temperature conditions. The facies are traced in space and time (in the cross section). In this way the circumstances of cryolithogenesis in former geological epochs are reconstructed.

The permafrost-facies analysis is based on a specific relationship: The cryogenous phenomena and corresponding processes are developed only in specific deposits. The environmental conditions and the mode of accumulation of the deposits determine not only their composition and water content but the type of freezing as well (in a seasonally thawing layer or a talik) and hence the nature of cryogenous features.

The relation of cryogenous features to the genesis of the country rock is established easily and unambiguously by comparing flood plain, lacustrine, and coastal-marine sediments that have just passed into a perennially frozen state. These sediments have different cryogenous textures, even if their grain-size composition is the same. Their individual facies are characterized by different ice inclusions, soil wedges, etc.

To understand the whole diversity of cryogenous phenomena and determine the genetic and zonal patterns of their development, it is essential to study different taxonomic environments of cryolithogenesis. The largest subdivision of this type is the sedimentation environment (subaerial or subaqueous); the most finely subdivided is the permafrost-landscape environment characteristic of various floodplain elements, slopes, etc.

Many investigators single out a large number of small permafrost-landscape subdivisions characteristic of "typical" sections of terrain. As new regions are being studied, the number of such subdivisions may become infinitely large, while the tasks of a scientific analysis require some sort of systematization. We group these subdivisions together on the basis of permafrost-geological characteristcis and single out sedimentary facies formed on floodplains, slopes, and so on. These facies are equivalent to specific permafrost-landscape conditions and have specific characteristics. They can be compared with each other, and, what is most important, they can be traced in meridional and latitudinal directions.

The permafrost-facies analysis made it possible to establish the cryolithogenetic types of the seasonally thawing layer. Each type has its own specific facies and is characterized by a specific regime of sediment accumulation and freezing, specific cryogenous processes and cryogenous features. We[8] distinguish the following types:

1. Seasonally thawing layer represented by marshy facies--peaty suglinoks, supeses, and peat. It is characterized by layered and banded cryogenous textures and by the presence of ice veins.*
2. Seasonally thawing layer consisting of sediments deficient of moisture formed on shallow parts of river channels, dry sections of floodplains, slopes, etc. Its water content is less than 15 percent, and it is characterized by soil wedges only (sketches and descriptions of these wedges can be found in the Katasonov and Solov'ev[9]).
3. Seasonally thawing layer consisting of loesslike and humic materials (water content 15-40 percent) formed on moderately moist elements of relief. Soil wedges are again the most characteristic features (filling wedges--see Katasonov and Solov'ev[9]).
4. Seasonally thawing layer represented by coarse-grained deposits (alluvial, fluvioglacial, etc.) formed under conditions of excessive moisture. This type is characterized by various cryoturbations, "boiling kettles," and involutions.

The examination of meridional cross sections compiled for each cryolithogenetic type of the seasonally thawing layer (Katasonov[8], Figure 1) revealed some interesting relationships. The marshy facies, for example, were found to be very conservative from the point of view of development of different cryogenous phenomena, which both in the north and south are represented by layered cryotextures and ice veins. The sediments in the third and fourth types of the seasonally thawing layer have more dynamic freezing regimes. The cryogenous phenomena characteristic of these types vary from south to north: Soil wedges and involutions are replaced by soil-ice and ice veins. These relationships become especially obvious if we single out and examine in

*Reference to ice veins in a seasonally thawing layer is puzzling (Editor).

the meridional cross section the more finely subdivided types of the seasonally thawing layer, such as marshy floodplain sediments and marshy sediments deposited on slopes, fluvial sediments with high water contents, and fluvioglacial deposits.

The geological approach to the interpretation of cryogenous phenomena has justified itself completely with respect to continental deposits. We attempted to adopt the same approach to the subaqueous deposits formed in conjunction with taliks. The cryolithogenetic types of the latter (see Figure 1) are based on the dimensions

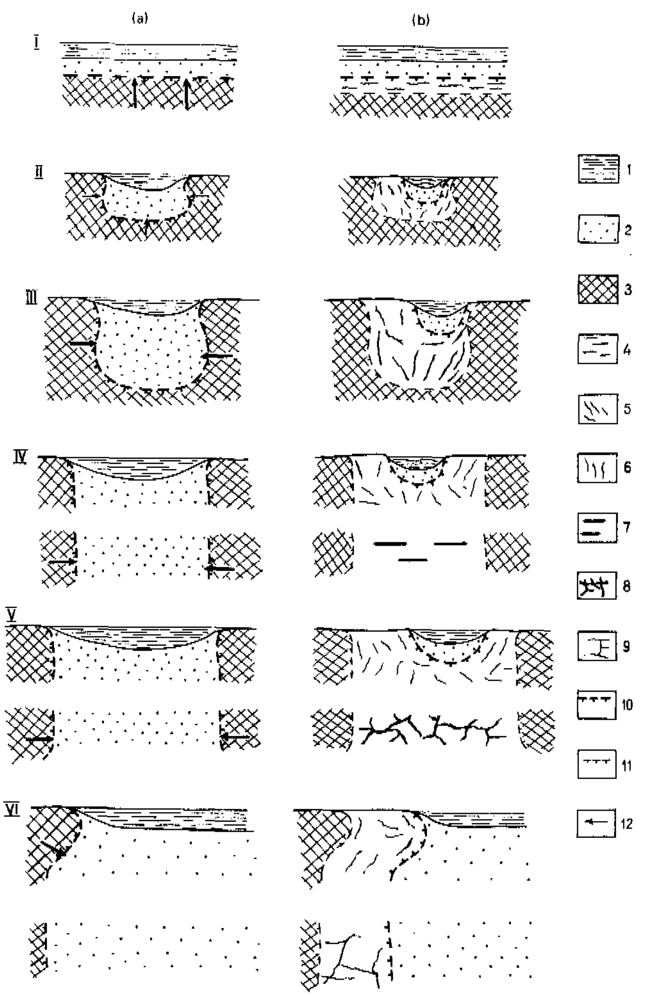

FIGURE 1 Cryolithogenetic types of taliks. (a) initial stages; (b) final stages of development.

I--subaqueous seasonally thawing layer; II--shallow closed taliks; III--deep thaw basins; IV--open taliks; V--large open taliks; VI--talik zones.

1--water; 2--unfrozen soils; 3--permafrost--frozen substrate; 4-6--cryogenous textures of syngenetically frozen subaqueous deposits; 7-9--cryogenous textures of parasyngenetically frozen subaqueous deposits; 10--permafrost table; 11--former permafrost table; 12--direction of movement of the freezing front.

of "thaw basins," since these determine the duration of the unfrozen state of soils, the nature of changes in the latter, their freezing regime, and the processes of ice formation. Figure 1 shows the cryogenous textures characteristic of subaqueous deposits in the taliks of various types, among which we include the subaqueous seasonally thawing layer. Later, Figure 1 will be supplemented by diagrams of large ice bodies, which in our opinion are formd in different taliks in different ways.

The permafrost-facies method accounts equally for geological and permafrost factors of earth-material formation. It helps to establish the basic principles of cryolithology as an independent science dealing with frozen sedimentary material.

First, it was established that the development of cryogenous phenomena in the present geological epoch depends on the permafrost-landscape conditions. This important and, in our opinion, quite obvious relationship was established for the Pleistocene deposits as well: Numerous data indicate that cryogenous textures, ice veins, etc, are related to the genesis of the accommodating deposits and the facies to which the latter belong.

Second, the isolation of facies, i.e., indicators of corresponding permafrost-landscape conditions of accumulation and freezing of sediments, made it possible to subdivide permafrost into genetic varieties with different cryogenous textures and ice contents. It became possible to determine the geological patterns of development of cryogenous phenomena, including buried ice.

Third, comparison of similar facies in Pleistocene and contemporary perennially frozen deposits provides a reliable basis for using the cryogenous index textures, soil wedges, and ice veins as *permafrost-facies characteristics* of Quaternary deposits. This makes it possible to recreate former conditions of accumulation and freezing of sediments.

In conclusion we should point out that the permafrost-facies analysis is being developed further. It would greatly benefit from mineralogical and petrographic studies of cryolithogenetic deposits.

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FROST WEATHERING

V. N. KONISHCHEV Moscow Lomonosov State University

Major characteristics of the composition of the sedimentary rocks forming in the cryosphere originate during the stage when matter is being mobilized in the weathering crust. It is as if weathering sets up frameworks in which the ensuing stages of lithogenesis take place.

In the view of many investigators, the main process leading to the disintegration of the rocks in the cryosphere is the disrupting action of the ice: frost weathering.[7-9,11,13,18] The effect of this action is in proportion to the number of cycles of freezing and thawing and the extent to which the material is moisture-laden. Under favorable conditions, frost weathering leads to the formation of very fine-grained residuum consisting chiefly of comminuted chips of the crystals making up the primary minerals. Experimental findings published in the literature and the results of a study of the composition of cryogenous residuum and the products of its immediate redeposition in various regions have made it possible to conclude that the comminution limit of the most widespread rock-forming minerals is a particle size of at least 0.01 mm. In the case of some minerals, it would appear that this limit may be even lower.

Some justification for this conclusion is afforded by the results of M. I. Sumgin's experiments, which establish that thin films of water (1.4 μm) freeze at temperatures that are very often observed in areas with a harsh climate.[12] Nevertheless, the actual freezing characteristics of the soil material (the beginning of water crystallization in the very large cracks and pores, and the migration of bound water from the thin cracks towards the centers of crystallization) place limits on the possibility of ice forming in the microfissures. This circumstance served as a basis for concluding that frost weathering can lead to the formation of very coarsely clastic residuum.[16] Comminution of the rocks without any appreciable change in mineralogical composition stems from a special process that is essentially physicochemical and is associated with the hydration and dehydration of the rocks (hydrational weathering). This type of weathering, in the opinion of some authors, is especially typical of the fairly humid regions where the climate is temperate.

On the other hand, where harsh climatic conditions prevail, it is not in itself of importance since it is suppressed by frost weathering.[16] Indeed, the thin aqueous films adsorbed on the walls of microfissures ranging in size from several micrometres to hairlike imperfections of molecular-sized crystals, generate a two-dimensional disrupting pressure amount to tens and hundreds of kilograms per square centimenter and are a powerful factor in the disintegration of the rocks.[2,15] Apparently, a peak disintegration effect should be observed in the presence of conditions giving rise to periodical regeneration or fluctuation of the disjoining pressure exerted by the thin aqueous films. One such condition is the alternate freezing and thawing of the rocks. Periodical ice formation in rocks is not only an independent factor in their disintegration but it also sets up highly favorable conditions for the emergence of hydrational weathering.

When a rock is in the process of becoming frozen, the commencement of crystallization is preceded by a stage of cooling from some positive temperature to 0°, or to the temperature at which water crystallizes. During the cooling process, the water contained in the soil undergoes important structural changes. There is an increase in the number of associated aggregates with an icelike structure and a decrease in its content of monomeric molecules.[1] The change that takes place in the properties of the water with the lowering of the temperature, and, in the view of some authors, the surface energy values of the skeleton,[15] increases the extent to which the soil particles are hydrated. This stems from the following adsorption formula, by means of which the binding of the water on the surface of the minerals can be written:[1]

$$\sigma = n\tau$$

where σ is the number of molecules accounted for by a unit of surface, n is the number of molecules striking a unit of surface in a unit of time, and τ is the average length of time that the molecules are in contact at the surface.

Since the hydration process has a dynamic character, with a lowering of the temperature there will be a rise in the value of τ, which is equivalent to an increase in the degree of hydration. The maximum hydration values of the soil skeleton are attained at 0°; this has been confirmed by experiments.[17] In this stage of cooling the value of the disrupting pressure of the thin films of water will also be a maximum. With the onset of water crystallization, beginning in the largest of rock pores, owing to the high hydrophility of the ice the water will migrate from the microfissures towards the ice crystals that have already begun to form. As the temperature continues to fall, the thickness of the films of water will accordingly diminish, as will their disrupting action.

Thus, the alternation that occurs in the freezing and thawing of the moist rock will result in the range of amplitudes of the disrupting pressure exerted by the aqueous films in the microfissures being the maximum possible and will set up highly favorable conditions for the hydrational disintegration of the minerals. The specific character of the circumstances giving rise to the fluctuation of the disrupting pressure (periodical ice formation) makes it possible to distinguish a zonal variant of this type of rock disintegration: cryohydrational weathering.

The disintegration of minerals through hydration occurs when there is an alternation of soil wetting and drying out. Under natural conditions, this is widespread in semiarid and arid climates. The results of the experiments cited later in this paper show that intensive disintegration of minerals takes place when there is cryohydration. Evidently, adsorption of the molecules of the gases on the microfissures when the water is evaporating from the soil is of great importance here. These gases tend to prevent the molecules of water from penetrating the microfissures during subsequent wetting. The water penetrates the microfissures more rapidly when the soils are wetted by melt water or low-temperature water by virtue of the increase in the water solubility of the gases as the temperature becomes lower. For this reason it is probable that at higher temperatures hydration of the walls of the microfissures will take place more slowly and will sometimes be impossible.

The data presented in Table 1 make it possible to draw the following conclusions, bearing in mind the accuracy of the grain-size analysis. In sand, particles measuring 0.5-0.25 mm are not only subject to disintegration during freezing and thawing, but also during temperature fluctuations in the positive-temperature. Moreover, the degree of disintegration differs but little. Compared with the control sample, the sand that was subject to periodical wetting and drying hardly changed at all.

In a suglinok, particles measuring 0.25-0.10 mm disintegrate more intensively. In comparison with the control sample, the decrease in the number of particles larger than 0.1 mm in the suglinok (+3°, +20°) was 10.3 percent; in the suglinok (-5°, +5°), the corresponding value was 14.7 percent. In the suglinok (-20°, +20°), which had undergone half as many cycles, the decrease in the number of particles larger than 0.1 mm was 5.2 percent. Alternate wetting and drying of the suglinok also led to disintegration of the sandy grains (the number of particles larger than 0.1 mm decreased by 3.2 percent), although this decrease was smaller than in the case of freezing and thawing.

The data derived from the grain-size analysis are confirmed by the results of a study of the soils in thin sections (Figure 1). The adduced data make it possible to state that the principal role in the fine-grained disintegration of sand during cryogenesis is played by the fluctuation in the disrupting pressure of the aqueous film.

Cryohydrational weathering leads to the accumulation of *melkozem*, in which the main share belongs to the coarse aleuritic fraction (0.05-0.01 mm).

In the general case, the condition determining the limit of comminution of the minerals is the relation between their surface energy, depending on the crystallochemical characteristics, and the properties of the surface-active medium (in this case, the water), depending on its temperature and mineralization.

With frost disintegration the fragments are much larger: Under favorable conditions they are probably not finer than coarse sand. This question, however, requires further study.

With frost weathering, the fissures and struc-

TABLE 1 Variation in the Grain-Size Composition of Fine-Grained Soils with Fluctuations in the Temperatures and Moisture Content[a]

No.	Soil	Temperature, °C	Moisture Content	Number of Cycles	Granulometric Composition, %						
					0.5–0.25	0.25–0.10	0.10–0.05	0.05–0.01	0.01–0.0005	0.005–0.001	<0.001
1	Control sample of suglinok		Soil saturated	--	9.4	19.4	8.0	32.5	10.5	12.6	7.6
2		-20; +20	Soil saturated	25	7.7	15.9	15.3	33.0	11.1	11.0	5.7
3		- 5; + 5	Soil saturated	50	7.7	6.4	24.0	32.0	10.6	12.1	7.2
4		+ 3; +20	Soil saturated	50	6.9	11.6	20.1	28.1	10.1	14.2	9.0
5		+20	Wetting and drying	25	9.5	16.0	10.0	33.5	10.9	11.8	8.3
6	Control sample of sand		Soil saturated	--	26.2	60.4	13.2	0.2	--	--	--
7		-20; +20	Soil saturated	25	22.2	64.3	14.0	0.5	--	--	--
8		- 5; + 5	Soil saturated	50	19.4	66.2	13.9	0.4	--	--	--
9		+ 3; +20	Soil saturated	50	20.2	72.7	6.7	0.4	--	--	--
10		+20	Wetting and drying	25	25.5	62.7	10.6	1.2	--	--	--

[a]The experiment was performed by M. I. Turbina in accordance with a program worked out by the author. The grain-size analyses were made by T. G. Fel'dman. The soils were moistened with distilled water.

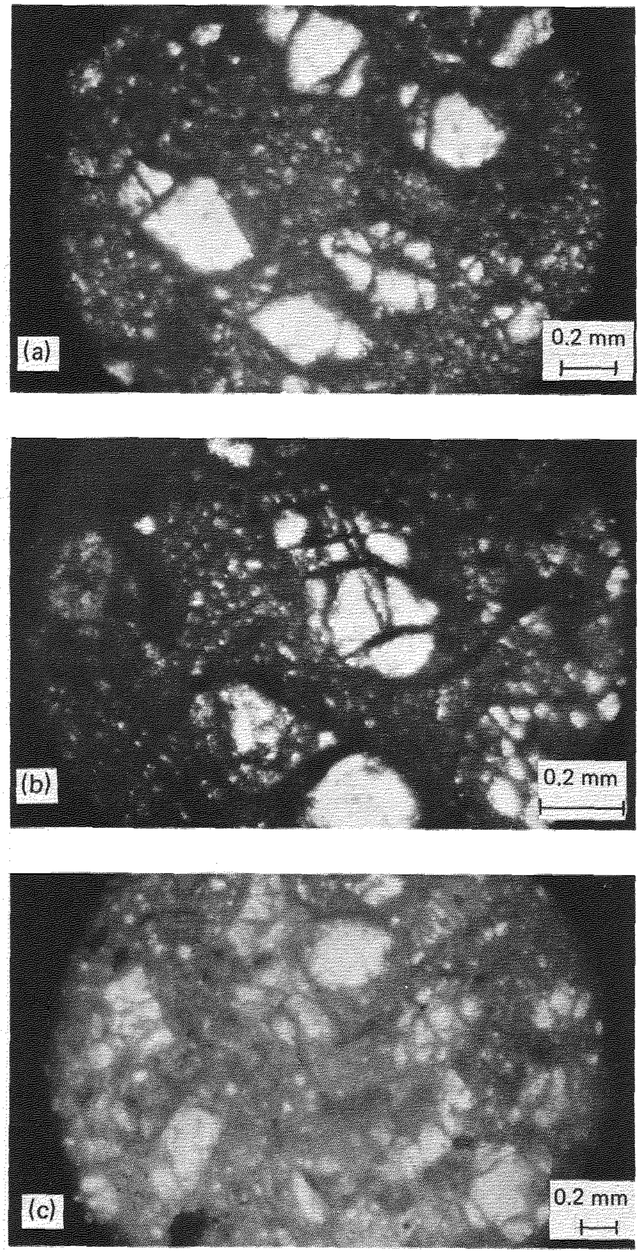

FIGURE 1 Comminution of sandy particles in a suglinok (photographed by M. I. Turbina). (a) t (+5°; -5°), 50 cycles; (b) t (+20°; -20°), 25 cycles; (c) t (+3°; +20°), 50 cycles.

tural defects that are present in the rocks and minerals are divided into three types, according to the character of the processes:

1. Macrofissures, in which crystallization of the water occurs and the ice exerts a disrupting pressure.
2. Microfissures, in which a change in the thickness of the aqueous films is observed, as well as a substantial fluctuation in their disrupting pressure.
3. Ultramicrofissures, in which the bond between the molecules of water and the mineral is so strong that the thickness of the film of water hardly reacts at all to changes in the external conditions, which are extreme for the region in question. No appreciable fluctuation in the disrupting pressure occurs in them.

The intensity of the disintegration and the sizes of the fragments when there is frost and cryohydrational disintegration are determined by the relation between the change in the external conditions (the amplitude of the temperature fluctuations, rate of freezing, and moisture content) and the surface properties of the minerals, which determine the mobility of the bound water.

Given favorable wetting conditions, the size of the fragments when there is frost weathering is inversely proportional to the rate of freezing.

The process of cryohydrational disintegration takes place with maximum intensity when the conditions are such that there is a cold, wet summer with a ground temperature close to 0° and a harsh winter. With the rise in the summer ground temperature, the intensity of the cryohydrational disintegration decreases and the average size of the fragments increases. The effect produced by cryohydrational disintegration reaches a maximum along the periphery of snow patches, glaciers, and icings. Throughout most or all of the warm season, the cold melt waters wet the rocks, thereby maintaining a high disrupting pressure in the microfissures. When freezing begins, these abruptly decrease in size in consequence of the water migrating in the direction of the growing ice crystals. It is this process that forms the basis of the nivation process, which occupies an important place in the morpholithogenesis of areas that have a cold climate.

Thus, frost weathering consists of at least two processes: frost disintegration, which leads to coarsely clastic disintegration of the rock, and cryohydrational disintegration, which is the cause of the fine-grained comminution. The sizes of the fragments that form when there is frost and hydrational weathering evidently correspond to the macro- and microfreezing distinguished by J. Tricart,[18] and also the coarse and fine comminution into which V. L. Sukhodrovskii subdivides frost weathering.[13]

Although these mechanisms act in conjunction with one another during the frost-weathering process, the relation between them may change, depending on the conditions. Henceforth it will evidently be possible to distinguish several types of frost weathering.

The question of the behavior of clay minerals when there is frost weathering will need to be given special consideration. A study of the weathering of massive crystalline rocks in cold humid regions has shown that the synthetic neogenesis of clay minerals is greatly hampered or absent altogether and that amorphous compounds of SiO_2, Al_2O_3 and Fe_2O_3 are the end product of the transformation of the endogenous minerals. This is due to the influence of the low temperatures that retard or preclude the majority of the chemical reactions and to the abundance of fine-grained organic matter that enters into an interaction with hydrates of sesquioxides and blocks their interaction with silica.[14]

As was indicated earlier, when the conditions are such that there is a constant alternation of freezing and thawing, a cyclical change in the degree of hydration of the mineral material takes place. This also tends to hinder the formation of clay mineral structures.

In the vast expanses of the north, various sedimentary deposits of Mesozoic-Cenozoic age fall within the sphere of frost weathering, included among which is a large number of clay minerals that had originated under dissimilar climatic conditions. The question of the behavior of secondary clay minerals in connection with cryogenesis has been little studied, and most of the investigators have not even mentioned it.

The existence of a limit of comminution of the hypogenous minerals that is much larger than the grain size of the clay minerals does not give grounds for concluding that when frost weathering is in progress they do not undergo important transformations. The distinctive features of the composition and texture and the crystallochemical properties of these two large groups of minerals give rise to a need for studying each of them separately. The fact that the clay minerals include molecules of water and OH groups, and the "sandwich" texture, high hydrophility, and "open" structural pattern of a number of the minerals as regards the molecules of water are the reason why they undergo major transformations when cryogenesis is in progress. The mobility of the bound water adsorbed by various minerals differs markedly. For example, on montmorillonite the bound water is more mobile than on kaolinite.[10] Generally speaking, the mobility of bound water on hypogenous minerals is less than that on clay minerals. In montmorillonite, even water that is present in the interstratal space reacts to a change in the external conditions. In a paper by D. M. Anderson and P. Hoekstra,[17] it is shown that there is a change in the interplanar distances in montmorillonite during the various stages of the freezing and thawing cycle (Figure 2). Typically, their peak values are observed at near-zero temperatures, i.e., we have experimental confirmation of the conclusion that the disrupting pressure exerted by the aqueous film is greatest prior to the beginning of freezing and after the ice melts.

The possibility of a fluctuation occurring in the disrupting pressure exerted by the aqueous films when there is a change in the temperature of the moisture content is directly proportional to the mobility of the bound water and the thinness of the microfissures. Consequently, in the case of argillaceous minerals, the limiting size of the microfissures and structural defects in which this process is possible is exceedingly small; it even includes the interstratal spaces.

As the temperature becomes lower, there is an increase in the dielectric transmissive properties of the water and the polarity of its structural associates, and, in consequence, in the degree of adsorption.[1] This enhances the possibility of water penetrating between the basal planes of the minerals, which becomes greater as the size of the particles and number of interstratal cations diminish. The repeated change in the degree of adsorption, caused by the freezing and thawing, leads to an increase in the interplanar distances and to hydration of the interstratal cations and inner basal planes and thereby makes it easier for organic molecules to penetrate the structure of the clay minerals. Ultimately, this process tends to split the single crystals of the latter, even to the point where they become separate lamellae.

It has been found by experiment that with repeated freezing and thawing a change occurs in the structural perfection of kaolinite and hydromica, which has probably been incorrectly interpreted as the formation of montmorillonite from these minerals.[3]

The diffractograms of the clayey mass and the morphology of its various (in particular, subaerial) deposits that were formed under harsh climatic conditions are characterized by a lack of distinctiveness and a lowered reflex intensity, which attests to the imperfectness of the crystalline structure, a predominance of degraded forms, and a high content of amorphous material.[4,6]

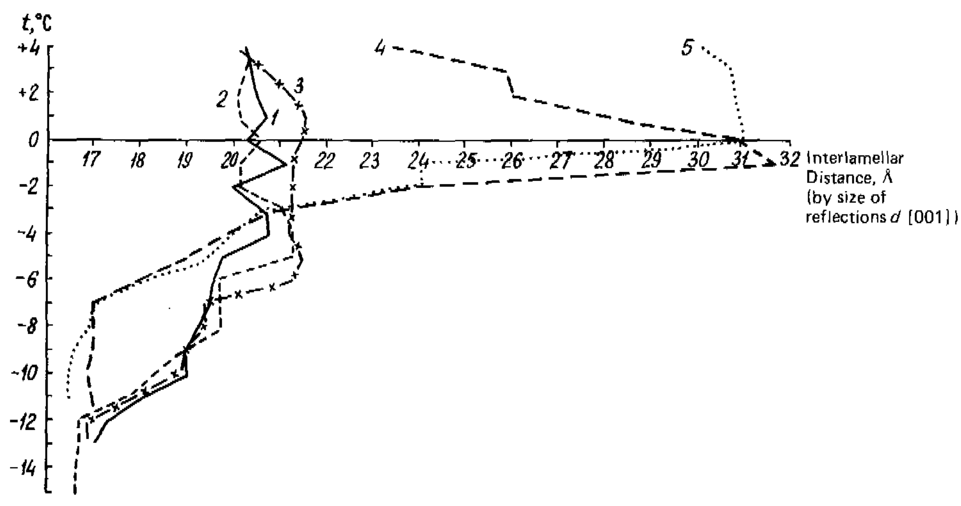

FIGURE 2 Change in the interlamellar distances in Wyoming bentonite under the influence of freezing and thawing. The graph is constructed from the data in M. I. Sumgin.[12]

Al-bentonite, cooling-freezing cycle: (1) w = 54 percent; (2) w = 80 percent; (3) w = 211 percent

Li-bentonite, thawing-warming cycle: (4) w = 75 percent; (5) w = 198 percent.

The fine-grained constitutent of cryogenous residuum contains a much larger quantity of free SiO_2, Al_2O_3, and Fe_2O_3 than do the underlying deposits.[5]

These findings make it possible to state that frost weathering causes disintegration of the crystalline structure of secondary clay minerals, i.e., their degradation takes place at a rate that is evidently not uniform in the various minerals. As a result, there is an increase in the content of hydrophilic substances that give amorphous X-ray diffraction patterns (allophanes, oxides of silica, iron and possibly aluminum, and organic clay complexes) in the cryogenous residuum of heterogeneous fine-grained soils.

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CRYOLITHOGENESIS

A. I. POPOV Moscow Lomonosov State University

Cryolithogenesis is defined as a combination of processes that determine the development of frozen earth materials characteristic of the frozen zone of the lithosphere. The most important factors of cryolithogenesis and the patterns of development of frozen earth materials have been repeatedly discussed above.[3,4] The publication of the main concepts of cryolithogenesis did not evoke any serious criticism in the literature, which does not mean, however, that the problem of cryolithogenesis has been completely solved, or that its scope, its zonal characteristics, the patterns of development of frozen earth materials, etc., are now fully understood. There are still many unclear and even controversial points that must be solved.

The aim of the present article is merely to summarize our ideas concerning cryolithogenesis as a natural phenomenon, and we shall not be able to elucidate all unclear aspects of the problem. We shall limit ourselves to a general review of the basic statements, stressing the most important points, and shall introduce a number of additions to and corrections of those concepts that were published earlier.

Cryolithogenesis as a natural phenomenon that manifests itself in geological processes related to freezing and thawing of earth materials, i.e., to formation and disappearance of ice in the latter, is the domain of a separate science--cryolithology.

It is important to note that cryolithology, which is a geological science, has one very important physicogeographical aspect. This is related to the fact that the very existence of ice, the most important component of frozen earth materials and an extremely unstable mineral in the upper horizons of the Earth's crust, depends on the heat exchange in the atmosphere-Earth's crust system. And the heat exchange, as is well known, is largely a function of physicogeographical factors, i.e., the transformation of the radiant energy of the Sun to the thermal energy, depending on the relief, vegetation cover, snow cover, and other factors.

Until recently there was no well-defined concept of freezing of earth materials as a geological process. Individual contradictory statements concerning this were not very convincing.

Our interpretation of freezing[3] as a residue-forming process in some cases and a diagenetic process in others has not met with any objections from geologists and permafrost scientists, and at the present level of our knowledge it is evidently acceptable. This author is well aware of the fact that in a number of instances such differentiation is arbitrary, although it is sound in its essential points.

The process of frost weathering (the so-called cryoeluvial process) represents a successive destruction of crystalline, metamorphic, and sedimentary rocks, which are gradually transformed to the final, residual, silty product. This process extends to unconsolidated materials (e.g., sand) as well, which are also transformed to residual silt when subjected to repeated alternation of freezing and thawing.

It should be emphasized that freezing of dense rocks (crystalline, metamorphic) is best regarded as an initial stage of frost weathering, which precedes a more substantial destruction of rock material, even if it is not accompanied by a successive change of phases. Therefore, crystalline, metamorphic, and other rocks that remain frozen for a long time (even in terms of geological time) must be regarded as being in the initial stage of frost weathering.

Frost weathering is most pronounced in the case of seasonal freezing and thawing of earth materials. Ice formation in this case is of a surging nature, and ice is the main agent of destruction. However, ice may not be present in material that is a product of frost weathering.

Cryodiagenesis is a process in which freshly deposited sediments and unconsolidated materials freeze without subsequent thawing, i.e., a process of long-term irreversible ice formation. Ice in this case behaves like an authigenous mineral. In the course of freezing and long-term irreversible ice formation, the fine-grained part of sediments or unconsolidated materials is dehydrated and compacted, which is undoubtedly a diagenetic phenomenon.

If freezing coincides with sedimentation and the sediment is not subjected to diagenetic changes of the usual type, cryodiagenesis is the only or, in any case, the main factor in the transformation of the sediment to rock, i.e., it is the principal factor of diagenesis.

However, if freezing starts after the diagenetic transformation of a previously unfrozen sediment to unconsolidated material, then cryodiagenesis must be regarded as a continuation of diagenetic changes and as a factor that superimposes its new diagenetic imprint on the already formed material. In this case cryodiagenesis is a factor of secondary diagenesis.

I. D. Danilov[2] refers to this type of freezing as cryoepigenesis, to differentiate it from cryodiagenesis, which is not really justifiable since freezing of unconsolidated materials is a continuation of diagenesis and not a transition to a new stage of a more substantial transformation. This follows from the fact that freezing of fine-grained earth materials (in the absence of seasonal thawing) is accompanied by further dehydration and internal compression of the mineral part, as well as by ice formation and formation of a superimposed cryogenous texture. The ability of water to migrate towards the freezing front in clays, suglinoks, and silts--the very fact of the mobility of loosely bound water--serves as an indication of a relatively weak lithification of soil. This is precisely the reason why in this case, as in the previous instance (freezing is the main factor of diagenesis of the sediment), superimposed and in principle similar layered and latticed cryogenous textures are formed in clays and similar soils.

It is obvious, of course, that such phenomena can take place only under conditions of low temperature and relatively low pressure, i.e., under conditions prevailing in the zone of diagenesis and not epigenesis or katagenesis. All this does not mean that there are no differences between the two cases of cryodiagenesis; the important thing is that basically they are similar.

In this connection we should point out that sediments transformed by katagenesis to solid rock (siltstone, mudstone) no longer contain bound water capable of migration on freezing. Therefore, no superimposed cryogenous textures are formed in them, and all phenomena related to diagenesis are absent. The same applies, of course, also to crystalline and metamorphic rocks. In all such dense rocks, freezing of free water in the fissures is the most typical phenomenon, and this is the process that is chiefly responsible for the destruction during frost weathering. Therefore, freezing of dense rocks, no matter how long, is always a weathering phenomenon.

Concluding the discussion of the geological nature of cryolithogenesis, it would be proper to define the most important characteristics of its two *main processes*: *frost weathering* and cryodiagenesis.

Frost weathering involves the following:

1. Disintegration of crystalline, metamorphic, and sedimentary rocks (mudstones, siltstone, sandstones, and limestones); successive change of stages--frozen fissured rock, cryoclastites and cryoclastopelites, cryopelite.
2. Predominance of mechanical destruction of rocks due to freezing of free water in the fissures in the initial stages of frost weathering.
3. No cryogenous textures can be formed in the initial stages of frost weathering, since there is no loosely bound water in the solid rocks and no migration of water on freezing.

Cryodiagenesis involved the following:

1. Freezing of fine-grained materials and formation of a fixed superimposed cryogenous texture in them in the course of long-term irreversible ice formation predetermined by the migration of loosely bound water toward the freezing front.
2. Dehydration and compaction by internal compression of mineral aggregates enclosed between the ice lenses during ice formation.
3. No breaking down of mineral particles during freezing; fixed initial mechanical composition of material.

In an earlier publication[3] we stated that it would be expedient to single out two main genetic cryogenous complexes, epicryogenous and syncryogenous, both of which are characterized by a set of specific characteristics within frozen or freezing earth materials. These characteristics are largely dependent both on the method and the temperature conditions of freezing.

It was found that freezing temperature and hence the intensity of the freezing process make it possible to single out genetic horizons in epi- and syncryogenous materials, which opens up new possibilities for a better interpretation of the genesis of these formations and of ice distribution in them, as well as for more meaningful permafrost mapping and surveys, identification of natural permafrost zones, and so forth.

As a result of additional research, our earlier scheme of genetic horizons has been modified by subdividing permafrost of the epicryogenous type into two subtypes and by considering the cases where seasonal freezing only is taking place. We now have: (a) a subtype corresponding to epicryogenesis on solid rock (crystalline, metamorphic); and (b) a subtype corresponding to epicryogenesis on unconsolidated earth materials.

After these modifications, the following genetic horizons (from top to bottom) should be distinguished in each subtype: (a) on bedrock--active cryohypergenesis (with a phase change), active cryohypergenesis (without a phase change), and passive cryohypergenesis; and (b) on unconsolidated earth materials--active cryohypergenesis (with a phase change), active cryodiagenesis, and passive cryodiagenesis.

It should be borne in mind that, in the case of seasonal freezing only, there is only one horizon of active cryohypergenesis with a phase change on both solid rock and unconsolidated materials.

The syncryogenous permafrost contains the following horizons (from top to bottom): seasonal cryodiagenesis, active cryodiagenesis, and relative conservation. In the case of sedimentation with seasonal freezing only, the horizons are: active cryohypergenesis (with a phase change) and relative conservation of traces of former seasonal freezing.

The processes that determine the formation of the aforementioned horizons were described by us earlier,[3,4] and there is no need to repeat them. We shall merely point out the properties characteristic of the new subtype of epicryogenous permafrost resting on bedrock.

The horizon of active cryohypergenesis (with a phase change) is characterized by seasonal freezing and thawing, i.e., there is a seasonal phase change within it. Intensive frost weathering is taking place on crystalline, metamorphic, and other rocks within this horizon. It is here that cryoclastites and cryoclastopelites, i.e., initial and intermediate residual products, are formed. The stage of rudimentary residual material, i.e., frozen material in solid rock, is very brief. Cryopelite, the final product of frost weathering, may also be formed on bedrock if there is practically no denudation, but this is rare.

Formation of dispersed and especially fine-grained material within this horizon leads to heaving and sorting, the most obvious result of which is the so-called patterned soil: stone polygons, stone nets, and other features. Solifluction and similar movement of material on slopes results in the formation of similar features: stone strips, rock streams, etc. Hence the processes characteristics of the horizon of active cryohypergenesis (with a phase change) are often reflected in the relief. Therefore, the characteristic relief forms may serve as indicators of various processes occurring in the given horizon.

The thickness of the horizon of active cryohypergenesis (with a phase change), i.e., the active layer, is closely related to the natural zonality and ranges from 0.5-1.0 m to 3 m and more.

The horizon of active cryohypergenesis (without a phase change) is below the preceding horizon, i.e., it is located within permafrost. It is characterized by more or less abrupt temperature fluctuations from season to season. The effect of temperature amplitudes on the continuity of solid rock within the negative temperature range has been studied most inadequately. However, field observations leave no doubt that within the zone of temperature amplitudes frozen solid rocks are weathered to a much greater extent than at large depths within the horizon of passive cryohypergenesis. There is no doubt that the factors responsible for "loosening up" the rocks are various coefficients of volumetric expansion and compression of rock and ice in the fissures. These factors invariably manifest themselves during temperature changes within the negative temperature range. General volumetric changes in the rock induced by temperature fluctuations have the same result. Rapid cooling from above so-called frost fissures may lead to formation of temperature fissures within this "horizon." The frost fissures are a very characteristic feature in the horizon of active cryodiagenesis in unconsolidated frozen materials.[3] They occur most frequently in mudstones, siltstones, limestones, and perhaps some sandstones. Crystalline and strongly metamorphosed rocks are subjected to frost fissuring to a much lesser degree.

Formation of frost fissures in the aforementioned rocks does not result in well-developed and thick polygonal ice wedges, because of very limited possibilities for compaction and plastic deformation (squeezing out) of semiconsolidated material.

We must also consider the more refined physicochemical weathering effect of unfrozen water on the walls of fissures in the bedrock, which invisibly manifests itself due to temperature changes in perennially frozen solid rocks.[1]

The thickness of the horizon of active cryohypergenesis (without a phase change) may reach 15-20 m and more.

Next comes the horizon of passive cryohypergenesis. Its specific characteristic is the absence or almost total absence of seasonal temperature fluctuations, which excludes the development of processes typical of the preceding horizon and leaves solid rocks containing ice-filled fissures in a relatively stable state. The long-term temperature fluctuations within this horizon are by far not sufficient to generate such an active cryohypergenesis as is the case with the preceding horizon.

The horizon of passive cryohypergenesis may be many tens or hundreds of meters thick, depending on the total thickness of permafrost.

All the above-mentioned genetic horizons may gradually shift downwards in the course of denudation, providing the latter is sufficiently pronounced. In this case freezing of rocks as a frost-weathering process becomes even more evident.

This process, which is essentially the opposite of syncryogenesis, although much slower than the latter, deserves much more attention than it is getting at present.

As was mentioned earlier, the genetic horizons proposed by us for epicryogenous permafrost initially consisting of sediments and unconsolidated materials and for syncryogenous permafrost were described in a previous publication.

Such are some, and in our opinion important, aspects of cryolithogenesis as a natural phenomenon, which deserves close attention of permafrost scientists and other specialists working in the field of geology and geography.

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EFFECT OF NEOTECTONIC MOVEMENTS ON FORMATION OF PERMAFROST REGIONS

N. N. ROMANOVSKII Moscow Lomonosov State University

Although the study of the effects of tectonics, including neotectonic movements, on permafrost was started relatively recently, there have been some noteworthy achievements in this field. The work of P. F. Shvetsov, A. I. Kalabin, and O. N. Tolstikhin dealing with the effects of faults on the distribution of pressure-seepage taliks and icings is well known. The same applies to the studies of A. I. Popov concerning the relation between syngenetically freezing deposits, including multiple ice wedges, and the regions of most recent subsidence. Additional important investigations are those of V. V. Baulin, G. B. Ostryi, and others,[2,3,18] who studied the effect of the depth of the base of the Siberian plate and the structures of the platform mantle on the thickness of permafrost, and the work of Yu. F. Andreev, A. N. Lastochkin, and others, on the relationship between lake thermokarst and neotectonic movements. E. B. Belopukhova[4] has recently published an interesting study on the tectonic control of formation of perennial frost mounds in the west Siberian plain.

The relationships between the permafrost processes and phenomena in the wide sense of the word and the neotectonic movements are varied. Some of them are of a direct nature; others are very complex. We shall concern ourselves mostly with the relationship in time between the start of perennial freezing and the neotonic movements, as well as with the peculiarities in the formation of permafrost, its cryogenous structure, and discontinuity, and so forth, related to these movements. The study of neotectonics includes the most recent movements during the Neogene-Quaternary stage of the Earth's History.[17] The time of existence of permafrost in Eurasia extends only to the Pleistocene, the last stage of the Cenozoic era, and perhaps also to some parts of the upper Pliocene. The age of permafrost is related to its zonal occurrence at high latitudes: in the north permafrost is relatively older than in the south; in the mountain areas of southern Siberia, central Asia, etc., the age of permafrost decreases with altitude. Therefore, to evaluate the effect of neotectonic movements on the structure, temperature regime, discontinuity, and other characteristics of permafrost, it is important to know the relationship between the time of its formation and the time of occurrence of tectonic movements. It appears expedient to single out the main cases of such relationships:

1. Perennial freezing started after the end of the active stage of neotectonic movements.
2. Perennial freezing of the lithosphere started before the active stage of neotectonic movements (or, more precisely, prior to their latest stage).
3. Perennial freezing was due to neotectonic movements.

In the first case, the neotectonic movements influenced formation of permafrost only indirectly. They were responsible for the main features of the present-day relief, some aspects of the fissuring of rocks, and the patterns of denudation and accumulation of Quaternary deposits, or they determined the lines of rejuvenated tectonic disturbances. Rapid movement of groundwater was related to the open part of the latter. These movements "prepared the substrate" for perennial freezing. They determined many characteristics of epigenetically freezing deposits, the cryogenous structure of the latter, the distribution of earth materials with a high ice content,[11] and the distribution of taliks, as well as a number of cryogenous phenomena occurring during perennial freezing. In some cases the effect of neotectonic movements on permafrost was about the same as that of older tectonic movements.

The Patomsk upland may serve as an example of the relationship between the stages in the development of neotectonic movements and the time of permafrost formation in eastern Siberia. According to Lungersgauzen,[16] the main neotectonic movements (the Patomsk phase) occurred here in the second half of the Pliocene and in the interval between the Pliocene and the Pleistocene. Permafrost composed of unconsolidated epigenetically frozen deposits is generally thick (up to 100 m), while frozen rocks are up to 200-300 m in thickness.[5] This points to their relatively young, upper Pleistocene age. During the Pleistocene, the strata of such thickness could have been subjected to repeated freezing and thawing.

In these epigenetically frozen strata, it is difficult to find direct signs of neotectonic movements, although for the purposes of permafrost regionalization, consideration of these movements is very important. This is done by identifying permafrost-geological regions on the basis of the most recent structures, which makes it possible to consider the effect of geostructural, geomorphological, hydrogeological, and other factors on permafrost. Therefore, it was neotectonic regionalization that formed a basis for permafrost-thermal regionalization of the Patomsk upland.[6]

In the second case, permafrost began to be formed prior to the occurrence of active tectonic movements in the region. Under such conditions, the effects of these movements are diversified.

In the mountains undergoing uplift, the permafrost conditions become generally more severe, due to absolute altitudes and changes in the relief of mountain areas. In the most recent stage, the uplifts are usually in the form of differentiated movements that lead to formation of strongly dissected relief of the alpine type. In this case the nature of cooling (freezing) of the lithosphere is changed and a large-scale cooling of rising mountain structures takes place.

As an example of the above, we can cite the central part of the Berkhoyansk Ridge and particularly the Kharaulakh Mountains. According to Lungersgauzen,[15] the main movements took place here beginning with the second half of the middle Pleistocene (the Verkhoyansk phase), when the Verkhoyansk Ridge was formed, and continued while shifting northwards during the first half of the upper Pleistocene (the Kharaulakh phase). Therefore, they occurred approximately from 150,000 to 50,000 yr ago. The uplifts in these regions ranged from several hundred to 1,000 m and more. This led to a drop in the mean annual temperature of earth materials in the upper parts of mountain structures ranging from 2-3°C to 5-7°C, and this is according to very modest estimates. Such a temperature drop in a relatively short period of the geological time was not only of the same magnitude as the amplitudes of the Pleistocene cooling and warming cycles in eastern Siberia, but also may have exceeded them. However, in spite of a general increase in the severity of the climate, the ground temperature and the discontinuity of permafrost at some altitudes in the temperature inversion regions may increase.

In structural forms subjected to relatively small uplifts and in those that had been stabilized, or were subsiding (the latter is especially important in the case of platforms and plates), the accumulated deposits were freezing syngenetically under subaerial conditions.[19] The ice content of these deposits depends on the temperature regime of permafrost in the substrate. The ice content of silty supes-suglinok alluvial and other deposits increases with the decreasing temperature of earth material due to both texture-forming[8,13] and multiple-vein ice.[20] Therefore, the ice content of continental deposits that accumulate in the areas of the most recent subsidence depends on the permafrost zonality. The higher the ice content of the deposits, the more pronounced the thermokarst, thermal erosion, and other processes that may take place later during changes in permafrost and geological conditions.

The neotectonic movements are reflected in the cryogenous structure of permafrost. Changes in the cryogenous texture of permafrost occur along the zones of active movements. This phenomenon was deduced from the cryogenous texture of epigenetically freezing soils, as well as from the structure of ice lenses in the Uyandina superimposed Cenozoic depression and in the adjacent part of the Selennyakh Ridge.[12] The most recent dislocations may leave traces in the cryogenous structure of syngenetically freezing deposits as well. For example, Velikotskii[7] pointed out the presence of fissures of tectonic origin in the middle and upper Quaternary alluvial deposits containing syngenetic multiple-vein ice. Lungersgauzen[16] mentioned the presence of young dislocations in the Lena delta, which disturbed the syngenetically frozen alluvial deposits and peat bogs widely occurring there.

The most recent movements may affect the discontinuity of permafrost and the taliks. The studies in the Yana-Indigirka interfluve have shown that these movements may displace the pressure-seepage taliks, change their type, and divide a single talik into two separate units belonging to different categories.[1] This facilitates the migration of icings formed in the valleys below these taliks, changes the form of the valleys in places where icings occur, and modifies the composition and cryogenous texture of alluvium. In one particular case (in the mouth of the Tekhon River on the southern slopes of the Cherskii Ridge), icings ceased to be formed due to the effects of neotectonic movements,[21] which we interpret as a change in the talik feeding these icings with water. In 1971 Afanasenko, Chizhov, and this author investigated a mineral spring in the northern part of the Ulakhan-Sis Ridge, which was fed by a pressure-seepage talik. The latter was located at the intersection of two large faults on the slope of a valley of a small creek, a tributary of the Ulakhan-Kyugyulyur River. Permafrost surrounding the talik was up to 390 m in thickness. The presence of recently dead trees and other indicators in the spring itself and on the icing proved that both the talik and the spring were formed not long ago. Mineral water broke through to the surface at the intersection of two rejuvenated faults evidently due to seismic phenomena (the seismicity of the region is 7 to 8). Recent formation of such taliks in thick and cold permafrost was previously unknown.

The relationship between permafrost and tectonic movements indicates that the time period during which these movements affected the properties of permafrost increased from south to north in accordance with the geocryological zonality. The structure of old permafrost indicates that the effect of neotectonic movements on it lasted longer than in young permafrost formations. In folded mountain areas, this relationship obeys the laws of geocryological altitudinal zonality: At higher altitudes the effect of tectonic movements on permafrost lasted longer than at lower altitudes.

Several facts indicate that this type of relationship exists in the vertical cross section of permafrost as well. In epigenetic permafrost that did not thaw since its formation and in polygenetic permafrost, the upper part of the cross section of the former is usually the oldest, and it is this part that contains most signs of tectonic movements. The age of permafrost decreases with depth; syngenetically frozen deposits are always younger also. Hence they contain fewer signs of tectonic movements. Therefore, the relationship between the age of syngenetic and epigenetic permafrost and the signs of tectonic movements in their structure may provide additional information on the time of occurrence of these movements.

In the third case, perennial freezing of earth materials was triggered by the most recent tectonic movements. In the mountains of southern Siberia, central Asia, and the Caucasus, permafrost, while absent on plains and low uplands, was formed in the upper parts of mountains in the course of ascending tectonic movements. The uplift of mountain massifs brought them to altitudinal belts where the heat exchange between the ground and the atmosphere was negative. The rate of uplift and the differentiation of tectonic movements had a considerable bearing on the nature of perennial freezing. Permafrost was very often formed simultaneously with glaciation. Permafrost of this type has not been adequately studied, but we may assume that its properties related to the effect of tectonic movements on the cryogenous structure, discontinuity, and other characteristics will lie halfway between those described for the first and second cases.

The relationship of neotectonic movements to transgressions and regressions of the sea is well known. It manifests itself especially clearly within lowlands stretching towards the polar basin. The sea transgressions related to neotectonic movements and changes in the ocean level led during the Pleistocene to a degradation of permafrost below the sea bottom. Dissolution of ground ice by saltwater, the temperature of which was usually below zero, played an important role in the thawing of permafrost. The temperatures in the water layers at the bottom and in the soil layers on the bottom varied very little throughout the area of the water basin. This led to a fairly uniform degradation of permafrost below the sea. The reduction in the frozen zone in the cross section depended mainly on the duration of existence of the water basin and the geological structure of the area. At present, such processes are occurring on a large scale in the shelf areas of the Laptev and East Siberian seas, which are Holocene thermal abrasion formations.

The sea regressions were accompnied by a long-term freezing of deposits exposed by the retreating sea. Formation of the *cryolithozone* took place under conditions that corresponded to the permafrost-temperature zones established at the time of the regression. The father north, the lower the thermodynamic level of the heat exchange at which freezing was taking place. From the point of view of geology, the replacement of subaqueous conditions by subaerial was proceeding rapidly. In the first approximation, the resulting temperature drop on the ground surface may be regarded as abrupt, especially at high latitudes. Therefore, perennial freezing under conditions of uplift and regression of the sea basin follows different laws of thermal physics than in the case of periodic changes of surface temperatures.[9] This generates considerable permafrost-geological effects. For example, freezing of water-saturated deposits, the lithification of which increases with depth, brings about a higher ice content in the surface horizons, but there is no gradual decrease in the ice content with depth. At considerable depths (100-200 m), fine-grained deposits contain large accumulations of injection and segregation ice, as, for example, in the northern part of western Siberia.[10] Let us recall that in the case of periodic changes in the heat exchange, the ice content gradually decreases with depth due to the attenuation of the heat exchange, while large accumulations of ground ice at considerable depths are impossible.[9] Cryohalogenic water is often present below permafrost in areas previously submerged under the sea, while intra- and interpermafrost water may be present in the permafrost itself.

Therefore, under severe conditions prevailing at high latitudes, the tectonic control of transgressions and regressions of the sea may also determine the characteristics of long-term thawing and freezing of earth materials and affect the cryogenous structure of permafrost.

It must be mentioned in conclusion that there are some very peculiar manifestations of neotectonic movements in the permafrost region, which are often very dependent on the severity of the permafrost conditions. This refers, first of all, to changes in the occurrence, shape, size, and even the very nature of water-bearing taliks. It is known that in the areas of occurrence of Quaternary deposits with a high ice content, thermokarst is much more pronounced in places where subsidence is taking place and the drainage conditions are deteriorating.[14] Progressive development of sublacustrine taliks is taking place here. On uplifted sections, where the drainage conditions are improving, the area occupied by lakes is reduced and the taliks below the lakes freeze.

Earlier on we talked about the neotectonic displacement of hydrogenic and hydrogeogenic pressure-filtration taliks that occur in conjunction with rejuvenated tectonic disturbances. The neotectonic movements usually alter the fissuring of rocks in the talik itself and in its immediate surroundings. This changes the movement of pressure water in the talik. The migration of the talik in the surrounding low-temperature permafrost appears to be as follows. Permafrost thaws near that part of the talik where the movement of water is becoming faster. The cross section of the talik increases, and the rate of water movement decreases. The talik begins to freeze in that part where the flow rate is slowest. Therefore, the displacement of taliks along the lines of tectonic disturbances is a result of two opposite processes: an increase in the fissuring and the resulting increase in the cross section of the talik, on the one hand, and the concentration of the ascending water flow resulting from "compression" of the cross section of the talik due to freezing, on the other. The displacement of the talik is accompanied by the displacement of the naleds formed below it. The latter in turn leaves behind it a strip devoid of trees and vegetation. As a result of this, the consequences of neotectonic movements are fairly easily deciphered on aerial photographs.

Under sufficiently "mild" permafrost conditions, the taliks evidently do not migrate, since there is no lateral freezing. Increase in the fissuring of rocks during neotectonic movements leads, under such conditions, to an increase in the total area of the talik, a change in the discharge of groundwater, formation or disappearance of naleds, and so on. Therefore, under severe permafrost conditions, the signs of neotectonic movements are often more distinct than under milder conditions. They are least distinct outside the limits of the permafrost region.

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FROST-INDUCED DEFORMATIONS OF ANTHROPOGENE DEPOSITS IN THE UKRAINE

M. V. VEKLICH, N. A. SIRENKO, and V. A. DUBNYAK Geography Section, Academy of Sciences of the Ukrainian SSSR

The upper Cenozoic deposits of the Ukraine are represented by several sedimentary formations: loess, Quaternary glacial-plain formations, red and brown Pliocene clays, and marine terrestrial and alpine Pliocene-Quaternary formations. With the exception of the alpine formations, all of them have been studied in detail (see Table 1). Eleven stratigraphic horizons have been identified in the Pliocene clays, 18 in the Quaternary loess, and so on.

The subaerial loess formations are represented by loess and buried soils; the red and brown Pliocene clays, by buried soils of various types that alternate in the cross sections in a well-defined sequence. Each of these horizons displays a specific combination of characteristic features that is absent in the other horizons. The subaqueous deposits, mainly alluvial and alluvial-lacustrine sediments, occur on six Pliocene and six Quaternary floodplain terraces, as well as on the high and the low floodplains. Each of the floodplain terraces and high floodplains has two stratigraphic alluvial horizons, which can be easily correlated with the horizons of the subaerial beds.

The stratigraphic differentiation of the aforementioned geological formations is based on well-founded general stratigraphic, geomorphological, paleopedological, physicochemical, spore-pollen, malacofaunal, engineering-geological, and other information. In particular, 42 reference sections and over a thousand other sections of the loess formation have been studied.

The marine Pliocene-Quaternary terrestrial deposits developed inland are differentiated to a smaller degree than are the continental deposits. Furthermore, many horizons of the Quaternary marine deposits cannot be identified inland (see Table 1).

The frost-induced deformations in the upper Cenozoic deposits of the Ukraine became an object of investigations as recently as the 1950's,[1-5,7-9] although forms that were similar to, or were in fact, such deformations were described without further comment as early as 1910-1920.[6] However, no special studies of these phenomena have yet been undertaken, and we too investigated them only in conjunction with other problems. Therefore, the available data are not of equal value: the stratigraphic information is very detailed, the morphological data are less so, and the genetic information is almost nonexistent. The best known are the loess formations, while subaqueous deposits have been studied to a smaller extent.

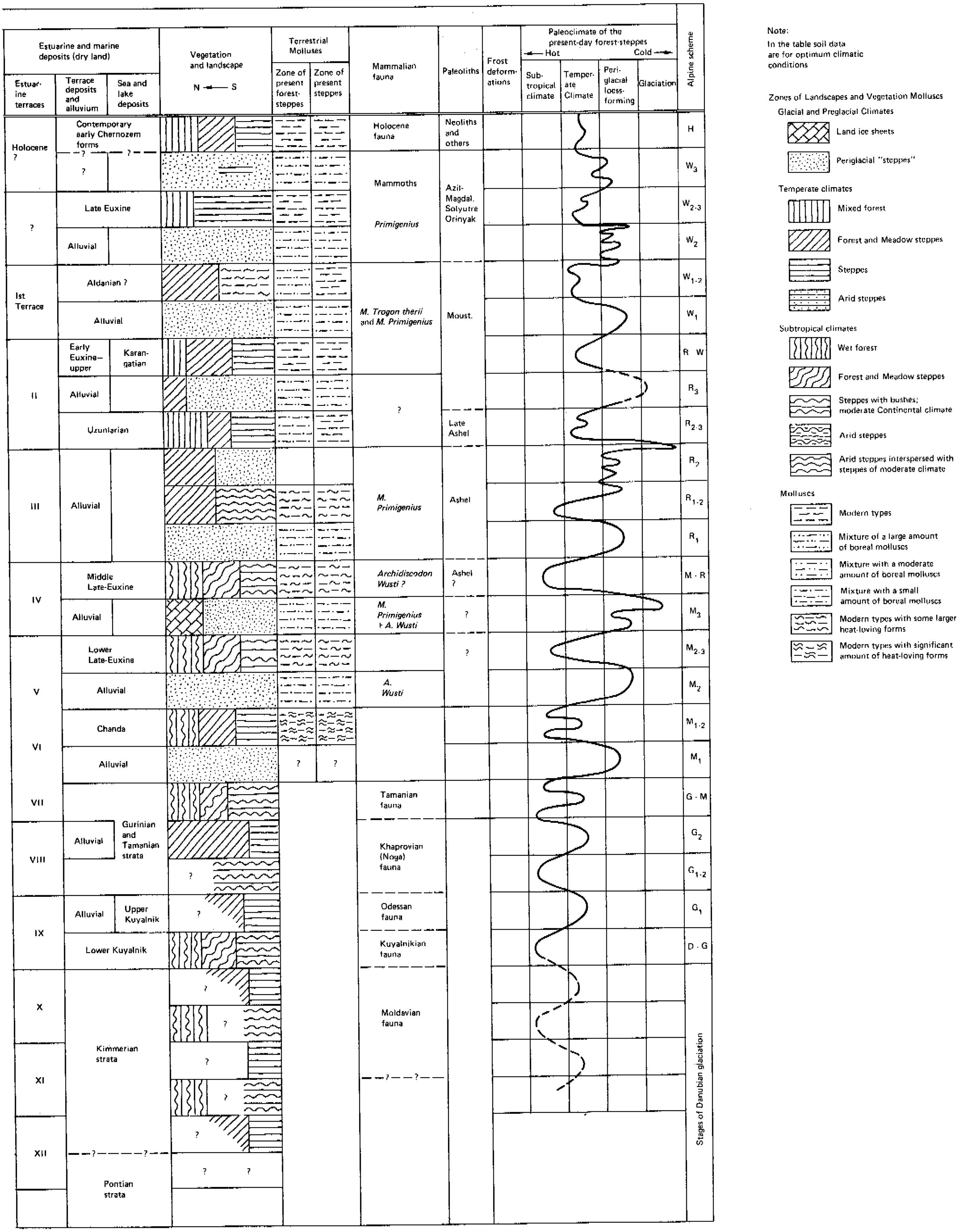
There are two main forms of frost-induced deformations: wedgelike fissures and small-scale crushing of layers. The fissures vary from 0.1-0.2 to 1-2 m in depth and from a few millimeters to 0.5-0.85 m in width. The upper part is usually below the loess horizon, while their middle and lower parts are located in the underlying subsoils. The fissures are filled with vertically layered loessial and other suglinoks, while in a cross section they may display bulges and a kettlelike form in their upper part. The upper layers of buried soils may bend away from the fissure. In plan view, the fissures form polygons of various, usually small, sizes (up to 1 or 2 m in diameter and usually less).

Small-scale crushing of layers was found in a few cross sections and stratigraphic horizons. They were observed either in the subaqueous deposits or in the hydromorphic formations. This type of crushing occurred most frequently in the hydromorphic buried soils and clay-suglinok subaqueous deposits of interglacial origin, as well as in the lower parts of the overlying layers. The horizontal dimensions of crushed structures were usually small.

Also observed were a peculiar microaggregation and differentiation of the mineral mass of individual horizons with respect to its grain-size composition, which could have been due to subsequent processes.

Since frost-induced deformations in the subaerial deposts are confined to the lower layers of the loess horizons and the underlying buried soils, while in the subaqueous deposits they occur throughout the cross section of alluvial or lacustrine-alluvial deposits, which are synchronous with the loess (and the moraine) horizons, and also in the upper parts of alluvial, lacustrine-alluvial, and lacustrine-marine deposits of the warm phase, these deformations are synchronous with the loess horizons. Nine generations of frost-induced deformations have been identified in the Ukraine. They are synchronous with the Azov, Sula, Tiligul, Orel, Dnepr, Tyasmin, Udai, Bug, and Black Sea loess horizons. There are also traces of these deformations of the Shirokoe (late Pliocene) time. The deformations display a geographic zonality: they are bigger, more distinct, and more frequent in the north than in the south. The best-developed deformations in the cross section are those of the Dnepr, Udai, Bug, and Orel age. Generally speaking, each age generation of frost-induced deformations is characterized by the predominance of definite types, forms of occurrence, and patterns of aerial distribution.

Unified stratigraphy of the Anthropogene of the European USSR (MSK. USSR 1964) and the Pliocene of the Black Sea region			Stages of development in the Ukraine in the Pliocene and Anthropogene (M. F. Veklich, 1965, 1968)				STAGES OF GROWTH AND DEVELOPMENT				
							Glacier and Subaerial development in glacial formations	Subaerial development of loessial red- and brown-colored deposits		River-valley and alluvial deposits	
System (period)	Basic sub-divisions	Horizons in the Pliocene strata	Period	Basic stages	Phase of the stage	Index of phases		Zones of maximum glaciation	Non-glaciated zones	Terraces	Alluvial horizons
QUATERNARY (Anthropogene) Q	Contemporary deposits Q_{IV}	Contemporary	Anthropogene (Quaternary) An	Late Anthropogene An_3	Holocene	h	Contemporary soils			Lower Poima	–
	Upper Quaternary Q_{III}	Baltian			Pre-chernozem	pts	Sands	Loesses; brownish, straw-colored	Loesses, brownish, straw-colored; locally with shallow chernozem-like soils	Upper Poima	Upper
		Ulasian			Dauphinian	d	Chernozem subaerial soils	Chernozem, with carbonates, and leached out, soils	Light-brown, carbonate soils, and underlying chernozem		Lower
		Kaliniuian			Burgalian	bg	Sands, supesses suglinoks	Straw-colored loesses	Straw-colored loesses, supesses, sands	I Nad-poimen terrace	Upper
					Vigochian	v		Weakly developed, brown forest-soils	Brown soils / Red-brown soils		Lower
					Udanian	ud		Shallow straw-colored loesses	Straw-colored loesses; brownish, straw-colored light gray sands	II	Upper
		Mikulinian		Middle Anthropogene An_2	Prelunian	p	Swamp podsolic soils	Brown-gray forest soils; podsolized, leached-out chernozems	Carbonate-chernozems / Grayish-brown soils		Lower
	Middle Quaternary Q_{II}	Moscovian			Tyaminian	ts	Sands, supesses, clays, moraines	Brownish, straw-colored loesses	Brownish, straw-colored loesses; light-brown supesses, sands	III	Upper
		Odinian			Kaidakian	k	Marsh podgolic soils	Marsh, light gray, gray and brown, forest soils	Leached-out podsolized chernozems / Carbonate chernozems		Lower
		Dneiperian			Dneiperian	dn	Sands, clays	Glacial deposits	Straw-colored, and brownish straw-colored loesses, sands	IV	Upper
					Potyaganian	po	?	Shallow chernozem soils	Brown and reddish-brown soils		
		Likhvinian			Orelian	or	?	Shallow straw-colored loesses	Straw-colored, light brown, loesses		
				Early Anthropogene An_1	Zavadian	z	Brown forest soils	Brown forest soils / Brown meadow / Brown soils	Brown and red-brown soils		Lower
	Lower Quaternary Q_I	Okanian			Tilirulian	ti	Sands, suglinoks	Loessial moraines(?). Glacial lake suglinoks, with pebbly, crystalline soils	Straw-colored, light-brown loesses; sands	V	Upper
		Beloveshian			Lubenian	l	Chernozem-like forest meadow soils	Red-brown meadow soils	Reddish, light-brown soils		Lower
					Sulian	su	Sands, suglinoks	Straw-colored loesses	Straw-colored and light-brown loesses; sands	VI	Upper
		(?)			Martonovian	m	Marshy, podsolic soils	Brown meadow loess-like soils / Thick, brownish and reddish-brownish	Reddish-brownish soils		Lower
					Preazovian	pa	Sands, supesses, suglinoks	Straw-colored loesses	Straw-colored and light-brown loesses; sands	VII	Upper
NEOGENE N	Pliocene N_2 / Upper Pliocene N_2^3	Gurinian and Tamanian strata	Neogene N / Pliocene N_2	Late N_2^3	Shirokinian	sh	Light gray, brown, leached out, and with carbonates, soils, and clays with light clay seams		Red-brown soils and clays with loessial clays		Lower
					Ilicherian	i	Light clays with light-clay soils			VIII	Upper
					Kryzhanian	kr	Brown, leached out, soils and clays		Red-brown soils and clays		Lower
		Kuyalnikian formation: Upper Kuyal.			Berezanian	br	Light clays with light gray soils			IX	Upper
	Middle Pliocene N_2^2	Kuyalnikian formation: Lower Kuyal.		Middle N_2^2	Beregovian	b	Red-brown, leached out, soils and clays		Red-brown, with carbonates, soils and clays		Lower
		Kimmerian formation			Siberian	si	Light clays with light gray soils			X	Upper
					Middle Bogdanian	bm	Red-brown, leached out soils and clays	Red-brown, with carbonates, soils and clays	Red soils and clays with gypsum		Lower
					Andarian	a	Light clays with light gray soils			XI	Upper
					Early Bogdanian	bi	?	Red-brown, with carbonates, soils and clays	Red soils and clays, with carbonates		Lower
					Oskolian	os	Light clays with light gray soils			XII	Upper
	Lower Pliocene N_2^1	Pontian formation		Early N_2^1	Ivankovian	iv	?				Lower


Estuarine and marine deposits (dry land)
Estuarine terraces
Terrace deposits and alluvium
Sea and lake deposits
Vegetation and landscape
N ← S
Terrestrial Molluscs
Zone of present forest-steppes
Zone of present steppes
Mammalian fauna
Paleoliths
Frost deformations
Paleoclimate of the present-day forest-steppes
← Hot
Cold →
Subtropical climate
Temperate Climate
Periglacial loess-forming
Glaciation
Alpine scheme
Holocene ?
Contemporary early Chernozem forms
?
?
Late Euxine
Alluvial
Ist Terrace
Aldanian ?
Alluvial
II
Early Euxine–upper
Karangatian
Alluvial
Uzunlarian
III
Alluvial
IV
Middle Late-Euxine
Alluvial
V
Lower Late-Euxine
Alluvial
VI
Chanda
Alluvial
VII
VIII
Alluvial
Gurinian and Tamanian strata
IX
Alluvial
Upper Kuyalnik
Lower Kuyalnik
X
XI
Kimmerian strata
XII
Pontian strata
Holocene fauna
Mammoths
Primigenius
M. Trogon therii and M. Primigenius
?
M. Primigenius
Archidiscodon Wusti ?
M. Primigenius + A. Wusti
A. Wusti
Tamanian fauna
Khaprovian (Noga) fauna
Odessan fauna
Kuyalnikian fauna
Moldavian fauna
Neoliths and others
Azil-Magdal. Solyutre Orinyak
Moust.
Late Ashel
Ashel
Ashel ?
?
?
H
W_3
$W_{2\text{-}3}$
W_2
$W_{1\text{-}2}$
W_1
R W
R_3
$R_{2\text{-}3}$
R_2
$R_{1\text{-}2}$
R_1
M - R
M_3
$M_{2\text{-}3}$
M_2
$M_{1\text{-}2}$
M_1
G - M
G_2
$G_{1\text{-}2}$
G_1
D - G
Stages of Danubian glaciation
Note:
In the table soil data are for optimum climatic conditions
Zones of Landscapes and Vegetation Molluscs
Glacial and Preglacial Climates
Land ice sheets
Periglacial "steppes"
Temperate climates
Mixed forest
Forest and Meadow steppes
Steppes
Arid steppes
Subtropical climates
Wet forest
Forest and Meadow steppes
Steppes with bushes; moderate Continental climate
Arid steppes
Arid steppes interspersed with steppes of moderate climate
Molluscs
Modern types
Mixture of a large amount of boreal molluscs
Mixture with a moderate amount of boreal molluscs
Mixture with a small amount of boreal molluscs
Modern types with some larger heat-loving forms
Modern types with significant amount of heat-loving forms

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CRYOGENOUS TEXTURE OF SOILS IN THE LAYER OF SEASONAL THAWING

E. A. VTYURINA Pacific Institute of Geography, Far Eastern Scientific Center, Academy of Sciences of the USSR.

Cryogenous texture may be complex, if the class of the cryotexture varies throughout the cross section, or simple, if the class remains unchanged (see Figure 1).

Complex cryogenous texture results from two factors: the type of freezing and a nonuniform composition of the seasonally thawing layer. Since supes and suglinok predominate in this layer, the type of freezing is the more important factor, i.e., the differences in the direction of freezing (whether from one or two sides) and its relative timing (first from above and then from below). The regular pattern of freezing of the seasonally thawing layer, which obeys the laws of latitudinal and altitudinal zonality, brings about equally strict relationships in the complex cryogenous texture of this layer. Therefore, we have termed the cryogenous texture of the seasonally thawing layer containing supeses and suglinoks a "regularly alternating or regular texture." This is the main subclass of the complex cryogenous texture that occurs most frequently owing to the predominance of the above-mentioned soils and in the seasonally thawing layer. There are four types of this texture corresponding to the main types of freezing:

1. Streaky-massive texture, which is formed if the layer freezes from above only. It is characteristic mainly of the southern part of the region with seasonally thawed ground where the permafrost temperature does not exceed -0.5°C.
2. Streaky-massive-streaky texture formed if freezing proceeds from two sides, but only if freezing from below starts simultaneously with or later than freezing from above. It is characteristic of the central part of the region, where the temperature of permafrost ranges from -0.5°C to -5°C.
3. Streaky-massive-streaky-ataxitic texture formed if freezing from below preceded that from above. It is characteristic of the northern part of the region, where the permafrost temperature is below -5°C.
4. Streaky-ataxitic texture formed if freezing from below precedes that from above and the thickness of the layer does not exceed 50 cm. It is characteristic of the northernmost part of the region.

The second reason for the complexity of cryogenous textures of the seasonally thawing layer is the nonuniformity of its composition and the alternation of soils that are capable or incapable of ice segretation. We have termed the complex cryogenous texture resulting from an abrupt change in the soil composition "the irregularly alternating or irregular texture." There are four main types of irregular cryogenous texture, which are as follows.

1. Massive-streaky type characteristic of a

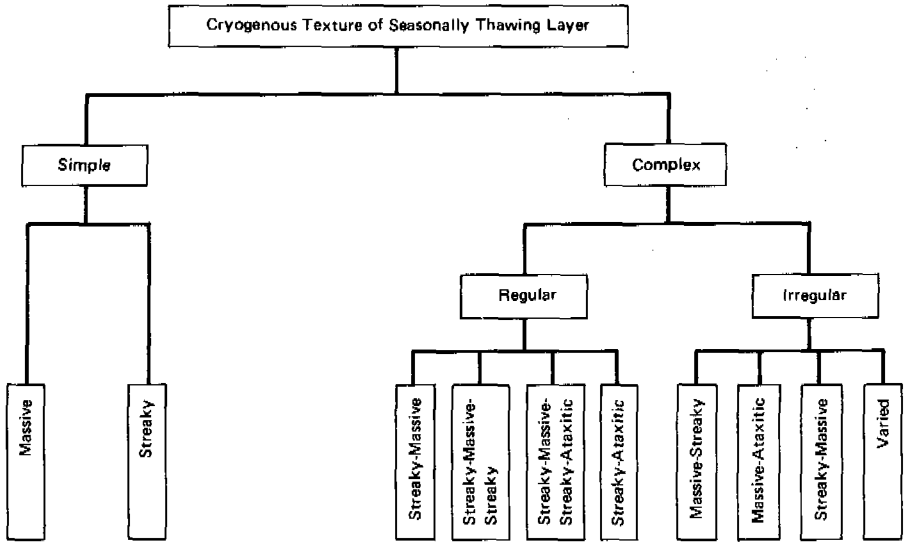

FIGURE 1

seasonally thawing layer, the upper part of which consists of slightly decomposed peat or coarse-grained soils. It is frequently formed in the zone of unilateral freezing and also in places where freezing from below starts later than that from above.

2. Massive-ataxitic type formed in a layer of the same composition as before but only if freezing from below precedes that from above. It is characteristic of the northern part of a region.

3. Streaky-massive type formed in a seasonally thawing layer, the lower half of which consists of soils incapable of segregational ice formation. It is found in the zone where the layer freezes from two sides.

4. Varied type formed in a layer consisting of alternating bands of soils that are capable consisting of alternating bands of soils that are capable or incapable of segregational ice formation. It can be found in any zone.

Although the complex nature of cryogenous textures is one of the characteristics of the seasonally thawing layer, the latter often contains only one class of texture. Frequently occurring types are:

1. Streaky type, which is characteristic of layers consisting of uniformly, and strongly-peaty, supeses and suglinoks, strongly-silty peat, and fractured bedrock. It can be found in any zone. In the southern part of the region, it is characteristic also of supeses and suglinoks without peat, which freeze only from above and have an additional supply of water.

2. Massive type which is characteristic of layers consisting entirely of sand and coarse-grained soils and without any fine-grained material. It can be found in any zone. Wet soils consisting of gravel, pebbles, and rock waste in the Far North and on sections where freezing from below precedes that from above may have a massive-basal cryogenous texture, which in our opinion should be regarded as a complex irregular texture.

The cryogenous texture is one of the most important characteristics of the seasonally thawing layer, the study of which helps to understand the reasons behind its formation. The studies of classes, subclasses, and types of texture indicate that this characteristic can and should be used as one of the criteria in the classification and mapping of the seasonally thawing layer. Previous attempts to map the textures of the seasonally thawing layer have not been successful, since they were limited to showing a single cryotexture as being typical of a given region or section.[3] However, this is acceptable only on mapping sections of the seasonally thawing layer with a simple texture. In the case of a complex texture, all cryotextures found throughout the cross section are equally typical, and, therefore, the indication of only one of them as the most important is incorrect and misleading.

Considerations of the class, subclass, and type of cryogenous texture help to show the diversity of the structure of the seasonally thawing layer correctly and more fully.[2] It is especially important to study and map the types of regular textures, since they indicate not only the distribution of ice in the soils of the seasonally thawing layer, but also its thickness, type of freezing, direction of water migration during freezing, and temperature of permafrost in a given section.

Knowledge of the main relationships in the cryogenous texture of the seasonally thawing layer and of factors that determine the formation of its various types helped us to develop the cryotextural method of studying permafrost.[1] Use of this method in conjunction with other paleogeographical methods makes it possible to reconstruct the paleogeocryological conditions of formation of permafrost, especially its syngenetic variety, which is formed as a result of the transition of the lower half of the seasonally thawing layer to a perennially frozen state as the accumulation of sediments increases.

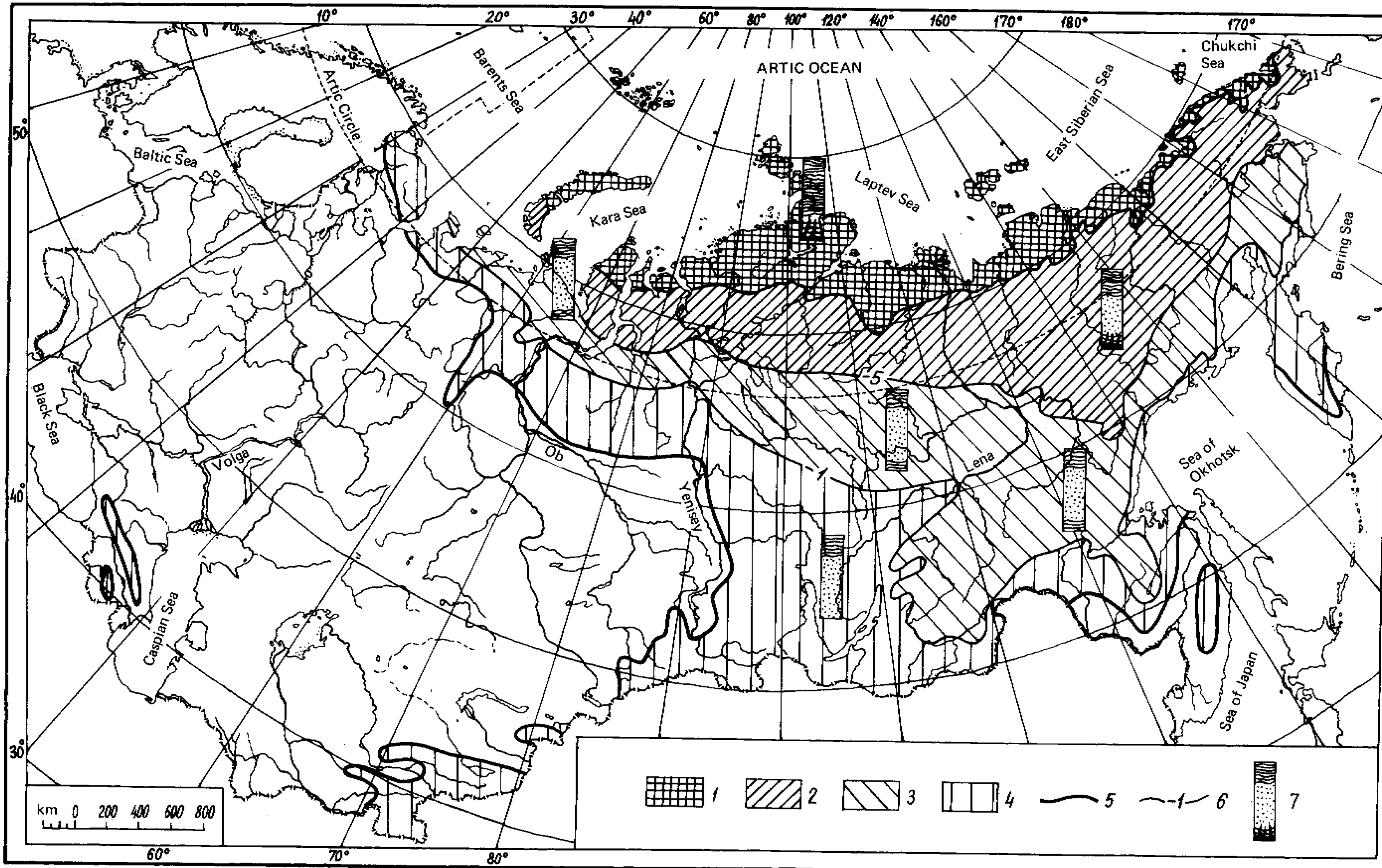

FIGURE 2 Schematic map showing the types of complex regular cryogenous textures of soils in the seasonally thawing layer. Compiled by E. A. Vtyurina. 1--streaky-ataxitic type, 2--streaky-massive-streaky-ataxitic type, 3--streaky-massive-streaky, 4--mostly streaky-massive, 5--the boundary of the region of seasonally frozen ground, 6--the boundaries of zones with different types of cryogenous textures, and 7--cross section showing the type of regular cryogenous texture of the seasonally thawing layer.

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SELF-REGULATING MECHANISM OF SIZE-LIMITED ICE WEDGES

SH. SH. GASANOV North Eastern Integrated Scientific Research Institute

The theory of secondary ice-wedge formation is among the number of well-studied problems of cryolithology. Nevertheless, a number of aspects of the process remain as yet unclear. Among these can be placed the possibility of limitations to the growth of ice wedges in width and the factors that regulate the process.

In publications on cryolithology, this issue is considered generally from the point of view of changes in external conditions. Thus, P. A. Shumski[6] linked the termination of the growth of the ice wedges to the exit of the terrace from the floodplain or to the strengthening of the slope. In contrast to this, S. V. Tomirdiaro[5] considered that the process of ice-wedge formation is at a certain point terminated by thermokarst. However, these views do not seem to be confirmed by observations carried out under natural conditions and cannot be considered to be of universal application, since the ice wedges do become formed on stable slopes and in other elements of the relief where thermokarst does not develop.

B. N. Dostovalov[3] pointed out the gradual change in physical and mechanical properties within the system, ice-frozen sediments occurring during the growth of ice wedges. It is assumed that in the process of growth of the wedges there occurs a gradual increase in the compactness of the system to such an extent that the cracks either do not develop or, when they are formed, they are so thin that water does not penetrate into them.

On the whole the pattern of reasoning is correct, except for the reasons that cause the decline in the thickness of the cracks. It is a well-known fact that the relative compression of supesses and suglinoks within the range of negative temperatures is only 3-5 percent, i.e., lower than the volume of macro-ice saturation within ice wedges almost of an order of a magnitude.[1] The excess of soil is squeezed out along the wedges to the seasonally thawed layer in a form of little rolls, and to a considerable extent it becomes diffused in the course of gravitational consolidation in the internal-polygonal space.[4] In this way there is practically complete relief of stresses that develop in the course of the growth of the wedges laterally.

A similar solution to this problem can be obtained, in our opinion, by considering the phenomenon from the point of view of the pattern of size changes in frost-induced wedges with the increase in the size of a wedge as a result of the drop in the combined coefficient of thermal expansion of ice and the frozen soils.

Such a relationship can be expressed in the following way:

$$l = \Delta t \sum_{i=1}^{t=2} \alpha_i L_i,$$

where Δt = difference in temperature between the maximum and at a given time at a given depth, α = combined coefficient of thermal expansion of ice and soils, and L = width (on the surface) of ice and soil in the elementary ice-polygonal block.

In the expression given above, Δt is a constant, while the remaining two are variable.

Approximate calculations of the change in the width of ice-induced cracks l, depending upon the width of the ice wedge, were made using the formula $i = \Delta t \alpha L$. In the course of the calculations, data supplied by I. N. Votiakov, S. E. Grechishchev, E. P. Shushirina, and others on temperature deformation of ice and of frozen supesses and suglinoks with a moisture content of 30-40 percent (respectively, 50×10^{-6} and 80 to 100 $\times 10^{-6}$) were used.

The calculations show that, with the growth of the ice wedge laterally, there occurs a gradual but at the same time consistent change in the linear dimensions of the frost-induced crack. With the increase in the size of the wedge, the crack becomes narrower and the depth of its penetration becomes less (Figure 1). It is essential, how-

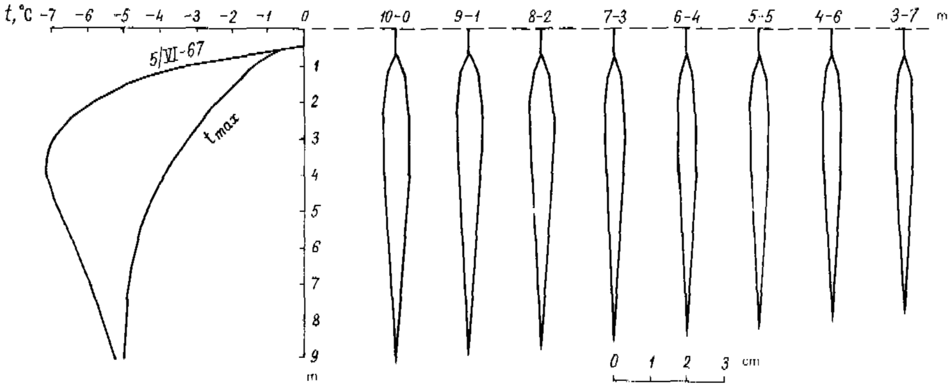

FIGURE 1 Changes in the size in the frost-induced crack, which has a depth depending upon the width of the ice wedge and the difference in temperature between the maximum and at the given time (June 5, 1957). The numbers above the frost-induced crack (10-0) denote: the first one--the width of the core of the polygon; second--width of the ice wedge. Temperature data from V. K. Riabchun for the Anadyr region.

ever, to pay attention here to two things. First, the calculations do not include the relaxation time of the stress, the temperature aftereffect, ice sublimation and convective heat exchange in the crack, expenditure of stress on reversible processes of compression-decompression, and other heterogeneously directed processes that modify the morphology of the crack, while retaining the general tendencies of its development. Second, it has been noted in the course of observations (B. I. Vtiurin, P. A. Shumski, and others) that with the increase in size of the wedge, the depth of penetration of the cracks does not diminish, as would follow from our calculations, but, on the contrary, it increases. The penetration of the cracks to a depth that is not caused by the magnitude of temperature deformation can probably be explained by the brittle properties of ice at low temperatures.

As we know, the main source of the infilling of the cracks that have been formed are melted snow and flood waters. In spring, water that has a temperature of +0.5-1.0°C begins gradually to seep through along both sides of the crack into the body of the wedge, where it becomes frozen. The front of the water, which is seeping down, has a characteristic bend in its profile caused by the phenomenon of hysteresis of wetting (Figure 2a). The water displaces air that flows upward along the edges of the crack. The penetration of water is accompanied by heat losses on partial melting of sublimation ice and the heating of the walls of the crack. Such is generally the course of the development of the process when the frost-induced crack is sufficiently thick.

Since, with the increase in the width of the wedge the crack becomes narrower at a certain critical width, there occurs a qualitative change in the process. The water that seeps along the walls of the crack within the seasonally thawed layer now fills the crack (see Figure 2b). There is then a sharp drop in the speed of penetration of water downward on account of the resistance of trapped air in the crack, the surface tension of the lower meniscus, an increase in the viscosity of water (up to 1.7 to $1.8 \cdot 10^{-2}$ at temperatures of +0.5-2.0°C), and also the mechanical resistance of the sublimed ice crystal on the walls of the crack which are close together.

As a result, the water before it has time to fill the entire crack becomes crystallized within the confines of the seasonally thawing layer and in the cover of the ice wedge, thus forming an ice plug.[2] The thawing of the seasonal layer and of the plug (May to June) is accompanied by the warming up of the entire mass of frozen soil and by the closing of the walls of the crack. By the time of the complete melting of the ice plug (August-September), the crack is closed completely, and water has no further access to it.

Thus, from the analysis of the mechanism, we can conclude that in the process of ice-wedge formation there is a gradual accumulation of factors, which at a certain stage reach a critical point that excludes further lateral growth of the wedge. Only by an automatic mechanism can facts be logically explained, facts showing

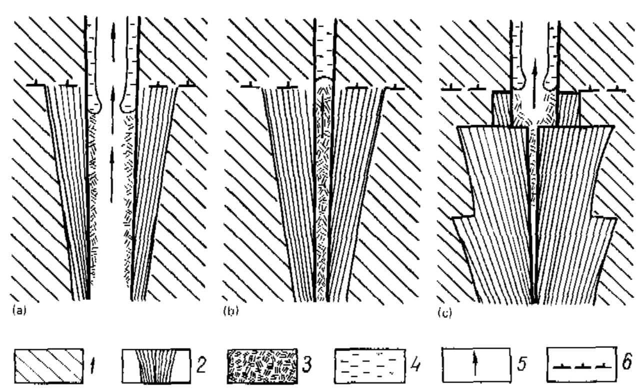

FIGURE 2 Conditions of penetration of water into a frost-induced crack depending upon the stage of ice-wedge formation. (a) Free access of water into the crack at the initial stage of ice formation. (b) Access of water is difficult in the final stage. (c) Renewed growth of the wedge at the new level and the outline of the cross section of the crack in the new system of a two-level wedge: 1--soil in which it is located; 2--ice wedge; 3--crystals of sublimed ice in the walls of the crack; 4--water that is seeping into the crack; 5--direction of movement of air out of the crack, air that is replaced by water; 6--floor of seasonally thawing layer.

the unchanging existence of ice wedges over periods extending through many tens of thousands of years--wedges that are located immediately underneath the seasonally thawing layer. Otherwise, with any rate of growth the ice wedges must have become fused in their upper part, forming a continuous ice shield underneath the seasonally thawing layer.

From the aspects of the self-regulatory mechanism presented above, can probably be also explained a number of issues discussed within the framework of the theory of the formation of thick syngenetic wedges.

Numerous studies (A. I. Popov, P. A. Shumski, B. I. Vtiurin, E. M. Katasonov, N. N. Romanovski, and others) have established a characteristic bend in the cryolithological layer formation in the cores of polygons and the corresponding bending of the side contacts of wedges, which indicate in an aggregate a rhythmic pattern of processes of ice and soil formation in the course of their vertical development with the unchanged maintenance of the polygonal thaw zone on the surface.

This type of rhythm of ice formation can probably be explained also independently of the pattern of soil formation, using the self-regulatory mechanism of the growth of ice wedges, which has been considered above. As we know, the rates of growth of wedges in lateral direction are somewhat greater (up to 1 cm/yr) than the rate of accumulation of soils on the surface of the thawing zone. As a result, at a certain point the process limits temperature deformation for ice-wedge formation, and the lateral growth of the wedge gradually dies down.

The renewal of intensive growth of ice wedges laterally occurs after the accumulation of sediments of a certain thickness and the corresponding increase in the cover of the frozen soils (Figure 2c). In horizontal cross section, the frost-induced crack has specific features that are conditioned by the varying values of the coefficient of thermal expansion of newly accumulated sediments and the underlying system of ice formation. In accordance with this, the rate of growth of wedges laterally in the newly deposited sediments and in the underlying wedge will also be differentiated. As a result of the cyclic repetition of this process in the cross section, there appears to occur a system of many-layered ice wedges superimposed one upon another, which have a steplike side contacts and rhythmically repeated series of unequally deformed cryolithological layers within the cores of the polygons. Such a cyclical nature of the process of ice formation is complicated by the superimposed rhythmic pattern of accumulation of sedimentation, which would cause certain deviations in the structure of the cross sections, as compared with the pattern presented above.

The mechanism of syngenetic ice formation, which has been considered here, while it explains the peculiar rhythmic nature of the formation of the veins and of the soils in which they are imbedded, does not, unfortunately, clarify all the issues arising in the complex problem of the mechanism of formation of thick syngenetic wedges.

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ICE FORMATION DURING THE DEVELOPMENT OF EPIGENETIC PERMAFROST

T. N. ZHESTKOVA Moscow Lomonsov State University

Soviet and foreign investigators have determined that the most important factors that govern the process of ice formation in the soils are the state, water content (and composition of the latter), and the rate of movement of the freezing front.

Studies of the ice content in cross sections cutting through the entire thickness of Quaternary deposits and comparisons of field observations with experimental data have confirmed the existing ideas concerning the effect of these factors on ice formation and formation of textures. At the same time, it was established that the ability of soils with a certain grain-size distribution and of certain chemical and mineralogical composition to form ice does not always determine the nature of ice saturation in a frozen massif. As a rule, changes in the mineralogical and chemical composition and in the grain-size distribution of fine-grained soil within a single genetic group of sediments are not reflected in their ice content and cryogenous texture. The dependence of the ice content of soils on their composition is very distinct in the uppermost permafrost horizons only and becomes less so with depth.

Increase or decrease in the ice content of uniform sediments at various depths and changes of cryogenous textures in the cross section are sometimes quite unrelated to changes in the composition of the sediments, just as changes in the ratio of critical fractions in the suglinoks and clays have practically no effect on the ice distribution in them.

In cross sections composed of heterogeneous soils, a nonuniform [degree of] ice saturation throughout the depth usually results from the presence in fine-grained soils of sandy or gravel layers the thickness of which exceeds the height of the capillary rise in them. However, even in this case, there are sections with an increased ice content whose presence cannot be explained by the composition of the soil. Field observations and calculations based on the diffusion equation have shown that changes in the freezing rate may lead to formation of ice-rich horizons in soils of uniform composition. In an open system, a nonuniform ice content throughout the depth may result from a change in the soil composition, the presence of water-bearing horizons in freezing sediments, and a variable regime of freezing on the surface.

In the case of the freezing of a closed system, the degree of ice saturation depends first of all on the water-retaining ability of soils and their permeability, while in an open system ice content depends not so much on the composition and the water-physical properties of soils as on the presence of water-bearing horizons, the abundance of water in the latter, and the water pressure in them.

Experimental freezing of water-saturated soils at different surface temperatures of samples have shown that the water content and the critical density of soils are not constant and depend on the conditions of freezing. There are several critical water contents and critical densities for homogeneous soils in a single genetic group of sediments. As the rate of movement of the freezing front increases, the absolute critical water content at which formation of ice lenses is possible increases, while the density decreases.

At very low rates of movement of the freezing front, the segregation ice formation is possible if the water content of soil is below that at the plasticity limit. This was confirmed by data obtained in deep cross sections.

Experimental freezing of suglinoks and clays with different initial water contents and under conditions of a closed system has shown that a change in cryogenous textures and their distribution throughout the depth of a homogeneous soil mass depends (with all other conditions remaining the same) on the rate of movement of the freezing front. For soils of different compositions and with different initial water contents, there are specific limits of the rate of movement of the freezing front. Changes in cryogenous textures occur above and below these limits. Formation of layered textures is possible only at very low rates of movement of the freezing front in soils with a high water content. Under the same conditions of the heat exchange, latticed and massive cryogenous textures are formed in soils with a low water content. Further growth of ice lenses and lattices occurs in the frozen zone and depends on the time of formation of permafrost.

Lately, the process of water migration and ice formation has been used to explain the cryogenous structure of thick permafrost layers.

Quantitative and qualitative descriptions of water migration towards the freezing front are based on the theory of moisture transfer in capillary-porous colloidal bodies. The main importance is attached to the migration of liquid water due to the moisture gradient in the thawed layer, which depends on the water content at the freezing boundary. The thermo-osmotic liquid water and vapor-diffusion water are ignored. The migration of water in a frozen layer is not examined either.

The critical water content is introduced as a limiting parameter at the boundary of the thawed zone. On the basis of these assumptions, the study of water migration makes it possible to determine the factors that affect ice formation in seasonally frozen soils and permafrost. Of these, the most important are the freezing rate, the temperature changes during freezing, and the depth of occurrence of water-bearing horizons. However, all these assumptions do not satisfy the contemporary ideas concerning groundwater and the mechanism of its movement, on the one hand, and do not explain a number of natural phenomena, on the other.

According to contemporary views, the movement in the liquid phase occurs mainly by the migration of film water. Free water compensates for the loss of film water, and the greater the amount of free water in freezing soil, the faster the migration of water and hence the greater the volume of moving water. Therefore, one may talk merely about a faster or a slower rate of water migration at a water content above or below the plastic limit, but not about the cessation of water migration. The validity of data on the critical water content and of experiments to determine the latter is limited by the duration of the experiments, the longest of which are those under natural conditions, and they are limited by the time of formation of seasonally frozen, or seasonally thawed, layers. In contrast to these conditions, permafrost is being formed over periods of thousands of years and during this time even the slowest process may result in considerable changes in the phase composition of soils. Freezing of soils in the active layer, and especially the experimental freezing of samples, occurs so rapidly that in a number of cases the soils cannot be regarded as models of long-term freezing, since quasi-equilibrium processes may be mistaken for the state of equilibrium.

It is known that the density and porosity of soils that passed through a prolonged stage of sedimentation and diagenesis decrease, and that such soils can be completely saturated with water at water contents equal to or below the plastic limit. In this case, the movement of film water, which is not compensated for by free water, occurs very slowly, and this leads to a considerable settlement of soils. Desiccation fissures formed in this way probably serve as a basis for the development of latticed cryogenous textures.

With allowances for the above, it is possible to suggest the following mechanism of ice formation in the case of epigenetic freezing of soil. In the upper part of the soil mass, where the water content is above the plastic limit, the freezing front moves toward the thawed zone, at the boundary of which the water content is equal to the critical value. Freezing of water at this boundary results in formation of primary cryogenous textures. A certain amount of unfrozen water remains in frozen soil and, when the soil temperature falls, migration of this water leads to the formation of secondary cryogenous textures with thinner ice lenses. The secondary texture appears to be included in the primary texture. The fraction of primary ice and the number of primary lens textures diminish with the rate of movement of the freezing front, with the decrease in the porosity of soils, and with a decrease in their initial water content.

If the soil is regarded as a continuous homogeneous medium, allowances should be made for the drop in the freezing point due to the pressure of overlying soils in accordance with the Clausius-Clapeyron equation. However, cavities and primary fissures present in the soils may serve as sites for ice accumulation. As was noted earlier, a network of fissures for accommodating cryogenous textures resulting from the desiccation of soils is essential at low water contents.

In the studies dealing with the forms of transfer in capillary-porous and colloidal bodies, particularly in freezing soils, vapor transfer, and thermo-osmotic and pseudo-osmotic migration of water due to the gradients in the salt concentrations are regarded to be of secondary importance.

Considering the long time required for the formation of permafrost, it may be assumed that these forms of transfer play an important role in the redistribution of water. Their precise role is still unknown, but all of them tend to dewater the lower horizons and increase the ice content in the upper horizons. The same can be said of sublimation and condensation in the frozen layer. Over a long period of time, this process should result in a certain reduction of layered and latticed cryogenous textures and an increase in massive textures with partial filling of soil pores with ice.

If the migration process involves the adsorption water as well, then such migration desorption must be accompanied by heat absorption, rapid cooling of soil, and accelerated freezing.

In unconsolidated deposits of limited thickness (i.e., in a closed system), the migration process during freezing always reduces the time of formation of permafrost. This is due to the temperature gradient throughout the depth of freezing soil.

In an open system, water migration prolongs the period of formation of permafrost.

CRYOGENOUS STRUCTURE OF QUATERNARY DEPOSITS ON THE YAMAL PENINSULA

M. M. KOREISHA Production and Research Institute for Engineering and Exploration in Construction

In the studies of formation of ice-rich permafrost, a distinction is made between epigenetic and syngenetic permafrost, which differ with respect to the time of sediment accumulation and freezing. The latest findings indicate that there are numerous forms of cryogenous structures within the two genetic types of deposits, and in each case the cryogenous structure is related to the specific conditions of sedimentation and freezing. It is not always possible to explain various aspects of the cryogenous structure or even decide to which genetic type a given permafrost formation belongs.

The marked success of the permafrost-facies analysis suggested by E. M. Katasonov[2] is due to the fact that, in the case of typical syngenesis, freezing of sediments from below is sufficiently uniform from the point of view of thermal physics. Therefore, it was possible to develop a geological method of investigations, leaving aside the physical aspects of ice crystallization.

The application of the permafrost-facies analysis is still restricted, in spite of its obvious advantages, because the role of freezing in the formation of facies is rarely known.

The study of the cryogenous structure of different facies of Quaternary deposits on the Yamal Peninsula revealed such a complex and sometimes confusing picture that the available methods of investigation were clearly inadequate to provide the necessary answers. Our method of studying the cryogenous structure was not very successful either.[3] The dependencies of numerical parameters of the ice structure and permafrost texture determined in the field on various properties of permafrost could not be explained unambiguously.[1]

We shall cite a few examples of complex cryogenous structures from the area around the Nurma-Yakha River on the Yamal Peninsula.

The cryogenous structure of marine upper Pleistocene (Kasantsev) deposits, the freezing of which was obviously of the epigenetic type, was related to the interaction between large ice sheets and an overlying system of ice inclusions that formed latticed and layered-latticed cryogenous textures. In one cross section through the sediments of the IV marine terrace, the overall thickness of which exceeded 30 m, an almost horizontal, slightly downwarping ice wedge (more precisely ice soil) between 1 m and 2.5 m in thickness, was resting at the base of a suglinok layer underlain by sand (Figure 1). The cryogenous structure of the overlying suglinoks was in some way related to the ice wedge. Their water content (without visible ice inclusions) increased from 25 percent in the upper part of the cross section to 35-45 percent at the ice soil. The unit ice content of the latter reached 60-70 percent. While the distribution of mineral impurities in the horizontal ice was very uneven, in the vertical ice inclusions higher up (2-5 cm in thickness and 2 m or more in length), the distribution of sand and clay particles was more regular. The average dimensions of ice crystals in the vertical inclusions almost doubled in the downward direction, while the structure became more uniform.

Just 100 m from the described cross section, the same suglinoks contained a system of bent

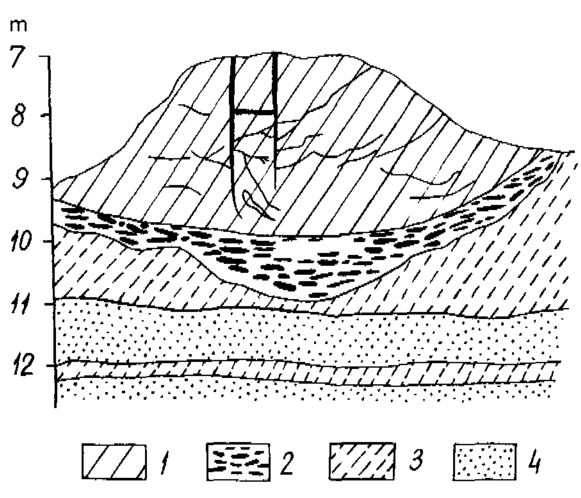

FIGURE 1 Cryogenous structure of Kazantsev deposits, IV marine terrace, the Nurma-Yakha River (the Yamal Peninsula). 1--suglinok, 2--ice-soil, 3--supes, 4--sand. Vertical scale indicates the altitude from the top of the terrace, the relative altitude of which is 40 m.

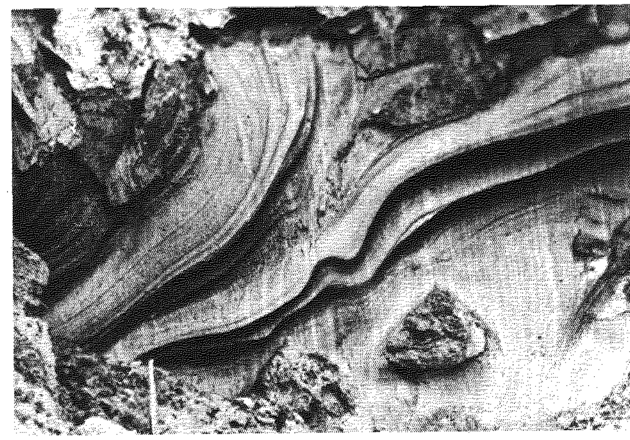

FIGURE 2 Ice in the Kasantsev deposits of the IV marine terrace. Inclusions of supes with a low ice content (>1 m in diameter) may be seen below bent ice layers containing mineral impurities.

ice and ice-soil layers of very peculiar and complex shapes resting directly on the sandy layer (Figure 2). The cryogenous texture of accommodating suglinoks was not related in any way to the ice layers except a narrow band along the contacts. The texture of bent ice sheets was very nonuniform along the strike. However, some specific features were noted also. The sandy particles were distributed in the ice more evenly than clay and silt particles that formed round or elongated inclusions measuring between 5 mm and 100 mm. The thickness of individual ice layers ranged from 0.1 m to 1 m. There were fewer mineral impurities in the middle part of each layer, so that these parts formed protruding ribs in the outcrops (Figure 2). The contact between the ice and the suglinoks varied from well defined to very indistinct, but in all cases a more or less gradual transition from single round suglinok inclusions in the ice to a denser arrangement of rectangular inclusions and then to the latticed cryogenous texture was clearly visible.

The shape, size, and texture of ice and ice-soil layers indicated that the latter were formed with the participation of water injections containing mineral particles (probably secondary) from the underlying water-bearing sandy horizon. Freezing after injection led to a differentiation of ice and mineral particles due to segregation processes. However, such an explanation is provisional. The terms "segregation" and "injection" are themselves provisional, since their mechanism is not clear even in simple cases. In our opinion, the discussion on the segregation or injection genesis of ice layers is not supported by a generally accepted theory of formation and development of segregation and injection ice. The same cautious approach should be adopted also toward the texture-forming ice, the purely segregation (migration) genesis of which is doubtful.

The study of the cryogenous structure of the upper Pleistocene alluvial-lacustrine sediments

(I-III terraces above the floodplain) encountered its own specific difficulties.

The alluvial-lacustrine sediments in the III terrace represent a complex interlayering of relatively thin (0.2-1.0 m) horizons of supeses, sands, and suglinoks. A still thinner, banded stratification occurs also. The general nature of the cryogenous texture and the genesis of such sediments do not indicate with any certainty either the epigenetic or syngenetic type of freezing. The textures of some suglinok horizons, which probably belong to the floodplain alluvial facies, are very distinct and are typical of sediments of the given type. The textures of sandy and supes horizons are not very clear, or are indistinct altogether. On examining the lithological differences in such cross sections, it is sometimes possible to detect some features of the cryogenous texture characteristic of epigenetic freezing. For example, the thickness of ice inclusions and the cells of the latticed texture in a homogeneous suglinok layer located between horizons of different compositions increase with depth. Frequent interlayering of lithologically different sediments evidently obscures the genetic features of the freezing process and makes the texture of each horizon dependent on its composition.

The typical syngenetic ice-rich sediments in the outcrop of the II terrace on the Nurma-Yakha River contained small wedges (up to 0.5-1.0 m in thickness) with a different texture. Evidently, under conditions of an elevated floodplain, freezing of a deep talik below a lake (after drying of the latter) in the syngenetic bed of flood plain sediments proper may proceed along different lines. It may be either of syngenetic or epigenetic type and may occur both from above and from below.

Generally, it may be stated that the cryogenous texture of sediments reflects all types of freezing: climatic, local, and so forth. The annual relationships between the facies and freezing are complex and varied.

Further development of the permafrost-facies analysis and of other methods of studying the texture of permafrost requires knowledge of how this texture is related to conditions that created it. One of the methods might be the correlation of known or given conditions of freezing with the texture of permafrost by means of petrographic analysis of both the ice structure and the cryogenous texture. This would require model studies of freezing under laboratory and field conditions. Of course, the experiments with the use of petrographic methods do not exclude but presuppose a strictly physical analysis of freezing, water migration, and ice crystallization.

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CHANGES IN THE COMPOSITION AND PROPERTIES OF CLAY SOILS DURING FREEZING

F. N. LESHCHIKOV and T. G. RYASHCHENKO Institute of the Earth's Crust, Siberian Branch of the USSR Academy of Sciences

In the present stage of their development under conditions of a severly continental climate, the Quaternary clay deposits in the zone of hypergenesis in eastern Siberia are subjected to strong effects of seasonal freezing and thawing, which is reflected in their composition and properties.[1-3]

The experimental results of repeated (10, 25, and 50 cycles) freezing and thawing of 23 specimen of alluvial, talus, and residual talus soils of supes-suglinok composition taken at depths ranging from 0.8 to 6.0 m near Irkutsk, Bratsk, and Ust-Ilim made it possible to establish the main changes in their composition and physico-mechanical properties. The nature of these changes was determined by calculating the coefficient of variability (K_z), which is a ratio of the final value of a parameter to its initial value. The most distinct changes were as follows: reduction in the content of the sand fraction, increase in the amount of silt and clay particles, reduction in the content of water soluble salts (S_{BP}) and changes in their qualitative composition, increase in the number of mobile forms of silica and sesquioxides (S_{TP}), decrease in the capacity of the cation exchange (E), transformation of the mineral composition of the colloids in the deposits, changes in plastic properties (γ_c) and in swelling (n), and general reduction in the density of soil (see Figure 1).

Repeated freezing and thawing results not only in a disaggregation of sandy material, but also coagulation of clay and silt particles, as indi-

TABLE 1 Main Indicators of the Composition and Properties of Clay Soils during Repeated Freezing

Number of Specimen and Depth	Number of Cycles	Fractions, %			Coefficient of Aggregation, %	Salt Content, %		Clay Minerals	Plasticity Index, I_p	Expansion (n), %	Water Content (W), %	Dry Unit Weight (a), g/cm^3	Coefficient of Relative Settlement, σ_{pr} at 3 kg/cm^2
		Sandy	Silty	Clayey		Insoluable, S_{TP}	Water Soluable, S_{BP}						
	0	34.3	56.2	9.5	29.0	14.24	0.20	M,K	5.0	--	39.3	1.62	0
8,882	10	22.3	61.0	16.7	22.0	29.85	0.10	M,K (?)	12.8	--	43.5	1.53	0
3 m	25	29.0	59.5	11.5	27.8	27.19	0.11	M	8.0	--	42.0	1.58	0
	0	66.0	22.9	10.9	10.9	22.7	0.29	H,V (?)	2.1	--	44.2	1.50	0.010
6,658	10	35.0	51.0	14.0	8.3	29.3	0.12	M,H,K	9.7	--	47.6	1.42	0.001
3 m	25	39.8	46.8	13.4	13.9	29.1	0.18	V,H	6.3	--	47.7	1.42	0.070
	0	42.8	48.0	9.2	14.5	26.22	0.53	K,H	8.6	6.8	39.7	1.61	0.010
6,346	10	35.6	56.1	8.3	15.5	--	--	--	8.4	9.3	38.1	1.67	0.015
	25	37.4	54.8	7.8	16.7	--	--	--	5.3	4.9	38.3	1.67	0.023
2 m	50	41.6	49.9	8.5	17.8	26.04	0.16	M,K	8.4	11.3	40.3	1.61	0.032

NOTE: M = montmorillonite, K = kaolinite, H = hydromica, and V = vermiculite.

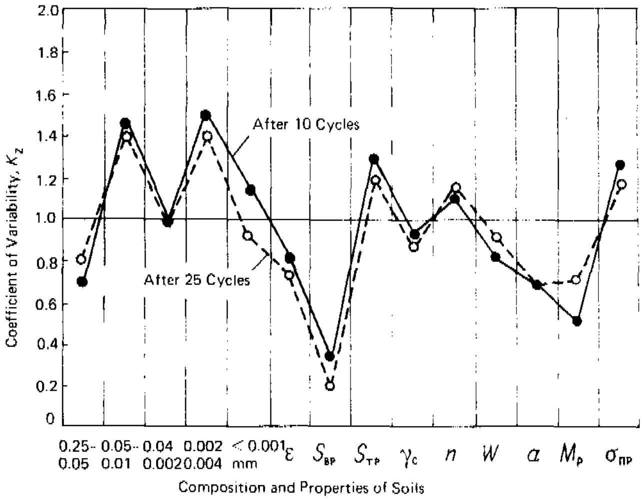

FIGURE 1 Coefficients of variability of the main indicators of the composition and properties of clay soils during repeated freezing. (a) coefficient of variability (K_z), (b) composition and properties of soils, (c) after 25 cycles, and (d) after 10 cycles.

cated by an increase in the aggregation coefficient after 25-50 freezing cycles (see Table 1). Coagulation reduces the free surface of particles and their exchange activity. At the same time, water, once free from surface effects, destroys the solid phase of soil still further.

The most significant changes occur in the salt content. For water-soluble compounds, the variability coefficients range from 0.2 to 0.6, while for insoluble compounds they are 1.2-2.6. Freezing out of salts is due to their transformation to insoluble compounds as a result of negative temperatures. The fact that these salts are enriched with mobile forms of silica indicates that chemical weathering of soils intensifies on repeated freezing and thawing.

The variability coefficients of physical and physicomechanical parameters of soils have relatively low absolute values. There is a tendency towards an increase in porosity. Changes in the water content may occur in either direction, and there is a direct relationship between water content and compressibility. The greatest changes occur in the plasticity and swelling of soils; it is known that these are dependent on the content of the clay fraction, its mineral composition, and the degree of aggregation of the soil.

Changes in plastic properties and swelling after freezing not only confirm that the grain-size composition of soils is altered, but also indicate possible changes in the mineral composition of their colloidal part.

Thermal and X-ray analyses of clay fractions in frozen specimens indicated stronger temperature effects and revealed the presence of new components, particularly kaolinite and montmorillonite (see Table 1). Qualitative changes in the clay fraction are related to its quantitative increase. Destruction of strongly pelitizised sand and silt grains in feldspars on freezing increased the kaolinite and montmorillonite contents in primary pelite.

During freezing and thawing the soils become less dense and this increases the leaching (M_p) and the absolute value of the coefficient of relative settlement (σ_{np}). The increase in the settlement is due to additional destruction of structural bonds in the soil, the strength of which depends on the content and composition of salts and clay components, which undergo the greatest changes on freezing. Hence, the change in the settlement characteristics is a logical result of physicochemical transformations of the soil substance, which occur at low temperatures.

Under natural conditions, a change in structural and textural characteristics of clay soils in the upper, 1-m-thick layer determines to a large extent the frost fissuring in winter and formation of dessiccation fissures in the forest-steppe areas on the Angara River in the first half of summer. These fissures accelerate the heat and water exchange, and this intensifies the cooling of soils in winter and heating and moistening in the summer. Due to fissuring, structureless soils are formed in a layer 50 cm in thickness.

The data obtained also explain certain other characteristics of the Quaternary clay cover in the southern part of eastern Siberia. Increase in the silt fraction and considerable settlement, which occur everywhere in the upper, 3-m-thick layer of soil, the aggregation, a low content of water-soluble salts and other characteristics, such as macroporosity and schistosity, are the result of seasonal freezing and thawing. All this should be considered during engineering and geological investigations of soil conditions in a given region.

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EFFECT OF CRYOGENESIS ON THE SOIL COVER AND SOIL PROFILE

E. M. NAUMOV V. V. Dokuchaev Soil Institute

Soil is a part of the lithosphere where cryogenous processes manifest themselves most strongly as a result of hydrothermal amplitudes and changes in the phase composition of water. Cryogenous processes take place in the soil under any low-temperature conditions (whether constant or temporary).

The effects of cryogenesis under different conditions are varied and manifest themselves in the sorting, cryoturbation, displacement, and mixing of soil; solifluction and frost fissuring; formation of polygons, thermokarst settlement, and heaving; transformation of the granulometric soil composition; various forms of migration of soil solutions; retention of surface water and solutions above the ice; dehydration and denaturing of chemical compounds; suppression of the biological activity, etc.[4,6,7] In the present paper, we examine the effect of cryogenesis on the soil cover and its profile in the continental part of the zone of maximum development of frozen materials of Pleistocene age.[2,6]

Universal development of polygonal wedge ice on the enormous northern coastal plains and in the adjacent river valleys of the northeast farther south is the determining factor of the lithogenesis. The enormous scale of the cryogenous lacustrine-thermokarst sedimentation makes it necessary to regard it as a special type of periglacial lithogenesis.[9] It has been established that the polygonal wedge ice on the Omolon-Anyui-Kolyma plain strongly affects the formation of relief and landscape. The uneven occurrence and thawing of ice wedges results in undulating (alass) relief[1,2,8] incorporating ridges and ravines. Our soil investigations on the Krestovaya River and the Omolon-Anyui-Kolyma plain have shown that permafrost, as one form of manifestation of cryogenesis, has a strong effect on the structure of the soil cover and soil profile.[5] Generally speaking, relatively more pronounced dry conditions of soil formation are created on the hills and ridges, while the ravines (alasses) become swampy, are covered with peat, and contain a large number of thermokarst lakes.

The soil cover on tops of ridges and hills is fairly complex. There is usually a combination of elevated surfaces that are polygons of various, often irregular, forms measuring 5-20 m (Figure 1). The polygons alternate with depressed sections which are lower by 50-100 cm or more), which usually stretch along the polygons and are about 1.5-5.0 m in width. Such depressions contain water, especially where they intersect. The depressions may surround a polygon from all sides. Published material[1,2] and our observations in the outcrops on the banks of the Kolyma and the M. Anyui rivers, along the coast of the East Siberian Sea, and in other places indicate that these elements of microrelief are the result of cryogenous processes. Evidently the elevated polygons are ice-rich earth "pillars" separated by thermokarst subsidence depressions on ice wedges. It was this type of "smallpox relief" of thermokarst origin that was described by permafrost investigators on flat summits and defined as the initial stage of the thermokarst destruction of swampy plains.[1]

The above-mentioned combination of two elements of relief leads to formation of two different landscapes (dry on the polygons and wet in thermokarst depressions) and determines the development of different soils. The surface of a polygon is nonuniform also. As a result of pronounced soil cryogenesis, the entire surface of a polygon is dissected by a network of fissures that separates and surrounds small mineral "pillars," i.e., small polygons (Figure 2). The latter are about 0.8-1.5 m in diameter, their form differs, and their height above the bottom of fissures ranges from 40 to 80 cm. The fissures are 30 to 50 cm in width (in the upper part). The small polygons are usually covered with lichen, and there is moss in the fissures. Water may be present below the peat in the fissures (usually where they intersect), while the polygons are, as a rule, dry at the top.

Hence the soil profiles of small polygons and fissures are vastly different. In the former a thin (2-5 cm) organic layer rests on mineral soil with various degress of gleyzation, which at a depth of 60-90 cm is replaced by frozen fine-grained soil containing veinlets and small lenses of ice (migration ice). Small polygons are dissected by fissures into which soil particles may fall in from the top. The soil profile has a prismatic-latticed-cellular texture (resulting from melting of ice veinlets in the fine-grained

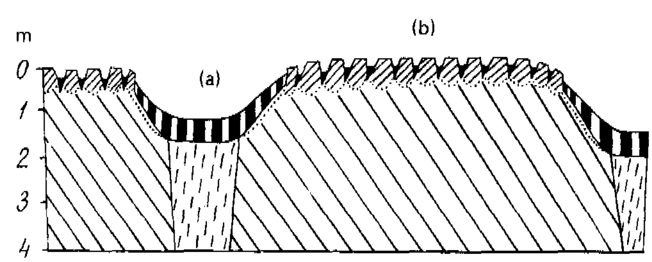

FIGURE 1 Schematic cross section through a cryogenous microcomplex: thermokarst settlement of soil (a) and polygon (b). For symbols, see Figure 2.

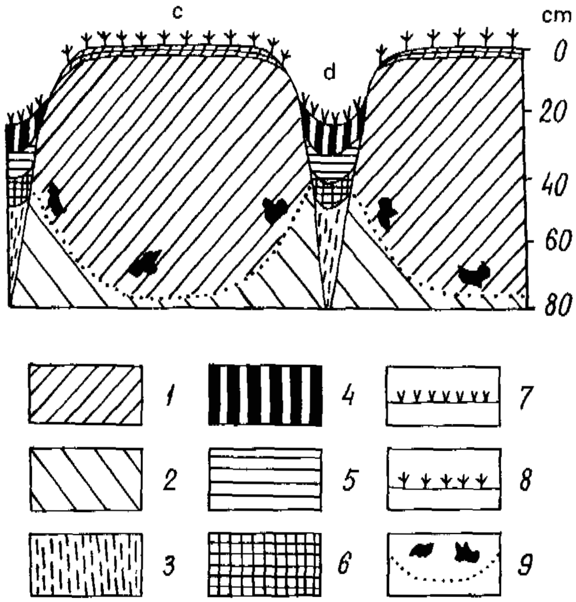

FIGURE 2 Schematic cross section through a cryogenous nanocomplex: small polygon (c) and frost fissure (d). 1--active layer, 2--frozen fine-grained soil with ice, 3--ice wedge in a fissure, 4--peaty horizon (A_0^1), 5--peaty horizon (A_0^2), 6--peaty organic horizon (A_0A_1), 7--moss cover, 8--lichen cover, and 9--fragments of organic horizons buried by cryoturbations and the lower boundary of seasonal thaw.

soil). The horizon is saturated with surface water. The profile rests on impermeable ice-rich fine-grained soil.

Therefore, the profile of a polygon represents a peculiar closed cryogenous system (cryogenous microcosm). The profile in a frost-fissure consists of two or three peat horizons (total thickness, 30-50 cm) with various degrees of decomposition. Ice (microvein) occurs at depths of 20-30 cm. Hence, cryogenous processes result in a peculiar combination of soils: marshy soils in places of thermokarst settlement and a cryogenous soil nanocomplex on the polygons. On soil maps, the areas with such pronounced cryogenesis must be indicated by contours with a complex soil cover (combinations, mosaics, etc.). Since a permafrost survey includes mapping of cryogenous processes and forms of permafrost relief,[3] it would be very expedient and even necessary to use permafrost maps on surveying the soils. This will result in a more accurate representation of the soil structure in permafrost regions. Although it is known that cryogenesis has a strong effect on soil processes and may even result in formation of soils with a specific permafrost regime,[4,7] this effect has not been adequately investigated. Therefore it is essential to continue the study of the relationship between cryogenesis and soil formation.

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FORMATION OF DIAMOND PLACERS UNDER CONDITIONS OF ARCTIC CLIMATE AND PERMAFROST

B. I. PROKOPCHUK Central Research Institute of Geological Exploration

Diamond placers formed under conditions of an arctic climate in permafrost regions are characterized by specific structural features, distribution of diamonds, and scale of occurrence.

A severe climate slows down considerably the destruction of kimberlite and intermediate collectors. They remain frozen for better part of the year and even the summer thaw does not extend beyond a depth of 1 m to 1.5 m. Physical and chemical weathering, release of diamonds, and formation of placers occur within this layer. Therefore, residual placers are small in area and depth. Furthermore, they contain a large amount of ice (up to 40-60 percent), which also reduces their diamond content. Absence of "yellow earth," i.e., the most disintegrated layer, is typical of residual placers in the Arctic. This layer is invariably present in temperate and tropical zones and is the most diamondiferous horizon in the cross section of a placer.

Destruction and transportation of diamond-bearing material on the slopes usually proceeds very slowly, since small temporary water-floods erode the upper thawing layer only. Therefore, placers on slopes are small also. They are usually found in association with solifluctional formations characterized by complex layering resulting from solifluction, freezing out of coarse fragmentary material, and other processes. Coarse fragmentary material and rock debris in which diamonds are concentrated occur on the brow of solifluction terraces. Solifluction enriches the alluvial placers with diamondiferous material. The latter is transported from the slopes and is later reworked by water. Other mechanisms of transporting this material from the slopes are of little importance. In the Arctic, residual and colluvial diamond placers formed as a result of destruction of bedrock deposits (kimberlite pipes) are small in area and depth and play a minor role in the formation of alluvial placers. Formation of large diamond placers in permafrost requires the presence of a primary source with a large washout area that compensates for the small thickness of the active layer. This requirement is satisfied by intermediate collectors rather than kimberlites, whose outcrops on the surface measure a few tenths of a square kilometer only.

The climate does not favor active transport of diamond-bearing materials by rivers either. In winter most rivers freeze right through or have an extremely low discharge. For the better part of summer, the rivers are shallow, so that coarse fragmentary material and the diamonds remain stationary on the riverbed. The discharge and the rate of flow increase abruptly during floods in the spring and fall. Transport of material and enrichment of placers occur during these brief periods of time (20 to 30 days). The material is transported for short distances, while the enrichment process involves small volumes of alluvium. Unconsolidated material near the channel is frequently frozen and does not move even during high floods. Therefore, alluvial placers display clear striations. The microstriae roughly corresponding to alluvium layers of various thickness are especially distinct, which is due to the morphology of the valleys.[3] The microstriae are related to the hydrodynamic regime characteristic of the arctic zones. Most diamonds are found in the microstriae of river sections where the rates of flow and the discharge are highest.

There are specific characteristics in the vertical distribution of diamonds as well. In residual and colluvial placers, the most enriched are the upper horizons, which is because of frost heave. Coarse fragmentary material and the diamonds migrate upwards. This was confirmed beyond doubt not only during exploration, but also during the development of the Yakut placers. The diamonds in thin alluvial beds (up to 1 m in thickness) are distributed evenly throughout the cross section. The richest horizons in thicker beds are the upper part and the horizon next to bedrock. Enrichment of the upper horizon is because, under conditions of arctic climate, only the upper alluvial beds are reworked by water. The lower parts remain frozen and are not enriched. The reasons for the enrichment of the horizon next to bedrock will be explained later.

It is important to point out an interesting phenomenon that favors the accumulation of diamonds and is related to deep freezing, i.e., formation of peculiar residual deposits below the riverbed. We noted that in the Lena diamond-bearing deposits in western Yakutia[1,2] the alluvial deposits on the riverbeds rest in some cases not on bedrock but on peculiar unconsolidated deposits that are genetically not related to alluvium. Their formation is due to destruction of bedrock (weathering) below the alluvium.

The structure of residual deposits below the riverbed is different on sections composed of different rocks and depends on the thickness and composition of overlying alluvial deposits, the depth and the regime of the river, and the time it took these deposits to form. The profile of residual deposits on carbonate rocks is as follows. On unweathered rocks there are usually three zones reflecting the stages of weathering of the initial rock: a zone of fissured limestones and dolomites 20-50 cm in thickness, which is gradually replaced by a 10-20-cm-thick detrital-clay zone, and finally a clay zone 10-30 cm in thickness.

The average thickness of residual deposits on carbonate rocks ranges from 0.3 m to 0.6 m. On sandstones these 0.3-m-thick deposits are represented by clay sand and on siltstone by clay. The characteristics of residual deposits are also dependent on the overlying alluvium. Their maximum thickness was observed where that of alluvial deposits did not exceed 1 m. They are thickest on Paleozoic carbonates and thinnest on Mesozoic sandy-clay formations. The residual deposits below the riverbed are thicker if the overlying alluvium consists of pebbles and are thinner if the latter is represented by sand. With all other conditions being equal, this difference is 2:1. The residual deposits are thicker in shallow creeks and in the rapids of large rivers that freeze in winter. They are absent in deep pools, oxbow lakes, and in sections without alluvial deposits.

All these facts lead to the conclusion that residual deposits below the riverbed are formed under conditions of Arctic climate in permafrost regions where seasonal thawing and freezing extend to bedrock. The water-filled fissures in the bedrock expand during frost weathering and this forms the detrital material. The latter is reworked by water in the summer. The annual occurrence of these processes leads to formation of residual clay deposits on carbonate rocks.

The residual deposits below the riverbed were examined in detail because the search for diamonds revealed that placers with similar hydrodynamic conditions and located at equal distances from the primary source contained vastly different amounts of diamonds. It was noted that the diamond content on sections with well-developed residual deposits of clay resting on bedrock was several times higher than on sections where such deposits were absent. The following are the results of sampling carried out in a 0.9-m-thick layer of alluvium on the Muna River (a left tributary of the Lena):

Sampling Interval, cm	Diamond Content, %
Alluvium	
0-30	2
30-60	16
60-90	20
Residual deposits on bedrock	
0-20	62

No such detailed layer-by-layer sampling was done on other rivers. However, it was noted on the Motorchuna River that more diamonds were present in the residual deposits than in the alluvial beds. It was found during exploration on the Molodo and Eekit rivers (left tributaries of the Lena) that concentrated samples taken from residual deposits below the riverbed contained much more heavy minerals including those found in association with diamonds than the samples from alluvial beds.

Therefore, the residual deposits below the riverbed may be regarded as diamond collectors. Due to their high specific weight, the diamonds sink through the alluvial bed and are trapped in the residual clay resting on bedrock. The residual deposits are formed only under conditions of severe climate in permafrost regions.

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SOME PATTERNS OF FORMATION OF THE HORIZON OF DISCONTINUOUS FROST WEATHERING

N. V. TUMEL AND YU. V. MUDROV Moscow Lomonosov State University

As is known, many investigators share the view that loesslike surface deposits in the contemporary periglacial zone are a product of frost weathering.

A complex combination of physicomechanical, physicochemical, and biogeochemical processes related to ice formation in the layer of seasonal thawing, i.e., the layer of discontinuous cryohypergenesis, predetermines the specific characteristics of soil weathering.[1,3,4] Frozen ground phenomena, including nivation, determine the main aspects of frost weathering. The latter is most pronounced in the horizon of discontinuous weathering.

The product of frost weathering on northern plains is cryopelite, i.e., loesslike surface deposits.[3] The predominantly suglinoklike material of these deposits is qualitatively different from the substrate on which it is being developed. Several specific characteristics of these deposits (their silty nature, aggregation, appearance of secondary colloidal minerals and their distribution, micro- and macroporosity, etc.) depend on the specific characteristics of frost action in the layer of discontinuous frost weathering.

The thickness of the mantle is usually equal to, or is greater than, that of the layer of discontinuous weathering, although some investigators feel that the mantle is usually not as thick as the latter layer.[1] We have in mind the thickness of the mantle on the surface of flattened water divides. In depressions the mantle may be considerably thicker due to redeposition by various processes occurring on the slopes.

In the course of formation, the mantle deposits acquire such specific characteristics as thixotropic properties. The thixotropic properties of the mantle suglinoks serve as evidence of their genetic origin.[2] The structural features of the loess-like deposits and their thixotropic properties are responsible for such specific surface phenomena as solifluction (typical examples of which are pseudoterraces, ridges, and lobes), clay boils, and hummocky microrelief (mineral hummocks). Similar but outwardly different features occur on the coarse-grained weathered crust of crystalline rocks with a fine-grained mineral filler: striations on slopes, stone polygons, and stone rings. The thixotropic properties of the mantle deposits facilitate the development of these phenomena. Consequently, the presence of clay boils, hummocky microrelief, solifluction, and corresponding sorted forms on coarse-grained residual material may serve as an indicator of a developed mantle horizon that had passed through a certain stage of frost weathering.

Mottled and hummocky forms of microrelief are of polygenetic origin. Their development (clay boils on sections with a poorly developed topsoil and a thin layer of turf, the mineral hummocks on sections with a relatively thick layer of turf, and a better developed topsoil) is predetermined by the formation of closed systems. The latter are due to fissuring during frost desiccation of soils in the course of development of a convective instability and settlements in unfrozen wet thixotropic soils in the layer of seasonal thawing. If rarefied soil is exuded to the surface, this leads to the formation of clay boils, but, if this does not happen, ice formation in combination with heaving leads to the development of hummocky relief. The uneven increase in the soil volume during freezing results not only from phase transitions of water, but also from a redistribution of the soil mass within the closed system.[4]

The absence of loesslike mantle deposits and genetically related microrelief forms on the relatively young Holocene surfaces indicates that the transformation of the substrate to the final products of cryohypergenesis (cryopelite or cryoclastopelite) requires considerable time. Observations show that these deposits and related microrelief features are developed on all surfaces that had been formed by the end of the upper Pleistocene.

However, in the permafrost region, the loesslike deposits are not relict formations. In the layer of seasonal thawing and freezing, the processes of cryohypergenesis are occurring constantly, although their intensity may vary in relation to changes in the physicogeographical conditions.

Structural microrelief on the loesslike deposits develops in distinct stages. There are active, developing forms, as well as stabilized forms, which are often in the degradation stage (e.g., shrubs on the surfaces of clay boils). The reasons for this are very complex. One of the main causes is the cyclic development of colloidal systems and their aging, as well as the processes of syneresis, i.e., spontaneous compaction of coagulation structures, which determines the thixotropic properties of the mantle deposits.

The cryogenous structure of the horizon of discontinuous frost weathering, as well as physical and physicomechanical processes in it, are largely a function of its temperature regime. The mean annual temperature of this horizon is one of the most important characteristics used in the analysis of its thickness. However, it does not completely reflect the peculiarities of the thermal state of soils throughout the year. To understand the processes that occur in this horizon and determine its main structural features and properties, it is important to consider the temperature regime from season to season. In this respect more and more use is being made of such characteristics of the temperature field as the temperature gradient.

A. I. Popov showed the importance of the maximum fall and winter gradients in the formation of cryogenous textures. In the case of the layer of discontinuous frost weathering, we should also note that the temperature gradient and hence the ice content depend on the thickness of the horizon itself. If the horizon is not thick (about 0.6 m), the maximum gradients hold throughout the depth and have the same value, although the arrival of their maximum values at the base of the horizon is somewhat delayed. Because of this, there are usually two kinds of texture: fine-layered or latticed texture in the upper part and massive texture at the base. If the horizon is very thick (1 m and more), the maximum gradients in its lower part are 1.5-2.0 times smaller than in the upper part. In the case of optimum water contents of soils in the horizon, this leads to formation of three kinds of texture: ice-saturated upper and lower parts of the profile, and the middle part with a low ice content. This type of ice distribution occurs within a wide range of mean annual temperatures at the base of the horizon of discontinuous frost weathering. In the Bol'shezemel'skaya tundra, for example, it occurs at temperatures ranging from -1.5°C to -4°C.

A nongradient or almost nongradient temperature distribution in the transition periods in the spring and fall, when a uniform temperature is established throughout the profile, is very characteristic of the given horizon. Apart from differences in the thickness, the duration and the thermal background of its nongradient (isothermal, according to D. V. Redozubov) state differ also in the spring and fall. The fall is 1.5 to 3 times longer than the spring. A nongradient temperature distribution in the fall is the result of the leveling off of temperatures near 0°C (in early fall just above 0°C, in late fall just below). We may assume that the daily temperature fluctuations between positive and negative values, which are so common in the fall, extend throughout the horizon of discontinuous frost weathering, the thickness of which is at its maximum at this time of the year. Therefore, the physicomechanical and the physicochemical processes related to these temperature fluctuations are developed simultaneously throughout the horizon, and frost weathering is most intensive.

In the spring, a nongradient distribution of temperatures occurs against the background of their negative values. The absolute temperatures decrease simultaneously and very rapidly throughout the horizon. For example, while the temperature in the Bol'shezemel'skaya tundra increases in a month from -10°C or -15°C to -1°C or -7°C, the gradient in the layer of discontinuous frost weathering remains the same and does not exceed 2 deg/m. Such a temperature background prior to thawing should not affect cryolithogenesis. The situation changes in the first half of summer during thawing, which is greatly dependent on the nature of mottled microrelief. Large differences in the depth of thawing below the medallions and the surrounding ground lead to formation of a cellular base of the seasonally thawing layer and hence to considerable differences in the temperatures and water contents of medallions and the surrounding ground. This time of the year is the most favorable for further development of mottled microrelief.

The temperature distribution throughout the profile in the summer is interesting, since soils are heated to a temperature of +4°C, which corresponds to the maximum density of water. Consequently, the summer temperature of soils affects the migration of solutions in the soils and this in turn determines the accumulation of iron and the intensity of thixotropy, which is one of the most important and characteristic properties of the horizon of discontinuous frost weathering.

Therefore, the temperature regime throughout the year strongly influences the processes in the horizon of discontinuous frost weathering, depending on the season. For example, physical weathering predominates in winter, chemical weathering in the summer, and physicochemical weathering in the spring and fall.

Mottled and hummocky microrelief, as well as the solifluctional features, may be of vastly different ages and may even be formed at present. However, they can be developed only on soils that had been properly prepared by frost weathering. The available data indicate that such soils are the mantle deposits on the subarctic plains, which are not younger than the upper Pleistocene and are represented by silty suglinoks and supeses with very distinct thixotropic properties.

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CLAY MINERALS IN PLEISTOCENE PERMAFROST AND CONTEMPORANEOUS BOTTOM DEPOSITS OF CENTRAL YAKUTIA (AN X-RAY STUDY)

M. N. USKOV Institute of Permafrost Studies, Siberian Branch of the USSR Academy of Sciences

Numerous investigators have shown that clay minerals are reliable indicators of physicochemical conditions, thermodynamic parameters, and the geological environment responsible for weathering, sedimentation, diagenesis, and all other stages of soil and rock formation in the zone of weathering. This is due first of all to the fact that clay minerals are the main product of authigenic mineral formation in these processes.* Determination of individual clay minerals, their quantitative distribution throughout a cross section, and their crystallochemical properties is especially important during stratigraphic differentiations of "barren" beds and in paleogeographic regional studies.[1,2,6,7]

The problem of reconstructing the paleogeographic environment can be divided into two adequately independent parts. First of all, the specific composition of the clay fraction (or clay cement) in soils depends on the paleogeographic circumstances, i.e., for each type of lithogenesis there is a specific paragenesis of clay minerals. On the other hand, the crystallochemical properties (the iron content, parameters of the elementary cell, degree of swelling, mixed layering, polytype modification, acidity or the $Fe^{2+}:Fe^{3+}$ ratio in the octahedra of the packet or in the brucite layer, structural ordering and structural properties proper) make it possible to reconstruct the facies environment within the lithogenetic types, i.e., the acidity or the alkalinity of the sedimentary environment, the degree of diagenetic reworking of soil, the oxidation-reduction potential of the environment, and subsequent epigenetic evolution of soil.

However, in spite of the abundance and importance of the data obtained, there is no unambiguous and reliable information on the precise mineralogical composition of finely dispersed soil fractions in the cryolithozone,† and no justifiable subdivisions of lithogenetic and epigenetic transformations of sedimentary material. In this connection, the "ice type of lithogenesis" suggested by N. M. Strakhov[5] (or cryolithogenesis as it is commonly referred to in permafrost studies) is a very general term without sufficient (for the establishment of the type of lithogenesis) mineral-lithological content. Moreover, there is no information on the mineral-crystallochemical mechanism of transformations (or on their absence under conditions of cryolithogenesis) of textures of layered minerals (in this case clay minerals) at the level of atomic crystal lattices in thermodynamic descriptions of this mechanism.

Closely related to this are studies of the state and properties of various types of water in the structure (especially in structural interpacket voids) of clay minerals throughout the entire temperature range, both during freezing and thawing. This problem, although purely theoretical in the intial stages of investigations (in spite of its experimental nature), is of fundamental importance for the explanation of low-temperature physical, mechanical, and other properties of natural soils and fine-grained systems in general.

Bearing in mind the above mentioned unsolved problems, it appears essential to carry out detailed studies of clay minerals from soils in the cryolithozone using contemporary physical methods of structural investigations with the aim of determining general physicochemical and thermodynamic conditions of the cryolithogenesis. We studied the clay fraction (<1μ) from perennially frozen Pleistocene sediments (the core samples* from boreholes in the upper reaches of the Tatta River in Central Yakutia) and from contemporaneous unfrozen bottom deposits in the thermokarst lakes (the cores† from the Khaiagatta Lake and the lakes along the Pokrovsk highway in Central Yakutia).

All soils contained more or less the same minerals: various hydromicas that made up the main mass of crystalline material (>25 percent, generally 40-55 percent), kaolinite (10-20 percent), montmorillonite (10-25 percent and up to 35 percent), chlorites (15-25 percent), and such nonclay minerals as quartz (up to 20 percent) and feldspars (5-10 percent), including both orthoclase and plagioclase. Occasional traces of calcite (up to 3 percent) were present also.

Let us examine the mineralogical characteristics of the given soils and the connection between differences in their mineral composition and differences in their cryogenous history. In contem-

*At the same time, M. A. Rateev[2-4] claims that clay beds, especially in marine facies, are usually of allothigenic origin.

†Permafrost studies occasionally contain detailed tables and descriptions of the mineralogical composition of heavy fractions, which has no relation at all to authigenous minerals in general and mineral formation in the cryolithozone in particular.

*Samples, geological descriptions of soils, and stratigraphic differentiation were provided by M. S. Ivanov.

†Sampled by N. V. Bosikov.

poraneous unfrozen deposits, the soils (or more precisely, the sediments) are still in the syngenetic stage, and this is clearly expressed in their mineralogical composition. All clay minerals in the sediments from the Khaiagatta Lake (in two type profiles through the entire sedimentary layer) and from the lakes along the Pokrovsk highway are very dispersed and poorly crystallized, and are therefore structurally not ordered. Because of structural deficiencies, hydromica and chlorite are strongly degraded and are partially (in blocks) transformed to montmorillonite, forming unordered mixed-layered phases. The traces of montmorillonite packets, which swell up in water and because of intrusions of organic polar molecules, are statistically distributed, and therefore the X-ray photographs reveal the presence of a fractional 001 series (reflections from basal atomic lattices). Evidently, such direction of the crystallochemical process is due to conditions prevailing at the bottom of the lakes, i.e., a low mineralization of water, which leaches out first the cations from the interpacket spaces (i.e., normal degradation of micas to montmorillonite) and then Fe^{3+}, Fe^{2+}, and Mg from the octahedra of micas, hydromicas, and chlorites.

We regard montmorillonite under given conditions as a newly formed mineral, i.e., we think that montmorillonization is taking place in the sediment, which is related to the dispersion of sedimentary material. Since in the contemporaneous bottom deposits of the thermokarst lakes this process can occur only at the bottom during early diagenetic transformations, montmorillonite as such does not yet exist. Instead, there are swelling montmorillonitelike packets incorporated in the mixed-layered mineral phases, i.e., something like metastable intermediate phases. Our study of two pairs of samples from the profiles in the Khaiagatta Lake showed that the number of swelling montmorillonitelike packets increased with depth right up to the appearance of montmorillonite proper.

Hydromicas from the sediments in the lakes along the Pokrovsk highway are close to glauconite, i.e., ferruginous trioctahedral hydromica, while in the sediments of the Khaiagatta Lake they resemble conventional illite, which we regard as a dioctahedral degraded muscovite (more precisely, a mixed-layered unordered intergrowth of various dioctahedral hydromicas).

Judging by the X-ray analysis, the interaction with HCl, and the reaction to heating at 600°C for 1 h, the chlorites are varieties of magnesia with a low iron-content resembling clinochlore-apennine, characteristic of chlorites formed on dark-colored minerals in magmatic and metamorphic rocks.

The same applies to the mineralogical composition of soils from the Pleistocene permafrost beds studied by us. However, between these two soil groups there are also some important differences.

First, the degradation of hydromicas in the earlier, unfrozen state advanced much farther. This is indicated by the absence of intermediate mixed-layered phases. Instead, there is a fairly large (10-40 percent) amount of montmorillonite. This is precisely the reason why hydromica in these sediments is free of swelling packets.

Secondly, the similarities in the distribution of montmorillonite and chlorite in relation to the amount of the pelitic component (see Figure 1a; the psammitic fraction level, 20 percent), as well as in the distribution of montmorillonite and kaolinite in relation to the psammitic component (Figure 1b; the pelitic component level, 0 percent), indicate that the montmorillonization of soil (evidently the main transformation process under conditions of cryolithogenesis) extends not only to hydromica but also chlorite and later even kaolinite.

On the basis of structural transformations of clay minerals, the montmorillonization process should be related to two subsequent stages of the diagenesis:

1. Transformation of hydromica to montmorillonite.
2. Transformation of chlorite and later kaolinite to montmorillonite.

According to the available data, the sediments in contemporaneous bottom deposits have not yet reached even the first stage (only the lowest zones of bottom sediments correspond to the initial phase of the first stage), while in the

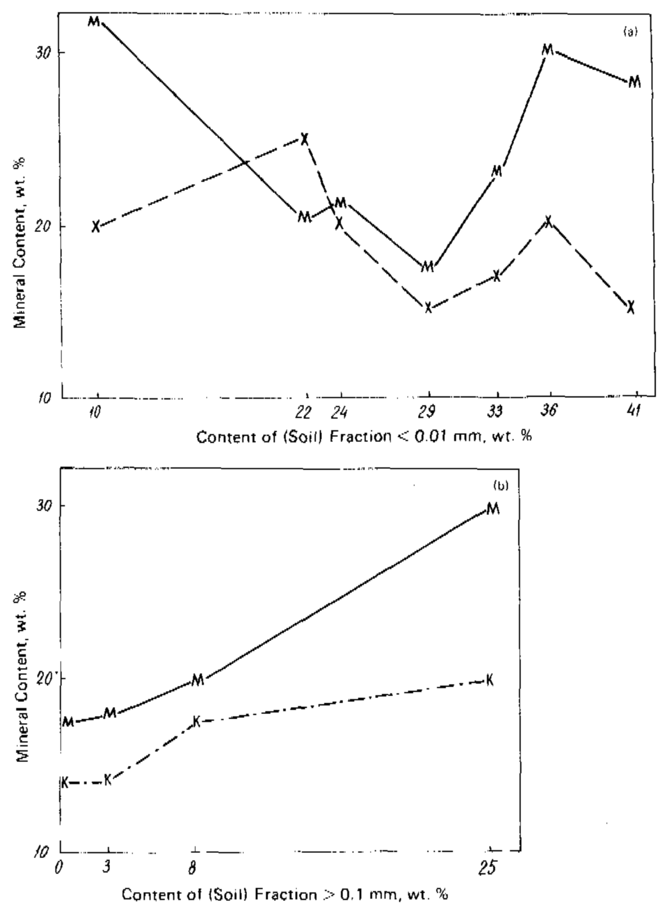

FIGURE 1 Clay mineral content vs. grain size of soils. M = montmorillonite, X = chlorite, K = kaolinite.

Pleistocene permafrost the montmorillonization has reached the second stage. On the basis of structural modifications, the transition of kaolinite to montmorillonite can be related to the late diagenetic-early epigenetic stage of soil transformation.

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MECHANISM OF FORMATION OF LENSLIKE ICE IN STREAMS AND SOILS

B. V. UTKIN All-Union Research Institute of Road Construction

Lensiform accumulations of seasonal ice occur characteristically within a wide range of negative temperatures and water discharges on freezing surfaces and in underground streams. The physical mechanism of ice-lens formation is one of the important problems in permafrost investigations that has been studied for more than 200 yr.

The greatest attention is paid to hydrolaccoliths, thick lenslike formations on crossings over mountain streams, and ice lenses on temporary streams, particularly in the earth beds of roads, formed during abrupt changes from winter thaw to cold weather.

The ability of an ice lens to develop an enormous uplifting force while growing in area and height leads to serious deformations of engineering structures. The complexity of this problem makes it imperative to solve it with the help of classical mechanics.

In nature the initial, "starting" act of the process is a nonelastic collision of two material systems (bodies) with masses m_1 and m_2 and velocities v_1 and v_2 at a certain angle β. The interacting systems cancel their kinetic energies, which are used up on the formation of the total mass of the qualitatively new geometric form.

The kinetic energy cancelled during a nonelastic glancing collision may be expressed by the following equation:

$$\Delta T = \frac{m_1 m_2 (v_1 \cos \alpha_1 - v_2 \cos \alpha_2)^2}{2(m_1 + m_2)}, \qquad (1)$$

where $\Delta T = T_0 - T$ is the cancelled kinetic energy, T_0 is the kinetic energy prior to collision, T is the total kinetic energy of the combined systems, $\beta = \alpha_1 - \alpha_2$ is the collision angle, and α_1 and α_2 are the angles between the direction of the movement of the systems and the line of collision.

Equation (1) describes the changes in the energy state of forming masses, including the flows (fluid and loose) that interact with each other with elements of the surrounding medium.

In the case of a direct, central collision,* where $\alpha_1 = 180°$ and $\alpha_2 = 0$, Equation (1) assumes the form of the well-known Carnot formula:

$$\Delta T = \frac{m_1 m_2 (v_1 - v_2)^2}{2(m_1 + m_2)}. \qquad (2)$$

The factors that cancel the kinetic energy of a fluid mineral flow (water, quicksand) may be any obstacles that have a slowing down or braking effect: solid bodies related to the morphology of the riverbed, additional water discharges, frozen barriers, obstacles related to human activies, etc. Allowances should be made for the fact that formation of an ice lens occurs in a freezing turbulent flow, i.e., a flow that carries suspended ice crystals. Such a flow obeys the hydraulic laws.

The cancelled kinetic energy of the flow is

*A direct central collision is regarded as a special case of a glancing collision.

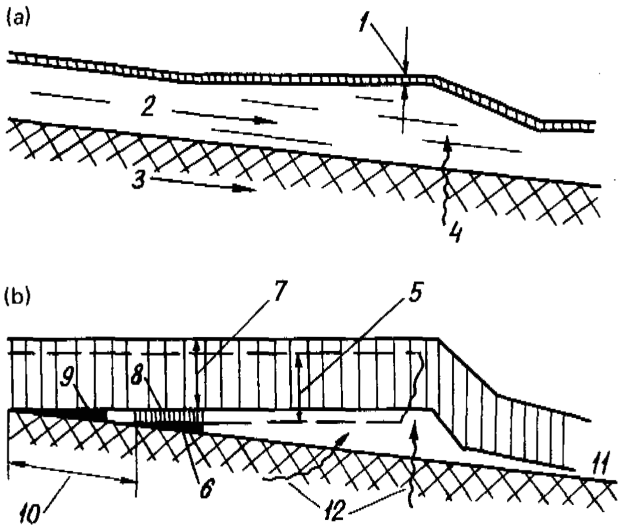

FIGURE 1 Formation of ice wedges in a freezing water flow. (a) First stage: 1--primary ice cover on the free surface of a water mound, 2--channel flow, 3--subchannel flow, 4--source of groundwater (its presence is not essential). (b) Second stage: 5--ice cover formed by freezing water in a wedgelike fissure (6), 7--ice cover after water filling the fissure (6) has formed the ice wedge (8) and after formation of a new fissure (9); 9--wedgelike fissure formed after the ice cover was pulled away on section (10) from the solid base (11), 12--water coming in due to pressure or suction forces. The process is shown schematically. In actual fact the height of the ice wedge (8) was only 0.1 times greater than that of the water wedge (6).

converted to the potential energy, which manifests itself in the formation of the dynamic backwater Δh. Using the Bernoulli equation

$$h_1 + \frac{v_1^2}{} = h_2 + \frac{v_2^2}{2g}$$

the dynamic backwater can be expressed by the following formula:

$$\Delta h = h_2 - h_1 = \frac{v_2^2 - v_1^2}{2g}, \qquad (4)$$

where h_1 and h_2 are the heights of the free surface of water, v_1 and v_2 are the average rates of flow without and with backwater, respectively.

Δh is the height of the water mound on the free surface of which the ice crystals in the water freeze during subsequent freezing of the channel to form a stable (primary) ice cover.

Among numerous types of lenslike ice, the best known is the hydrolaccolith, which is an ideal object for investigation.

In the initial stage of formation of a hydrolaccolith, where the process is characterized by the cancellation of the kinetic energy of the freezing mineral mass flow, the temperature factor is of secondary importance: The ice crystals in the water merely adfreeze to the convex free surface of the water mound.

In the second stage the crystallization forces begin to play an important role.

The forces developed by ice crystals formed on the lower surface of the ice cover where it touches the solid riverbed push it upwards and a [partial] vacuum is formed underneath. Water sucked into this "vacuum" ensures continuous crystallization.

The hydrolaccolith grows because a wedgelike fissure is always present between the solid base and the curved ice cover (at the point where the even ice surface becomes curved) (Figure 1). Water freezes in this fissure to form an ice wedge and due to a large expansive force developed by ice crystals (2,000-3,000 kg/cm^2), which greatly exceeds the adfreezing force of ice (6-7 kg/cm^2), the even ice cover on some sections is pulled away from the base. A new wedgelike fissure is formed, which is again filled with water to form another ice wedge, and so on. The hydrolaccolith grows in area and height.

Water enters the hydrolaccolith owing to hydrodynamic pressure from below the riverbed or from underwater springs. Water sucked in is of less importance, since suction occurs only during very cold weather that intensifies the growth of crystals at the base of the ice coating.

Sucked in melt or capillary water plays an important role in the formation of ice lenses in the soils, and particularly at the base of roadbeds, since the hydrodynamic pressure there is low and is produced only by passing vehicles.

If the channel is very deep and the primary ice cover on the hydrolaccolith does not reach the solid base, the uplifting force of crystals is not realized and the hydrolaccolith cannot grow. An ice dam is an example of such an ice lens (Figure 2).

Other types of ice lenses (pingos, peat mounds, etc.) are developed in a similar way.

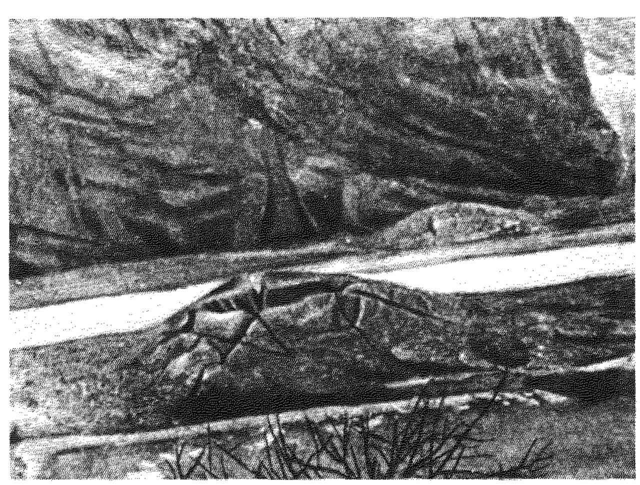

FIGURE 2 Ice domes on the surface of an ice dam on the Naryn River. The domes are filled with slush ice and are coated with a 10-cm-thick ice cover (Photo by A. N. Chizhov).

PART IV

Physics, Physical Chemistry, and Mechanics of Frozen Ground and Ice

N. A. TSYTOVICH,
B. A. SAVEL'EV, and
I. I. VOTYAKOV, *Editors*

MECHANICAL PROPERTIES OF ICE AS A FUNCTION OF THE CONDITIONS OF ITS FORMATION

K. F. VOYTKOVSKIY AND V. N. GOLUBEV   Moscow Lomonosov State University

The mechanism of deformation, strength, and all the mechanical properties of ice are determined to a significant degree by its structure, which, in turn, depends on the thermodynamic conditions of the formation and existence of the ice.

Depending on the degree of supercooling of the water, the magnitude and direction of the temperature gradient, the rate of the supply of the primary phase at the interface, and the presence of impurities, ice of one structure or another is formed. The phase transition of water into ice presupposes the occurrence of crystal nuclei in supercooled liquid and their subsequent growth, which requires preliminary separation of the molecules settling on their surface from the molecules of the surrounding liquid. The activation energy required for this must correspond to the activation energy of self-diffusion in the liquid ΔU characterizing the temperature dependence of the diffusion coefficient $D = Ae^{-\Delta U/kT}$ and the viscosity of this liquid $\eta = C'e^{-\Delta U/kT}$, where k is the Boltzmann constant and T is the temperature °K. Then the crystallization rate I of the supercooled liquid must vary proportionally to the ratio of the number of critical nuclei Ng to the viscosity η:[1]

$$I = C \exp\left(-\frac{\Delta U}{kT}\right) \exp\left(-\frac{4}{3}\frac{\pi\sigma_{iw}}{kT}\cdot r^2\right), \quad (1)$$

where the radius of the critical nucleus $r = 2\sigma_{iw}T_0/\rho_i\lambda\Delta T$, σ_{iw} is the surface tension at the ice-water interface, ρ_i is the density of the ice, λ is the [latent] heat of fusion and $T = T_0 - \Delta T$ where T_0 is the phase transition temperature and ΔT is the [degree of] supercooling of the system. Analysis of this formula shows that under other equal conditions the rate of occurrence of the centers of crystallization must increase with an increase in T, passing through a maximum when $\Delta T = \Delta T = 70$ deg. If we consider that the number of crystals per unit volume N of the ice body formed is proportional to I, then under the condition $\Delta T << T_0$, N will increase, and the volume of the average crystal V_{cr} will decrease as ΔT increases; that is, the greater the supercooling in the system, the finer grain the ice formed will be. The experimental data demonstrate that the dependence of N_{cr} on the magnitude of the supercooling ΔT can be approximated by the following empirical formula:

$$N_{cr} = m(\Delta T)^{1/3}, \quad (2)$$

where m is a proportionality factor that depends on the freezing conditions and in the case of a homogeneous process is equal to $(0.5 + 01.0)10^2$.

In the case of freezing of saltwater, the rate of occurrence of the crystallization centers I must increase in connection with the corresponding applications of T_0, σ, and ΔU. Furthermore, an additional increase in N_{cr} must be observed as a result of retardation of the growth rate of the crystals during formation of concentric zones of ions and molecules of the dissolved salts around them. For values of $\Delta T > 1$°C as a result of this process alone, the number of crystals per unit volume must increase proportionally to the cube of the salinity of the initial water.[2] Studies of the ice formed from water of equal salinity but with different supercooling have demonstrated the following nature of variation of structure: With an increase in salinity and degree of supercooling of the freezing water, the crystals decrease; for the same salinity, the volume of the average crystal turns out to be proportional to $(\Delta T)^{-1/3}$. On the basis of the statically uniform distribution of the nuclei occurring with respect to volume of the supercooled liquid during the homogenous process, more or less uniform-grain ice is formed with isometric anhedral crystals, the mean size of which is determined by the degree of supercooling of the liquid and the salinity of the freezing water.

Under natural conditions, however, it is necessary, more frequently to deal not with homogeneous but with heterogeneous ice formation, which is connected primarily with a significant decrease in the energy of formation of the nuclei of the crystalline phase in the presence of a prepared interface and, in practice, with the constant presence of mechanical admixtures in the water. The thermodynamic calculations[3] show that, even on an ideally smooth infinite surface, a decrease in ΔW of the energy of formation of the ice crystal nucleus by comparison with the homogeneous process is

$$\Delta W = \alpha[Ar^2\sigma_{iw}], \qquad (3)$$

where r is the radius of the equilibrium nucleus, A is a geometric parameter, σ_{iw} is the surface tension at the ice-water interface, and α is a dimensionless factor varying from 0 to 10 with an increase in the degree of cryophilic nature of the material of the base from -1 to +1. The presence on the surface of a solid state of "microroughnesses" leads to a still greater increase in ΔW as a result of which the heterogeneous process of ice formation takes place quite strongly, even with very low supercooling of the water. Since the probability of homogeneous formation of the nuclei of the ice crystals with such supercooling is extraordinarily small, the growth of the ice crystals occurring on the surface is continued in the volume of the liquid in practice without limit. Therefore, in the case of a heterogeneous process in almost all cases, the ice of columnar or long-prism structure is formed. The basic factors controlling the length of the crystals are the thickness of the fluid layer and the conditon of heat and mass exchange at the crystallization front. On the basis of anisotropy of thermal conductivity and growth rate in the various crystallographic directions, the unfavorably oriented crystals taper out and, on going away from the interface, the ice body is made up of larger and larger ice crystals of prismatic shape with orientation of the optical axes perpendicular to the long side of the prism.

Under natural conditions, it is most frequently necessary to deal with the following cases of heterogeneous ice formation:

1. The occurrence of ice on the surface of a cold solid body in contact with a volume of supercooled water.
2. The formation of ice also at a water-air interface.
3. The growth of ice on the surface of a cooled solid body on impingement of drops of supercooled water on it.

In the case of freezing of water at the interface with a solid body, the structure of the contact layers of the ice depends above all on the physicochemical properties of the surface of this body. Usually the contact layers of the ice formation are made up of uniform-grain ice with more or less isometric crystals, the dimensions, shape, and orientation of which are determined by the degree of "cryophobicity" ("hydrophobicity") of the material, its thermal diffusivity, the number of active centers on the surface, and the epitaxy of the ice in the given material. The studies made demonstrated that the dimensions of the crystals increase with an increase in cryophobicity of the material, with a decrease in its thermal diffusivity, and with a decrease in the number of active centers on its surface. With equality of these conditions, the number of crystals per unit volume N_{cr} turns out to be proportional to $(\Delta T)^{1/3}$; however, the proportionality factor m increases by 1 to 2 orders by comparison with the values of m for a homogeneous process. In each individual case m is a unique function of the base material.

In the case where orthotropic growth of the ice occurs, its structure varies in accordance with the laws of geometric selection, as a result of which, at a distance of 3 cm to 5 cm from the contact, the basic structural controlling factors become conditions of heat and mass exchange at the ice-water interface, and the role of the base material in practice reduces to zero. In this stage of ice formation, as a rule, large-grain ice of columnar or long-prism structure is formed with predominant orientation of the optical axes parallel to the surface of the solid body (Figure 1c).

In the case of ice occurring at a water-air interface, the structure of its surface layers depends not only on the degree of supercooling of the water, its salinity, and the magnitude of the temperature gradient, but also the presence of agents accelerating the crystallization process, namely, the number and degree of dispersedness of the foreign admixtures, the falling of drops of supercooled water, or solid precipitation. The formation of the primary ice crystals takes place usually either as a result of the growth of snowflakes, which are the most active introduced nuclei of crystallization, or near the particles of foreign admixtures at the points of more significant local supercooling occurring primarily as a result of falling of drogs of supercooled water. Here, large acicular or dendritic crystals are formed, growing both on the surface and into the layer of water and forming a type of ice cover. The space between these crystals is usually filled by finer-grained ice as a result of which the surface layers of the ice covers are characterized by significant variety in the structural characteristics, both with respect to area and with respect to thickness. However, this variety is observed only in the first centimeters of the ice cover. As it grows, the ice structure varies in accordance with the laws of geometric selection in the presence of constrained growth of the crystals and, at a depth of 5 cm to 6 cm, large-grained ice of columnar structure usually is formed with orientation of the optical axes parallel to the freezing surface. This mechanism of ice formation is characteristic of large bodies of water with large area of the air-water interface.

For small volumes of water, we are dealing either with water drops, which move in the air and, on the basis of significant curvature of the interface, are capable of intense supercooling (to -40°C) or with a volume of water with air-water and solid-water interfaces. In both cases the formation of the ice begins, as a rule, at the points of contact between the supercooled water and the solid surface. In the water drop, the phase transition process begins either around foreign particles after the drop's corresponding supercooling (when atmospheric precipitation such as snow or hail is formed) or after its collision with an obstacle, which, running along its surface, causes icing. The structure of the ice is in this case analogous to the structure of the ice formed when freezing at the contact of a sufficiently large volume of water on a solid surface except that as a result of the high degree of supercooling of the liquid the ice has a finer-grained structure. It is characteristic that in this case the number of crystals per unit volume N_{cr} varies proportionally to $(\Delta T)^{1/3}$.

At the contact between an insignificantly small volume of water and the air or a solid body, as a rule freezing begins with the active centers at the surface of this solid state and extends into the volume of liquid or along its surface. The greatest development is found in the case of crystals when the direction of the highest growth rate conincides with the direction of the temperature gradient.

If the water layer is between two cold surfaces, then crystal nucleation occurs on both surfaces; however, the strengths of the occurrence of the crystal and their growth rate are functions of the physicochemical characteristics of the base material and the intensities of the heat fluxes from each of the surfaces. Depending on the relation between these variables in the central part of the ice interlayer formed or somewhat shifted from the center, there is a contact zone between the two ice bodies growing opposite to each other. With sufficiently great thickness of the interlayer (several centimeters to several tens of centimeters), the contact zone is represented by a weakly undulating surface along which an increased salt concentration and concentration of admixed elements is noted as a result of which no persistent growth of the large prismatic ice crystals from opposite surfaces is observed, as a rule. In the case of a small thickness of the interlayer of water (millimeters to a few centimeters), the stage of orthotropic crystallization cannot develop to its full extent, and the contact zone between the two ice bodies is located with great difficulty by crystallo-optical or chemical research. As a rule, it is represented by a sharply broken surface. No increase of soft or impure elements is observed in it. The structure of the ice formation in this case varies weakly with respect to thickness, and the dimensions, the shape and orientation of the crystals are determined primarily by the physicochemical properties of the materials of the base and the degree of supercooling of the water.

One of the most widespread variants in the formation of ice at the contact with a solid substrate is ice formation during freezing of the moisture in a soil. Here two basic kinds of ice are formed: (1) cement-ice located between the grains of the soil or coating individual particles not connected to each other and in this way stabilizing their mutual arrangement; (2) interlayers and lenses of the ice formed during freezing of the water, filling the voids and cracks in the soil, or arising as a result of the moisture migration to the freezing front.

In the first case, a small amount of water freezes between two or several solid surfaces characterized by significant roughness and the presence of a large number of active centers. Under such conditions, small-grained ice is usually formed with random orientation of the isometric or short-prismatic crystals (Figure 1).

In the second case with a small thickness of the interlayers, the ice is also made up of small (about 1 mm in diameter) crystals of isometric or rod shape with random or weakly ordered orientation of the optical axes of the crystals. With significant thickness of the interlayers and lenses, the structure of the ice body varies from fine-grained with random orientation of the crystals at the contact with the soil grain to medium- or large-grained with columnar or prismatic crystals in the central part of the interlayers and lenses. The predominant orientation in the central part is parallel to the surface of the interlayers with horizontal arrangement of them, and it is accumulated at an angle to these surfaces with a vertical arrangement.

Thus, in all cases of heterogeneous phase transition, the ice-formation process passes through two stages: (1) the formation of a primary layer of ice, the structure and growth rate of which are determined by the physicochemical characteristics of the base material and the heat transfer at the solid-water interface; (2) subsequent growth of the ice formation when the structure of the ice and the process rate are determined by the heat and mass transfer at the ice-water interface and the growth conditions of the ice on the ice base (usually with an ice layer thickness of more than 1 cm to 2 cm). Depending on the engineering problem solved, we must deal either with fine-grained or medium-grained ice made up of crystals of isometric or rod shape with random or weakly ordered orientation (the contact layers of the ice formations and the ice-cement in the frozen ground) or with large- and medium-grained ice made up of columnar or prismatic crystals with predominant orientation of the optical axes perpendicular to the long axis of the prism (the central parts of the interlayers and lenses in the frozen ground removed from contact of the ice coatings).

The above-described three basic versions of ice formation encompass the entire complex of cases of heterogeneous phase transitions; however, the possibility of variation with time of the basic factors determing the process and the existence of numerous additional factors (freezing in the open or closed system, capillary phenomena, moisture migration, and so on), which can lead to significant variation of the structure of the ice formed, are not taken into account.

The structure of the ice formed during the freezing process undergoes certain changes with time as a result of the processes of metamorphism of the crystals. Basic changes take place during deformation of the ice.

The detailed crystal optical studies of deformed ice demonstrated that the basic mechanisms of deformations of the single crystals, or the aggregations made up of several large crystals, are sliding along the base and prismatic planes, whereas during deformation of polycrystalline ice the primary role is played by microshifts of the crystals with respect to each other and the occurrence of microcracks with translational sliding of the groups of crystals and recrystallization processes.

The study of light interference at the boundaries of the deformed crystals demonstrated that the larger the crystals the more intense the stress concentration at the contacts between them and, correspondingly, the more intensely the probabil-

TABLE 1 Compressive Strength of Ice (kbar/cm^2) (compression rate, ε = 0.03/min; temperature, -10°C)

Mean crystal size, mm	Orientation of the crystals		
	Random	Ordered	
		σ''[a]	σ'
Fine-grained ice ($d \leq 1$)	49.5	52	47.5
Medium-grained ice (d = 2 to 5)	—	42.7	36.4
Large-grained ice ($d > 5$)	—	32.2	27.9

[a] σ'' is the direction of compression in the direction of the optical axes; σ' is the direction of compression perpendicular to the optical axes of the crystals.

ity of disturbance of the continuity of the crystals and the crack formation at the contacts increases. This leads to weakening of the ice and a reduction in its strength. Here, it was discovered that in the fine-grained ice the cracks that occur damp out at the adjacent contacts without causing brittle fracture of the ice sample, whereas in coarse-crystalline ice and in single crystals the cracks that occur rapidly progress and the deformed specimen splits into individual blocks. Therefore, the fine-grained ice has the highest strength (see Table 1).

In order to discover the dependence of the strength characteristics of ice on its structure, it is necessary to consider at least three basic factors: the dimensions of the crystals, their isometric nature, and the orientation of the optical axes. At the present time it is possible to consider it proved that the greater the compressive strength of the ice, the smaller the dimensions of the crystals, and that the greater the distortion of the boundaries, the larger the number of crystals, the orientation of the optical axes of which coincides with the direction of the deforming force (Figure 1).

The structure of ice is not a stable characteristic, and, even under constant external conditions, significant changes that encompass the entire complex of structural characteristics take place with time: the dimensions, shape and orientation of the optical axes of the crystals, the nature of distribution of the air inclusions, the brine cells, and the mineral particles vary. With a temperature close to the phase transition point, the rate of increase in the mean diameter of the grains turns out to be inversely proportional to the time of existence of the ice t: $dD/dt = K(1/t^{1-n})$, where K = 1 to 5; n = 0.07 to 0.28. The presence of a temperature gradient and a deforming force accelerate the process of structural rearrangement of the ice.

Experiments with respect to compression of polycrystalline ice with random arrangement of the optical axes under the effect of constant load demonstrated that the variation of its structure during the deformation process is defined by the deformation time and the magnitude of the applied force. In accordance with the three basic stages of creep, it is possible to isolate three recrystallization stages: crystal growth, relative stabilization of the structure, and fracturing of the crystals (Figure 2b, c, d). During the first moments after application of the

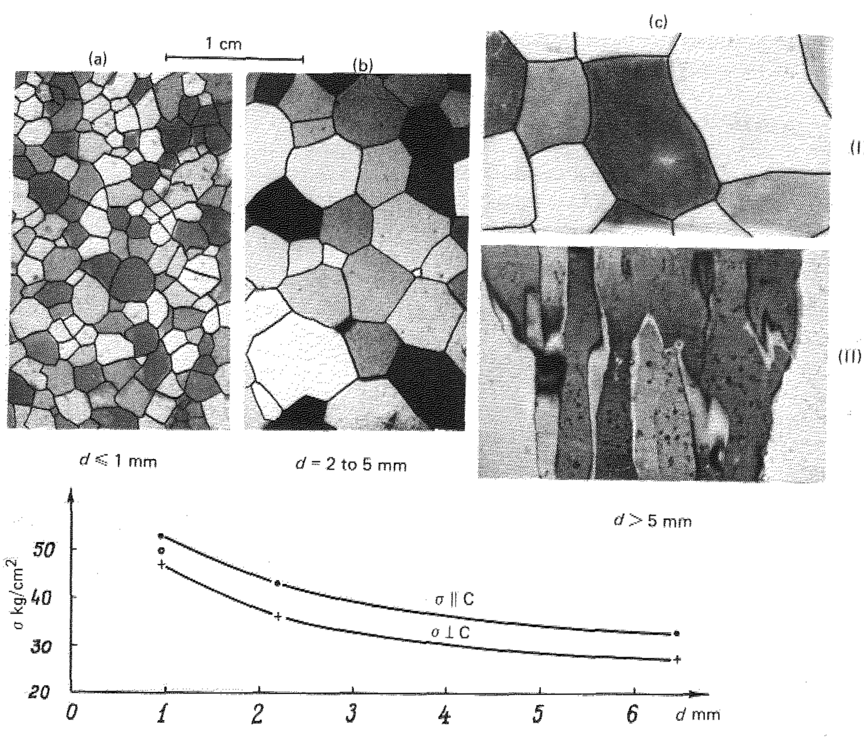

FIGURE 1 Structures of ice subjected to mechanical testing. (a) Fine-grained ice made up of isometric crystals with random orientation of the optical axes. (b) Medium-grained ice made up of columnar and isometric crystals with predominant orientation of the optical axes parallel to the plane of shear. (c) Large-grained columnar ice, optical axes oriented perpendicular to the long axis of the crystals (I--section parallel to the freezing surface; II--section perpendicular to the freezing surface). (d) Compressive strength of the ice as a function of crystal dimensions (d_{mm}) and orientation of the optical axes (C) (temperature, -10°; strain rate, ε = 0.30 min^{-1}).

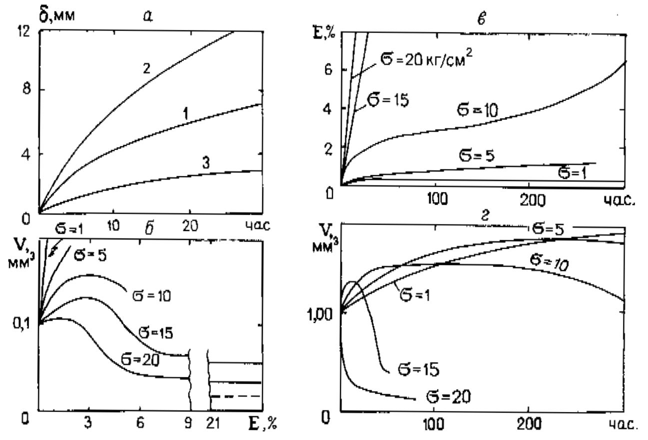

FIGURE 2 Creep curves for polycrystalline ice. (a) In bending (beams 10 × 10 cm in cross section, span 100 cm, load 40 kg). (b) Magnitudes of the deformation at a temperature of -5°C. (c) In compression with the possibility of lateral expansion at a temperature of -5°C and curves of recrystallization of the ice during the creep process. (d) Compression and deformation time. The direction of the crystal axes in the bean is vertical (1) and horizontal (2). The remaining curves are for ice of random structure. On the graph (b), the dotted line shows the size of the crystals in the ruptured specimens.

load, the crystals adapt to the new thermodynamic conditions. Relative rotations and displacements of the crystals and their partial destruction are observed. This leads to simplification of the shape of the crystals. The intensity of this process gradually damps out, and further changes in the structure of the deformed ice are caused by the effort of the system to achieve minimum surface and [minimum] free energy. The less stressed crystals grow at the expense of the more stressed crystals, the small crystals are absorbed by the larger ones, the total number of crystals decreases, and the volume of the average crystal increases significantly (Figure 3a). As deformation takes place, the growth rate of the crystals slows down; and, after some time, relative stabilization of the structure takes place, which is characterized by significant ordering of the orientation of the crystals, combination of a series of small intracrystalline displacements into translational group displacements with the formation of cracks in the crystals (Figure 3b), and growth of certain parts of the fractured crystals at the expense of others. Further deformation leads to progressive growth of the fissures and fracture of the crystals (Figure 3c). This stage is observed only in the presence of shear stresses greater than some limit, about 1.5 kg/cm^2 in the ice at an ice temperature of -1°C and 3 kg/cm^2 at an ice temperature of -4°C [4,5]. With an initial size of the ice crystals of 1 mm^3, the compressive force of 20 km/cm^2 is the limiting load at which the brittle fracture of the ice during some period is still compensated for by the recrystallization processes. A further increase in the deforming force leads to the diminution of the first two stages of the total recrystallization process. Experiments with ice of the same structure demonstrated that the magnitude of the limiting force decreases with an increase in temperature. The same effect is observed also with an increase in volume of the average initial crystal size. The discovered law permits the conclusion to be drawn of a gradual reduction in the strength of the deformed ice as a result of an increase in the crystal size, an increase in the isometry of the crystal, and an increase in the degree of ordering of the crystal orientation during the recrystallization process.

The rate and magnitude of the relative deformation of the ice during the creep process depend both on the size of the crystals and on the directions of the shear forces with respect to the predominant direction of the crystal axes [4,5]. The least creep rates are noted for fine-grained ice with random arrangement of the crystals (Figure 2a, curve 3); the highest creep rate is noted for ice of prismatic structure where the maximum shear stresses operate in the planes coinciding with the directions of the prism axes (curve 2). For the other directions of effect of the shear stresses, the creep rate has an intermediate value between the two indicated cases (curve 1).

The mechanical properties of the ice as a function of the conditions of its formation is of a highly complex nature, since the structure of the ice determining these properties can vary significantly depending on a combination of various factors acting during the ice formation. Therefore, in each special case it is necessary to discover the dependence of the structure on

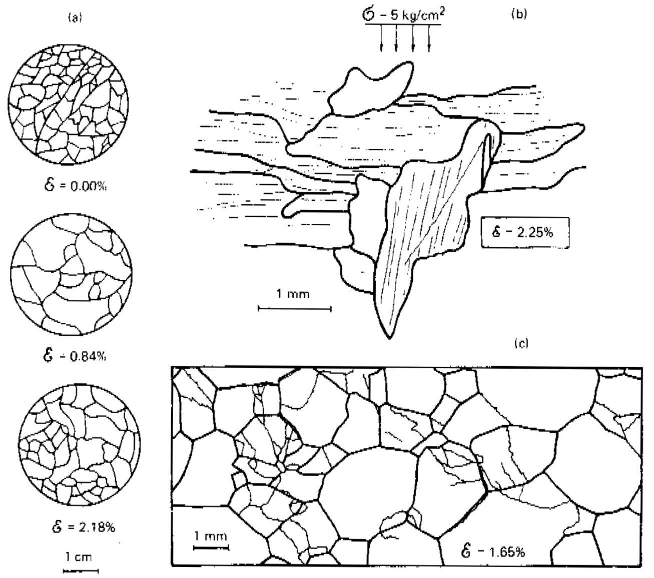

FIGURE 3 Variation in ice crystal size. (a) Formation of intracrystalline fissures. (b) Formation of transcrystalline fissures during the creep process at $\sigma = 5$ kg/cm^2 and a temperature of -5°C.

specific factors, taking into account all such phenomena, and to judge the mechanical properties of the ice with respect to structure. The dependence of the structure on the conditions of formation of the ice has been insufficiently studied at the present time. We consider that when studying the process of heterogeneous ice formation and the mechanical properties of the ice, it is necessary to pay attention primarily to the solution of the following problems:

1. The discovery of the degree of cryophobicity of the various materials.
2. The determination of the conditions of surface arrangement of the ice on the various materials.
3. The study of the process of formation of the primary ice crystals in soil of different fineness and moisture content.
4. The mechanism of deformation and fracture of ice of different structure.
5. The processes of destruction of the continuity and crack formation in the ice during its deformation.

The discovery of the dependence of the physicomechanical properties and structure of the ice on the conditions of its formation permit the forecasting of the strength characteristics of the ice during the solution of a number of engineering and technical problems of ice engineering or purposeful variation of these characteristics creating the corresponding thermodynamic conditions during the ice-formation time.

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LONG-TERM RUPTURE OF FROZEN SOIL AS A THERMALLY ACTIVATED PROCESS

S. S. VYALOV Scientific Research Institute of Foundations and Underground Structures

It is known that under relatively large loads in frozen soils undamped creep develops, which includes three stages: damped deformation, steady state flow, and progressive flow. This last stage ends with rupture occurring over a longer time, depending upon the smallness of the load. The law of reduction in strength is characterized by the relation between the magnitude of the load and the time before rupture. The strength calculations of the soil are reduced to determining the load, which does not lead to rupture during the time of its effect (for example, the service life of the structure).

The existing analytical laws of long-term strength begin with phenomenological arguments. At the same time it appears possible to derive these laws on the basis of the discovery of the physical principles of the processes of deformation and delayed rupture and the examination of these processes as thermally activated processes.

The study of the kinetics of structural deformations of unfrozen[1] and frozen[2] soils* and also ice[3-5] permits isolation of three basic types of these deformations: mutual displacement of the mineral particles, flow of the ice and movement of the adhesive water, and formation and development of microcracks and other structural "defects."

The mutual displacement of the particles begins at the weakest points and is accompanied by re-

*These studies consisted in performing creep tests on soils with simultaneous investigation of the variations in microstructure originating during the deformation and long-term rupture of the soil.

composition of the particles and their reorientation in the direction of shear.

The process of flow of the pore ice-cement still has not been investigated. As for the larger ice inclusions, they have both intracrystalline slip of the crystal "platelets" and bending of the platelets accompanied by decomposition of the crystals and their reorientation--for the crystals in which the basal planes do not coincide with the direction of shear. Here, the work of viscous deformation is converted into the energy that is dissipated in the body and leads to the heating of it. This causes melting of the ice-cement at the contacts, an addition to the unfrozen bound water, and its displacement to the less stressed points. The indicated phenomenon was established experimentally.[6] In the case of saturated clay-soil with a temperature close to zero deg, the displacement of the particles is primarily caused by flow of the water film encompassing the colloidal and clay particles. For the soil at a lower temperature, in which the displacement of the particles takes place as a result of rupture of the bonds, melting of the ice facilitates these disturbances.

The most important factor causing fracture of soil during the creep process is the occurrence and development of microcracks and other structural defects in the soil. The microcracks obviously also appear directly in the large ice inclusions analogously to the way in which this happens in ice crystals.[7] During slow flow, along with the occurrence of new microcracks, "healing" of those that were formed earlier takes place. With a high velocity gradient, crack development prevails, which leads to fracture.

The above processes can be considered as simultaneous occurrence of the phenomena of stress-relief and strengthening. The stress-relief is caused by rupture of the interparticle bonds, melting of the ice, and development of the microcracks. The strengthening is caused by tighter packing of the particles during displacement, the occurrence of new bonds, the freezing of the excess bound water at the points with lower stress concentration in accordance with the principle of dynamic equilibrium of the unfrozen water and ice,[8] and also "healing" of the microcracks.

By analogy with the data from structural investigations of the unfrozen clay soil,[1] it is possible to propose that rupture takes place when the degree of damage to the cross-sectional area of the soil by microcracks and other "defects" $\omega = S_d/S$ (where S_d is the area occupied by the "defects," and S is the total cross-sectional area) reaches a critical value that is a constant for the given soil $\omega = \omega_r$ = const.

According to the Frankel-Eyring molecular-kinetic theory (rate theory) which has found application for ice,[5,9] unfrozen,[10] and frozen,[1,11] soils, the elementary particles (molecules) of a viscous medium are held in a state of equilibrium by an energy barrier that separates the adjacent fields of force. In order to overcome this barrier, the particle must acquire "free activation energy," U, equal to or greater than the potential energy of the bond U_0. In the general case the "activation energy" depends on the magnitude of the applied stress τ, that is, $U - U(\tau)$. The number of "jumps" of the particles over the barrier per unit time is

$$j = j_0 \exp\left[-\frac{1}{kT}U(\tau)\right], \qquad (1)$$

where $j_0 = XkT/h$; X is a particle concentration function, k is the Boltzmann constant, h is the Planck constant, and T is the absolute temperature (°K).

In molecular-kinetic theory, by elementary particles we mean molecules of a viscous medium, and the flow is considered as their continuous displacement. However, the deformation of the soil is caused not only by movement of the water film, but also by displacement of the solid particles accompanied by significant structural changes. Accordingly, the interparticle bonds change, and, consequently, the energy, analogous to the activation energy required for breaking of them. Thus, the form of the dependence of the energy of rupture of the interparticle bonds on the stress U (τ) must be discovered from an investigation of the kinetics of the structural variations of the soil.

In S. S. Vyalov *et al.*[1] the derivation of the generalized rheologic equation of clay soil is presented. Here, we shall present a somewhat different derivation of the equation of long-term strength as applied to frozen soils.

We shall consider Equation (1) as the expression describing the displacement rate of the solid particles of the frozen soil, and the "activation energy" $U(\tau)$ as the energy that must be communicated to these particles to destroy the bond between them and displace them from one position to another. Because with an increase in stress this "activation energy" decreases, it is possible to set $U(\tau) = \alpha\, U_0/\tau$ where U_0 is the initial potential energy of the bond (before the beginning of deformation), and α as a coefficient characterizing the variation of the structure.

It is obvious that the time up to rupture is inversely proportional to the specific rate of displacement of the particles $t_r = a/j$, that is, considering Equation (1) we can write*

$$t = Be^{\beta/\tau} \text{ or } \tau = \frac{\beta}{\ln t/B}, \qquad (2)$$

where $\beta = \alpha U_0/kT$, and $B = \delta/j_0 = \delta h/kTX$.

Inasmuch as the specific displacement rate of the particles j is directly proportional to the flow rate, that is, $d\gamma/dt = A_j$, Equation (2) is identical to the condition established from the experiments that the time to rupture is inversely proportional to the steady-state flow rate $t = \text{const}/d\gamma/dt$.

For more correct observation of the initial conditions ($t \to \infty$; $\tau \to \tau_0$), Equation (2) must be written in the form

$$\tau = \frac{\beta}{\ln \dfrac{t + t^*}{B}} \quad \text{or} \quad \ln(t + t^*) = \nu\,\frac{\tau_0 - \tau}{\tau} \qquad (3)$$

*Here and hereafter the subscript r on t will be omitted.

where t^* is the arbitrary very small value of the time (it is possible to set $t^* = 1$); $\tau_0 = \beta/\ln t^*B$ is the provisional instantaneous strength (depending on U_0 and T) and $\nu = \ln 1/B$.

Beginning with the above mechanism of deformation and rupture of frozen soils, let us discuss the physical meaning of the parameters entering into Equation (3). From the experiment it follows[1] that the variation in degree of damage per unit time with respect to the undamaged cross-sectional area is $\Delta\omega/(1 - \omega = \kappa - \tau\ \Delta t/(t + t^*)$, where κ is a parameter and τ is some stress function. Considering that ω varies from ω_0 to ω_r for $0 \leq t \leq t_r$ and integrating the equation presented within these limits, we obtain

$$\kappa\tau\ \ln(t_r + t^*) = \ln \frac{1 - \omega_0}{1 - \omega_r}. \qquad (4)$$

From comparing Equations (4) and (3), it follows that

$$\tau = \frac{\tau}{\tau_0 - \tau}, \quad \nu = \ln \frac{t^*}{B} = \frac{1}{\kappa} \ln \frac{1 - \omega_0}{1 - \omega_r}.$$

Thus, the parameter $B = [(1 - \omega_r)/(1 - \omega_0)]^{1/\kappa}$ has a strict physical meaning characterizing the ratio of the undamaged cross-sectional area of the soil at the time of rupture $(F_0 - F_r)$ to the area in the initial state F_0 (for $t^* = 1$); this ratio is a constant for the given soil. In exactly the same way

$$\beta = \tau_0 \ln \frac{t^*}{B} = \frac{\tau_0}{\kappa} \ln \frac{1 - \omega_0}{1 - \omega_r} \text{ (for } t^* = 1).$$

The indicated parameters can be defined directly by experimental data.

Formula (3) proposed in its time[6] as phenomenological was checked by the author and a number of other researchers on a large amount of experimental data. In Figure 1 we have one of these comparison; the parameters of Equation (10) for this case are presented in Table 1.

Formula (3) was obtained under the assumption that the disturbance of the relations between the particles takes place for any stress if it acts for an unlimited time and, correspondingly, $\tau \to 0$ for $\tau \to \infty$.

If we assume that the breaking of the bonds takes place only when the applied force exceeds some value, then the equation for the long-term strength is written in the following form proposed by Yu. K. Zaretskiy:[1]

$$t = \nu \frac{\tau_0 - \tau}{\tau - \tau_\infty}, \qquad (5)$$

where τ_∞ is the long-term strength up to which the deformations are damped and rupture does not occur.

It must be noted that both the separation of the creep into damped and undamped and the concept of the long-term strength are arbitrary. Actually, on the one hand we know that the soil (the more so ice) flows, even for small stresses, if we consider these processes on geological time

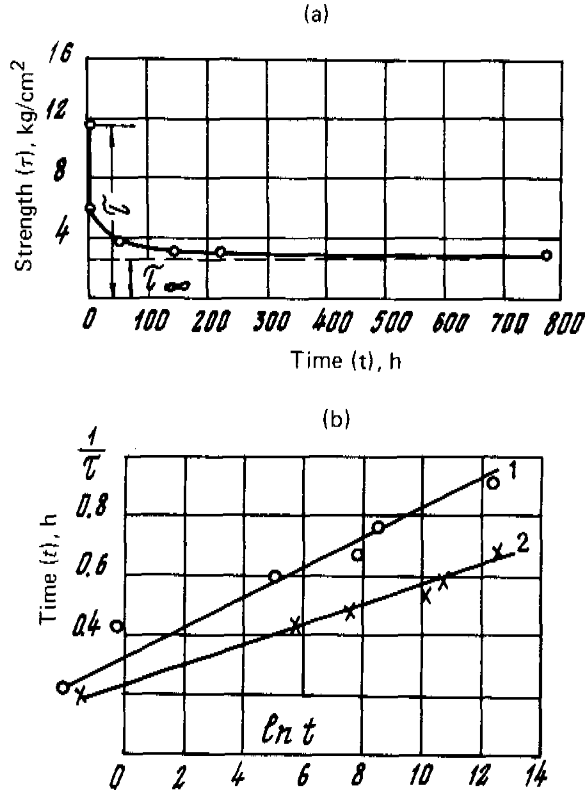

FIGURE 1 Curves of long-term strength (a) and their rectification in the coordinates $(1/\tau)$ - $\ln t$ (b). Shear testing of frozen soil along the lateral surface of foundations (adfreeze strength) for $\Theta = -0.4°$. 1--silty clay; 2--silty supes.

scales. On the other hand, we see the fact of practical stabilization of the deformations everywhere. It is possible to consider that after a sufficiently long time the reduction in strength becomes so insignificant that this reduction can be neglected in engineering calculations. Correspondingly, when using Formula (3) it is necessary to set

$$\tau_\infty = \frac{\beta}{\ln \frac{t_{lim}}{B}}, \qquad (6)$$

where t_{lim} is set either equal to the service life of the structure or it is determined from the condition that the difference in the values of τ for $t = t_{lim}$ and for $t = 100$ years does not exceed 3 percent, that is:

$$t_{lim} = (100B^{0.03})^{1/1.03} \qquad (6a)$$

where B is in years.

For the example presented in Figure 1 (experiment 2), Formulas (6) and (6a) give: $t_{lim} \simeq 48$ yr and $\tau_\infty = 1.28$ kg/cm^2. If we set $t_{lim} = 100$ yr, then we obtain $\tau_\infty = 1.26$ kg/cm^2, that is, for $t_{lim} = 50$ and 100 yr the values of τ_∞ are in

TABLE 1 Experimental values of the parameters of Equation (10)

Soil	Type of Tests	B, h^{-1}	a, $kg \cdot cm^{-2}$	b, $kg \cdot cm^{-2} \cdot deg.^{-n}$	n
Sand	Compression	$1.01 \cdot 10^{-2}$	14.9	58.3	0.5
Suglinok	Compression	$2.3 \cdot 10^{-1}$	11.5	22.5	0.7
Supes	Adfreeze	$6.5 \cdot 10^{-6}$	15.02	46.5	0.5

practice equal and coincide with the experimental value of $\tau_\infty = 1.35$ to $1.25\ kg/cm^2$. If we apply Formula (5) to forecast the reduction in strength for a geological time interval, for example, after 1,000 and 10,000 yr, then we obtained $\tau_{1,000} = 1.14\ kg/cm^2$ and $\tau_{10,000} = 1.06\ kg/cm^2$.

The parameters β and B of Equation (3) according to Formula (2) are inversely proportional to the absolute temperature T. In turn, the activation energy also depends on the temperature and can be expressed by the equation known from thermodynamics $U = H - TS$, where H is the heat of activation (enthalpy), and S is the entropy of the system. Then the dependence of the strength on temperature is defined by the expression

$$\tau = \frac{\alpha(H_0 - TS_0)}{kT \ln\left[kT\frac{x}{\delta h}(t + t^*)\right]} \approx \frac{\alpha(H_0 - TS_0)}{kT \ln \frac{t + t^*}{B}} = \frac{\beta(T)}{\ln \frac{t + t^*}{B}}. \tag{7}$$

The approximate equality follows from the fact that the effect of the factor in $\ln \kappa T$ is appreciably less than the factor κT, and, therefore, for simplification it is possible to set $\delta h/\kappa Tx = B \simeq$ const.

The dependence of τ on T undergoes a discontinuous variation when $T = T_0$, where T_0 is the ground thawing temperature. At this point both the potential energy of the bond U and the values of the parameters B and β (Figure 2) vary. However, in practice the indicated variations occur not instantaneously but in some time interval

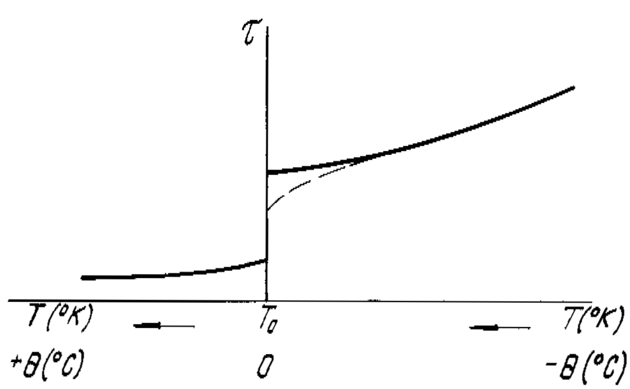

FIGURE 2 Soil strength and temperature.

where in this interval the influx of heat is felt significantly during dissipation of the mechanical energy. Correspondingly, the actual dependence of the strength on the temperature differs somewhat from the theoretical, as is shown by the dotted line in Figure 2, and it is well described by the empirical equations of the type:

$$\beta = a + b\Theta^n \text{ or } \beta = a(1 + \Theta)^m, \tag{8}$$

where Θ is the negative temperature in °C without the minus sign.

The first of the presented formulas is preferable for the higher (from 0° to -10°) and the second, for the lower, temperature. Taking the first of these formulas, we obtain the following form for Equation (3):

$$\tau = \frac{a + b\Theta^n}{\ln \frac{t + t^*}{B}}. \tag{9}$$

In order to consider the variability with time of the load $\tau = \tau(t)$ and the temperature $\Theta = \Theta(t)$, it is possible to apply the so-called principle of linear summation of the damageability:

$$\int_0^{t_r} \frac{dt}{t(\tau,\Theta)} = 1.$$

Substituting the value of t from Formula (2) in this equality and then expressing the parameter β in terms of the temperature dependence (8), we obtain

$$1 = \frac{1}{B}\int_{-t^*}^{t_r} \exp\left[-\frac{\beta(\Theta)}{\tau(t)}\right] dt = \frac{1}{\beta}\int_{-t^*}^{t_r} \exp \left\{-\frac{a + b[\Theta(t)]^n}{\tau(t)}\right\} dt. \tag{10}$$

Variable loads were considered in Equation (3). Here let us consider the problem of taking temperature variation into account.

A. V. Nadezhdin and V. A. Sorokin, under the direction of the author, performed experiments to determine the long-term strength in the case of uniaxial compression of sand ($w = 18.9$ percent) and suglinok ($w = 16.1$ percent) at different temperatures. The parameters of Equation (10) were calculated on the basis of the experimental data; they are presented in Table 1. In

TABLE 2 Time before Rupture t_r and Long-Term Strength τ_∞ for Different Temperatures Θ

Θ (°C)	Sand, Compression at σ = 14 kg/cm^2			Suglinok, Compression at σ = 14 kg/cm^2			Supes, Adfreeze Strength (see Figure 2)		
	t_r, h (Experiments)	$t_{r,2}/t_{r,1}$		t_r, h (Experiments)	$t_{r,2}/t_{r,1}$		τ_∞, kg/cm^2 (Experiments)	$\tau_{\infty(2)}/\tau_{\infty(1)}$	
		Experimental	Theoretical		Experimental	Theoretical		Experimental	Theoretical
-1	1.9	1.0	1.0	2.6	1.0	1.0	1.0	1.0	1.0
-2	10.9	5.75	5.5	7.1	2.73	2.83	1.5	1.5	1.3
-4	121.6	64.6	62.2	36.3	13.9	14.2	2.5	2.5	2.4

this table we have the experimental data from Figure 1 (experiment 2).

Let us consider several cases of temperature regime.

Case 1

From the experiments the time before rupture $t_{r,1}$ was determined for the given value of the load τ = const and for a soil temperature Θ_1 = const. It is necessary to determine the time before rupture $t_{r,2}$ for the same value of τ = const but for a different temperature Θ_2 = const. On the contrary, if the strength τ_1 is determined for Θ_1, then it is necessary to determine the strength t_r for Θ_2.

In accordance with Equation (10) and for simplicity setting $t^* = 1$, we have

$$\frac{t_{r,2}}{t_{r,1}} = \exp\left[\frac{b}{\tau}\left(\Theta_2{}^n - \Theta_1{}^n\right)\right] \quad \text{and} \quad \frac{\tau 2}{\tau 1} = \frac{a + b\,\Theta_2{}^n}{a + b\,\Theta_1{}^n}\,.$$

A comparison of the experimental and the calculated data is presented in Table 2.

From the given data it follows that with variation of temperature the time before rupture varies to a significantly greater degree than the strength.

Case 2

The soil to which the load was applied, τ = const, had for some time a negative temperature Θ_1. Then the temperature varied discontinuously to a value of Θ_2 = const. It is necessary to determine the time to rupture t_r.

Integrating Equation (10) within the limits from $t^* = 0$ to t_1 and from t_1 to t_r, we find that $t_r = t_1 + (t_m - t_1)\exp[(1/\tau)b(\Theta_2{}^n - \Theta_1{}^n)]$ where $t_m = B\exp[(1/\tau)(a + b\Theta_1{}^n)]$.

From the given expressions it is possible to determine the value of τ_∞ if we set $t_r = t_{lim}$ and solve the equation with respect to $\tau = \tau_\infty$.

Let us propose, for example, that we are required to determine the long-term strength of adfreeze of the supes in the case where the temperature $\Theta_1 = -1°$ was maintained for t_1 = 1 yr and then reduced to $\Theta_2 = -2°$. Substituting the values of the parameters from Table 1 in the above formula and setting t_{lim} = 50 yr, we obtain τ_∞ = 1.51 kg/cm^2. If the temperature $\Theta_1 = -1$ deg was kept constant, then we would have τ = 1.28 kg/cm^2, and, if from the very beginning the temperature were $\Theta_2 = -2°$, then τ_∞ = -2.03 kg/cm^2.

In order to check the correctness of the solution obtained, experiments for the compression of the frozen sand were performed under a load of σ = 14 kg/cm^2; during the time Θ_1 = 2.5 h, a temperature of $\Theta_1 = -4°$ was maintained in

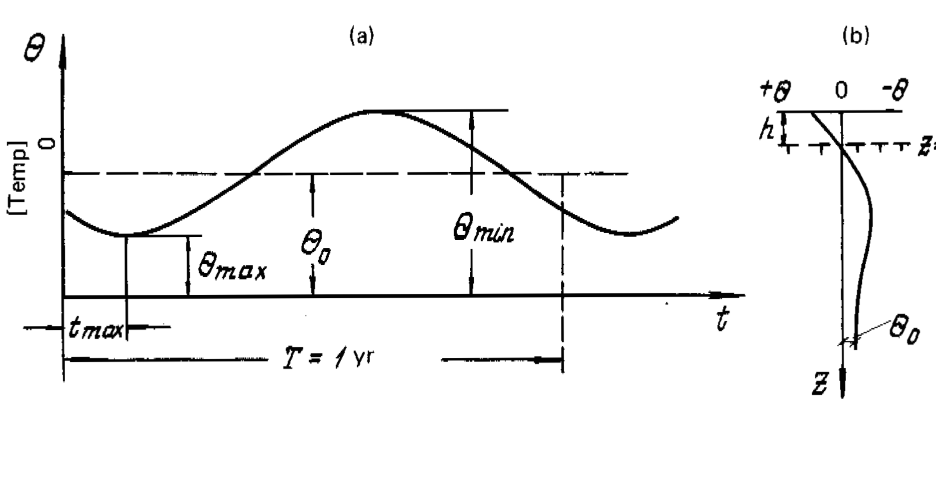

FIGURE 3 Curve of the annual temperature variation of permafrost: (a) at a given depth z; (b) with respect to depth.

the soil, which was then raised to $\Theta_2 = -1°$. The theoretical time to destruction in this case should be (for the parameters presented in Table 1) t_r = 5.4 h; in practice, it was 5.9 h.

Case 3--Periodic Temperature Variation

The most interesting case is that of periodic variation of the temperature of the frozen soil, inasmuch as this case corresponds to the actual temperature regime of frozen ground. It is known that the ground temperature variation at a depth z is determined by the expression (Figure 3a):

$$\Theta = \Theta_0 \left[1 - e^{-D} \cos \frac{2\pi t}{t_1} - D\right], \qquad (11)$$

where Θ_0 is the mean annual temperature of the permafrost (°C), t_1 = 1 yr (7,860 h) is the temperature variation period, $D = z\sqrt{\pi/a_t t_1}$ is the damping decrement, and a_t is the coefficient of thermal diffusivity of the frozen soil (m^2/h).

In Formula (11) it is assumed that when $z = 0$ and $t = 0$ the temperature $\Theta = 0$, that is, the time $t = 0$ is reckoned from the time when the seasonal thawing of the ground reaches its greatest value (Figure 3c).

The solution of the problem is obtained by substituting Expression (11) in Formula (10),

$$1 = \frac{1}{B} \int_{-t^*}^{t_r} \exp \left\{-\frac{1}{\tau(t)} \left[a + b\Theta_c^{\ n}(1-e^{-D}\cos\left|\frac{2\pi t}{t_1} - D\right|\right]^n\right\} dt. \qquad (12)$$

It is possible to define the time before rupture t_r under load τ as constant or variable with time or setting $t_r = t_{lim}$ to find the long-term strength $\tau = \tau_\infty$. These data will pertain to the state of the ground at the depth z, for example, under a footing. If it is necessary to determine the strength properties of the ground with respect to the entire investigated depth from $z = 0$ to $z = l$, for example, when estimating the adfreeze strength with respect to the length of the pile, then it is necessary to sum the diagram of τ_∞; analytically, this summation is written in the form

$$1 = \frac{1}{Bl} \int_{-t^*}^{l} e^N dt dz, \qquad (13)$$

where N is the expression in the braces in Formula (11).

For the soil the parameters of which are presented in Table 1, the solution of Equation (12) was obtained using a [digital] computer*.

In Figure 4 we have curves of long-term strength with respect to depth of the frozen ground constructed from the data of these calculations. In addition to the curves calculated by

*The calculations were performed by L. N. Krustalev, L. A. Sukhodol'skaya, and M. S. Gayday.

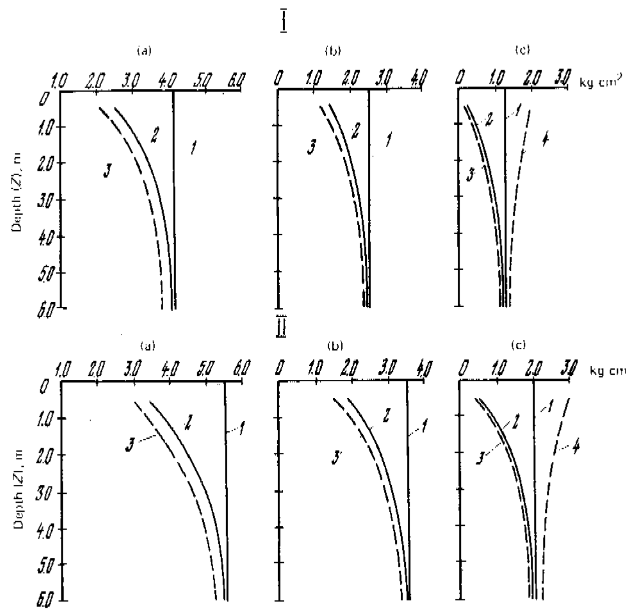

FIGURE 4 Distribution of the long-term strength with respect to depth. (a) compressive strength, sand; (b) the same, suglinok; (c) adfreeze strength, supes. The curves were constructed for temperature values as follows: 1--mean annual Θ_0; 2--variable $\Theta(t)$; 3--maximum Θ_{max}; 4--minimum Θ_{min} with an initial mean annual temperature equal to (I) $\Theta_0 = 1°$; (II) $\Theta_0 = -2°$.

Formula (2) for the variable temperature $\Theta(t)$, the graphs also show the strength diagrams constructed for constant values of the ground temperature corresponding to Θ_0, Θ_{max}, and Θ_{min} (the last case is investigated only for the adfreeze strength); the values of Θ_{max} and Θ_{min} are determined from Expression (11).

In Figure 5 we have the soil strength at different depths as a function of mean annual temperature Θ_0.

As is obvious from Figures 4 and 5, consideration of the annual behavior of the ground temperature significantly changes the values of the long-term strength by comparison with the strength determined for the mean annual value,

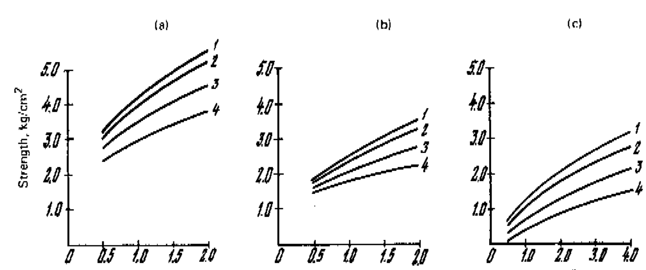

FIGURE 5 Long-term strength as a function of the mean annual (negative) ground temperature. (a) compressive strength, sand; (b) the same, suglinok; (c) adfreeze strength, supes. The strengths are given for various depths: 1--Z = 10 m; 2--Z = 4 m; 3--Z = 2m; 4--Z = 1 m.

and the more so for the minimum temperature. These divergences are especially great within the limits of the ordinary depths of foundations (1 m to 2 m below the depth of seasonal thawing), but with depth decreasing and when l = 6-8 m the temperature fluctuations are almost not reflected in the long-term strength.

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BASIC LAWS OF THERMORHEOLOGY AND TEMPERATURE CRACKING OF FROZEN GROUND

S. YE. GRECHISHCHEV All-Union Scientific Research Institute of Hydrogeology and Engineering Geology

Recently attention to the studies of temperature deformations and temperature fissuring in frozen ground has increased. This is connected with the establishment of a number of factors indicating the effect of frost fissures in the ground on structures and also the occurrence of the necessity for building new, nonstandard structures in the north, the maintenance of which is connected for technological reasons with the occurrence of additional temperature stresses in the frozen ground.

Cases are known of inadmissible deformations of buildings and structures as a result of frost cracking of the ground.[1-3] Frost cracking has a great effect on underground cables.[4] When laying buried pipelines, it is necessary to consider possible additional tensile stresses by the gripping of the pipe in the frozen ground and the formation of frost cracks perpendicular to the axis of the pipeline.[5,6] The necessity for predicting frost cracking of the frozen ground is obvious when planning and designing underground reservoirs for low-temperature liquefied gases[7] and also when constructing underground cold-storage facilities.[8] The phenomenon of thermal cracking of frozen ground is acquiring special significance in connection with the developing trend toward the construction of roads, airports, embankments, and dams from local materials in the North.

The scientific basis for forecasting frost cracking of frozen ground is studies in thermorheology to investigate the physical interrelation among temperature, deformation of the ground, stress, and time. At this time a number of basic laws of temperature deformation of frozen ground have been established.

Experimental studies[9-12] have demonstrated that frozen soils have anomalously large coefficients of thermal expansion: to $2{,}000 \cdot 10^{-6}$ per deg and more for clay, 100 to $400 \cdot 10^{-6}$ per deg for suglinoks and supesses, and about 20 to $50 \cdot 10^{-6}$ per deg for sand. For comparison let us note that ordinary building materials (steel, concrete, wood, and so on) and also rocks and minerals have coefficients of thermal expansion of 5 to $20 \cdot 10^{-6}$ per deg. The magnitude of the coefficient of thermal expansion of frozen soils strongly depends on the temperature, decreasing as the temperature decreases. The thermal deformation as a function of temperature is sharply nonlinear, which distinguishes frozen soil from other materials. In Figure 1 we have the basic experimental types of the relationships between thermal deformation and temperature.

The moisture content of the frozen soil has a great effect on the magnitude of the thermal deformation. For ordinarily encountered values of the moisture content of clay, the coefficient of thermal expansion increases with a decrease in moisture content. With an increase in moisture content, the coefficient of thermal expansion in sand increases, and it decreases in suglinok.[13]

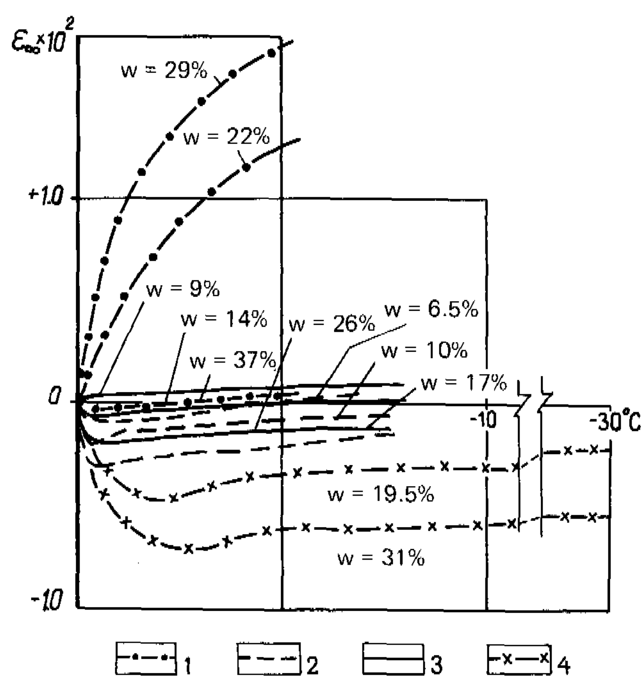

FIGURE 1 Magnitude of the stabilized-temperature deformations (ε_∞) as a function of temperature (initial temperature, 0°). 1--kaolin clay; 2--medium-grained sand; 3--mixture (kaolin clay 70 percent, medium-grained sand 30 percent); 4--supes (according to the data of I. N. Votyakov).

The effect of the granulometric composition of the soil on the magnitude of its coefficient of thermal expansion is quite large. With a decrease in the content of the clay fraction and, correspondingly, with an increase in the sand, the coefficient of thermal expansion decreases exponentially.

The phenomenon called the "temporary thermal aftereffect" has great significance for estimating the thermorheologic properties.[14] This aftereffect is that after a change in ground temperature, and its stabilization, ground deformations continue to develop for a prolonged period of time in spite of the fact that the temperature remains constant.

Processing of the experimental data demonstrated that the basic laws of thermal expansion of frozen soils in the temperature range from 0° to -30° can be expressed by the following type of empirical nonlinear hereditary function:[11]

$$\delta = \int_0^t K(t - \tau)\Phi[\Theta_K(\tau),\ \Theta_H]d\tau, \qquad (1)$$

where

$$\Phi(\Theta_K, \Theta_H) = B_\infty \left[\left(e^{-\frac{\Theta_K}{\Theta_0}} - \lambda e^{-\frac{\Theta_K}{\Theta_1}}\right) - \left(e^{-\frac{\Theta_H}{\Theta_1}} - \lambda e^{-\frac{\Theta_H}{\Theta_1}}\right)\right] \qquad (2)$$

$K(t - \tau)$ is the hereditary function of the temperature effects, which can be written in the form

$$K(t - \tau) = \frac{1}{t_0} \exp - \frac{t - \tau}{t_0}; \qquad (3)$$

δ is the temperature deformation; Θ_H, Θ_K are the initial and final temperatures (the absolute value of the negative temperature, °C); B_∞, λ, Θ_0, and Θ_1 are the empirical constants of the frozen soil; t_0 is the time of the thermal aftereffect; and t is time.

The parameters B_∞ and λ in Formula (2) are an analog of the coefficient of thermal expansion. They depend to a very significant degree on the moisture content. B_∞ decreases and λ increases with an increase in moisture content by the following law:

$$B_\infty = \frac{B'_\infty}{1 + W/W_1}, \qquad \lambda = \frac{\lambda' W}{1 + W/W_1}, \qquad (4)$$

where B'_∞, λ', and W_1 are empirical constants, which do not depend on the moisture.

The time of the temperature after effect t_0 in Expression (3) decreases almost exponentially with an increase in the soil moisture content,[15] and it varies within the limits from 10 h to 100 h for different soils.

The use of Expression (1) for practical calculations is quite difficult in view of its nonlinearity. This expression can be reduced to a linear function if the calculated temperature range causing the thermal stress is known in advance. Then for the given temperature range, the magnitude of the coefficient of thermal expansion can be averaged, and, instead of the nonlinear Expression (2), it is possible to take the linear function

$$\delta \simeq \alpha_\infty \int_0^t K(t - \tau)\Theta(\tau)d\tau, \qquad (5)$$

where α is the stabilized coefficient of thermal expansion; Θ is the temperature; $K(t - \tau)$ is given by Expression (3).

In Expression (5), the coefficient of thermal expansion is constant, but it depends on the temperature gradient, which, generally speaking, makes the medium nonuniform. Some approximate values of the coefficient α_∞ for different temperature drops and different soils are presented in Table 1, which is compiled from the author's data and from published sources.

The use of the linear model [Equation (5)] for design is entirely admissible in cases where it is necessary to predict frost cracking at temperatures below -2° and if it is known in advance that in the temperature range from 0° to -2° no cracking has occurred.

In addition to the information about the peculiarities of the development of temperature deformations in the permafrost, in order to predict the thermal fissuring it is necessary also to have the strength condition for variable loads and temperature available. The study of the problem of the fracture of frozen ground under variable loads is especially important in cases where the load temporarily exceeds the long-term strength, and it is not clear whether it, being a variable, can lead to rupture in a short time. The latter situation is encountered with temperature stresses in frozen ground in the case of harmonic variation of the temperature with time. As the studies demonstrated, the breaking of the samples of frozen soil under tension is determined by the reduction in the "immediate" strength under the applied load. If the latter is constant, then the rupture takes place at the time when the "immediate" strength drops to the magnitude of the applied load, which can be expressed by the corresponding equation of the hereditary type:

$$\sigma(t) \leq \sigma_{inst} - \int_0^t [\sigma(\tau) - \sigma_{1t}]K_1(t - \tau)d\tau, \qquad (6)$$

where $\sigma(t)$ is the variable stress with time, σ_{inst} is the "immediate" strength; σ_{1t} is the long-term strength, and $K_1(t - \tau)$ is the function of a succession of effects.

The function $K_1(t - \tau)$ is an increasing function. It is determined on the basis of the experimental data and can be approximated to by one of the following expressions:

$$K_1(t) = \frac{n+1}{T_1}\left(\frac{t}{T_1}\right)^n \text{ or } K_1(t) = \frac{1}{T_1} e^{\frac{t}{T_1}}, \qquad (7)$$

where T_1, n [> 1] are empirical constants.

TABLE 1 Some Values of the Coefficient of Thermal Expansion α_∞ and the Time of the Aftereffect t_0 of Frozen Soils

Soil	Moisture Contents, w, %	$\alpha_\infty \cdot 10^6$ per deg with a temperature drop of: -2° to 5°	-2° to 10°	-2° to 15°	-2° to 20°	t_0,h
Sand	20-25	50.0	35.0	35.0	35.0	0
	15-20	30.0	20.0	20.0	20.0	-
	10-15	20.0	16.0	16.0	16.0	-
	5-10	15.0	12.0	12.0	12.0	-
Supes	30-35	150.0	96.0	72.0	60.0	-
	20-25	230.0	150.0	110.0	90.0	20
	15-20	300.0	190.0	150.0	120.0	-
	10-15	400.0	260.0	190.0	160.0	36
Suglinok	40-45	125.0	80.0	60.0	50.0	-
	30-35	200.0	130.0	95.0	80.0	-
	20-25	380.0	240.0	180.0	150.0	50
	15-20	520.0	340.0	250.0	210.0	-
	10-15	750.0	480.0	360.0	300.0	-
Clay	35-40	230.0	-	-	-	50
	25-30	1,800.0	-	-	-	-
	20-25	1,200.0	-	-	-	120

The characteristic rupture time T_1 turned out to be related to the relaxation time of the frozen soil T_r by the following approximate expression[16]

$$T_1 \simeq 2T_r. \tag{8}$$

Let us consider the thermal stress field in a continuous mass of frozen ground that we have schematized as a continuous half-space with harmonic variation of the surface temperature. In order to describe the temperature field in the ground let us use the following known expression:

$$\Theta = \frac{Ame^{-\mu z}}{\sqrt{(m+\mu)^2 + \mu^2}} \cdot \sin(\omega t - \mu z + \gamma), \tag{8a}$$

where $m = \lambda_C/\lambda_M H_C$; $\mu = \sqrt{\omega/2\kappa_M}$; A is the amplitude of the negative air temperatures; ω is the frequency of temperature fluctuation; H_C is the thickness of the snow cover: λ_C, λ_M are the coefficients of thermal conductivity of the snow and the frozen ground; κ_M is the coefficient of thermal diffusivity of the frozen soil; z is the vertical coordinate reckoned from the surface downward; t is time;

$$\gamma = \arcsin\left[(m+\mu)/\sqrt{(m+\mu) - \mu^2}\right].$$

If we assume that the frozen ground acts as an elasto-visco-plastic medium,[17-20] then in order to calculate the horizontal temperature stresses σ_x considering the linear function [Equation (5)], it is possible to obtain the following expressions:

for $\sigma_x < \sigma_{1t}, \sigma_x$

$$= \frac{\alpha_\infty A_z E(\Theta)}{(1-\nu)\sqrt{1+\omega^2 t_0^2}} \sin(\omega t - \mu z + \gamma - \beta),$$

for $\sigma_x > \sigma_{1t}, \sigma_x - \bar{\sigma}_{1t}$

$$= \frac{\alpha_\infty A_z E(\Theta)\omega T_r \sin(\omega t - \mu z + \gamma - \beta + \Delta)}{(1-\nu)\sqrt{(1+\omega^2 t_0^2)(1+\omega^2 T_r^2)}}, \tag{9}$$

where

$$A_z = \frac{Ame^{-\mu z}}{\sqrt{(m+\mu)^2 + \mu^2}};$$

$$\beta = \arcsin\ \omega t_0/\sqrt{1 + \omega^2 + t_0^2};$$

$$\Delta = \arcsin(1/\sqrt{1 + \omega^2 T_r^2};$$

$$\bar{\sigma}_{1t}E(\Theta) = \int_0^{t/T} = \frac{\sigma_{1t}[\Theta(\nu)]}{E[\Theta(\nu)]} e^{-\left(\frac{t-\nu}{T_r}\right)} d\nu. \tag{10}$$

Expression (10) shows that in the plastic region the increase in stresses over the long-term strength is proportional to the frequency of the thermal stresses and, consequently, the low-frequency temperature fluctuations (for example, the annual ones) cannot in practice cause significant overstress. Therefore, the dynamics of the formation of frost fissures obviously looks like the following: long-period cooling leads to stresses equal to the long-term strength, and rupture is caused by secondary short-period temperature fluctuations.

Substituting Expression (10) in the strength condition [Equation (6)] and considering the fact that the long-term strength is exceeded more than once during the winter as a result of the secondary temperature fluctuations, we obtain the following condition of the possibility of the formation of a single frost crack:

$$A_z^1 + A_z^2 \geq \frac{(1-\nu)\sigma_{1t}(\Theta)}{\alpha_\infty E(\Theta)} + \frac{(1-\nu)(\sigma_{inst} - \sigma_{1t})\Phi(\omega_2)}{\alpha_\infty E(\Theta)[1 + A_z^2\omega_2/A_z^1\omega_1]}, \tag{11}$$

where A_z^1 and A_z^2 are the amplitudes of the long-period annual and short-period (several days) temperature fluctuations at the depth z; ω_1 and ω_2 are the frequencies of the long-period and short-period temperature fluctuations;

$$\phi(\omega_2) = 2(1 + 16\omega_2^2 T_r^2)/\omega_2 T_r (e^{\frac{\pi}{4\omega_2 T_r}} + 1)$$

$$= 12.5\omega_2^2 T_r^2.$$

Expression (11) essentially depends on the frequency of the secondary temperature fluctuations and shows that the higher this frequency, the larger the amplitude must be in order that the formation of the initial fissure become possible.

After the occurrence of the primary crack in the massif, the formation of the other fissures closest to it will be determined by the intensity of relief of the thermal stresses next to the primary fissure.[21] Assuming that the frost cracks are sufficiently deep, the problem of thermal stresses near a single fissure can be schematized as the problem of stresses in the massif having the shape of a quarter-space ($y \geq 0$, $z \geq 0$, z is the vertical coordinate), the boundaries of which are free of external loads, the air temperature varies according to a harmonic law, and at the boundaries there are boundary conditions of the third type, that is,

$$\left.\begin{aligned} z = 0;\ \frac{\partial\Theta}{\partial z} = m_1(\Theta - \Theta_A) \\ y = 0;\ \frac{\partial\Theta}{\partial y} = m_2(\Theta - \Theta_A) \end{aligned}\right\}, \tag{12}$$

where $m_1 = \lambda_i/\lambda_M H_i$; λ_i, H_i are the coefficient of thermal diffusivity and the thickness of the thermal insulation (for example, snow) at the boundaries of the quarter space; Θ_A is the air temperature; Θ is the ground temperature.

The solution of the thermoelastic problem by the Galerkin variation method gives the following values of the maximum tangential stresses at the boundaries of the massif

$$\left.\frac{(1-\nu)\sigma_y}{\alpha_\infty AE(\Theta)}\right|_{z=0} = \left[\frac{m_1}{\sqrt{(m_1+\mu)^2+\mu^2}} + \frac{m_2 e^{-\mu y}}{\sqrt{(m_2+\mu)^2+\mu^2}}\right](1-e^{-\mu y})^2$$

$$\left.\frac{(1-\nu)\sigma_z}{\alpha_\infty AE(\Theta)}\right|_{y=0} = \left[\frac{m_2}{\sqrt{(m_2+\mu)^2+\mu^2}} + \frac{m_1 e^{-\mu z}}{\sqrt{(m_1+\mu)^2+\mu^2}}\right](1-e^{-\mu z})^2, \quad (13)$$

where α_∞, A, E, and μ are the same as in Formulas (8-10).

By using the expressions in Equation 13 and the strength condition [Equation (6)], it is possible to solve the problem of the minimum distance between two adjacent cracks, that is, the problem of the width of the fissure polygons. In Figure 2 we show the graph for calculating the minimum distance between the frost cracks. On this graph, along with the notation already introduced above, additional notation is used: $A_0^{(1)}$ is the mean minumum surface temperature of the ground in the coldest month, $A_0^{(2)}$ is the amplitude of the secondary (6 to 7 days) temperature fluctuations of the ground surface during the coldest month, and Θ_1 is the test temperature at which the values of σ_{1t} and E are obtained. As is obvious from the graph, the width of the polygons depends sharply on the amplitude of the secondary temperature fluctuations at the ground surface during the winter: The smaller the amplitude, the greater the distance between the adjacent fissures. Let us also note that the solution of the same problem permits calculation of the maximum width of the opening of a vertical crack at the top S by the simple formula:

$$S = 4.5\alpha_\infty A_0^{(1)}/\sqrt{\omega_1/2\kappa_M} \quad (14)$$

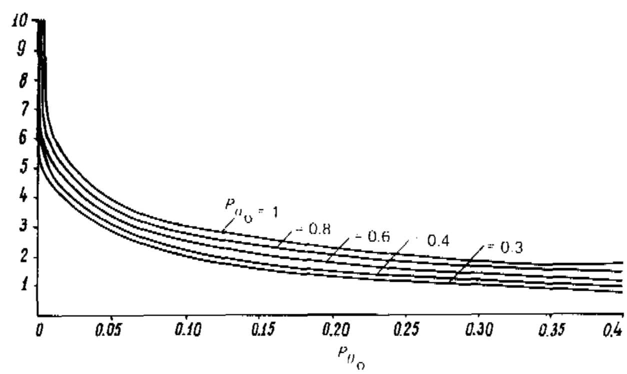

FIGURE 2 Reduced distance between frost cracks l^* as a function of the reduced temperature conditions P_{Θ_0} and the reduced accuracy [*sic*] of the frozen ground, P_{σ_0}, where:

$$l^* = l\sqrt{w_1/ZH_M};\quad P\sigma_0 = \frac{\sigma}{2\alpha_\infty A_0^{(1)}E}\sqrt{\frac{\Theta_1}{A_0^{(1)}}}$$

$$P_{\Theta_0} = \frac{A_0^2}{A_0^{(1)}\sqrt{1+w_2^2t_0^2}}.$$

The thermorheology of permafrost for underground storages cooled to negative temperatures is of special interest. The creation of an artificial frozen zone around the underground storage leads to the necessity for checking the possibility of frost fissuring of the soil, especially in the roof of the storage. In addition, the question of the optimal (from the point of view of excluding the possibility of thermal fissuring) regime of cooling the underground storage insuring the achievement of the given operating temperature Θ_K is both interesting and important.

For simplicity, considering the massif uniform (that is, a half-space), its properties not dependent on the temperature, and assuming as before that the frozen rock is an elasto-visco-plastic medium, it is possible to write the solution in the plastic region with a surface temperature of the pit $\Theta(t)$, which varies arbitrarily in time in the following way:

$$u(t) = \alpha_\infty \int_0^t \Theta(\tau)e^{-\frac{t-\tau}{T_r}}d\tau \text{ for } \sigma>\sigma_{1t} \quad (15)$$

where $u(t) = (1-\nu)(\sigma - \sigma_{1t}/E$; σ is the thermal stress.

The conditions imposing restrictions on the cooling regime, that is, on the function $\Theta(t)$ will be found by substituting Expression (15) in the strength condition [Equation (6)]:

$$u(t_K) << \frac{(\sigma_{inst} - \sigma_{1t})(1-\nu)}{E} - \int_0^{t_K} u(\tau)K_1(t_K-\tau)d\tau, \quad (16)$$

where t_K is the time in which the given operating temperature Θ_K is achieved; $u(t)$ is the temperature function defined by Expression (15).

In the following calculations we have used both the exponential and power forms of the function $K_1(t)$ in accordance with Expression (7). The studies were performed for the following possible cooling regimes:

1. The cooling rate of the surface of the pit is constant.
2. The air temperature in the storage is constant and equal to Θ_K, that is, $\Theta(t) \simeq \Theta_K$ [1 -

$\exp(-\kappa_M \lambda^2_{is}\, t/h^2_{is}\, \lambda)]$, where λ_{is} and h_{is} are the coefficients of thermal conductivity and thickness of the insulated layer on the walls of the pit.

3. The constant heat removal from the surface of the pit q = const, that is,[22]

$$\Theta = \Theta_K \sqrt{mt}, \qquad \text{where} \qquad m = 1.27q^2\kappa_M/\lambda^2\Theta^2 K.$$

The corresponding calculations by Formula (16) established that for elasto-visco-plastic media such as rocks and frozen soils there are characteristic critical temperature drops Θ^I_K and Θ^{II}_K defined considering the initial stresses by the following expressions:

$$\Theta^I_K = (\sigma_{inst} - \sigma_0)(1 - \nu)/\alpha_\infty E$$

$$\Theta^{II}_K = [k\sigma_{inst} - (k - 1)\sigma_{1t} - \sigma_0](1 - \nu)/\alpha_\infty E, \qquad (17)$$

where k fluctuates from 1.15 to 2.2 and can, on the average, be taken equal to 1.5.

The resistance of the rock to thermal fissuring is determined by the relation between the given operating temperature Θ_K and the critical values of Θ^I_K and Θ^{II}_K. The following cases occur here: (a) if $\Theta_K > \Theta^{II}_K$, then the deterioration of the rock is unavoidable for any cooling regime; (b) if $\Theta_K < \Theta^I_K$, then the deterioration does not take place for any cooling regime; (c) if $\Theta^I_K < \Theta_K < \Theta^{II}_K$, then the deterioration does not come only in the case where the optimal parameters of the cooling regime are observed. The optimal parameters for the investigated three cases are the following:

(1) For Θ = const

$$\Theta \cong 0.5[\Theta_K - (\sigma_{1t} - \sigma_0)(1 - \nu)/\alpha_\infty E/_{T_r}$$

(2) For the air temperature in the pit equal to Θ_K

$$h_{is} = \lambda_{is}\sqrt{\kappa_M T_r}/1.38\lambda$$

(3) For q = const

$$q = 0.7\lambda[\Theta_K - (\sigma_{1t} - \sigma_0)(1 - \nu)/\alpha_\infty E/\sqrt{\kappa_M T_r}$$

The above-investigated examples indicate a number of characteristic features that are imposed by the thermorheologic properties of frozen ground on predicting the thermal stresses and frost fissuring, both with harmonic variation of the temperature in the massif and artificial cooling of it.

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BASIC LAWS OF THE FORMATION OF THE COMPOSITION AND ENGINEERING-GEOLOGIC PROPERTIES OF FROZEN SOILS DURING FREEZING AND THAWING

V. A. KUDRYAVTSEV, N. F. POLTEV, N. I. TRUSH, AND YE. P. SHUSHERINA
Moscow Lomonosov State University

In northern regions under the effect of severe climatic conditions, soils are formed with characteristic composition and properties caused by the peculiarities of their genesis. Here we have in mind the fact that the genesis includes cryolithogenesis, that is, the processes of the formation of soil connected with the phase transition of water.

One of the peculiarities of soil deposits in the north having great engineering-geologic significance is the increased content of silt fractions. It is known that the upper (soil) layers subject to annual freezing and thawing are rich in silty particles. The results of cryologic studies in recent years indicate that the deeper layers of the permafrost also contain an increased amount of silt. The syngenetic series of permafrost, which on accumulation of the deposits have frozen many times, passing through the stage of seasonally thawed deposits and then buried in the frozen state under a new layer of deposition, are especially silty.

The variation in fineness of the loose soil during the process of cryolithogenesis takes place, on the one hand, as a result of breakdown of sandy and larger particles,[1] and, on the other hand, as a result of coagulation of colloids and clay particles.[2]

In the case of sandy and larger fractions, primary silty fractions are formed. On coagulation of the colloidal and clay fractions, secondary silt particles--microaggregates--are formed that remain after the rock thaws and exhibit the properties of a silty fraction.[1]

The sandy and larger particles experience thermal stresses during the weathering process and are subjected to brittle fracture at low temperature under the effect of stresses arising as a result of expansion of the water on its conversion to ice. Under the effect of these stresses, the sand particles do not always fracture. Sometimes only microcracks arise in them into which the water penetrates when the rock thaws, being absorbed on their free surface. The wedging pressure of the absorbed layer that completes the breakdown arises,[3,4] and as a result silty fractions are formed. The further process of breakdown diminishes, since with a decrease in size of the particles their strength increases. In addition, on the formation of the particles their deformation takes place along the free surfaces without rupture, without disturbance of the continuity; that is, it is realized as "cold treatment" of the particles, and their strength grows still more. Simultaneously with this, some silt particles become multicrystalline, which also increases their strength.

The secondary silty particles or silty microaggregates formed as a result of coagulation of colloidal and clay particles during the freezing of the soil also have stable dimensions. It is known[5] that the probability of cohesion during coagulation decreases with an increase in the size of the aggregates formed. The indicated phenomenon is explained by the fact that the discharging (loss of charge by a particle) takes place nonuniformly with respect to the entire surface of the particle. The cohesion is realized as a result of the sections of the particle surface where the discharge has gone farther, where the potential is below critical. As a result, the more strongly charged sections turn out to be outside, and the aggregates repel each other. "The greater the dimensions of the aggregates, the lower the possibility that other particles will cohere to them."[5]

During the process of cryolithogenesis, the ice content, structure, and texture of the soil change significantly. Aggregation and frost heave of the soil during freezing have important engineering-geologic significance. During thawing of icy soils, significant, nonuniform settlement occurs, and their bearing capacity is lowered. In other words, the negative freezing processes in the engineering-geological sense are connected with the iciness of the soils.

At the present time the Geocryology Department of Moscow State University has developed and made broad use of the methods of calculating heat and mass-transfer processes on a digital computer in freezing fine-grained soils. The calculations performed permit quantitative establishment of a number of laws of formation of the cryogenous structure essentially important from the point of view of the formation of engineering-geologic properties of freezing soils.

As has already been stated, the properties of loose frozen deposits are determined primarily by the composition and cryogenous structure and also the temperature regime of the frozen series.

The general laws of the dependence of the engineering-geologic properties on the indicated factors and, in particular, on the granulometric composition, the water content-iciness, the structure, and temperature can be illustrated in the example of mechanical properties from the data of various authors and including the materials of the Geocryology Department of Moscow State University.

Numerous experimental data obtained for various types of deformation and in a wide temperature range (to -60°C) indicate a reduction in resistance of the frozen soil with an increase in fineness.[6-11] This can be indicated by an analysis of the creep curves, the curves for the relation between the stress and strain for a different time of loading, the viscosity coefficients, and the short-term and long-term strengths.

The effect of the grain size is also exhibited in the nature of the failure of the frozen ground both for short-term and long-term load effects. With an increase in fineness, the failure of the frozen soil acquires a more plastic nature. This indicates the fact that the dependence of the mechanical properties of the frozen clay soil on its granulometric composition is caused to a significant degree by variation of the amount of unfrozen water and iciness, correspondingly. The form of the elementary (sandy, silty, and so on) soil particles and their aggregates is also felt in an essential way.

The degree of effect of grain size of the frozen soil on its properties significantly depends on the moisture content, temperature, and composition. With an increase in moisture, the effect of the granulometric composition on strength diminishes; the same thing is noted with an increase in temperature.

The interrelation of the basic characteristics of the composition of the frozen soil, such as grain size, moisture content, degree of saturation, and so on, and also temperature can be traced when investigating the dependence of the mechanical properties of the frozen soil on its total moisture content.

The experimental material[8-15] indicates a different nature of the dependence of the strength of the soil on the total moisture content, w.

An increase in the total moisture content of the frozen, incompletely saturated soil, independently of its grain size, temperature, and form of deformation, leads to an increase in strength connected primarily with the cementing effect of the ice.

With an increase in the total moisture content w of the frozen saturated soil, in the majority of cases a reduction in strength is observed at a diminishing rate. A reduction in strength with an increase in moisture content under the conditions of practically complete saturation and an increase in the porosity, respectively, is caused primarily by the weakening of the structural bonds between the particles of the matrix and their aggregates, as a result of separation by the ice and weakening of the granular framework, which is not compensated for completely by the cementing effect of the ice.

Independently of the indicated conditions of variation of the total moisture content, the frozen soil can be both stronger and weaker than the ice. The grain size and temperature have a significant effect on the degree of the noted relations.

As a rule, the variation in the total moisture content *w* occurs under appreciably more complicated conditions by comparison with those investigated above. In the various ranges of variation of *w*, both the porosity and the degree of saturation can vary. Accordingly, the dependence of the strength on *w* can have a highly complex nature, which, however, can be analyzed on the basis of the above-investigated cases.

It must be noted that the above-investigated general form of the dependence of the strength of the frozen soil on its moisture content pertains to the case of the short-term loading. With a prolonged loading, the indicated relation can be different, because the ice is gradually converted from a strengthening component to a weakening component.

The structure formed both by freezing and during the process of cryogenesis has a significant effect on the mechanical properties of the frozen soil.

The layering formed in the soil before freezing is also reflected in the mechanical properties after freezing.[6] From a comparison of the shear creep curves of the frozen varved clay, it follows that, when the shear plane passes along the layering, the deformation of the soil at any point in time is greater than when the shear plane is perpendicular to the layers.

The dependence of the strength of the frozen soil on the cryogenous structure is especially clearly exhibited in the temperature range close to 0°C.[16] The experimental data in the temperature range from -0.4° to -2° indicate that for close values of the total moisture content clay soil of massive texture is characterized by smaller instantaneous shear strength, cohesion, and friction than in the case of reticular texture; here, with an increase in thickness of the ice inclusions, the soil strength increases. The indicated relation also occurs for a sufficiently long action of the load (up to 2.5 to 3 months). The noted laws can be explained on the basis of comparing the strength of the ice and the contacts of the frozen soil. At temperatures close to 0°, the contacts of the mineral grains with the ice cement are, by comparison with the ice, weaker sections because of the presence of unfrozen water. However, under the prolonged effect of a load and also with a reduction in temperature, the effect of the cryogenous texture can be different, as a result of possible variation in the ratio of the ice strength and the contacts of the frozen soil.

The cryogenous texture of the frozen soil has an effect on the compressibility of the soil, both in the negative temperature range (above -4° -5°)[17] and in the settlement of the frozen soil during thawing.[18] The role of the cryogenous texture is also significant in the formation of the soil strength during and after thawing.[19]

The effect of the temperature on the mechanical properties of the frozen soil can be traced when investigating deformation characteristics.[6,8] From a comparison of the creep curves and the viscosity coefficients at various temperatures, it follows that a reduction in temperature causes the same effect as a reduction in grain size--an increase in strength of the frozen soil. An analogous law is noted also when comparing the curves reflecting the relation between the stress and strain of frozen soils of different grain size at different temperatures. The dependence of the deformation properties of the frozen soil on the temperature can, according to S. S. Vyalov and S. E. Gorodetskiy, be described by the following expressions:

$$\sigma = A(t)\varepsilon^{m},$$

where σ is the stress, ε is the strain, $A(t)$ is a parameter that depends on the time t and the temperature Θ;

$$A(t) = \xi t^{-\lambda},$$

where λ in practice does not depend on Θ, and

$$\xi(\Theta) = \omega(\Theta + 1)^{\kappa}$$

As is known, on transition through 0°C, a sharp increase in strength of the soil occurs that is connected with the conversion of water to ice. A further reduction in the temperature of the frozen soil to -20°, according to numerous experimental data, is characterized by a smooth increase in strength occurring independently of the grain size with damped intensity.[6-9,20,21] This nature of variation of the strength of the frozen soil, just as the above-noted dependence of the deformation properties on the temperature, is caused by an increase in the strength of the ice and viscosity of the unfrozen water and also further crystallization of the water and internal structural transformations.

The studies in the field of low negative temperatures (to -60°C) performed at the Department of Geocryology of Moscow State University[20,21] demonstrate that below -20° the smooth increase in strength with damped intensity can be observed only for frozen sand. For frozen clay soils (supes, suglinok, clay) at a temperature below -20°, significantly more intense increase in strength σ takes place than in the temperature range from 0° to -20° (the σ--Θ curve has an inflection point). Obviously, this is connected with the partial freezing of new portions of water.

The above-investigated relations for the strength σ as a function of the temperature Θ pertain to the short-term action of the load. With a long-term action of the load, the dependence of the strength of the frozen soil on the temperature attenuates somewhat,[6] which is connected primarily with the rheologic properties of the ice. With an increase in the time of action of the load, the σ--Θ curves become more gentle (to -20°). However, the temperature and long-term strength dependence of the frozen soil is exhibited quite clearly. The relation between the short- and long-term strength of the permafrost as a function of temperature, according to S. S. Vyalov, can be expressed by equations of the following type:

$$\sigma(t) = \frac{\beta}{\ln(t + 1)/B}, \text{ where } \beta = \omega(\Theta + 1)^{\alpha}.$$

As is obvious from what has been discussed above, the basic laws of formation of engineering-geologic properties of the frozen soil are determined by their composition and structure, which are caused by their genesis and also the thermal regime of the frozen beds. With multiple freezing and thawing, grain breakdown, coagulation, and aggregation of the soil particles are noted. This leads to a significant change in the soil itself, its composition, and, consequently, its properties. The moisture and ice content of the rock and its cryogenous structure and texture vary significantly during the process of cryolithogenesis. The temperature regime of the frozen beds also varies to a significant degree. The freezing conditions determine the nature of the cryogenous structure and texture of the frozen soil.

Thus, with investigation of the properties of the frozen soils, it is necessary to consider the nature of their cryolithogenesis and the modern conditions of existence determined by the entire complex of the natural environment. For each genetic type of soil, it is necessary to distinguish the nature of the cryolithogenesis, the type of cryogenous texture, and cryogenous structure. The study of the history of formation of soils, and especially their freezing, offers the possibility of discovering the laws of spatial arrangement (with respect to areas and depth) of the types of cryogenous structure and probable variation of the properties of the soil. The subsequent variations of the complex of natural conditions can lead to some variation in the primary cryogenous textures, both as a result of thawing and secondary freezing and as a result of redistribution (migration) of the moisture with respect to the section in connection with variation in the temperature regime of the frozen soil.

In the case of construction or other economic development of a region, the natural conditions change significantly. This leads to variation in turn in the temperature regime of the soil and, consequently, to a change in soil properties.

Therefore, in the case of engineering-geological evaluation of the frozen soil, on the basis of studying the entire complex of natural conditions, it is necessary to compare the freezing prognosis for the direction of control of the soil properties for construction purposes.

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PHYSICAL AND MECHANICAL PROPERTIES OF SALINE SOILS

N. A. TSYTOVICH, YA. A. KRONIK, K. F. MARKIN, V. I. AKSENOV AND M. V. SAMUEL'SON Moscow Kuybyshev Construction Engineering Institute, LenZNIIEP Institute

In many permafrost areas, especially along the coast of the northern seas and oceans and in continental regions with a moisture shortage, saline soils are often used as the base for buildings or the materials of earth structures.[1-5] Artificially salinized soil[5-10] is finding broader and broader application in construction in the Far North. Depending on the amount of salt contained in it and the temperature, the salinized soil can be in the permafrozen, seasonally frozen, or seasonally thawed state, and it can be characterized correspondingly by significantly different strength and deformational and permeable characteristics. The properties of such soil vary with the time under the effect of the climatic, hydrogeological, and other factors and also under the effect of the economic activity of man and the development of the northern regions by man. The specific peculiarities of the saline soil require that the types of soil be distinguished.

The problem of studying saline soil in the geocryolithozone and use of this soil in construction has been insufficiently studied. At the present time in geocryological practice, there are no standard classifications for saline soil and no united procedures for determining the properties of this soil. Construction practice also lacks planning, design, and construction Norms for building on saline soil and Norms for the application of saline soil in earthern structures. Moreover, in certain forms of construction (for example, in road and hydroengineering), the use of strongly saline soil is not recommended.

In this report the classification and procedure for determining the physical characteristics of saline soils are proposed. The properties of frozen saline clay soils and the strength properties of saline frozen soils are described.

CLASSIFICATION OF SALINE SOILS

The basic index of saline soils--the salinity z--characterizes the salt content per unit weight of dry soil. The minimum salinity such that when exceeded the soil must be of the saline type is determined by the significant variation of its properties by comparison with the analogous nonsaline soil, and it depends to a significant degree on the granulometric composition of the soil. Therefore, it is proposed that the difference in granulometric composition of the soils be used as the basis for classifying saline soils.

On the basis of many years of experience in studying the properties of saline soils and analysis of the published data,[2-4,8-12] it is proposed that four basic types of saline soil be isolated, each of which, in turn, must be subdivided into groups and subgroups with respect to genesis, degree of salinity, structural-textural attributes, and the degree of settlement when the salt is washed away (chemical suffusion) and when thawing (in the case of frozen saline ground): (1) rocks and semirocks with salinity $z \geq 3$ percent in the thawed state and with $z \geq 2$ percent in the present state; (2) coarsely clastic saline soils with $z \geq 1$ percent thawed and with $z \geq 0.5$ percent frozen; (3) sandy saline soils with $z \geq 0.5$ percent thawed and with $z \geq 0.1$ percent frozen; (4) clayey saline soils with $z \geq 0.25$ percent thawed and $z \geq 0.1$ percent frozen.

In this paper basically a study is made of coarse clastics with fine-grained filler and sandy and clayey soils salinized with fast-soluble and medium-soluble salts,[2] and peat not investigated in the report as special forms of saline soils.

The presence of salt in the soil changes its properties significantly, especially in the case of the excess-saline and strong-saline and high-temperature frozen soils. Thus, for example, the presence of 0.1 percent salt in frozen sandy soils lowers the strength indexes by 10 percent to 30 percent at a temperature of t_r = -2°C to -4°C and by 40 percent and more at $t_r \geq$ -1°C. The freezing point of the pore soil solution at z = 0.1 percent decreases to -0.2°C to -0.6°C, which for high-temperature frozen soils has very great significance. The permeability of saline soil can vary by 1-2 orders and more.[4,8,9,11]

It is possible to subdivide (in the first approximation) soils the following way by the degree of salinity (Table 1).

For an approximate determination of the freezing point of the pore soil solutions of saline soil with sufficient accuracy for construction practice (to 10-20 percent), it is possible to use solubility tables for the salts and the freezing points of their solutions, determining the concentration of the pore solution C_{ps} by the previously discussed procedure. In the case of predominance in the pore solution of NaCl salt, for example, for the saline soils of the coastal regions where sodium chloride is 60-80 percent, the proposed procedure provides sufficiently high accuracy.

BASIC PHYSICAL CHARACTERISTICS OF SALINE SOILS

The basic physical characteristics of saline soils are γ_s--unit weight; γ_{ss}--unit weight of solids; γ_c--unit weight of crystalline salt (in the case of excess and strongly saline soils); w_s--moisture content; z--salinity determined by the dry residue of aqueous extract--for easily soluble salt, or hydrochloric acid extract--for medium-soluble salt (for example, for gypsum soils); w_{uv}--moisture content with respect to nonsolution pore water; p_s--the specific content of coarsely clastic particles (more than 2mm) as a function of the dry weight of all the soil (for coarsely clastic saline soils); w_{us}--amount of unfrozen water in the frozen saline soil.

The studies by Ya. A. Kronik[5,8,11] established that w_{uv} = (0.9-0.95) w_{nc}, where w_{nc} is the amount of tightly bound water according to A. F. Lebedev for nonsaline soil. In construction practice it is possible with sufficient accuracy to set $w_{uv} = w_{nc}$ and for saline sandy and

TABLE 1 Subdivision of Saline Soils by Degree of Salinity

Degree of Salinization of the Soil	Salt Content of the Soil z, %					
	Coarsely Clastic		Sandy		Clayey	
	Thawed	Frozen	Thawed	Frozen	Thawed	Frozen
Weak	1 to 2	0.5 to 1	0.5 to 1	0.1 to 0.25	0.25 to 0.5	0.1 to 0.25
Medium	2 to 5	1 to 3	1 to 3	0.25 to 0.5	0.5 to 2	0.25 to 0.5
Strong	5 to 10	3 to 5	3 to 8	0.5 to 2	2 to 5	0.5 to 2
Excessive	> 10	> 5	> 8	> 2	> 5	> 2

TABLE 2 Physical and Physicochemical Characteristics of Saline Soils

Characteristics Determined Directly	Characteristics Determined by Calculation	Calculation Formulas
1	2	3
Basic		
γ_s--unit weight of saline soil	γ_{ds}--dry unit weight of the saline soil solids	$\gamma_{ds} = \gamma_s/(1 + w_s)$
γ_{ss}--grain unit weight of the saline soil	γ_d--dry unit weight of the soil matrix matrix (without considering salt)	$\gamma_d = \gamma_{ds}(1 - z)$
γ_c--unit weight of crystalline salt	γ_y--dry unit weight of the soil (without considering salt)	$\gamma_y = \gamma_{ss} \cdot \gamma_c(1 - z)/(\gamma_c - \gamma_{ss})$
w_s--moisture content of saline soil	w--soil moisture content (without considering salt)	$w = w_s/(1 - z)$
z--salinity of the ground	C_z--unit salinity of the ground	$C_z = z \cdot \gamma_{ds}$
w_{uv}--moisture content of the "nonsolute volume"	C_{ps}--concentration of the pore solution	$C_{ps} = C_z/V_{ps} = z\gamma_{ps}/[w_s - w_{uv} + (1 + w_{uv})z]$
w_u--amount of unfrozen water	For soil salinized with NaCl	$C_{ps} = z/[w_s - w_{uv} + (0.34 + w_{uv})z]$
p_s--number of coarse clastic particles	For soil salinized with $CaCl_2$	$C_{ps} = z/[w_s - w_{uv} + (0.26 + w_{uv})z]$
Auxiliary		
C_{ps}--concentration of the pore solution	γ_p--unit weight of the salt solution or the pore	$\gamma_{ps} = G_{ps}/V_{ps}$
γ_p--unit weight of the solution or γ_{ps} is the pore solution	For NaCl salt	$\gamma_{ps} = 1 + 0.66 \cdot C_{ps}$
	For $CaCl_2$ salt	$\gamma_{ps} = 1 + 0.74 \cdot C_{ps}$
	n--porosity of the ground	$n = (\gamma_y - \gamma_d)/\gamma_y$
C_t--equilibrium concentration of the solution	ε--void ratio	$\varepsilon = (\gamma_y - \gamma_d)/\gamma_d$
C_{BB}--concentration of the aqueous extraction of the ground	w_{ts}--moisture content of the saline ground	$w_{ts} = (n \cdot \gamma_{pr} - C_z)/\gamma_{ds}$
w_a--moisture content of the strongly bound water according to A. F. Lebedev	I_s--saturation coefficient of the saline soil	$I_s = w_s/w_a$
C_{ae}--concentration of the acid extract of the ground	n_D--diffusion porosity of saline soil	$n_D = \gamma_{ds}[w_s - w_{uv} + (1 + w_{uv})z]/\gamma_{ps}$

TABLE 2 (Continued)

Characteristics Determined Directly	Characteristics Determined by Calculation	Calculation Formulas
1	2	3
Formulas for calculating the salinity by the data from titrating an aqueous extract of the ground:	V_{os}--volume of air per cm^3 of saline soil	Simplified formula: $n_D = \gamma_{ds}(w_s - w_{uv})$
Exact		$V_{os} = 1[\gamma_d((1/\gamma_y) + (w_{uv}/\gamma_{uv})) + n_D]$
$z = (w_{ae} - w_{uv}) \cdot C_{ae}/(\gamma_B - (1 + w_{uv}) \cdot C_{ae})$	γ_s--total volumetric ice content of frozen saline soil	$\lambda_s = \gamma_s(w_s - w_u)/0.9 \cdot (1 + w)$
Simplified		
$z = (w_{ae} - w_{uv}) \cdot C_{ae}$ for the ae ratio of soil:water-1:10	i_s--relative ice content of the frozen saline soil	$i_s = 1 - (w_u/w_s)$
$z \approx 10 \cdot C_{ae}$		

NOTE: In the table the volumetric concentrations are assumed ($C_{vol} = C_{weight} \cdot \gamma_p$).

coarsely clastic soil with sandy filler $w_{uv} \approx w_{nc} \approx 0$.

If it is possible to determine the concentration of the pore soil solution without preparing the extract, for example, by a salinometer, in place of the salinity the basic characteristic of the salt content in the soil can be the concentration of the pore solution C_{ps} determined considering the nonsolute volume of moisture by the formulas presented in Table 2. In coarsely clastic soil with fine-grained filler, the salinity is determined for their fines filler.

According to the above indicated basic physical indexes, all of the remaining physicochemical and physicomechanical characteristics of saline soil are determined. Therefore, the procedure and accuracy of determining them have especially great significance. In Table 2 we show the calculation formulas for determining all the characteristics of saline soil.

The amount of unfrozen water in the frozen saline soil was determined by the calorimetric method. For soil temperatures corresponding in practice to the frozen state, according to N. A. Tsytovich,[13] the amount of unfrozen water can be taken equal to the amount of adsorbed water $w_u \approx w_{nc} \approx w_{uv}$.

The above method of determining the physical characteristics of saline soils, sufficiently checked out during 7 yr in geotechnical practice in northern dam construction,[5,8,11] permits compilation of the salt balance in the soil and, on the basis of this, solution of a broad circle of engineering problems with respect to salt transport in the soil under various thermodynamic conditions and also approximate evaluation of the structural parameters of saline soils and physical characteristics of their components.

CLASSIFICATION OF SALINE CLAYEY SOILS

Saline clayey soils must be divided into (1) supesses, (2) suglinoks, (3) clays, and (4) structurally unstable macroporous soil (silts, loess, and so on).

The isolated types of soil having a number of identical properties caused by the general physicochemical nature of interaction of the salt solutions with the clay particles also have some distinguishing features caused by their genesis, mineralogic composition, temperature, type and ratio of the salt ions contained in the soil, the composition of the exchange complex, the degree of settlement, the structural-textural peculiarities, and so on. Therefore, each of the isolated types of saline soil requires special study and consideration for engineering research and construction.

Let us consider the most general properties characteristic of the majority of saline clay soils. Depending on the degree of salinity and the temperature, the clay soils can in the first approximation be characterized by the following phase composition of the water and the structural-textural variations (Table 3).

The studies by the authors[5,8,10,11] demonstrated that the structural properties of the saline soils depend primarily on their moisture content, density, temperature, and concentration of the pore solution. It has been experimentally established that the concentration of the pore solution C_{ps} is one of the main regulators of the processes of structure formation in clay soils predetermining their aqueous-physical and mechanical properties.

In thawed ground with a gradual increase in concentration of the pore solution, an increase

TABLE 3 Phase Composition of the Water and Structural Properties of Saline Soils

Degree of Salinity of the Soil	Thermal State of the Ground	Salinity, %	Temperature Range of the Phase Transitions, °C			Form of the Presence of Salt in the Ground		Predominant Types of Structures (According to P. A. Rebinder)
			Min[a]	Max[b]	Mean	Easily soluble	Mediumly soluble	
Weak	Frozen	0.10 to 0.25	-0.6 to -2.5	-0.2 to -0.8	-0.4 to -1.6	In the frozen pore solution (the saline pore ice)[c]		Crystallization[d] and coagulation-condensation)[e]
	Thawed	0.25 to 0.50	-0.8 to -5.5	-0.4 to -1.5	-0.6 to -3.5	In the pore solution		Dispersion and coagulation
Medium	Frozen	0.25 to 0.50	-0.8 to -5.5	-0.4 to -1.5	-0.6 to -3.5	Saline pore ice	Saline pore ice and crystals	Coagulation-crystallization and crystallization
	Thawed	0.50 to 2.0	-1.7 to -39.5	-0.8 to -4.1	-1.2 to -21.8	In the pore solution	In the pore solution and in crystals	Coagulation and coagulation-condensation
Strong	Frozen	0.50 to 2.0	-1.7 to -39.5	-0.8 to -4.1	-1.2 to -21.8	Saline pore ice and crystals		Crystallization and coagulation-crystallization
	Thawed	2.0 to 5.0	-7.1 to -55.0	-2.7 to -27.0	-4.9 to -41.0	In the saturated pore solution and in crystals		Crystallization and coagulation-condensation
In excess	Frozen	>2	$t_s \leq -4$ to -8[f]			Saline pore ice, crystals, aggregates, layers of salt		
	Thawed	>5	$t_s \leq -8$ to -12[f]			Saturated pore solution, crystals, aggregates, layers of salt		Coagulation and coagulation-crystallization

[a] Values are close to the artificial dry soil.
[b] Values closer to the natural wet soil.
[c] For ground temperature below the temperatures of the intense phase transitions but above the temperatures of the eutectic points of NaCl and CaCl ($t_{eutectic}$ = -21.0°C and -55.0°C).
[d] For solidly-frozen soil.
[e] For plastically-frozen soil.
[f] For a large content of crystal salt and supersaturated pore solutions, the ground can freeze at higher (negative) temperatures.

in fineness of the soil aggregates, an increase in the plasticity limits, and a reduction in the permeability (Figures 1,2) are observed. The maximum increase in strength is achieved for concentrations of the pore solutions close to 1 *N*. A further increase in C_{ps} causes intense development of coagulation and aggregation of the soil particles, which causes a reduction in strength and plasticity limits. Therefore, the optimal concentration C_{ps} = 1 *N* is called the aggregation threshold.[4,8,11]

In the mean concentration range C_{ps} = 0.06 to 0.25 g/cm^3, dynamic equilibrium of the coagulation and dispersion is observed, and the strength and deformational properties of the thawed saline soil essentially do not differ from those of the nonsaline soil (Figures 3,4).

For high concentrations (C_{ps} > 0.25 g/cm^3) when for solutions of certain types of salts the state of complete saturation is achieved, the intense aggregation of the soil particles begins causing a significant reduction in the Atterberg indexes, a decrease in strength, an increase in compressibility, and an increase in the adsorption, water-absorptive soils,[4,8,11,14] which must be considered in practical use in the construction of strongly and excess-saline soil.

It is possible to achieve an improvement in

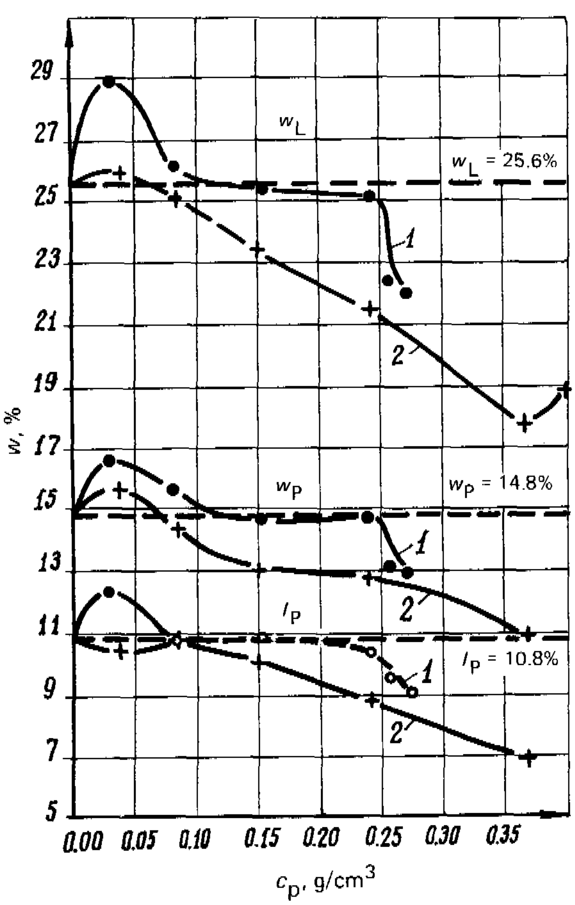

FIGURE 1 Variations of the [Atterberg] indexes (w_L, w_p, and I_p) of Vilyuy suglinok as a function of the concentration C_{ps} of introduced solutions of the salts NaCl (1) and $CaCl_2$ (2).

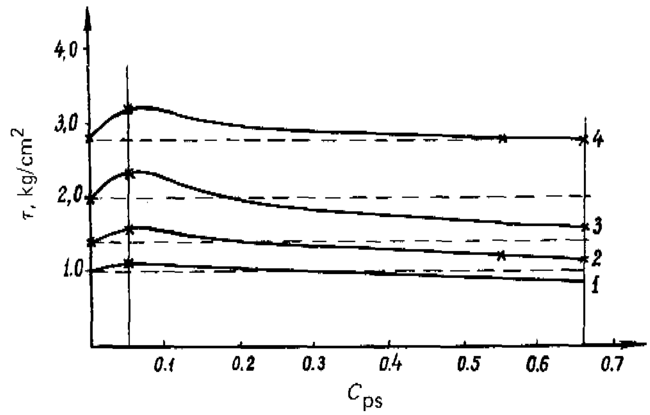

FIGURE 2 Shear strength τ_s of Vilyuy suglinok, made saline with $CaCl_2$, and pore solution concentration:

1. σ_1 = 1 kg/cm^2; γ_0 = 1.73 to 1.81 g/cm^3
2. σ_2 = 2 kg/cm^2; γ_0 = 1.74 to 1.81 g/cm^3
3. σ_3 = 4 kg/cm^2; γ_0 = 1.74 to 1.83 g/cm^3
4. σ_4 = 4 kg/cm^2; γ_0 = 2.05 to 2.33 g/cm^3

the structural properties of a saline soil by reducing its water content and increasing the density. The change in moisture content causes a corresponding change in concentration of the pore solution, which also predetermines the above-described processes of microstructuring in clay soils and the formation of coagulation structures. At the same time an increase in the soil density also accelerates the processes of structure formation. Here, in the density range equal to 0.98 to 1.03 of the maximum standard density, intense formation of coagulation-condensation and partial-cementation structures cause a sharp increase in salinity of the soil and its compaction begins. Therefore, the supercompacting of the thawed saline clay soil above 1.03 of the maximum standard is recommended in construction practice as one of the effective means of strengthening the soil, ensuring the creation in the supercompacted soil of a practically waterproof high-strength structure with sufficiently stable and, in practice, long-lasting (at least up to 10 yr) irreversible noncementation bonds.

The highly significant variations of the structural properties of the saline soils with time take place during their desalting, the nature and rate of which are determined by three basic factors: (a) percolation leaching out, (b) diffusion removal of the salt, and (c) desalting under the effect of climatic factors--variations in the temperature-moisture regime and especially with multiple freezing and thawing of the soil. The results of the investigations on the desalting of soil are presented in references 4, 11, and 12.

The variations in the concentration of the pore solution also cause sharp variations of the freezing and melting points of the soil and therefore to a significant degree affect the strength and deformational properties of the frozen and thawing soil.

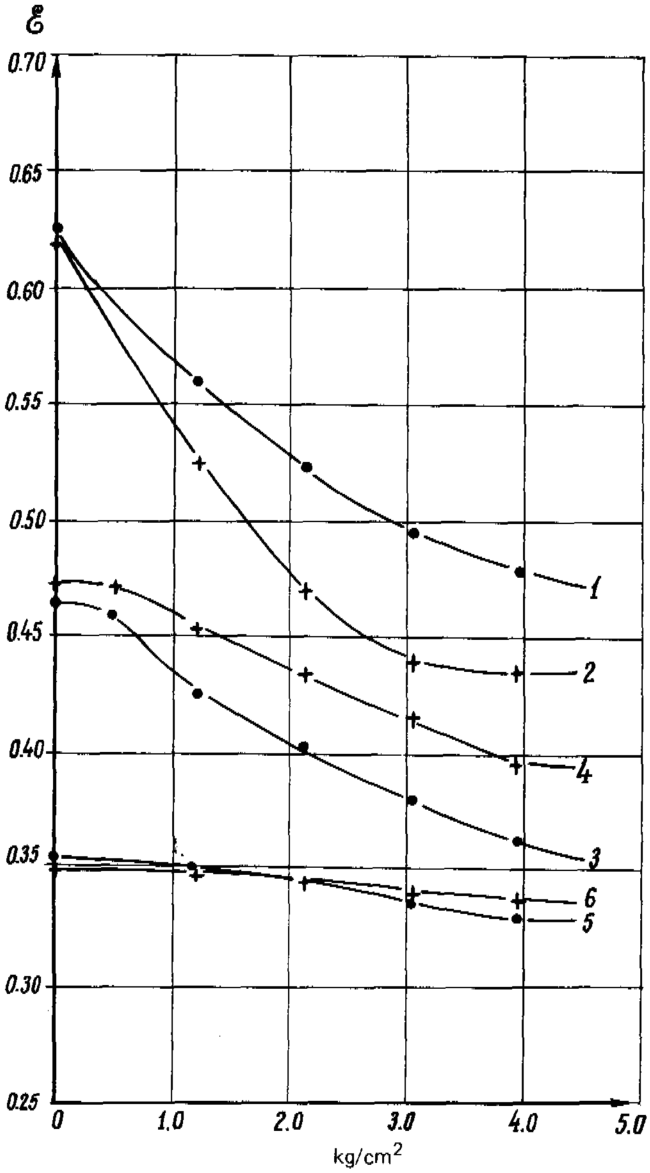

FIGURE 3 Compression curves for nonsaline (2, 4,6) and saline (1,3,5) solution of $CaCl_2$ with C_{ps} = 0.37 to 0.40 g/cm^3, of detrital Vilyuy suglinoks (compression under water):

1,2--γ_0 = 2.0 g/cm^3; w_0 = 11.0%; w_{0k} = 16.2%
3,4--γ_0 = 2.22 g/cm^3; w_0 = 11.1%; w_{0k} = 11.3%
5,6--γ_0 = 2.40 g/cm^3; w_0 = 11.0%; w_{0k} = 11.0%

STRENGTH PROPERTIES OF SALINE FROZEN SOIL

There are extremely few data on the experimental study of the strength of frozen saline soils. In the early studies,[1,3] the absence of theoretical data did not permit a full analysis of the effect of the concentrations of the pore solution. The minimum natural salinity of the investigated soil, as a rule, did not exceed 0.25 percent. A comparison of the data obtained with similar nonsaline frozen soil demonstrated that the strength with this salinity is reduced by 1.5 to 2 times.

The results of the experimental studies performed in order to discover the effect of the concentration of the pore solution on the strength of the frozen ground are then presented. In the experiments the ball-indenter method of N. A. Tsytovich[5] was used to determine the equivalent cohesion as the most available evaluation method of the strength and rheologic properties of the soil.

The experiments were performed with spherical indenters with a diameter of 17 mm in the chambers of the Amderma underground laboratory with constant temperatures of -1°, -2°, -3°, and -4°C. The soil samples were prepared in large lots for simultaneous study in all chambers. The experiments were performed on fine silty sand of two kinds: supes and suglinok. As a material for manufacturing the paste of the sandy soil, nonsaline fine quartz sand and a silt-clay mixture obtained from suglinok by grinding it in dry form and sifting through a sieve with holes 0.1 mm; distilled water and salt crystals were also used. The granulometric composition and plasticity limits of the investigated soils are presented in Table 4.

The samples of sandy, silty sand of the first type contained 25 percent of silty-clay fractions, and the second type, 45 percent. In both cases during preparation, after careful mixing a defined quantity of solution corresponding to the total moisture capacity without compacting was added to the dry mixture. The wet mixture was again mixed and held at a temperature of 18°C to 20°C for 24 h. The moisture content of the supes was assumed to be at the liquid limit w_L = 23.5 percent.

The mineralogic composition of the fine frac-

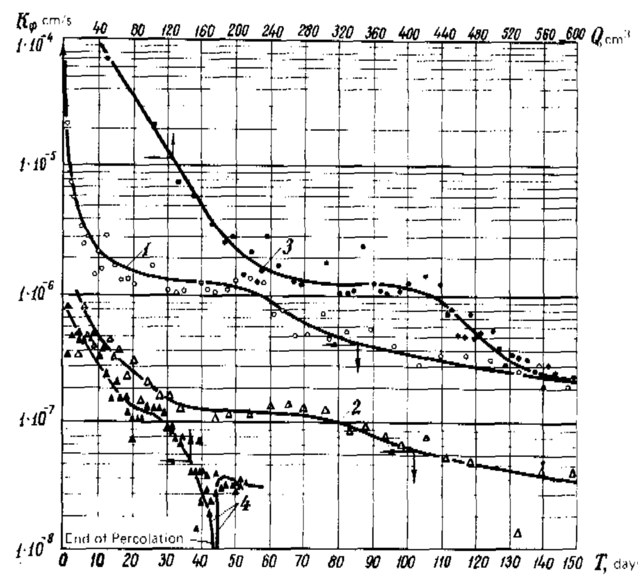

FIGURE 4 Variation of the permeability k_ϕ of the Vilyuy detrital suglinok with desalting and as a function of the amount of percolated water Q_ϕ. Samples were first made saline by percolating a saturated NaCl solution to C_{ps} = 0.275 g/cm^3. 1,3--for specimens with γ_d = 1.54 g/cm^3; 2,4--for samples with γ_d = 2.16 g/cm^3.

TABLE 4 Characteristics of the Soil Used

Name of of Soil	Granulometric composition, % 2-1 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.1 mm	0.1-0.05 mm	0.05-0.01 mm	0.01-0.002 mm	<0.002	Limits Liquid w_L	Limits Plastic w_P	Plasticity Index I_P
Quartz sand	trace	trace	trace	98	1	1	--	--	--	--	--
Silt-clay mixture	--	--	--	--	44	34	18	4	25	17	8
Hydromica supes	--	trace	1	7	50	30	7	5	23.5	17.5	6
Quaternary suglinok	--	trace	trace	7	31	17	26	19	37	20	17

tions of supes and suglinok is uniform, with predominance of hydromica and with an insignificant admixture of montmorillonite. The salt composition for the studies was assumed close to that of the local soil and corresponding to the composition of the salts in seawater. In the vicinity of Amderma, where seawater was taken for evaporation, the ion composition of the salts was as follows:

Ca^{+2}	Mg^{+2}	K^{+1}	Na^{+1}	Cl^{-1}	SO_4^{-2}	CO_3^{-2}
1.2 to 1.4%	3.5 to 7.5%	0.5 to 1.0%	31 to 40%	42 to 55%	3 to 7%	to 5.0%

The adopted concentrations are indicated in Table 5. The experimental studies gave the following results: The presence in the frozen ground of water-soluble salts assumed in the experiments to have a composition even of insignificant concentration not exceeding 0.2 percent (which corresponds to 0.03 *N* for NaCl) significantly reduces the long-term equivalent cohesion of the ground determined by the Tsytovich ball test.

The sharpest reduction in strength is observed in sandy soils with a concentration up to 0.5 percent (Table 5). In supes and sand with a large content of silt-clay fractions, a reduction in strength with an increase in concentration appears to be less marked. This fact is explained obviously by the fact that with an increase in the salt concentration the ice-cement strength is reduced appreciably more sharply than the cohesion as a result of interaction of the mineral particles. The role of these strength components in the soil with different content of clay fractions is different.

The reduction in strength as a function of the

TABLE 5 Values of C^8_{eq} for Soil of Different Granulometric Composition

Type of Soil	Salt Concentration %	C^8_{eq}, kg/cm^2 for values of t, °C: -1	-2	-3	-4
Quartz sand with a silt-clay fraction content of 25%, w = 19%	0.0	4.7	5.4	7.9	9.3
	0.26	1.5	3.1	5.7	7.1
	0.52	0.6	1.2	2.3	3.5
	1.04	--	0.6	1.2	2.3
	2.6	--	--	0.4	0.6
The same with a silt-clay fraction content of 45%, w = 22%	0.26	2.6	3.7	6.5	7.4
	0.52	1.3	2.6	--	--
	2.6	--		0.6	0.7
Hydromica supes $w = w_L$ = 23.5%	0.23-0.27	2.6	6.1	0.8	12.5
	0.64	2.5	3.7	7.0	11.3
	1.1-1.2	1.2	2.1	5.3	8.7
	2.1-2.5	--	0.4	0.9	1.5
	4.4-5.0	--	--	--	0.4
Quaternary suglinok (hydromica with montmorillonite admixture), w_P = 17, $w = w_L$ = 37%	1.35	--	0.64	1.56	2.35
	2.2	--	0.46	0.78	1.05
	3.2	--	--	0.54	0.87
	4.3	--	--	--	0.5

concentration of the solution appears at temperatures close to zero, corresponding to the significant phase transformation. Thus, when $t = -4°C$ the solution concentration of 2.5 percent causes a reduction in strength in silty sand by 15 to 20 times. In supes with the same concentration the strength drops over the entire temperature range. It is possible to propose that at this concentration the reduction in temperature does not cause phase transformations as a result of total suppression of the diffused layer, and the cohesion caused by interaction of solid particles varies insignificantly in this temperature range.

The rheologic properties of the investigated frozen ground within the concentration limits from 0 percent to 2.5 percent vary insignificantly. The characteristic rheologic parameter--the ratio of the long-term equivalent cohesion C_{eq} to the 8-h cohesion--$C^8{}_{eq}$--varies from 0.4 to 0.9 in all experiments with clayey soil. Here, the law of variation of this ratio as a function of the concentration of the pore solution or temperature was not established. The great scattering of this coefficient indicates that when determining the strength of saline frozen ground using ball indenters in each series of experiments it is necessary experimentally to establish the ratio $C_{eq}/C^8{}_{eq}$.

An important property of saline clay soil is the strengthening of it under load established in the experiment. When determining C_{eq} of the supesses under a load from 1 to 9 kg by a ball 17 mm in diameter, it was established that the magnitude of $C^8{}_{eq}$ varies 2 to 2.5 times. In the sand the hardening phenomenon is expressed significantly more weakly.

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PHYSICAL AND MECHANICAL PROPERTIES OF FROZEN AND THAWING COARSE CLASTIC SOILS

N. A. TSYTOVICH AND YA. A. KRONIK Moscow Kuybyshev Engineering-Construction Institute

Quaternary surface deposits of the Far North are frequently represented in the form of permafrozen (or, frequently, seasonally thawing) coarse clastic soils distinguished from typical soils by a number of specific peculiarities. This soil has been studied insufficiently so that at the present time in practice there are no generally accepted procedures for determining the physical and mechanical characteristics or Norms with respect to their use in construction. In the last decade, in connection with the beginning of intensive development of the Far North, many researchers have started deeper studies of the given problem.[1-11]

In this report basic results are generalized from many years of studies of the genetic peculiarities, composition, and properties of coarse soils with a clay filler of certain regions of the Far North (the Kola and Taymyr peninsulas; the Khantaya, Vilyuy, and Kolyma rivers) performed at the MISI Kuybyshev Institute at the Laboratory of Hydroengineering Geocryology (Hydrogeocryology). The work was started in 1964 in connection with the problems of the practice of hydroengineering construction in permafrost.

GENETIC PECULIARITIES OF COARSE CLASTIC SOILS IN THE FAR NORTH

According to the results of the studies of the authors[4,5,10,11] performed basically on Vilyuy, Khantay, and Kola coarse clastic soils with fine-grained filler, the following basic genetic peculiarities were discovered, that essentially distinguished them from standard cohesive soils and coarse noncohesive soils:

1. Occurrence as a thin bed on the bedrock and significant nonuniformity of the soil with respect to moisture and granulometric composition.
2. Permafrost (more rarely seasonally frozen) state, excess supersaturation and transition during thawing to the fluid-plastic and fluid state with the phenomena of subsidence and thermokarst.
3. The appearance of thixotropic-quick properties (thixotropic liquefaction during dynamic effects (for example, from the effect of construction machinery and through transportation).
4. Significant content of coarse particles--up to 60-70 percent.
5. Severe weathering of coarse particles and high water-absorbing capacity (up to 4-5 percent for the Khantay pebble-gravel particles and up to 6-7 percent for the Vilyuy detrital particles (Figure 1).
6. High specific weight of the mineral particles, γ_S = 2.80 to 3.00 g/cm^3 and higher.
7. Predominance in the fine-grained component of the soil of silt fractions, the clay mineral montmorillonite, and multivalent cations in the exchange complex.
8. High heave of the investigated soil (for example, up to 25 percent for Vilyuy soil and up to 50 percent for Khantay soil).

The established and investigated peculiarities considered in more detail in references[4,5,10] are explained basically by genesis of the given soil--the Quaternary surface deposits of the regions of the Far North, and they are not considered, as a rule, in engineering-geological research, in the case of geocryologic research, and in the case of geotechnical control in construction. Accordingly, the authors propose a classification of coarse soils in the Far North and methods of determining their basic properties.

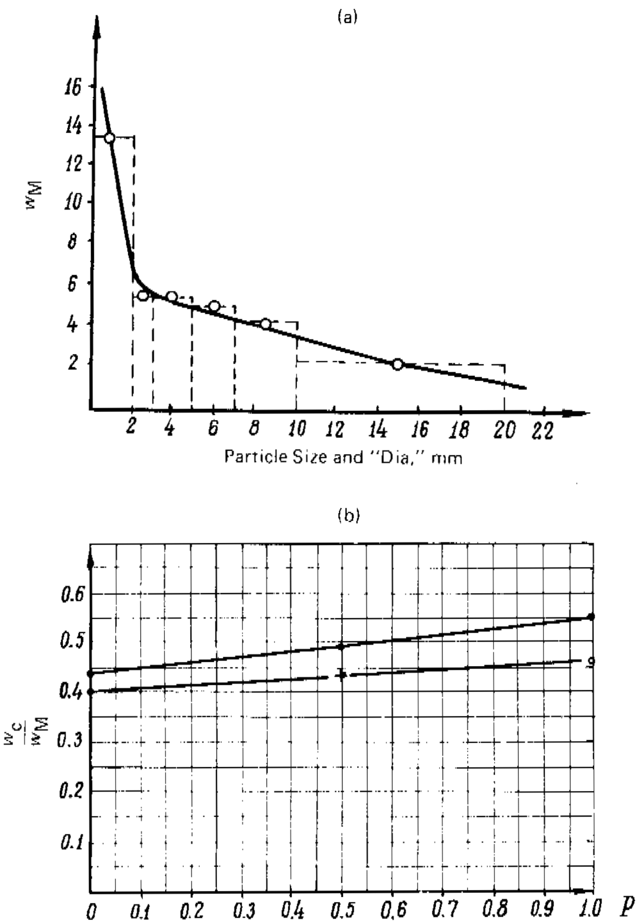

FIGURE 1 Water-physical properties of Vilyuy coarse soils: (a) amount of physically held moisture w_M for different fractions of Vilyuy coarse soils; (b) relative content of strongly bound water w_B as a function of the content of coarse fragments p.

CLASSIFICATIONS OF COARSE CLASTIC SOILS

The most important characteristic of the coarse soils with fine-grained filler is the index of the content of coarse particles (more than 2 mm)--p--which permits evaluation in the first approximation of the property of all the coarse soil (or soil mixture) and establishment of the close interrelation between the physical and mechanical properties of all of the soil and its fine-soil filler. It is recommended that the content by weight of the coarse particles p in the coarse soil with fine filler be determined with sufficient accuracy for construction purposes by the results of screening the dried weighed samples of soil through a 2 mm mesh. For more exact studies and in geotechnical practice, the value of p must be determined by wet screening-washing of the wet coarse soil with water through the screen with subsequent separate drying of the fines and fragments.

Depending on the magnitude of p, the properties of the coarse soil differ quite significantly, and with a high fines content they are predetermined basically by the filler properties. With a defined ratio of clastic particles and fines where the fragments are directly in contact with each other, by forming a solid framework (dense packing) and the fines are in the pores, the properties of all the soil depend in practice to a small degree on the properties of the fine-grained filler. In frozen coarse soils to a certain value of p, the presence of ice, as a new component phase, is felt primarily in the variation and properties of the filler, which, in turn, will predetermine the properties of all the frozen coarse soils. However, for some p corresponding to a dense packing of the coarse fragments, the ice will influence the properties of the coarse particles directly, separating them during heave or being a plastic binder (plasto-viscous lubricant) between the fragments, significantly changing the properties, especially the strength and deformational properties of all the frozen coarse soil.

On the basis of a generalization of the data of,[1-20], it is proposed that the coarse soils be classified by a function of the content of coarse rock fragments p and their relation to the fine-grained filler in the following four classes:

Class I--Coarse clastic soil with fine-grained filler with a content of coarse (more than 2 mm) particles p from 10 to 35 percent.

Class II--the same with p = 35 to 70 percent.

Class III--the same with p = 70 to 90 percent.

Class IV--Coarse soil without filler--p > 90 percent.

The soil of Class I is characterized by the fact that the main volume of the soil is fine-grained, and the coarse inclusions are distributed quite uniformly in it ("they float in the fines"), not coming into contact in practice. All properties of such soil are predetermined by the properties of the fine-grained filler, and in the case of deformation of the coarse soil the fragments do not in practice come into contact with each other.[4,5,17-19] The water-physical, permeability, and mechanical characteristics of this soil correlate quite well with analogous indexes of their fines filler with recalculation of the latter for the entire coarse soil considering the content of coarse particles p or filler $(1 - p)$.[2,4,5,10,16] The compaction coefficient* of the particles [18,19] in such soil is positive, and the strength and deformational characteristics of the soil are appreciably higher than those for standard fine-grained soil without coarse clastic inclusions. The cryogenous texture and properties of the frozen and thawing soil of this class are also predetermined by the texture and properties of the frozen-fines filler, and the thermophysical characteristics of the soil are essentially (30 to 80 percent)

*Undefined--probably a coefficient, including a critical void ratio and measuring dilatancy.

different from the indexes for the standard frozen soil without coarse clastic inclusions.

The properties of the soil in Class II are also predetermined by the properties of the fine-grained filler, and in this sense the soil is close to Class I soil. However, there are significant differences. Above all, the quantitative indexes of the properties of soils in Class II differ sharply from the characteristics of soils in Class I: The water-physical properties vary significantly, the strength increases sharply, the permeability increases, and the values of the thermophysical characteristics increase. For the majority of coarse clastic particles, the packing coefficient has a negative value,[18,19] and the fragments, forming comparatively tighter packing and taking up appreciably greater volume in the filler surrounding them, can during deformation of the rock enter into contact with each other, that is, they partially cohere. Here, along with the sliding friction of the particles, rolling friction also is exhibited. The soils in this class include the majority of permafrost soils and thawing coarse clastics of the surface deposits of the Far North of residual and residual-colluvial origin, [4,5,10] and their soil mixtures are widely used as the materials of the impervious elements of rock-fill dams.

With a large-particle content of about 65-70 percent by weight, all of the fragments come into close contact, forming a tight skeleton, and the filler fills only the pores between the fragments. In this case the content and porosity of the fines are determined by the content and volume of the pores in the coarsely clastic framework. The properties of such coarse soils differ sharply from the soils of Class I and II, and therefore they are separated into a separate Class III with p = 70-90 percent.

For the soils of Class III, relatively little effect of the filler on the properties of all the soil and a significant increase in the role of the most coarse clastic particles are acharacteristic. The packing coefficient of the particles is negative,[18,19] and the free displacement of the clastic particles in the fines, as occurs in soils of Class I and II, is in practice impossible. In compression, the clastic skeleton takes all the load, and the deformability of the ground essentially decreases, acquiring the nature of quasielastic deformations. Simultaneously, the strength of the soil and the permeability increase significantly. For the development of shear deformations in such soil it is necessary to overcome cohesion of the particles; here the rolling friction of the clastic particles one with respect to the other prevails, and the ground in the shear zone breaks up.[7,10,13,19] For frozen clastic soils of Class III with complete filling of the pores with frozen fines, the development of significant creep deformations is characteristic.[3,7,8] The ground in this class includes the frequently encountered alluvial and coarse clastic soil similar to it of bimodal granulometric composition[15] and also the frozen coarsely clastic soil of placer origin, which is widespread in the northeastern part of the USSR.[1,7,8]

The coarse clastic soils of Class IV with p > 90 percent are a discontinuous medium of practically uncemented coarse clastic particles that form a rigid skeleton with voids. The properties of these soil types are predetermined by the properties of the coarse clastic particles (their density, shape, strength, and so on) and do not in practice depend on the filler, the small content of which can be neglected. For frozen ice-saturated soil of this class, the development of creep deformation to some degree is also characteristic.[3,7,8] The thawed coarse clastic soils of Class IV are distinguished by high permeability, and the mechanical properties of these soils do not in practice depend on the moisture content and degree of saturation.

Each of the indicated classes of coarse clastic soil is subdivided in turn into types depending on the shape of the particles (scree, gravel, detritus, pebbles) and the type of filler (sand, supes, suglinok, clay), and it requires independent study and consideration for engineering research and construction.

Let us consider in more detail the physical and mechanical properties of the coarse soils with fines filler of certain regions of the Far North in the example of the Vilyuy detrital suglinoks and the Khantay pebbly gravel supesses and suglinoks that belong to Classes I and II of coarse fragmented soil.

WATER-PHYSICAL PROPERTIES OF COARSE CLASTIC SOILS WITH CLAY FILLER

For Classes I-III of coarse clastic soils, the water-physical properties are predetermined by the water-physical properties of the filler. The soils in Classes I and II with a clay filler have plastic properties that are entirely determined by the plasticity of the filler. In the case of sandy filler of Class I and II soils and also Class III and IV soils, independently of the type of filler, the soils in practice have no plasticity. In order to determine the plasticity limits of the coarse clastic soil in Classes I and II, it is necessary primarily to note exactly how to determine the total moisture content of the coarsely clastic soil w, the moisture content of its filler w_f and the moisutre content of the coarsely clastic particles w_K. The procedure for determining these indexes have been discussed in detail.[4,5,10] The moisture content of the coarse clastic soil is determined by the standard method--drying at t = +105 to 110°C of an average sample of coarse clastic soil weighing 2 kg to 4 kg.

The importance of w_f and the moisture content of the coarse clastic particles w_K is determined by calculation by the following formulas:

1. For the case of small amounts of moisture (all the moisture is bound) $w \leq w_M$,

$$w_K = \frac{p_w}{p}, \qquad (1)$$

$$w_f = \frac{1 - p_w}{1 - p}, \qquad (2)$$

2. For the case of large amounts of moisture (the moisture is bound and free) $w > w_M$,

$$w_K = w_M = \text{const}, \qquad (3)$$

$$w_f = \frac{w - p \cdot w_{MK}}{1 - p}, \qquad (4)$$

$$w_f = w_{MK} + \frac{w - w_M}{1 - p}, \qquad (4a)$$

where p is the content by weight of the coarse fragments in the entire soil (as a fraction), p_w is the moisture content coefficient of the coarsely clastic particles

$$p_w = \frac{p \cdot w_{MK}}{w_M}, \qquad (5)$$

and w_M, w_{MF}, and w_{MK} are the maximum molecular moisture content of the entire soil, its filler, and coarse particles, respectively, determined with sufficient accuracy for construction practice by the MISI Institute procedures and the method of moisture-containing media of A. F. Lebedev.

The plasticity indexes of the coarsely clastic ground are predetermined by the plasticity limits of the filler, and they are calculated by the following formulas:

1. The plastic limit of the coarse clastic soil w_P:

$$w_P = w_{PF}(1 - p) + w_{MK} \cdot p, \qquad (6)$$

2. The liquid limit of the coarse clastic soil w_L:

$$w_L = w_{LF}(1 - p) + w_{MK} \cdot p, \qquad (7)$$

where w_{PF} and w_{LF} are the plastic limit and liquid limit of the filler determined by standard procedures.

The use of the given procedure for soils in Classes I and II demonstrates its quite high reliability and satisfactory agreement with the measured data (Figures 2 and 3).

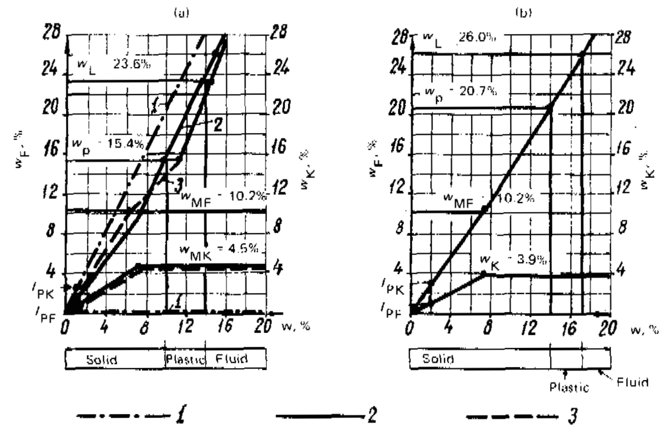

FIGURE 2 Diagrams of the physical state with respect to moisture (consistency) of the Vilyuy detrital suglinoks (p = 0.5). (a) deposits 11; (b) deposits 12. 1--curve calculated by the existing procedure, 2--curve calculated by the MISI procedure, and 3--experimental curve.

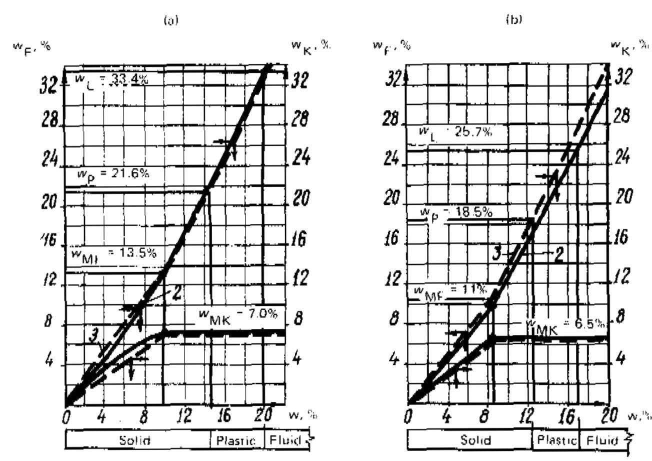

FIGURE 3 Diagrams of the physical state with respect to moisture (consistency) of Khantay coarse clastic soils. (a) gravelly suglinok of deposit 8 (p = 0.5); (b) pebbly gravelly supes of deposit 9 (p = 0.4). The provisional notation is the same as in Figure 2.

DETERMINATION OF THE DENSITY OF A COARSE CLASTIC SOIL

For coarse clastic soil, it is necessary to distinguish the density of all the soil--the unit weight γ and the dry unit weight γ_d--and the density of its filler--the dry unit weight of the filler γ_{df}, which is an important geotechnical index of the coarse clastic soil used in construction. The volumetric weight of a coarse clastic soil is determined by methods analogous to those for cohesive soils with the exception of the cutting ring method, the application of which gives significant errors. It is preferable to determine the volumetric weight of the coarse clastic soil by the "lune" method,* selecting the lune no less than 20 × 20 × 20 cm, covering it with polyethylene film and filling with water or sand, or by measurements of the weight, size, and volume of a frozen chunk of the soil.

For coarse clastic ground of Classes I and II, the dry unit weight of the filler γ_{df} is determined by the following MISI formula:[4,5]

*The field density test.

$$\gamma_{df} = \frac{\gamma_{dk} \cdot \gamma_d \cdot (1 - p)}{(\gamma_{dk} - \gamma_d \cdot p)\left(1 + \frac{w_k \cdot \gamma_y}{\gamma_w}\right)}, \qquad (8)$$

where γ_{dk} is the unit weight of the coarse particles, γ_w is the unit weight of the water contained in the coarse clastic particles, assumed in construction practice to be equal to 1 g/cm^3.

COMPACTION OF COARSE CLASTIC SOILS WITH FINE-GRAINED FILLER

The coarse clastic soils Classes I and II have increased capacity for being compacted under the effect of the equipment used when putting it into earthern structures and fills. The basic characteristics of compaction of an investigated soil is the maximum dry unit weight $\gamma_{d \cdot max}$ and the optimal moisture content, w_{opt}, determined by the DORNII Institute procedure or by the modified AASHO method[4,5]--are presented in the Table 1.

The compaction indexes with respect to the modified AASHO procedure coincide more exactly with the data with respect to compaction of Vilyuy and Khantay coarse clastic soil under natural production conditions with layer-by-layer rolling by heavy compaction equipment, in particular, the MAZ-525 dump trucks. The correlation between the maximum densitites obtained by the above-indicated procedures was found in the form

$$\gamma_d(\text{AASHO max}) = (1.03 \text{ to } 1.06)\gamma_d(\text{DORNII max}).$$

Considering the significant complexity, the amount of labor and not always high accuracy of the laboratory compaction testing of the coarse clastic soils in Classes I and II, it is proposed that the maximum density and optimal moisture content be determined by calculation using the standard compaction data for the filler $\gamma_{ddf \cdot max}$ and $w_{F \cdot opt}$. Here the calculation is performed with the following formulas:

$$\gamma_{d \cdot max} = \frac{\gamma_{df\ max}\gamma_k}{\gamma_k(1 - p) + \gamma_{df\ max}p(1 + \gamma_k w_k)} \qquad (9)$$

$$w_{opt} = w_{f \cdot opt}(1 - p) + w_k p. \qquad (10)$$

An experimental test demonstrated the entirely satisfactory agreement of the data obtained experimentally and calculated by the proposed formulas (Figure 4).

TABLE 1 Compaction Indexes of Coarse Clastic Soils

Item No.	Type of Soil	Content of Large Particles, p	Compaction Indexes: By the DORNII Institute Procedure w_{opt}, %	$\gamma_{d \cdot max}$, g/cm^3	Compaction Indexes: By the AASHO Method w_{opt}, %	$\gamma_{d \cdot max}$, g/cm^3
1	Vilyuy detrital suglinok (deposit 11)	0.5	11.2	2.13	10.4	2.20
2	Vilyuy suglinok filler (deposit 11)	--	14.9	1.853		
3	Gravelly suglinok Khantay (deposit 8)	0.5	8.6	2.20	8.0	2.34
4	Khantay gravelly suglinok (deposit 8)	0.3	9.6	2.13	--	--
5	Khantay suglinok filler (deposit 8)	--	13.2	1.99	--	--

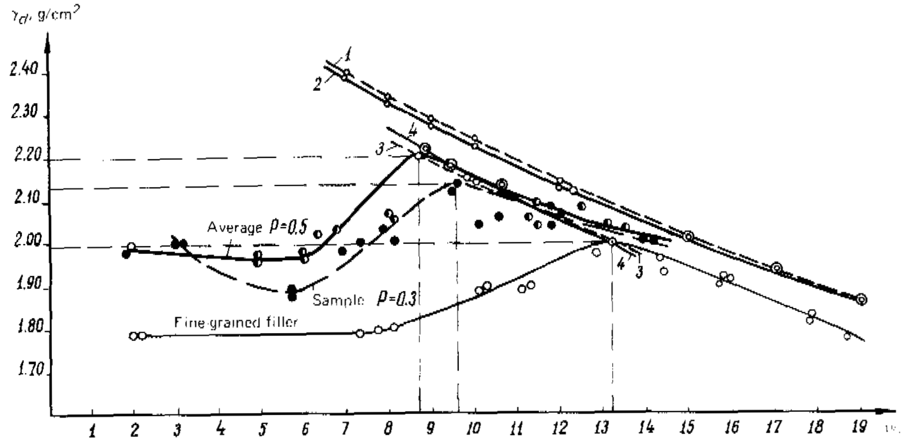

FIGURE 4 Maximum standard density $\gamma_{d.max}$ of Khantay pebbly-gravelly suglinok as a function of the moisture content w with standard compaction by the DORNII procedure. 1 and 2--curves for the maximum densities with water saturation for filler (1) and all the coarse clastic soil (2), 3--curve for the maximum standard densities according to the experimental data, and 4--curve for the maximum standard densities calculated by the MISI formula (9). (1) fine grained filler, (2) average sample, and (3) γ_d, g/cm^3.

WEATHERING AND HYGROSCOPIC WATER CAPACITY OF COARSE CLASTIC SOILS

The above-discussed methods of determining the physical and mechanical characteristics of coarse clastic soils with fine-grained fillers take into account the weathering of the clastic particles in terms of the maximum molecular moisture content of the fragments w_{MK} directly characterizing the moisture adsorption of the clastic particles as a result of microcracks and macrocracks formed by weathering.

The magnitude of w_{MK} is determined by the data from 3-day water absorption of dry fragments placed in water,[4,5,10] and for the given case of a coarse clastic soil it is in practice a constant characterizing the degree of weathering of the clastic particles with respect to water adsorption.

For alluvial and similar weakly weathered, coarse clastic soils, the water adsorption and degree of weathering are very small, and they can be neglected in construction practice, assuming that all of the moisture in such ground is contained in the filler. However, the coarse clastic soils of the regions of the Far North, as a rule, are strongly weathered and consideration of the degree of weathering of them with respect to water adsorption is mandatory.

For the approximate rapid method of determining the degree of weathering with respective water adsorption (with accuracy to 0.5 to 1 percent in the decreasing direction), the value of w_{MK} can be determined by the water absorption of a weighed sample of large fragments in 1 h to 2 h. It is necessary especially to note that with this procedure for water adsorption, the moisture capacity characteristic obtained for the fragments will correspond to the degree of water adsorption i, called the parameter of state with respect to moisture,[17,20] and it is the basic index of the degree of weathering of the residual coarse clastic soil used in foreign engineering-geologic practice.[20]

In this report new criteria are proposed for the degree of weathering of coarse clastic soils--a total weathering coefficient K_B and relative weathering B_K. Inasmuch as the basic and final result of the weathering processes is the formation of a network of microcracks and macrocracks, in the fragments predetermining all the physical and mechanical properties and in the final analysis causing rupture of the fragments, it is logical to take the integral characteristics of the total K_B and the relative B_K volume occupied by the cracks per unit volume of individual fragments as the weathering coefficient of the coarse clastic fragments. Assuming with some approximation that in the case of maximum moisture content the moisture adsorbed by the fragments fills all the micropores and macropores, for determination of the total weathering coefficient K_B and the relative weathering coefficient B_K corresponding to the average value of the void ratio of the individual fragments ε_K, the following formulas are obtained:

$$K_B = \frac{w_{MK} \cdot \gamma_K}{\gamma_w + w_{M\lambda K}} \quad (11)$$

$$B_K = \frac{w_{MK} \cdot \gamma_K}{\gamma_w} = \frac{K_B}{1 - K_B} \quad (12)$$

By the values of the coefficient of total weathering K_B, it is proposed that the coarse clastic soils be subdivided in practice: unweathered with $K_B < 2.5$, weakly weathered K_B = 2.5 to 5 percent, average weathered K_B = 5 to 10 percent, strongly weathered K_B = 10 to 20 percent, very strongly weathered $K_B > 20$ percent. According to the classification data, the Vilyuy and Khantay coarse clastic soils are strongly weathered (K_B = 11.5 to 17.5 percent and B_K = 13 to 21 percent.

Further studies by the authors are being performed at the MISI Kuybyshev Institute in the direction of considering the degree of weathering and genetic peculiarities of the coarse clastic soils of the Far North when studying their physicomechanical, percolation, and thermophysical properties.

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BASIC LAWS OF CRYOGENOUS PROCESSES IN SILT-CLAY AND SANDY FORMATIONS

P. F. SHVETSOV, S. YE. GRECHISCHEV, AND L. V. CHISTOTINOV All-Union Scientific Research Institute of Hydrogeology and Engineering Geology

The physicogeographic processes that are caused by energy and mass exchange in freezing, frozen, and thawing soil and rock are called cryogenous. These processes promote nonuniform changes in the temperature, pressure, and concentration of the solutions causing phase transformations of the water and its movement and also additional stresses* in the soil-ground complex. The results of these processes--structural and textural new forms in the soil and rock, the movement of their macrovolumes, and, as a result, the occurrence of forms of micro- and mesorelief--are called cryogenous phenomena.

The freezing, frozen nonequilibrium, and thawing soil complexes, the series and massifs of rock are highly complex physicochemical systems connected with other systems (the environment) by three forms of interactions (contacts): (1) mechanical, (2) thermal, and (3) material (mass exchange). It would be more correct to call them physicogeological systems.

The laws of variation of state and movement of the components of these systems can be understood by investigating them using four methods: the lithologic-petrographic, mechanical, statistical (beginning with the molecular-kinetic representations), and thermodynamic, understood in the broadest sense of the word. The value and the results of the lithographic-petrographic method are shown in the literature,[1-4] and so on.

The mechanical method is a component part of the general thermodynamic method. In spite of this, when investigating cryogenous processes, it is isolated as independent. On this basis results that are important to practice and that have long been used by engineering geocryology were obtained. Something analogous occurs with statistical and thermodynamic methods, which mutually complement each other. However, the basic principles and the possibilities of applying them to the ground systems are different.

Thermodynamics investigates the microscopic systems, the spatial dimensions and the time of existence of which are sufficient for performing the normal measurement operations. The apparatus of thermodynamics is constructed so that it is easy to consider all of the phenomenological laws. The methods of statistics as applied to liquids and especially to such systems as wet-sand and silty-clay formations are less fruitful: the mathematical apparatus is awkward. Now let us make a comment regarding the peculiarities of the classical thermodynamics preventing broad application in geocryology.

The processes of heat exchange and, in particular, thermal conductivity of classical thermodynamics have not been investigated at all.

*The additional stresses, which are called external forces, are connected with the gravitational acceleration, the acceleration of the circular motion of the entire macrosystem, magnetization, and external loads on the system.

Therefore, geocryologists, using the Fourier theory of thermal conductivity, which appeared before thermodynamics, did not see any necessity for using the methods of thermodynamics to study the laws of freezing and thawing of soil and rock. The problem of the occurrence of the processes in time was not stated by classical thermodynamics. As a result, it was found that heat transferred between the topsoil-ground complex and the atmosphere was the natural result of the spontaneous process of freezing of this complex.

Indeed, the freezing of the soil-ground complex, just as any spontaneous process in nature, is accompanied not only by heat exchange with the environment but also the work done on the external environment as a result of internal energy of the soil and rock.

The basic conditions of the development of cryogenous processes must be considered to be heat losses by the soil-ground complex or heat influxes to them connected with the heat exchange cycles of different duration. Let us divide these processes into three types.

1. The processes caused by the heat losses of the deposits to the atmosphere as a result of internal energy. They take place in the freezing and cooling frozen series and generate ascending vertical movements opposite to the gravitational force and interruptions of continuity of the soil series (frost heave and fissuring and also hydroeffusive processes).

2. Processes caused by heat influxes to the soil series from the external environment and completed at the expense of solar energy and gravitational and kinetic energy of the moving water. They take place in thawing and heated frozen series and are accompanied by the descending movements of the ground mass. The most widespread morphogenetic effective and, in practice, important form of this type of cryogenous process is considered to be the thermokarst process.

3. The processes caused by both components--outgoing and incoming (as a result of internal and external energy with interaction of a number of geophysical fields). They take place in the seasonally freezing and thawing layer. They include the slope processes caused above all by periodic heating and thawing of the ground series.

In the development of modern cryogenous processes, the manifestation of the law of broad zonality is observed. The thermal cycles of the soil-ground complex and especially their increments caused by 2- to 3-yr warmings increase from the south to north within the permafrost area. This is connected with a decrease in the relative equilibrium of the soil and ground complex in the same direction.[5-6]

With all the difference of the silty-clay formations from sandy ones, they turn out to be similar in thermodynamic respects with moisture somewhat less than the maximum molecular moisture content of one formation or another. The processes of crystallization of film water and thawing of the ice formed from it are nonisothermal. In this case the cryogenous processes are weakly expressed and do not have great practical significance.

The sharpest difference in sandy, and silty-clay, formations comes when both are saturated or supersaturated with water and ice. The most intense cryogenous phenomena are connected with the freezing and thawing of these ground masses.

Even in the case of freezing of the sandy ground mass containing in practice only gravitational water, part of the internal energy, although a small part by comparison with that which goes to the environment, is spent on doing the work of expansion of the water, when it freezes in the pores, and compression of the mineral particles. The freezing of weak aqueous solutions in the pores of the sandy formations and the freezing of macrovolumes of the latter is thus represented by the isothermal process corresponding to a decrease in enthalpy H of the ground mass:

$$-\Delta H = \Delta u + p\Delta v, \qquad (1)$$

where u is the internal energy, p is the pressure, and v is volume.

The second term in (1) expresses all of the work of the so-called heave of the sandy ground mass. In the case of thawing, this component has opposite sign--the work of compaction is done as a result of the external energy sources (heat and compression of the overburden).

This elementary representation is not at all justified when describing thermodynamic processes in the macrovolumes of silty-clay formations. They are complex, colloidal systems. A similar formation taken in the macrovolume is characterized by the average parameters of state, and it is considered by a number of researchers as a single-phase molecular-colloidal solution.[7,8]

In the silty-clay ground mass, along with the attached water coating individual mineral particles, there is a significant amount of free aqueous solution. The combination of film and gravitational moisture of the silty-clay skeleton is a required condition of the development of intense cryogenous processes in the soil-ground complex. Applying the methods of classical thermodynamics and the thermodynamics of the irreversible processes to the study of the laws of such processes, it is possible to obtain a number of important results.[5,9]

In a number of cases during the thawing of silty-clay as a result of solar heat, only the work of swelling of the mineral interlayers and the formation of the new interfaces between the particles of the system can be completed. The fact itself of gradual rupture (dispersal) and swelling during thawing of the strongly packed and low-ice mineral interlayers under the effect of the influx of heat and water to them were experimentally established by F. G. Bakulin and V. F. Zhukov.

In the general case, the variation of the specific surface Q and free energy F during the freezing or thawing process in the silty-clay soil is expressed by the Gibbs equation:

$$-dF = sdT + pdv + \sigma dQ - \mu dn, \qquad (2)$$

where Δ_h = heave. The last condition (6) appears quite important since it establishes the relation between the heaving of the soil and the possibility of thermal cracking: namely, the greater the heave, Δ_h, the more the ground is subject to cracking, which fully coincides with the general thermodynamic arguments discussed above.

It is also necessary to note that the amplitude and frequency of the temperature fluctuations have a great effect on the possibility of frost cracking: the greater they are, the more probable the crack formation. This follows from Formula (5) and agrees with the energy concept of cryogenous processes.[16]

Up to now we have approached the analysis of the laws of cryogenous processes, considering wet ground as a unique system, not subdividing it into individual components. At the same time it is known that the nature of the course of many of the processes in soil is determined by the properties and the state of the water in the soil. On the other hand, the adsorption, wetting, capillary nature, and other phenomena on the surface depend to a great extent on the physical and physicochemical properties on the surface of the soil particles. The comparatively large radii of the spheres of effect of the mineral particles on the water molecules provide a basis for assuming that the electric, dipole, and van der Waals forces of interaction taken together can insure the possibility of the exhibition of far-acting effects, and, consequently, the volumetric nature of the surface phenomena in the soil. Accordingly, it is possible to apply the principle of the general theory of capillary effects type II to the ground.[17] By using the mathematical apparatus of this theory, it is possible to obtain an explicit expression for the chemical potential μw of, the attached water having important significance when analyzing the properties and behavior in different processes, not only of the ground moisture but also the ground as a whole.

In the case where the interaction of the water with the soil is accompanied by a reduction or increase in the intital surface of the particles Q as a result of certain processes (for example, aggregation, coagulation, dispersal, and so on),

$$\mu_{aw} = \mu_w + \frac{\omega_\infty Q}{d_w w} - \frac{(\omega_T - \omega_\infty)\,\chi^2\,Q^3}{d_w w^3} \qquad (7)$$

for $Q \neq$ const.

When Q = const,

$$\mu_{aw} = \mu_w - \frac{2\,(\omega_T - \omega_\infty)\,\chi^2\,Q^3}{d_w w^3}, \qquad (8)$$

where μ_w is the chemical potential of the ordinary free water, ω_T and ω_∞ are the specific free surface energy of the mineral particles at the boundary with the air and the total surface energy at the contact boundary of the particles with the water and the water with the air respectively, d_w is the specific weight of the water, w is the water content, and χ is a minimum thickness beginning with which the properties of the water are exhibited as a liquid.

For simplicity of further discussion let us write Equations (7) and (8) in general form

$$\mu_{aw} = \mu_w + \Delta\mu_{aw} \qquad (9)$$

where $\Delta\mu_{aw}$ is the reduction in chemical potential of the bound water by comparison with free water, and let us use (9) to derive the expression describing the reduction in temperature of the beginning of freezing of wet ground. We obtain

$$\Delta T_{aw} = T_0 - T = -\frac{T_0}{L_0}\,\Delta\mu_{aw}, \qquad (10)$$

where T_0 is the normal freezing point of the water in °K and L_0 is the heat of crystallization of the water.

On the basis of (7)-(9), considering in (10) the explicit dependence on moisture and all of the other variables entering into these formulas, by denoting for convenience of notation the coefficients A and B_1, we have

$$\Delta T_{aw} = \frac{A}{w^3} - \frac{B_1}{w}. \qquad (11)$$

Considering the dependence of the freezing point of the soil water on the concentration of dissolved salts

$$\Delta T_c = \frac{B_2}{w}, \qquad (12)$$

finally we obtain

$$\Delta T = \Delta T_{aw} + \Delta T_c = \frac{A}{w^3} + \frac{B}{w}. \qquad (13)$$

In the general case, B in (13) can also be negative.

In Figure 2 we have a comparison of the experimental termperatures of the beginning of freezing and thawing for sand with fractions of 0.25-0.5 mm as a function of moisture and the values calculated by Formula (13). Here, for coefficients A and B, the following values were obtained: for freezing $A = 1.25 \cdot 10^{-6}$, $B = 5.0 \cdot 10^{-2}$; for thawing $A = 3.0 \cdot 10^{-6}$, $B = 2.76 \cdot 10^{-2}$. In addition to the fact that Expression (13) quite accurately describes the experimental nature of the relations for t_1 [temperature at start of freezing] and t_2 [temperature of thawing] as functions of w, let us also consider the following significant facts. From the experiments it follows that the coefficients A and B have different values for freezing and thawing. On the basis of (7) and (8), this means that the processes of freezing and thawing are accompanied by a decrease or increase in the total specific surface of the mineral particles. Inasmuch as the freezing-thawing is connected with the appearance of new interfaces as a result of the occurrence or disappearance of ice, the total specific surface energy of the attached water depending on the number and quality of the interfaces also varies during the freezing and thawing process.

Thus, the variation of the surface of the mineral particles during freezing of thawing fol-

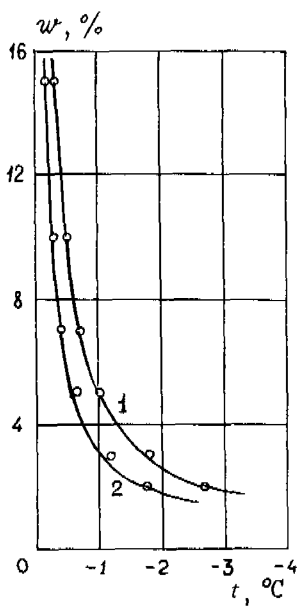

FIGURE 2 Temperatures of the phase transformations of water in sand (0.25-0.5 mm) as a function of water content. 1--temperature of the beginning of freezing; 2--temperature of thawing.

lows not only from the thermodynamics of the ground mass as a whole, but also from the results of a thermodynamic analysis of the properties of the bound (unfrozen) water during transition of it from one state to another.

Finally, an important practical consequence of the above results is the possibility of obtaining quantitative relation of the chemical potential of the unfrozen water to its content by the data from cryoscopic measurements on the basis of (10) and (11):

$$\Delta\mu = -\frac{L\rho}{T_0}\left(\frac{A}{w^3} + \frac{B}{w}\right). \qquad (14)$$

The basis for cryogenous processes is the phase transformations of water. At the present time a number of distinguishing features have been discovered in the course of freezing and thawing of the soils themselves. However, many of these features frequently cannot be explained on the basis of the developed representation of these processes as equilibrium processes without studying the characteristic features of the kinetics of these processes.

The available experimental data show that the process of freezing of the bound water obviously is of a relaxation nature. Thus, in 1949 P. V. Vershinin *et al.*[18] discovered that the equilibrium amount of unfrozen water in the soild is reached only after several days. The latter means that under other identical conditions the crystallization of the moisture in the ground takes place appreciably more slowly than ordinary free water. Later it was found[19] that the final equilibrium value of the melting point of the frozen soil depends both on the time of the melting point of the frozen samples at a given negative temperature and the holding temperature of these samples (see Table 1). The noted peculiarities in the behavior of the water in the ground during phase transformations reduce not only to delaying the beginning of its freezing as a result of the possibility of more prolonged existence in the supercolled (metastable) state in isolated, small and very small pores,[20] but primarily to complicate the further course and development of crystallization after the ice crystals appeared in the frozen soil.

According to the available information, primarily ice-I is formed in the soil. Therefore, the freezing of part of the bound water must be preceded by rearrangement of its structure into an icelike structure.[21] Since time is needed for this rearrangement, in the general case in Expressions (13) and (14), the coefficients A and B will depend on time

$$A = A\left(\frac{\tau}{\tau_r}\right), \quad B = B\left(\frac{\tau}{\tau_r}\right), \qquad (15)$$

where τ_r is the relaxation time of the crystallization process of the film of bound water. The different rates of rearrangement of the structure of the bound water into an icelike structure and the rate of heat release during freezing can cause a different kind of segre-

TABLE 1 Thawing Temperature t_2 of the Soil as a Function of the Freezing Point (the Temperature of the Medium) t_m and the Time τ of Holding the Specimens at the Given t_m

Ground	t_m, °C	τ, h	t_2, °C
Sand, fine,	-3	1	-1.17
w = 5%	-3	21	-0.92
	-10	1	-0.98
	-10	18	-0.85
	-10	45	-0.67
Suglinok,	-4.5	1	-0.67
w = 7%	-10	1.5	-0.55
	-10	22	-0.45

gated ice in a soil, leading to uniform (cement type) or excess ice deposition.

In the general case the functions in Equation (15) mean slow variation with time, not only of the specific free surface energy of the attached water, but also the magnitude of the interacting surface of the water with mineral particles.

Therefore, the processes of dehydration, coagulation, etc., occurring during the phase transformations of the water in the ground, can also be of a relaxation nature.

This nature of freezing of the bound water follows not only from the presented arguments of the thermodynamics of irreversible processes, but also from a theoretical investigation of the mechanical interaction of the phase components of the freezing soil. Let us consider this phenomenon in the example of the simplest uniform model of freezing in the vicinity of an individual soil particle. Let the medium near the soil particle be made up of individual infinitely extended layers of differenct thickness (Figure 3): films of unfrozen water of variable thickness h depending on the ground temperature, a frozen layer of water Δ thick, and pore ice. Let us assume that at the time $\tau = 0$ the temperature of the entire medium has been lowered discontinuously by the amount $\Delta\Theta$. This must lead to freezing of part of the layer of unfrozen water that had the initial thickness h_0. During freezing, the layer of water (layer 2, Figure 3) tried to increase in volume by the amount δ_m--the relative magnitude of the linear expansion on transition of the water into ice. As a result of this, mechanical stresses occur that can be calculated by the ordinary methods of the mechanics of a continuous medium.

For simplicity let us assume that all three layers (1, 2, and 3 in Figure 3) are viscoelastic layers and have identical characteristics: Young's modulus E, Poisson's ratio ν, and relaxation time τ_0. From the effective forces, let us consider the volumetric forces $R(z)$ caused by the presence of the surface of the soil particle and decreasing with distance from this surface by the law

$$R(z) \approx \frac{An}{z^{n+1}}, \tag{16}$$

where A, n are constants characterizing the interaction of the particle surface with the film of unfrozen water.

Considering that the reduction in temperature of the phase transitions is determined by the known Clapeyron-Clausius equation, and the hydrostatic pressure inside the investigated medium is obtained as the solution of the corresponding problem (not presented here), then in order to determine the thickness of the frozen layer Δ it is possible to obtain the following integral equations

$$\left[\xi_m + \frac{E\delta}{(1-2\nu)H}\right]\Delta(\tau) - \frac{E\delta}{(1-2\nu)H}\cdot\frac{1}{\tau_0}\int_0^{\tau} e^{-\frac{\tau-\phi}{\tau_0}}\Delta(\phi)\,d\phi \approx \frac{\Delta\Theta}{\kappa}, \tag{17}$$

where H is half the distance between the ground particles; ξ_m is a constant which depends on A and n and has dimensionality kg/cm^3; $\kappa = T_0(v_w - v_I)/L_0$; and v_w, v_I are the specific phase volumes of water and ice respectively.

Solving this equation, we obtain:

$$\Delta(\tau) = \frac{\Delta t}{\kappa\cdot\xi_m}\left\{1 - \frac{E\delta_m}{E\delta_m + (1-2\nu)\xi_m H}\exp\left[-\frac{(1-2\nu)\xi_m H}{E\delta_m - (1-2\nu)\xi_m H}\cdot\frac{\tau}{\tau_0}\right]\right\} \tag{18}$$

From (18) it follows that with a discontinuous reduction in temperature by the amount Δt, the decrease in thickness of the film of unfrozen water follows an exponential law, that is, the freezing process must be of a relaxation nature. One of its manifestations is the experimentally observed temporary after effect of thermal deformations of frozen ground.[22]

It is interesting to note that the characteristic time τ_p of the investigated relaxation process in accordance with the solution in Equation (18) is described by the expression

$$\tau_p = \tau_0\cdot\frac{E\delta_m + (1-2\nu)\xi_m H}{(1-2\nu)\xi_m H} \tag{19}$$

and, consequently, it depends sharply on the compressibility of the ice (the Poisson's ratio, ν). For ν close to 0.5, the aftereffect time τ_p increases and will always be greater than τ_0--the relaxation time of the ice.

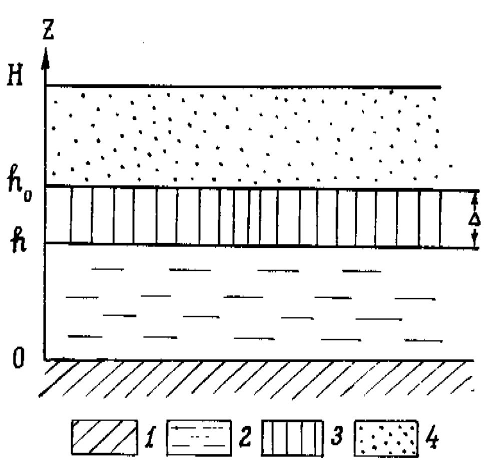

FIGURE 3 Uniform freezing in the vicinity of a ground particle. 1--surface of the ground particle, 2--film of unfrozen water, 3--frozen layer, and 4--pore ice.

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EFFECT OF COAGULATORS ON THE MAGNITUDE OF FROST HEAVE OF FAR EASTERN SUPESSES AND SUGLINOKS

G. D. VOROSHILOV Birobidzhanstroy Trust

In the Far Eastern part of the USSR, most soils are heaving suglinoks and supesses. Their deformation on freezing causes serious damage to unloaded foundations during the construction process, lightly loaded foundations of various buildings and structures, electric power transmission lines and communications lines, artificial structures, roadbeds and railroad beds, and also landing strips.

At the present time many methods of controlling frost heave have been successfully developed and are being widely used. Among them are well-known engineering-ameliorative recommendations aimed at draining the soil in the winter-freezing zone and also in the 3- to 4-m zone below the normal depth of freezing, and the reduction of the moisture in general, especially during freezing fo the ground in the fall-winter period.

However, under the peculiar natural conditions of the Far East (the monsoon climate, the prolonged rains when most of the precipitation comes in the summer-fall period) and under the complex hydrogeological conditions, these measures are in practice unrealizable or have little effect. In addition, frequently the structural measures connected with assigning foundations for frost heave, structural adaptation, and various methods of improving the foundation work are frequently used. In the conditions of the Far East with a severe winter and deep seasonal thawing, these measures are expressed in the depth of the foundations below the freezing depth.

In recent years, physicochemical methods of controlling heave, which, in our opinion, are the most promising, have begun to be used. These methods are connected with the variation of the physicochemical properties of the soil, its capacity for adsorption, and the composition of the exchange cations, with an increase in the amount of unfrozen moisture in the soil and variation of its specific surface and structure.

By varying the magnitude of the specific surface energy of the soil, it is possible to control the process of moisture migration and soil heave. This is achieved by physiochemical processing of the soil by coagulators and materials changing the surface quality of the particles.

In order to check the effect of the coagulators on the magnitude of the frost heave of Far Eastern supesses and suglinoks, especially the materials that are waste from local production, a number of laboratory experiments were performed.

The problem of comprehensive study of the possibility of utilizing sulfite alkalis--waste from local cellulose production--was stated.

The experiments were performed on local suglinoks taken at Amursk, the vicinity of Komsomol'sk-na-Amure, the Semistochnaya Station of the Yevreyskaya autonomous region, and on kaolin as the most heaving soil. The experimental results are presented in Tables 1, 2, and 3.

From the data presented in Tables 1 and 2, it is obvious that the suglinoks heaved 14.38 mm, and after treatment the heave was reduced significantly. Here, the soil without treatment has a reticular structure with clearly expressed interlayers of ice (Figure 1), and after treatment they have a massive texture (Figure 2).

As is known, kaolin is the most typical heaving-soil type. In our experiments it is also characterized by the greatest heave of 22.19 mm. However, after treatment with sulfite alkalis, the kaolin becomes nonheaving. Usually natural kaolin in the frozen state has a reti-

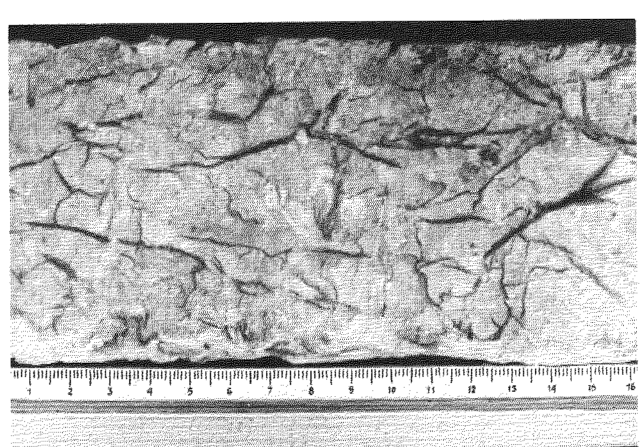

FIGURE 1 Suglinok without treatment.

TABLE 1 Magnitude of the Heave of Clay Soils Treated with Various Reagents

Fractions, % Sand, 1.0-0.1 mm	Silt, 0.05-0.005, mm	Clay, 0.005-0.001, mm	Atterberg Limits of the Soil w_L	w_P	Plasticity Index I_P	Initial Moisture Content, %	Reagent Used for Treatment, % Content	Magnitude of Heave, mm
52	24	24	38.6	19.9	18.7	28.8	Shale tar, 3%	2.41
52	24	24	38.6	19.9	18.7	20	Epoxy resin, 3%	0.08
52	24	24	38.6	19.9	18.7	18	Water re-pelling liquid, 3%	1.07
52	24	24	38.6	19.9	18.7	31	Sulfite alka lis evapo-rated 3%	0.23
52	24	24	38.6	19.9	18.7	30.1	Urea-formal-dehyde resin, 3%	1.31
52	24	24	38.6	19.9	18.7	26.9	—	12.98
49	27	24	37.7	19.0	18.7	20.9	Sulfite al-kalies evap-orated, 3%	1.06
49	27	24	37.7	19.0	18.7	18.8	Sulfite al-kalies evap-orated, 3%	0.29
49	27	24	37.7	19.0	18.7	22.9	—	14.38

TABLE 2 Magnitude of the Heave of Kaolin Treated with Various Reagents

Fractions, % Sand, 1.0-0.1 mm	Silt, 0.05-0.005 mm	Clay, 0.005 0.001 mm	Atterberg Limits of the Soil w_L	w_P	Plasticity Index, I_P	Initial Moisture Content, %	Reagent Used for Treatment, % Content	Magnitude of Heave, mm
1.6	94.4	4.0	69.1	18.8	50.3	38.3	Sulfite al-kalis evap-orated 3%	0.78
1.6	94.4	4.0	69.1	18.8	50.3	40.7	Urea formal-dehyde resin, 3%	3.42
1.6	94.4	4.0	69.1	18.8	50.3	41.2	—	22.19

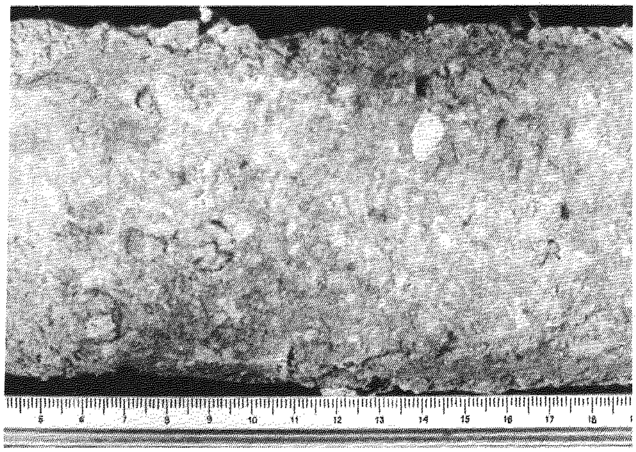

FIGURE 2 Suglinok treated with sulfite alkalis.

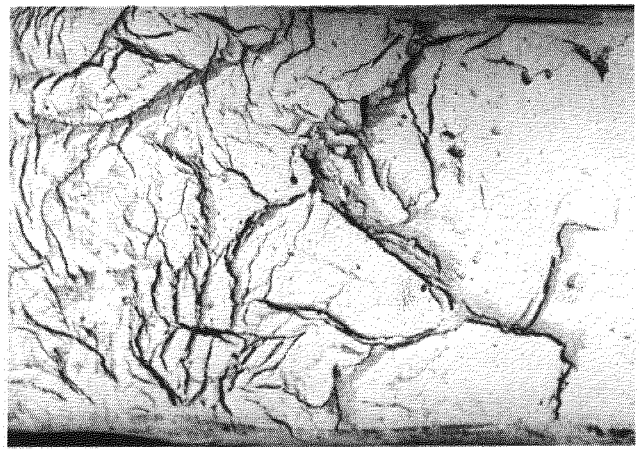

FIGURE 3 Natural kaolin.

cular texture with celarly expressed interlayers of ice (Figure 3). After treatment with sulfite alkalis, it acquires a massive texture, which has been established by numerous experiments. The most effective reagents are epoxy resins and evaporated sulfite alkalis.

Under the conditions of the Far East, the supesses are heaving, as confirmed by the data presented in Table 3. They have a reticular texture with clearly expressed interlayers of ice. After chemical treatment they in practice become nonheaving (the heave is reduced by 15 or more times), and they acquire a massive texture without visible ice inclusions, which has been established by numerous laboratory experiments.

The data from microaggregate analysis presented in Table 4 show that soil treated by reagents contains appreciably fewer small fractions by comparison with the untreated soil.

CONCLUSIONS

1. Soil treated with coagulators essentially show a decrease in the magnitude of the frost heave; sulfite alkalis turn out to be most efficient.
2. The dosage of the sulfite alkali in the amount of 1.5% of the weight of the soil essentially decreases the magnitude of the heave and can be sufficient for practical application.
3. A soil, whatever its moisture content, treated by chemical reagents does not heave.

TABLE 3 Magnitude of Heave of Supes (Mineralogic Composition of Hydromica) Treated with Evaporated Sulfite Alkali

Fractions, %			Plasticity Limits		Plasticity Index, I_P	Initial Moisture, Content, %	Percentage Content	Magnitude of Heave mm
Sand, 1.0-0.1 mm	Silt, 0.5-0.005 mm	Clay, 0.005-0.001 mm	w_L	w_P				
57.5	29.1	13.4	28.8	23.8	5.0	15.1	—	15.01
57.5	29.1	13.4	28.8	23.8	5.0	14.6	3.0	0.03
57.5	29.1	13.4	28.8	23.8	5.0	16.8	3.0	0.12
58.0	29.3	12.7	29.1	24.0	5.1	18.2	—	19.30
58.0	29.3	12.7	29.1	24.0	5.1	21.0	3.0	0.40
58.0	29.3	12.7	29.1	24.0	5.1	18.9	3.0	1.00
77.0	11.0	12.0	23.9	20.0	3.9	18.6	—	16.30
77.0	11.0	12.0	23.9	20.0	3.9	18.7	2.5	1.12
77.0	11.0	12.0	23.9	20.0	3.9	18.3	1.5	0.95
67.1	23.1	9.8	22.6	19.9	3.7	21.0	—	11.67
67.1	23.1	9.8	22.6	19.9	3.7	22.3	1.5	0.67
67.1	23.1	9.8	22.6	19.9	3.7	21.0	1.5	0.61
67.1	23.1	9.8	22.6	19.9	3.7	21.3	1.5	0.23

TABLE 4 Microaggregate analysis of soil treated with sulfite alkalis

Fractions, %						Reagent Content, %
More than 0.25 mm	0.25-0.05 mm	0.05-0.01 mm	0.01-0.005 mm	0.005-0.001 mm	Less than 0.001 mm	
33.05	32.95	19.24	5.48	6.84	2.44	1.5
31.96	36.92	16.86	5.52	6.08	2.66	2.5
25.43	29.48	19.04	10.68	10,88	4.48	Without treatment
27.42	28.02	20.96	7.72	11.40	4.48	Without treatment

STRUCTURAL TRANSFORMATIONS IN FROZEN SOILS ON VARIATION OF THE GROUND TEMPERATURE

I. N. VOTYAKOV Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

The physicomechanical properties of frozen soils are related in the closest way to the problem of structure formation in dispersed systems characterized by sharply developed surfaces and, consequently, a high excess of free surface energy. Here, both the coagulation structures--the steric grids occurring by disordered (random) cohesion of the particles of the dispersed phase through the water film and crystallization-condensation structures formed as a result of direct accretion of the particles or the crystals,--have the defining significance.

It is obvious that the structure of a fine-grained frozen soil with water present in it in the form of solid and liquid phases is a highly complex combination of crystallization-condensation and coagulation structures.

If we begin with the principle of the equilibrium state of the water in frozen soils advanced in 1945 by N. A. Tsytovich, according to which, with any increase or decrease in temperature, partial melting of the ice or crystallization of the unfrozen water takes place, then from the point of view of variation of the structure of frozen soil it is also possible to talk about partial transition of the crystallization-condensation structures to coagulation structures and back.

The dynamics of the structural transformations in frozen soils on variation of their temperature are obviously determined by many factors among which the prevailing ones are the fineness of the soil, its moisture content, and mineralization.

The partial transition of certain structures to others is accompanied in one way or another by variation of all the physicomechanical properties of the frozen soil.

We have established that the basic indicators most clearly characterizing the dynamics of the structural transformation in frozen soils on variation of their temperature are the sonic conductivity and thermal deformations.

The close correlation between the structural transformations in frozen soils, their sound conductivity, and thermal deformations is caused by the fact that the propagation rates of the longitudinal ultrasonic waves increase to a significant degree with an increase in the elasticity of the body, that is, with transition of the coagulation structure to crystallization-condensation on reduction of the ground temperature, and, vice versa, the transformations are accompanied by very large volumetric changes.

The maximum propagation rate of longitudinal ultrasonic waves characterizes frozen soil with a moisture content corresponding to saturation in the range of temperature from -30° to -10°, and the minimum corresponds to dry air, especially in the free-flowing state. In the first case the velocities reach 4,000-4,300 m/s, and in the second case they drop to 350-400 m/s. During the process and after complete thawing of the frozen saturated soil, that is, on transition of the crystallization-condensation structures to coagulation structures, the propagation rates of the ultrasounds decrease by 4-5 times. Here, the intensity of reduction of the velocity as the permafrost temperature increases is determined by the intensity of the increase in the content of unfrozen water. Therefore, in clay soils and also in artifi-

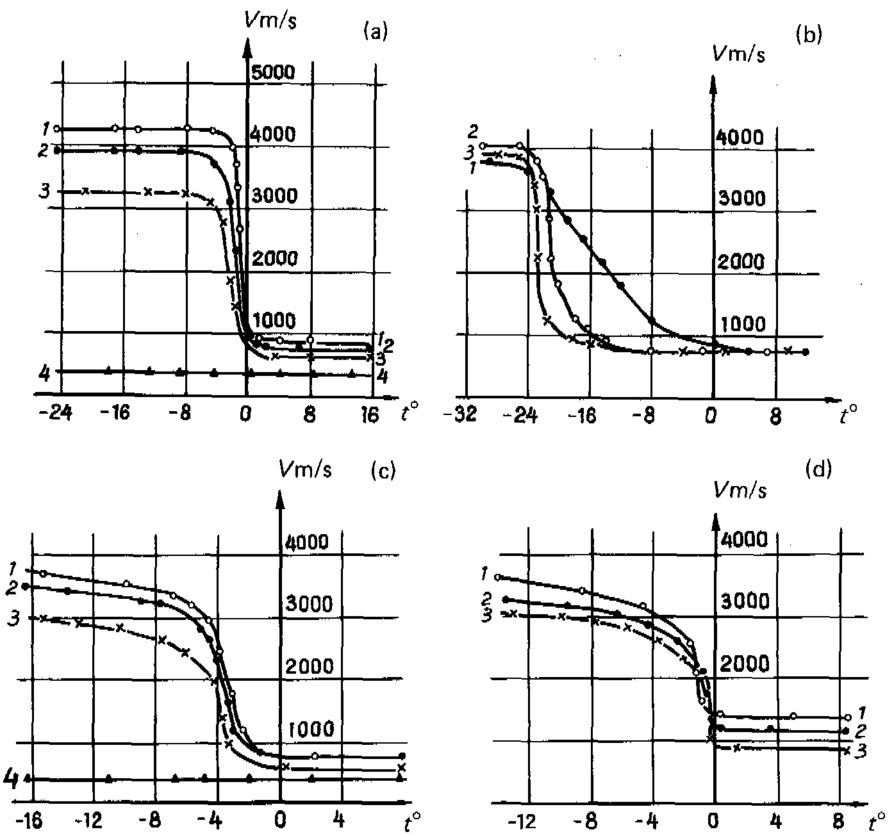

FIGURE 1 Curves showing the variation of the propagation rate of ultrasonic longitudinal waves in frozen and thawing soils as a function of their fineness, moisture content, mineralization, and temperature. (a) Fine-grained sand: 1--*w* = 21 percent, 2--*w* = 16 percent, 3--*w* = 10 percent, 4--*w* = 1 percent. (b) Fine-grained sand artificially salinized with NaCl at *w* = 24 percent: 1--mineralization 1 percent, 2--3 percent, 3--5 percent. (c) Silty supes: 1--*w* = 30 percent, 2--*w* = 25 percent, 3--*w* = 15 percent, 4--*w* = 1.5 percent. (d) Silty suglinok: 1--*w* = 28 percent, 2--*w* = 34 percent, 3--*w* = 40 percent.

cially salinized sand the propagation rate of the ultrasound decreases noticeably with an increase in temperature beginning with -20° to -10°, whereas in the unslainized sand the rate remains constant (4,300 m/s) with an increase in temperature to -2° to -1°.

In Figure 1 we show the experimental results with respect to determining the propagation rates of longitudinal ultrasonic waves in frozen and thawing soils as a function of the degree of their fineness, moisture content, mineralization, and temperature variation.

From the curves characterizing the variation of the transmission rate of ultrasonic waves through frozen and thawing soils, it is obvious that the maximum velocity V = 4,300 m/s characterizes the unslainized, completely saturated, fine-grained sand in the range of temperature variations from -30° to -5°, when all of the moisture in them is in the frozen state and there are no air spaces. As the moisture is reduced and the air porosity is increased, the velocity V decreases and reaches its minimum value (about 400 m/s) in the dry-air state. It remains constant both in the negative and positive temperature ranges.

The very sharp decrease in the velocity V at temperatures close to 0° is a result of the intense process of thawing in the ice and the transition of the crystallization-condensation structures into coagulation structures. The velocity stabilization indicates completion of these processes.

Inasmuch as the structural transformations in frozen and thawing soils are directly connected with the phase transitions of the water, they can occur in any ranges of variation of the negative temperatures, except in this case the moisture in the soil will be contained in the form of two phases in dynamic equilibrium with each other.

A clear illustration of the stated arguments is the experiments performed by us with artificially salinized fine-grained sand, supes, and suglinok, which are represented in the form of the curves for the variation of the ultrasonic velocity with an increase in temperature in Figure 1b, c, d. By analyzing the curves, it is possible to note that in the salinized sand a very sharp drop in velocity V is observed beginning with a temperature of -22°, and in the supes-suglinok there is a quite smooth velocity drop beginning with -20°, which indicates the presence of unfrozen water in the soils at the same temperature.

As the temperature increases, partial melting of the ice takes place along with transition of the crystallization-condensation structural bonds to coagulation bonds.

Thus, the nature of variation of the propagation rate of the longitudinal sound waves in the frozen ground on variation of their temperature and especially on thawing can serve as an indicator of the structural transformations taking place as a result of phase transitions of the water in the soil.

The frozen clayey soil is characterized by very large values of the temperature deformations that develop for a prolonged period of time and are defined by the fineness of the soil, its moisture content, and the temperature drop. The thermal coefficients of expansion (compression) of the frozen clayer soil increase with an increase in temperature and degree of fineness of the soil, and they have maximum values with a moisture content corresponding to their plastic limit or maximum molecular moisture capacity in the thawed state. For example, the coefficient of thermal expansion α with 1-day standing of the samples in the region of temperature drops from -1.6° to -0.4° ($t°_{av}$ = -1) varies: for varved clay, from $345 \cdot 10^{-6}$ with a moisture content of 40 percent to $205 \cdot 10^{-6}$ with a moisture content of 65 percent; for surface suglinok, from $290 \cdot 10^{-6}$ at a moisture content of 30 percent to $210 \cdot 10^{-6}$ at a moisture content of 40 percent; and for supes, from $110 \cdot 10^{-6}$ with a moisture content of 20 percent to $70 \cdot 10^{-6}$ with a moisture content of 35 percent. With a reduction in the mean temperature to -5° to -6° the indicated coefficients decrease by 3-5 times with corresponding moisture contents.

Graphical representation of the laws of variation of α_{av} for the indicated soil as a function of their moisture content and average temperature, $t°_p$ give the curves depicted in Figure 2a.

The very large values of the coefficient α_c for frozen clayey soils, especially in the region of comparatively high temepratures, are explained by us only by the structural transformations connected with the coagulation-colloidal processes that are accompanied by very large volumetric variations of the mineral grains of the soil with an increase or reduction in its temperature. The corresponding calculations show that the coefficients of linear expansion (compression) of the hydrated grain structure of frozen clayey soil α is an entire order higher than the experimentally obtained values of α. This especially pertains to the region of the temperature drop close to zero, where the most significant and intense phase transformations of the water and, consequently, the structural transformations take place. Here it is necessary to note that the coefficient α_s depends only on the soil temperature and does not depend at all on its moisture content, inasmuch as the coagulation-colloidal processes and the corresponding thermal transformations are caused indirectly by phase transitions of the attached water in the soil.

The graphical representation of the laws of variation of α_c of the frozen clayey soil as their temperature is reduced gives the curves depicted in Figure 2b, which are entirely analogous to the curves for the variation of the percentage content of unfrozen water in this soil with a reduction in temperature.

During the course of further experimental studies, it was established that after freezing of the samples at one and the same temperature and transport of them to the conditions with another temperature regime, they continued to be deformed (expanded or compressed) during a quite prolonged time reaching 2 or 3 months.

The curves for the development of the temperature deformations of the frozen clay with time depending on the moisture content and the range of the temperature drop are presented in Figure 2c.

It was also established that after a 1-day freezing of the samples they continued to be deformed for a prolonged time period at the same temperature. The course of the relative deformations of the suglinok with time at constant

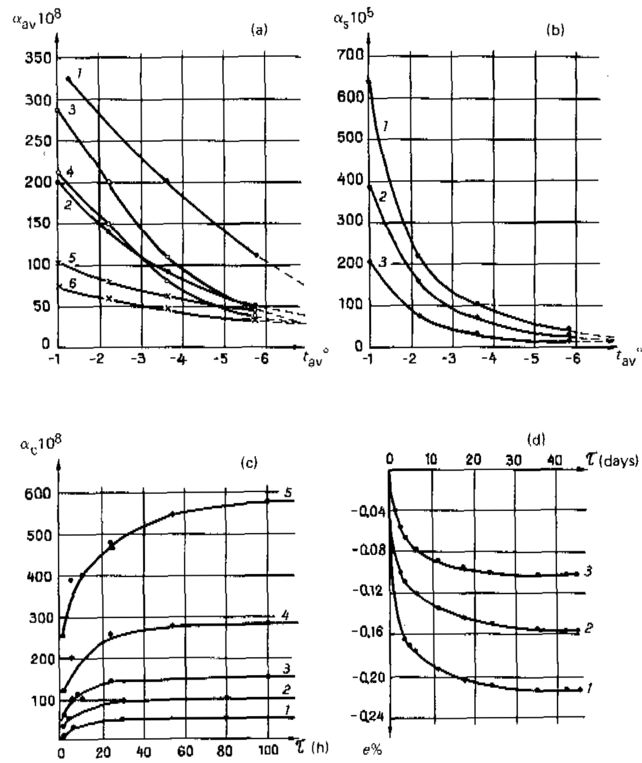

FIGURE 2 The coefficients of thermal expansion of frozen soils as a function of their fineness, moisture content, and temperature, and also the curves for the development of thermal deformations of the soil with time. (a) Variation of the coefficient of thermal expansion of the frozen soil as a function of moisture content and temperature: 1--clay at w = 40 percent, 2--clay at w = 65 percent, 3--suglinok at 2 = 30 percent, 4--suglinok at w = 40 percent, 5--supes at w = 20 percent, 6--supes at w = 35 percent. (b) Variation of the coefficients of thermal expansion α_s of the hydrated mineral grains of the frozen soils as a function of their mean temperature: 1--clay, 2--suglinok, 3--supes. (c) Variation of the coefficient of thermal expansion of the frozen clay, α_c, with time as a function of its moisture content and the temperature drop: 1--w = 55 percent (t_1 = -17°, t_2 = -3°); 2--w = 45 percent (t_1 = -17°, t_2 = -3°); 3--w = 45 percent (t_1 = -7.5°, t_2 = -3°); 4--w = 35 percent (t_1 = -7.5°, t_2 = -3°); 5--w = 25 percent (t_1 = -7.5 percent, t_2 = 3 percent). (d) Development of the temperature deformations, e percent, of frozen suglinok with time after its one-day freezing for t = -3°: 1--w = 24 percent; 2--w = 30 percent; 3--w = 40 percent.

temperature of -3° as a function of the moisture content is shown in Figure 2d.

The voluntary development of the deformations of the frozen soils with time, both at the freezing point and after transfer to other temperature conditions, inertia was caused by internal structural transformations occurring in the frozen soil after freezing or variation of the temperature regime of the environment. These processes of deformation of the frozen soil developing with time will be called thermorheologic.

Thus, the frozen and freezing clayey soil is an exceptionally mobile and simultaneously inertial system with a very complex mechanism of interaction among its components.

The thermorheologic properties of the frozen soil obviously arise from the inertia of the internal structures of the transformations taking place in the clayey soil with temperature variation.

STABILITY OF EARTH STRUCTURES AND EMBANKMENTS

A. I. DEMENT'YEV Production and Scientific Research Institute for Engineering Research in Construction

An earth structure quite frequently is subjected to deformations under the effect of various natural factors. This occurs especially frequently and most intensely in permafrost regions where the deformations are even destructive.

The deformation of earth structure, including embankments and excavations, can be considered to be the result of the occurrence and the effect of natural factors. The deformations can be subdivided into the following basic groups: (a) deformations caused by cyclically-recurring seasonal freezing of the ground; (b) deformations connected with irreversible changes in state of the frozen soil; and (c) deformations caused by climatic, geologic, and hydrogeologic factors.

This classification scheme reflects most completely the genetic structure of the deformations that occur and the nature of the effect of natural factors on earth structures and embankments.

The cyclically recurring seasonal freezing processes cause heave of the ground during winter freezing and subsidence during summer thawing. These processes, which regularly follow each other with the time of year, do not have a significant effect on the state of the structure, inasmuch as the ground heave occurring during freezing is completely removed during the summer thaw.

The most significant deformations during frost heave are observed with roadbeds, causing temporary interruption of the normal operation of the roads.

Under the effect of the winter heave of the soil, artificial structures are subjected to still larger deformations expressed in distortion of bridges, the uprights of auxiliary structures, enclosures, culverts, and so on. These processes are accompanied, in contrast to the roadbed, by residual distortion that increases from year to year.

The deformations of the earth structure of the second group are caused by irreversible changes in the state of the permafrost located in its foundations. It is necessary to note that the erection of the earth structure itself introduces very large changes in the natural conditions, which lead to its subsequent deformations. Excavations destroy the protective plant cover and the upper ground layer, denuding the permafrost under them; fills change the conditions of the heat exchange between the ground and the atmosphere, causing cavein. All of this causes thawing of the permafrost under the roadbed, which is the natural subgrade, saturation of the permafrost, and subsidence and reduction in bearing capacity. This leads to deformations of the structure and loss of stability.

The greatest changes take place with high ice content of the frozen ground and especially in the presence of underground ice. In addition to the general settlement, thawing can lead to large nonuniform local subsidences that reach the greatest magnitude in peat and, especially, the heave mounds. The subsidence in these cases sometimes reaches values of 2 and more meters, and the roadbed is characterized by frequent alternation of rises and depressions (Figure 1).

The height [of fill] has a very great effect on the amount of thawing of the ground under a fill. In the case of low fills, thawing of the permafrost under them takes place very intensely; the fills with great height usually protect the permafrost from thawing, at the same time preventing it from settlement, which to a significant degree insures conservation of the fill itself. The deformations of the earth structure caused by thawing of the permafrost usually take place most intensely during the first 2 or 3 yr after building the structure; thereafter, they gradually diminish and 5 to 6 yr later basically cease.

FIGURE 1 Settlement of a roadbed caused by thawing of the permafrost in the subgrade (photograph by the author).

Quite frequently the deformations of an earth structure occur not as a result of the manifestation of the above-described freezing processes, but as a result of the continuously occurring geological and hydrogeological processes and also the effects of surface water and atmospheric precipitation. The latter cause the appearance of deformations belonging to the third group and are also one of the most important causes of disturbance of the stability of the earth structure.

The earth structure always disturbs the natural surface runoff, which leads to accumulation of water at the structure somehow being converted into a type of dam, which becomes the cause of deformation of the bearing ground and erosion of the fills. The stability of the fills in these cases is dependent on the location and size of drainage structures. With irregular arrangement of them, without the required consideration of the local relief and the conditions of the surface runoff, frequently the washouts of the embankments are formed at places where such a structure is needed. The erosion of the structure by surface water and currents frequently takes place on flooded or periodically flooded sections. Water erosion sometimes reaches very large dimensions, and in individual cases the destruction of the structure by the surface water assumes great scales (Figure 2).

The cause for deformation of an earth structure can also be another form of erosion caused by falling rain. The rainwater accumulating from the top of the fill and also the water formed from thawing snow runs off along its embankments (or along the embankments of the excavation) and destroys them, creating numerous washouts, especially on the unprotected banks without cover. In the case of high fills, the washes also sometimes reach large sizes--to several meters. However, such deformation of the banks is primarily observed on their surface and does not destroy a basic portion of the structure. Their danger for the stability of the earth structure is always appreciably less by comparison with the erosion by accumulated surface water at the fill, such as, for example, in the case presented in Figure 2.

Deformation of an earth structure and its banks can also be caused by wind erosion. The most significant destruction takes place in cases where the body of the structure is made up of fine, easily blown sand. For deep excavations in such ground, along with the banks blowing away, there is silting up of the culverts and drains, which sharply disturbs the normal removal of surface water. The eolian processes are most developed in the sections running through dry and open terrain and also on the higher fills in open river valleys. In the forested and swampy sections with low fills, the eolian processes are expressed more weakly and have almost no practical significance. As a rule, wind erosion is of a surface nature and presents appreciably less danger for stability of the earth structure by comparison with all types of water erosion.

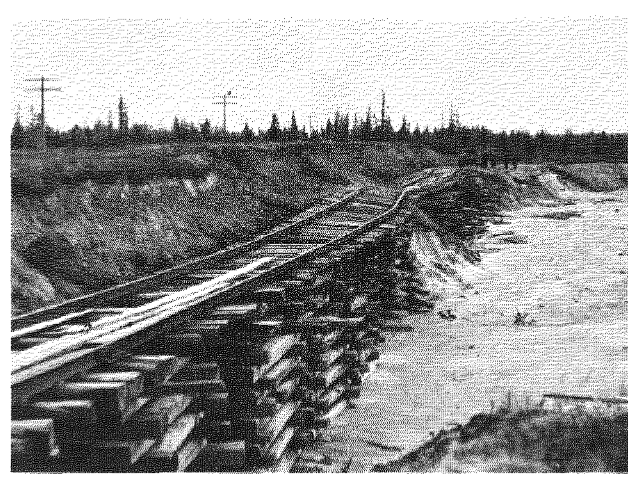

FIGURE 2 Erosion of a roadbed by surface water (photograph by the author).

SOME HEAVE LAWS OF FREEZING SOILS

V. O. ORLOV Production and Scientific Research Institute for Engineering Research in Construction

Among the set of numerous interrelated factors affecting frost heave, grain size, moisture content, and the degree of cooling of the soil during the freezing period play an important role. From the point of view of the physics of surface phenomena promoting recognition of soil heave theory, the interaction of these factors forms the force field of the freezing medium, predetermining in it the processes of moisture transport and ice accumulation on which the intensity of frost deformation of the soil depends.

The effect of the gain size on the heave is most sharply exhibited when there are fractions finer than 0.1-0.07 mm in the ground having a sufficiently large free energy reserve concentrated in the double (according to Tyutyunov) surface layer at the interface between the mineral particle and the water. The additions of these fractions to a coarse soil transforms the latter into frost-sensitive (heave) systems. Therefore, it is important to have a concept of the quantitative relation of the component groups of the individual fractions characterizing their transition to the frost-susceptible systems along with their porosity.

The problem of establishment of the relation between the physicomechanical properties of the ground and heave has been investigated by many researchers, especially foreign researchers. However, the information obtained about the boundary of the frost-susceptible soil with respect to granulometric composition does not give the same answer to this question. The opinion of the majority of researchers reduces to the fact that the transition of the ground to the category of being frost susceptible is defined by the content of silt-clay particles less than 0.02 mm in diameter in the amount of more than 1 to 3 percent.

The studies by the author[1] demonstrate that the capacity of the soil for adsorbing film moisture required to develop the heave process can be expressed in terms of the integral contact surface of the mineral particles and water (S_0) proportional to the free energy of the ground system and capillary pressure (P_k) relative to the porosity of the ground. In this case the frost susceptibility of any soil can be determined by its grain-size criterion (D), the basis for which is an energy potential of the ground system expressed in terms of the above-indicated values:

$$D = \kappa P_K S_0 = \frac{24K\sigma}{\varepsilon d_0^2} = \frac{1.8K}{\varepsilon d_0^2}, \qquad (1)$$

where σ is the surface tension of the water, assumed equal to 0.077 g/cm at a temperature of 0°C; ε is the void ratio; K is a proportionality factor equal to 10^{-4} g/cm^3; and d_0 is the mean diameter of the mineral particles of the soil (cm) comprising the series of functions

$$d_i = \left(\frac{P_1}{d_1} + \frac{P_2}{d_2} + \dots \frac{P_i}{d_i}\right)^{-1}, \qquad (2)$$

where d_1, d_2, ... d_i are the diameters of the particles of individual fractions, cm; P_1, P_2, ... P_i are the percentage content of the fractions expressed in fractions of a unit.

In practical calculations the particle diameters of the individual fractions to 0.001 mm are taken with respect to their minimal dimensions increased by 30 percent. For particles less than 0.001 mm, the calculated diameter is taken equal to 0.0005 mm. The percentage content of clay fractions is determined in the following intervals: 0.005-0.002, 0.002-0.001, and less than 0.001 mm.

The estimate of the grain-size criterion made on the basis of the research results of J. F. Haley and C. W. Kaplar[2] demonstrated that the transition of the soil into frost-susceptible systems corresponds in the first approximation to $D \cong 1$. When $D < 1$, the soil is non-frost-susceptible and, when $D > 1$, it is frost-susceptible. The variation of the values of D within the limits from 1 to 5 corresponds to the degree of frost sensitivity of weakly heaving ground for which the heave modulus (the magnitude) of the heave in cm referred to a freezing ground layer of 1 m) is no more than 2.0-3.5 cm/m.

The basic condition of heave of any soil is the excess of the volumetric increment of the frozen and unfrozen water accumulated in the given mass of frozen ground, above the volume of pores free of water (n) of the same mass. In the case of absence of migration moisture accumulation in the frozen soil, a unit volume of soil during crystallization in it of an amount of water of the initial content (q) increases by the amount $\Delta V_1 = 0.09\, q\, (1 - n)$, where 0.09 is the coefficient of volumetric expansion of the water turning into ice.

The increment of migration moisture in the freezing ground in the same volume causes an additional increase in the soil volume during its crystallization by the amount $\Delta V_2 = (1 + 0.09)q$. If, as a result of the initial moisture content, the heave modulus does not exceed 2.0-3.5 cm/m, with migration moisture accumulation it is capable of reaching 15-20cm/m and more.

Therefore, when investigating the frost susceptibility of the soil with respect to the moisture regime, it is necessary to distinguish two moisture indexes, independent of each other, determining the beginning of ground heave, starting with its physical state and the capacity of the liquid phase for redistribution (migration).

The first index--the heave limit moisture content w_{nn}--characterizes (according to N. A. Tsytovich) the conversion of the soil into the state for which all of the pores are filled with ice and unfrozen water, but ground heave is absent. The value of w_{nn} defines the first initial condition of heave expressed by the inequality

$$w > w_{nn} \tag{3}$$

for

$$w_{nn} = 0.91 \frac{\gamma - \gamma_d}{\gamma \gamma_d} + 0.09 w_u(t), \tag{4}$$

where w is the prewinter moisture content of the thawed ground (as a fraction); 0.91 is the unit weight of the ice, g/cm^3; γ is the unit weight of the soil, g/cm^3; γ_d is the unit weight of the dry soil, g/cm^3; $w_u(t)$ is the unfrozen water content in the freezing ground at a mean temperature t.

The second index--the critical moisture content w_{cr}--characterizes the largest content of bound water for which the displacement of the liquid phase to the freezing boundary is curtailed. The inequality

$$w > w_{cr} \tag{5}$$

expresses the second condition of heaving of the clay soil--as a result of migration of the moisture in it. In the first approximation w_{cr} can be equated to the moisture content corresponding to the plastic limit.

The phase transitions of the free water in the initial crystallization stage leading to segregated ice formation and the disturbance of the equilibrium state of the liquid phase in the adjacent layers of thawed ground predetermine the conditions of redistribution of the film water inside the freezing zone, that is, the boundary zone of the frozen ground.

The heave process occurring in this zone is characterized by the temperature interval, the upper limit of which determines the curtailment of the redistribution of the film moisture capable of causing an increase in volume of the frozen ground. The temperature at which the heave is curtailed is caused by the physicochemical properties of the soil and for supes it is -1.5° to -2°; for suglinok, it is -2° to -4°; for clay, it is -4° and lower.

The continuity of the bonding of the film water during its movement in the freezing zone is maintained as a result of the presence of the temperature gradient causing the formation of a concentration gradient of this moisture. The increase in the temperature gradient to a critical value causes an increase in the migration rate of the water as a result of an increase in its concentration gradient. The effect of the temperature gradient on the heave is illustrated by the results of laboratory tests with respect to freezing the silty suglinok with an initial moisture content of 28-30 percent presented in Table 1.

With a decrease in temperature, thermal deformations of the grain structure occur, which are accompanied by a decrease in volume of the ground. The accumulation of the free surface energy of the mineral particles in the presence of a difference in the hydrothermal and, in individual cases, chemical potentials in the freezing medium defines work expended on putting the film mechanism of water migration in the liquid phase into effect in the freezing zone. Here, the amount of film moisture increases as a function of the variation of the adsorption capacity of the mineral grains in the case of an increment of free energy as a result of the thermal deformation of the mineral particles.

TABLE 1 Effect of the Magnitude of the Temperature Gradient on the Migration Moisture Accumulation and Ground Heave

No. of Experiment	No. of Freezing Cycle	Time of Freezing of the Ground, h	Mean Temperature of the External Environment, °C	Mean Temperature Gradient of the Ground during the Freezing Period, deg/cm	Amount of Water Entering the Sample of Freezing Soil, mg/cm^2	Magnitude of the Frost Heave, mm
4	1	125	-6.0	0.67	397	3.20
5	2	188	-9.1	0.91	1770	16.96
6	1	81	-7.0	0.86	618	8.81
7a	2	52	-6.5	0.76	458	5.88
7σ	1	52	-6.5	0.80	0	1.30

Another circumstance characteristic of ground heave is the process of coagulation of the mineral and organic particles voluntarily taking place in the freezing zone, the intensity of which also increases as the free-energy increment. On the contrary, by promoting differentiation of the soil components, the process of coagulation of the ground particles along with the formation of microaggregates causes a reduction in the specific surface and energy potential of the particles, which leads to their dehydration. The law of variation of the free surface energy of the particles in the process of freezing of the ground has a complex nonlinear nature that depends on the dynamic interaction of the grain size and aggregation processes of the ground particles.

The crystallization of the adsorbed water formed during the process of dehydration of the particles increases the amount of film water and leads to excess ice formation, which causes stress relief of the mineral part of the ground and its heave.

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ROLE OF PERIGLACIAL CLIMATE IN DEEP DISTURBANCES OF THE TERTIARY DEPOSITS OF THE BOHEMIAN MASSIF

YA. RYBARZH Geological Institute of the Academy of Sciences of the Czechoslovakian SSR, Prague

The initial occurrence of the deposits was studied in the Tertiary basins of the Bohemian forests in construction foundation pits and the walls of open-pit mines. The clearly expressed unevennesses of the foot of the river terraces were described in the base of which the clay or aleurolite deposits or the brown coal layer emerged to the surface. The sharp disturbances are noted primarily at the interface of the two geotechnically different media, for example, at the clay and sand contacts.

The occurrence and development of deformations of the Pleistocene age were for the most part of a prolonged and periodic nature. The Bohemian massif in the Pleistocene was in the zone of continental glaciation, that is, it was subjected to the effects of the periglacial climate. This led to deep disturbance of the deposits in the permafrost zone. The freezing clay aleurolite deposits were often disturbed to a depth of several tens of meters, being covered by a dense network of smooth tracks which probably caused the occurrence of so-called jointed clays.* Under favorable conditions, deep-folded structures also arose.

The especially intense disturbance of the initial structure of the deposits and variation of its properties occurred in the seasonal freezing and thawing layer (in the active layer). The climatic peculiarities of the periglacial period were also felt in the increased activity of the slope processes.

As an example, let us consider the deformations that were studied in the vicinity of the Nekhranitse Dam in the northwestern part of Czechoslovakia. When exploring the deposits of the sand-gravel for the construction of the earth dam, definite unevennesses were detected in the base of the terrace gravel. Small unevennesses were noted also when exploring directly in the dam profile. It was proposed that they are due to erosion. Only after sinking more than 800 boreholes and wells and systematic documentation of the walls of open-pit mines were the processes of the occurrence of unevennesses better understood.

The foot of the river terraces of the Ogrzhe River is represented by Miocene silty argillites, frequently aleurites* or a layer of brown coal. The silty argillites have the nature of fissured clays, forming fragments of irregular shape. Along with the prevailing clay mineral--montmorillonite--there were also chlorite and

*The dense network of fine cracks penetrates the Tertiary clays to a depth of 30-50 m similarly to the type of so-called frost fissures from the permafrost zone.[1,2]

*Silts.

illite. The natural moisture content of the unweathered argillites varies within the limits of 36-42 percent, the liquid limit is 75-95 percent, the plastic limit is 35-50 percent, and the plasticity index is 30-40 percent. The sandy gravel of the terraces with a total thickness of 8 m is covered with suglinoks and sometimes loess.

The base of the river terraces is very uneven. The unevennesses are concentrated in several zones that are represented in the form of hollow convex arches with a width of 40-80 m. These zones are formed by a system of brachyanticlines with axes in the direction of the strip. Depressions are noted in the foot of the latter. The nature of the deformation determines the profile directed perpendicular to the brachyanticlinal zone (Figure 1). The southern wings of the folds are steeper than the northern wings, which are sometimes vertical, and they are tilted in the upper part. The upper part of the folds is raised perpendicular to the surface. The sandy clastic materials were advanced together with the deforming argillites. The amplitude of the folds diminishes with depth, and only at a depth of about 10-12 m is deformation observed.

The disturbed zone is noted, as a rule, at the boundary of two terraces with different height of the base at the points where the argillites are covered by terrace gravels of relatively little thickness.

Along with such folded structures, which we have denoted as the first degree deformations, near Nekhranits cryoturbation phenomena of smaller dimensions were observed connected with the surface layer of the deposit subjected in the Pleistocene to processes of seasonal thawing and freezing (second-degree deformation).

The first degree unevennesses at Nekhranits are difficult to explain. Let us denote as initial unevennesses all those which occurred at the surface of the pre-Quaternary ground as a result of the deep river erosion. The flow of water with rapid advancement of the meanders easily dug a pool in the erosion stage at the banks and left depressions in the core of the meanders. The erosion and development of unevenness were completed during the period of accumulation of clastic materials. Later, in the uneven places after deposition of the cover, favorable conditions were created for the development of secondary deformations, which, with respect to scale and size, forced the primary disturbances to the background. The secondary unevennesses thus occurred at the points where the freezing clay-aleurolite were covered by the thinnest layer of clastic accumulation. The development of the disturbances could take place during the deposition of the terrace gravel or only after accumulation of it. However, the syngenetic origin is difficult to prove in the majority of cases.

The development of the first-degree secondary unevennesses, depending on the local conditions, can be explained by an entire series of factors, the most important of which are the following:

1. When the defined stressed state is exceeded, the plastically sensitive deposit was forced out of the zone loaded under the heavy weight of the roof into the more lightly loaded zone.[3]

2. After the onset of cold periods of the Pleistocene, initially a permafrost layer is formed. On freezing of sufficiently wet clay-aleurolite, the growth of the ice and the uplift of its surface takes place, which is caused by an increased migration of water to the cooled places on the surface of the clayey ground.

3. During the cold periods of the Pleistocene, the plastic deformations of the permafrost with a large ice content developed more rapidly.

4. With absence of the cold periods of the Pleistocene during thawing of the frozen beds, the strength of the argillaceous rock decreased. Even with a small difference in weight of the overburden, their plastic extrusion into the unloading zones can occur.

5. The volume of rock after unloading can also increase on a small scale.

6. At the point of decrease in weight of the overburden, deformations can occur with relief from horizontal stress (for example, the residual horizontal stress in the already consolidated clay, the stresses of seismic and tectonic origin).[4]

An analysis of the development of the basic type of deformations near Nekhranits is illustrated in Figure 2. As the basic factor it is necessary to isolate the effect of the different weight of the overburden and the effect of increased migration of the water in the direction of the temperature drop in freezing clay-aleurolite deposit (a). The redistribution of the mass occurs (b). The upper part of the protruding fold within the limits of the active layer is characterized by an explicit uplift of the argillites to the surface (c). The denuda-

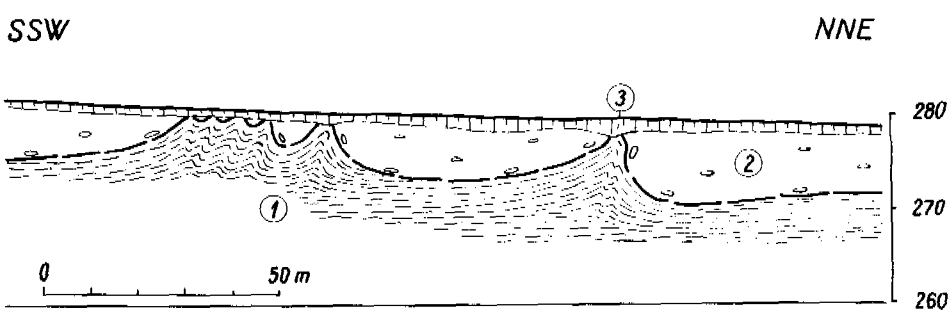

FIGURE 1 Geological profile of the folded zone near Nekhranits. 1--silty argillites, 2--sandy gravels, and 3--eolian-alluvial deposits.

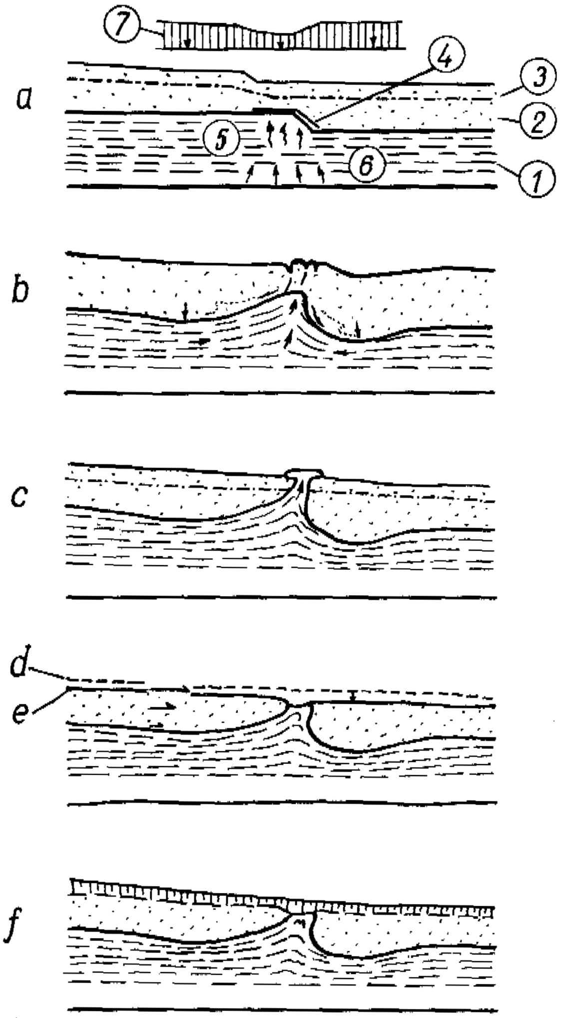

FIGURE 2 Stages of development (top to bottom) of the basic type of deformations detected near Nekhranits. 1--silty argillites, 2--sandy gravel, 3--depth of seasonal thawing and freezing, 4--exposed section of the argillite surface, 5--increased water migration, 6--direction of the temperature drop, and 7--nonuniform vertical loading that has an effect on the argillite surface.

tion leads to lowering of the surface of the territory (d), and retreat of the cold period as a result of thawing of the frozen rock, to settlement of the surface (e). The settlement is nonuniform and develops most intensely at the points where the argillites more subject to pressure protrude to the surface. The asymmetric nature of the fold directed along the slop is explained by the development of solifluction. The surface unevennesses are simultaneously filled with loess (f).

The Pleistocene distrubances of the rock in the Tertiary deposits complicate extraction of the raw material and engineering construction. The unevennesses in the base of the deposits complicated the exact calculation of the reserves of raw material. Part of the deposit even with careful selective extraction was not completely used. This situation was observed in the open-pit mines near Frantishkovyy Lazni and also in the Ogrzhe River basin.

The evennesses of the bedding of the sandy or clastic rock can have a detrimental effect on the possibility of working the lower-lying rock if it develops as deposits. This is especially felt in the surface of the brown-coal layer distrubed by deformations. The outside upper part of the layer is not used and remains, as, for example, in the open-pit brown-coal mine of Silvestr in the vicinity of Sokolov or the Nastup Open-pit Mine near Kadan'.

Freezing and thawing impair the physico-mechanical properties of the deposit. In the clay and aleurolite deposits, heave and plastic creep deformations develop, the shear strength is reduced, and infiltration increases. After damping of the deformations, the initial properties of the rock frequently are restored, but not fully. Obtaining the rock at great depths, in addition, also has other unfavorable consequences. The rock subjected to freezing and thawing is weathered more intensely.

Construction on the disturbed beds requires increased care,[5] and this implies, as a rule, an increase in expenditures on construction. The difficulties on a large scale can occur when building dams on a sharply disturbed clay base.

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NATURE OF DEFORMATION AND VARIATION IN STRENGTH OF FROZEN COARSE CLASTICS AS A FUNCTION OF THEIR COMPOSITION AND STRUCTURE

V. N. TAYBASHEV All Union Scientific Research Institute of Gold and Rare Metals

Tests for uniaxial compression, shear, and tension indicate that frozen coarse clastics of massive texture in the state of complete saturation have rheologic properties and develop creep deformations, lowering the resistance to long-term loads.

When explaining the peculiarities of deformation and destruction of such soil, it is expedient to consider two systems as a set: large mineral inclusions and small-grained filler (sand, silt, clay).

In this paper a study was made of the laws of variation of the strength and deformation properties of frozen coarse clastics depending on the ratio (by weight) of the coarse clastics and the filler (supes) and the granulometric composition of the filler (sand, supes, suglinok). For this purpose, at a temperature of -3° and -5°, seven soil mixtures were tested. In four of them the content of supes filler was as follows: 5 percent II mixture, 20 percent III mixture, 30 percent IV mixture, and 48 percent VI mixture; in three of them the filler is represented by sand--mixture V, supes; mixture VI and suglinok; and mixture VII. In addition, gravel and shingle were tested--mixture I cemented with pure ice (Table 1).

The strength of the frozen coarse clastics as a function of the ratio of the coarse clastic and fine-grained fractions is exhibited in the following way.

The uniaxial compressive and shear strengths during rapid fracture increases with a decrease in filler content and a corresponding increase in ice content of the fine-grained fraction; in the case of prolonged fracture, on the contrary, the ground with a high filler content has greater strength. The soil with the densest composition of the coarse clastics and with the largest ice content per unit volume of filler has the greatest strength during rapid fracture and the least during prolonged fracture (Figure 1a).

The nature of fracture of the samples during rupture indicates that the strongest bonds in coarse clastic soils are the adfreeze forces at the contact of the surface of large grains and the ice film surrounding them. Therefore, frozen coarse clastics similar to mixture III are characterized by optimal composition for which the strength of the filler is the greatest, and the adfreeze forces have subordinate significance in the general resistance to fracture; the soil is also characterized by the greatest rupture strength (Figure 1b). The different nature of variation of strength as a function of the type of deformation is caused by different operating conditions of the filler and coarse clastic inclusions during compression and rupture.

For the tested soil the intensity of the reduction of strength with time increases as the filler content decreases and its ice content increases.

The effect of the composition of the filler on the strength of the frozen coarse clastic is exhibited in the following way. The soil with sandy filler has the greatest strength, and the soil with suglinok filler has the least.

A reduction in strength of the frozen soil with time is described quite well by the function that we obtained:

$$\left[\frac{\sigma_1 - \sigma_\infty}{\sigma_2 - \sigma_\infty}\right]^4 = \frac{t_2}{t_1}, \quad (1)$$

TABLE 1 Physico-Mechanical Characteristics of Frozen Saturated Coarse Clastics

Soil [specimen]	Granulometric Composition,% Fractions,mm 20-40	2-20	2	Moisture Content,% Soil	Filler	Unit Weight,g/cm^3 Soil	Filler	Porosity Soil	Coarse Clastics	Filler
Mixture I	15.0	85.0	--	18.3	--	2.03	0.92	0.34	0.34	--
Mixture II	14.0	81.0	5.0	16.7	333.9	2.07	1.09	0.32	0.35	0.91
Mixture III	12.0	68.0	20.0	14.9	75.0	2.12	1.50	0.29	0.43	0.68
Mixture IV	10.5	59.5	30.0	12.3	40.9	2.18	1.72	0.26	0.48	0.55
Mixture V	8.0	44.0	48.0	13.0	27.0	2.17	1.90	0.28	0.62	0.44
Mixture VI	8.0	44.0	48.0	12.9	26.9	2.18	1.91	0.27	0.62	0.44
Mixture VII	8.0	44.0	48.0	13.6	28.4	2.16	1.89	0.28	0.62	0.46

where t_1 and t_2 are the time of transition from the stage of steady state creep to the stage of progressive creep, respectively, under the effect of the loads σ_1 and σ_2.

The proposed function indicates that the frozen ground is characterized by a constant $N = (\sigma_i - \sigma_\infty)^4 t$ defining the nature of the variation of the strength with time.

The peculiarities of the deformation of the ice-saturated coarse clastics are considered by the test results for uniaxial compression on a prism of soil 20 × 20 cm in cross section and 40 cm high. The development of the deformation of the frozen soil with time with constant stress can be characterized graphically by the creep curves (Figure 2a), which are conventionally separated[1] into several sections depicting the various stages of deformation: "instantaneous" strain (ε_0) (completely restored after unloading), the strain of nonsteady state creep (ε_1), steady-state creep (ε_2), and progressive creep (ε_3).

Let us consider the relation for the specific

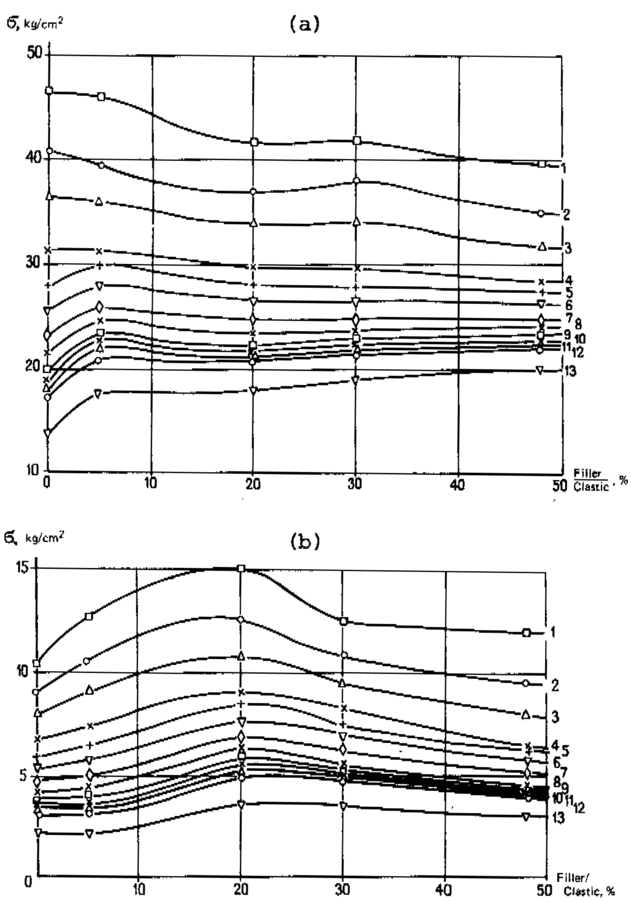

FIGURE 1 Strength of frozen coarse clastics as a function of the filler (supes) content for various points in time at a temperature of −5°. (a) Uniaxial compression. (b) Shear. 1--2 min, 2--5 min, 3--10 min, 4--30 min, 5--1 h, 6--2 h, 7--5 h, 8--10 h, 9--25 h, 10--50 h, 11--100 h, 12--200 h, and 13--∞.

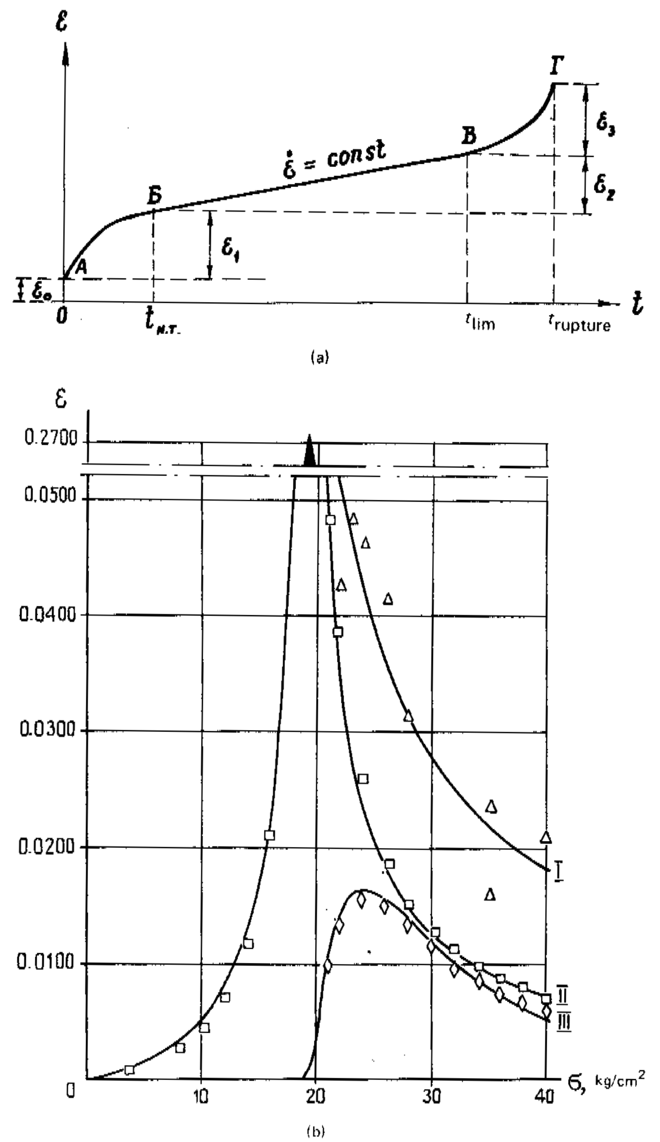

FIGURE 2 (a) Creep curve [typical] for frozen soil. (b) Creep deformation as a function of stress for coarse clastics (mixture VII) at a temperature of −5°. I--limiting compressive deformation $\varepsilon(t_{1im})$, II--deformation of non-steady-state creep ε_1, and III--deformation of steady-state creep ε_2.

values of the compressive deformation as a function of the magnitude of the load. The graphs (Figure 2b) constructed in the coordinates $\varepsilon_1 - \sigma$ (curve II) and $\varepsilon_2 - \sigma$ (curve III) depict the dependence of the maximum possible deformation on the stress in the stages of non-steady-state and steady-state creep, and the graph $\varepsilon(t_{1im}) - \sigma$ characterizes the dependence of the limiting value of the total compressive deformation on the magnitude of the load at the time of transition of the deformation process to the third stage of creep. From the graph it is obvious that the limiting deformation of the damped creep (curve II) increases from zero when $\sigma = 0$ to its maximum when $\sigma = \sigma_\infty$. A further increase in load causes

TABLE 2 Values of the Creep Parameters of Frozen Coarse Clastics in Uniaxial Compression

Soil [specimen]	Temperature, °C	Non-Steady-State Creep Deformation, ε_1: $C \cdot 10^3$, cm^2/kg	Non-Steady-State Creep Deformation, ε_1: a $\left(\frac{cm^2}{kg}\right)^{1/4}$	Steady-State Creep Deformation, ε_2: K $\left(\frac{cm^2}{kg}\right)^{1/4}$	Steady-State Creep Deformation, ε_2: b $\left(\frac{cm^2}{kg}\right)^{1/4}$
Mixture	-5	74.27	3.2	54.60	10.0
Mixture	-5	49.79	3.1	83.93	10.2
Mixture	-3	55.02	3.8	257.20	10.7
Mixture	-5	55.02	3.0	109.95	10.4
Mixture	-5	40.76	2.4	90.02	10.0
Mixture	-5	40.76	2.5	134.29	10.0
Mixture	-3	22.37	2.2	714.00	11.6
Mixture	-5	40.76	2.5	403.40	11.0
Mixture	-5	45.05	2.6	232.80	10.7

a decrease in the absolute magnitude of the limiting deformation of nonsteady creep. Thus, the loads very similar with respect to the long-term strength cause unlimited long ($t \to \infty$) deformation processes with gradually diminishing rate. This stress-strain state of the soil corresponds to the largest limiting values of the deformation of non-steady-state creep.[2]

The dependence of the limiting deformation of non-steady-state creep on the stress can be described by the expression:

$$\varepsilon_1 = C\sigma \exp\left[- a \mid \sigma - \sigma_\infty \mid^{1/4}\right], \qquad (2)$$

where C and a are coefficients, the values of which are presented in Table 2.

The relation between the strain of steady-state creep (ε_2) and the stress is illustrated in the graph (Figure 2b) of curve III. The limiting value of ε_2 of frozen coarse clastics varies from zero at $\sigma \leq \sigma_\infty$ to the maximum at $\sigma = \sigma_{opt}$. Then each successively increasing value of the stress corresponds to an ever-decreasing magnitude of the steady-state creep strain, which at the limit approaches zero when for very large loads the frozen soil is deformed as an elastic body.

The dependence of the stead-state creep strain (ε_2) on the stress is satisfactorily described by the expression:

$$\varepsilon_2 = K\left(\sigma - \sigma_\infty\right)^4 \exp\left[- b\left(\sigma - \sigma_\infty\right)^{1/4}\right], \qquad (3)$$

where K and b are empirical coefficients (Table 2). The limiting deformation of the frozen coarse clastics $\varepsilon(t_{lim}) = \varepsilon_0 + \varepsilon_1 + \varepsilon_2$, corresponding to the time of the transition from the steady-state creep stage to the progressive creep stage, decreases from the maximum value at $\sigma = \sigma_\infty$ to the minimum value for very large stresses (Figure 2b, curve I).

The limiting values of the compressive deformation $\varepsilon(t_{lim})$ essentially (by 2-4 times) increase with an increase in the filler content in the soil and correspondingly with an increase in porosity of the coarse clastic material.

In the soil with an identical percentage content of filler but different granulometric composition, the magnitude of the limiting deformation $\varepsilon(t_{lim})$ decreases in the series sand-supes-suglinok; that is, the maximum compressive deformation corresponds to soil with the strongest filler.

Thus, coarse frozen clastics can be considered as elasto-plasto-viscous bodies with characteristic forms of deformation. The strength and deformation properties of this type of soil are determined, on the one hand, by the composition of the soil and, on the other, by the effect of such external factors as the temperature, the form of the stressed state, and the magnitude and time of effect of the load.

Calculations with respect to strength and creep of coarse frozen clastics using the characteristics and formulas obtained [above] give a good comparison with the limiting deformation of the pillars in the placer mines of northeastern USSR.

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EFFECT OF FREEZING AND THAWING ON THE STRUCTURE, COMPOSITION, AND PROPERTIES OF COHESIVE SOILS

I. A. TYUTYUNOV, M. V. AVEROCHKIMA, AND V. P. TITOV Scientific Research Institute of Foundations and Underground Structures, All-Union Scientific Research Institute of Railroad Transportation

In the regions with severe climatic conditions, the so-called loess-like suglinok is widespread. The study of this type of suglinok has received a great deal of attention, and there are in the literature various points of view on the genesis of it. Some consider the loess suglinok to be the product of physical disintegration;[3] others, not refuting the participation of the physical destruction, consider that its origin is cryogenous physicochemical processes.[5] For a long time these points of view were based on field investigations. The first point of view was proved by the fact that the silty fractions formed under the effect of physical destruction cannot decompose into finer particles during chemical preparation for mechanical analysis. The second is proved by the fact that the chemical treatment, nevertheless, gives an increase in the finer fractions, but their low yield is explained by the irreversibility of the cryogenous coagulation of the colloidal particles. The fact that the colloids can be formed in the ground not only when freezing and thawing but also in the permafrozen state was proved earlier.[4] Nevertheless, the problem of the origin of loess-like suglinok cannot be solved without special experiment. Since formation takes place under the predominant effect of cryogenous physicochemical processes, obviously any very fine-grained soil placed under specific conditions must change its composition and approximate loess-like differences.

In checking this assumption in 1963, I. A. Tyutyunov reported on a special experiment, the essence of which consisted in the following: the "askangel"* and kaolin were placed under suglinok polar conditions in a series of loess-like suglinoks and left for 6 yr. The experiment was performed so that the askangel and kaolin exchanged solutions with the solutions of the loess-like suglinok surrounding them, but they could not exchange colloidal particles. For this purpose, the askangel and kaolin were first treated with salt solutions or substitution: in the first, univalent cations for calcium, and in the second, multivalent cations for calcium. The prepared pastes were used to fill double parchment cylinders 80 cm high and 10 cm in diameters. These cylinders were placed in double canvas bags of somewhat greater dimensions. Then two 5-in. holes 170 cm deep were drilled. The cylinders with the askangel were put into one hole, and the cylinders with kaolin in the other. The holes were filled with diluted loess-like suglinok and left in this state for 6 yr.

In 1969 the cylinders of askangel and kaolin were extracted. X-ray structural studies and analyses of the infrared spectroscopy do not reveal noticeable changes in the crystalline structure of the initial materials. However, the assumptions of the possibility of variation of the microaggregate and granulometric composition of the soil were confirmed. From the presented data (Table 1), it is obvious that the kaolin underwent significant transformations in the direction of a decrease in grain size: As a result of destruction of the 0.05-0.01-mm fraction, an increase in the 0.005-0.001 and less than 0.001-mm fractions took place. The variation of the microaggregate composition of the askangel took place in the other direction. The general trend here is characterized by significant aggregation of the particles, primarily the 0.25-0.05-mm and 0.05-0.01-mm fractions. The number of fractions < 0.001 mm sharply decreased and in practice turned out to be the same as in the loess-like suglinok.

The changes in the microaggregate composition as a result of chemical treatment characterized by the granulometric composition of the soil are quite significant. In kaolin and in the askangel, the fraction of less than 0.001 mm increased by about 10 percent. However, it is necessary to note that the initial grain sizes of the natural askangel were not restored by chemical treatment. In the 6 yr under natural conditions of the polar regions, the askangel underwent significant irreversible changes.

Changes of an opposite nature took place in the kaolin. Here, the effect of saturation with the potassium ion was felt. These transformations are confirmed to a certain degree by the variation in absorption capacity. In the kaolin the total amount of calcium and magnesium in the upper meter layer doubled. In the askangel the consumption decreased somewhat, especially in the lower layers. As should be expected, the composition of the absorbed bases of the kaolin and askangel was balanced with the composition of the loessial suglinok.

The studies that were performed are important to study the phenomena directly connected with practice. It was demonstrated theoretically and experimentally[2,5] that the soil subjected to the least heave is the soil in which significant aggregation of the particles during the freezing process is observed. This is confirmed by observations under natural conditions. When study-

*Highly-active Na-montmorillinitic clay.

ing the phenomena of heave formation in the northern and northwestern parts of the country, a number of interesting principles were discovered. When freezing fine-grained soil, the formation of ice interlayers takes place accompanied by dehydration of the mineral skeleton, volumetric shrinkage, and an increase in the soil density. This process takes place the more intensely the higher the soil moisture content before freezing. During thawing of the ice inclusions in the ground massif, voids remain for some time into which there is displacement of part of the thawed moisture and also the moisture filtering in from the surface. Accordingly, secondary variation of the initial granulometric composition of the soil can occur. The redistribution of the soil particles of different coarseness was observed under natural conditions when studying solifluction phenomena. Here, during thawing, which is the characteristic of the intensity of the investigated process, the soil moisture content was as follows: The total moisture content considering the ice inclusions reached 46.52 percent, the moisture content of the soil aggregates did not exceed 22-26 percent at the same time as the liquid limit was 32-34, and the plasticity index was 12-14.

An important characteristic of such soil is the fact that during the freezing process it is subject to aggregation, not as a result of coagulation of the elementary particles, but in connection with the combination of the smaller aggregates. Experience shows that the silty ground in the freezing zone acquires a higher permeability, low bearing capacity[1] during thawing, and the capacity for thixotropic relief.

The stress relief in the soil during thawing was estimated by special experiments in the laboratory, which stated as their goal the investigation of the effect of the natural structural bonds on the degree of stress relief in the soil during thawing. Silty surface-suglinok with disturbed and undisturbed structure was tested. The samples with undisturbed structure were cut out of blocks using rings 2 cm high and 40 cm^2 in area, and then they were placed in wet sand to complete their saturation. The samples with disturbed structure were prepared from the same monoliths in the form of a paste, with density and moisture content corresponding to the state of the samples with undisturbed structure after saturation. The soil was frozen at a temperature of -3°C (without the supply of water from below). The shear strength was determined immediately after thawing of the specimens, without stabilization of the compressive load. The shear was made with gradual application of the shearing force. In the first series of experi-

TABLE 1 Microaggregate Composition of the Soils

Depth of Sample, cm	Content of Fractions,%					
	1-0.25	0.25-0.05	0.05-0.01	0.01-0.005	0.005-0.001	<0.001
	Loess-like Suglinok					
0-10	2.67	19.05	55.69	5.16	10.74	6.69
40-50	6.12	14.30	55.41	9.12	6.73	8.32
70-80	6.51	11.70	61.79	8.94	4.09	6.97
90-100	10.37	21.63	41.48	7.59	13.71	5.92
130-140	7.01	23.71	38.61	8.90	14.92	6.85
	Kaolin					
0-10	0.19	13.95	19.79	1.60	24.49	40.00
40-50	0.26	8.03	11.17	19.06	29.78	31.00
70-80	0.26	12.34	14.32	9.17	32.95	30.96
90-100	0.07	16.25	10.41	10.72	29.37	33.18
110-120	0.29	14.59	15.93	7.58	31.47	30.43
150-160	0.13	14.06	20.84	3.31	32.82	28.84
K-kaolin	traces	19.20	70.60	6.00	3.20	--
	Askangel					
0-10	5.90	52.24	16.34	11.37	6.02	8.13
40-50	5.61	10.75	60.69	1.61	13.66	7.69
70-80	2.20	52.51	22.38	3.43	12.96	6.52
90-100	0.92	42.25	31.69	2.36	17.58	6.03
110-120	1.32	65.73	7.19	8.65	9.37	7.76
150-160	1.70	55.65	22.23	4.90	11.19	5.23
Ca-acid-activated clay gel	10.30	30.90	28.60	8.80	10.80	10.60

NOTE: The analyses were performed by M. Ye. Semina.

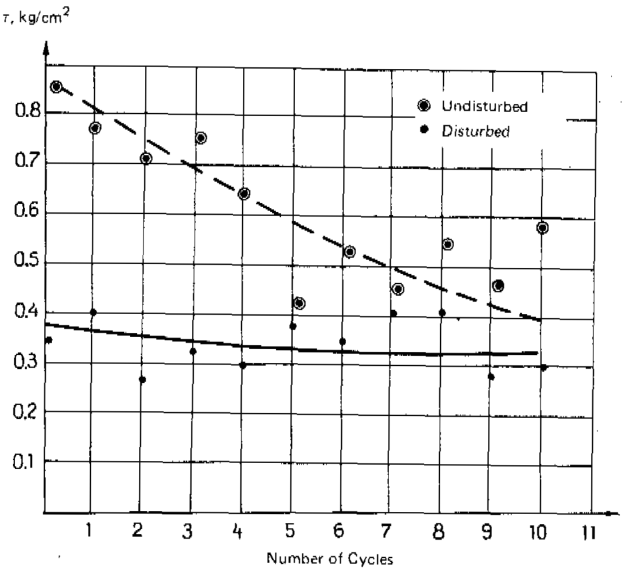

FIGURE 1 Variation of the shear strength during cyclic freezing and thawing.

ments, the shear time was 7 min; then it was reduced to 1 min to avoid consolidation of the soil. The dependence of the shear strength on the number of freeze-thaw cycles is presented in Figure 1.

The results of the experiments demonstrated that the reduction in soil strength with an increase in the number of freeze-thaw cycles takes place to some limit, after which it weakens. For the paste samples this reduction in strength takes place less intensely than for the block samples. The greatest reduction in strength is observed after one or two freezing cycles. This is also confirmed by construction practice: The first one or two seasons of deformation of the ground, as a rule, are the most intense. In addition, it is noted that the values of the strength of the soil samples with undisturbed structure approach each other as the number of freeze-thaw cycles increases.

Thus, the experimental data confirm that the freeze-thaw processes lead to a reduction in strength of the ground as a result of disturbance of the natural strutctural bonds.

The reduction in strength of the ground with thawing can be evaluated using the "coefficient of freezing sensitivity," that is, the ratio of the strength of the unfrozen, and thawing, soil samples. As the experiments by Ye. P. Shusherina revealed,[6] the relative reduction in strength is greater the higher the initial density of the soil. According to the data from the experiments that we performed, this relation is more clearly expressed for disturbed samples (Figure 2). The experiments with samples with undisturbed structure demonstrated with respect to the closed system that in the density range of 1.50-1.67 g/cm^3, the effect of the latter is felt insignificantly. Consequently, when forecasting the degree of possible stress relief of the soil, it is necessary to consider the density, the structure, and freezing conditions.

For the cases of inadmissible reduction in the bearing capacity of the soil, it is expedient to strengthen it with chemical additives. The latter must be a means increasing the previous, and creating new, structural bonds. The corresponding experiments were performed by the procedure discussed above, but aluminim sulphate, ferric chloride, calcium chloride, and quickline were added to the soil.

As a result of the introduction of these additives, partial dehydration of the ground particles and the formation of cementing gels were observed. Among the additives used, the most effective were ferric chloride and calcium chloride, after the introduction of which the moisture content of the specimen remained in practice constant. On the whole, all the chemicals used demonstrated that the soil density is made stable with time and stable even during cyclic processes of freezing and thawing.

As is known, during the freezing process there is heave, which is determined by the intensity of the moisture migration to the freezing front. In turn, the nature and intensity of the migration depend to a significant degree on the soil composition, the type of structure, and the magnitude and number of the pores and capillaries. The disturbance of the natural soil structure causes a reduction in the heave, which was demonstrated by field experiments. They consisted in the fact that in special areas the ground is dug out to a different depth from the surface, and then it is packed to the density of the soil in the control sites with undisturbed structure. The results of the observations (leveling of the surface of the site) appear in Table 2. In the indicated experiments, the ground in the sites is excavated not to the entire freezing depth. The latter, which varies in different seasons, fluctuated within the limits of 1.0-1.2 m. However, the disturbance of the structure is felt in the magnitude of the heave quite significantly. The number of thickness of the horizontally-

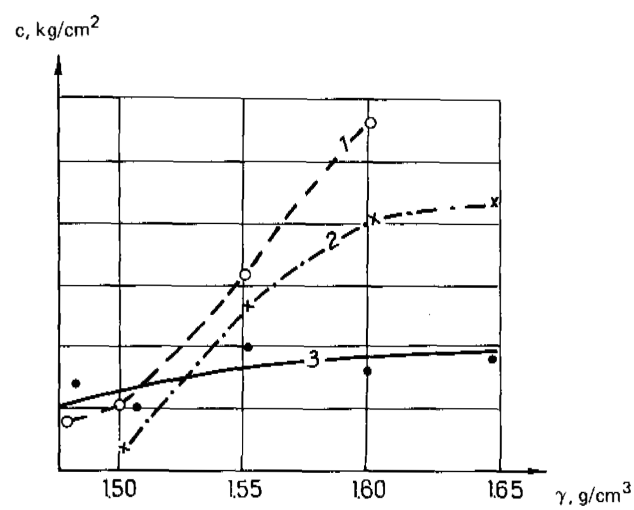

FIGURE 2 Cohesion of the soil as a function of density with a different number of freeze-thaw cycles: 1--without freezing, 2--one freezing cycle, and 3--five freezing cycles.

TABLE 2 Effect of Disturbance of the Natural Structure on Ground Heave

State of the Soil Structure	Magnitude of the Heave, mm: Observation Seasons 1	2	3	4	Thickness of the Excavated Layer H, m
Undisturbed	156	219	261	241	--
Disturbed	76	167	277	257	0.9
Undisturbed	--	159	282	272	--
Disturbed	--	88	210	229	0.8
Undisturbed	103	--	--	--	
Disturbed	53	--	--	--	0.6

oriented interlayers of ice decreases sharply within the cores taken in the sites with the soil dug out for which the reticular cryogenous texture is characteristic. With time the effect of the disturbance of the structure attenuates, since the formation of the secondary structure in the ground takes place which, with respect to its nature, gradually approaches the natural "frozen" structure.

The data presented show that the study of the dynamics of the process of variation of composition and structure of the soil during freezing and thawing permits us to obtain the necessary solutions to a number of practical problems.

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COMPRESSOMETRIC STUDIES OF THE STRENGTH AND DEFORMATION OF FROZEN SOILS DURING THAWING

V. B. SHVETS, V. V. LUSHNIKOV, I. A. MARENINOV, AND B. I. SUKHANOV Ural Kirov, Polytechnic Institute

The compressometric method of testing thawing soils is becoming more and more widespread. The obvious advantages of this method are simplicity and compactness of the devices, the broad range of their application with respect to depth, and also low energy consumption. They make it possible to use the method when performing experimental work on undeveloped sites.

The compressometric equipment and procedure for determining the strength and deformation of thawing ground developed by the authors was used when testing some permafrozen suglinok (Chitinskaya Oblast), permafrozen sand and supes (Nadym), and seasonally thawing suglinok and clay (Sverdlovsk). The tests were run at depths of 2-10 m. The investigated ground had primarily homogeneous and frequently reticular cryogenous texture. The thawed clay soil was, with respect to consistency, in the firm- and soft-plastic states.

In order to determine the compressibility of the frozen ground during thawing, an air-electric compressometer,* type PEV-127,[1] was used in holes 127 mm in diameter. The operating chamber of the compressometer was equipped with a heating element, the power supply to which was from a portable station with a drive from an internal combustion engine. The heating regime was controlled and regulated by the readings of a temperature gauge fastened to the outside surface of the shell of the operating chamber

During the process of testing in the operating chamber (Figure 1a), the initial pressure required to expand the shell to the initial diameter d_m was created in order to bring it against the walls of the hole and to transmit a small residual pressure to the ground--on the order of 0.1 kg/cm^2. Then the operating element was heated at a temperature of 70°-90° for 6-10 h to form a cylinder of thawing ground around the well with an outside diameter D equal to 30-50 cm.

On completion of heating, tests were run on the thawing ground with increasing load stages from 0.2 kg/cm^2 to 0.5 kg/cm^2 as a function of the form and state of the ground with maintenance of each stage to provisional stabilization of the deformation (0.1 mm in 15-30 min). The actual dimensions of the thawing zone were more precisely defined after testing with a probe with mechanical drive. The schematic of the formation of the thawing zone on the test graph in the coordinates of the pressure p and the well wall deformation Δd for different duration of heating appears in Figure 1b.

*Also termed "Pressuremeter."

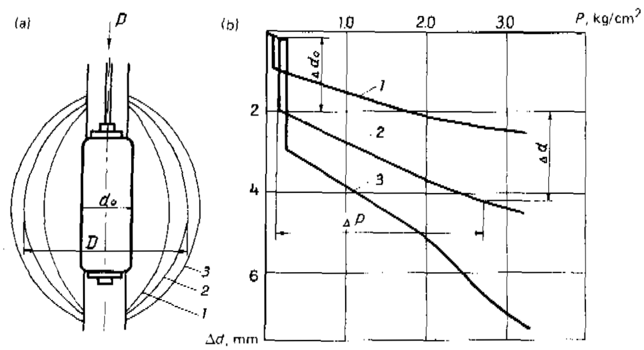

FIGURE 1 Schematic and graphs of compressibility tests run on permafrost during thawing: (a) schematic of the operating chamber and boundary of the thawing zone 1-3 after heating the ground 2, 5, and 10 h, respectively; (b) test graph 1-3 corresponding to the indicating heating times. d_0--initial borehole diameter (after heating), D--outside diameter of the cylinder of thawing ground, Δd_0--increment of the initial diameter of the well from the effect of heating only, and Δd--increment of the borehold diameter in the thawing section in the pressure range ΔP. The graph gives the representation of the magnitude of the increment in the well diameter Δd_0 as a result of thawing of the ground and the dependence of the deformations of the thawed ground on the transmitted pressure.

In cases where the moisture of the frozen soil was high and during the process of thawing of the ground the danger of loss of stability arose, the borehole was reinforced with casing. The operating chamber was placed in the lower part of the casing equipped with a deformable section[2] 1,200-1,500 mm long, which was made of individual elastic split elements* attached to the ends of the casing. The experiment was performed in the heated section directly inside the deformed part of the casing, which could simultaneously be used for the removal of excess moisture.

The compressibility characteristics of the thawing ground--the coefficient of relative compression during thawing A_0 and the deformation modulus of the thawing ground E_0--were determined by the formulas that take into account the

*For the structural design of the device, the Committee on Inventions and Discoveries under the USSR Council of Ministers adopted a resolution on 18 December 1971 to issue an author's certificate according to claim No. 1610347/29-14.

settlement and compression deformation within the limits of the heated ground cylinder of limited size. The formulas are:

$$A_0 = \frac{d_0 - d_m}{D - d_m},$$

$$E_0 = kwd_0 \frac{\Delta p}{\Delta d},$$

where d_m, d_0 are the wall diameter before and after thawing, respectively; D is the outside diameter of the cylinder of thawing ground; Δp, Δd are the pressure increment and the increment in the diameter of the well in the linear section of the test graph, respectively; k is a correction factor established on the basis of a comparison of the results of the compressometric and standard testing of the soil in the thawed state by indenters; and w is the coefficient taking into account the limited size of the compressed zone around the chamber of the compressometer.

As a result of testing the permafrozen suglinok, A_0 = 0.01-0.05 and E_0 = 50-80 kg/cm^2 would obtain, which corresponds to the order of these magnitudes according to the data from laboratory testing of the thawed ground by ordinary compressometers (without the heating element).

In order to determine the strength characteristics of the soil in the frozen and thawed state, the translational shear PPS-89 compressometer* was used in boreholes 89 mm in diameter.[3] The servomechanism of the instrument (Figure 2a) equipped with a heater is assembled from individual unit elements advancing together with the shell of the compressometric chamber when a pressure is created inside the cavity. The rigid elements of the shell have transverse ribs of small thickness inserted into the ground. The shear is realized under the given pressure normal to the wall of the borehole by shearing the ground in the longitudinal direction.

After lowering the compressometer into the borehole, a pressure on the order of 1 kg/cm^2 is created inside the servomechanism, which is necessary for insertion of the blades into the ground until the surfaces of the longitudinal elements are flat against the walls of the borehole. In order to facilitate the introduction of the blades, the servomechanism is heated simultaneously. The heating regime is regulated in order to obtain a given positive or a negative temperature in the shear zone. When testing the ground during thawing, the cylinder diameter of the thawed ground must exceed the diameter of the servomechanism by 5-10 cm, which usually is reached in 15-30 min. of heating.

The shearing of the ground is realized after achievement of the given temperature by a gradual increase in the longitudinal withdrawing force. After the first shear, the second and subsequent experiments are performed with the same soil for other, significantly different, values of the shear normal to the surface and pressures. The compacting time under normal pressure and the rate of application of the shearing force are taken as a function of the adopted test procedure (fast or slow shear). The values of the pressure normal to the shearing surface were taken from 1 kg/cm^2 to 5 kg/cm^2 when testing frozen ground and from 1 kg/cm^2 to 2-3 kg/cm^2 when testing thawed ground.

*For the structural design of the device, the Committee on Inventions and Discoveries under the USSR Council of Ministers adopted a resolution on 6 October 1971 to issue an author's certificate according to claim No. 1439028/29-14.

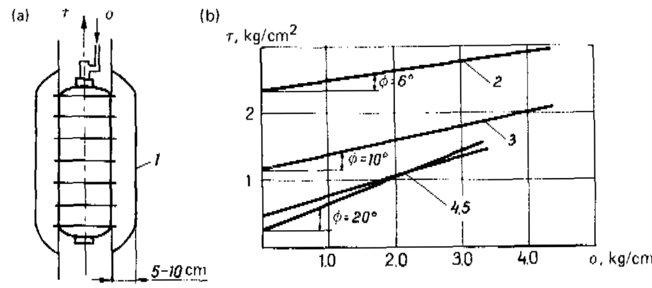

FIGURE 2 Schematic and graph of the shear testing of frozen ground during thawing: (a) schematic and graphs of the servomechanism and the boundary 1 of the thawing zone; (b) graph of the shear $\tau = f(\sigma)$ of the soil 2-5 at a temperature in the shear zone of -2, -1, +1, and 5°C, respectively.

The limiting shear strength was determined considering the actual area of the sheared cylinder and the resistance of the prism of soil to the upper blade. The shear parameters c and ϕ were determined by graphical or analytical processing of the results of several experiments. In Figure 2, we have the graphs of the shear $\tau = f(\sigma)$ for soil with different temperatures in the shear zone (from -2°C to + 5°C). The graphical data indicate the sharp reduction in the ground strength during the thawing process under natural conditions.

Thus, the use of the compressometric method to study the deformation and strength characteristics of the permafrost during its thawing turned out to be highly effective. This method can be one of the new areas of field research--thermocompressometric measurements of permafrost.

CONCLUSIONS

1. On the basis of the compressometric method, the apparatus and procedure for determining the compressibility characteristics of permafrost during thawing and the strength characteristics of the ground with given positive and negative temperature were developed.

2. The tested apparatus is distinguished by compactness, broad range of use with respect to depth, and applicability for testing permafrost of homogeneous and reticular structure.

3. Comparative tests performed by the standard and proposed thermocompressometric methods

demonstrated satisfactory coincidence of the order of magnitude of the compressibility and strength characteristics of the ground.

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NATURE OF WATER IN VERY FINE-GRAINED DEPOSITS AND ITS CRYSTALLIZATION CHARACTERISTICS

A. A. ANANYAN *Moscow Lomonosov State University*

Studies by soil scientists and geocryologists have demonstrated that the properties of water in fine-grained deposits and, in particular, in the topsoil, differ from the properties of water in a tank. It has been demonstrated, for example, that the freezing point of this water is below 0°C. When the temperature drops to the freezing point, the formation of the first ice crystals takes place; the rest of the water remains in the unfrozen state and is in dynamic equilibrium with the released ice crystals. On further reduction of the temperature, some of the water molecules crystallize further, and so on.[1,2] However, a certain amount of water does not convert to ice in general but remains in the unfrozen state.[3]

The dynamic equilibrium between the ice crystals and the unfrozen water is an entirely stable state, which is confirmed by permafrost studies where this equilibrium has been maintained for millennia.

Therefore, the water contained in fine-grained deposits is considered as a special phase, the properties of which differ from the properties of water in a tank.[4]

STRUCTURE OF WATER IN FINE-GRAINED DEPOSITS

Let us consider the system made up of the water and fine-grained soil. For simplicity, let us assume that the water does not in practice contain water-soluble salts.

The basic components of most of the soil are represented by oxides [and silicates]. Their surface is hydrated and contains hydroxyl groups. On interaction with water, the active centers of the surface of the mineral particles are primarily the hydroxyl groups and the coordinated-unsaturated silicon atoms on the surface of the silicates. Along with the coordination [ionic] bonds, part of the water molecules form hydrogen bonds with the surface atoms of the oxygen and hydrogen.[5] The role of the exchange cations in binding the water to the surface of the mineral particles is much less.[6]

On the basis of experimental and theoretical developments, it has been demonstrated that in the system made up of water and fine-grained mineral particles the active centers of the surface of the mineral particles affect the translational motion of the water molecules nearest to them--they orient them. This orienting effect is transferred to the "relay" mechanism to the subsequent layers of water, diminishing with distance from the surface.[7,8] This leads to variation of the potential barriers separating the adjacent positions of equilibrium in the structure of the liquid, to variation of the mobility and frequency of the molecule exchange, which, in turn, causes energy inhomogeneity of the adjacent water molecules.

Since the normal structure of the water is characterized by the energy equivalent of all the molecules in the tetragonal pattern, the variation in mobility of part of the molecules leads to variation in structure. Therefore, the structure of the water is distorted. The topography of the active centers on the surface of the mineral particles makes its own contribution to the distortion of the structure of the water, since the arrangement of the active structures does not coincide with the arrangement of the molecules in the water structure.[9] As a result of this effect, some ordering of the structure of the nearest layers of water takes place with respect to the active centers of the surface of the mineral particles; the heat of wetting is released here.

As has been noted,[10,11] the distortion of the structure is a kinetic obstacle complicating the crystallization of the water at 0°C, since for crystallization it is necessary, in addition to lowering the temperature, to have a definite arrangement of the molecules close to their arrangement in the structure of the ice.

The representation of the distortion of the water structure in fine-grained soils (the oriented state of the water molecules) was introduced into science for the first time in 1958.[12] The basis for representations of the distortion of the structure is discussed in part by A. A. Ananyan.[13] Let us present some of the principles.

1. The reduction in the freezing point of the soil-water, also the presence of the liquid phase of the water in a frozen soil along with ice crystals at a temperature below the freezing point of the soil, indicates that the unfrozen water differs with respect to structure from the part of the water which has crystallized. A gradual decrease in the amount of unfrozen water with a reduction in temperature indicates nonuniformity of the structure.

2. The capacity for supercooling is one of the indirect indications of the nature of the structure of the liquid phase. The less water in the soil, the more its structure is disturbed by the effect of the active centers of the surface of the mineral particles, the lower its energy level, the greater the difference of it from the water in a tank, and the more prolonged the supercooled state, which, as experiments have demonstrated, in a number of cases is measured in weeks and more.[11]

3. On percolation of water through fine-grained soil, an increase in the soil temperature led to an increase in the permeability, not only as a result of a decrease in the viscosity of the water with temperature, but also as a result of disturbance of the oriented state of the water in the soil.[14] This again confirms our concepts of the oriented state of the water in fine-grained soil.

4. The studies of the times of transverse relaxation of the water protons (T_2) by the nuclear-magnetic resonance-spin-echo method demonstrated that as the moisture content increases the relaxation time increases monotonically. This indicates an increase in the energy level of the water molecules in the soil. For example, we have demonstrated that with an increase in the moisture content of kaolin samples from 5 to 52.3 percent T_2 increased from 1.7 ms to 6.8 ms, respectively. On variation of the moisture of askanite* from 10 to 83 percent, T_2 varied from 0.10 ms to 1.0 ms, and so on.

The following questions arise: (1) What causes the reduction in freezing point in practice of unsalinized fine-grained soil (including topsoil)? (2) If the distortion of the structure of the water is a kinetic obstacle to crystallization, then why with a further reduction in temperature does crystallization of the unfrozen soil take place in the frozen rock? (3) What causes the coexistence of ice crystals with liquid phase of the water (with the unfrozen water)? (4) What is the structure of the ice formed from the unfrozen water; if this is ice, then why is it formed and why does it melt not at 0°C but at different, lower, temperatures?

In order to answer these questions, it is necessary to consider the ratio of the forces active in the water contained in the fine-grained soil.

In the liquid phase made up of water and the fine-grained mineral-particle system, the basic active forces of interest to us are the following: (a) the long-range ion-dipole forces between the active centers of the surface of the mineral particles and the nearest water molecules distorting its structure. These forces decrease relatively slowly with an increase in spacing between the particles. (b) The short-range forces sharply decreasing with distance. These are interactions among the water molecules striving to group the molecules into tetrahedrons.

With a reduction in temperature, in connection with a decrease in the translational motion, orientation effects increase, and the water viscosity increases. The distortion of its structure complicates the crystallization at 0°C; the water in nature remains in the liquid state.

It is highly significant that, with a further reduction in temperature, the effect of the long-range forces increases slowly, and the short-range forces, comparatively rapidly.[15] The interaction of the molecules least oriented by the surface turns out to be stronger than the ion-dipole interaction distorting the structure of the water. Therefore, the least-oriented molecules go out of the sphere of orientation of the surface and are bound into tetrahedrons. With sufficiently low temperature, they are grouped into the ice structure. The isolation of the first ice crystals corresponds to the freezing point of the rock. The rest of the water remains in nature in the unfrozen state. With further reduction in temperature, the interaction among the water molecules increases. This leads to a decrease in the amount of ice, and so on. Some water molecules corresponding to 1-2 molecular layers generally do not crystallize at lower temperatures and remain in the vitreous stage, since they are so oriented that they are not capable of being grouped into the ice structure.[3]

The crystals formed from the unfrozen water are in dynamic equilibrium with the rest of the water. With a reduction in temperature and a decrease in thermal motion, the predominant increase in the short-range forces by comparison with the increase in ion-dipole interaction leads to an increase in the amount of ice and a decrease in the unfrozen water. With a reduction in temperature, on the contrary, the

*Montmorillonitic clay.

effect of the short-range forces decreases, some of the molecules decrease again under the effect of the surface forces, and the amount of unfrozen water increases.

Thus, in the frozen fine-grained soil depending on the direction of the external thermal effect, the formation or melting of ice crystals takes places. Under defined temperature conditions, the solid and liquid phases of the water in dynamic equilibrium coexist.

The problem of the structure of ice formed from unfrozen water has not been investigated up to now. If we assume that for any temperature stage a special configuration of the ice is formed, then it turns out that there are an enormous number of configurations that contradict nature. Obviously, it must be assumed that during crystallization of the unfrozen water, Ice-I is formed, possibly with a defective structure that occurs and melts not at 0°C but at different lower temperatures.

The coexistence of the solid and liquid phases of the water in the frozen fine-grained soils and the phase transitions of the water in them arise from the fact that the ice crystals are in equilibrium, not with ordinary water but with water having a distorted structure, the molecules of which are so oriented by the surface effect that under the given temperature conditions they are not capable of grouping into an ice structure.

CRYSTALLIZATION-DIFFUSION MECHANISM OF MOISTURE MIGRATION

The freezing of fine-grained soil is connected with redistribution of the moisture and with heave phenomena. The necessary condition for this is the presence of unfrozen water in the freezing zone, minimum mean thickness of the water films in this zone of the soil (relatively high energy level of the water molecules), and sufficient permeability of the soil.[1]

It has been noted that the closer the layers of water are located to the active centers of the surface of the mineral particles, the more sharply the structure of the water is distorted--in the general case, the lower the mobility of its molecules and the lower their energy level.[11,16]

As practice has demonstrated, in the majority of cases where the mean thickness of the water film in the thawed zone is <9-10 molecular layers, the moisture migration process in the liquid state practically does not occur, since the molecules of this water are characterized by relatively low energy level.

When the fine-grained soil freezes, the least oriented water molecules turn into ice. In the remaining unfrozen water, the thickness of the film decreases. The smaller thickness of the film of unfrozen water, by comparison with water films in the thawed zone and the lower temperature in the freezing zone, cause less mobility of the molecules of unfrozen water. Therefore, in the case of one-way freezing, the frequency of the jumps of the water molecules from the films in the thawed zone into films in the frozen zone turns out to be greater than in the opposite direction. There is predominant displacement of the water molecules by self-diffusion from the thicker films in the thawed zone into the thin films in the frozen zone, that is, from the sections with higher structural temperature into the sections with lower. For example, in the Moscow suglinoks with a specific surface of 101 m^2/g and at a moisture content of 31.4 percent, the number of molecular layers was on the average 11.3, and T_2 was about 0.7 ms; at a temperature of -0.5°C, the amount of unfrozen water turned out to be 9.7 percent, and the [average] number of molecular layers, 3.2. This number of layers corresponds to T_2 of about 0.3 ms.

At the present time there is no theory that would permit direct relation of the experimentally observed values of the relaxation times to the characteristics of the thermal motion of the molecules in the adsorbed state; only qualitative estimates are possible.[17] Therefore, the presented values of T_2 only characterize the qualitative difference in the energy levels.

The water molecules entering the frozen zone turn out to be "excessive" by comparison with the number that can remain in the liquid state under the temperature conditions of the frozen zone; therefore, they are grouped into an ice structure,[1,12] which in a number of cases leads to the heave phenomena.

CONCLUSIONS

1. The water in fine-grained soils represents a special phase characterized by a distorted structure and properties differing from the properties of water in a container.

2. The cause of the reduction in freezing point of the unsalinized fine-grain rock (in particular, soils and topsoils) is the distortion of the structure of the water in it.

3. The crystallization of the unfrozen water in the presence of a reduction in temperature takes place because, in the liquid phase of the system made up of the water and the fine-grained soil, the ratio of the active forces changes. The oriented effect between the water molecules (dipole-dipole, short-range forces) grouping them into the ice structure increase more rapidly than the orienting interaction among the active centers of the surface of the mineral particles and the nearest water molecules (ion-dipole, long-range forces) distorting the water structure.

4. The coexistence of the liquid and solid phases of the water and the phase transitions of it into ice in frozen fine-grained soils are caused by the fact that the ice crystals are in dynamic equilibrium, not with ordinary water but with water having a distorted structure, the molecules of which for the given temperature are not capable of being grouped into the ice structures; hence they are oriented by the surface forces of the mineral particles.

5. The migration of the moisture in the liquid

state from thawed sections to the freezing sections of the soil is caused by the crystallization-diffusion mechanism of displacement of the water molecule.

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STRUCTURES, PHASE TRANSITIONS, AND PROPERTIES OF FREE AND BOUND WATER

B. N. DOSTOVALOV   Moscow Lomonosov State University

VARIATION OF THE STRUCTURE OF ICE DURING HEATING--THE CONDITIONS OF ITS STABILITY AND MELTING

The phase transitions and corresponding changes in the structure of matter are caused by the disturbance of a stable or equilibrium relation between the energies of motion and interaction of its component elementary particles

The crystal lattice of the ice is a regular alternation in space of "potential barriers" and "potential wells" in which the atoms (molecules) are arranged (Figure 1a). The energy required for tearing the particle out of the "potential well" is called the "activation energy." When heating a material, the mean kinetic energy of the particles E_k increases, and the mean activation energy E_{act} decreases correspondingly. As is obvious from Figure 1a, $E_{act} = E_0 - E_k$, where E_0 is the total potential energy of the "near interaction of the particles" measured by the work of removal of the particle from the potential well at a temperature of $T = 0°K$.

When heating a single ice crystal in the temperature range of $T < T_{melt}$, the kinetic energy E_k will increase and the activation energy E_{act} will decrease (Figure 1b). Under the condition where $E_{k\ max} > E_{act\ min}$, the jumping of the molecules from certain nodes of the lattice to others begins. Simultaneously empty nodes (vacancies) appear in the ice lattice, and the torn-out molecules will move in the cavities of the lattice between the lattice nodes, creating an internal vapor pressure P_{int}. Part of these molecules will again occupy the vacancies, healing the lattice defects. If the heat flux is stopped, then the internal evaporation and healing of the lattice will reach equilibrium with equality of the internal and external vapor pressure. Thus, the equilibrium and stability of the ice lattice are expressed by two conditions:

$$E(T)_k < E(T)_{act}, \tag{1}$$

$$P_{int} = P_{ext}. \tag{2}$$

the temperature range of the crystal stability on satisfaction of relations (1,2) is shown in Figure 1b.

When the mean kinetic energy exceeds the mean activation energy (Figure 1b), the crystal lattice begins to deteriorate, that is, the ice begins to melt. The melting condition can be written in the form of the following equations:

$$\bar{E}(T_{melt})_k > \bar{E}(T_{melt})_{act}, \tag{3}$$

$$P(T_{melt})_{int} = P(T_{melt})_{ext}. \tag{4}$$

From condition (3,4) and Figure 1b, we have the following conclusions: A reduction in the melting point at constant pressure is possible only with a decrease in $\bar{E}(T)_{act}$ and, on the contrary, an increase in $\bar{E}(T)_{act}$ must increase T_{melt}; the increase in P_{int} and P_{ext} increases, and their decrease lowers the melting point correspondingly.

STABILITY OF FREE WATER, ITS STRUCTURE, PHASE TRANSITIONS, AND PROPERTIES

With further heating of the ice, all the pulses with maximum energy are expended on breaking the bonds; therefore, the temperature does not rise.

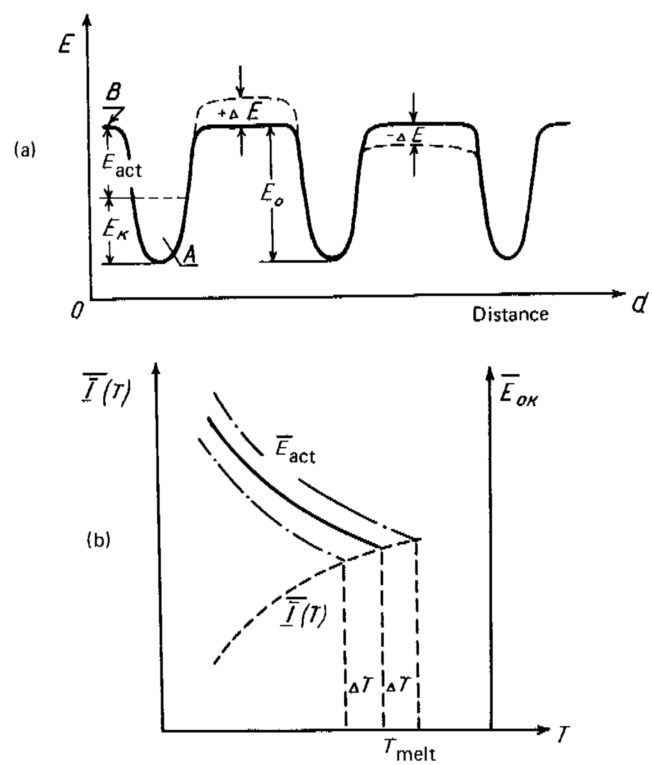

FIGURE 1 Alternation of the potential barrier (B) and the potential well (A) in the material (a). E_o is the total potential energy of the barrier (0°K), E_k is the kinetic energy, and E_{act} is the activation energy of the particle. The temperature range of ice stability is defined by the condition $\bar{E}(T)_{kin} \leq \bar{E}(T)_{act}$ (b).

When the influx of heat reaches 30 cal/g, the apparent isothermal melting ends, and the water begins to heat up.

Here, 9-11 percent of the bonds break. This is obvious from the relation of the specific heats of melting and evaporation of the ice. It is possible to answer the basic question of why the isothermal melting ends so quickly in spite of the complete impossibility of breaking the bonds (Formula 3) and what stops it is the following: When the ice melts, such a large number of vacancies are formed that the pulses, along with activation of the individual molecules, tear out entire pieces of the lattice. The jumps or Brownian motion of the molecule aggregates (clusters) begin. Also, the pulses can accelerate the motion of the clusters without breaking the bonds inside them; that is, the possibility of an increase in temperature arises. Thus, during melting a qualitative change in the translational motion takes place, and the diffusion in the water D is expressed by the formula:

$$D = \sum_{i=1}^{i=n} A_i e - \frac{E(T,i)_{act}}{RT}, \qquad (5)$$

where A_i are some of the constant coefficients and n is the number of molecules in the clusters completing the jumps.

The mean size of the clusters $\bar{r}$ must decrease with an increase in the mean pulse energy $I(T)$ and temperarure, which is illustrated in Figure 2a. Thus, the isothermal melting of the ice stops, because, when the heat of melting is communicated to it, the mean size of the clusters becomes so small that the mean pulses $I(T)$ cannot destroy them; but they only increase kinetic energy, that is, they increase T.

The correctness of the indicated relation (Figure 2a) can be explained by the following arguments: If a cluster is in collision, the part of it directly receiving the acceleration will entrain the rest of the cluster as a result of the cohesive forces in some cross section $f(r^2)$, and the entrained part under the effect of the forces of inertia $\phi(r^3,a)$ will counteract this. If $f(r^2) < \phi(r^3,a)$, then rupture of the cluster takes place. Since with a decrease in the dimensions $r(n)$ of the clusters, the forces of inertia $\phi(r^3,a)$ decrease more rapidly than the cohesion $f(r^2)$, with a constant pulse (temperature) there must be a minimum size of the cluster r_{k1} below which the given pulse cannot destroy a cluster. Thus, the inequality

$$f(r^2) \geq \phi(r^3,a), \qquad (6)$$

where a is the acceleration of the entrained part of the clusters, is the condition of stability of the mean size of the clusters and simultaneously the condition of cessation of their further melting H_2O at given temperature and vapor pressure. In Figure 2a, we also have the depedence of the mean size of the clusters $\bar{r}$ on the vapor pressure P. With a decrease in vapor pressure (P), the clusters evaporate, and at the same temperature it decreases (curve 2); with an increase in the vapor pressure, the dimensions of the clusters increase (curve 3). The vaporizing points vary correspondingly.

Hence we have the following conclusions:

1. The Brownian motion of the crystal clusters of H_2O and the secondary structural framework formed by them are the main distinguishing feature of the liquid state and the structure of the water. If there is no movement of the clusters and their secondary structure, then there is no liquid state.

2. The approximate model of the structure of the water (Figure 2b) comprises six elements: H_2O molecules at the lattice nodes of the ice inside the clusters, the H_2O molecules completing the jumps, the vacancies at the lattice nodes of the ice inside the clusters, the H_2O clusters

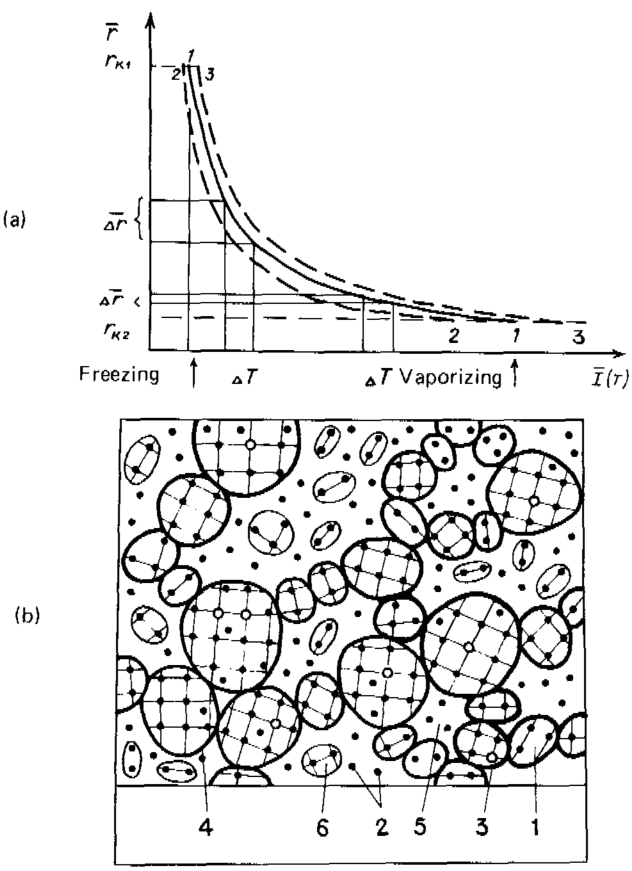

FIGURE 2 (a) Mean size $\bar{r}$ of the clusters [of Molecules] as a function of the temperature T and the kinetic energy of the pulses $I(T) = E_k$. Curve 1--for normal vapor pressures; curves 2 and 3, respectively--for reduced and increased vapor pressure. (b) Models of the water lattice. 1--H_2O molecules at the lattice nodes of the ice inside the clusters, 2--H_2O translating molecules, 3--vacancies in the ice lattice inside the clusters, 4--cluster in the amorphous structure of their framework, 5--cavities in the framework of the clusters, 6--clusters knocked out or shifted from the framework to the cavity.

in the amorphous secondary framework of the clusters, the cavities in the framework of the clusters, and the clusters knocked out of the framework or shifted by the pulses into the cavity of their secondary structure.

3. The water is crystallized by the structure inside the clusters and amorphous with respect to their secondary structure (Figure 2b) where $\bar{r}$ is the mean size of the elements of structure of the clusters which varies as a function of the temperature and vapor pressure from $\bar{r}_{k1}$ (freezing) to $\bar{r}_{k2}$ (boiling) (Figure 2a).

4. The water exists under the following conditions $E(T)_k > E(T)_{act}$;

$$P_{int} = P_{ext} \text{ and } \bar{r}_{k1} \geq r(n) \geq \bar{r}_{k2}.$$

If on cooling of the water $\bar{r}(n)$ becomes larger than $\bar{r}_{k1}$, the pulses cannot knock the clusters out of their framework, their motion ceases, and the liquid is converted into a solid state. If on heating the water $\bar{r}(n)$ becomes less than $\bar{r}_{k2}$, the velocities of the clusters increase so much that the bonds between them cannot hold them in the secondary structure. The latter decays, the water ceases to hold its volume, and boiling starts.

5. The evaporating water vapor contains monomers and small H_2O clusters. The vapor becomes purely monomeric only at the critical temperature.

6. The dimesions $\bar{r}(n)$ of the clusters are smaller and more stable (Figure 2a) at high T than at low T; therefore, the broken structures of the water (small $\bar{r}(n)$) can be retained for a significant time at reduced T. This explains the reduction in the anomalous properties of the superheated water with subsequent cooling (experiments of Letnikov, Klassen).

STRUCTURE AND PROPERTIES OF BOUND WATER

When the active positively hydrated surfaces are wet with water, the following structural changes take place in it (Figure 3a).

The wetting water goes into the field of adsorption forces of the surface f_1 directed opposite to the adsorption forces of the framework of the clusters j_{III}. The adsorption forces f_1 primarily liquefy the vapor at the surface, and at some distance from the surface, as a result of the oppositeness of the effect and compensation of the forces f_I and f_{III}, an intermediate zone is formed with minimum vapor pressure P, where P becomes less than in the free water.

Under the effect of the anisotropy and zonality P_{int}, the mean size of the clusters $r(n)$ and their secondary structure (Figure 3a, top) become anisotropic and zonal. Here, three zones (I, II, III) are formed. In zone I close to the surface, the heat of wetting is released (80-110 cal/g of bound water). This means that the water in zone I has again formed the bonds broken during melting, that is, it has gone into the solid state. This happens because the increase in vapor pressure in zone I increases the equilibrium mean size of the clusters to values of $\bar{r}(n) > \bar{z}_{k1}$ (Figure 2a), and it decreases their velocity so much that the translational motion stops and "hot ice" is formed. Therefore, the "wetting" is partially made up of "icing."

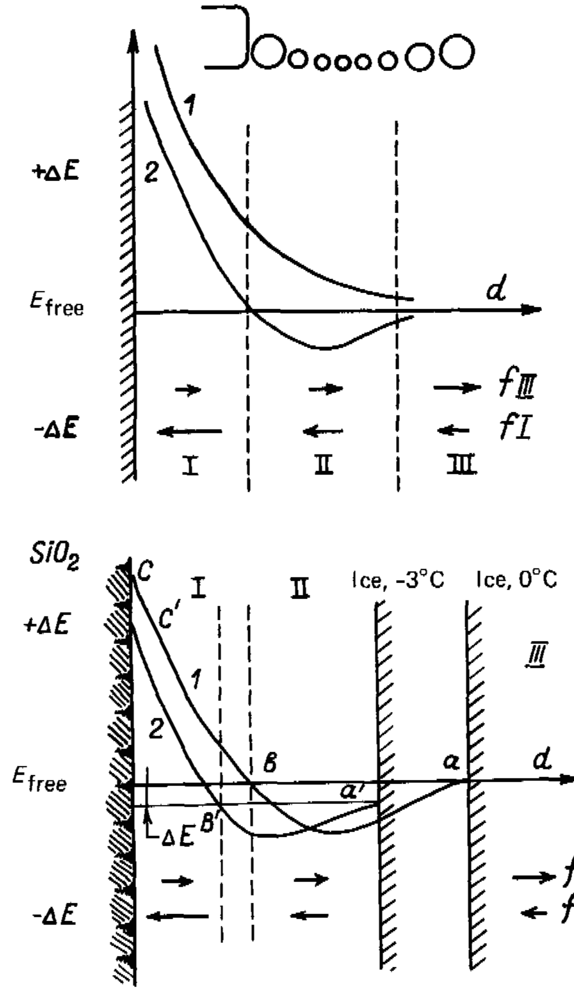

FIGURE 3 Variation of the activation energy $E(T)_{act}$ and the structure of the water with removal (d) from the surface. (a) E_{free} is the mean activation energy of free water; f_I is the adsorption force of the surface and the solid water in the zone I; f_{III} is the opposite force of the structure of the water in zone III. I--variation of the activation energy with only positive coupling; 2--variation of the same value with coupling-decoupling; I, II, III--zones of variation of the activation energies. Top--variation of the size of the clusters ($\bar{r}$). The freezing of the water between the ice and the SiO_2 surface. (b) I--activation energy in the bound (I), detached (II), and frozen water (III) at 0°C; 2--the same at -3°C; (a-B)--zone of detached unfrozen water at 0°C; a'-B'--the same at -3°C; (a-a')--freezing of the detached water; (B-B')--submelting of the bound water; ΔE--decrease in free energy at the surface of the ice (c-c')--the same at the surface; f_I and f_{III}--opposite adsorption forces of the SiO_2 and ice surfaces causing ΔE and (c-c').

In the intermediate zone II, a decrease in P causes the endothermal process of a decrease in sizes of the clusters $r(n)$ by comparison with the water and an increase in their mobility with the corresponding variation of all properties. This process proceeds as a result of a decrease in the surface energies of the active surface and the water during their interaction.

In zone III, the adsorption surface forces are insignificant; the water is almost free and retains its structure and properties. These structural changes during coupling of the water are illustrated in the upper part of Figure 3a.

An increase in mobility and a decrease in the activation energy in the intermediate zone of II by comparison with the free water must lower the temperatures of the phase transitions in it; increase the ion concentration, the heat capacity, the thermal conductivity, the dielectric constant; decrease the viscosity and the density; and correspondingly influence its physical properties and, via these properties, the physical properties of the en-tire dispersed wet system. At the present time the formation of intermediate mobile "detached" layers is confirmed by an ever increasing number of experiments.

Let us consider the results of some of these experiments.

REDUCTION OF THE PHASE TRANSITION TEMPERATURE OF BOUND WATER

The crystallization (boiling) point of a material is lower the lower the total energy of interaction between its particles and the activation energy, respectively (Figure 1a)--the experimentally detected reduction in the freezing or boiling point (with invariant external vapor pressure) directly indicates a decrease in activation energy and the formation of an intermediate mobile zone II (Figure 3a). It is widely known that wet dispersed media freeze in the negative temperature range.

The process and conditions of freezing of bound water between the surfaces, for example, SiO_2 (on the left) and ice (zone III on the right), is schematically depicted in Figure 3b. Here, the distances are plotted on the x-axis, and the activation energies, on the y-axis. Zone I is bound water, zone II is detached water at 0°C between points (a-B), and zone III is ice. At 0°C the ice is in equilibrium with the water under the following conditions:

$$I(T = 0°C) = \bar{E}(T = 0°C)_{act}, \quad (7)$$

$$P(0°C)_I = P(0°C)_W, \quad (8)$$

where P is the vapor pressure, and the symbols I and W denote ice and water.

In the case of cooling, for example, to -3°C, the following processes take place.

1. The energies $I(T)$ decrease; the activation energies $E(T)_{act}$ increase; the equilibrium conditions at the ice-water boundary (7) and (8) are disturbed, the ice beings to go from point a to point b. However, at the same time at the ice-water boundary, the force f_{III} remains constant and the force f_I increases inversely to some power of the distance between α and the ice. Therefore, the pulses in the ice-water direction increase, and in the opposite direction they decrease the more strongly the closer to the surface. Therefore, when the cooling stops, for example, at a temperature of -3°C, a new equilibrium is established which is expressed by the following conditions:

$$\bar{I}(-3°) + \Delta I = \bar{E}(-3)_{act},$$

which is equivalent to

$$I(-3°) = E(-3°)_{act} - \Delta E, \quad (9)$$

$$P(-3°)_I = P(-3°)_W. \quad (10)$$

From Figure 3b, we draw the following conclusions:

1. The exothermal freezing of the water in the (a-a') zone is accompanied by endothermal decoupling or suppression of part of the bound water between the points (B-B').
2. The zone of detached liquid water from the position (a-B) at 0°C moves with cooling to -3° and partial freezing to the position (a'-B').
3. With a further reduction in the temperature, the decoupling zone does not completely disappear, but becomes an intermediate zone of increased mobility of the elementary particles between the regions of two interacting structures. This explains why the melting of the crystals begins near the foreign inclusions or at the contact boundaries of single crystals.

REDUCTION OF THE BOILING POINT OF WATER IN WET CLAY (According to the Report by M. V. Voronova)

The results of the drying of clay at a constant temperature (T - 105°C) and simultaneous measurement of the sample temperature are presented in Figure 4, and they show that the drying curve (below) runs monotonically in accordance with the reduction in moisture and that the temperature rising to the point 1 (about 60°) then drops to the point 2 (about 50°) and then rises monotonically to T = 105°C.

The form of the thermograms of M. V. Voronova indicates the decoupling of the water in the sense the endothermal effect takes place at reduced temperature, that is, with reduced values of the pulses.

The thermogram with minimum temperature at point 2 indicates boiling with heating. It is possible to think that the pores in the clay fall into two groups: (1) the capillaries between the clay particles and (2) the micropores in the particles themselves. The coupling-decoupling process of the water in the micropores leads to condensation of the water near the surfaces and to expansion with a decrease in the vapor pres-

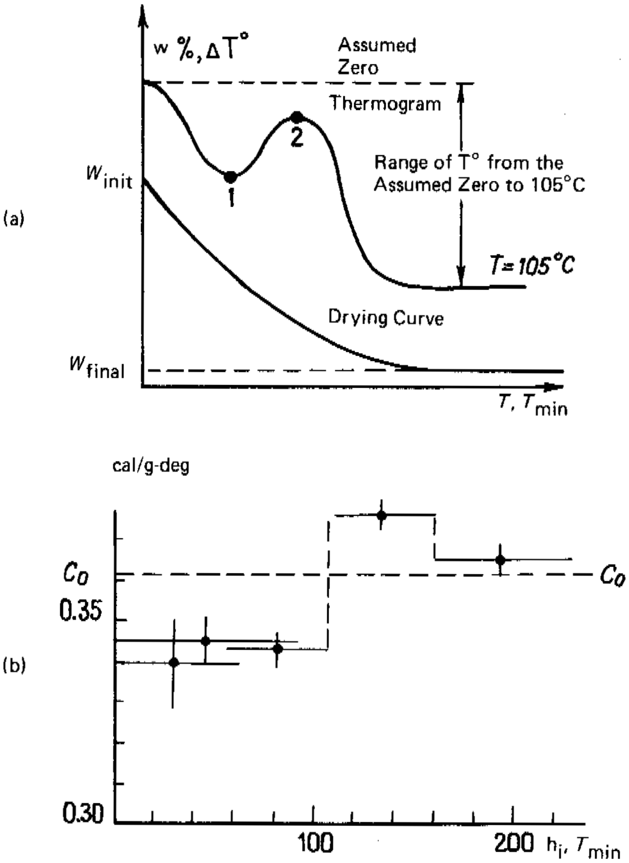

FIGURE 4 (a) Thermogram (A) and the curves (B) for drying of clay according to M. V. Voronova. (b) The specific heat capacity c of nitrobenzene in a disperse nitrobenzene-water-powdered glass system as a function of the distance h_i to the glass at $t = 20.2°C$, according to Yu. M. Popovskiy; line c_0--specific heat capacity of free nitrobenzene.

sure in the central parts of the capillaries and ultrapores. Here, the effect of the external vapor pressure is compensated for by the effect of the menisci and the layers of strongly bound water blocking up the more narrow capillaries. In the case of vaporization of the bound water at $T = 60°C$, the pressure in the capillaries increases and the vapor "explodes" into the outer space overcoming the external pressure. Then the temperature decreases to point 2, and the heating proceeds monotonically.

VARIATION OF THE SPECIFIC HEAT CAPACITY AND THERMAL CONDUCTIVITY OF THE ATTACHED LIQUID

The specific heat capacity (c) is made up of the part of the heat directly increasing the temperature or the kinetic energy c_K and part of the heat going to breaking or decreasing the bond energy c_B:

$$c = c_K + c_B. \tag{11}$$

With a decrease in the bonds and the activation energy, c_B must increase sharply, increasing the total value of c. Therefore, the values of the heat capacity of the bound water and, in general, the bound polar liquid must vary on removal from the binding surface opposite to the variation of the activation energy according to curve 2 in Figure 3a.

This result was obtained by Yu. M. Popovskiy* (Figure 4b). Comparing the curve in Figure 4b with curve 2 in Figure 3a, we see that they are the mirror image of each other, which must occur if the decoupling in fact exists.

In the case of bonding of the water to the surfaces, the thermal conductivity also increases. In the experiments with respect to heat and mass exchange in the drying processes, it is noted that the thermal conductivity of brick wetted by moisture is appreciably greater than the thermal conductivity of dry brick and moisture. This effect is naturally explained by the disproportionately large increase in mobility of the particles and the thermal conductivity of the liquid in the case of its decoupling as a result of the opposite effect of the walls of the pores on its structure.

The processes of coupling-decoupling of the water in the disperse capillary media and the corresponding variations of its structure explain many of the other "anomalous" variations in the properties of attached water from the same point of view.

CONCLUSIONS

1. The proposed trial approximate model of the strucutre of water explains: (a) the structural-genetic differences between the solid, liquid, and vapor states of the water; (b) the determinancy of the temperatures and heats of phased transitions; (c) the stability of the metastable states of the water.

2. From the model of the structure of free water, the model of the bound water and the representation of the process of bonding and release of the water and the formation of the intermediate "bound" layers are naturally obtained. These representations are confirmed experimentally and offer the possibility of predicting new effects.

3. The unfrozen water exists at negative temperatures because it is more mobile and is activated more easily than the free water. The solid phase of H_2O in fine-grained frozen soil is represented by two versions: ice-I, formed during freezing of portions of the free water, and "hot ice," formed during wetting of the surfaces and positive hydration of the ions in the case of a positive temperature. In the case

*Yu. M. Popovskiy, "Study of the Transition of the Boundary Phase to the Volumetric Liquid," *Issledovaniya v Oblasti Poverkhnostnykh sil* (Studies in the Field of Surface Forces), collection of reports of the Third Conference on Surface Forces, Moscow, Nauka, 148-153, 1967.

of freezing of the unfrozen water, the ice-I increases and the "hot ice" submelts. With an increase in T, the process goes in the opposite direction. Therefore, the heat of crystallization of the unfrozen water not only is viable but also differential.

4. With further development of the molecular-kinetic theory, the liquids in the water model, taking into account the existence of clusters and their secondary structure, must be considered as the most promising.

MIGRATION OF MOISTURE IN FINE-GRAINED SOIL OF DIFFERENT COMPOSITION, STRUCTURE, AND PROPERTIES

V. A. KUDRYAVTSEV, E. D. YERSHOV, AND V. G. CHEVEREV
Moscow Lomonosov State University

Many researchers have studied the laws of the migration of moisture in fog and frozen and freezing ground. The efforts of the geocryologists, thermophysicists, soil scientists, agrophysicists, and physical chemists concentrated in the same area have permitted the achievement of significant results in the development of the moisture-transfer theory in capillary-porous bodies.

In spite of this, there are a number of problems that are in need of a detailed study as applied to the peculiarities of the moisture-migration process in thawed, freezing, and frozen fine-grained soil. The most urgent of them are the following: the discovery of the relation between the moisture content and the moisture potential for different geological-genetic types of rock, the determination of the moisture-transfer coefficients of thawed and frozen ground, and the laws of variation of these coefficients as a function of composition; structure, nature of the cryogenous textures, and properties of the fine-grained soil; and a quantitative estimate of the role of the different gradients of the partial potentials of the moisture transfer during the moisture-migration processes in the thawed and frozen zones of the freezing ground.

At the present time, it is assumed that the basic motive force of the moisture transfer in the ground is the gradient of the free surface energy of the ground system not expended on interaction with the water. In practice, however, it is more expedient to operate, not with this variable, but with the gradient of the total thermodynamic potential of the soil moisture (grad μ_W)$\cdot\mu_W$. This is called the chemical or isobaric-isothermal potential and is represented in the form of the sum of a number of partial potentials: ionic, osmotic, gravitational, hydrostatic, electric, magnetic, and so on. The total potential of the ground moisture reflects the magnitude of the reduction of free energy of the water on its interaction with the solid state under specific thermodynamic conditions, and it is the work that must be done in order reversibly and isothermally to convert 1 g of ground moisture into volumetric (free) water. Both the total potential and its partial components are a function of temperature. The majority of researchers take the position of numerical equality of the potential μ_W and the ground moisture pressure. This permits experimental determination of the value of $\mu_W = f(v)$ when studying the dynamics of the moisture-transfer process in fine-grained soil.

As is known, the relation between the moisture potential and the moisture content of the ground in certain processes (such as, for example, the drying and wetting of the ground, sharp variation in density of the migration moisture flux, and so on) is ambiguous, that is, it is characterized by hysteresis. The basic cause of this phenomenon consists in different variations of the structure and physicomechanical properties of the soil during the occurrence of these processes.[8] As a result of such variations, the differential moisture content of the ground defined by the relation $(C_W = \partial w/\partial \mu_W)_T$ turned out to be different. This causes the possibility of the existence of different soil moisture with the same value of the moisture potential in it.

The experimentally established fact of the existence of hysteresis shows that the moisture-content gradient in the soil cannot be taken without stipulations with respect to the motive force of the moisture transfer. Consequently, when analyzing moisture migration in fine-grained soil, it is proper to use (in the case of the stationary regime) the expression

$$I_W = \lambda_W \cdot \text{grad}\ \mu_W, \tag{1}$$

where I_W is the density of the ground moisture flux; $\lambda_W^W = K_W \cdot C_W \cdot \gamma_d$ is the coefficient of moisture conductivity of the soil; K_W is the diffusion coefficient of the ground moisture; γ_d is the dry unit weight.

The use of the equation in moisture form

$$I_w = K_w \cdot \text{grad}\, W_{vol} \qquad (2)$$

is possible only in the case of an invariant differential moisture content of the ground. The above discussion also pertains to an analysis of the nonstationary regime of the moisture exchange described by the ordinary differential equation of ground moisture transferred.[11]

During freezing of the ground under natural conditions a number of processes occur that lead to significant ambiguous relations of μ_w and W. It is necessary to consider among these processes the processes of drying-wetting and heave settlement connected with this in the thawed zone of the frozen ground, variability with time of the migration moisture flux to the freezing front, increase in volume of the frozen soil as a result of the phase transformations of the ground moisture, and so on. The appearance and the role of hysteresis in the freezing and frozen soil have been investigated quite poorly. This complicates the discovery of the laws of migration of soil moisture and the application of the mathematical method of studying the process of freezing of fine-grained soils considering moisture migration.

The second, but no-less-important, problem when studying the moisture migration in freezing fine-grained soils turns out to be the development of the methods of determination and the laws of variation of the moisture-exchange characteristics of the thawed and frozen soil (λ_w, K_w, C_w) depending on the composition, structure, and properties of the soil and the water in them. Serious studies in this area have been carried out at the present time in the geocryology department of Moscow State University. They are studied on a specially created experimental setup permitting simultaneous experimentation with 10 soil samples under the conditions of open and closed hydrologic cycles in the presence and absence of temperature and gravitational field gradients in the soil samples. The laboratory method of determining the moisture-transfer coefficients is based on the principle of stationary moisture exchange in the soil.

The experimental setup that measures the parameters of the moisture-migration process in the soil is constructed on the basis of two refrigerators (Figure 1). One of them is designed to maintain the given temperature in the air conduit to the openings in which the investigated soil samples are connected. The regulation of the temperature and the velocity and moisture of the air flow in the air duct ensures the required intensity of evaporation of the moisture from the open surface of the samples. Moisture from a graduated capillary tube is fed to the opposite end of the sample through a sand interlayer. As the meniscus of the water in it moves, the moisture flux density through the transverse cross section of the sample I_w is recorded, and the time of the establishment of the stationary regime is determined. The presence of thermostats in both refrigerators permits assignment of both the isothermal conditions in the investigated samples (in the -20°C to +80°C range) and the temperature gradient. During the experiment using strain gages, the potential distribution of the ground moisture is recorded along the length of the sample. At the time of establishing the stationary regime, the moisture distribution and specific weight are determined, after which the moisture transfer coefficients are calculated by Formulas (1) and (2). The moisture-migration parameters experimentally obtained permit investigation of the laws of variation of the entire set of moisture exchange characteristics (λ_w, C_w, K_w, μ_w) as a function of the composition, structure, and properties of the thawed ground. By means of procedural ex-

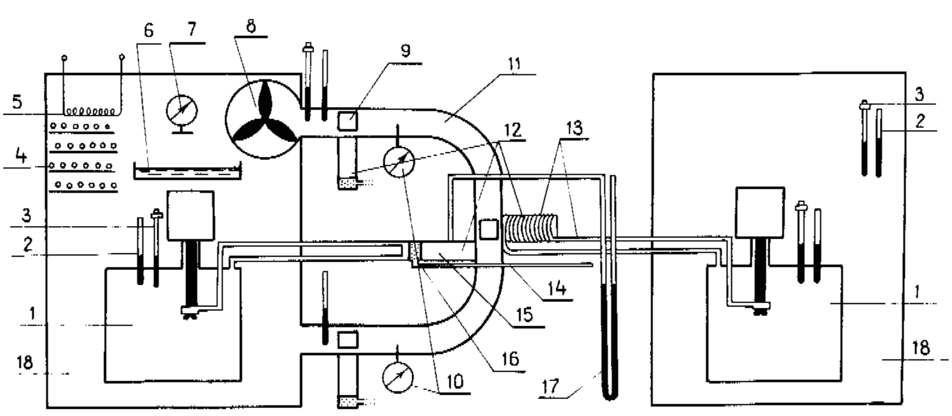

FIGURE 1 Schematic of the setup for determining the parameters of the moisture-transfer process in thawed, frozen, and freezing capillary-porous soils: 1--thermostat, 2--laboratory thermometer, 3--contact thermometer, 4--air dryer, 5--air heater, 6--air humidifier, 7--hygrometer, 8--fan, 9--window for attaching a mold with a sample, 10--anemometer, 11--air duct, 12--mold with the investigated soil samples, 13--hose with the thermostated liquid, 14--graduated capillary tube, 15--investigated soil samples, 16--sand, 17--strain gauge, and 18--refrigerator.

periments, the optimal dimensions of the tested soil samples were established, and high reliability of the results obtained by the stationary moisture exchange regime method was demonstrated.[5] The use of the laboratory method of determining the moisture-transfer coefficients as the standard led to the development of a procedure for approximate determination of these coefficients under field conditions (the field method).[10]

The laboratory study of the moisture-transfer process in different geological-genetic types of deposit permitted a sufficiently broad explanation of the dependence of the ground moisture diffusion coefficient (as the basic moisture transfer coefficient) of the composition, structure, and properties of the soil and the thermodynamic conditions of the environment. The experimental discovery of the dependence of K_W and I_W on each of the ground parameters was made with invariability of all other ground characteristics.

It was established that K_W is a function of the moisture content of the soil and can vary as a function of the moisture content by one or two orders. The curves for the moisture-transfer coefficient as a function of the moisture in individual sections only have an exponential nature; on the whole they can be represented by an n-degree polynomial. Their nature is predetermined by the differential porosity of the soil, and it is satisfactorily explained on the basis of analyzing the combined mechanism of moisture transfer in the soil.

The curve for the coefficient K_W as a function of the moisture in the first approximation can be divided into three sections (Figure 2, curve 3).* In the first section of curve 1, comparatively small values of the moisture-transfer coefficient of the investigated soils with respect to absolute magnitude are observed in a moisture range of less than MDC.† Such small values of the coefficient are characteristic for the film mechanism of moisture transfer in the soil when all of the moisture is distributed in the form of water films on the surfaces of the soil particles and frequently can fill only the non-moisture-conducting ultracapillary pores. The results of the studies made by Globus[2] also indicate an analogous mechanism of displacement of the moisture with a soil moisture content less than MDC.

The values of the coefficient K_W in the third section of the experimental curve are determined in practice wholly by the capillary mechanism of the moisture transfer in the soil. The great mobility of the capillary moisture by comparison with film moisture in the given case causes the existence of significantly larger absolute values of the moisture-transfer coefficient. The faster growth of K_W in the section is explained by the

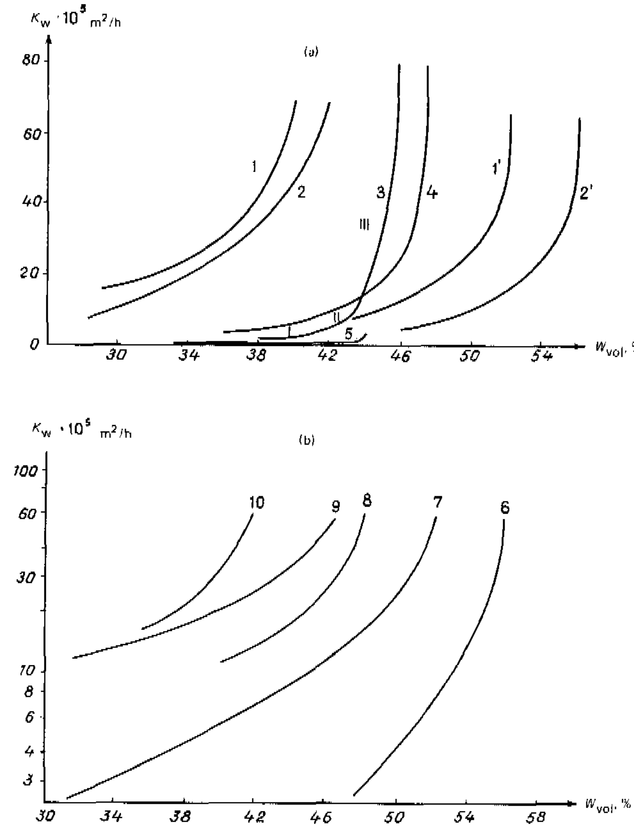

FIGURE 2 Moisture diffusion coefficient as a function of the volumetric moisture content of soils of different granulometric (b) and mineralogic (a) composition. Dudinskaya supes natural (1) and disturbed (1^1), Dudinskaya suglinok of natural (2) and disturbed (2^1), kaolin clay (3), hydromica clay (4), bentonite clay (5), Na-montmorillonitic clay (6), Moscow suglinok (7), Dudinskaya suglinok (8), loess-like suglinok (9), Tyumen' supes (10). I, II, III--sections of the curve $K_W = f(W)$ characterized by film, film-capillary and capillary mechanisms of moisture transfer in the soil respectively.

fact that in the given soil-moisture range small additions of moisture lead to intense formation of moisture-conducting capillaries.

The second section of curve II can be represented as the transitional section from the film mechanism of soil moisture migration to the capillary mechanism with an increase in moisture content.

The granulometric, microaggregate, and mineralogic composition, and also the composition of the exchange cations of the soil, determine to a significant degree the mobility of the moisture in the soil. Thus, with a fixed value of the moisture content, K_W decreases with an increase in the content of clay particles in the soil (Figure 2b) and minerals of the montmorillonite group (Figure 2a). In the region of film moisture transfer, this is connected with an increase of the active surface in a unit volume

*The properties of the investigated soils can be found in Yershov *et al.*[3] and Kudryavtsev and Yershov.[7]

†By MDC we mean the moisture for which discontinuity occurs in the free (capillary) moisture.[2]

of the soil (S_{vol}) and the peculiarities of the extensible crystal lattice of the minerals, which leads in the final analysis to a large binding energy of the film moisture. With a moisture content greater than the MDC when the capillary mechanism of transfer is basically in operation, a decrease in the coefficient K_w with an increase in the fineness of the soil and the mineral content in it of the montmorillonite group is explained by an increase in the porosity and the decrease in the mean effective radius of the pores in the ground, which leads to a decrease in the mean moisture-filled radius of the active capillaries.[9] The mobility of the capillary moisture in the ground is determined in the first approximation by the square of the radius of the active capillary. An analogous effect is also detected when replacing the multivalent exchange cations in the ground by univalent ones. The decrease in the values of K_w and I_w fixed in this way eventually changes the nature of the ice composition in the soil.[13]

The peculiarities of the composition and structure of fine-grained soils have no less significant effect on the moisture-migration process. Thus, in undisturbed soil in comparison

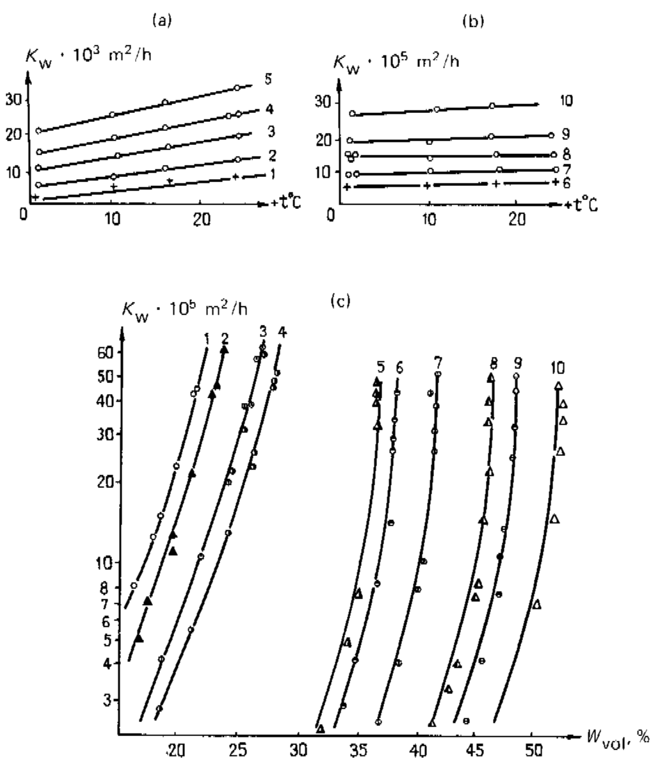

FIGURE 3 Moisture diffusion coefficient as a function of temperature. (a) Moscow supes (for W_{vol} equal to: 1-16, 2-18, 3-20, 4-21, and 5-22 percent). (b) Kiev clay (for W_{vol} equal to: 6-41, 7-42, 8-43, 9-44, and 10-45 percent). (c) The dry unit weight and moisture content of the Moscow supes (for γ_d equal to: 1--1.98, 2--1.93, 3--1.82, and 4--1.77 g/cm) and Kiev clay (for γ_d equal to: 5--1.69, 6--1.65, 7--1.55, 8--1.41, 9--1.35, and 10--1.25 g/cm).

with disturbed soil, as a rule the values of K_w turn out to be large (Figure 2a). This is caused by a smaller magnitude of S_{vol}, a sharply nonuniform pore-space of the undisturbed soil, and also an increased content in the soils of larger pores-capillaries. The above-noted law is evident in soils of different composition but with identical total porosity.

If the soil porosity decreases, the values of K_w increase with constant moisture content (Figure 3c). The discovered law agrees well with modern concepts of the nature of variation of the structure of the pore-space on compaction of the soil, and it is explained by an increase in the thickness of the water films and intensification of the role of the capillary mechanism of the moisture transfer as a result of the formation of an additional number of moisture-filled capillaries.[4] An analysis of the experimental data with respect to the effect of the ground density (porosity) on the variation of K_w permits proposal of a method of calculating the function $K_w = f(\gamma_d)$ in the presence of a "standard" curve $K_w^e = f(W)$ for the investigated soil of known density (γ_d^e). The method is based on the fact that the curves $K_w = f(W)$ are similar for soil with different dry unit weight, and they are shifted one relative to the other with respect to the x-axis by the amount $W_k^x - W_k^e$ (Figure 3c). The values of W_k^x and W_k^e are the values of the capillary moisture capacity of the soil with a density γ_d^x and γ_d^e. The calculation formula has the form:

$$W\Big|_{K_w^e}^{x} = W\Big|_{K_w^e}^{e} + \left(\gamma_d^e - \gamma_d^x\right)\tan\alpha,$$

where

$$\tan\alpha = \frac{W_k^x - W_k^e}{\gamma_d^e - \gamma_d^x}$$

In Figure 3c we have the experimental points, and the curves are constructed using Formula (3).

The water-conducting properties of the soil are determined by the composition, structure, and properties, not only of the grain-size, but also the migrating moisture itself and, primarily, the concentration of salts dissolved in it and the variation of the energy state of the soil water under the effect of the temperature, magnetic, electric, and other fields. However, these problems have at the present time been investigated far from adequately.

The set of studies performed under the conditions of isothermal moisture regimes have demonstrated that the parameters K_w and I_w Increased directly in proportional to an increase in the soil temperature. This relation must be determined by the linear decrease in the magnitude of the surface tension of the soil moisture causing its mobility with an increase in temperature in the 0°-30°C range. The most significant effect of the temperature on the variation of K_w is exhibited in soils with a large pore radius. Thus, in the supesses the variation of K_w is 3-5 percent, and in clay it is 0.5 percent per

°C (Figure 3a,b). The experimentally discovered nature of the function $K_w = f(t)$ permits the proposal of the following formula in order to calculate $K_w = f(W,t)$:

$$K_w = K'_w + (t' - t) \tan \alpha,$$

where K'_w is the moisture diffusion coefficient taken with the known curve for the relation for K_w as a function of W at a temperature t'. In order to determine the angle α, the function $K_w = f(W)$ obtained for two temperatures of the investigated soil is used.

An analysis of the experimentally obtained values of the diffusion coefficient of the soil moisture for various geological-genetic types of rock demonstrated that the variation of these characteristics, both as a function of moisture content, composition, and structure, and as a function of a number of the soil properties can reach two and more orders. Such a high range of variation of K_w does not permit use of a constant value of the coefficient for the solution of the joint problems of heat and mass transfer in the present soils, but requires knowledge of the specific function $K_w = f(W, \gamma_d)$ for the given type of ground under similar thermodynamic conditions of the environment.[12] Therefore, one of the basic problems in the field of studying the moisture-exchange characteristic of thawed ground today is the development of approximate and exact methods of calculating these characteristics by the known water-physical properties of fine-grained soils. The calculation methods must obviously be based on analyzing the energy state of the moisture in the soil determined by the differential porosity of the soil, the volumetric active surface, the nature and the crystal lattice of the minerals, the composition of the exchange cations, the thickness of the water film, the size and number of the waterfilled (active) capillaries in the ground, etc.

The coefficients of the moisture transfer in frozen soil have been poorly studied up to the present time. The individual studies performed in this area by N. S. Ivanov, L. V. Chistotinov, V. A. Kudryavtsev, E. D. Yershov, V. V. Gurov, and so on, are still preliminary, indicating, however, far from a secondary role of the moisture migration in the frozen zone of the soil during freezing and frost drying. Such results indicate the necessity for serious study of the diffusion coefficient, the moisture conductivity, the differential moisture content, and the moisture-transfer potentials of the frozen ground as a function of their composition, cryogenous structure, and properties.

The third area in the field of study of the moisture-migration law in thawed, freezing, and frozen fine-grained soil is the discovery in quantitative respects of the role of the various gradients of the partial potentials of the moisture transfer in the total migration flux. It is necessary to note that for frozen soil at this level there is an extremely limited number of experimental papers[1,6,13,15] that, actually, only state the problem. As applied to thawed ground, there are an entire series of interesting laboratory studies that permitted the discovery of the dependence of the density of the migration flux on the magnitude of the gradients of the various partial moisture potential. However, only in individual cases were these results tied to the potential gradients of the moisture transfer observed under natural conditions.

Thus, for example, the theoretical estimate of the role of the temperature gradient in the moisture transfer in thawed fine-grained soil was made comparatively long ago.[11,14] However, the experimental test of this effect for the case of freezing ground was not carried out. Accordingly, at the geocryology department, special experiments were set up for a comparison of the migration-moisture flux in samples of clay, suglinok, and supes for the cases of gradient and gradientless temperature fields in them. The results of the experiments demonstrated that the existence of the temperature gradient on the order of 2-3 deg/cm does not lead to a noticeable increase in the flux-density of the migrating moisture in the soil. However, at a temperature gradient of 4-5 deg/cm, the density of the migration flux increases by 10-20 percent in comparison with the isothermal moisture-transfer process.

According to numerous observations,[14] the natural temperature gradient in the thawed zone of the soil near the freezing front does not exceed 0.5 deg/cm. Consequently, the magnitude of the thermogradient moisture flux to the ice deposition front when solving the problem of seasonal and perennial freezing of the ground can be neglected. In the case of artificial freezing of the fine-grained soil characterized by high thermal potential, the differential approach to estimating the magnitude of the moisture flux is necessary.

The consideration of gravitational forces is no less important when calculating the migration of the soil moisture to the freezing front. In this case the moisture migration is realized along the gravitational force sector and the moisture-transfer potential under the equal conditions must be represented by the sum of two partial potentials: the capillary (Ψ_k) and the gravitational (Ψ_z).

For film moisture migration when the soil moisture is less than the *MDC*, the effect of the gravitational force on the moisture is very low since $\Psi_k >> \Psi_z$. In the case of the capillary mechanism of moisture transfer ($W > MDC$), the moisture flux density caused by the gravitational gradient will be determined by the radius of the capillaries of the ground (r_{ave}) and the height of the soil mass (l). The laboratory studies that we performed on clay and supes soils ($r_{ave} < 0.5 \cdot 10^{-3}$ cm and l<15 cm) demonstrated the absence of such noticeable difference in the density of the migration-moisture flux for vertical and horizontal arrangement of the specimens. An analogous result (on the basis of the theoretical arguments) was obtained by A. V. Lyukovov, who demonstrated that in the moisture of a porous body about 10 cm high, the effect of the gravitational force can be neglected, with an error of approximately 6 percent, if the radius of the capillary of this body is less than $1 \cdot 10^{-3}$ cm. If the body dimensions increase by 10 times (l = 1m), then the dimensions of the capillary pores are not subject to a significant effect of the gravitational field ($r < 10^{-4}$ cm), decrease correspondingly, and so on.

When estimating the effect of the gravitational force on the heat-transfer process in a soil massif, it is possible to use as a basis the first approximation in the above-presented relation for r_{ave} and l, assuming that r_{ave} is the mean effective radius of the soil capillaries.

The effect of the natural gradients of the electric and magnetic fields and also the osmotic gradient and the soil density gradient for moisture transfer have been poorly studied. However, laboratory efforts in this direction are permitting the statement of the significant role in the effect of the noted factors on the process of migration of moisture in the ground and intensifying this effect with a reduction in temperature and moisture content (Panchenko, 1972; Taylor and Carey, 1960; and others).

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PROBLEMS IN THE THERMODYNAMICS OF FROZEN DEPOSITS

B. A. SAVEL'YEV Moscow Lomonosov State University

In closed dispersed systems during the freezing of which there is no removal or addition of material, the possible variations are defined by the complete differential of the thermodynamic potential:

$$d\phi\left(P,T\sum_{i=1}^{K} x_i^{t}\,(w-1),\ \sum_{i=1}^{v_1} x_i^{zh},\ l,\ \sum_{i=1}^{v_2} x_i^{l},\ \sum_{n=1}^{m} x_i^{z},\ S_k\right) = \frac{\partial\phi}{\partial P}\,dP + \frac{\partial\phi}{\partial T}\,dT$$
$$+ \sum_{i=1}^{k}\frac{\partial\phi}{\partial x_i^{T}}\,dx_i^{T} + \frac{\partial\phi}{\partial(w-1)}\,d(w-1)$$
$$+ \sum_{i=1}^{v_1}\frac{\partial\phi}{\partial x_i^{zh}}\,dx_i^{zh} + \sum_{i=1}^{v_2}\frac{\partial\phi}{\partial x_i^{l}}\,dx_i^{l} + \frac{\partial\phi}{\partial l}\,dl$$
$$+ \sum_{i=1}^{m}\frac{\partial\phi}{\partial x_i^{z}}\,dx_i^{z} + \frac{\partial\phi}{\partial S_k}\,dS_k \cdots \qquad (1)$$

Here x_i^T are the molar concentrations of the solid components of the organic-mineral mass; x_i^{zh} is the concentration of the components of the dissolved materials; x_i^l is the ice content in nature;* x^z is the concentration of the gas components, P is the pressure; T is temperature; w is the moisture content; l is the ice content, S_k is the surface area; $\partial\phi/\partial S_k$ characterizes the thermodynamic potential, which depends on the surface coefficients;

$$\frac{\partial\Phi}{\partial x_i^{T}},\quad \frac{\partial\Phi}{\partial(w-l)},\quad \frac{\partial\Phi}{\partial l},\quad \frac{\partial\Phi}{\partial x_i^{z}},\quad \frac{\partial\Phi}{\partial t_i^{l}},\quad \frac{\partial\Phi}{\partial S_k}$$

are the partial derivatives of the complete thermodynamic potential.

Inasmuch as the ice inclusions and unfrozen water are phases of the H_2O component determining the structure, composition, and the entire set of properties of the frozen soil, before discussing the conditions of phase equilibrium it is necessary to consider the physical essence of the process of melting of the ice, considering that melting and crystallization processes are ideally reversible. Beginning with the established principles that the structure of the ice lattice is molecular and a water molecule has a clearly expressed dipole, it is possible to imagine that the melting process is determined by the effect of the forces disturbing the attraction of the molecules in the ice lattice. The relation between the molecules is caused in turn by the spacing between them. The expression for the potential energy of two interacting dipoles in the lattice will have the form:

$$\Phi = -\frac{p^2}{r^3} + \frac{\beta}{r^n}, \qquad (2)$$

where p is the magnitude of the electric moment of the dipole, r is the distance between two interacting molecules, and β/r^n reflects the attenuating effect from the point of view of the other molecules of the lattice on the interaction of the two investigated dipoles.

The force of interaction between the molecules F is, by differentiation:

$$F = \frac{\partial\Phi}{\partial r} = \frac{3p^2}{r^4} - \frac{n\beta}{r^{n+1}}. \qquad (3)$$

For a nonzero temperature, the mean distance between centers of the oscillating particles will differ from the spacing between the particles at absolute zero. Let us stipulate that the thermal expansion of the crystal is proportional to the temperature. Then it is possible to apply the expression defining the expansion coefficient:

*The value of x_i^l is variable. It depends on the thermodynamic and physicochemical conditions and on the presence of liquid and gas inclusions and at the same time it is the primary factor determining the texture of frozen soil and its properties.

$$\alpha = \frac{1}{r_0} \cdot \frac{\partial r}{\partial T} = \text{const}, \quad (4)$$

where r_0 is the minimum distance between the centers of the two closest molecules, and T is the absolute temperature. The spacing between the two molecules as a function of temperature is represented by the following expression:

$$r = r_0 \ (1 + \alpha T). \quad (5)$$

Using Equation (5), let us make the substitution in the same (3); then we obtain the expression defining the force of interaction between two molecules as a function of temperature T:

$$F = \frac{\partial \phi}{\partial r} = \frac{3p^2}{r^4{}_0 (1 + \alpha T)^4} - \frac{n\beta}{r_0{}^{n+1} (1 + \alpha T)^{n+1}}. \quad (6)$$

When $T = 0$, the particles are in equilibrium; the resultant force is 0, and it is possible to write:

$$F_0 = \frac{\partial \phi}{\partial r} = 0,$$

$$\frac{3p^2}{r_0{}^4} = \frac{n\beta}{r_0{}^{n+1}}. \quad (7)$$

Then (6) considering (7) assumes the form:

$$F = \frac{\partial \phi}{\partial r} = \frac{3p^2}{r_0{}^4} \left[\frac{1}{(1 + \alpha T)^4} - \frac{1}{(1 + \alpha T)^{n+1}} \right]. \quad (8)$$

Expression (8) for the force of interaction between two particles as a function of the distance between can be represented graphically. The behavior of the curve shows that within the temperature range from $T = 0$ to T_s corresponding to the spacing between the molecules r_s, the curve is rising. Within these limits the forces of attraction increase with an increase in temperature. In the 0 to T_s temperature range, the ice has the quasielastic property of a solid. Transition through the maximum, together with an increase in temperature, decreases the attractive forces; here the properties of the solid state are lost. Obviously, T_s can be taken as the melting point, that is, the phase transition point. In order to determine the melting point, it is necessary to establish the position of the maximum:

$$\frac{\partial F}{\partial T_s} = \frac{\partial^2 \phi}{\partial r \, \partial T_s} = 0. \quad (9)$$

Then Equation (8) assumes the form

$$\frac{n + 1}{(1 + \alpha T_s)^{n+2}} - \frac{4}{(1 + \alpha T_s)^5} = 0, \quad (10)$$

transforming which, we obtain

$$4 \ (1 + \alpha T_s)^{n+2} = (n + 1) \ (1 + \alpha T_s)^5$$

or

$$(1 + \alpha T_s)^{n-3} = \frac{n + 1}{4}. \quad (11)$$

In the final form,

$$\alpha T_s = \sqrt[n-3]{\frac{n + 1}{4}} - 1,$$

where n is the structural parameter of the ice lattice reflecting the packing effect. Using (11), it is possible to define the value of n by setting $T_s = 273°$ and $\alpha = 0.000169/[\text{deg}]$.

The processes of the occurrence of ice from a liquid or vapor medium and its deterioration are completely reversible. However, the intensity of the phase transformation depends on the geometry of the body itself. The molecules located at the vertices have the highest free energy; the molecules located on the edges have somewhat less, and there is still less for molecules on the face. The molecules separated in the thawing process from the face of the crystal can move predominantly in a direction perpendicular to the plane. The molecules torn away near the nodes and the edges of the polyhedron have the possibility of moving at different angles to the crystal lattice. Thus, the molecules located on the vertices and edges of the crystal have a larger number of degrees of freedom than the molecules located on the faces.

At 0°C the free water has an energy content equal to 152 cal/g, whereas ice has 72.32 cal/g. The difference in the energy between them is 79.68 cal/g, which is the latent heat of fusion. In other words, this is the work of formation of the ice lattice caused by the change in spacing on going from 3.00 Å between the water molecules to shifting of the molecules in the ice lattice to spacings of 2.76 Å. In the given case for water, a spacing between the instantaneously attached molecules is assumed, and the molecules attached at the nodes are not taken into account (the translational displacements). The melting of the ice leads to almost doubling of the internal energy of the molecules. The sublimation of the ice increases the energy of its molecules by 10.6 times (the heat content for vapor at a temperature of 0° and a pressure of 760 mm Hg is 767 cal/g). The free energy of the ice and the free energy of water at 0° are equal, but with respect to the entire internal energy its percentage will not be equal. For ice the free energy is 92 percent of the internal energy, but for water it is only 44 percent.

The basic thermodynamic condition of phase equilibrium in the water-ice system is equality of the temperatures, pressures, and chemical potentials of the liquid and solid phases in the soil. The relation between the temperature, pressure, and the ratio of the water and ice is defined by the Clapeyron-Clausius expression:

$$\frac{\partial P}{\partial T} = \frac{\lambda_I}{T\,(V_w - V_I)}, \qquad (12)$$

where λ_I is the latent heat of fusion of 1 g of ice and V_w and V_I are the specific volumes of the water and ice.

Analogously for the process of evaporation and condensation

$$\frac{\partial P}{\partial T} = \frac{\lambda_w}{T(V_w - V_v)}, \qquad (13)$$

where λ_w is the heat of evaporation and condensation of 1 g of water and V_v is the specific volume of the vapor.

For the process of sublimation:

$$\frac{\partial P}{\partial T} = \frac{\lambda_s}{T(V_v - V_I)}, \qquad (14)$$

where λ_s is the heat of sublimation of 1 g of ice.

The presence of water in a deposit always leads to solution of some of the material in it. Generally, the liquid phase of the deposit is weak solutions. The mean concentration of the natural pore solutions is: for limestones, 0.36 g/l; marls, 0.25 g/l; sand, 0.2 g/l; clay, 0.16 g/l; anhydrides 0.14 g/l. However, rock saturated with highly mineralized pore solutions does occur. From weak molecular binary solutions, the chemical potentials of the solvent (μ_{ws}) and the dissolved material (μ_x) are, respectively,

$$\mu_{ws} = \mu_w - RT_x,$$

$$\mu_x = RT \ln x + \phi\,(P,T), \qquad (15)$$

where $\mu_w = (\partial\mu/\partial m)_{S,V}$ is the chemical potential of the free water; there is an increment in the internal energy (u) with an increase in mass (m) by one if the volume (v) and the entropy (s) are constant, and x is the concentration of the dissolved material; $\Psi(P,T)$ is the function of pressure and temperature; R is the universal gas constant. With an increase in concentration of the weak solutions of the dispersed material, their freezing point drops as defined by a function:

$$\Delta T = \frac{kT_0^{\,2}}{\lambda I}\,(x_w - x_I), \qquad (16)$$

where k is the Boltzmann constant, T_0 is the freezing point of pure water, and x_w and x_I are the amounts of water and ice expressed in relative values.

For more concentrated pore solutions, the increase in temperature is defined by a more complex expression:

$$\Delta T = \frac{kT_0^{\,2} \ln\,(1 - x)}{kT_0 \ln\,(1 - x) - \lambda_I}. \qquad (17)$$

In contrast to the molecular solutions, the variation in concentration of the aqueous solutions of electrolytes is especially felt in the freezing point. In such cases, instead of the concentration, the factor of activity is introduced by means of which the effect of the ions on the thermodynamic properties of the material is taken into account. Robinson and Stokes[1] established the relation between the activity A and a reduction in the freezing point ΔT:

$$-\ln A = 0.0004207\;\Delta\,T + 2.1\cdot 10^{-6}\;\Delta T^2, \qquad (18)$$

where $\ln A = Q_\phi^{\,2}\xi/6EkT$, Q_ϕ is the charge on the ions, $1/\xi = 3.06\cdot 10^{-8}\sqrt{v}$, v is the volume of the solution of 1 mole of electrolyte in liters, and $E = 81.79$ is the dielectric constant of the solvent (water).

For ionic pore solutions, it is possible to express the values of the chemical potentials of the solvent (μ_w) and the dissolved materials (μ_{xs}) by the equation:

$$\left.\begin{aligned} \mu_{wm} &= \mu_w - xkT + \frac{1}{3}\sqrt{\frac{8\pi}{E3}\cdot\frac{\rho_w Q_\phi 6}{kT}} \\ \mu_{xm} &= kT \ln x + \phi(P,T) - \sqrt{\frac{8\pi}{E3}\cdot\frac{\rho_w Q_\phi 6}{kT}} \end{aligned}\right\}, \qquad (19)$$

where ρ_w is the density of the water.

N. S. Ivanov[2] expressed the relation between the temperature, the pressure, and the concentration of the ionic pore solution by an equation of this type:

$$\frac{\lambda_{Im}}{T}\,dT + (v_{wm} - v_{Im})\,dP = d\left[\left(x_w - x_I\right)kT - \frac{1}{3}\sqrt{\frac{8\pi Q_\phi 6}{E3\;T}}\left(\sqrt{\rho_w x_w 3} - \sqrt{\rho_I x_I 3}\right)\right], \qquad (20)$$

where λ_{Im} is the specific heat of melting in the rock, v_{wm} and v_{Im} are the specific volumes of the water and ice in the rock, x_w and x_I are the water and ice concentrations, and ρ_I is the ice density.

The chemical potential of the bound water, μ_{wb}, is defined by the expression:

$$\mu_{wb} = \mu_w + u_c, \qquad (21)$$

where μ_w is the chemical potential of the free water, and u_c is the potential energy of the water molecules in the force field of surface particles.

If the chemical reaction takes place in the investigated medium, then the chemical potential that characterizes the energy of the system on variation of its composition is applied to the thermodynamic potential. The magnitude of the chemical potential can be written in the form of this expression:

$$du = \sum\left(\frac{\partial u}{\partial n_i}\right)_{v,S,\,n_j} dn_i + \left(\frac{\partial u}{\partial S}\right)_{v,N} dS + \left(\frac{\partial u}{\partial v}\right)_{S,N} dv, \qquad (22)$$

where $(\partial u/\partial n_i)_{v,S,n_j}$ is the increment of the internal energy of the given phase of the system with an increase in mass of material per unit at constant volume v, entropy S, and mass n; N is the number of molecules of all the material; n_j is the number of moles of all material except the number i.

On displacement of the phases with respect to each other as a result of the presence of the double electric layer, an electrokinetic potential arises that will be exhibited in the process of migration of the liquid near the stationary solid phase during electroosmosis, during settling of solid particles in the liquid, and certain other processes. Thanks to the defined dependence of the electrokinetic potential on the grain size of the mineral particles, the composition of the cations of the diffused layer, and the degree of hydration of the surface, the electrokinetic potential can serve as the determinant, not only of the migration of material, but also variation of time of the physiochemical processes and the structure of the double electric layer. The electrokinetic potential is always less than the thermodynamic potential. With an increase in the surface of the soil, its electrokinetic potential increases as a result of an increase in the number of ions on the surface of the particles. It is defined by an expression of the type:

$$\Phi_{ek} = \frac{kV\eta}{E\varepsilon}, \tag{23}$$

where V is the linear rate of displacement of one phase with respect to the other, η is the viscosity of the liquid, E is the voltage of the electric field, ε is the dielectric constant, k is the coefficient reflecting the contact of one phase with the other, and Φ_{ek} is the electrokinetic potential.

The total internal energy U and the entropy S of the system comprising the liquid phase of the water, the ice, the attached water, and the mineral part of the frozen soil is defined as the sum of the energies and entropies of each phase of H_2O and the energy and entropy of the mineral skeleton (not differentiating the latter into its components):

$$U = U_w + U_I + U_C + U_G,$$

$$S = S_w + S_I + S_C + S_G, \tag{24}$$

where the indexes w, I, C, and G indicate the medium, that is, free water, ice, bound water, and soil, respectively.

The simplified version of the phase state was derived by N. S. Ivanov[1] from the Clapeyron-Clausius equation considering the empirical relations between the compression and the residual moisture:

$$w(T) = w(T_0) - C_w \ln \frac{\lambda_I}{\Delta v P_0 T_0} \Delta T, \tag{25}$$

where w is moisture content, Δv is the increment in volume, P_0 is the normal pressure, T_0 is the temperature of the beginning of freezing, C_w is an experimental parameter, and ΔT is a temperature increment.

The molar thermodynamic potential of the attached water is defined by the equation (1):

$$\Phi_{wc}(P_0T) = \Phi_w(P_0T) + v_w e^{\frac{w_0 - w}{C_w}} \left[1 + 0.5k_p \, C\, \frac{2w_0 - w}{Cw}\right], \tag{26}$$

where Φ_w is the molar potential of the free water, v_w is the volume of the free water, w_0 is the amount of bound water at the pressure P_0, and k_p is the compressibility coefficient.

Equations (25) and (26) permit determination of the composition of the pore moisture, that is, the bound water and free water moving into the ice and they make it possible to obtain the characteristics of the thermodynamic potential of the bound water.

In the thermodynamics of such a complex system as frozen soil, the energetics of the surface layers has great significance. Obviously it is possible to refer the adsorbed water to the surface layers that is stored at negative temperatures as a result of a reduction in its vapor pressure to the vapor pressure of ice by surface binding. On the basis of these principles, each negative temperature for the given adsorbent corresponds to a defined liquid phase content in the equilibrium with the ice. If the adsorbent is not dissolved in the adsorbed phase, then a clear interface arises between them. The thickness of the surface layer can be determined from the expression describing the variation of the density ρ of the liquid as a function of the distance h to the boundary surface:[3]

$$\rho = \rho_0 + \frac{\pi\rho_0{}^2 x_0 (a'\ \rho' - a\rho)}{6} \cdot \frac{1}{h^3}, \tag{27}$$

where ρ_0 is the density of the liquid; x_0 is the isothermal compressibility of the liquid; a is the van-der-Waals' constant of the interaction of the liquid molecules with each other; a' is the constant for the interaction of the liquid molecules with the molecules of another phase, the value of which for water is estimated by the heat of wetting and is 10.5 kcal/mol or 118 erg/cm^2; and ρ' is the density of the other phase. Analogously to Equation (27), it is possible to compile new expressions using the laws of variation in the surface layer of other properties, for example, the variation of the pressure tensor, and the dielectric properties with the chemical potentials; here it is important that the variations undergo a jump over the boundary surface. For calculating the constant a from expression (27), the following equation is used:

$$a = \frac{27R^2\ T^2{}_K}{64P_K},$$

where $R = 8.3147°\cdot 10^7$ erg/deg-mole is the gas constant, T_K is the critical temperature for which the gas can still be converted to a liquid, P_K is the kinetic pressure required for conversion of the gas into liquid at the critical temperature.

By using the results of determining the density of the adsorbed liquid ρ by Equation (27), it is possible to calculate the value of the tangential component of the stress tensor:

$$\tau = P + \frac{\pi \rho_0 (a'\rho' - a\rho_0)}{8} \cdot \frac{1}{h^3} \qquad (28)$$

For a uniform liquid in the absence of deviation from Pascal's law, $\tau = P_N = P$, where τ is the tangential component of the stress, and P_N is the principal normal stress. The value of h entering into Equations (27) and (28) can be established from the direct measurement of the amount of adsorbed liquid by a foreign body[4] and the determinations of the specific surface of the adsorbents.[5]

For the two-dimensional surface layer or the two-phase system in the presence of the two-dimensional interface, the following generalized equation is used:

$$Ad\sigma = -SdT + vdP_N - \sum_i n_i du_i, \qquad (29)$$

where A is the area of the interface, and its increment is determined from the expression

$$\Delta A = \frac{\Delta w + P_N \Delta v}{\sigma},$$

v is the volume of the system; Δw is the work of deformation, $\Delta w = -\iiint_v dv\, \Sigma_{ik} P_{ik} \Delta\varepsilon_{ik}$; P_{ik} and ε_{ik} are the tensor components of stress and strain; $\sigma = \int_0^h (P_N - \tau)\, dz$ is the [surface] tension in the layer; n is the mass (the number of moles); and μ is the total chemical potential.

The increment of the total energy of the system dU will be defined by an equation of the type:

$$dU = TdS + dw + \sum_j \bar{P}_j d\bar{z}_j + \sum_i \bar{\mu}_i dn_i, \qquad (30)$$

where $\Sigma_j \bar{P}_j dz_j$ is the work done on the system by external forces, $\bar{P}_j$ is the generalized external force, and $\bar{z}_j$ is the generalized external coordinate.

A group of basic thermodynamic potentials for a flat superficial layer or two-phase system with a flat interface characteristic of frozen soil can be represented by the following expressions:

$$F = -P_N v + \sigma A + \sum_i \mu_i n_i,$$

$$dF = -SdT - P_N dv + \sigma dA + \sum_i \mu_i dn_i,$$

$$H = U + P_N v - \sigma A = TS + \sum_i \mu_i n_i,$$

$$dH = TS + vdP_N + Ad\sigma + \sum_i \mu_i dn_i,$$

$$\sum_i \mu_i n_i = U - TS + P_N v - \sigma A, \qquad (31)$$

where F is the free energy, and H is the enthalpy.

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NEW CONCEPTS OF THE NATURE OF FROZEN SOILS

I. S. TYUTYUNOV Scientific Research Institute of Foundations and Underground Structures

The ground as the foundation for various structures is formed as a result of the interaction of the solid material of the earth with water. In the opinion of V. I. Vernadskiy, there is no natural body which could be compared with water "with respect to effect on the course of the basic, grandest geological processes."

In addition, all these phenomena and processes turn out to be possible and take place on the basis of the fact that the interacting free water is subject to significant variations and is converted into bound water. The bound water is the most active converter of rocks and minerals in physical and chemical respects.[1] This most important property and effect are not attenuated in the frozen part of the lithosphere,[2] converting its other part into monolithic frozen ground. By changing the rock and introducing its generating contribution in the formation of the ground, the attached water is, nevertheless, obligated to the solid material of the earth for its properties, state, and enormous reserve of free energy.

It is quite obvious that in order essentially to improve the properties of the ground and, in turn, increase its strength it is necessary to investigate the nature of the bond of the water by the solid particles and to discover the physics of the conversion of these solid particles into a monolithic natural body.

There are numerous papers on bound water. A great deal of attention has been given to the description of certain of its properties, although no exhaustive answer has as yet been given to the direct question of what bound water is. However, in order to control the properties of the ground in general and frozen ground in particular, information about the bound water is clearly inadequate.

At the beginning of the 1920's, it was established[3] that a film of adsorbed water is characterized by great surface energy at the interface with the solid surface and appreciably less surface energy at the interface with the free water or air.

Obviously, this film must be nonuniform in physicochemical, including thermodynamic, respects. It is possible to assert that the large surface energy is explained in the first case by special orientation of the molecules and their increased density caused by the great effect of the force field of the foreign surface. This agrees well with the data of Table 1.[4]

The low surface energy in the latter case is probably caused by orientation and packing, not of the molecules but of their structural aggregates [clusters].

The noted difference in the surface energy permits the assumption that the process of binding of the water by the foreign surface is realized by two mechanisms--with and without rupture of the bonds between the molecules. This provides a basis for considering the bond of the water by the particles as collective hydration of the ions forming the surfaces and consideration of this hydration identical to the hydration of the individual ions in the solution.

In the chemistry of solutions, as is known, although there are different points of view,[5] nevertheless the opinion prevails that each ion initially forms an aqua-complex, a component part of which is a sphere of fixed molecules. After originating, the aqua-complexes in turn have an orienting and compacting effect on the structured molecules surrounding them.

The calculations performed using the M. Born equation[6] demonstrate (Table 2) that the "short-range" hydration--the formation of aqua-complexes --for various ions is accompanied by significantly different expenditures of energy. For univalent ions, they are from 30 (Cs^+) to 90 (Cu^+) kcal; for bivalent ions, from 157 (Ba^{2+}) to 318 (Cu^{2+}) kcal; for trivalent ions, they are from 522 (Sc^{3+}) to 687 (CO^{3+}) kcal.

The energy of "long-range" hydration (hydration of the aqua-complexes of all bivalent and

TABLE 1 Magnitude of the Free Surface Energy (ergs/cm^2) of the Water at the Interface with Solid Material

Solid Material	σT Water	Solid Material	σT Water
PbI_2	130	$SrSO_4$	1,400
$CaSO_4 \cdot 2H_2O$	370	$BaSO_4$	1,200-3,000
Ag_2SO_4	575	CaF_2	2,500
PbF_2			

trivalent ions) is appreciably less than the energy of short-range hydration. However, for univalent ions these differences are not so great, and some times they are absent. It is of special interest that all ions within their groups expend equal amounts of energy on further hydration.

This fact, just as the differences in energies of the short-range and long-range hydration, permits advancement of an important principle according to which the total sphere of hydration is identical for all ions. The difference consists only in the energy expended on the distortion of the structure of the water in this sphere. Correspondingly, the water of short-range hydration differs essentially from the water of long-range hydration of the ions.

Beginning with the above assumption, it is possible to make the following assumption. Each particle of the soil material, independently of its nature, binds a certain amount of water by a unit of its surface. The nature of the particles, and, in turn, their surface energy, is reflected in the structural distortions: the greater the energy of the particles, the stronger the structural deformation. Since the surface forces of the particles are external forces with respect to the molecules, the work of them with respect to deformation of the hydrogen bonds causes an increase in the free potential energy of the water adsorbed by the particles in the hydration process.

Thus, the part of the film of bound water that is in contact with the surface of the particles is formed from a special phase of the water. The molecules in it are characterized by an increased electron density. There are nonclassical forces of repulsion between them, the measure of which is the energy required to complete the work of bringing them together.

It is quite obvious that such molecules cannot be rearranged into the structure of ice, and therefore the phase formed by them, which we shall call the "boundary phase," is not converted into a crystal body by a reduction in temperature.

The boundary phase and the surface ions together are a surface aqua-complex, the capacity for hydration of which determines the existence of another part of the film of bound water. However (according to the data of Table 2), the hydration energy of this complex depends on the nature of the ions. Nevertheless, inasmuch as this long-range hydration encompasses an appreciably greater volume of water than the short-range hydration, the molecular bonds in it do not break, but are only deformed although to a different degree.

Beginning with this and based on the general concepts of variation of the molar free surface energy of the water, it is possible to consider that the second part of the film consists wholly of "liquid crystals."[3,4]

This is a special phase of the water, and we shall call it the boundary phase. It is true that it is nonuniform in physical respects, inasmuch as the structural distortion of the liquid crystal decreases from the surface of the aqua-complex. However, this nonuniformity also determines the possibility of its simultaneous equilibrium state with the boundary and volumetric phases of the water.

On proceeding to an explanation of the strength

TABLE 2 Energy of Hydration of Ions and the Aqua-Complexes Formed by Them

Ions	Total Energy of Hydration of the ion, kcal	Energy of Formation of the Aqua-Complex, kcal	Thermodynamic Parameters of Hydration of Aqua-Complexes, kcal		
			ΔG_B	ΔH_B	$T\Delta S_B$
H+	-263	--			
Li+	-125	-78	-46	-47	-0.894
Na+	-100	-55	-44	-45	-0.894
K+	- 79	-38	-40	-41	-0.596
Rb+	- 75	-36	-39	-39	-0.596
Cs+	- 68	-30	-37	-38	-0.596
Cu+	-139	-94	-44	-45	-0.894
Mg^{2+}	-464	-274	-185	-188	-2.980
Ca^{2+}	-382	-205	-175	-177	-2.980
Ba^{2+}	-316	-157	-156	-159	-2.980
Mn^{2+}	-445	-267	-178	-181	-2.980
Fe^{2+}	-468	-282	-182	-186	-2.980
Co^{2+}	-497	-311	-183	-186	-2.980
Cu^{2+}	-507	-318	-188	-191	-3.278
Al^{3+}	-1122	-670	-442	-450	-7.450
Sc^{3+}	-947	-522	-418	-425	-7.152
Ti^{3+}	-1027	-597	-423	-430	-7.152
V^{3+}	-1053	-619	-427	-434	-7.152
Fe^{3+}	-1072	-636	-429	-436	-7.152
Co^{3+}	-1126	-687	-432	-439	-7.152
Mn^{+3}	-1098	-660	-431	-438	-7.152

of frozen soil and a model representation promoting discovery of its nature, it is necessary primarily to consider the following. The freezing of the ground is accompanied not only by the formation of ice but also aggregation of particles. Here, the thawed crystals are formed from liquid characterized by relatively weak distortion of the structure and making up the basic mass of the boundary phase, which freezes in a small temperature range of ~0° to -1°C. The liquid crystals with strong distortion of the bonds between the molecules do not freeze even at very low temperatures. They also make up the unfrozen water determined by the calorimetric method. However, the amount of this water depends basically on the degree of aggregation of the particles when freezing, a careful investigation of which belongs to Z. A. Nersesova.[7] The fact is entirely natural that in cases where tha absorbing complex is occupied by multivalent cations, the aggregation takes place more intensely, and the amount of unfrozen water at the given temperature decreases significantly. On the contrary, on saturation of the absorbing complex with univalent cations dispersing agents, the unfrozen water undergoes smaller quantitative changes at the same temperature.

If the amount of unfrozen water does not vary with a reduction in temperature, this means that the soil has arrived in dynamic equilibrium with the environment. In it all the particles and aggregates, although they are separated by the effect of the wedging pressure, at the same time are bound into a whole by the capillary pressure of the film of unfrozen water. In this sense the frozen soil is a special natural body reinforced by thin aqueous films of unfrozen water, the strength characteristics of which are determined by the surface properties of the solid material. Nevertheless, the strength of frozen soil depends not only on the primary mineral particles but also on the strength of the ice bodies and mineral aggregates. This is connected with the fact that the ice bodies and mineral aggregates, although they are reinforced by thin film of unfrozen water, are nevertheless characterized by low strength.

In accordance with the known thermodynamic representations,[8] the strength of the ice crystal is determined by the work of its formation U_I equal to

$$U_I = U_{Is} - U_{Iv}, \qquad (1)$$

where U_{Is} is the energy of formation of the crystal surface without considering the work of formation of its volume and: U_{Iv} is the energy of formation of the volume of the crystal without considering the energy of formation of its surface.

The energy expended on the formation of the volume is a kind of gained energy; therefore, the small ice crystals must have great strength by comparison with the large ones.

Unconditionally, the mineral aggregates differ essentially from the crystals. Nevertheless, for the general estimate it is possible to use the discussed representation for crystals and consider that the strength of aggregation is determined from the expression

$$U_a = U_{as} - U_{av}, \qquad (2)$$

where U_a is the energy of formation of the aggregate, U_{as} is the energy of formation of the surface of the aggregate without considering the energy of formation of the volume, and U_{av} is the energy of formation of the volume of the aggregate without considering the energy of formation of its surface.

From Equation (2) it follows that in the first approximation the small aggregates must have greater strength than the large aggregates. For this reason the deformations of the frozen soil accompanied by the formation of intravolumetric interfaces begin with the deterioration of the larger aggregates, the energy of formation of which is comparatively small.

In accordance with Equation (1), the soil with massive texture where there is only ice-cement[4] must be characterized by the greatest strength by comparison with the soil of layered and layered-reticular texture. Inasmuch as the large accumulations of ice are observed in the freezing soil, its mechanical strength must be least. This is also confirmed by the experimental data (Table 3). In this connection it is of great interest that the fractionation of the sand particles causes a significant increase in the volume of the soil in the frozen state.

According to Equation (2), the strength of the clay soil subject to significant aggregation during freezing must be less than the strength of the sandy soil for identical but quite low temperatures.

If this principle is true, then among the cohesive soils in general the greatest strength must characterize the soil that is saturated with univalent cations and the least strength with multivalent cations. The correctness of this propostion is proved by the data presented (Table 3).

It is known that among the univalent cohesive soils in general the effect on the energy of aggregation of the particles is felt from the trivalent minerals; then it is natural that the strength of the ground saturated with the latter must be greater by comparison with that saturated with bivalent cations. This is also observed in reality. Under the effect of trivalent cations practically all the soil loses its heave properties over a prolonged period.

Thus, the theoretical generalizations and experimental data permit us to draw important conclusions having great significance for practical purposes.

Above all, the bound water is two-phase water. At the present time the following are isolated in it in an entirely well-founded manner: the boundary phase comprising molecules characterized by increased electron density and the presence of nonclassical forces of repulsion and the boundary phase that is the set of liquid crystals and therefore not increasing in volume on solidification of the latter (during crystallization).

The bound water in the soil characterized by

TABLE 3 Mechanical Strength of Different Soils at a Temperature of -10°C

Name of the Soil	Mechanical Strength, kg/cm^2	Soil Moisture Content %
Quartz sand of medium size (unwashed)	98	20
Quartz sand of medium size (washed)	141.8	22.4
Quartz sand, fine	223	20
Natural kaolin	59.4	53.3
K--kaolin	80.4	45.2
Ca--kaolin	66.8	52.5
Fe--kaolin	75.8	58.3
Natural Fullers' Earth	59.1	100.0
K--Fullers' Earth	75.1	100.0
Ca--Fullers' Earth	50.1	100.0
Fe--Fullers' Earth	72.3	100.0
Natural heavy suglinok	62.0	40.0
Ice	23	

the exchange-adsorption capacity has a large free energy content, a reduced freezing point, and, as a consequence, a high migration capacity. In all of the natural ground it is not subject to the effect of the gravitational field and differs from the free water by the wedging and capillary pressure making up its physical essence and nature in its unity.

In the soil saturated with univalent cations, as follows from the data presented (see Table 2), the particles cannot expend any large amounts of energy on distortion of the structure of the bound water. Nevertheless, they contain a greater unfrozen part of it than soil saturated with multivalent cations. The specific free surface energy of the particles of fine sand and the medium sand is identical. However, there is more unfrozen water in it than in sand. Nevertheless, such soil differs from all of the others [*sic*] by the greatest strength that, as can be proposed, is connected with a high degree of reinforcement with thin films of unfrozen water. Here, by the degree of reinforcing we mean the ratio of the total surface of the thin films to the total surface of the initial (elementary) particles.

However, it follows from the above comparison, in order to increase the strength of the soil in the frozen state it is necessary not only to change the energy state of the bound water but also the physicochemical properties of its solid phases. This is possible if we decrease the amount of chemically active particles in the ground to a minimum, including the particles with high surface energy. It is important to retain the very fine grain in certain cases, preventing coagulation and significantly increase it in others without increasing the chemical activity of the newly formed particles. The freezing of the water (solidification of the liquid crystals) in place in the form of ice-cement has special significance.

The strength characteristics of the films of unfrozen water, as noted above, are determined by the surface properties of the solid material. This is the general principle, and it is not justified when the solids, having high surface energy, on the basis of the presence of a significant number of defects, are themselves characterized by comparatively low strength. For this reason, considering Equations (1) and (2), it is necessary to consider that the real strength of the frozen ground is expressed by the relation

$$z_M = (W_{Sa} - W_{Va})S_a + W_{Si} - W_{Vi})S_i, \qquad (3)$$

where S_a is the total surface of the mineral particles (aggregates), cm^2; S_i is the total surface of the ice bodies, cm^2. Other notation has the former meaning.

Considering that (2)

$$W_{Va} = \frac{2}{3} W_{Sa} \qquad (4)$$

and

$$W_{Vi} = \frac{2}{3} W_{Si}, \qquad (5)$$

Equation (3) can be written as follows:

$$z_M = \frac{1}{3} W_{Sa}S_a + W_{Si}S_i. \qquad (6)$$

In accordance with the above, the closest problem of physicochemistry of frozen ground [6] is the development of methods of improving the strength of the ice formed in the structural aggregates if in the first pores it turns out to be impossible to exclude the coagulation of the particles.

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ELASTIC AND ELECTRICAL PROPERTIES OF FROZEN GROUND

A. D. FROLOV Moscow Geological Exploration Institute

The broad development of geological exploration operations, mining industry, and construction in the permafrost zone requires the expansion of the studies of physicomechanical properties of cryogenous rock. The electrical and elastic properties are the most important characteristics of any solid state and frozen soils in particular. The frozen soil is multiphase and a frequently multicomponent medium subject to the dynamics of transformations among which the leading role belongs to the phase transitions of water. When soil freezes (or thaws), the variations of its physical properties are caused primarily by two basic factors: (1) the composition, that is, the content and distribution of the unfrozen water, and (2) the ice content and the cryogenous texture, that is, the quantity and nature of the ice distribution. These factors depend on the characteristic features of phase transition of the liquid-solid state of the pore moisture, which are most complicated and varied in the sandy-clay soils. The ice content and, the amount and the composition of the unfrozen water under the given thermodynamic conditions are determined by the magnitude of the specific surface (the degree of grain-fineness) and the mineralogic composition (the adsorption activity of the matrix) of the soil and also the magnitude and nature of mineralization of the pore solution and moisture.

An analysis of the available data on the composition and structure of frozen soils[1,2,3] indicates the great variety of states and variations of natural cryogenous formations connected with variability of the granulometric, mineralogic, and phase compositions. All of this causes complexity in studying their physical properties under natural conditions. Therefore, laboratory studies are acquiring very important significance in establishing the basic laws of the formation and variation of the physical properties of frozen deposits.

The first experiments with respect to studying the elastic properties of frozen soils, which we performed at the Geocryology Department of Moscow State University, demonstrated that the propagation rate of elastic waves and the elastic moduli can vary significantly because of many factors.[4] Further studies offer the possibility of laying down the physcial principles and noting the paths of application of acoustic methods for the study of frozen soils and also the discovery of the basic features of the propagation of elastic waves in frozen sandy-clayey soils.[5] It is necessary to consider the latter when performing geophysical operations by seismic and acoustic methods in permafrost regions and when studying the processes occurring in frozen soils.

Several years ago, the studies of the electric properties of frozen soils in variable electromagnetic fields were started. It must be noted that up to this time in the literature, with the exception of the work of B. N. Dostovalov,[6] there were no data in this important and interesting field of the physcis of frozen soils. At the present time, definite results have been obtained that will permit the establishment of the basic relationships between frequency and temperature of the electric properties of frozen soils.

The information obtained and available in literature on the electrical resistance of frozen soils[7,etc.] permits the creation of the physical principles of electrometric methods of studying the phase composition of frozen soils and also of more well-founded consideration of the specific nature of their properties when interpreting the data from field electric logging and electrical exploration of boreholes using direct and alternating current, when evaluating the conditions of propagation and absorption of electromagnetic waves, and so on.

Above all, for such studies monomineral materials--quartz, sand, and kaolin--were selected. The last is especially convenient since it permits the study of a quite broad range of moisture content without the application of the effects of swelling and offers the possibility of performing experiments on clay soils in the purest form.

ELASTIC MODULI OF FROZEN SOILS AND OF ICE

The mechanical properties, in particular elasticity (rigidity), of a frozen soil are determined to a great degree by the process of ice formation. The ice formed in the soil is, on the one hand, cement, ensuring additional cohesion, that is, monolithic growth of the medium and, on the other hand, one of the rock-forming minerals of the polycrystalline frozen soil. Depending on the degree of saturation of the voids and the nature of the ice distribution with respect to the particles of the mineral matrix, a number of characteristic types of ice and cement are isolated.[2,3] With a small degree of filling in the pores, the ice plays the role only of cement of the contact or film type. In this case the soil in the frozen state remains a gas-saturated porous medium; however, its compressibility on reduction of the temperature decreases sharply as a result of the cementing effect of the ice. With complete saturation of the soil formed when it freezes, the ice fills the central parts of the pores and also the regions at the contacts of the particles and forms polycrystalline or multicrystalline grains. The unfrozen water remains in the form of films or discrete accumulations in the boundary zones between the mineral grains and the ice. A large part of dissolved salts, the bases and other admixtures, are forced into these zones during the process of freezing of the pore solution. Thus, a fine (sand-clay) soil in the frozen state becomes close to a polycrystalline medium, the strength, elastic, and other properties of which are determined by the dynamics of the defects and dislocations and, the structure and composition of the boundary intergranular zones. The ice crystals formed often are broken, which has an additional effect on the temperature dependence of the mechanical properties of the rock. Thus, in both cases the mechanical properties (strength, elastic moduli, and so on) of the soil must vary significantly on variation of the temperature and moisture content.

Let us consider the experimental results obtained on the values and the corresponding relations of the elastic moduli for frozen sand and kaolin that were determined from the propagation rates of the longitudinal V_p and Rayleigh V_r waves obtained by the method of longitudinal profiling by the ultrasonic pulse method[8] on frequencies of ~100 Hz. The dynamic elastic moduli obtained by the ultrasonic method are in view of the short-term nature of the effect and small (appreciably less than the elastic limit) stresses the "purest" characteristics of the elasticity of the materials. In Table 1 we have the values of Young's modulus (E), the shear modulus (G), the bulk modulus (K), and Poisson's ratio (ν) of cryogenous soil and ice averaged with respect to many samples, which can be used for appropriate calculations to estimate the strength and the behavior of frozen soil under dynamic loads.

The characteristic features of the dependence of the moduli on temperature and moisture content were also established. Thus, with a reduction in temperature from -1° to -20° for frozen kaolin (Figure 1), there is an increase in the values of the module E, G, and K. For sand this dependence is observed only for moduli E and G, but with a different intensity of the variation than for kaolin for the other absolute values. The bulk modules for frozen sand remains practically constant in the investigated negative temperature range. The greatest variations are observed for the modulus E, kaolin, and K; the least, for ν. The intensity of the variation of the elastic moduli for sand decreases noticeably for -5° to -10° and in practice is diminished approximately at a temperature of -15° and lower at the same time as for clay up to -20° is observed. For saturated kaolin (a degree of saturation of ~1), it was found that in the first approximation the

TABLE 1

Soil	Elastic Moduli, kg/cm² $E \cdot 10^{-4}$	$G \cdot 10^{-4}$	$K \cdot 10^{-4}$	ν
Quartz sand	36 to 32 30 to 26	15 to 13 13 to 10	21 to 19 15 to 15	0.25 to 0.2
Kaolin	12 to 3 13 to 4	4 to 1 5 to 1.5	14 to 4 16 to 7	0.40 to 0.33
Fresh ice	10 to 8	4 to 2.5	9 to 7 12 to 10	 0.4 to 0.3

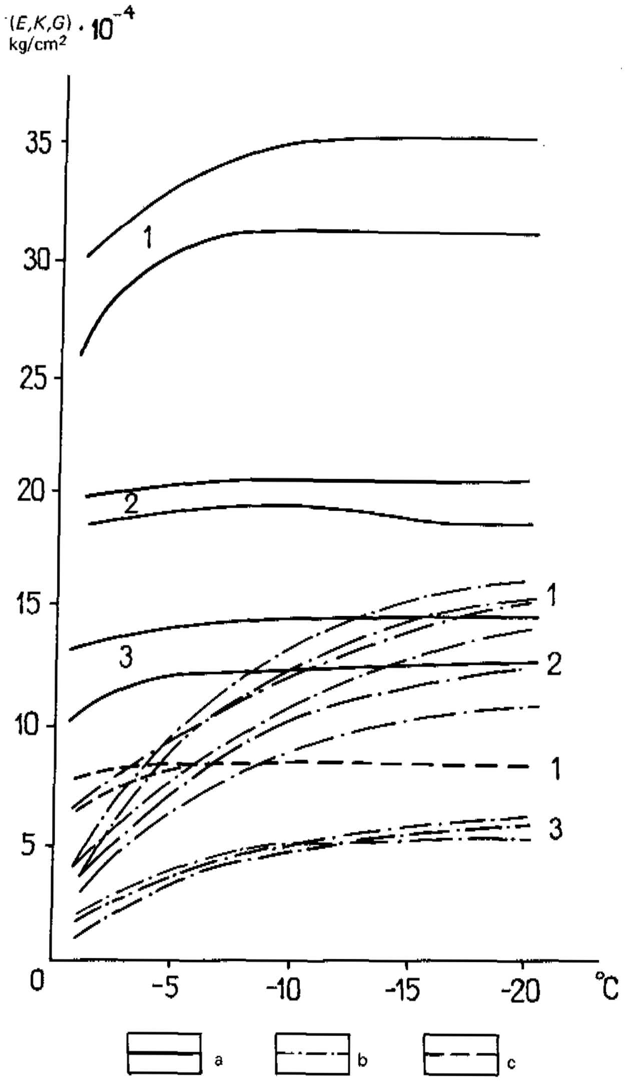

FIGURE 1 Temperature and elastic moduli of cryogenous materials. (a) sand; (b) kaolin; (c) ice. 1--Young's modulus, *E*; 2--bulk modulus, *K*; 3--shear modulus, *G*.

elastic moduli depend linearly on the volumetric moisture content where an increase in the total moisture (ice) content of the soil can cause both an increase and a decrease in their values (Figure 2). This depends on the ratio for a defined fixed temperarure of the elasticity (rigidity) of the ice and frozen soil.

If the elasticity of the ice is higher than for frozen soil, then with an increase in moisture (volume of the ice in the soil) its elastic moduli will increase, and otherwise they will decrease. Hence, it follows that for frozen sandy-clay soils of almost any composition it is possible to discover the temperature for which its elastic moduli turn out to be equal to the corresponding values for ice and will not depend on the volumetric moisutre content (ice content) of the soil. It was established that these temperatures will differ for soils of different lithology and degree of saturation. Thus, for example, for the modulus *G* of the kaolin (Figure 2) this temperature is -13° to -14°; for the modulus *E* it is -9° to -10°; and for the modulus *K* it is ~-5°. This opens up the prospects for studying the nature of the physicomechanical and natural properties of the frozen soil.

The effect of the degree of saturation on the values of tne elastic moduli has been especially studied in samples of quartz sand and clay in the temperature range of 0° to -40°. It has been established that an increase in the degree of saturation causes a significant increase in the elastic moduli. The intensity and nature of this increase depend both on the range of variation of the degree of thawing in the pores and on the lithology of the soil. For frozen sand it was found that the intensity and nature of increase in the values of the elastic moduli are determined only by the *range of variation* of the degree of saturation, and the effect of the temperature on this function is insignificant

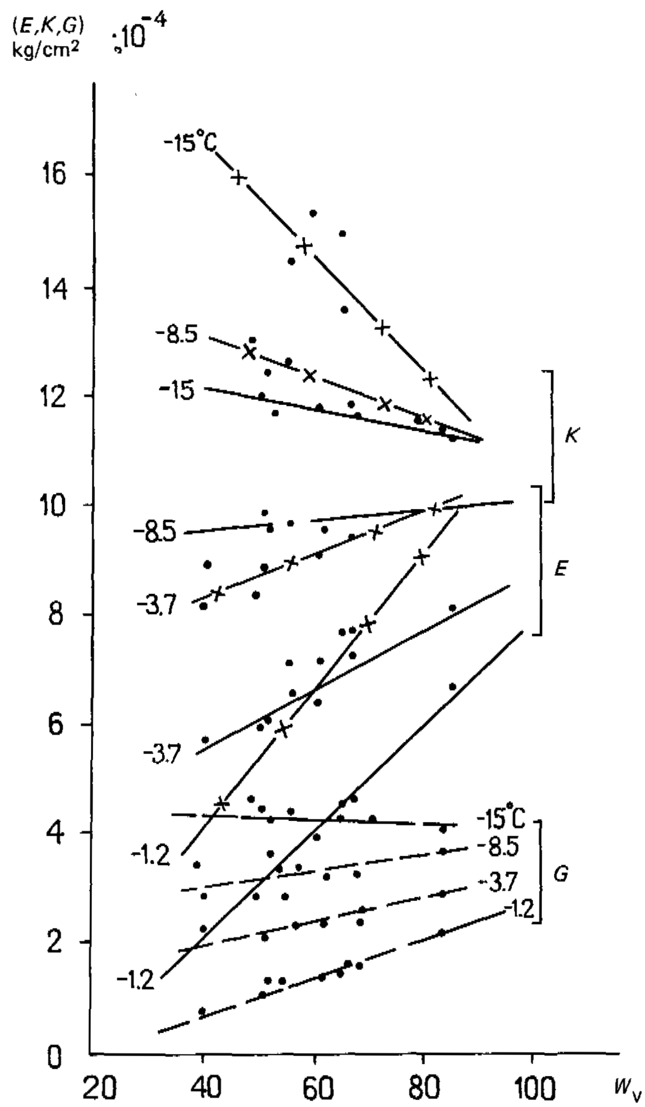

FIGURE 2 Elastic moduli of kaolin as a function of degree of saturation. Young's modulus, *E*; bulk modulus, *K*; shear modulus, *G*.

and is reduced only to variation of the absolute values of the moduli. In the very fine-grained soils (heavy suglinok) in contrast to sand, a reduction in the temperature causes an increase in the elastic moduli with an increase in the degree of saturation.

ELECTRICAL PROPERTIES OF PERMAFROST

The electrical properties of frozen sand-clay soils are specially determined in a variable electromagnetic field, not so much by the properties of the individual components of the soil as by the mutual effect. However, the quantitative consideration of these effects is still impossible to realize as a result of the insufficient degree to which they have been studied. Therefore, it is necessary to begin with simplified representations that, in our opinion, can be reduced to the following:

1. The electrical properties of the frozen soil are determined by the ice and unfrozen water contents, since these components are characterized by significantly greater electrical conductivity and polarizability than the material of the mineral grains and air.
2. The ice and unfrozen water are incomplete electrical materials, the properties of which must be described by the complex characteristic (the conductivity $\sigma = \sigma' + j\sigma''$ or the permeability $\varepsilon = \varepsilon' - j\varepsilon''$) and certain differences in their relaxation times, which depend on the temperature.
3. For pure ice[9,10] it was found that depending on the field frequency and its temperature the dielectric constant ε' can vary from 90 to 3.5, and the loss factor ε'' from 60 to 2.
4. For unfrozen water (the pore solution) it is possible to assume that the distribution of the relaxation time was determined by the mobility of the ions contained in it and the dipole groups, which had to vary noticeably with variation of the temepratire and concentration. The latter was determined both by the composition of the pore moisture and the film thickness or the discrete inclusions of unfrozen water, which depends on the mineralogic and granulometric composition of the soil and also on its temperature.

An unavoidable conclusion from these principles is the conclusion of the presence in the frozen soil of frequency and temperature variation of the electrical properties.

The experiments performed permitted data to be obtained that confirm the stated representations and offer the possibility of establishing the characteristic features of both types of variation for rock of different mineralogic composition and degree of saturation.

As examples in Figures 3 and 4, we have some of the relations obtained for frozen sand and clay soil. The experimental results and their analyses are investigated in more detail in individual papers, part of which have been published.[9,11,12]

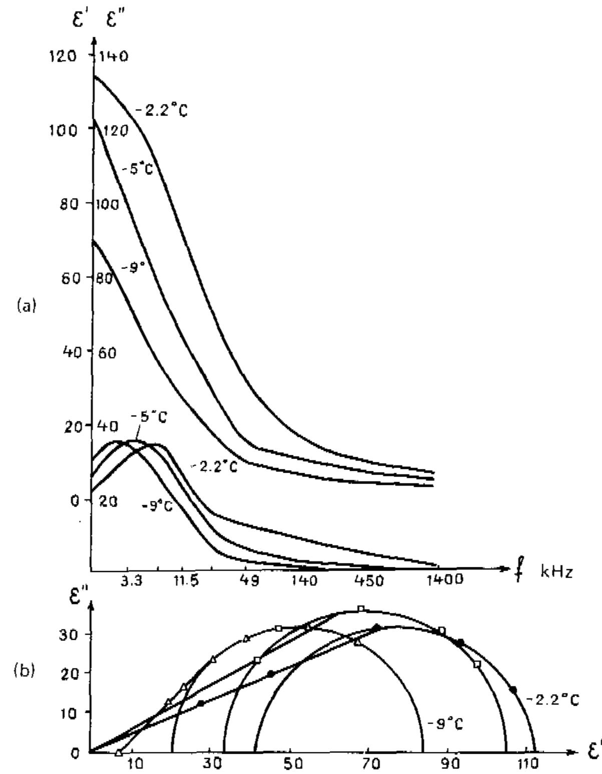

FIGURE 3 Frequency effects on the electrical properties of sand with a moisture content by weight of 19.2 percent for various temperatures. (a) curve; ε'--1; ε''--2. (b) Cole-Cole diagram.

FREQUENCY EFFECTS (Figure 3)

1. The dielectric constant ε' of frozen soil decreases significantly with an increase in the frequency of the electromagnetic field. The range of the strongest variations of ε' with a reduction in temepratire shifts to the low-frequency region. The values of ε' vary from 80 to 110, with the highest temperature of the experiment (-2°) and frequency 1.5 kHz to 4-6 at a frequency of 1.5 MHz for sand and to 4-30 for clay soil in the temperature range of -70° to -2°.
2. The loss factor ε'' has a maximum at a frequency that depends slightly on the total moisture content (ice content) of the soil and is lower the lower the temperature.
3. The relaxation time is asymmetrical with respect to the most probable distribution; the Cole-Cole diagram is asymmetric (Figure 3c).

TEMPERATURE EFFECTS (Figure 4)

1. The dielectric constant ε' decreases with the decrease in temperature, asympototically approaching a constant value at sufficiently low temperatures (for quartz sand, below -15°; for kaolin, below -45° to 50°; for mixtures of them, in the intermediate range).
2. The loss factor ε'' has a maximum at a

temperature that is shifted to the range of lower values with a decrease in frequency of the electromagnetic field.

3. For frozen sand one range of variation has been established in the temperature range of about -5°; for frozen clay soils, two variation ranges have been established in intervals of about 0° and about -30° to -40°.

4. The relaxation time distribution is symmetrical with respect to the most probable [distribution curve]; the interdependence diagram ε' (t°) and ε"(t°) is a semicircle (Figure 4c).

The relaxation nature of the frequency and temperature effects on frozen soils has been experimentally proved, and their characteristic features have been established.[9,12]

The experiments performed also permit evaluation of the effect of the moisture (ice) content and the mineralogic composition on the electrical properties of the frozen soil. The effect of the total moisture content has been studied in greater detail and the quartz sand

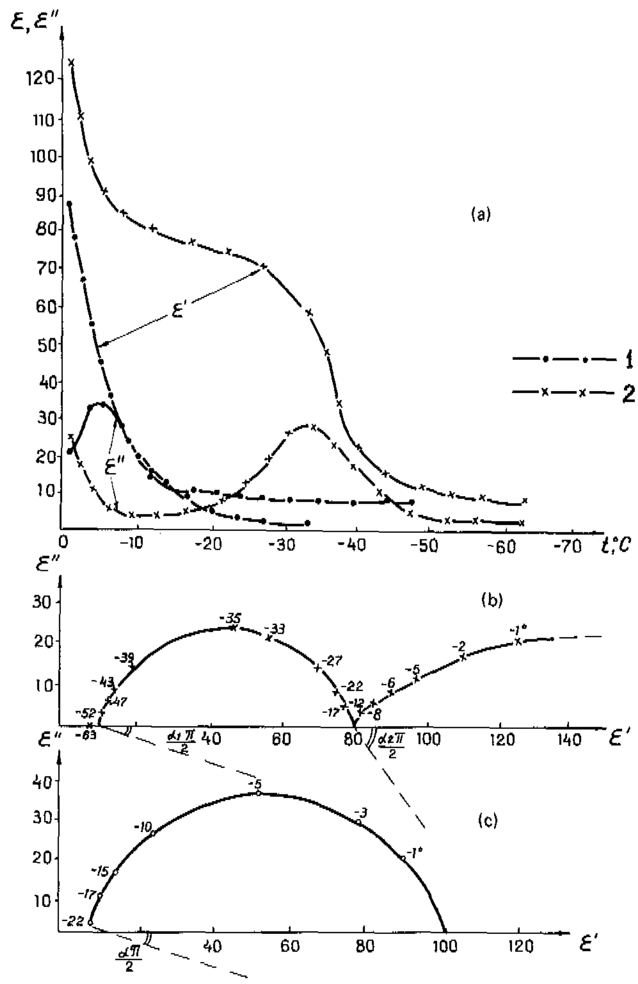

FIGURE 4 Temperature effects on the electrical properties of frozen soil (moisture content by weight 5 percent; frequency, 1.5 kHz). 1--sand; 2--kaolin. (a)-Graphs of ε'--1; ε"--2. (b,c)-Diagrams of the interdependence of ε' (t°) and ε" (t°) for kaolin (b) and sand (c).

samples in the range of values of the moisture content by weight from 1 to 19 percent, and the mineralogic composition and the samples of mixtures of sand and kaolin or variation of the kaolin content in the sample from 5 to 100 percent. The basic results reduce to the following:

1. The variation of the total moisture (ice) content has a slight effect on the electrical properties of the frozen sand at negative temperatures close to 0°, which differs significantly from the properties at positive temperatures.[13] At sufficiently low temperatures, when the amount of unfrozen water in the frozen rock is negligibly small, there is a clear dependence of ε' and ε" on the ice content.

2. The dependence on the mineralogic composition of the soil is excluded primarily in terms of the unfrozen water content in the soil of different lithology and granulometric composition, that is, it is possible to consider that under other equal conditions the electrical properties significantly depend on the effective (considering the adsorption activity) specific surface of the mineral components entering into the rock composition.

Thus, the unfrozen water content and also the kinetics of its phase transitions have a primary effect on the electrical properties of the the frozen soil.

In conclusion, let us note that the material investigated in the report permits compilation of the basic concepts of the nature and characteristic features of dynamic, elastic, and electrical properties of frozen ground that are the most sensitive parameters reflecting the specific variations occurring in cryogenous formations and the different thermodynamic conditions. Knowledge of them permits discovery of many aspects of the physicochemical processes occurring in frozen soils and the development of the physical principles of various electrometric and acoustic methods of studying them, and it permits a better-founded approach to the interpretation of the geophysical data. Of course, the studies made are only an initial step, creating the basis for planning and realization of further studies in this field, which, in our opinion, must be expanded and deepened.

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ELECTRICAL CONDUCTIVITY OF FROZEN ROCKS*

V. S. YAKUPOV Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

The electrical conductivity of frozen and thawed rocks is identical in nature: The basic current-carriers are the ions of the pore solution and the ions of the external shell of the double electric layer at the mineral-solution interface, primarily adsorbed. The relative role of these groups of ions is determined by their numerical ratio, which depends on the porosity, the mineralization of the pore solution, and the magnitude of the active specific surface. The total number of ions in the pore solution of the rock as a result of freezing does not vary and, if its consequence were only a decrease in thickness of the liquid solution films with corresponding growth of concentration, then the conductivity of the soil would also have to remain constant. In reality, some of the ions of the pore solution are engulfed by the ice as a result of isomorphic replacement of the water molecules by ions and the formation of liquid solution films in the ice inclusions; and the role of the adsorbed ions increases, in particular, on the basis of better conservation of the basic means of their movement, since the strongly bound water does not freeze. However, the phase transitions of the water in the freezing rock change not only the amount of liquid pore water and the concentration of material dissolved in it but simultaneously the structure of the pore space of the rock, and, in the case of loose deposits, frequently their total moisture content and specific surface in accordance with the decrease in proportion of mineral particles in the volume of the rock and aggregation of some of them. As a result, the primary cause of the development of the electrical conductivity of thawed and frozen rock turns out to be an in-

*In this paper, the word "rock" is used in the geological sense, meaning "rocks and soils" to engineers.

crease in the tortuous nature of their pore channels, that is, the lengthening of the current-conducting paths of the rock caused by the formation of the ice inclusions. The disturbance of continuity of the conducting medium as a result of separation of the individual parts of the rock by ice inclusions is especially large in the case of statistical systems (loose deposits; and exception is the very fine-grained soils) and much less in the case of matrix systems (rock). The process and magnitude of the variation of the electrical conductivity of freezing rock and its final value are a function of the granulometric composition in the case of unconsolidated rock (soils) and a function of the structure and volume of the pore space in the case of solid rock, the moisture content, the type of cryotexture and cryocomposition, mineralization of the saturation water, and temperature. The basic results of studying the specific electrical resistance of permafrost in its natural occurrence using the DC VEZ method are reduced to the following.*

INCREASE IN ELECTRICAL RESISTIVITY OF ROCK AS A RESULT OF FREEZING

In rock the volume and shape of the inclusions formed when the ice freezes are rigidly bound by the pore space of the rock. As a result of this, the increase in electrical resistivity of the freezing rock is limited; it increases on the average not more than 10 times. The magnitude of the discontinuity is greater, the greater the proportion and absolute amount of free water, that is, the greater the percentage of relatively large pores and the more coarsely clastic the rock. The resistivity of the microporous rocks (some varieties of aleurites, clay shales, extrusives, possibly some intrusive rock, monolithic quartz veins, and so on) is practically identical both for positive and negative temperatures.

The freezing of the unconsolidated deposits usually is accompanied by deep changes in their physical properties. In the unconsolidated deposits, the ice separating the particles and the entire aggregates and blocks of rock is capable of increasing its total moisture content by many times, entirely converting the structure of the pore space, and, in accordance with the decrease in the proportion of the mineral particles, also decreasing the magnitude of the specific mineral surface. Depending on the degree of transformation of the rock during the course of freezing--the type and degree of development of the cryogenous texture and cryogenous composition--its resistivity also varies. The freezing of the very fine-grained soils with the formation of a massive cryotexture increases the resistivity by about 10 times; while the formation of a taxitic† cryotexture, the resistivity increases by about 100 times; with the formation of epigenetic multiple ice veins, about 400 times; with the formation of syngenetic ice veins, about 2,000 times. The resistivity of the very fine-grained soils containing in practice only the strongly-bound water is identical both for positive and negative temperatures. The resistivity of the coarsely fragmented soils during freezing increases by hundreds of times.

The electrical resistivity of frozen rock varies, depending on the lithologic peculiarities, the moisture content, the type of cryotexture, and the cryocomposition within the broad range from 10 to 10 million Ω-m. *It is essential that the relative* differentiation of the rock with respect to electrical conductivity after freezing is retained, and among sedimentary rocks, even increased.* This exceptionally important principle opens up broad possibilities for studying permafrost by electrologging methods.

EFFECT OF MOISTURE ON THE RESISTIVITY OF FROZEN SOILS

The behavior of the resistivity of frozen soils as a function of moisture content depends on their granulometric composition, inasmuch as the type of cryotexture under other equal conditions is determined by the joint effect of these two factors. The disturbance of the continuity and the coarsely fragmented and large-grained deposits takes place even with low moisture content in the stage of formation of the ice-cement at the contacts of the individual particles. As a result, in spite of the comparatively high electrical conductivity of the frozen rock particles taken separately, its total resistance increases sharply. The increase in the initial moisture content changes the resistivity of the coarse deposits little, inasmuch as the nature of the contacts of the adjacent particles is retained.

In fine-grained soils at a moisture content to a certain limit, and in solid rocks, the ice-cement is formed in sufficiently large and usually isolated pores. In this stage, with an increase in moisture, the resistivity increases slowly. At a moisture content corresponding to the beginning of the occurrence of the taxitic cryogenous texture, the resistivity increases sharply and discontinuously, and then its growth again is delayed. The magnitude of the jump reaches tens of thousands of Ω-m.

RESISTIVITY OF FROZEN ROCKS AS A FUNCTION OF LITHOLOGIC COMPOSITION

In addition to the composition, the magnitude of the resistivity of a frozen rock is also affected by the moisture content; for fine-grained

*Yakupov, V. S. Elektroprovodnost' i geoelektricheskiy Razrez merzlykh tolshch (Electrical conductivity and geoelectric section of the permafrost zone). Moscow, Nauka, 1968.

†Taxite is a clastic lava.

*The distribution of the electrical conductivity of frozen rocks is described by a lognormal law.

soils, from the time of formation of the taxitic cryotexture, its effect becomes decisive: The upper limits of the resistance of the coarsely plastic and the most ice-rich fine-grained deposits turn out to be of the same order. Therefore, only rock with a maximum cryotexture is investigated.

The resistivity of the frozen solid rock depends on the porosity and specific surface determining the ratio of the amount of free water to the amoutn of strongly bound water. Therefore, the resistivity can be relatively large, and in the case of rock with insignificant porosity and specific surface (some granitoids, metamorphic rock, quartz veins, and so on) and in the case of rock with great porosity and a large percentage of large pores, inasmuch as the increase in resistivity is large in the case of freezing (sandstones).

The unconsolidated deposits with maximum cryotexture with respect to magnitude of the resistivity decompose into two sets (in the final state they form one set). One of them belongs to the fine-grained lacustrine-alluvial and floodplain deposits in their base; the other belongs to the coarse deposits (channel and oxbow facies of alluvium). Hence, it follows that the resistivity of the loose deposits with massive cryotexture as a function of the size of the particles initially increases slowly; then it increases discontinuously by several tens of thousands of ohm-meters and then again increases slowly. The discontinuity of the resistivity corresponds to structural variations in the rock: ice contained in the rock in the form of individual inclusions as converted into the surrounding medium. The rock from the matrix system, if we consider the conducting medium--water--as the matrix, becomes statistical.

EFFECT OF TEMPERATURE ON THE RESISTIVITY OF PERMAFROST

The specific resistance of frozen rocks with low concentration of the pore solution increases discontinuously on crystallization of the free water, that is, at the time of formation of the cryogenous texture. With a further reduction in temperature only somewhat increasing the dimensions of the previously occurring ice inclusions, the resistivity practically does not change: According to the results of studies by the VEZ method, the frozen lithologically uniform rock of the most varied composition with insignificant mineralization of the pore water independently of the temperature distribution in them is practically a uniform electric horizon, the lower boundary of which coincides within the limits of accuracy of observations with the position of the 0°C isotherm.

If there is only bound water in the rock, then the resistivity increases apparently continuously, increasing in the final analysis by no more than 2 or 3 times. If almost all the water in the rock is strongly bound, then the resistivity is identical both for positive and negative temperatures.

According to laboratory studies on the samples, the electrical resistivity of frozen rocks varies noticeably with temperature. The cause of the divergence of the results of the study in the samples and the large massifs of rock consists obviously in the fact that the laboratory studies of the electrical conductivity of the freezing and frozen rock as functions of temperature are carried out without considering supercooling of the water in the rock and the duration of its transition to ice and without observing the similarity principle, which does not permit extension of the results obtained to the large volumes of rock in the natural occurrence.

EFFECT OF MINERALIZATION OF THE PORE WATER ON THE ELECTRICAL RESISTIVITY OF FROZEN ROCKS

An increase in the mineralization of the pore water lowers the freezing point of the rock and its resistivity, which in the defined concentration range becomes a function of the temperature, affects the type of cryotexture that occurs in the freezing rock, and creates the microcharacteristics of the spatial arrangement of the ice and frozen rock in nature and in the ice inclusions themselves. Just as in the case of saturation with fresh water, the variation of the resistivity of the freezing rock with mineralized pore water is determined by the variation in the tortuous nature of the pore channels of the rock. The effect of the mineralization of the pore water is greater for the coarser soils and for the wetter rock, inasmuch as the free water content in it is greater with respect to the strongly bound water.

The effect of the mineralization of the pore water with sufficiently high concentration is extraordinarily great and entirely changes the picture of the distribution of the resistivity; it prevails over the effect of the lithologic composition and converts the process of freezing of the rock and the nature of the variation of the resistivity of the rock accompanying it.

The usually great difference in the thawed and frozen rock with respect to electrical conductivity, the high degree of differentiation with respect to electrical conductivity of frozen rock of different composition, and the cryogenous texture and cryogenous composition make electrologging a powerful means of studying the morphology, composition, and cryogenous structure of the frozen deposits.

STUDY OF THE STRUCTURAL CHANGES IN SOILS DURING FREEZING

J. AGUIRRE-PUENTE, M. VIGNES, AND P. VIAUD Laboratory of Aerothermics of the National Research Center, France

The problem of heat transport with variation of the free state of the water, known in literature as the Stefan problem, has been the object of numerous studies. However, there are analytical solutions only for a small number of conditions that are of definite interest from the point of view of physics. Among them only the solutions of Neuman and Stefan are strict.[1] They pertain to the uniform semi-infinite model with constant surface temperature or constant freezing rate, and they describe conditions that in practice are not encountered in nature. Other researchers using the mathematical approximation gave solutions applied under defined conditions: Portnov[2] gave the solution for the case of a semi-infinite medium with a variable surface temperature and an initial temperature equal to the freezing point; the solutions obtained by the aerothermics laboratory[3,4] pertain to finite and semi-infinite media with variable temperature at the cooling surface and any initial temperature. The authors[5] have considered this problem in more detail for the case of freezing of ground.

All of these solutions pertain to the case where the medium is assumed uniform. The Stefan problem was solved for the case of a medium with variable characteristics.[6,7] Numerical solutions were given for a number of special cases.[8]

The heat-transport laws using phased transformations are connected with macroscopic phenomena, for example, for interpretation of the position of the freezing front in the ground massif.[9] However, the freezing of the dispersed medium is frequently accompanied by variations of its structure: lensing in the freezing zone, heaving, destruction of biological cells, and so on, which cannot be entirely completely characterized by the indicated laws. In connection with certain assumptions that occur in them, it is not possible to consider the microscopic phenomena at the water-ice and water-solid interfaces.

These phenomena cause the mass-transfer potential in the medium still not covered by freezing. The process rate is low and subject to Darcy's law. However, for the case of an unsaturated medium, the problem of water displacement in two phases arises (thermomigration).

The problem of the isothermal two-phase flow can be considered from the point of view of the mechanics of liquids in a porous medium, whereas the thermomigration requires a more complicated approach.[10]

For a number of years the authors have performed systematic studies of soils of different granulometric composition[5] and obtained the correlation between the intensity of the cryogenous phenomena and the temperature and moisture conditions.[11]

A study was made of the microscopic phenomena in order to establish the interrelation with the macroscopic transport phenomena. After discussion of some of the basic principles of the phenomena at the interface, a study was made of certain models permitting the discovery of the value of the bound water where the freezing front approaches the surface of the solid and also the role of the bound water in the freezing of a capillary system.

It is possible to consider a dispersed medium in the macroscopic aspects as continuous. However, inasmuch as the problem of phase transformations must also be considered in microscopic aspects, it is impossible to use the hypothesis of continuity of a dispersed medium.

In microscopic aspects, systems are used that permit disturbance of the continuity.

The authors call the transition zone at the contact of two phases the interphase zone, where the variation of their characteristics takes place continuously, and the surface of the disturbance of continuity is called the "interface" as proposed by Gibbs.

The physicochemical interrelations of the water and the solid substrate and the thickness of the interphase zone depend on the molecular state of the water and the nature of the solid body.

On going away from the surface of the solid substrate, the intensity of this interrelation decreases, and at a defined distance from it the water has the properties of free water. One of the manifestations of the interaction of water and the solid substrate is a reduction in the equilibrium temperature of the ice and attached water, T_E, as the distance from the surface of the solid state δ decreases. For pure quartz powder this relation was expressed by the curve in Figure 1. This curve has been constructed considering the specific surface of the investigated object by the experiments of Szanto and Aguirre-Puente,[12] who gave the relation for the unfrozen water content as a function of temperature. The continuous variation of the properties of the water near the surface of the solid substrate permits, by analogy with hydrodynamics, characterization of the interphase zone as the boundary layer. The thickness of this layer can be determined with respect to the given parameter. If we take as this parameter the equilibrium tem-

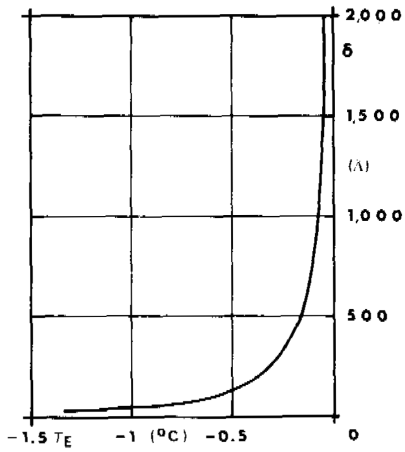


FIGURE 1 Temperature of the ice and bound water, T_E, as a function of the distance from the surface of the solid substrate δ (in Å).

perature of the ice and the attached water, T_E, the thickness of the boundary layer can be defined as the distance from the surface of the solid substrate $\delta = \delta_0$, where the equilibrium temperature $T_E = T_0 - \varepsilon$; here T_0 is the equilibrium temperature of the ice and the free water; $\varepsilon = 10^{-2}$ deg.

In the case of pure quartz, the authors obtained the thickness of the layer of bound water to be on the order of 5,000 Å.[13]

For the majority of systems it is possible to neglect the effect of the layer of bound water, since its thickness is very small by comparison with the dimensions of their components. Thus, for the case of freezing of the water in the vessel at the Gibbs interface having special properties (the energy of the boundary layer, the surface tension), it is possible to take the contact zone of the water and the wall of the solid body and the water and ice. The movement of the freezing front is a Stefan's problem. If the surface of the solid state is nondeformable, then near it the water-ice interface assumes a form that is defined by the equilibrium of three surface tensions: water-ice, water-solid, and ice-solid. This equilibrium is determined by Young's equation. (Young introduced the concept of the contact angle.)

If crystallization of the water takes place in a porous saturated medium, then the peculiarities of this phenomenon depend on the structure of the medium.

In a porous medium with large particles and large pores, the thickness of the interphase zones is very small by comparison with the size of the particles. When the freezing front is advanced in this medium, the curvature of the water-ice interface in the pores is small. Therefore, it is possible to neglect the effect of the bound water and the capillary phenomena, and the solution of Stefan's problem is applicable.

In a porous medium where large particles are close together, the water-ice interface has great curvature. The role of the capillary phenomena in the pores increases, but in connection with the particle size, the number of capillaries per unit volume is small, and their effect on the whole remains insignificant.

In a porous medium with small particles and large pores, the effect of the bound water is great, but the water-ice interfaces have a low curvature, and the role of the capillary phenomena remains insignificant. However, repulsion of the particles by the ice-formation front takes place.

In a porous medium with small particles, the effect of the bound water is exhibited to a greater degree, since the pore dimensions can have the same order of magnitude as the thickness of the bound water layer. The curvature of the water-ice interface increases sharply, and the capillary phenomena are expressed very clearly. The number of particles and capillaries per unit volume is large: The effect of the bond of the water and the capillary nature are exhibited on microscopic scales.

Thus, as the particle and capillary sizes decrease, the role of the water adsorption on the surface of the substrate increases.

The specific surface of the dispersed medium is the parameter that characterizes the increase in the role of the capillary and boundary phenomena with an increase in the degree of dispersion.

MODEL FOR A TWO-DIMENSIONAL INTERFACE

It is assumed that the wet medium is semi-infinite and is bounded by a solid surface (Figure 2a); it is possible to neglect the gravitational force, and the uniform heat flux ϕ is perpendicular to the surface so that the water-ice interface is shifted in the same direction. While the freezing front is far from the ice-water and water-solid state interfaces, they are characterized respectively by the energies f_{iw} and f_{ws}; the latter depends on the nature of the solid state. The system is converted to the direct contact of the ice and solid substrate, which is characterized by the energy of the interface f_{is} (Figure 2b).

If the molecular structure of the water at the surface of the solid substrate corresponds to the ice structure, then, on removal of the heat when the temperature of the solid surface reaches a value of T_0, ice is formed on it $f_{is} \le f_{iw} + f_{ws}$.

If the molecular structure of the water on the surface of the substrate does not correspond to the structure of the ice, then when the freezing front reaches a layer of bound water, its advancement stops (Figure 2c). In this case the absence of the ice formation on the surface of the substrate cannot be expressed using the energies of the interfaces: f_{iw} and f_{ws}, which are defined independently for the cases of ice-water and water-substrate. Actually, the interphase zones of the ice and water and the water and the surface of the substrate come together, creating a region of interaction of three phases; for the approach of the ice and substrate surfaces, it is necessary to have further heat loss.

With a further reduction in temperature, the freezing front will occupy a new position of equilibrium at a distance of $\delta < \delta_0$ (Figure 2d). The analysis of this process from the energy point of view is possible if we assume that in the freezing process the pressures in each of the phases are constant and their volumes vary.

The expression for the free energy per unit surface in the initial stage where $T = T_0$ and $\delta = \delta_0$ (Figure 2c) has the form:

$$F_0 = \int_0^{\delta_0} F(\delta)d\delta + f_{iw} + f_{ws}, \tag{1}$$

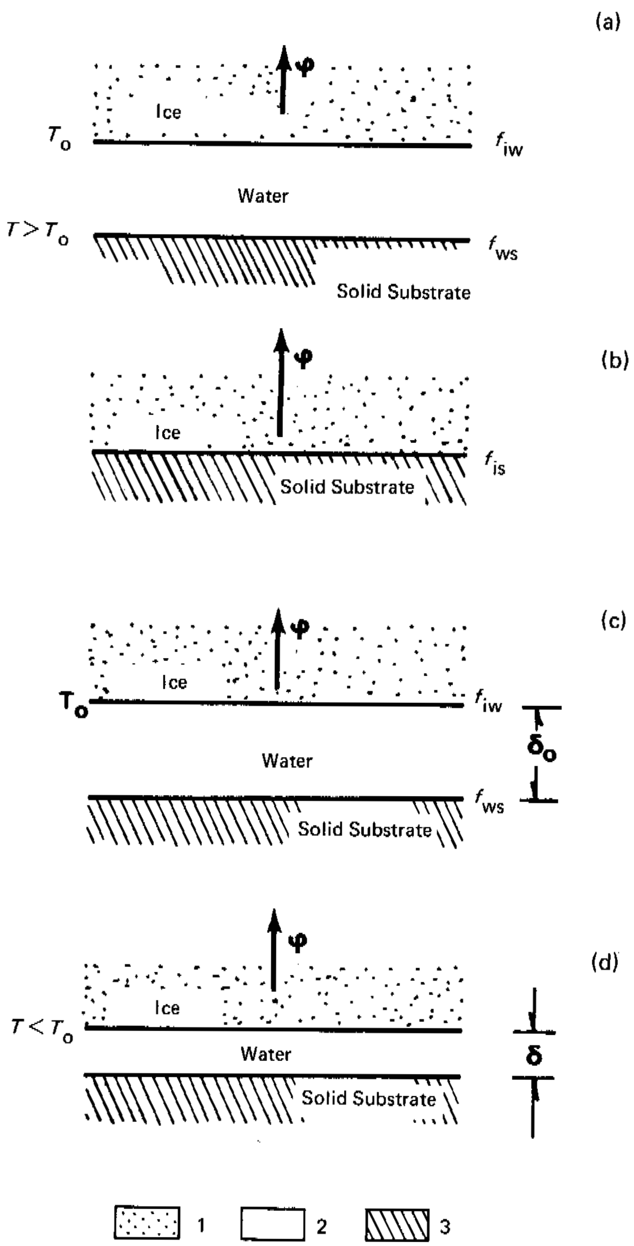


FIGURE 2 Advancement of the freezing front in the case of a flat interface: 1--ice; 2--water; 3--solid substrate.

where $f(\delta)$ is the free energy arriving per unit volume of bound water.

If we assume that the ice can be in direct contact with the substrate where the loss of heat takes place, then

$$F_1 = f_{is}, \tag{2}$$

where F_1 is the free energy per unit surface of the interface of the ice and the substrate. Then it is possible to assume that for the system depicted in Figure 2c, the free energy barrier is

$$\Delta F = F_1 - F_0 = f_{is} - \left(\int_0^{\delta_0} F(\delta)d\delta + f_{iw} + f_{ws}\right). \tag{3}$$

The absence of the ice formation on the surface of the substrate can be expressed using equations (1) and (2) which show that the energy barrier exists only under the condition where

$$f_{is} > \int_0^{\delta_0} F(\delta)d\delta = f_{iw} + f_{ws}. \tag{4}$$

During the process of approach of the ice and substrate surfaces, the spacing between them, δ, becomes less than δ_0 (the interphase zones of ice and water and water and substrate come together); the bound water will become the basic component in the complex system made up of ice, bound water, and the substrate. The use of the concept of the Gibbs interface as applied to this system[14] permits determination of the equivalent interface: It is characterized by the free energy f_{iws}, the magnitude of which depends on δ. The expression for the free energy barrier for a value of $\delta \le \delta_0$ has the form

$$\Delta F = f_{is} - f_{iws}(\delta). \tag{5}$$

Where $\delta = \delta_0$, that is, where the freezing front at T_0 is at the boundary of the layer of bound water, the value of $f_{iws}(\delta)$ is obtained on the basis of Equations (3) and (5):

$$f_{iws}\delta = \int_0^{\delta_0} F(\delta)d\delta + f_{iw} + f_{ws}, \tag{6}$$

and since $\int_0^{\delta_0} f(\delta)d\delta > 0$, then we have:

$$f_{iws}(\delta) > f_{iw} + f_{ws}. \tag{7}$$

Experience shows (see Figure 1) that the distance δ decreases when the temperature drops. The heat loss leads to a reduction of the energy barrier, which, beginning with Equations (3), (5), and (6), gives:

$$f_{is} - f_{iws}(\delta) < f_{is} - f_{iws}(\delta_0), \tag{8}$$

hence

$$f_{iws}(\delta) > f_{iws}(\delta_0). \qquad (9)$$

Expressions (7) and (9) make it possible to predict the shape of the curve $f_{iws}(\delta)$ shown in Figure 3. If we admit that a direct contact between the ice and the substrate is possible, then there is a value of δ^* where the first term of Equation (4) is equal to 0 (curve 1, Figure 3). From the time when $\delta = \delta^*$, $\Delta F(\delta) = 0$, the energy barrier disappears and the process of approach of the ice and mineral particles depends only on the removal of the latent heat released on conversion of the remaining water to ice. The condition where $\Delta F(\delta^*) = 0$ where $\delta^* = 0$ and $f_{iws}(\delta) = f_{is}$ (curve 2, Figure 3) is a special case.

It is necessary, however, to remember that Equation (4) permits the possibility of a direct contact between the ice and mineral particles; if the crystalline structures of the ice and particles are incompatible, then even for very low temperatures between the ice and the surface of the particles there will always be an energy barrier $\Delta F > 0$ localized in the bound water layer (curve 3, Figure 3). This is confirmed by the experiments of Antoniou,[15] who established the presence of two or three molecular layers of attached unfrozen water on the surface of porous glass.

What has been discussed above can be applied to the dispersed systems supplemented by various capillary models.

SIMPLE CAPILLARY MODELS

The classical representations pertaining to simple capillary models contain the following assumptions:

1. The water-ice, water-substrate, and ice-substrate interphase zones are considered as the Gibbs interface.

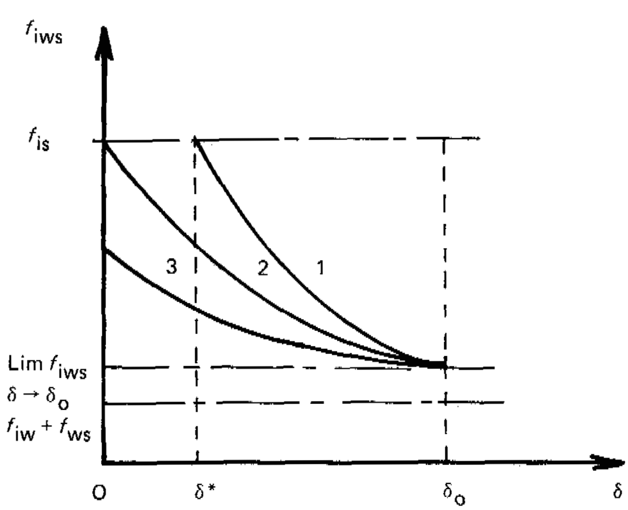


FIGURE 3 Variation of the free energy of the interface f_{iws}: curve 1—for a value of δ^* when in Equation (4) $\int_0^{\delta_0} F(\delta)\,d\delta = 0$; curve 2—$\delta^* = 0$, $f_{iws}(\delta = 0)\ f_{is}$; curve 3—$\Delta F > 0$.

2. On the solid state, which is neutral from the physicochemical point of view, there is no bound water layer.

3. The curvature of the water-ice meniscus is constant, and the water completely wets the surface of the substrate ($\sigma_{is} = \sigma_{iw} + \sigma_{ws}$, where σ_{is}, σ_{iw}, and σ_{ws} are the values of the surface tension of the ice-substrate, ice-water, and water-substrate, respectively).

4. The surface tensions σ, the latent heat of fusion of the ice L, and the physical properties of the two phases do not depend on the pressure and temperature.

5. The stress tensor in the ice is isotropic, uniform, and reduces to the pressure. The effect of the gravitational force can be neglected.

Considering these assumptions, the equilibrium of the interface is subject to the Laplace law

$$P = P'' - P' = \frac{2\sigma}{r}, \qquad (10)$$

where P'' and P' are the pressures in phase " and phase ', respectively; σ is the surface tension at the interface of the two phases; and $1/r$ is the curvature of the interface. In Equation (10), the phase " is with respect to the convex side of the surface and r is positive; consequently, $P'' \leqslant P'$.

In the special case of variation of the phase stage,[16] the displacement of the equilibrium is expressed by the equation

$$-\frac{L}{T}\,dT + (v' - v'')\,dP' - v''\left(d\frac{2\sigma}{r}\right) = 0, \qquad (11)$$

where v' and v'' are the specific volumes of the phases ' and phase ". The system of Equations (10) and (11) contains four variables of which only two are independent. Hereafter, the phases ' and " will be water (W) and ice (i), respectively.

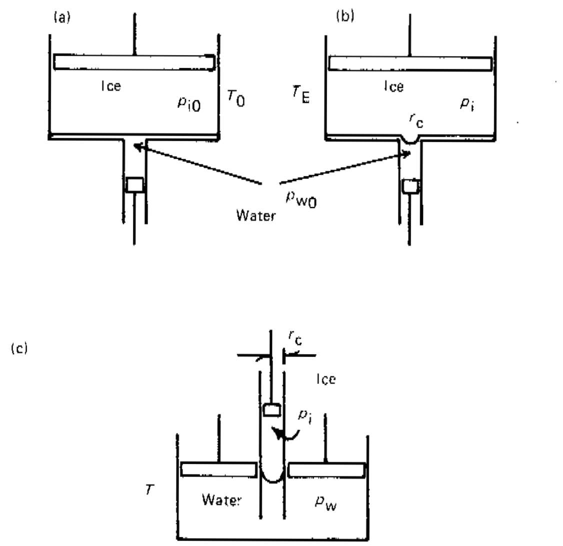
In the systems depicted schematically in Figure 4, a study is made of the phase transformation in the water with constant pressure P_{w0}.

The initial position (Figure 4a) is characterized by the plane interface ($1/r = 0$), $P_{w0} = P_{i0} = P_0$ (normal pressure) and $T_0 = 273°K$ (the equilibrium temperature of the free water and ice). The final position (Figure 4b) is characterized by the hemispherical interface with the radius r_c (the radius of the capillary), the pressure in the water $P_w0 = P_0$, the pressure in the ice P_i, and the temperature T_E.

As was indicated above, the transformations accompanying the transition from the initial state to the final state must be subject to Equation (11), where the second term is zero; the differentials are equivalent to the finite differences:

$$T_E = T_0\left(1 - \frac{v_i \cdot 2\sigma_{iw}}{L \cdot r_c}\right). \qquad (12)$$

Then the cryoscopic reduction in the temperature in the capillary is:



(d)

	Initial Position	Final Position	
		Transformation 1 T Constant	Transformation 2 T Constant
p_w	p_{w0}	$p_{w1} = p_{w0}\ \ \Delta p$	$p_{w2} = p_{w0} - \Delta p$
p_i	$p_i = p_{w0} - \Delta p$	$p_{i1} = p_{w0}$	$p_{i2} = p_{w0}$
T	T_E	T_{E1}	T_E
r	r_c	r_c	r_2

FIGURE 4 Phase transformations of water at constant pressure: (a) initial position (flat interface); (b) final position (hemispherical interface).

$$\Delta T = T_0 - T_E = \frac{v_i}{L} \cdot \frac{2\sigma_{iw}}{r_0} \cdot T_0$$

(the Thomson formula). Using the Laplace Equation (10), the pressure in the ice is determined:

$$P_i = P_{w0} + \frac{2\sigma_{iw}}{r_c}. \qquad (13)$$

Analogously, in the capillary system with a convex water-ice interface, the phase transformations with constant curvature of the latter can be investigated using Equation (11). The initial and final states of the system appear in Figure 4c. From Equation (11), in which the third term is zero, and from Equation (12) we obtain

$$T_{E1} = T_0 \left[1 - v_w \cdot \frac{2\sigma_{iw}}{L \cdot r_c} + 0 \left(\frac{2\sigma_{iv}}{L \cdot r_c} \right)^2 \right]. \qquad (14)$$

For the case of water-ice where $v_1 \simeq 1.1\ v_w$ the expression (14) can be written in the form

$$T_E < T_{E_1} < T_0. \qquad (15)$$

It is possible to represent another phase transformation (variation 2), which is given in Figure 4c, d. Equation (11), where the first term is zero, gives:

$$r_2 = r_c \left(\frac{v_i}{2v_i - v_w} \right). \qquad (16)$$

When $v_i \simeq 1.1\ v_w$ we have $r_2 < r_c$, which excludes the equilibrium inside the capillary under the condition of constancy of the radius of curvature.

According to the adopted assumptions, the capillary diameter is quite large, so that the thickness of the layer of bound water can be neglected.

NEW CAPILLARY MODELS

Let us consider the system depicted in Figure 5 analogous to the model proposed by Everett[17] and formed by the capillary C with the radius r_c connecting two reservoirs A and B closed by pistons.

In addition, let us assume the existence of a layer of bound water on all the inside surfaces.

From the system through the reservoir A, a one-way loss of heat ϕ takes place that causes displacement of the ice deposition front downward.

The water-ice interface maintains the equilibrium temperature T until it reaches the bound-water layer, δ_0, on the inside wall of the reservoir A.

If $\delta_0 < r_c$, then inside the capillary there is a cylindrical column of free water with the radius r_{eff}. The presence of bound water creates the possibility of its displacement between the ice and substrate; therefore, the physicochemical interaction acquires greater significance than the radius of the capillary.

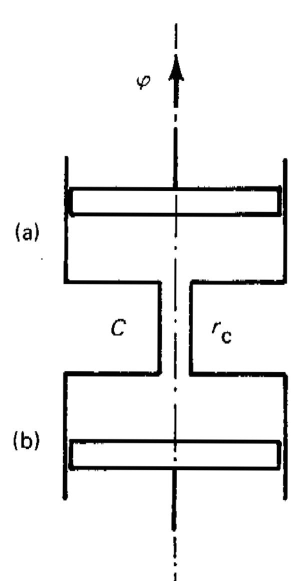


FIGURE 5 Reservoirs A and B covered by pistons and connected by the capillary C with a radius r_c.

CONSEQUENCES OF THE POSSIBILITY OF DISPLACEMENT OF THE BOUND WATER BETWEEN THE ICE AND THE SUBSTRATE

When in the case of a constant heat loss the freezing front reaches the bound-water layer at the bottom of the reservoir *A* (Figure 6a), it occupies the following positions successively $\delta_1, \delta_2 \ldots \delta_j$, which corresponds to the freezing points $T_1, T_2, \ldots T_j$ lower than T_0 (Figure 6b). In the mouth of the capillary and the free water zone of the ice surface, a bulge is formed with a radius r_j.

It is assumed that the phase transformation of the water takes place under the condition of constancy of the initial pressure P_0 in the reservoirs *A* and *B*.

Formula (14), where r_c is replaced by r_j,

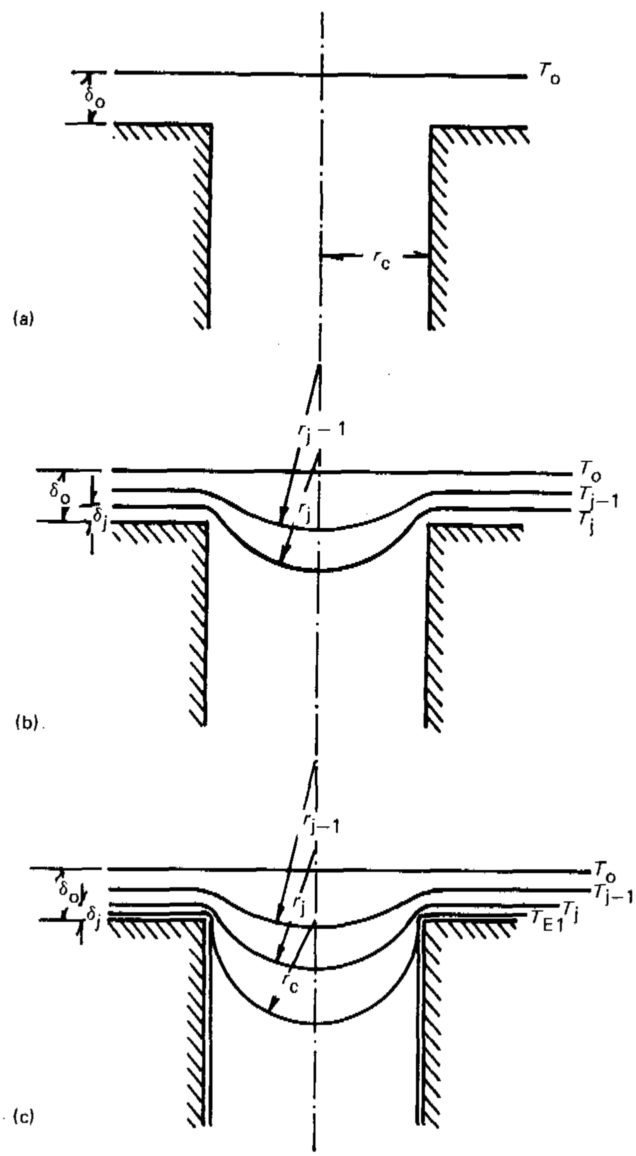


FIGURE 6 Displacement of the ice formation boundary during freezing of the water in the vessel *A*: (a) first stage; (b) second stage; (c) third stage.

gives the equilibrium temperature for the water-ice interface with a radius r_j:

$$T_j = T_0 \; 1 - \frac{v_w \cdot 2\sigma_{iw}}{L \cdot r_j} . \qquad (17)$$

Formula (17), where r_j is replaced by r_{eff}, gives the equilibrium temperature T_{E_1} in the capillary:

$$T_{E_1} = T_0 \left(1 - \frac{v_w \cdot 2\sigma_{iw}}{L \cdot r_{eff}}\right)$$

and if $r_{eff} \simeq r_c$, then

$$T_{E_1} = T_0 \left(1 - \frac{v_w \cdot 2\sigma_{iw}}{L \cdot r_c} .\right)$$

Then the system of curves with identical temperature equilibrium can be expressed by the graph (Figure 6c).

The Laplace equation permits the following equation to be written:

$$\Delta P = \frac{2\sigma_{iw}}{r_j} . \qquad (18)$$

Assuming uniformity of the pressure in the ice and equality of the pressures $P_i = P_0 = P_W$ in the reservoirs *A* and *B*, we obtain a depression in the water near the convexity, the pressure gradient in the capillary, and the water flow from the reservoir to the freezing front. Assuming that this flow follows Poiseuille's equation, we obtain the water flow rate in the capillary

$$q = \frac{\pi r_c^4}{8\eta} \cdot \frac{\Delta P}{l} = \frac{\pi r_c^4}{8\eta} \cdot \frac{2\sigma_{iw}}{l \cdot r_j} , \qquad (19)$$

where q is the volumetric flow rate of the water in the capillary, η^* is the viscosity of the water, and l is the length of the capillary.

If all the water is converted to ice, the latent heat released per unit time is:

$$\phi_L = \frac{q}{v_w} \cdot L . \qquad (20)$$

If we neglect the heat flux necessary to maintain the temperature gradient in the ice, then $r_j(\phi)$ can be expressed as follows:

$$r_j(\phi) = \frac{\pi r_c^4}{8\eta} \cdot \frac{2\sigma_{iw}}{l \cdot \phi} \cdot \frac{L}{v_w} . \qquad (21)$$

If $r_j(\phi)$ defined by Equation (21) exceeds r_c, then for constant ϕ a steady-state freezing regime is established until the radius of the convexity reaches a value of r_j (Figure 7).

This steady-state regime can be characterized in the following way. The influx of water from the reservoir *B* through the capillary is equal to the amount of ice formed per unit time on the

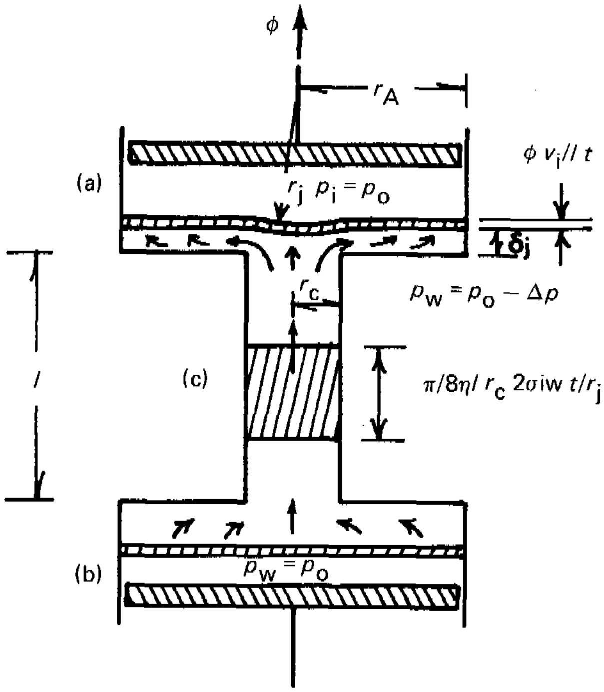


FIGURE 7 Rise of ice (heave) in the reservoir *A* with a steady-state freezing regime.

interface, which keeps the geometric shape, the position, and temperature T_j constant, and the ice in the reservoir *A* will rise. As a result we obtain the magnitude of the heave (= the rise of the ice), V_g:

$$V_g = \frac{\phi \cdot v_i}{L} = \frac{\sigma_{iw}}{4\eta l} \cdot \frac{v_i}{v_w} \cdot \left(\frac{r_c}{r_A}\right)^2 \cdot \frac{r_c^{\,2}}{r_j(\phi)}, \tag{22}$$

where r_A is the radius of the reservoir *A*.

If $r_j(\phi)$ is r_{eff}, the interface remains in a state of equilibrium until the entry into the capillary and the arrival of the water will be a maximum, and the magnitude of the heave will be

$$V_{g\max} = \frac{\sigma_{iw}}{4\eta \cdot l} \cdot \frac{v_i}{v_w} \cdot \left(\frac{r_c}{r_A}\right)^2 \cdot \frac{r_c^{\,2}}{r_{\text{eff}}} \simeq \frac{\sigma_{iw}}{4\eta l} \cdot \frac{v_i}{v_w} \cdot \left(\frac{r_c}{r_A}\right)^2 \cdot r_c. \tag{23}$$

If $r_j(\phi)$ is less than r_{eff}, the radius of the interface passes through the value of r_{eff}, and it penetrates into the capillary. Under the conditions where $\phi_{T}(q) < \phi$, the position of equilibrium is thus established and the meniscus reaches the reservoir *B*. However, the transition from $r_j = \infty k r_j$[sic] $= r_{eff}$ requires a definite time $\Delta t(\phi)$ during which the non-steady-state heave regime is maintained in the reservoir *A*.

It is assumed that the phase conversion of the water is realized with a stationary piston (variable pressure in the reservoir *A*).

The heat loss ϕ from the system leads to advancement of the ice deposition front in the reservoir *A*. As soon as it reaches the attached water layer, the interface is deformed, a bulge is formed with the radius r_j, and a pressure difference arises just as in the preceding case. The volume *A* remains constant, no movement of the water or heave take place, and ΔP appears in this case as the excess pressure in the ice. The maximum pressure of the heave is

$$\Delta P_{\max} = \frac{2\sigma_{iw}}{r_{eff}} \simeq \frac{2\sigma_{iw}}{r_e}.$$

RESULTS OF INTERACTION BETWEEN THE ICE, THE BOUND WATER, AND THE SOLID SUBSTRATE

If the radius of the capillary is very small, then only bound water will appear in it, and the physicochemical phenomena and not the conditions of equilibrium of the phase interface will acquire predominant significance.

The family of curves with identical equilibrium temperature T_j is presented in Figure 8a.

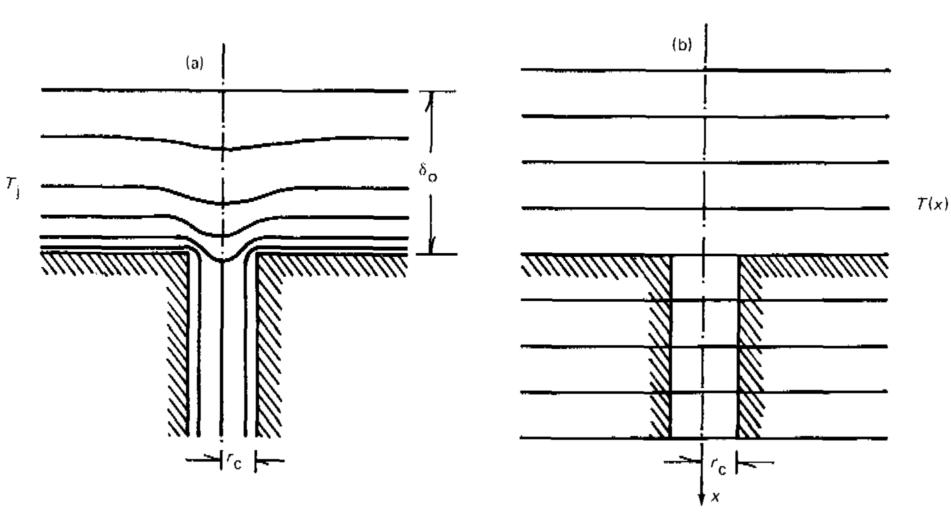


FIGURE 8 Freezing of the water with a small capillary radius: (a) family of curves with identical equilibrium temperature T_j; (b) family of isotherms $T(x)$.

Let us assume that the flux ϕ creates a temperature gradient in the system and that the thermal diffusivity of all the phases (the substrate, water, and ice) is equal; then the family of isotherms $T(x)$ can be represented by Figure 8b, where x denotes the distance from the mouth of the capillary. The combination of the two figures (Figure 8a and 8b) permits us to obtain the shape of the water-ice interface $\delta(x)$; in Figure 9a, we have this interface. The nature of the attached and the free water is different; therefore, it is impossible to begin only with the values of the surface tension; it is necessary to consider the complex interface zone of ice, bound water, and substrate.

This interface zone is equivalent to the interface that is characterized by the energy:

$$f_{iws}(\delta) = f_{w \cdot bw}(\delta) + f_{bw \cdot s}(\delta) + \int_0^{\delta} F(\delta)\, d(\delta).$$

where the bound water is denoted as bw. This function was investigated above (see Figure 3). In Figure 9b, we have the function $f_{iws}(x)$ for the case of the capillary.

Assuming that within the limits of the layer of bound water its density varies little, it is possible to assume that the number of moles of bound water, n_{bw}, is proportional to δ and thus enters into f_{iws}. The chemical potential $\mu_{bw} = \partial f_{iws}/\partial n_{bw}$ is proportional to the derivative for $\partial f_{iws}/\partial\delta$ and its value decreases with δ. In Figure 9c, we have the shape of the curve $\mu_{bw}(x)$.

In this case the migration of the water takes place in the direction from the sections with a high chemical potential to the sections with a low chemical potential, that is, from thick films of bound water to thin films.

If in the investigated system the piston can move freely (the ice pressure is constant and equal to the normal pressure), then the water from the reservoir B goes through the capillary and feeds the ice, which causes it to rise in the reservoir A. The interface approaches the position of equilibrium, which depends on the heat flux reaching the system and the characteristics of the displacement of the bound water in the capillary and also between the ice and the vessel walls.

For a quantity of $\phi = \phi_L = qL/v_W$, the value of q is not determined by Poiseuille's equation, but the properties of the bound water have still been insufficiently studied to evaluate this quantitatively.

If the piston of reservoir A is fixed and rising of the ice in it is not permitted, then the heat flux removed from the system lowers the temperature in the capillary, and the interface is not in the position of equilibrium. The ice in the capillary reaches the reservoir B at the same time as the thickness of the bound water decreases. Here, the chemical potential gradient of the bound water between the ice and the surface of the substrate is compensated for by an increase in pressure in the ice.

What has been discussed indicates the value of the physicochemical effects of the layer of bound water: intake at the interface, dispersement of the water to the freezing front, and heaving. If the medium is rigid, then they lead to the excess pressure in the ice.

The above-indicated method of analysis also permits explanation of the phenomenon of repulsion of the particles observed during displacements of the freezing front.[13]

CONCLUSIONS

It has been demonstrated that the phenomena at the interface in the porous medium leads to the appearance of a pressure gradient and the chemical potential near the freezing front. Depending on the deformability of the medium, the freezing causes displacement of the water to the freezing front and the deposition of ice or excess pressure in the ice; both of these phenomena depend directly on the intensity of heat removal from the system.

If the mechanical conditions (deformability of the medium) permit heaving, then with constant thermal conditions and a defined position of the interface, a steady-state hydrodynamic regime arises with constant ice formation. Under these conditions the heave is proportional to the freezing time. If the thermal regime is not steady state, then the ice formation continues until the freezing limit reaches the capillary.

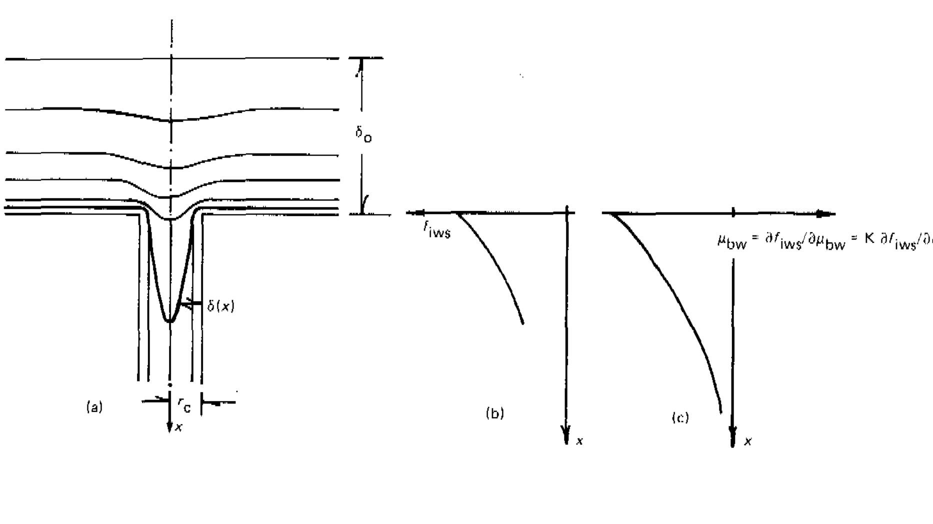


FIGURE 9 Movement of the freezing boundary in the capillary (a). Surface energy of the interface (b) as a function of the distance from the mouth of the capillary $f_{iws}(x)$. Variation of the chemical potential of the attached water μ_{bw} as a function of the distance from the mouth of the capillary (c).

The magnitude of the heave during the time of penetration of the freezing front depends on the shape of the interface and the local heat flux.

The behavior of the grains during freezing depends on the nature and structure of the porous and solid phases and the external effects. The parameters that determine the macroscopic displacement of the liquid in the massif of frozen ground are the permeability and the position of the ground level; they are analogous to the capillary and hydrodynamic properties of the above-investigated models.

Laboratory tests established the relation between the freezing conditions and the manifestations of freezing; the interrelation between the permeability of the ground and the heat flux reaching the phase transformation zone has been confirmed. However, in connection with the complexity of this phenomenon, not all of the questions can be answered at the present time. For example, the frost susceptibility of the dispersed system cannot be characterized by a simple relation or independent parameters. The difficulties of quantitative evaluation of the phenomena arise as a result of lack of knowledge of the properties and behavior of the bound water. However, the scientific approach adopted in the present report permitted explanation of the causes of the variation in structure of the dispersed systems during their freezing.

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SOME CHARACTERISTIC FEATURES OF WATER MIGRATION DURING FREEZING OF FINE SOILS

YE. V. VOLKOVA Moscow Lomonosov State University

The process of water migration during the freezing of fine soils is the result of the effect of a crystallization-diffusion mechanism that takes into account the migration process from the point of view of the molecular-kinetic representations of the water structure and its interaction with the surface of the solid particles.

Under the effect of the active centers of the surfaces of the mineral particles, the structure of the nearest layers of water is distorted, the translational motion of the molecules decreases, and the water properties change, including an increase in the viscosity of the water. The effect of the surface forces is transferred to the subsequent layers of water, attenuating as the thickness of the water film increases.

On a reduction in temperature in the fine soil, part of the water with the least-distorted structure is converted into ice. The rest with more-distorted structure and greater viscosity remains in the unfrozen state.

As a result of self-diffusion from the difference in mobility, displacement of the molecules takes place from the sections with the more mobile molecules (in thick films of the thawed zone) to the sections with less mobile molecules (to the thin films of the frozen zone), that is, from the sections with the higher temperature to the sections with a lower temperature. The water molecules that move into the frozen zone are in excess there by comparison with the number that can be found in the soil in the unfrozen state at the given temperature; therefore, they are grouped into the ice structure.[1,2]

Inasmuch as the migration process is determined by self-diffusion of the water molecules contained in the soil, obviously the nature and intensity of the migration will depend on the degree of mobility of the molecules, that is, the energy state of the water.

It is possible to judge the energy state of the water by the average thickness of the water films and by the relaxation times of the protons of its molecules. Beginning with this, it is possible to try to explain the different nature of the water migration with one-way freezing of the soil such as kaolin, bentonite, and suglinok. Their plasticity properties and the magnitude of the specific surface defined by the adsorption of the water vapor[3] are presented in Table 1.

The investigated soils with different initial moisture contents froze from the top in a closed system with the surface temperature of about -5°C. The test results are presented in Tables 2 and 3.

The suglinok samples with initial moisture content from 9.7 to 29.6 percent froze with a different duration of experiments--from 4 h to 168 h (Table 2, nos. 10-11; Table 3, nos. 5-8). The experiments confirmed the position that in suglinok with a moisture content equal to or less than the plastic limit water migration does not take place in practice in the case of one-way freezing. For example, in the samples with an initial moisture content of 9.7 and 19.0 percent, there was no redistribution of water (Table 2, nos. 10-11; Table 3, no. 8).

With an increase in moisture content from 23 to 29.2 percent, the mean magnitude of the migrational flow with a freezing time of 4 h increased from 0.027 gm/cm^2 to 0.112 $gm/cm^2 \cdot h$ (Table 2, nos. 12-14). With an increase in the freezing time, the moisture content of the thawed zone under the thawing boundary approaches the plastic limit. Thus from Table 3 (nos. 5-7), it is obvious that 4 h later the moisture content under the thawing boundary was 23.9 percent; after 72 h, it was 22.2 percent; and after 168 h, 20.2 percent. After 168 h the moisture content equalized with respect to depth.

A somewhat different kind of water migration was observed in the case of one-way freezing of

TABLE 1 Plasticity and Specific Surface

Type of Soil	Plasticity Limits: Plastic Limit	Plasticity Limits: Liquid Limit	Specific Surface, m^2/g
Kaolin	34.0	52.0	27.3
Suglinok	18.6	34.2	101.0
Bentonite	70.5	98.0	560.0

TABLE 2 Variation of the Migration Flux as a Function of the Initial Moisture Content of the Specimen and the Freezing Time

Sample Number	Type of Soil	Initial Moisture Content, %	Freezing Time, h	Migration Flow, $gm/cm^2 \cdot h$
1	Kaolin	5.9	53	0.004
2	the same	9.9	24	0.003
3	the same	28.9	25	0.010
4	the same	36.4	24	0.045
5	the same	46.1	24	0.120
6	the same	36.6	2	0.204
7	the same	36.1	8	0.122
8	the same	36.1	12	0.084
9	the same	36.2	95	0.019
10	Suglinok	9.7	4	--
11	the same	19.0	4	--
12	the same	23.0	4	0.027
13	the same	27.1	4	0.105
14	the same	29.2	4	0.112
15	the same	28.2	26	0.024
16	the same	27.0	72	0.020
17	the same	29.6	168	0.014
18	Bentonite	96.6	24	0.013

the kaolin. In the kaolin samples redistribution of the water took place for all the investigated values of the initial moisture content--within the limits from 6 to 46 percent. The mean magnitude of the migration flow with an increase in moisture from 9.9 to 46.1 percent increased from 0.004 to 0.120 $gm/cm^2 \cdot h$ (freezing time was 24 h, Table 2, nos. 2-5).

The most intense migration was observed in the first hours of freezing. From Table 2 (nos. 6-9), it is obvious that the mean magnitude of the migration flux after 2 h was 0.204; after 8 h, it was 0.122; after 12 h, it was 0.084; and after 95 h, 0.019 $gm/cm^2 \cdot h$.

The moisture content of the thawed zone of the sample after freezing depended linearly on the initial moisture content. The coefficient of water accumulation, that is, the ratio of the final moisture content of the thawed zone to the initial moisture content of the specimen with respect to all the experiments turned out to be 0.7-0.8 (Table 3).

During the process of one-way freezing of the bentonite with an initial moisture content close to the plastic limit, only insignificant redistribution of moisture was observed. The mean magnitude of the migration flux 24 h after the beginning of freezing was a total of 0.013 $gm/cm^2 \cdot h$ (Table 2, no. 18).

The differences in the nature of the redistribution of the water in the case of one-way freezing of the kaolin, bentonite, and suglinok can be explained if we consider the mean thickness of the water film (n) in the ground and the time of transverse relaxation of the protons of its molecules (T_2) as a function of the soil moisture (Table 4).

From the table it is obvious that in suglinok with a moisture content below the plastic limit of 18.6 percent, the mean thickness of the water film turned out to be a total of 6.5 molecular layers; the transverse relaxation time is low, about 0.55 ms.

In kaolin with a moisture content of the plastic limit (34.0 percent), there are about 38 molecular layers of water characterized by significant relaxation times T_2 = 5.0 ms. Even with a moisture content of 6.9 percent there are still about nine molecular layers of water.

In bentonite with a moisture content below the liquid limit, the mean thickness of the water films is a total of about 6.5 molecular layers; the transverse relaxation time was about 2.00 ms.

Thus, in kaolin the mean thickness of the water films is high. Its molecules are characterized by significant relaxation times; therefore, in the case of one-way freezing, redistribution of water is observed also for moisture contents appreciably less than the plastic limit. Therefore, obviously, in kaolin by comparison, for example, with suglinok, there is rapid equalization of the moisture with respect to height of the specimen.

In suglinok with a moisture content below the plastic limit, the water film six molecular layers thick obviously is so bound that the displacement of its molecules is difficult, and therefore moisture migration practically does not occur.

In bentonite even with a moisture content

TABLE 3 Redistribution of Water with Respect to Depth of the Specimens after Freezing as a Function of the Initial Moisture and the Freezing Time

Number	Type of Soil	Duration of Experiment, h	Depth of Freezing, cm	Coefficient of Water Accumulation	Initial Moisture Content, %	Moisture Content after Freezing with Respect to Sample Depth, cm							
						0-1	1-2	2-3	3-4	4-6	6-8	8-10	10-14
1	Kaolin	8	2.0	0.8	36.1	55.3	58.6	30.8	30.0	30.3	30.4	30.2	
2	the same	24	3.0	0.7	46.1	38.5	130.2		33.6	33.0	33.1	33.0	
3	the same	53	--	0.8	5.9	27.8	7.3	4.7	4.4	4.7	4.2	4.4	
4	the same	143	2.0	0.8	45.2	95.6	127.6	36.0	35.5	35.8	35.6	35.8	
5	Suglinok	4	3.0	--	28.3	40.8	36.4	35.9	23.9	24.6	24.9	25.8	26.6
6	the same	72	4.0	--	27.0	52.3		39.1		22.2	23.0	23.6	24.8
7	the same	168	4.0	--	29.6	49.2		64.9		20.2	20.0	20.2	20.3
8	the same	4	--	--	19.0	18.9	--	18.5	18.9	18.9	18.9	19.0	19.1

TABLE 4 Number of Molecular Layers (n) and Relaxation Time (T_2) as a Function of Moisture Content (w)

Kaolin			Suglinok			Bentonite		
w,%	n	T_2, ms	w,%	n	T_2, ms	w,%	n	T_2, ms
6.0	9.2	--						
34.0	38.0	5.0	18.6	6.5	0.55	70.5	4.6	1.68
52.0	65.5	6.8	32.0	11.5	0.70	98.0	6.5	2.00

below the liquid limit, the thickness of the water film is small, which obviously causes insignificant migration of the water during one-way freezing.

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ICE STRUCTURE AS A FUNCTION OF SALINITY OF THE FREEZING WATER

V. N. GOLUBEV Moscow Lomonosov State University

Crystallo-optical studies of ice formed under natural conditions and during laboratory simulation of the process of freezing of fresh and salt water have demonstrated that the salinity of the water is one of the primary factors determining the structure of the ice.[1,2]

In the presence of dissolved salts, the phase transition of the water to ice is characterized by an entire series of specific features, among which the most important are the following: (1) variation in mobility of the water molecules near the hydrated ions, (2) a reduction in temperature of the phase transition, (3) conversion of the flat surface of the crystallization front to a cellular surface, (4) the formation of interlayers and cells of brine inside the crystals and on the boundary between them, and (5) the migration of the brine and precipitation of the crystal hydrates of the salts dissolved in the water. Here, the variation in mobility of the water molecules and reduction in temperature of the phase transition are most sharply felt in the course of the crystallization process in the stage of occurrence of ice crystal nuclei, whereas the conversion of the flat surface of the crystallization front to cellular and the formation of the cells and interlayers of brine has a defining effect on the growth process of the nuclei already formed and on the crystal habit. The processes of migration of the brine and precipitation of the crystal hydrates of the salts take place primarily in the last stages of hardening of the ice formation, and they must be considered more processes of syngenetic and epigenetic metamorphism of the ice and not crystallization processes. When studying the dependence of the ice structure on the salinity of the freezing solution, the investigation of the first two groups of factors having a direct effect on the occurrence and growth of the ice crystals appears to be the most interesting.

According to the kinetic theory of liquids,[3] the rate of occurrence of crystallization centers in the liquid as a function of supercooling ΔT is described by the equation:

$$I = Ce^{-\frac{\Delta U}{kv_0 - \Delta T}} e^{-\frac{4\pi\sigma}{3k(T_0 - \Delta T)}\left(\frac{2\sigma T_0}{\rho_i \lambda \Delta T}\right)^2}, \quad (1)$$

Where C is the integration constant, ΔU is the activation energy of self-diffusion of the water molecules, k is the Boltzmann constant. σ is the surface tension at the ice-water interface, ρ_i is the ice density, and λ is the latent heat of fusion.

T_0 is the temperature of the phase transition. Here the first exponential factor is a function inverse to the dependence of the viscosity η of the liquid on the temperature: $1/\eta = C_1 e - \Delta U/kT$, and the second factor determines the number of critical nuclei of the new phase as a function of ΔT: $Ng = C_2 e^{-4\pi\sigma/3kTr_\star^2}$, where the critical nucleus radius $r_\star = 2\sigma T_0/\rho_i\lambda\Delta T$.

The variation of the structural properties of the water in the prisms of dissolved salts finds its expression in the variation of an entire series of physical parameters, including the values directly determining the rate of occurrence of the crystallization centers: namely, η and T_0. Thus, an increase in the NaCl concentration in the water from zero to 60 percent leads to an increase in viscosity by about 10 percent. At the same time T_0 drops from 273° to 269°K, which also leads to an additional increase in viscosity at the time of formation of the ice crystal nuclei. As a result of the added effect of these factors a viscosity of 60 percent of the solution with equal ΔT turns out to be almost 40 percent higher than the viscosity of fresh water. Thus, as a result of an increase in the viscosity of the solution with an increase in its salinity, the rate of occurrence of the crystallization centers must decrease. However, the second exponential factor varies proportionally to e^{-T_0}, that is, a result of a reduction in the phase transition temperature with an increase in salinity, Ng, and together with it also I must increase. For analysis of the relation between these two opposite trends, let us differentiate I with respect to T_0, and let us substitute the numerical values, setting in this case ΔT so small by comparison with T_0 that $T_0/T_0 - \Delta T_0) \cong 1$. Then when $dI/dT_0 = 0$, $T_0 \cong 3\Delta T$. According to the condition adopted by us, $T_0 >> \Delta T$. However, in this case $dI/dT_0 < 0$, that is, for identical values of ΔT and the condition $T_0 > 3\Delta T$, an increase in the salinity of the solution must lead to an increase in the rate of occurrence of the crystallization centers. The numerical solution of Equation (1) in the range of values of T_0 = 273-269°K and ΔT = 0-10° shows that with variation of the salinity from zero to 60 percent, I increases almost rectilinearly. The relative increment in I decreases somewhat with an increase in ΔT. If we consider that the number of crystals in the ice body is proportional to the number of crystallization centers that arise, then in the ice formed from salt water the number of crystals per unit volume N will be greater, and the volume of the average crystal will be smaller than in fresh ice formed with analogous supercooling.

Immediately after occurrence, the growth of the ice crystal nuclei takes place with an initial rate v_0, which depends on the degree of supercooling of the solution and the diffusion factor of the water molecules in it. Since the isomorphic intrusion of the salt ions into the crystal lattice of the ice does not take place, during the crystal growth process its faces reject the ions and molecules of the dissolved salts, and a zone with increased salt concentration is created in front of the crystallization front. Part of the salt from this zone is carried away by diffusion; however, the concentration in the layers following the boundary layers increases as a result of which, with a constant growth rate of the crystal, the salt concentration in the boundary zone must gradually increase with time.

However, the growth rate of the crystals as the salt concentration in the boundary layer of the liquid grows is retarded, and the diffusion of the salt from this layer as a result of an increase in the concentration difference of the salt in the boundary layer and the volume of the liquid rises. Between these two processes, after a time t from the beginning of growth or, in other words, at some distance x from the point of occurrence of the crystal, dynamic equilibrium sets up after which under invariant thermodynamic conditions the crystal growth takes place with a finite rate v_t', in the first approximation proportional to the diffusion coefficient of the salts D and the concentration gradient $\Delta S/e$ and inversely proportional to the concentration of the solution S_0:

$$v'_t = D\,\frac{\Delta S}{e}\cdot\frac{1}{S_0}\,. \qquad (2)$$

During the growth of the initially occurring crystals in the space between them under the condition of removal of the heat of crystallization, conditions are created for the occurrence and growth of new crystals. The relatively fast growth of these crystals continues until the formation of concentration zones, after which further growth of the crystals takes place with a steady state rate v_t''. Since as the occurrence and growth of the crystals takes place the concentration of the unfrozen part of the solution increases, the steady-state growth rate of the crystals v_t decreases, and the rate of occurrence of the nuclei of the new crystals at ΔT = const increases. Then in the ice formed from the water of greater salinity, the number of crystals N must be greater and the size of the average crystal smaller. Only as a result of a change in v_t should the volume of the average crystal V in ice of different salinity vary proportionally to S_0^{-3}. The dimensions of the individual crystals must also differ from each other, both as a result of difference in time of formation and as a result of a decrease in v_t during the crystallization process. The experiments demonstrated that the greatest variety of crystal dimensions is observed in ice formed from less-concentrated solutions, which agrees with the above-stated theoretical principles.

For very small ΔT = 0.01-0.1°C and correspondingly small $v_0 = 10^{-5}$ to 10^{-7} cm/s, between the process of the increase in concentration in the boundary layer of the solution as a result of crystal growth and the process of reduction of it at the expense of salt diffusion, there is dynamic equilibrium as a result of which the formation of the concentration zones does not take place or takes place very slowly, and the reduction in the growth rate of the crystals v_0 during crystallization is connected only with an increase in salinity of the unfrozen part of the solution. In this case with variation of the salinity of the initial solution S_0, the number of crystals N must vary proportionally to the function I of T_0, that is, an increase in N close to linear must be observed with an increase in S_0.

The results of the experimental studies with respect to freezing of the water of different salinity with different supercooling of it indicate the nature of variation of N as a function of S_0 (see Figure 1). For small supercooling of the water and, consequently, small v_0, the increase in N with an increase in S_0 has almost a linear nature (curves 1, 2, and 3) corresponding to the nature of the dependence of I on T_0 (curve A). As ΔT increases, the dependence of N on S_0 becomes stronger, acquiring an exponential nature with the exponent on S_0 varying from 1.5 to 3 with an increase in ΔT from 0.2° to 3°C. This indicates that with an increase in ΔT the process of the formation of concentration zones and, as

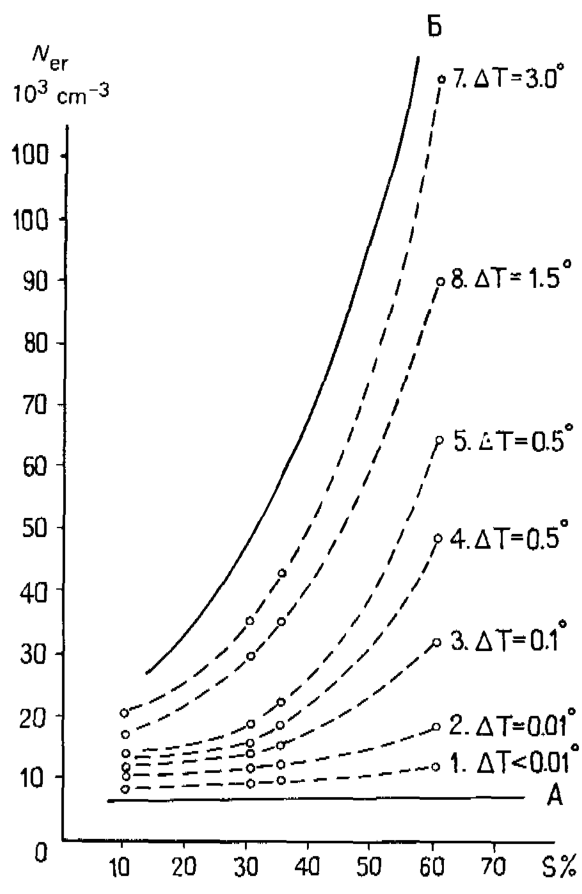

FIGURE 1 Variation of the number of crystals per unit volume of ice N_{cr} as a function of the salinity of the water S_0 for different degrees of supercooling ΔT.

a consequence, the rate of steady-state growth of the crystals v_t acquire a more and more important role as the structural-controlling factor.

The growth of the crystals stops at the time when the concentration zones of the individual crystals unite, and the concentration of the solution inside these zones reaches S_p, which is in equilibrium with the ice for a given temperature. Here, the two-phase system is formed, which comprises crystals of practically fresh ice and concentrated salt-brine solution, that is,

$$v_0\rho_0 = v_i\rho_i + v_p\rho_p, \tag{3}$$

where ρ_0, ρ_i, and ρ_p are the density of the initial solution, the ice, and the brine, respectively; v_0, v_i, and v_p are the volumes of the initial solution, ice, and brine where $v_p = v_0\rho_0 S_0/\rho_p S_p$. Since the thickness of the concentration zones and the salt concentration within them vary relatively little, the thickness of the intercrystalline interlayers of brine d in the same ice body must be approximately identical, and practically all the brine must be distributed more or less uniformly with respect to the crystal surface. The crystallo-optical studies of the ice samples demonstrated that only 5-10 percent of the brine is concentrated in the cells located primarily at the junction of 3 to 4 crystals, whereas the remaining 90-95 percent of the brine is uniformly distributed with respect to the intercrystalline interlayers. The differences in thickness of the interlayers usually does not exceed 10 percent.

Between the total surface of the ice crystals per unit volume P^{-1} cm and the volume of the idealized crystal there is a relation $P = 6v_i V_{cr}^{-3} = 6v_i^{2/3}V^{1/3}$. Under the condition that migration and leakage of the brine do not occur during the crystallization process and that the brine is uniformly distributed with respect to the intercrystalline interlayers, we obtain:

$$v_p = P\frac{d}{2} = 3v_i^{2/3}N^{1/3}. \tag{4}$$

Substituting the values of v_i and v_p in (4) from (3) and solving it with respect to N, we obtain the expression for N as a function of the salinity of the solution:

$$N = \left(\frac{1}{3d}\cdot\frac{S_0}{S_p}\right)^3\left[-\frac{\rho_i^{\,2}\rho_0}{\rho^3_{\,p}(1 - S_0/S_p)^2}\right], \tag{5}$$

or since part of expression (5) enclosed in brackets differs little from 1,

$$N \cong \left(\frac{1}{3d}\,\frac{S_0}{S_p}\right)^3, \tag{5a}$$

the relation obtained is analogous with respect to its structure to the dependence of N on S_0 following from the investigation of the processes of ice crystal growth occurring in solutions of different salinity for significant ΔT. Curve B

in the Figure 1, which represents expressions (5a), is quite close to the experimental curves 6 and 7 for N as a function of S_0 in ice formed from solutions with a salinity of 15-60 percent at ΔT = 1.5-3.0°C. Some difference in the calculated values from the experimental data is explained by the deviation of the form of the real crystals from the idealized crystal and the inclusion of part of the brine in the cells located both within the individual crystals and at the junction of three or four crystals. In the ice formed with small ΔT, the relations for N as a function of S_0 differ significantly from the function (5a), which indicates weakness of the effect of the process of formation of the concentration zones on the growth and occurrence of the crystals, and it can serve as an indirect indication of the possibility of significant redistribution of the brine when ΔT = 0.01-0.1°C.

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ICE SUBLIMATION IN FINE-GRAINED SOILS OF VARIOUS CRYOGENOUS STRUCTURE DURING THEIR INTERACTION WITH AN AIR CURRENT

E. D. YERSHOV, V. V. GUROV, AND B. N. DOSTOVALOV
Moscow Lomonosov State University

The process of the sublimation of ice in fine-grained soil has been studied little, although its role in the understanding of the physical principles of the processes occurring in frozen soils is difficult to overestimate (for example, with respect to phase transformations and the laws of migration of soil moisture). The practical importance of studying this process is determined by significant variation of the strength and rheologic characteristics of the soil during its freezing-desiccation.

The study of the laws of the sublimation process of pure ice and ice in fine-grained soil is being made at the Department of Geocryology of Moscow State University using a laboratory device that permits many months of experiments with samples of different soils of 10-fold repetition under the conditions of uniformity of the process and with variable parameters of the air current (velocity, temperature, and relative humidity). The procedure used in performing the experiments, and processing the data has been presented.[1,2]

The microscopic study of the surface of the ice sample demonstrated that the development of the sublimation process takes place nonuniformly from the entire surface of the ice and only in the active centers of sublimation, which are the defects in the crystal structure of the ice and characterized by the presence in them of associations of more mobile water molecules. The ice surface is spotted with shallow cuplike depressions at the nucleation points in time. Subsequently, they become deeper, wider, and partially merge. This leads to the formation of disjointed microrelief on the surface of which the nucleation and development of the small depressions is again traced. With time, obviously, the ice surface undergoes similar elementary development cycles. The experimental data from studying the intensity of the sublimation of the ice (I_c) with time (Figure 1a) agree with the indicated laws of development of the sublimating surface. The analogous results were obtained for small-, medium-, and large-crystal ice formed at temperatures of -5°, -50°, and -150° respectively. The difference consists only in the fact that with a decrease in the dimensions of the ice crystals, that is, with an increase in the lattice defects, the elementary cycle of development of the microrelief of the ice surface accelerates.

By comparison with pure ice, the ice-sublimation process in fine-grained soils turns out to be appreciably more complicated, inasmuch as it is necessary to assume that the ice sublimation is evaporation of a quasiliquid film located between the surface of the ice crystals and the mineral particles with constant filling of the pores by ice.[2] A confirmation of this is the results of the studies of a number of authors.[3-6] It is possible to state that in fine-grained soils the intensity of the ice sublimation is determined by two components: the moisture flows in the liquid and gas phases. This agrees with the experimental data of the existence of a gradient of the unfrozen moisture in the zone of clay soil desiccated during the sublimation process in contrast to the sandy zone where this process is absent, and the moisture transfer is wholly determined by the vapor flux (Figure 2). The special studies have demonstrated that in clay a portion of the vapor transfer in the general moisture flux through the sample is 50-80 percent, and 20-50 percent goes to the proportion of the migration of unfrozen water.

An experimental study of the parameters of the ice sublimation process in various geological-genetic types of soil (with a degree saturation with ice close to 1 demonstrated that the value of I_C increases with an increase in the content in the soil of clay particles and minerals of the kaolinite group and with a decrease in the dry unit weight (Figure 1a). This law is explained by an increase in porosity and the area of the ice and unfrozen water in the transverse cross section of comparative soil samples[2] and also a smaller magnitude of the bond energy of the moisture to the surface of the kaolinite-group minerals.

The temperature of the air interacting with the ice (Figure 1b) has a significant effect on

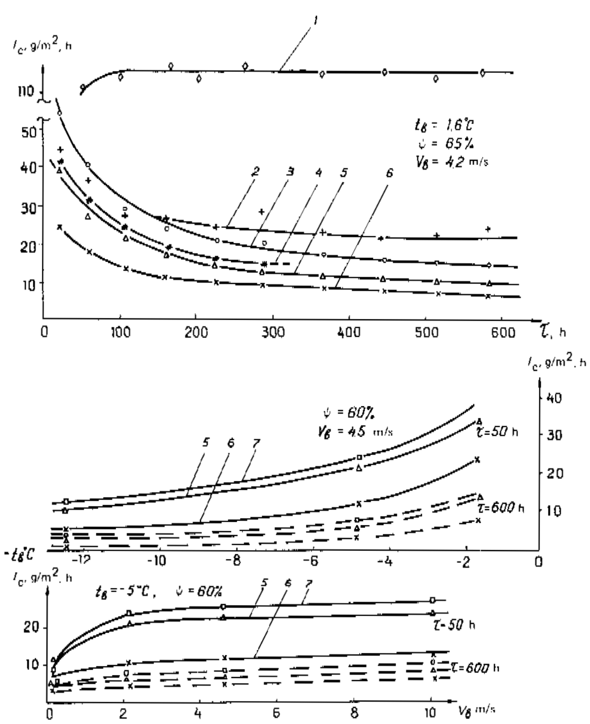

FIGURE 1 Intensity of the ice sublimation as a function of the mineralogical and granulometric composition of the soil (a), the temperature t_a (b), and the velocity V_a (c) of the air flux: [S_a is relative humidity]. 1--ice; 2--Glukhov kaolin; 3--Oglanly bentonite; 4--hydromica clay; 5--paleogenic (Kiev) clay; 6--Moscow supes; 7--Moscow suglinok.

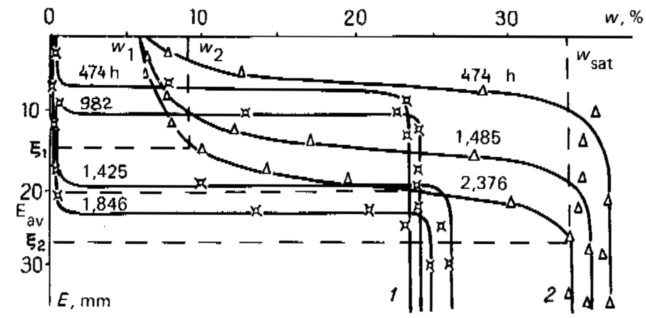

FIGURE 2 Distribution of the total moisture content with respect to height of the soil samples during the process of sublimation of the ice in sand (1) and in clay (2) for t_a = -5°C, S_a = 70 percent, and V_a = 0.

the intensity of the ice sublimation in the soil. This is related to the fact that with a reduction in temperature, along with a decrease in the partial pressure gradient of the water vapor and the amount of unfrozen water in the soil, the bond energy of the quasiliquid film to the ice surface and the soil particles increases.

A study of the magnitude of I_C as a function of the air flow rate demonstrated that the air flows with a velocity of 3-5 m/s have the greatest effect on the sublimation intensity (Figure 1c). Further, increase in the flow rate leads to only insignificant variation of the value of I_C. It must be noted that with time the desiccation zone increases and the intensity of ice sublimation decreases as a result of a decrease in the thermal conductivity of the upper layer of the soil and a corresponding decrease in the heat flux to the sublimation zone (Figure 1b, c--dotted curves).

The effect of the relative humidity of the air (S_a) on the magnitude of I_C was investigated in the range of S_a = 50-100 percent. When S_a = 100 percent, no sublimation of the ice was observed. A comparison of the experimental data permits the conclusion that I_C is proportional to the gradient of the partial pressure of the water vapor in the desiccation zone.

The dynamics of the moisture field of the soil samples during sublimation of the ice were investigated by successive division of repeated specimens with respect to height and determination of the total moisture (w) in them. With respect to nature of distribution of w in the soil, it is possible to isolate the zone that has been desiccated during the process of sublimation of the ice (ξ_2 thick) and the zone that has not been desiccated in which $w = w_{init}$ = const (Figure 2). The first of them, in turn, can be divided into a region with small (ξ_1 thick) and large ($\xi_2-\xi_1$ thick) gradients [of] w. The visual sublimation front recorded by the variation in color of the ground in the sample practically coincides with the magnitude of $\xi_{mean} = 0.5$ ($\xi_1 + \xi_2$). In accordance with the indicated zones and sublimation regions, it is possible to isolate three critical zones in the soil with respect to their moisture content:

w_1, w_2, w_{init}. These values of the moisture, which practically remain constant with time, are determined by the composition and properties of the soil and also the parameters of the air flow.

For a quantitative evaluation of the effect of a number of factors on the advancement of the sublimation zone (ξ_{av}) in the frozen ground, the following approximate formula is proposed

$$\xi_{av} = \frac{\int_0^{\tau} I_c d\tau}{\gamma_d [w_{init} - 0.5 (w_1 + w_2)]},$$

where γ_d is dry unit weight. It was derived on the basis of determing the amount of moisture lost by a unit area of the given soil as a result of sublimation of the ice in it during the time τ.[2] This formula was checked with the experimental data and demonstrated sufficient accuracy for practical calculations.

A study of the physicomechanical properties of the soil demonstrated that their strength is reduced during the process of ice sublimation. In the case of massive cryogenous texture, the loss of strength is greater in the coarse-grained soils than in the fine-grained soils. The soil with reticular, laminar, and laminar-reticular cryotexture loses its cohesiveness on the removal of ice during the sublimation process, and it disintegrates into blocks of desiccated soil.

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APPLICATION OF OPTICAL METHODS IN STUDYING THE STRUCTURE OF FROZEN SOILS

KY. ZIGERT Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

Optical studies of the microstructure of frozen soils have been limited up to now by the fact that researchers[1-5] have used exclusively petrographic sections by which it is possible to study only comparatively large (>0.1 mm) ice inclusions. However, it is not to be doubted that the study of the microstructure of the ice cement of fine-grained soils also has primary significance, since the nature of the distribution of the ice-cement has a significant effect on the mechanical properties of the frozen soil and, in addition, the data on the structural peculiarities of the ice-cement in frozen soils can be used to solve the problem of their genesis.

In order to study the microstructure of the ice cement in a fine-grained soil, it is necessary to use other optical methods. In our opinion the most promising is to study polished sections of frozen soil in reflected light by the dark-field method and also the application of the replica method (surface impressions).

STUDY OF POLISHED SECTIONS OF FROZEN SOILS IN REFLECTED LIGHT BY THE DARK-FIELD METHOD

Smooth polished surfaces of ore minerals and metals are studied in reflected light using ore or metallographic microscopes. Recently this method has been used also to study the microstructure of various silicates.[6]

According to P. A. Shumskiy,[5] polished sections are not suitable for the study of frozen soil under a microscope, since with ordinary direct illumination the ice quickly evaporates and the polished surface is destroyed under the effect of the light beam. This can be avoided if we use the dark-field method.[7]

In the case of illumination of the surface by the dark-field method, the light is directed at the sample at a particular angle. The light reflected by an ideal mirror surface, that is, a structureless object, is not returned to the objective, and the field of view remains dark. Objects of a polished section with high reflectivity having grain boundaries, growth zones, microcracks, and so on give an image in the form of fine light lines against a dark background. When studying poorly reflecting materials, the light penetrating into them is reflected uniformly by the boundaries, the cracks, inclusions, and so on. Thus, the researcher has the possibility of observing the internal structure of transparent minerals.

A significant advantage of the dark-field method in reflected light by comparison with the direct illumination method (the light-field method) consists in the fact that it gives a clear colored image, even from an unpolished surface.

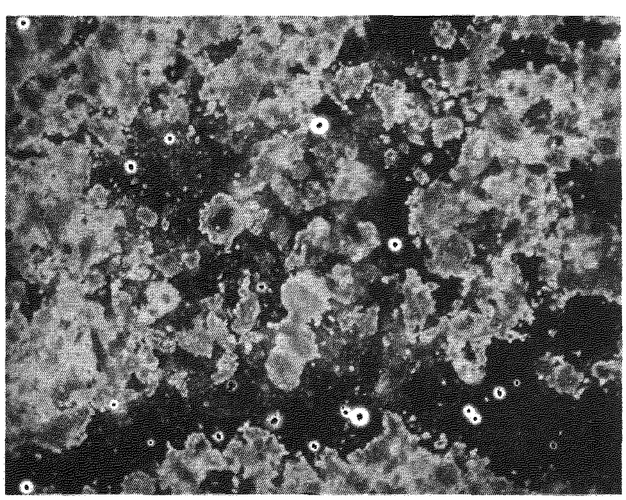

FIGURE 1 Frozen clayey aleurite [silt]. The ice (dark sections) is contained in the form of bunched basal cement. The mineral particles are light gray. Polished section, reflected light, dark field (160 magnification).

In Figure 1 we have the structure of frozen clayey aleurite in reflected light with illumination by the dark-field method. The sections of ice turn out to be almost structureless: They remain dark. On the ice surface small light crystals of sublimation ice are noticeable, which occur during the course of studying a specimen under a microscope. The particles of the mineral skeleton are light gray as a result of numerous internal reflections.

The preparation of the polished sections was carried out in the following way: Specimens about 2 × 2 × 3 cm were cut out of the frozen soil. The surface of interest was ground by emery cloth while cooled by carbon dioxide or by liquid nitrogen. After this the specimen was held for several hours in the air. As a result of evaporation of a thin surface layer, the scratches on the ice, which unavoidably occur during polishing, are smoothed out. The preparation of the polished sections and also their study under a microscope were carried out in an ambient negative temperature (-3°, -5°).

STUDIES OF FROZEN SOILS BY THE REPLICA METHOD

The replica method of studying specimens on an optical microscope was first used by V. Schaefer[8] in order to discover the microstructure of metals and alloys. Later it found application in studying various technical silicate materials[6,9] and also ice.[10,11]

The essence of the method is that, by using an organic polymer solution, castings (replicas) were taken from the surface of the specimen, recording all the fine detail of its structure.

Polymers dissolved in dichloroethane are suitable for making replicas of ice and frozen rock. Formvar,[10,11] polystyrene,[12] and, according to our experience, triafol are such polymers. The other solvents widely used in the manufacture of replicas for electron microscope research like amyl acetate and acetone do not give positive results, since they dissolve the ice.

We followed the example of R. V. Maksimyak[12] and used a solution of polystyrene in dichloroethane. The replicas were taken from differently prepared surfaces. When studying soil with a massive structure, a fresh fragment was used, which, after cleaning to remove adhered particles, was held for several hours in the air. As a result of evaporation, surface relief and ice-grained boundaries are clearly revealed. The soil surface of recticular or layered cryogenous texture was prepared just as for studying it in reflected light by the dark-field method.

A solution of polystyrene having the same negative temperature as the sample was applied to the prepared horizontal surface. The viscosity of the solution was selected experimentally as a function of the nature of the specimen. The solution must completely fill the surface relief of the specimen, but not envelop the individual particles. The film hardens after 5 to 8 h. During this time the specimen was held at a negative temperature. If the surface was prepared qualitatively and the viscosity of the solution was selected correctly, the film was easily separated from the specimen. Mineral particles always adhere to the replicas of frozen soil.

If they could not be washed away in water, then for removal of the particles the replica was first placed in hydrochloric acid (1:1) and then in concentrated hydrofluoric acid. The prepared replicas were placed with the impression up on a slide. The space between the replica and the slide was filled with water or other weakly refracting liquid. If the replica is very thin, then it is easily distorted. Therefore, it is expedient to cover it with a cover glass and again fill the space between the film and the slide with water or other liquid.

Under the microscope in the transmitted light, the replicas give a clear image of the structure of the soil (Figure 2). The mineral grains and aggregates are distinguished by a very uneven surface: The ice has a significantly lower relief and smooth surface, and the grain boundaries of the ice stand out as thin lines.

The primary advantage of the replica method when studying frozen soils consists in the fact that the finished replica can be kept for an unlimited long time, and it can be studied under a microscope at room temperature. In contrast to the polished sections, the replica usually covers a smaller area. Therefore, the combination of both methods is expedient. In reflected light it is possible to obtain information about the quantity and nature of distribution of the ice and mineral grains. The replica method permits a more detailed study of the structure of the ice itself in the soil. In addition, when necessary the polystyrene replica serves as a base for preparing two-stage replicas that can be studied under an electron microscope.

Both methods are advantageously distinguished from the examination of petrographic sections by the fact that the manufacture of the specimens from the frozen soil for the application of these methods is less difficult and offers the possibility of studying a large number of samples.

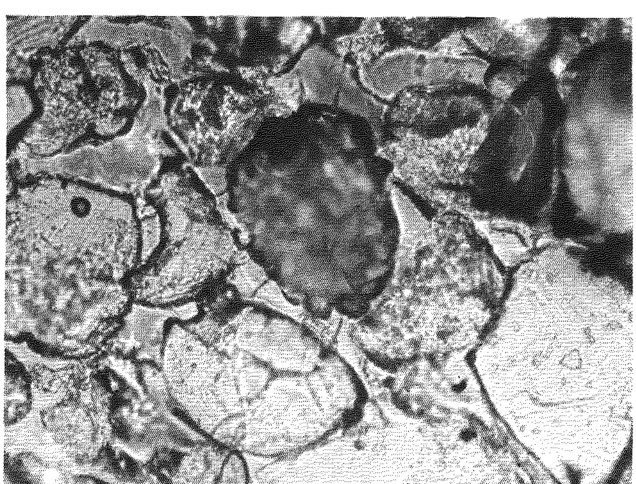

FIGURE 2 Frozen fine-grained sand. Ice (low relief) forms pore cement with hypidiomorphic-grain structure. Mineral particles of high relief and an uneven surface. The polystyrene replica transmits light (magnification 212).

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APPLICATION OF ULTRASONICS FOR EVALUATING THE PHASE COMPOSITION OF WATER AND STRENGTH CHARACTERISTICS OF FROZEN SOILS

YU. D. ZYKOV, A. D. FROLOV, AND YE. P. SHUSHERINA Production and Scientific Research Institute for Engineering Exploration in Construction, Moscow Geological Exploration Institute, Moscow Lomonosov State University

The strength characteristics of frozen soils are determined primarily by tedious static tests. Therefore, the development of faster and cheaper methods of determining these characteristics is highly urgent. One of the prospective areas of the given field is the application of methods based on determining the propagation rate of the elastic waves in frozen soils.

When freezing the liquid phase, the monolithic nature of the soil increases, which causes corresponding changes in the mechanical properties that can be recorded by the variations in propagation rates of elastic waves.[1] However, for realization of these possibilities it is necessary to establish the relations of the velocities and dynamic elastic moduli of the cryogenous materials to their strength characteristics.

Ultrasonic laboratory measurements of the acoustic and elastic characteristics performed on frozen sand, heavy suglinok, and kaolin; the study of their rupture and compressive strength; and also determination of the unfrozen water content permit establishment of the defined relations among the parameters.

All the experiments were performed on artificially prepared frozen soil specimens with massive cryogenous texture for different stages of saturation at temperatures to -40°C. The elastic moduli were calculated from the propagation rates of the longitudinal (V_p) and Rayleigh (V_R) elastic waves measured using the procedure of longitudinal ultrasonic sectioning.[2]

The determinations of the unfrozen water content by the calorimetric procedure, and strength characteristics on presses, were made at the laboratory of the Geocryology Department of Moscow State University.[3,4]

Let us consider some of the experimental results. In Figure 1 we have the propagation rate of longitudinal elastic waves (V_p) and the Young's moduli (E) as functions of the unfrozen water content [w_u] in the frozen saturated kaolin. The parameter of each graph is the moisture content [w] of the specimens, which varied from 25 percent (sample 1) to 149 percent (sample 5). As follows from Figure 1, within the limits of the investigated range, the unfrozen water content corresponds to changes of temperature of the sample from approximately -1 to -15°C; for kaolin an inverse proportional relation exists for the velocity V_p and the Young's moduli as functions of w_u. Here, the greater the total

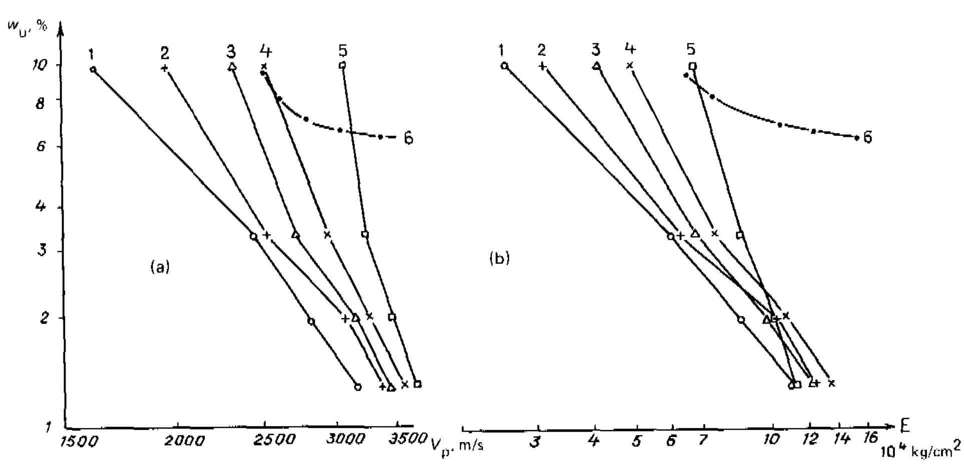

FIGURE 1 Velocity V_p (a) and Young's modulus E (b) as functions of the unfrozen water content w_u in the frozen saturated kaolin. 1-5--kaolin of various moisture contents: 1--25 percent; 2--35 percent; 4--63 percent; 5--149 percent; 6--suglinok. w = 25 percent.

moisture (ice) content of the soil, the more weakly is the investigated function expressed, since more of the proportion of specific weight of the rock is ice. Thus, the relation of the acoustic and elastic characteristics of the frozen soil to the content of unfrozen water in it depends essentially on the total moisture content of the soil, which must be considered when developing acoustic methods of estimating the phase composition of the water. Here it is necessary to note that the data we obtained corresponding to the natural negative temperatures characterize only a comparatively small section of the graph. In fact this function is not necessarily rectilinear in the entire possible temperature range. This is indicated by the data obtained for heavy suglinok consisting to a significant degree of montmorillonite and containing about 6 percent water in the unfrozen state, even at -60°C.[1]

In Figure 2 we have graphs characterizing the

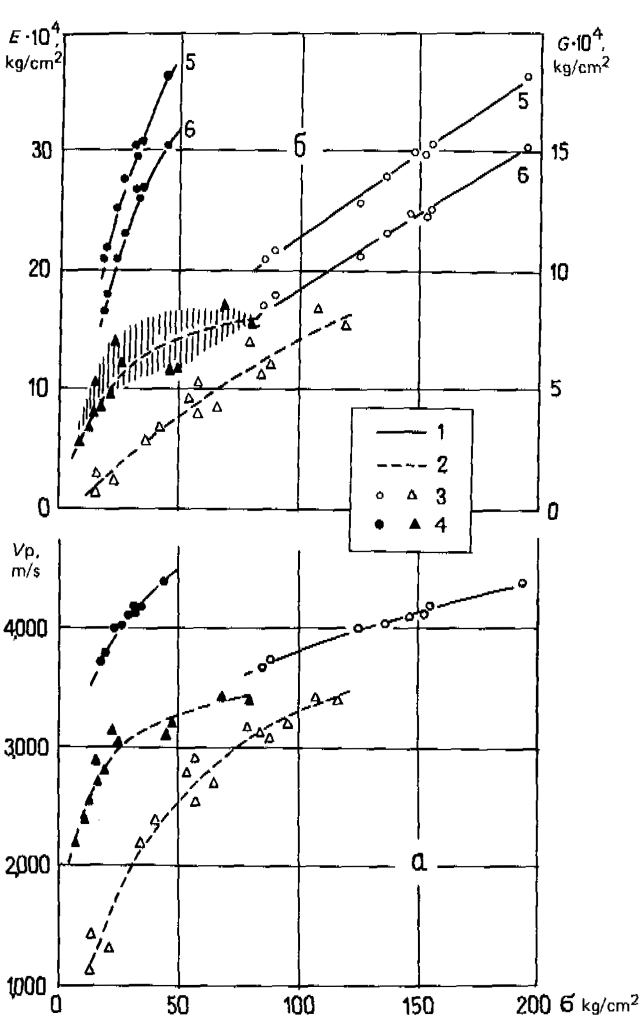

FIGURE 2 Relation of the ultimate compressive and shear strengths to the velocity V_p (a), Young's modulus E, and the shear modulus G (b) for frozen sand and suglinok. 1--sand; 2--suglinok; 3--compression; 4--shear; 5--Young's modulus, E; 6--shear modulus, G.

relation between the velocity V_p and the moduli E and G with the compressive and rupture strengths for the specimens of the frozen heavy suglinok and sand with a porosity of about 40 percent and varying degrees of saturation at temperatures of -10°C to -40°C. The points characterized by the small values of the strength correspond to specimens with the lowest degree of saturation (on the order of 50 percent). Just as should be expected, the frozen soil having high rigidity, that is large values of E, G, and V_p (the degree of saturation, with ice, is about 100 percent and temperatures are low) has maximum strength. However, the graphs indicate the significant differences in the investigated relation for soils of different lithology.

Thus, in the case of compression of the saturated specimens at -40°C, the strength of the sand is about 1.5-2.0 times greater than the strength of the suglinok. The values of V_P, E, and G, characterizing the rigidity of the rock, are as many times greater. At the same time, the ultimate shear strength for analogous specimens of sand is approximately half that for suglinok, although the relations of the rigidities of these rocks remain as before. For rock with low rigidity and, correspondingly, low strength, the noted direct proportionality between the compared variables in the case of compression turns out to be valid also for rupture.

This indicates the different mechanism of the processes of deformation and rupture of frozen soils of different lithology, which is caused by the specific nature of the phase transitions.

In conclusion it is necessary to note that the data presented, in spite of the limited experimental material, indicate the presence of an explicit relation between the strength characteristics of the frozen soil, Young's modulus and shear modulus, and, consequently, the propagation rates of the elastic waves. Thus, there is a real possibility of estimating the strength characteristics of frozen deposits by the results of ultrasonic and seismic data.

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THEORY OF TWO-PHASE POROUS MEDIA

A. V. KARAKIN Earth Physics Institute

This paper contains a discussion of a number of phenomena connected with the deformation of the crystal lattice of ice, the ice-water phase transitions and percolation of thaw water known in the literature as the Tyndall regelation phenomenon. Essentially it lies in the fact that at temperatures close to the melting point, under the effect of external pressure, ice melts and the thaw water is forced into the low-pressure region and refreezes. Externally, the phenomenon somewhat resembles the process of viscous deformation, except that the flow characteristics depend on the temperature and spatial distribution of the flow of the thaw-water. Let, for example, a wire loop on which a weight is suspended be thrown over a block of ice. The wire will pass through the body of the ice without disturbing its intergralness. If the places where the wire comes out of the ice are split, and the thaw water is allowed to run out of the deformation zone, then the rate of passage of the wire with the same load is significantly (about an order) retarded, inasmuch as the heat is removed along with the thaw water. If the holes are filled with ice and at the same time the thaw water is allowed to filter out of the region of increased pressure into the region of reduced pressure and freeze in the deformation zone itself, then the ordinary flow rate is established. The facts of thawing, efflux, and freezing of the water are observed visually in certain cases and they are confirmed by the characteristic arrangement and morphology in the freezing zone. A survey of the experimental facts and references to the original works can be found in Weinberg *et al.*[1]

In the experiments on extrusion, the ice in a cylinder was subjected to compression using a piston that moved from top to bottom. The extrudate, the transverse cross section of which matched the shape and size of the hole, was forced through the hole. If the water was basically moved along the ice-vessel boundary and freezing could occur on the surface of the ice (as is usually assumed when explaining regelation[1]), then the extrudate would expand near the hole and would not grow in length. Actually, the process would take place in such a way that the freezing would be localized in the cylinder, in the inside points of the medium near the hole. This is indicated by the characteristic arrangement of the crystals, which would be forced through the hole and, as a result of this, ruptured in a defined way and oriented. This permits us to conclude that the regelation occurs primarily at points inside the medium.

In the tunnel under "Blue Ice," La Chappelle[2] noted that the glacier flows around large (on the order of a meter or more) unevennesses of the underlying surface using the mechanism of viscous deformation, that is, shift of the ice layers parallel to the outer boundary of the glacier along the basal planes of the crystals was observed. In a section of the sample, these layers were obviously visible. For small sizes (less than 10 cm), thawing of the bottom protrusions through the glacier was observed. Consequently, when we have the characteristic length, $L_R \approx 50$ cm; when $L \gg L_R$, only viscous deformation of the ice lattice takes place; and when $L \ll L_R$, pure regelation (without viscous deformation). The existence of this dimension L_R can explain the infiltration of the thaw water through the pores and cracks and through the ice, which also have their characteristic dimension l_R. The ratio L_R/l_R must be related to the physicochemical constants of the medium.

The basic equations of regelation based on the representations of the mechanics of a two-phase heterogeneous medium, the general principles of the construction of which are discussed in the literature,[3,4] are formulated below.

Let us make the following assumptions: The ice and water will be considered uniform and isotropic media; water is a Newtonian liquid with a viscosity of 2 cp, and ice is a visco-elastic medium with a viscosity $\approx 10^{12}$-10^{16} poise and a relaxation time of ≈ 10-10^5 s; and the water moves through the pores and fissures in the ice. On the basis of the indicated difference in the viscosities and also the small relative dimensions of the pores for water, it is possible to use the concepts of the mechanics of a liquid percolating through a porous solid or elastic body according to Darcy's law. Within the framework of the hypothesis of local uniformity, the two-phase medium is assumed to be continuous and homogeneous, that is, the ice is polycrystalline, free from vapor air and other inclusions. In the equations of motion, we neglect the inertial and external forces. It is proposed that the volumetric proportion of the water is small, that is, $\alpha \ll 1$. Let us consider the elementary volume of the continuous two-phase medium dV, the linear dimensions of which are much greater than the dimensions of the pores but less than the dimensions of the paths of ice flow. We shall consider that the boundaries of this volume are "frozen" into the crystal lattice of the ice and deformed together with it. The idea of the hypothesis of local uniformity consists in the fact that the averaging of all the phase characteristics is carried out with respect to this volume. Correspondingly, we shall distinguish the true, mean, effective, and physical variables in this volume. The meaning of the true values of each of the phases follows from the definitions. They are related by the known rheologic equations. The mean values are obtained as a result of averaging the true values with respect to volume of the given phase within the limits of the elementary volume dV. Averaging permits simplification of some of the relations; for example, instead of the tensor equation of state for the water we obtain Darcy's law. The effective values correspond to a fictitious homogeneous medium in which both phases are uniformly distributed with respect to the entire elementary volume. The effective values permit us to write the equations of mechanics of the two-phase medium in the customary form for use, in the form of equations analogous to those that are used in the mechanics of a single-phase medium. The physical variables enter in different ways into the phenomenological relations available for experimental testing.

The basic equations following from the general principles of mechanics[3,4] are the following:

$$\frac{\partial \rho^1}{\partial t} + \operatorname{div}\left(p^{(1)} v^{1(1)}\right) = I_{21} - I_{12}, \quad \frac{\partial \rho^2}{\partial t} + \operatorname{div}\left(\rho^{(2)} v^{1(2)}\right) = I_{12} - I_{21},$$

$$\rho = \rho^{(1)} + \rho^{(2)}, \quad \rho^{(1)} = \rho^{2(1)} = (1 - \alpha)\rho^{1(1)}, \quad \rho^{(2)} = \rho^{2(2)} = \alpha\rho^{1(2)}$$

$$\sigma_{ij,j} = 0, \quad \sigma_{ij} = \sigma_{ij}^{2(1)} + \sigma_{ij}^{2(2)}, \quad \sigma_{ij,j'}^{2(1)} = r_i^{(12)}, \quad \sigma_{ij,j}^{2(2)} = r_i^{(21)}$$

$$\sigma_{ij}^{(1)} = (1 - \alpha)\sigma_{ij}^{1(1)}, \quad \sigma_{ij}^{(2)} = \alpha\sigma_{ij}^{1(2)}, \quad r_i^{(12)} = -r_i^{(21)} = r_i; \quad i, j = x, y, z. \tag{1}$$

Here the variables pertaining to the entire elementary volume of the mixture are physcial, and they are described by the symbols without superscripts, for example, σ_{ij} is the total stress tensor of the mixture and ρ is the density of the mixture. The indexes in the parentheses denote the numbers of the phases: (1) is ice and (2) is water. The 1 in front of the parenthesis corresponds to the mean value; the 2 corresponds to the effective value; the absence of an index in front of the parentheses corresponds to the physical variable; v_j is the velocity; t is time; I_{ij} are the nonnegative rates of the phase conversions $i \to j$; $r_i^{(mn)}$, r_i are the forces of the interphase interaction.

Equations (1) must be supplemented by the rheologic equations, which must be considered as phenomenological, that is, following from the experiment:

$$\tau_{ij}^{(1)} = 2\eta_1^{(1)}\left(1 + \Theta_1^{(1)}\frac{d}{dt}\right)\left(\overset{\smile}{\gamma}_{ij} - \frac{1}{3}\sigma_{ij}\overset{\smile}{\gamma}_{kk}\right)$$

$$P^{(1)} = -\eta_2^{(1)}\left(1 + \Theta_2^{(1)}\frac{d}{dt}\right)\overset{\smile}{\gamma}_{kk}, \quad P_i^{(2)} = -\gamma^{(2)} j_i^{(2)} \tag{2}$$

Here, τ_{ij} is the stress deviator; P is the the pressure; $\overset{\smile}{\gamma}_{ij}$ is the deformation-rate tensor of the volume dV; σ_{ij} is the Kronecker delta; j_i is the thaw-water-flow density vector; $\eta_k^{(i)}$, $\Theta_k^{(i)}$ are the rheological constants; $\gamma^{\smile}$ is the Darcy coefficient; and d/dt is the operator of the convective derivative with respect to time.

System (1) contains the mean and effective values; system (2) contains the physical variables. The basic difficulty of the problem, which consists in relating the physcial and effective variables, can be overcome using partial physical models of the two-phase medium. For example, the following models are constructed: (1) a system of elastic rods is placed in the viscous liquid, each of which is parallel to one of the mutually perpendicular coordinate axis, and (2) the system of parallel cylindrical channels is drilled in an elastic medium in which the viscous liquid is located. Here, it is possible to establish simple relations between the effective and physical variables and at the same time to close the basic equations of regelation and also find the explicit expressions for the rheologic constants. Under the assumption of constancy of the Darcy coefficient, a complicated system of thermohydrodynamic equations reduces to one hydrodynamic equation (for $\Theta_i^{(1)} = \Theta_2^{(1)} = 0$:

$$\left(\eta_2^{(1)} + \frac{4}{3}\eta_1^{(1)}\right)\Delta v_{kk} = \gamma \operatorname{div}(\rho v) + \gamma\frac{\partial \rho}{\partial t}. \tag{3}$$

From Equation (3) we have the existence of the characteristic regelation dimension

$$L_R = \sqrt{\eta_2^{(1)} + \frac{4}{3}\eta_1^{(1)} \Big/ \rho\gamma}$$

having the above-described properties.

Within the framework of the hypothesis of local equilibrium and also the hypothesis of additiveness with respect to mass of the input phases using the apparatus discussed in the literature,[3-5] it is possible to construct the thermodynamics of the nonequilibrium, kinetic processes occurring during regelation. The proposed theoretical scheme permits efficient consideration of the surface phenomena occurring at the phase contact boundary,[5] and generalization to the case of permafrost by introduction of a third solid component is also permitted if the ice component is at the melting point. In a short report it is difficult to discuss the numerous correlaries of this theory. For example, in the case steady state, the equation is easily integrated, and the corresponding characteristics can be checked using laboratory experiments. The results of the experiments to determine the speed of sound in thawing ice are also explained and calculated.[6]

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SURFACE PHENOMENA AT THE ICE-GAS AND ICE-SOLID INTERFACES

V. F. KISELEV, V. I. KVLIVIDZE, AND A. B. KURZAYEV
Moscow Lomonosov State University

The hypothesis of the existence of a quasi-liquid film on an ice surface at temperatures below the melting point (T_m) was first stated by Faraday, and it was the object of discussion between Faraday and Tyndall, on the one hand, and J. Thomson and Kelvin, on the other.[1,2] Beginning with the concept of the disturbed structure of the surface layers of solid states, Weil[3] qualitatively evaluated the thickness of the quasi-liquid layer at several hundreds of angstroms. Fletcher,[4] considering that in the quasi-liquid film covering the ice there is a transition from the approximate structure of the liquid surface to the structure of the ice crystal, the former made quantitative estimates of the dependence of the thickness of this film, *d*, on the temperature, *T*. According to the last model of Fletcher,[5] for variation of the temperature from 267°K to 272°K, the thickness of the quasi-liquid film on the ice surface increases from 10Å to 40Å.

In recent years, when studying polycrystalline ice by the nuclear magnetic resonance (NMR) method,[6-10] a narrow signal was detected in the spectra, indicating the existence of the moving phase into the ice for $T < T_m$, and estimates were made of the relative content of the moving phase as a function of the temperature. In order to discover the nature of the narrow signal in the NMR spectrum of the ice, the following possible hypotheses were advanced: (1) the presence in the ice of either molecules not connected by hydrogen bonds or admixtures,[9,10] (2) the lowering of the melting point caused by the excess of free surface energy at the nodes of the ice grains,[9,10] and (3) the presence of a quasi-liquid film on the ice-air interface.[6,9,10]

In order to obtain a more precise response regarding the nature of the narrow signal, we performed studies of the NMR spectra in different ice samples.[11] It was discovered that, for temperatures above 200°K against a background of the broad NMR signal from the protons of the water of the crystal lattice of the ice, a narrow component of the signal appears, indicating the appearance of the moving water phase. The concentration of the moving molecules increases sharply as the temperature approaches the melting point T_m. The measurements of the temporary relations demonstrated that the moving phase is highly stable and is not connected with the processes of restructuring of the crystal lattice. By some special experiments,[11] it was proved that the admixtures and defects in the ice are not the cause of the appearance of the moving water phase. An analysis of the published data and our results uniquely indicate the presence of a quasi-liquid film on the ice surface for $T < T_m$.

The measurement of the line width of the narrow component in the ice spectra led to a value of $\Delta\nu \simeq 5$ Hz. This corresponds to a correlation time of $\tau_c \simeq 10^{-11}$ s, which is only an order greater than the correlation time for water at 293°K. The great mobility of the molecules on the ice surface at negative temperatures explains the high sintering rate of the ice particles, which plays a significant role in the process of metamorphism of snow.[12]

Qualitatively the nature of variation of the NMR spectrum of the ice at the interface between the ice and they hydrophobic material on variation of the temperature is analogous to the corresponding relation for the ice-gas boundary. In the case of the boundary between the ice and the hydrophyllic material, there is a sharp reduction in temperature at the appearance of the mobile phase which turned out to be 90° to 100° below T_m of the volumetric ice.

The effect of the surface on the structure of the surface layer of the particle is common to dispersed bodies.[13] This pertains both to the characteristic interface between the solid state and gas and the material adsorbed on the surface. The numerous data on the diffraction of the slow electrons directly indicate the existence of a disordered layer on the atom-clear surfaces both of the standard covalent crystals with diamond lattice and the ionic crystals.[13] The mobility of the lattice elements in these disordered layers is essentially higher than in the volume of the crystal. The strong effect of the surface of the solid state on the structure of the adsorbed films is indicated by the data on lowering the melting point ΔT_m of the adsorbed phase by comparison with the normal liquid.[13-15] It is interesting to note that the value of ΔT_m does not depend only on the nature of the substrate surface, but also on the structure of the liquid. Thus, according to Kvlividze *et al.*,[15,16] under other equal conditions the reduction in temperature of the phase transition for adsorbed ammonia ($\Delta T_m \simeq 50°$) is less than for water ($\Delta T_m \simeq 90°$ to 100°). As is known, in the solid phase ammonia molecules form two hydrogen bonds, and water molecules, four. These data confirm the viewpoint of Fletcher[5] regarding the presence of a mobile phase on the surface of the crystals of the materials, the molecules of which have asymmetry and are joined by hydrogen bonds.

In the case of dispersed bodies under other equal conditions, the surface structure and the specific surface energy must depend on the particle dimension.[13] In the literature,[13-15,17] we have noted the effect of the dispersedness on the adsorption properties of the surface, the specific values of the surface energy and the interphase energy at the solid-state-liquid interface, and also the melting point of the adsorbed material. The number of mobile water molecules (N_W) at the ice-gas interface in the case of specimens with a specific surface $S \cong$ 20-30 m^2/g is significantly higher than for samples with a specific surface $S \cong 5$ m^2/g. The mobility of lattice elements is qualitatively characterized the Tammann temperature, $T_{Tamm} \simeq 0.5\ T_m$, but T_m, according to Folmer, drops with a decrease in the particle size.[13] Thus, for small particles for which the number of atoms in the destroyed surface layer becomes comparable to the number of atoms in the volume of the particle, T_{Tamm} can be essentially reduced.

In the case of small particles, the effect of the surface extends to the depth of the particle volume, which we noted when investigating the NMR of the freezing water in the adsorbed state.[15] This effect will be greater the smaller the particle. The studies of the line width of the wide component of the NMR signal and the second moment of the line demonstrated that when $T > 220°$K these values for dispersed ice are much lower than for polycrystalline ice. This indicates an increase in the mobility in the volume of the dispersed ice particles. However, as follows from a comparison of the line widths of the broad and narrow components, the mobility of molecules in the volume of the crystal is many orders lower than in its surface layer.

In the case of dispersed ice, there is a comparatively stable quasi-equilibrium between the mobile phase on the surface and the defect volume of the particle. In frozen soils the stable coexistence of the bound water and ice with a negative temperature is well known.

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CHARACTERISTIC FEATURES OF ICE FORMATIONS IN CONTACT WITH FROZEN SOILS

I. B. SAVEL'YEV Production and Scientific Research Institute on Engineering Research in Construction

In the development of strength of frozen soils, the binding forces between the soil particles and the ice have great significance. The nature of these forces depends on the temperature, the soil composition and the surface properties of its particles, structure and composition of the ice, and the presence of the liquidlike interlayer and its properties as a function of the thermodynamic conditions. In order to obtain a quantitative estimate of this function, a series of elementary experiments was performed. The soil was selected by the principle of the greatest difference in properties--kaolinite and montmorillonite clays, surface suglinok (salekhardskiy), and sand. The moisture content was so selected that when the samples froze redistribution of the moisture did not occur, and a massive texture resulted. The wet samples were frozen in refrigerators under three temperature regimes: -5°, -15°, and -20°C. Then water containing the ions of the given soil was frozen on the mechanically polished surface layer by layer or by the deposition method. Further tests at the same temperatures were performed in a cold chamber.

The structure of the frozen ice and its cation composition were studied layer by layer on removal from the contact with the frozen soil. In order to estimate the effect of the interlayer of unfrozen water on the ice, the frozen soil was investigated by the nuclear magnetic resonance method. Experiments were also performed to separate the ice from the soil.

As a result of studying the structure, using a polarizing microscope, a Fedorov table, and a camera, it was established that on the formation of congelation ice on the surface of the frozen soil of massive texture the mineral particles have an active effect on the formation of the ice structure as a function of their composition

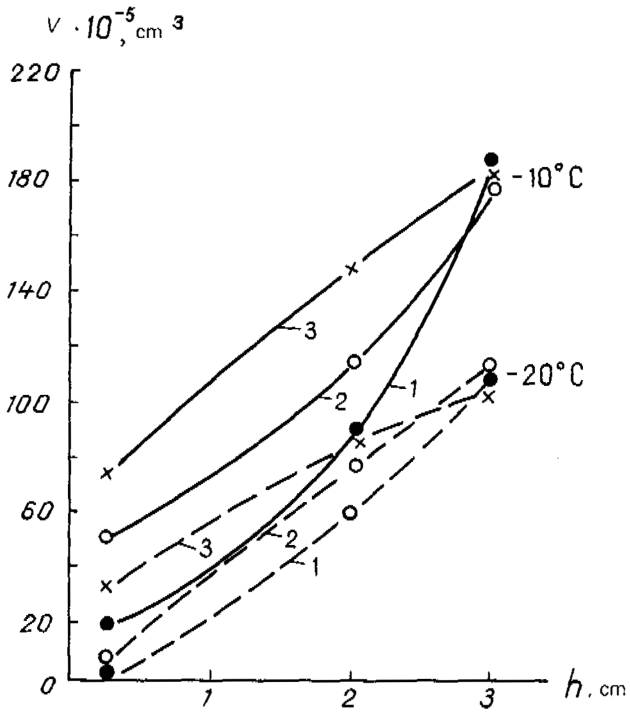

FIGURE 1 Variation of the average volume of the adfrozen ice crystals (*V*) on going away from the contact (*h*) with the surfaces of (1) frozen askangel [highly active Na-montmorillonite clay], (2) kaolin, and (3) sand at temperatures of -10°C and -20°C.

and the processes occurring on the surface.[1] In frozen montmorillonite clay, for example, appreciably more ice crystals grew per unit surface than on kaolinite or sand (Figure 1). With a reduction in the freezing point the crystal dimensions decreased, and the effectiveness of the surface effect on the ice structure noticeably diminished. The indicated phenomena were clearly exhibited in a layer of contact ice no more than 2 mm to 4 mm thick.

The chemical composition of the ice was studied using an atomadsorption spectrophotometer made by the Hitachi Company (Japan), having high selectivity and sensitivity. In the water samples taken layer by layer, on going away from the contact (2, 4, 6, 8 mm, etc.), the content of the cations Ca", Mg", Na', K' was determined. As a result, data were obtained (Figure 2) from which it is obvious that on going away from the contact the cation content increases.* In the paper by O. S. Konnova,[2] it was demonstrated that the cation composition of the ground affects the formation of the ice structure. However, here it is necessary to consider that the cations are distributed nonuniformly in the ice. The layers of ice near the contact having special structure are also isolated with respect to the reduced content of the dissolved ions. Thus, the effect of the mineral grams is felt not only on the structure of the contact ice but also in the redistribution of the microelements into the ice.

All the interactions between the freezing water and the mineral particles take place in an average way through the layer of unfrozen attached water or in the case of an inert substrate through the liquidlike transition layer, the existence of which was postulated by Weil.[3] Accordingly, it is necessary to evaluate the energetics of the unfrozen water. The energy parameter can be the spin-spin relaxation time of the protons (T_2) determined by the NMR method, inasmuch as the time T_2 is inversely proportional to the degree of bonding of the water.[4] The specimens in specially prepared test tubes made of polyethylene fluoride submerged in a thermostatic attachment operating on liquid nitrogen were placed in the magnetic field of the YaMR-1 [NMR-1] relaxameter. An additional digital recording unit[5] was introduced into the instrument, which sharply reduced its accuracy. As a result of the measurements, a series of exponential curves was obtained by which the amount of unfrozen water and the time T_2 were determined. The data on the amount of unfrozen water defined by the NMR method agree in the majority of cases (see Table 1) with the results of the calorimetric measurements of Z. A. Nersesova.[6] It is possible to explain the existing divergences obviously by the low sensitivity of the calorimetric method to the most closely related fractions of the unfrozen water. It must be noted that the small error in measuring

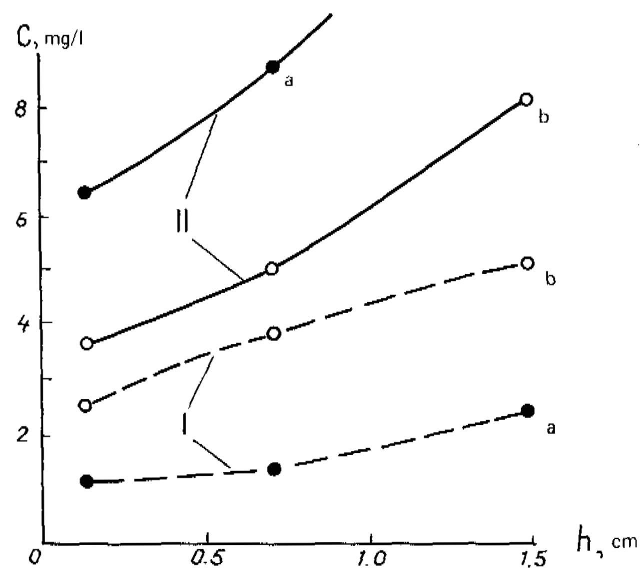

FIGURE 2 Variation of the concentration (*C*) of the cations K' (a) and Mg" (b) on going away from the contact (*R*) with frozen kaolin (I) and montmorillonite (II).

*Analogous studies were performed by the author of this report for water samples obtained from natural ice (from the regions of Tiksi and Igarka). The data are still in the processing stage, but there are grounds for considering that this method offers the possibility of compiling a well-founded opinion regarding the genesis of the ice inclusions.

TABLE 1 Comparison of Results Found by Calorimetry and by NMR

Soil	Amount of Unfrozen Water (%) per Weight of Dry Soil at Temperatures (°C)					
	-5		-10		Below -10	
	a	b	a	b	a	b
Clay containing montmorillonite	15-25	34.3	15-20	25.4	15-20	20.2
Suglinok	5-10	13.9	4-8	11.7	4-8	8.3
Sand	0.5	4.2	0.5	2.4	0.5	1.3

NOTE: a--calorimetric data; b--NMR data.

T_2 occurs as a result of the effect of paramagnetic admixtures. From the curves in Figure 3 it is obvious that the water is most energetically nonuniform in montmorillonite clay. It is possible to propose that the binding forces between the ice and the soil are determined primarily by the interlayer of unfrozen water, the thickness of which decreases with temperature. Obviously, under different temperature conditions this interlayer can be both a strengthening and a weakening factor of the frozen soil. It is necessary to consider that the primary layers of ice are probably formed partially from the bound water and, therefore, differ from ordinary ice.

The separation of the ice from the frozen soil was done on the IPG-2 instrument with a deformation rate of 2 mm/min and a separation area of 75 cm. A special attachment was made to test the samples for shear, since the IPG-2 is for compression. During these tests a control sample of concrete (300) was added. In all cases the break was of a cohesive nature, that is, a layer of ice 1-3 mm thick was left on the sample surface. Thus, the ice formed near the contact with the soil not only has a special structure and composition, but it also has special mechanical properties differing from ordinary ice. As for the cohesive forces in the ice at a distance of 1-3 mm from the contact, they, as should be expected, increase with an increase in the number of crystals and decrease in temperature (Figure 4).

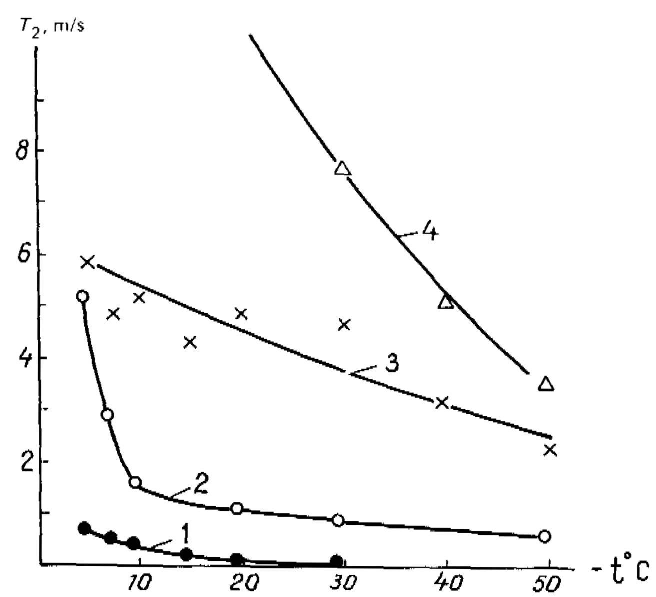

FIGURE 3 Temperature dependence of the spin-spin relaxation time of the protons of unfrozen water (T_2) in frozen samples of montmorillonite (1), kaolinite (2), running sand (3), and Moscow sand (4).

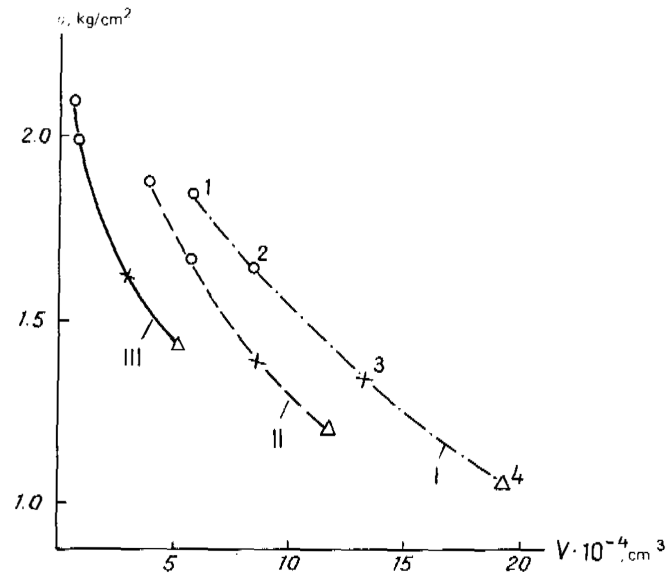

FIGURE 4 Shear (cohesion) strength of the contact ice (σ) as a function of the average crystal volume (V) on the surface of frozen (1) montmorillonite, (2) kaolinite, (3) concrete, and (4) sand at a temperature of -5°C (I), -10°C (II, and -15°C (III).

CONCLUSIONS

1. The effect of a dispersed substrate caused by its surface activity on the structure of the contact layer of adfrozen ice is established.

2. With freezing of the water contacting the soil for a long time, ice is formed in which the cation content, Ca", Mg", Na' and K' increases on going away from the contact with the ground.

3. The application of the NMR method permits determination of the amount of freezing water with greater accuracy than does the calorimetric method. The NMR method offers the possibility of estimating the energy nonuniformity of the adsorbed water on different dispersed substrates.

4. The resistance to separation of ice from the frozen soil under the conditions of our tests was of a cohesive nature. In all cases the contact layer of ice remained on the surface of the soil sample.

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EVALUATION OF SOME PHYSICAL PROPERTIES OF FROZEN SOILS AS A FUNCTION OF THE UNFROZEN WATER IN THEM

R. M. SARKISYAN Production and Scientific Research Institute on Engineering Research in Construction

This report contains a discussion of a number of definitions of the properties of frozen soils based on the dependence of the amount of unfrozen water w_u on the temperature. The reduction of problems of different nature to one base is economical, since the calorimetric measurements that are used to obtain w_u are complex, labor-consuming, and expensive, and the effort to reduce their number and improve the quality of such experimentation is natural.

According to the principle of the equilibrium state of water phases in a frozen soil of N. A. Tsytovich, frozen soil of given composition contains unfrozen water w_u in a wide temperature range, the amount of which with constant external pressure depends exclusively on the temperature.

The first calorimetric measurements of the amount of unfrozen water in frozen soil in the temperature increase cycle were made by Z. A. Nersesova.[1] The presence of unfrozen water in frozen soils was provided a basis by the crystal-chemical arguments of G. B. Bokiy,[2] and then it was confirmed by the measurements of the dielectric characteristics of the frozen soil.[3]

Z. A. Nersesova determined the amount of unfrozen water by the indirect method with respect to the relation of the heat of melting of the pore pore ice and the latent heat of fusion. Although she assumed the latent heat was constant, nevertheless the relation obtained arouses no doubt since in the experiments the temperature variation range of the ground was small and more frequently exceeded 10° C.

The function $w_u(T)$ has not up to now been subjected to a comprehensive analysis of the information included in it.

An investigation of the temperature at the beginning of freezing by physicochemical means presents difficulties primarily as a result of the absence of the practical possibility of recording the first formation of ice. Nevertheless, the function $w_u(T)$ obtained in the temperature-increase cycle for salinized ground when the frozen sample is heated in advance permits of graphical determination of the temperature of the beginning of freezing of the pore solution as a function of its concentration or the amount of water (solvent) contained in the ground.

If $w_u(T)$ is given, which expresses the dependence on the temperature of the amount of unfrozen water that is the solvent (the pore solution with a fixed amount of dissolved salts), then the temperature of the beginning of freezing of this solution in the ground with a moisutre content w will correspond to the point T_{fr} of intersection of the straight line AB with the curve $W_u(T)$, for which $w_u(T) = w$ (Figure 1). This is based on the fact that with a reduction in the soil temperature fresh ice is formed in its pores, which is accompanied by an increase in the concentration of the pore solution. To the left of the freezing point T_{fr}, the amount of solvent (its ratio to the weight of the mineral particles of the soil moisture) exceeds w; to the right, on the contrary, it is less than w, which indicates the appearance of the first ice crystals at $T = T_{fr}$.

The joint investigation of the curve AB (w = const) and the curve $w_u(T)$ in the temperature range $T < T_{fr}$ permits determination of the ratio of the water in the forzen soil in the liquid and solid phases: the segment BC on the straight line BD (Figure 1) defines the moisture w_u as a result of the liquid phase (according to the construction of the w_u curve itself), and CD determines the moisture w_{ice} as ice. A. M. Pchelintsev considered this in 1948.

The true specific heat capacity of the frozen soil $c(T)$, which is an additive variable, is made up of the heat capacity of its components--the mineral grains, ice, and liquid phase of the water.

The experimental determination of the variable is connected with great difficulty, since the heating (cooling) of the soil judging by the curve $w_u(T)$ is in practice always accompanied by phase transformations of the water.

It is possible to retain the phase composition of the water in frozen soil to determine the magnitude of the true specific heat capacity only with the cooling of the soil in the process occurring with respect to the polytropic regime, that is, accompanying the cooling (heating) of the soil by an increase (reduction) in pressure. The complexity of performing this experiment insures its practical meaning, since it requires knowledge of a number of curves $w_u(T)$ obtained for other pressures and necessary for the assignment of the polytropic regime in the experiments.

The indicated difficulty again draws attention to the $w_u(T)$ curve. By using the above-adopted notation for the moisture content of the water phases, we find the true specific heat capacity of the frozen soil:

$$c\ (T) = \frac{Q}{\gamma} = \frac{1}{1 + w}\left[c_s + c_u w_u(T) + c_i w_i\right] \quad (1)$$

$$w_u + w_i = w$$

or

$$c(T) = \frac{1}{1 + w}\left[c_s + c_u w_u(T) + c_i\left(w - w_u(T)\right)\right] \quad (2)$$

When using the expression $c(T)$ for geotechnical purposes where we deal with a narrow range of temperature variations of the frozen soil, it is assumed that the heat capacities (the mineral skeleton c_s, the unfrozen water c_u, and ice c_i) have a constant value. Then the determination of the value of $c(T)$ can be reduced to measuring segments on the curve $w_u(T)$ and their multiplication (BC times CD) by the heat capacity c_u and

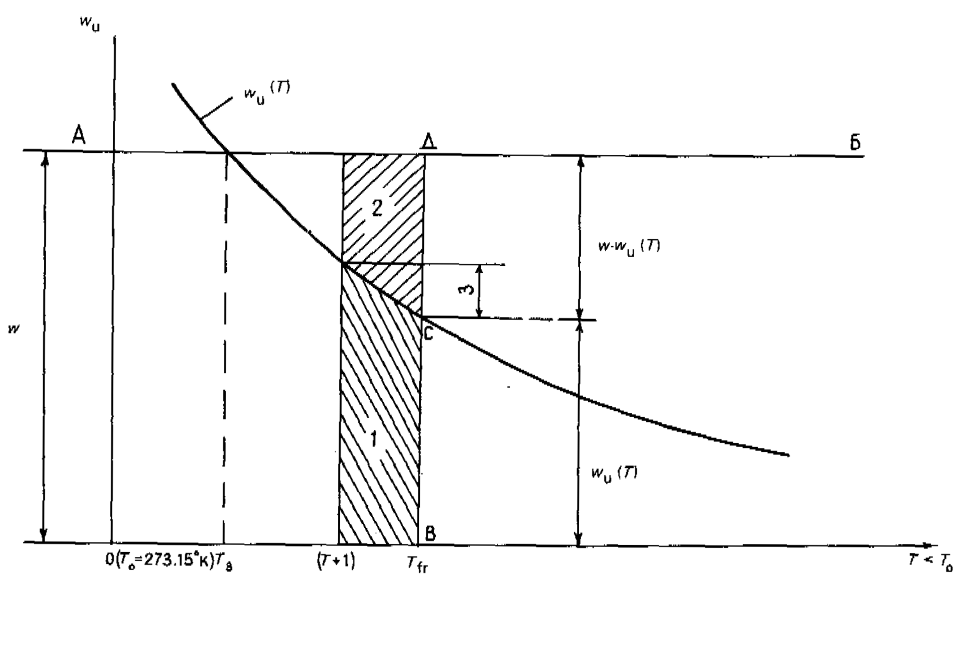

FIGURE 1 Form of the curve for the amount of unfrozen water w_u as a function of the temperature T:

$\int_T^{T+1} w_u(T)dT$ = area 1;

$\int_T^{T+1} [w - w_u(T)]dT$ = area 2;

$\int_T^{T+1} [\partial w_u(T)/\partial T]dT$ = segment 3.

c_i or only to the measurement of segments in the coordinate plane not transformed in a special way.

By using an analogous transformation of the coordinate plane, it is also possible to determine the effective heat capacity of the frozen soil, which is its natural calorimetric characteristic. It is measured by the amount of heat required to change the temperature of a unit weight of frozen soil by 1°. Part of the communicated heat is spent on heating the soil components and thawing part of the ice.

The effective heat capacity $c_{eff}(T)$ can be determined by the calorimetric method. However, if the investigated soil is studied with respect to $w_u(T)$, then it is defined graphically:

$$c_{efi}(T) = \frac{Q_{eff}}{\gamma} = \frac{1}{1+w}\left[\int_T^{T+1} c_u w_u(T)\,dT + \int_T^{T+1} c_i[w - w_u(T)]\cdot dT + \int_T^{T+1} L\frac{\partial w_u(T)}{\partial T}dT + c_s\right]. \quad (3)$$

Considering the latent heat of melting of the ice and the heat capacity c_u and c_i constant, we find the values of the integrals graphically (see Figure 1). Then

$$c_{eff}(T) = c(T) + \frac{L}{1+w}\frac{\partial w_u(T)}{\partial T}. \quad (4)$$

Let us show that the curve $w_u(T)$ also contains appropriate information for estimating the increment of the thermodynamic functions required to solve the physical problems.

In agreement with the first principle of thermodynamics, let us express the element of heat communicated to the unfrozen water

$$dQ = du - p'dv = du + \frac{p'}{\rho_u^2}\,d\rho, \quad (5)$$

where p' is the surface tension of the mineral grains and bound water and ρ_u is the density of the unfrozen water.

Let us represent the same element of heat as a function of the temperature T and the density of the unfrozen water ρ_u:

$$dQ = c_{eff}dT - a d\rho_u. \quad (6)$$

From the expression for dQ, we determine the form of the internal energy differential

$$du = c_{eff}dT + \left(\frac{p'}{\rho_u^2} - a\right)d\rho_u, \quad (7)$$

where a is a force factor, conjugate with the density of the unfrozen water ρ_u.

In addition, let us write out the expression for the entropy S of the unfrozen water using the second form of the element of heat:

$$dS = \frac{c_{eff}}{T}\,dT - \frac{a}{T}\,d\rho_u. \quad (8)$$

Let us use the reciprocity relations of the differentials u and S, that is,

$$\left(\frac{\partial c_{eff}}{\partial \rho_u}\right)_T = \frac{1}{\rho_u^2}\left(\frac{\partial p'}{\partial T}\right)_\rho - \left(\frac{\partial a}{\partial T}\right)_\rho \quad (9)$$

and

$$\frac{1}{T}\left(\frac{\partial c_{eff}}{\partial \rho}\right)_T = -\frac{T\left(\frac{\partial a}{\partial T}\right)_\rho - a}{T2}, \quad (10)$$

and let us determine the value of the force factor conjugate with p_u

$$a = \frac{T}{\rho_u^2}\left(\frac{\partial p'}{\partial T}\right)_\rho. \quad (11)$$

Thus, the form of the element of heat dQ and the differentials du and dS are established:

$$dQ = c_{eff}dT - \frac{T}{\rho_u^2}\left(\frac{\partial p'}{\partial T}\right)_\rho d\rho, \quad (12)$$

$$du \quad \frac{c_{eff}}{T}dT + \frac{1}{\rho_u^2}\left[p' - T\left(\frac{\partial p'}{\partial T}\right)_\rho\right]d\rho \quad (13)$$

and

$$dS = \frac{c_{eff}}{T}\,dT - \frac{1}{\rho_u^2}\left(\frac{\partial p'}{\partial T}\right)_\rho d\rho. \quad (14)$$

Let us only demonstrate an example of solving the problem of the dependence of ρ_u on the temperature using the expression obtained for dS. On communication of heat to the system, as is known, $dS > 0$. Let the unfrozen water and the ice in contact with it be reduced to the contact with the heat source. Then

$$dS = \frac{c_{eff}}{T}\,dT - \frac{1}{\rho_u^2}\left(\frac{dp'}{dT}\right)_\rho dp > 0$$

and

$$\frac{d\rho}{dT} < \frac{c_{eff}\rho_u^2}{T\,\frac{\partial p'}{\partial T}_\rho}. \quad (15)$$

The derivative $(\partial p'/\partial T)\rho$ is the variation of the surface tension of the bodies relating the water to the temperature, and, as is known, it is a negative value. Therefore, the complete derivative $\partial p/\partial T$ is less than a negative value. This result means that the density ρ_u and the temperature of the unfrozen water are related by an inverse relationship: With an increase in the temperature, the density of the unfrozen water decreases, and, on the contrary, it increases with a reduction in temperature.

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STUDY OF THE INDUCED POLARIZATION OF SYSTEMS SIMULATING SOILS DURING FREEZING

M. P. SIDOROVA AND D. A. FRIDRIKHSBERG
Leningrad State University

The induced polarization technique is widely used in modern geophysical practice for exploration of ore bodies, groundwater, and other purposes. The background effect caused by polarization of the ion-conducting country rock turns out to be highly significant in many cases, amsking the polarization of the electron-conducting bodies. The study of this effect has been the subject of a large number of paper.

Nevertheless, the temperature dependence of the induced polarization of the ion-conducting rock has almost not been investigated. This creates known difficulties when interpreting the data obtained by the induced polarization method, inasmuch as the method is used at a different time of the year and in different climatic zones of our country--from Zapolyar'ye to the Caucasus. The problem of the variation of the induced polarization during freezing is especially significant, since in this case the structure of the conducting paths varies--a two-phase system (saturated soil) is converted into a three-phase system where the conductivity is realized over thin films of "bound" water along the grain surfaces. In the laboratory studies[1] in the model of the poorly-graded clay (moisture content w = 30 percent), it was demonstrated that the induced polarization increased during freezing. Measurements made under field conditions[2] in the permafrost zone demonstrated that the values of η,* which vary within a broad range (from 0.3 to 5.8 percent), obviously increase at -2.5° by comparison with zero degrees. The mapping of the contact of the frozen and thawed rock (similar with respect to composition) revealed a zone of anomalous values of η.

In order to obtain the quantitative dependence of η on [temperature] T, experimental studies were made of the induced polarization in model systems of constant composition and structure, permitting us to obtain a representation of the nature of changes taking place. These changes are analogous to a known degree to the decrease in moisture content of the system, and they lead to the appearance of the third phase and then to the film transport. The study of the dependence on w† demonstrated[3] that with a decrease in w the values of η increase sharply and reach a maximum at $w \simeq 3$ percent, and then they decrease sharply whereas the variation in the transport numbers of the ions Δn increases monotonically. The explanation of the nature of the dependence of η on w follows from the representations of the nature of the induced polarization developed previously by the authors[4] for the system of series and parallel connection of the narrow (I) and wide (II) channels simulating the capillary-porous body. The narrow channels are electrochemically active, changing the transport numbers of the ions as a result of significant and uneven participation of the excess ions of the double electric layer in the transport of electricity.[5] In the wide channels the effect of the ions of the double layer can be neglected, and $\Delta n = 0$. As a result, in the microsections at the contact zones of the narrow and wide channels, concentration variations occur, leading to the appearance of diffusion potentials, the sum of which is E_{ip}. The theory developed for this model and confirmed experimentally leads to the following expression:[4]

$$\eta = \frac{4\ (\Delta n_I - \Delta n_{II})^2}{(L_I + L_{II})(1/\alpha_I + 1/L_{II})}, \qquad (1)$$

* $\eta = (E_{ip}/E) \cdot 100$ is the polarizability factor equal to the ratio (in percent) of polarization EMF (measured after inclusion of the current) to the primary EMF.

† w, the transport number of the i-th ion, (n_i) is defined as the proportion of the electricity transported by a given type of ion. The variation in the transport number (Δn_i) is equal to the difference of the transport numbers of the ion in the pores of the capillary system (n_i) and in a free solution ($n^0{}_i$):$\Delta n_i = n_i - n^0{}_i$.

where $L = S/l$, S is the cross-sectional area of the channel; l is its length; $\alpha = (x_v + x_s)/x_v$ is the efficiency; x_s and x_v are the surface and volumetric electrical conductivities. The values of Δn and α increase with a decrease in the concentration of the electrolyte and the radius of the pore channel (the thickness of the film). The theory explains the dependence of n on w by a decrease in thickness of the surface films. The increase in Δn_I and α occurring here leads to an increase in η according to (I). With a further decrease in w, the thin films are formed and in wide channels the difference between Δn_I and Δn_{II} is eliminated ($\Delta n_{II} \to \Delta n_I$) and $\eta \to 0$ for a continuous increase in Δn. Thus, the extremal nature of η as a function of w is well explained on the basis of the model. Recently the authors have studied the limiting induced polarization occurrring on passage of a current at the capillary-porous boundary with the free electrolyte also explained by the concentration variations at the interface of media with different η and introducing a noticeable contribution to the total induced polarization.[6-9]

The measurements of the induced polarization were taken by the four-electrode scheme in a cell of the usual type (l = 13 cm, d = 2.4 cm, MN = 6 cm) described previously.[4,10] The Ag/AgCl electrodes (nonpolarizing) were used as the feed electrodes, and $Cu/CuSO_4$ electrodes connected to the cell by bridges made up of agar jelly (prepared in a 0.01 molar solution of KCl) were used as the receiving electrodes. The charge time t_3 is 2 min; E_{ip} varied every 0.25 s after inclusion of the current. For measurements at +20°C, +40°C, and +60°C, the T-16 thermostat was used; the measurements at -2°C and -12°C were taken in the thermally insulated chamber filled with a mixture of ethylene glycol and dry ice. The studies were made on filled baffles made of glass beads (d = 60–90 μm), quartz (d = 100–130 μm) in $1 \cdot 10^{-4}$ normal KCl and SBS cation-exchange resin (in $5 \cdot 10^{-4}$ normal KCl), in a suspension of Oglanly bentonite (in distilled water), and in a quartz system (95 percent) + bentonite prepared by mixing the quartz with a bentonite suspension with subsequent drying and grinding.

The data obtained for positive T are presented in Table 1 (the error, $\Delta\eta/\eta < 0.1$).

Thus, for glass and quartz an increase in η is detected with an increase in T explained by the attenuation of the adsorption interaction of the counter ions with the solid phase as a result of which their mobility, the proportion of their participation in the charge transfer, and, consequently, the values of α and Δn increase.[11] A decrease in η with T for the ion-exchange resin can be explained by the fact that in dilute solutions $n_k \simeq$ I, and an increase in T can only decrease n_k as a result of an intensification of the Donnan absorption with an increase in diffusion. A decrease in η with an increase in T for bentonite is possibly connected with an increase in mobility of the particles of the nonrigid framework (Brownian motion) leading to erosion of the concentration variations, that is, to a decrease in η. The quartz-bentonite system is also not rigid as a result of segregation of the bentonite particles.

In the experiments with negative T, the values of E_{ip} were measured successively as the sample froze for 1 to 2 h. The criterion for the variation in structure of the pore space (the appearance of the ice phase) were the values of J and E on MN ($uR = E/J$); the cessation of conductivity (I = 0) corresponded to disturbance of the continuity of the conducting films; after this, the cell was placed in a thermostat at 20°, and E_{ip} was measured during the thawing process. The data obtained are presented in Table 2.

Thus, when freezing takes place, the difference in nature of variation of η is observed for variation of T for a coarse (beads) rigid system and structured highly dispersed suspension (bentonite). A significant increase in η during freezing in the first case is explained analogously to the effect of the moisture by thinning of the liquid film leading to an increase in α, Δn, and, in accordance with (1), also α. The measurements of E_{ip} for different t (to 15 s) demonstrated that the freezing significantly reduces the decay rate E_{ip}. This fact agrees with the concepts of the nature of induced polarization, inasmuch as the formation of the ice phase prolongs the diffusion path (with respect to films instead of the channel axes) and smoothens the concentration and potential gradients.

For a suspension of bentonite, on the contrary, a significant decrease in η was detected with freezing. This interesting fact can be explained either by an increase in the electro-

TABLE 1 Variation of the Polarizability (η, %) with Variation of the [Positive] Temperature

	20°	40°	60°	20°[a]
Glass beads	0.60	0.78	1.21	0.69
Quartz	0.26	0.28	0.42	0.25
SBS cation-exchange resin	10.0	9.0	8.0	10.0
Bentonite suspension	10.6	8.7	6.4	10.0
Quartz + bentonite	4.7	4.3	3.6	4.7

[a]Repeated measurements after heating to 60°.

TABLE 2 Variation of the Polarizability Factor (η) during Freezing

Glass Beads					Suspension of Bentonite (7%)				
T, °C	*I*, μA	*E*, V	E_{ip}, mV	η, %	T, °C	*I*, μA	*E*, V	E_{ip}, MV	η, %
20	83	24	36	0.15	20	500	1.15	95	8.3
-2	45	28	55	0.19	-12	300	1.25	88	7.0
	38	28	90	0.32		260	1.35	93	6.9
	36	28	112	0.40		200	1.35	85	6.3
	29	28	175	0.62		150	1.45	95	6.5
	25	33	213	0.64		80	1.45	93	6.4
	8	33	925	2.8		50	1.5	85	5.7
	4	33	1350	4.1		1.5	1.5	63	4.2
	0	0	0	0					
20	25	30	130	0.43		8.5	12	282	2.4
	29	24	50	0.21					
	39	24	50	0.21					
	63	24	27	0.11					

lyte composition at the particle surface on freezing of the H_2O* or by coagulation of the particles during the freezing process leading to an increase in grain size and, consequently, polarizability.

For a more complete understanding of the mechanism of this phenomenon, it is necessary to study the dependence of η on the electrolyte composition and the proportion of the mineral phase during the freezing of bentonite. These studies are being performed at the present time.

The results of this paper show that for sandy soils it is possible to expect an increase in the induced polarization during the freezing process; for clay soil it is possible to expect a decrease in induced polarization.

CONCLUSIONS

The induced polarization of coarse-grained rigid systems simulating sandy soil (glass beads) appreciably increases during freezing, whereas for structured suspensions of bentonite, simulating clay soil, a noticeable decrease in induced polarization is observed.

*The concentration of soluble salts in the bentonite suspension was about 10^{-3} normal, that is, an order greater than for beads, and the total amount of salt was appreciably greater as a result of the large volumetric fraction of the liquid phase in the initial suspension w = 0.93 (for beads, w = 0.44).

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ULTRASONIC METHOD OF DETERMINING THE PHASE COMPOSITION OF ICE FORMATIONS FROM SEAWATER AND SOLUTIONS

YU. YE. SLESARENKO Moscow Lomonosov State University

Results of experiments to determine the characteristics of the elasticity of salt and sea ice obtained by the ultrasonic method in the temperature range from 0°C to -25°C and with salinity from 0 to 35 percent are compared with the definitions of the phase composition. The established relations are the physcial basis for developing the ultrasonic method of evaluating the phase composition of ice formations from saltwater and seawater.

The variation in relations between pure ice and dissolved and solid salts occurring during the process of physicochemical transformations of the sea and salt ice affects the intercrystalline zones, breaking them down or strengthening them. The study of the phase composition permits control of the course of this process and is highly important in practice, inasmuch as the dynamics of the intercrystalline boundary zones are in this case the basic defining factor of the variation of strength and other physicomechanical properties of the salt-containing ice. However, the phase analysis did not permit the required development in view of the nonoperativeness and labor consumption of the methods of performing it.

In this paper an effort has been made to provide a basis for the ultrasonic method of evaluating the phase composition on the basis of studying the characteristics of elasticity of seawater and saltwater of different phase and chemical compositions.

It has been demonstrated[1] that, with a decrease in salinity and a reduction in temperature, the elasticity and other strength indexes of the seawater and saltwater increase. In the papers by Savel'yev[2] there is an indication of the defining effect of the phase composition on the physicomechanical properties of sea ice. Only Langleben[3] and Pounder[4] were able to perform experiments to compare the characteristics of the elasticity of sea ice and its phase composition, as a result of which a linear relation was obtained between the values found. However, these data had limited significance in view of certain inaccuracies of the experiments and the low range of variation of the brine content. Brown[5] provides a basis for the possibility of determining the bending strength of large ice fields by measuring the longitudinal wave velocity. As the defining factor affecting the elasticity and strength of the ice, an effort is made to provide a theoretical basis for the laws obtained on the basis of a stressed states of flat perforated plates. However, it is necessary to note the absence of data on the nature of variation of the elastic properties of sea ice in a broad range of variation of its phase composition. The data obtained when studying the salt and sea are of definite interest.

The propagation rates of longitudinal and Rayleigh elastic waves, the signals of the periods corresponding to these waves, and the calculated values of the Young's modulus and the shear modulus[1] are considered as functions of the same parameter--the liquid-phase content. The values of this parameter for sea ice were determined as a function of its salinity (the hydrochemical method) and temperature by the Weeks monogram[6] and also by the formulas of Frankenstein and Garner.[7] The efflux of brine during the measurement time is estimated by the salinity determined before and after the measurement cycle.

In Figure 1 we have the graphs of the propagation rates of the periods of the recorded signal and the dynamic Young's modulus for seawater and artificially salinized ice as a function of the liquid-phase content in them. As is obvious from Figures 1b and 1c, the nonlinear dependence of Young's modulus E and velocity V_p on the liquid-phase content in the ice is encountered. This change in velocities and Young's moduli is explained by weakening of the bonds between the ice crystals caused by an increase in the liquid-phase content. Simultaneously, a regular variation of the signal period is observed both for longitudinal T_p and for Rayleigh waves, T_R. In Figure 1a, we have graphs of $1/T$ that vary from 71.5 μ/s to 14 μ/s.

From Figure 1a, b, c, it follows that for one and the same percentage content of liquid phase there are significant differences in the absolute values of the velocities V_p, V_R of the periods T and the modulus E for salt (curves 1) and sea (curves 2) ice, which causes a different dependence of the measured parameters on the liquid-phase content in the ice.

The salt ice frozen from NaCl solutions ($S_{\text{init. water}}$ = 5, 10, 15, 25, 35 ppt) at a temperature of -15°C and -25°C is a simple two-component ice-brine system with a eutectic point of -21.1°C. With a decrease in the liquid-phase content with a reduction in temperature and salinity of this system, a smooth increase in V_p, V_R, $1/T$, and E is observed, which is caused by gradual freezing of the NaCl solution in the cells of the salt ice and a gradual increase in the liquid of the intercrystalline zones. The intersection of the curves with the coordinate axis gives values of the moduli and velocities that agree well with the data available in the literature for fresh ice containing insignificant mixtures of mineral salts.

Sea ice made from seawater with a salinity of

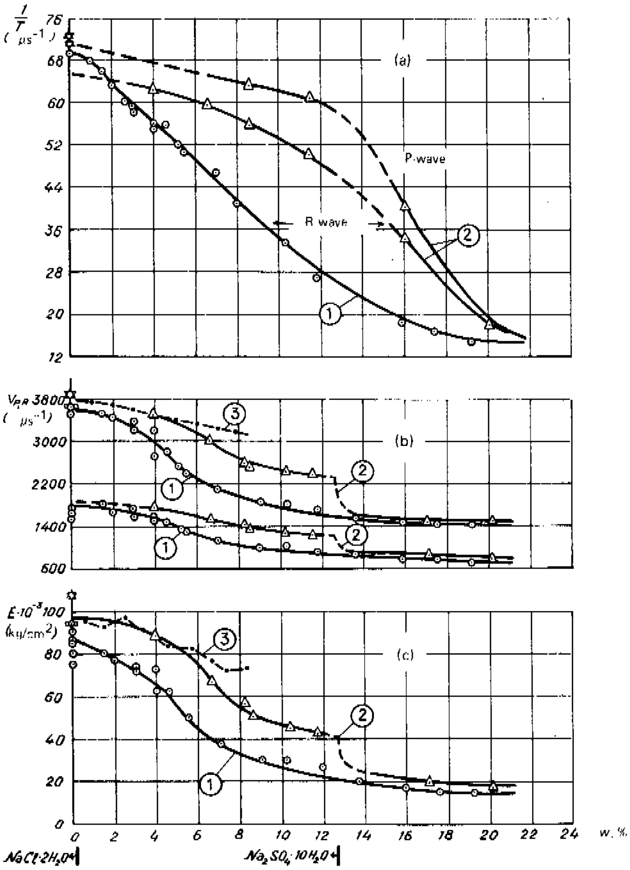

FIGURE 1 Parameters of the ultrasonic signal and Young's modulus as a function of the volumetric content of the liquid phase (*w*, percent) in ice formations according to Lavrov's data (river ice), the data of Bogorodskiy (lakes), Brokampf and Kverfurts (lakes), Berdennikov[2] (laboratory); ice from distilled water. (a) inverse values of the periods of the signals corresponding to the longitudinal (*P*) and Rayleigh (*R*) waves--$1/T$; (b) propagation rates of longitudinal (V_p) and Rayleigh (V_R) waves; (c) dynamic Young's modulus *E*. 1--ice from NaCl solution; 2--ice from seawater; 3--Arctic ice.[8]

30 ppt (the Bering Sea), is characterized by a similar relation between the above-indicated parameters, although not completely analogous. For velocities of V_p and V_R, values of $1/T$, and the Young's modulus *E* there is discontinuous variation with a brine content of $w = 12$ to 14 percent. The latter is connected with the method of preparing the ice specimens by lowering their temperature with given salinity, and the values of $w = 12$ to 14 percent corresponds to the transition to temperatures to below -8.2°C, where the Glauber's salt, $Na_2SO_4 \cdot 10H_2O$, precipitates out of the brine. When the solid salt precipitates out, there is healing of the skeletal defects of the sea ice, which leads to partial and, apparently, significant increase in rigidity of the boundary zones of the crystallites. This causes a noticeable increase in cohesion of the ice and its monolithic nature, which is felt in an increase in the elastic modulus by 85 percent and, according to Assur's data,[8] by one-third. For temperatures below the critical point for the solution of Glauber's salt, which corresponds to the brine content in our samples $w < 12$ percent, the absolute values of V_p, V_R, $1/T$, and *E* for marine ice are appreciably greater than for salt, although the nature of their variation in both cases is approximately identical with a decrease in *w*. This comparison of the relations for the artificially produced salt and sea ice permits establishment of the range within the limits of which their elasticities remain similar. From this it follows that when simulating sea ice with an artificially prepared ice can be referred to the sea ice at a temperature no lower than -8.2°C.

It has been proved experimentally that the peculiarities of the propagation of elastic waves and the variation of the elasticity modulus in the salt and sea ice are primarily determined by the specific nature of the physicochemical transformations of the ice, as a result of which there is variation of their phase and chemical compositions of the brine when the solid salts precipitate out at the corresponding temperatures. The relations obtained permit evaluation of the phase composition of the salt ice in the temperature range from 0°C to -23°C and sea ice at temperatures no lower than -8°C.

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APPLICATION OF THE METHOD OF INDUCED POLARIZATION FOR STUDYING FINE-GRAINED FROZEN SOILS

A. M. SNEGIREV, L. L. LYAKHOV, AND V. P. MEL'NIKOV Moscow Geological Exploration Institute, Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

Electromagnetic fields created in frozen soils are determined to a significant degree by the physicochemical properties of the latter, which gives rise to the possibility of using electrologging techniques for investigating the strength, permeability, and other characteristics of the rock in the cryolithozones. These methods include the induced polarization method, as recent papers have demonstrated.

The successful application of the method of induced polarization when solving engineering-geological and hydrogeological problems arose from the dependence of the induced polarization of ion-conducting rock on the configuration of the current-conducting pore space, its structure, the concentration of the electrolyte [pore] solution, ξ-potential, and so on. In the frozen deposit a temperature variation leads to variation of both the configuration of the current-conducting pore space, as a result of formation of a new phase--ice--and the structure, as a result of the occurrence of the double electric layer at the ice and frozen water boundary; variation of the concentration of the electrolyte; and the formation of cryohydrates. Depending on the initial characteristics of the medium and the freezing conditions, the transition of the deposit from the thawed state to the frozen state is accompanied by an increase or decrease in the magnitude of the polarizability. Therefore, the application of the induced polarization method under permafrost conditions required special studies, the first step in which was to study the polarizability of the ice and discover the degree of its effect on the polarizability of frozen ground as a whole.

In Figure 1 we have the results of studies of the electrical resistance and polarizability of ice samples prepared from KCl solutions of different concentration. The measurements were taken with a gradual increase in temperature from -15°C to 0°C. From the figure it is obvious that, with an increase in the electrolyte concentration, the specific electrical resistance of the ice decreases. With a sufficiently low negative temperature (-14°C), the resistance of the ice obtained when freezing solutions with high concentration is quite insignificant. Since the ice specimens before measuring were held in the frozen state not less than 2 days before measuring, the pores filled with the unfrozen water assumed the forms of cylinders, and the deviation from the cylindrical shape can be presupposed only at points where the crystals of salt precipitating out of the solution were found among the ice. This can be used to explain the fact that the polarization of the ice prepared from a solution of 10^{-4} N KCl is low, for 10^{-3} N KCl it is maximal, and with a further increase in concentration it decreases continuously as a result of an increase and equalization of the capillaries with respect to cross section. The variation in polarizability of the ice with an increase in the salt concentration is also connected with the variation of the ξ-potential on the ice and frozen electrolyte interface.[1]

By comparing the curves for ρ and η, we see that the thawing of the ice with an increase in

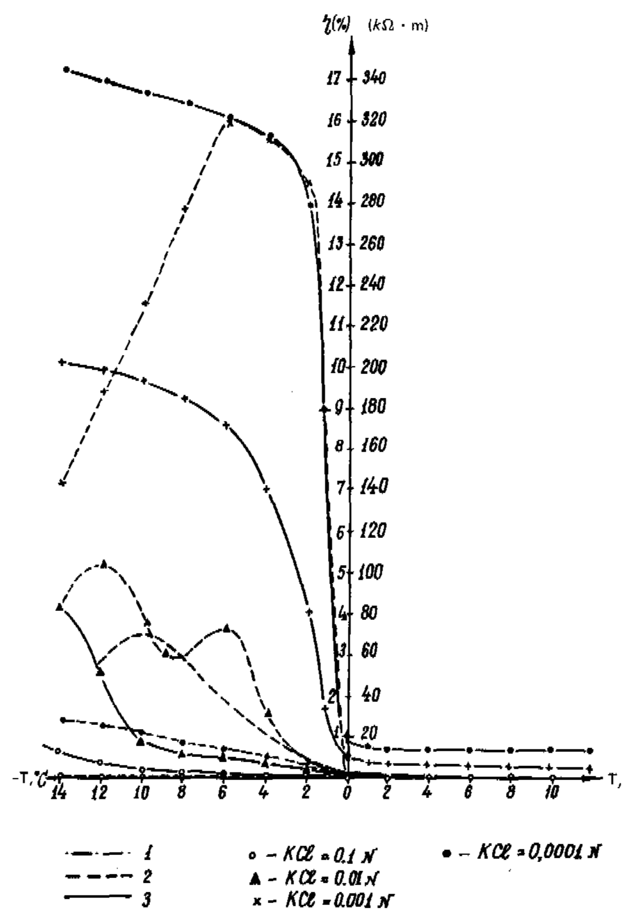

FIGURE 1 Resistivity and polarizability of ice as a function of the KCl concentration and temperature: 1--η for distilled water; 2--η and 3--ρ for solutions.

temperature taking place above all near the crystals of salt precipitating under the electrolyte leads to an increase in differentiation of the pore cross section along the current-conducting channels and, correspondingly, to a sharp increase in polarizability (this is especially well seen for an electrolyte concentration of 10^{-3} N KCl). At the same time the electrical resistivity varies quite weakly, since the cavities formed furing thawing are perpendicular to the basic direction of the current-conduction channels. A further increase in the temperature leads to equalization of the pore cross sections and a decrease in length of the electrolytic cell (according to A. F. Postel'nikov,[2] which leads to a decrease in polarizability, measured every 0.5 s after shutting the current off, and simultaneously to a decrease in the electrical resistance. Fr low electrolyte concentrations (10^{-4} N KCl), the thawing of the ice near the fine salt crystals cannot lead to noticeable differentiation in the cross section of the current-conducting channels; therefore, the polarizability decreases monotonically with an increase in temperature. The polarizability also varies for higher electrolyte concentrations (for example, for 0.1 N KCl), since the thawing of the ice in the comparatively wide pores having weak electrochemical activity takes place without a noticeable increase in their differentiation, which is indicated by the smooth decrease in electrical resistance. For an explanation of the experiments, it is also necessary to consider the variation of the electrolyte concentration in the pores and variation of the degree of solubility of the salts in the electrolyte with variation of the temperature.

Experimental studies of the ice demonstrated the following:

1. Their polarizability can reach values of more than 20 percent, and it is determined to a significant degree by the structure of the ice, the temperature, and the nature of its variation.
2. Ice as a rock-forming mineral in frozen deposits can have a significant effect on the electrochemical activity.
3. The fact of the polarizability of ice indicates the presence of an ice and unfrozen water phase boundary.

Judging by the results of the laboratory experiments, the presence of unfrozen water in the thickness of the ice is even noted with T = -14°C which theoretically agrees with the data of Nakaya and Matsumoto.[3] Thus, the possibility is discovered for determining the amount of unfrozen water in the soil with joint utilization of the data from measuring the resistance, polarizability, temperature, and salt concentrations.

The small amount of research on induced polarization in frozen deposits has permitted the selection of a theoretical model for describing the physicochemical processes leading to the occurrence of the induced polarization in the frozen soil. The theoretical estimates made for this model, taking into account the experimental studies, demonstrate that the polarizability of frozen soil can reach 30 percent and that it is quite clearly differentiated with respect to polarizability as a function of granulometric composition, degree of saturation, salt concentration, and pH of the electrolyte.

Field studies performed by the authors in a number of regions of central Yakutia, some results of which are presented in Figure 2, confirmed the presence of significant differentiation of the frozen soils with respect to polarizability by comparison with the thawed soil, and they permitted a preliminary evaluation of the possibility of the induced polarization method[4-6] for engineering-geological research.

In particular, it was established that:

1. The maximum values of the polarizability to 30 percent correspond to sections made up of

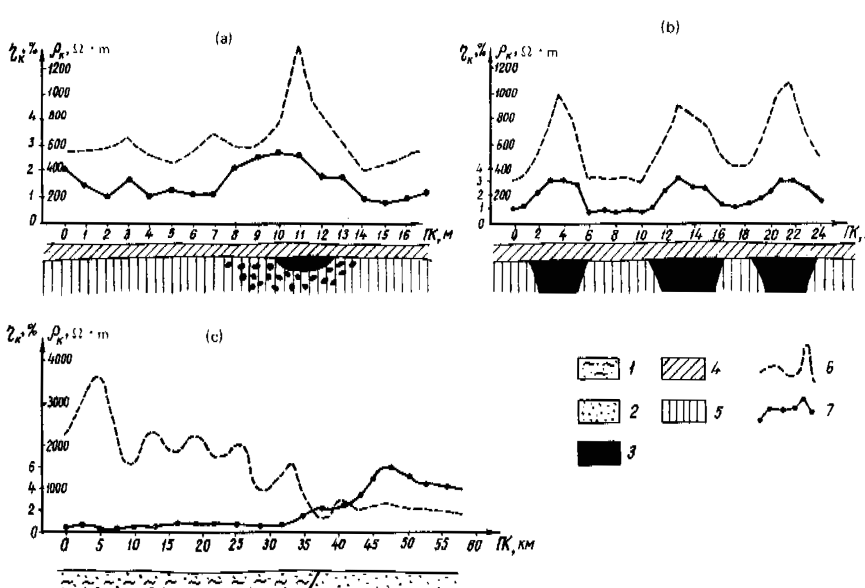

FIGURE 2 Graphs of the apparent polarizability η_k and the apparent resistance ρ_k by the symmetric profiling method: (a) in the section with increased ice content in the supes; (b) above the repeated ice veins in the sand; (c) above the suglinok and sandstone contact. 1--suglinok; 2--sandstone; 3--ice; 4--frozen soil; 5--thawed soil; 6--graph of ρ_k; 7--graph of η_k.

coarse-grained frozen soils. As the fineness of the soil increases, under other equal conditions its polarizability decreases.

2. Frozen and thawed soils, independently of granulometric composition, as a rule differ sharply with respect to electrical composition. The measurements of the polarizability indicate a decrease in the degree of differentiation of the frozen and thawed soil as a function of its grain size.

3. The frozen soil is characterized by a sharply retarded drop in the field of induced polarization with time where this rate is minimal for a coarse-grained soil.

4. With joint integration of the ρ_k and η_k graphs, the zones of occurrence of polygonal-vein ice and sections of ground with increased ice content are clearly distinguished.

5. In connection with the established dependence of the polarizability on the total mineralization of the soil water when using the induced polarization logging method of separating aquifers, in a number of cases it is possible to evaluate the quality (mineralization) of the water.

However, in spite of the fact that it is now possible to talk about the expediency of using polarizability to study cryogenous processes and to solve certain practical engineering-hydrogeological problems, many problems still remain unsolved.

Thus, for example, a far from complete study has been made of the process of the formation of the double electric layer at the interface of the ice and the unfrozen water, the variation of its parameters under the effect of the polarizing current, and the freezing and thawing rates. It is of definite interest to know the nature of the properties of double electric layers occurring at the boundaries of the mineral particles and ice.

From the procedural point of view, it is necessary to further study the dependence of the induced polarization on the current density in a broad range of its values and discover the possibilities of the induced polarization method when using current of very low frequencies, as well as to study other problems.

The results obtained can be of significant assistance during field research and also theoretical experimental studies of the electrokinetic phenomena having a great effect on the induced polarization of soils and seismo- and piezo-electric effects.

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CHARACTERISTIC FEATURES OF THE THEORETICAL SIMULATION OF ELECTROCHEMICAL PROCESSES IN ION-CONDUCTING FROZEN DEPOSITS

A. M. SNEGIREV, V. P. MEL'NIKOV, AND B. I. GENNADINIK Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

Experimental studies performed at the Geocryology Institute of the Siberian Department of the USSR Academy of Sciences jointly with the Moscow Geological Exploration Institute demonstrated that the polarizability of frozen deposits as a function of the structure and physical state of the deposits varies within broader limits (the maximum values reach 30 percent) by comparison with the thawed deposit. This fact is a favorable factor for using induced polarization and electrochemical activity in general to study the structure and composition of the frozen materials, both in the laboratories and under field conditions.

From the electrokinetic point of view, the multiphase and multicomponent frozen medium is a system comprising different types of combinations of solid (uniform or nonuniform mineral framework and ice), liquid, and gas phases. On passage of an electric current through the system, electrochemical (electrokinetic) processes can occur in it with which the slowly varying electric field is connected, the study of which can lead to conclusions regarding the structure of the medium and the state of its individual phase components. In order to understand the complex electric field, it is necessary to create mathematical and physical models of the medium properly reflecting the basic laws of the process.

The peculiarities of electrokinetic processes in a thawed ground are theoretically investigated in the model of A. F. Postel'nikov and D. I. Marshall.[1,2] The authors proposed the presence in nature of active and passive cylindrical capillaries, the electrokinetic properties of which are determined to a great extent by the structure of the double electric layer at the phase interface of the solid dielectric and the electrolyte solution. In frozen soils this model is inadequate, since a new phase appears--ice. A double electric layer occurs at the interface between the ice and the unfrozen electrolyte solution.[3] This fact forces isolation in the model of frozen ground of capillaries filled with unfrozen water and distinguished not only with respect to cross section, but also with respect to composition of the walls forming these capillaries: namely (1) the solid dielectric of the same composition, (2) ice-dielectric, (3) ice-ice, (4) ice-gas, and (5) solid dielectric-gas. (The last two types are more frequently encountered in incompletely saturated soils.)

The double electric layers on opoosite walls of the capillaries of the second, fourth, and fifth types can be characterized by different ζ-potentials, both with respect to magnitude and with respect to sign.[3,4] This causes the necessity for investigating the model of a thin capillary with asymmetric structure of the double electric layers at opposite boundaries.[5] This model permits evaluation of the electrokinetic properties of the capillaries of these types and, in particular, shows that in the second type of capillary the double electric layers occurring at the surface of the ice and solutions and the dielectric and solution can prevent complete freezing of the liquid in the capillary; but at the same time the capillary is characterized by a weak mean electrokinetic activity with respect to cross section, and it does not participate in the electrokinetic processes.

The presence of five types of capillaries in the frozen soil leads to the fact that the concepts of active and passive coupled voids become relative, since in the general case they are determined not only by the magnitude of the variation of the transport numbers, but also the sign. For example, in the capillary of the first type, as a rule, the transport number of the cations with respect to the free solution increases at the same rate; in the pores of the second type the transport numbers can differ insignificantly from those in free solution; and in the third type of capillary an increase in the transport number of the anions is possible. The structure of the capillaries is not stable with temperature variation, and therefore an analysis of more complex models than the model of A. F. Postel'nikov and D. I. Marshall is necessary.

When investigating the electrokinetic properties of the model of thawed soil, it was possible to limit ourselves to some average parameters characterizing the individual voids or all of the soil, for example, the excess surface conductivity[1] or the experimentally measured transport numbers on diaphragms.[6] In frozen soil, as a result of the variety of types of models of capillaries, it is necessary to tie their properties to the structure of the double electric layers developed on different phase interfaces and characterized by different ζ-potentials.

By investigating the effect of the structure of the diffusion part of the double electric layer on the electrokinetic properties of capillary systems (electroosmosis, flow potentials), it is possible to demonstrate that in the capillary outside the dense part of the double electric layer in practice there is no nonstationary liquid,[7,8] although it is capable of moving with much lower velocity than the liquid outside the diffusion part of the double electric layer. This explains the erroneous representation of the weak electrokinetic activity of the fine-capillary systems (when the capillary cross-section is commensurate with the thickness of the double electric layer) noted when using the Smolukhovskiy formulas to find the ζ-potential, which are valid when considering the structure of the double electric layer according to Helmholtz. Therefore, some authors proposed the presence of a nonstationary layer of solution near the phase interface with a thickness commensurate with the thickness of the double electric layer.

In conclusion, it is necessary to note that for successful theoretical simulation of the electrochemical processes in the frozen ion-conducting soil, theoretical studies must be combined with laboratory studies. The results of the experimental studies can be a representation of the electrochemical properties of individual elements of the cells and possible variations in the phase ratio, whereas the theoretical calculations give versions of the possible ratios of these elements in the model and the possibility of their experimental determination.

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DIELECTRIC METHOD OF DETERMINING THE UNFROZEN WATER CONTENT IN FROZEN SANDY-CLAY SOILS

A. D. FROLOV AND B. V. GUSEV Moscow Geological Exploration Institute, Industrial and Scientific Research Institute of Engineering Research in Construction

The experimental data we obtained on the temperature and frequency dependence of the dielectric constant ε' and the dielectric losses ε" of frozen dispersed sandy-clay soils indicate its relaxation nature.[1-3] We performed a special study of these relations and made a comparison of the results obtained with the corresponding data for fresh polycrystalline ice. It is demonstrated that the dielectric properties of cryogenous soils are determined primarily by their phase composition, that is, the content and type of unfrozen water. The direct effect of the mineralogic composition and total moisture (ice) content of the frozen soil has subordinate significance.

A detailed study of the temperature variation of the dielectric porperties has permitted us to obtain the relaxation equations of the temperature dependence of the complex dielectric constant of frozen soils in the general case for any number of temperature changes.[3] On the basis of this equation, the expression was obtained permitting calculation of the effective relaxation times of the system made up of the soil, unfrozen water, and ice, at any temperature.[3] In addition, the experimental data are used to calculate the effective relaxation times and the relaxation time distribution for different temperatures, and it was established that with a reduction in temperature there is a noticeable reduction in the changes. This can be related to a decrease in the unfrozen water content and transition of the soil to a more uniform composition. The experimental data obtained and their analysis indicate the presence of characteristic thermal variations of the dielectric properties of cryogenous soils caused by the kinetics of the phase transitions of water into ice and variation of the unfrozen water content: namely (1) a decrease in the value of the dielectric constant ε' with a reduction in soil temperature, and (2) the presence of a maximum loss factor ε'' and its independence of the total moisture (ice) content of the soil. This permitted transition to the development of the physical principles of the methods of evaluating the variation of unfrozen water content in frozen soil based on the study of its dielectric properties.

It is possible to evaluate the variation of the unfrozen water content by the results of these measurements using two procedures.

EVALUATING THE AMOUNT OF UNFROZEN WATER BY FORMULAS FOR MIXTURES

In the literature a number of formulas are known[4,5] that permit calculation of the dielectric constant of mixtures by known values of the content and dielectric constant of the components making up the mixture. We have transformed these formulas to calculate the volumetric content of unfrozen water as a result of which we found that

$$c = \frac{\varepsilon - \varepsilon_1}{\varepsilon_2 - \varepsilon_1} \qquad (1)$$

where c is the volumetric content of the unfrozen water; ε is the dielectric constant of the frozen soil; ε_1 is the dielectric constant of the system made up of the ice, the mineral matrix, and the air in the voids of the soil; and ε_2 is the dielectric constant of the unfrozen water. The value of ε_1 is defined by the formula

$$\lg \varepsilon_1 = A_{ma(mai)} \cdot A_{m(ma)} \cdot \lg \varepsilon_m + A_{i(mai)} \cdot \lg \varepsilon_i,$$

where $A_{ma(mai)}$ is the volumetric content of the system made up of the mineral matrix and the air in the system made up of the mineral grains, air and ice; ε_m is the dielectric constant of the mineral grains; $A_{i(mai)}$ is the volumetric content of the ice in the system made up of the mineral grains, air, and ice; ε_i is the dielectric constant of the ice; and $A_{m(ma)}$ is the volumetric content of the mineral grains in the system made up of mineral grains and air.

The value of ε_2 is more difficult to define. First, it can depend on the temperature, the nature of the mineral matrix, and so on. Secondly, there are no reliable, highly accurate determinations of the amount of unfrozen water at various temperatures. Therefore, at any step it is possible to determine the amount of unfrozen water only by making some simplifying assumptions. For this purpose, let us denote $\varepsilon_2 = k(t)\varepsilon_w$, where ε_w is the dielectric constant of pure water. The coefficient $k(t)$ can be described as follows: (1) its magnitude depends on the temperature; (2) $k(t) \gg 1$, which follows from the presence of uncompensated charges distributed in the layers of unfrozen water, the possibility of displacement of which is limited by the double layers, which leads to significant volumetric polarization.

On the basis of this, the above formula assumes the form:

$$c(t) = \frac{\varepsilon - \varepsilon_1}{k(t)\varepsilon_w} \qquad (2)$$

since $\varepsilon_2 \ll \varepsilon_1 (\varepsilon_w > \varepsilon_1)$.

By this formula the values of $k(t)c(t)$ were calculated for quartz sand samples. The results of these calculations are presented in Figure 1. They permit discovery of the following peculiarities of the variation of the amount of unfrozen water with a reduction in temperature:

1. The amount of unfrozen water in the frozen wet sand does not in practice depend on the total moisture (ice) content of the samples. The graphs constructed by the results of the measurements on different frequencies and, in sand samples with a moisture from 0.96 to 19.2 percent, practically coincide.

2. The variation in the amount of unfrozen water in the frozen sand occurs at a temperature of -25°C, although the intensity of these variations decreases sharply with a reduction in temperature.

If we assume that the coefficient $k(t)$ does not depend on the temperature and take as the

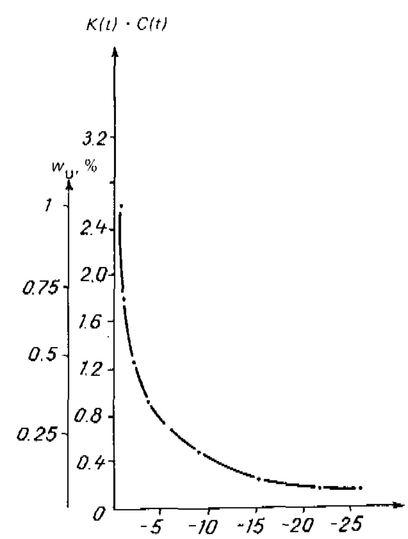

FIGURE 1 Magnitude of $k(t) \cdot c(t)$ and the unfrozen water content (w_u percent) in frozen quartz sand as a function of temperature.

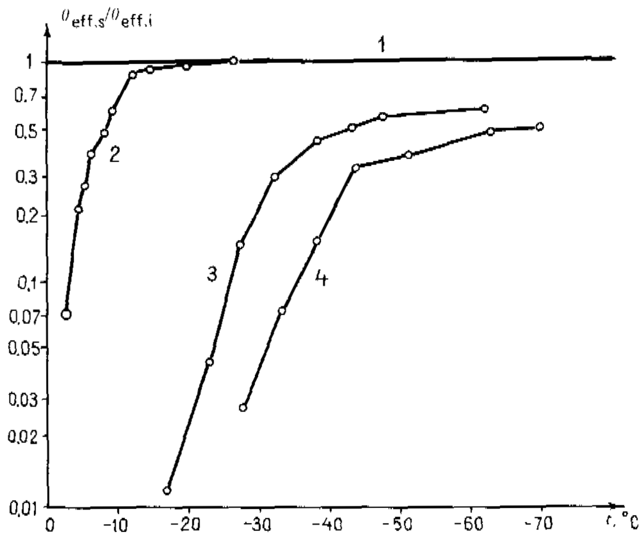

FIGURE 2 Ratio of the effective relaxation times Θ_{eff} of the soil and of ice as a function of temperature. 1--ice; 2--quartz sand; 3--kaolin; 4--suglinok.

base the calorimetric data by which at a temperature of -0.5°C the frozen sand contains about 1 percent unfrozen water (obviously, a high value), then it is possible to make a quantitative evaluation of the unfrozen water content at various temperatures. In accordance with the graph presented in Figure 1, at a temperature of -15°C in the frozen quartz sand (the 0.1-0.25-mm fraction), there is about 0.09 percent unfrozen water.

ESTIMATING THE AMOUNT OF UNFROZEN WATER BY COMPARING THE EFFECTS OF TEMPERATURE VARIATION ON THE EFFECTIVE RELAXATION TIMES OF ICE AND PERMAFROST

It is known that the effective relaxation time for frozen soils--$\Theta_{eff}(t)$--is determined by the relaxation of the ice and unfrozen water.[2,3] Therefore, in some temperature range (to the time of practically complete phase transitions), the ratio $\Theta_{eff}(t)_s/\Theta_{eff}(t)_i$ [subscript "s" for soil solids, "i" for ice] must characterize the contribution of the unfrozen water to the total relaxation effect, that is, the value of $k(t)c(t)$. Therefore, the graphs for the ratio $\Theta_{eff}(t)_s/\Theta_{eff}(t)_i$ as a function of temperature can by means of the earlier assumptions be converted into graphs of the dependence of the amount of unfrozen water in the frozen soil on the temperature. In this case the time of practically complete phase transitions of the water into ice is characterized by the emergence of the graph of the values of $\Theta_{eff}(t)_s/\Theta_{eff}(t)_i$ as a function of the temperature at the asymptote in the special case approaching 1 for sand. It is characteristic (Figure 2) that in the case of frozen clay the value of this ratio even at the temperature of about -70°C is still far from 1. This indicates that in the clay even at this low temperature there is a quite large amount of unfrozen water, which agrees well with the data of the phase transitions of the water in this soil.[6]

This permits recommendation of the dielectric method of investigating frozen soils when studying the peculiarities of the phase transitions of water and for estimating the amount of unfrozen water in the frozen soil at different temperatures. A significant advantage of the dielectric method is the possibility of taking measurements using quite simple devices and also its application both under laboratory and natural conditions, including continuous observations of the variation of the phase composition of the water in the frozen soil.

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EFFECT OF NATURAL ELECTRIC POTENTIALS ON WATER MIGRATION IN FREEZING SOILS

I. G. YARKIN Scientific Research Institute of Foundations and Underground Structures

When a soil freezes, a difference in electric potential arises between the frozen and thawed part of the sample, which can be measured using electrodes.

First, the measurement of the electric potential difference in the freezing "washed glacial material" was taken by Jumikis.[1] He measured the potentials on the order of several tens of millivolts. The study of the electric potential difference occurring during the freezing of soil was also performed by V. P. Borovitskiy.[2]

Our problem consisted in discovering the interrelations between the magnitude and the dynamics of the electric potentials and redistribution of the moisture in the freezing soil. The electric potential difference was measured for freezing of samples of different grain size and mineralogic composition for different moisture contents and freezing rates. A different nature of the dynamics of the electric potentials was discovered in the sand and clay and also a significant difference in the dynamics of the electrical potential difference in clay freezing without significant redistribution of the moisture (low-moisture and fast-freezing soil) and in a soil that froze with significant redistribution of the moisture (very wet clays and suglinoks with slow freezing). The interrelation between the magnitude of the electric potential difference and the intensity of the moisture redistribution in the freezing ground was observed.

The measurement of the electric potential difference was taken using grid platinum electrodes. The samples froze from the top down. The electric potential difference between the electrodes was measured using LPU-01 potentiometers with an accuracy to ±2 mV, and it was recorded on disphragm tape by the KVT-12 pen recorder. The comparison electrode was the top electrode. Therefore, the voltage of the lower electrode by comparison with the upper electrode was recorded on the graphs.

The standard curves reflecting the variation of the electric potential difference in the process of freezing of the samples are presented in Figure 1. The freezing conditions were in all three cases approximately identical. The electrodes were placed at the upper and lower ends of the samples and were buried 1.0-1.5 cm in the soil.

From the figure it is evident that curve 1 for sand has a minimum and maximum. These points correspond to the times of freezing at the upper and lower electrodes. The experimental result indicates that the freezing part of the sample was charged negatively with respect to the unfrozen part. The redistribution of the moisture in the freezing process with respect to the sandy sample profile was practically not observed.

Curve 2 does not have a clearly expressed minimum corresponding to the time of freezing in of the upper electrode. The rise of the curve upward begins after passage of the freezing front through the upper electrode. The freezing part of the soil acquires a negative potential with respect to the unfrozen part, which is retained to the end of the freezing process. The freezing of the kaolin appeared without noticeable redistribution of the moisture with respect to the sample profile.

Curve 3 indicates an entirely different trend in the dynamics of the electric potential difference. After freezing of the upper electrode, the curve begins to bend downward. The frozen part of the soil is charged positively with respect to the unfrozen part. The freezing of the sample took place with significant redistribution of the moisture: The moisture content of the upper third of the sample after freezing rose to 87.6 percent as a result of drying to

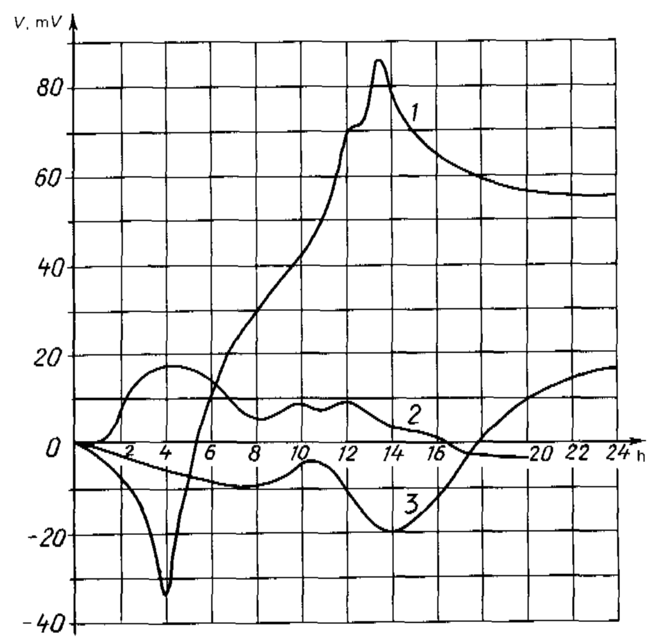

FIGURE 1 Variation of the magnitude of the electric potential difference during the freezing process of the soil samples. 1--sand with a moisture content of 18.5 percent; 2--kaolin with a moisture content of 40 percent; 3--the same, 60 percent.

44.7 percent of the middle third of the sample. The lower part retained its initial moisture content.

Thus, the dynamics of the electric potential difference in sandy soil are characterized by one type of curve (curve 1, Figure 1). For clay soil, two basic types of curves are characteristic. The first can include curve 2 (Figure 1). This type is characteristic for the soil in which significant redistribution of the moisture in the freezing process does not occur.

The second type can include curve 3 (Figure 1). This type of curve is characteristic of the soil in which redistribution of the moisture is noted by the profile of the samples during the freezing process.

Between the curves of the first and second type, there is a gradual transition. With an increase in the soil moisture content and with a decrease its freezing rate, the curves of the first type were initially gently sloping, and then they began to bend downward, that is, they became curves of the second type. The degree of moisture redistribution also increased, which indicates the interrelation between the soil migration and the dynamics of the electric potentials.

The distribution curve of the electropotentials in the clay soil, freezing without moisture redistribution, is similar in general features to

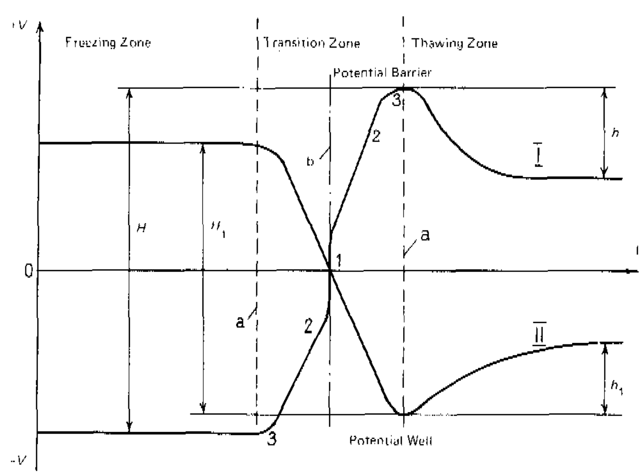

FIGURE 2 Variation of the magnitude of the electric potential with respect to the profile of the frozen soil samples. I--sand; II--clay soil freezing with significant redistribution of the moisture; H--height of the potential barrier in the sand above the frozen part of the sample; h--height of the potential barrier in the sand above the thawed part of the sample; H_1--depth of the potential well with respect to the frozen part of the clay soil; h_1--depth of the potential well with respect to the thawed part of the clay soil; 1, 2, and 3--the region of the jump of the electric potential to the region with a constant potential gradient and the region with a decreasing potential gradient in the phase transition zone of the sandy samples; a--zone boundaries; b--freezing front.

the curve for the sand and therefore it is not presented.

From Figure 2 it is obvious that in the frozen soil three zones have been isolated: the frozen soil, the region of phase transitions, and thawed soil. The maximum drop in magnitude of the electric potential is observed in the phase-transition zone. In sandy soil before the freezing front, an excess of positive charges was observed. The width of the phase transition zone is about 20 mm.

In clay soil freezing with significant moisture redistribution, the phase-transition zone is more extended, and the excess negative charges were formed before the freezing front.

Inasmuch as in natural soil the water usually migrates in the direction of the negative pole (cathode), the excess of positive charges is the potential barrier for the water molecules and prevents moisture migration to the freezing front. In the clay soil, the excess negative charges play the role of a "potential well" and promote movement of the moisture to the freezing front. The experiments with sand demonstrated that, as the electrodes come together, the electric potential difference decreases. However, as the electrodes come together, the rate of increase in the potential difference increases. At very short electrode spacings, the potential gradient becomes so large that it can be considered as a potential jump. This jump can be considered part of the phase interface between the ice and the pore solution before the freezing front (section 1 on curve 1, Figure 2). On both sides of this boundary, there are zones in which almost linear variation of the electric potential is observed (section 2 on curve 1, Figure 2), and at the boundary of the phase-transition zone, sections at the limits of which the potential gradient rapidly decreases (section 3 in curve 1, Figure 2). The potential jump at the phase interface can be considered as the contact potential between the phases.[3] The potential drop on both sides of the interface can be considered as the "volumetric charge" occurring for "selective separation of the ions" between the solid (ice) and liquid (solution) phases known by the name of "freezing potential"[4] and the Workman-Reynolds potential.[5,6] The relative sign (polarity) of the charge of the frozen and thawed part of the sample is determined by the chemical composition of the pore solution. The small values of the potential difference, by comparison with the Workman-Reynolds potentials occurring during freezing of the pure solutions, are a consequence of inadequate development of this effect under the conditions of the ground system.

The main cause of the occurrence of the electric potential difference in the freezing sand must be considered to be the Workman-Reynolds effect.

The basic cause of the appearance of an electric potential difference in clay soil is the interface potential gradient, which occurs as a result of the exchange-absorption processes on the surfaces of the mineral grains of the soil.

I. A. Tyutyunov[7] derived the analytical formula from the fundamental Gibbs equation[8] reflecting the dependence of the interphase potential V on the magnitude of the specific surface energy of the soil σ and the electric capacity of a unit surface of ground E:

$$dV = -\frac{d\sigma}{E}, \quad (1)$$

from which it follows (for a constant E) that

$$V = V_0 - \frac{\sigma}{E}, \quad (2)$$

where V_0 is the integration constant.

In the majority of natural soils, the surface of the mineral matrix is charged negatively; therefore V_0 is also negative. The absolute magnitude of the negative value increases with a reduction in temperature as a result of the fact that the magnitude of the specific energy surface σ increases with a reduction in temperature. For this reason, in clay soil there are excess negative charges on the colder end of the sample. If the moisture migration does not take place, the frozen part of the sample acquires a negative charge with respect to the thawed part. When measuring the electric potential difference, curves of the first type are obtained that are analogous to the curve 2 in Figure 1.

If the moisture migration in the freezing sample takes place quite intensely, flow potentials arise that compensate for part of the excess negative charge. The frozen part acquires a positive charge with respect to the unfrozen part of the sample and before the freezing front excess negative charges are retained, which form the "potential well." When measuring the electric potential difference, curves of the second type are obtained, which are analogous to curve 3 in Figure 1.

Thus, moisture migration in freezing clay soils takes place under the effect of that part of the electric interphase potential gradient that remains uncompensated for by the flow potentials.

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PART V

Groundwater in the Cryolithosphere

N. I. TELSTIKHIN and
O. N. TELSTIKHIN, *Editors*

CRYOGENOUS METAMORPHIZATION OF CHEMICAL COMPOSITION OF SUBSURFACE WATER (EXEMPLIFIED IN CENTRAL YAKUTIA)

N. P. ANISIMOVA Permafrostology Institute in Siberian Branch of USSR Academy of Sciences

Among the factors determining the chemical composition of subsurface water in permafrost regions, a specific role is played by the cryogenous processes, to which in recent times the ever-increasing attention of researchers has been devoted. In connection with the diversity of frost-hydrogeologic conditions within the permafrost and their variability through time, and also with the difficulties in conducting laboratory experiments simulating the natural processes, the investigation of this question becomes complicated to a considerable degree.

At the present time, we have studied most fully the effect of cryogenesis on the composition and distribution of salts between the ice and liquid phases of fresh surface- and sea-water.[2-12,16] The general tendencies obtained as a result of these studies are also valid for the freezing subsurface water; however, for the latter, there exists a series of specific features.

Among the wide diversity of factors determining the formation of the subsurface water's chemical composition, the role of cryogenous processes cannot always be successfully traced. This is particularly so because, under some conditions, they have leading significance, while in others their role is slight or has an ephemeral reversible nature.

The directivity and stage of cryogenous metamorphization of the underground water's chemical composition depend chiefly on temperature and freezing rate of the water-bearing (or thawing icy) soils, on original chemical composition of water and soils, thickness of the water-bearing horizon, and intensity of water exchange.

The variations in the chemical composition of underground waters in the process of freezing and subsequent thawing are manifested most distinctly on their emergence to the surface and formation of naleds. The physicochemical processes occurring at this time have been studied in the Ulakhan-Taryn and Bulus Springs in central Yakutia and under laboratory conditions.[1,14] The composition of water in the springs is calcium-magnesium hydrocarbonate with mineralization ranging from 0.10-0.14 g/l.

In winter the mineralization of the water flowing along the frozen base (soil or ice) increases; moreover, it does so all the more intensively the more gradually the freezing takes place and, consequently, the longer the path of its travel along the surface. The increase in mineralization of water occurs owing to its enrichment chiefly by hydrocarbonates of sodium, magnesium, and by sodium chloride. Such a change in composition is determined first by the transition to a precipitate on crystallization of water of difficultly-soluble calcium carbonates and second by solution of the salts contained in the surface layer in the water flowing along the naled. The solution of the magnesium hydrocarbonates is promoted by an increased content of carbon dioxide in the water, while the water's enrichment by chlorides and hydrocarbonates of sodium is favored by their migration into the surface layer (warmed by water) of the naled from the layers occurring deeper. This is confirmed by laboratory experiments: During drops in temperature in a column of ice, the chlorides, hydrocarbonates, and carbonates of sodium migrate intensively toward the "warmer" layers (see Figure 1).

During the spring season at maximum area of water flowing from the springs, its mineralization increases quite appreciably. Even at an increase of twice the water's mineralization, the content of sodium hydrocarbonates in it increases by 3 times; magnesium hydrocarbonates increase by 2.6 times, while the sodium chlorides increase 2 times.

In accordance with the increase in water mineralization, there is an increase in the mineralization of ice forming from it, i.e., it increases over the area of naled from its upper end to the lower. At the end of each encrustation* where the mineralization of ice increases, as compared with the mineralization of springs feeding the naled by 10-45 times, the content of sodium chlorides and magnesium hydrocarbonates increases by almost as many times, whereas the content of sodium hydrocarbonates increases accordingly by 25-150 times, although in the

*The interruptions in the arrival of new batches of flowing water cause the stratification, while an irregular water flow in various directions leads to the formation of "nateks" (encrustations) of excrescences, or of cascade-type benches, separated by smooth sectors).

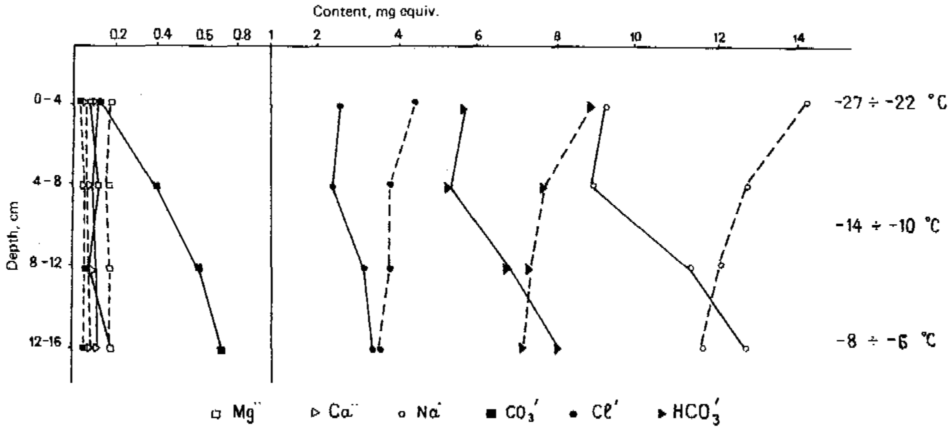

FIGURE 1 Migration of ions in ice specimen of chloride-hydrocarbonate composition and mineralization of 0.74 g/l at temperature-difference at its ends from -8 to -6 to -27 to -22°C.

spring water their percentage is about 10 percent.

During the winter season, the cryogenous variations in the composition of spring water occur only within the limits of the naled body, and, in general, all the salts prove to be, so to speak, "preserved" in some form or other in it. During the thawing period of a naled, these cryogenous changes are reflected appreciably on the mineralization and composition of melt water. The latter, mixing with the spring water, in its turn modifies the mineralization and composition to some degree or other.

Even during a partial melting of ice at the onset of the thawing of a naled, a large percentage of the readily soluble salts changes to solution, migrates downward toward the naled base, and is transported with the melt water beyond its limits; and, as a result of this, the naled becomes greatly "freshened" so that, in the first stage of the naled thawing (usually in the second half of April), the mineralization of melt water is maximum at an increased content of hydrocarbonates of sodium, magnesium, and sodium chloride. As a result, there is an increase in the content of these components also in the stream formed by the spring water.

In the second stage (May-June), the mineralization of the melt water decreases, since essentially only the calcium and magnesium hydrocarbonates having remained in the naled become dissolved. The calcium and magnesium carbonates, having precipitated on the crystallization of water, are partly carried away in the form of suspension by the surface flow, again become partly dissolved by the cold spring water containing carbon dioxide, and, by the aggressive summer rains, are partly assimilated by microorganisms and also remain on the valley flow.

During naled formation, the maximum quantities of calcium carbonate are lost by underground water of calcium hydrocarbonate composition with increased carbon dioxide content. According to data furnished by N. A. Vlasov, L. I. Pavlova, and A. V. Ivanov,[4] during the naled-formation process, the underground water in Yakutia loses 2-4 × 10^3 tons/yr of carbonate salts. An even higher figure has been obtained by N. N. Indoleva (unpublished data). According to her calculations, just for the southern Verkhoyansk region, the quantity of salts extracted by naleds from the naled-forming water reaches almost 15 × 10^3 tons/yr. Without evaluating here the accuracy of the values listed, we can consider that the fallout of carbonates into precipitation during the freezing of subsurface water outcropping to the surface constitutes a significant value and has appreciable importance in the variation in mineralization and composition of naled runoff. For the freshwater underground flow of the indicated composition, the naleds play the parts of unique types of geochemical cryogenous barriers. The effectiveness of the "impeding" role of such barriers is conditioned by the physicochemical processes during the freezing of water and thawing of ice.

In this manner, during the change of subsurface water to surface water through naled formation, on the one hand freshening takes place, while on the other hand the redistribution of the ions occurs.

In seasonally freezing-and-thawing layers under conditions of the most abrupt annual temperature fluctuations, the cryogenous changes in chemical composition and mineralization of suprapermafrost water are appreciable over the extent of one annual cycle.

During the freezing of suprapermafrost water, the rise in its mineralization is accompanied by changes in the ratios of ions. The calcium and magnesium hydrocarbonates convert to normal carbonates, which form eutectics with the ice. At this time, in the water there is a rise in the content of sodium hydrocarbonates. Even during the initial freezing spell (October-November) of the very fresh (30-60 mg/l suprapermafrost water having chlorides and hydrocarbonates of magnesium-calcium or sodium-calcium in the sands of active layer, the content of sodium hydrocarbonates and of sodium and magnesium sulfates in the water rises; moreover, it does so all the more, the more poorly the sector is drained (see Table 1, No. 1).

In the composition of the water sometimes preserved in the thin pereletok lenses, the

TABLE 1 Chemical Composition of Subsurface Water

Number in Series	Place of Sample Extraction	Date	Depth, M	Total Mineral Substances, g/l	Content of Ions, mg-equiv						Content of Ions, equiv %					
					Ca^{2+}	Mg^{2+}	$Na^{+}+K^{+}$	HCO_3^-	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^2	$Na^{+}+K^{+}$	HCO_3^-	SO_4^{2-}	Cl^-
1	Second above-floodplain terrace on Lena River in the vicinity of the city of Yakutsk	Nov. 20 1950	1.2	0.31	1.85	1.00	2.00	3.95	1.2	0.6	32.2	17.4	50.4	68.6	21.0	10.0
2	First above-floodplain terrace on Lunkha River (in residual lens)	July 1 1960	1.7	0.56	1.91	1.01	7.14	7.20	1.62	1.23	19.0	10.0	71.0	72.0	16.0	12.0
3	Khanchala River Valley	July 24 1964	1.5	0.80	3.19	5.73	4.54	4.33	8.51	0.62	23.7	42.6	33.7	32.2	63.2	4.6
4	Alluvial plain, natural landmark	July 24 1964	1.5	1.89	4.51	19.92	8.13	6.20	19.9	6.44	13.9	61.2	24.9	19.0	61.2	19.8
5	Lunkha River Valley, depression	July 12 1960	2.5	1.16	11.00	10.00	0.25	3.24	6.03	12.0	52.0	47.1	0.9	15.2	28.4	56.4
6	Alluvial plain	June 23 1964	1.0	0.06	0.23	0.30	0.70	0.69	Het	0.54	19.1	23.7	57.2	56.0	Het	44.0
7	Fifth above-floodplain terrace on Lena River	June 15 1963	0.6	0.03	0.18	0.18	0.31	0.39	Het	0.28	26.9	26.9	46.2	58.2	Het	41.8
8	Same place	Aug. 23 1963	1.7	0.26	1.99	1.48	1.19	4.38	Het	0.21	43.3	32.3	24.4	95.4	Het	4.6
9	Same place, natural landmark	Aug. 12 1960	1.8	1.03	2.20	4.30	13.10	14.20	1.00	4.40	11.2	21.9	66.9	72.4	5.1	22.4
10	Second above-floodplain terrace on Lena River	Sept. 21 1969	5.1	14.73	32.80	93.20	128.06	1.80	31.17	230.40	12.4	35.4	48.6	0.7	11.8	87.5
11	First above-floodplain terrace on Lena River	Oct. 30 1966	5.0	59.92	128.54	650.35	327.09	10.38	89.99	1005.61	11.6	58.8	29.6	0.9	8.1	91.0

relative content of sodium hydrocarbonates is still higher (see Table 1, No. 2).

In the ice-rich sands, having formed in the final freezing stage of the suprapermafrost water, the relative content of sodium and magnesium sulfates is usually greatly increased. Their role is particularly great during the freezing of water-saturated supesses and suglinoks in the poorly drained sectors (see Table 1, Nos. 3 and 4).

The chloride salts in the pore solutions, the concentration of which in the layers frozen in the last stage increases, partly migrate into the subjacent permafrost. Therefore, in such icy mineralized layers, the sulfates often predominate, while, in the subjacent permafrost layers, the content of chlorides is intensified.

During the freezing of saturated soils, in topographic depression (especially the ones without outflow), where chlorides and sulfates normally predominate in the salt composition, this migration is most appreciable. The movement of concentrated solutions into the underlying permafrost occurs under the effects of gravitational and thermal gradients and diffusion.

The results of chemical analyses of water extracts from the soils selected during the drilling of boreholes testify to the extensive distribution of the described phenomena in central Yakutia.

Mineralization of the suprapermafrost water increases in proportion to the thawing of frozen soils. From May to June, in the poorly drained sectors, the relative content of chlorides is increased (see Table 1, Nos. 6 and 7); from July to August, at total thawing of the frozen layer, mineralization of suprapermafrost water increases, chiefly owing to the calcium and magnesium hydrocarbonates, once again converting to a solution being enriched by carbon dioxide (see Table 1, No. 8). The significantly high solubility of magnesium carbonate salts, as compared with the calsium salts, frequently determines their predominant content in the suprapermafrost water.

The mineralization of suprapermafrost water in the depressions increases, chiefly owing to their enrichment by readily soluble salts (see Table 1, No. 9). The chlorides, having migrated (during the complete freezing of suprapermafrost water) into the permafrost, on thawing do not return completely, and the part that does return does so only under an appreciable temperature gradient.

The indicated modifications in the chemical composition of suprapermafrost waters occurring during their freezing and thawing have great importance in the chemical composition of the taliks' underground water into which they flow.

In the suprapermafrost and interpermafrost taliks occurring at depths ranging from 2.5 m to 20 m, where temperature throughout the year is variable, the appreciable cryogenous changes in the composition and mineralization of subsurface water occur over decades. In the process of the freezing of water-bearing soils in the taliks, mineralization of water increases and its salt composition varies, chiefly owing to the precipitation of individual salts and exchange reactions with the soil. The degree of metamorphization in the composition of this water depends both on the severity of the freezing conditions and on the dimensions of the talik, as well as on orignial composition and mineralization of water and the water circulation conditions.

In all cases of the freezing of saturated fine-grained soils in taliks at high negative temperatures, the increase in mineralization of water in the boundary layers with frozen soils occurs not only owing to the slightly migrating chlorides, sulfates, and sodium hydrocarbonates, but also to a considerable extent owing to the magnesium and calcuim hydrocarbonates. Obviously this feature is explained by the fact that at high negative temperatures of soil freezing, only a small proportion of the carbonates forms eutectic solutions with the ice. In addition, the calcium content in the solution increases due to being expelled by sodium from the exchange complex of soils. I. A. Tyutyunov[13] has established that, during the passage of soil temperatures from positive to negative, the normal pattern in the process of cation absorption Ca > Mg > Na changes to the opposite, i.e., during freezing of saturated soils, we have an intensification in the absorption, by soil, of sodium from the equilibrium solution, forcing the calcium and magnesium ions from the exchange complex. In the warmer layers of the talik located deeper, in proportion to accumulation of sodium hydrocarbonates, calcium precipitates in the form of carbonate. The magnesium and sodium hydrocarbonates accumulate in the solution.

In the water in the freezing sublacustrine taliks, even at a mineralization of 1.5-2.0 g/l, the sodium hydrocarbonates become predominant, while, at a mineralization of 5 g/l, the sodium carbonates become predominant.

On the freezing of subsurface water containing sulfates, their concentration in the layers bordering the frozen soils increases considerably. However, in the sublacustrine taliks of great thickness (even enclosed), sulfates normally do not accumulate because of the processes of desulfatization. Their cryogenous accumulation is possible in the thin taliks beneath shallow lakes freezing to the bottom and also in the taliks forming in the topographic depressions in the regions where the saline soils occur.

As already indicated, in the final stage in the freezing of the suprapermafrost water (enriched by easily soluble salts) of the active layer in the topographic depressions, the chlorides and often the sulfates also migrate into the subjacent permafrost. At a new thawing of the suprapermafrost layer, under the effect of a thermal gradient, the chlorides can migrate back into the warmer layers. However the process of reverse migration of chlorides is possible only at their relatively small concentrations in the solution. As experimental studies have indicated, in ice samples with an overall mineralization of 1.5-2.0 g/l, notwithstanding the higher temperature in the upper layers, migration of chlorides contained in the intercrystalline solutions predominates downward under the effect of gravitation. Taking into account that even in the slightly saline frozen soils the concentration of chlorides in the pore solution is above this value, we can theorize that reverse migration is insignificant.

Consequently in such sectors, chlorides gradually accumulate in the frozen layer, the ice-cement melts, and the frozen soil becomes thawed at a negative temperature. The highly mineralized waters with negative temperature have been accepted in [Soviet] literature under the name of "cryopegs."

In central Yakutia, the cryopegs of the described origin occur fairly often in the sandy deposits beneath the slight depressions in the low terraces along the Lena River, into which the readily soluble salts are transported by the surface, ground, and suprapermafrost water. Mineralization of water in them reaches 60-80 g/l and more. In the water, the sodium and magnesium chlorides predominate, sometimes at increased content of sodium sulfates (see Table 1, Nos. 10 and 11).

These taliks occur in the range of the most significant annual temperature fluctuations (2.5-6.0 m). Beneath the cryopegs, downward along the section, the concentration of the soil pore solution decreases sharply, but the relative content of sodium and magnesium chlorides will remain dominant over the entire section of granular deposits. Their absolute content in the pore solution is determined by the concentration in the talik. Thus, under the cryopegs with a mineralization of 51-20 g/l and a chlorine concentration of 8-10 g/l, its content in the pore solution downward along the section of the subjacent frozen layer of unconsolidated deposits decreases from 2.0 g/l to 0.1 g/l; beneath the cryopegs with a mineralization of 60 g/l and chlorine concentration of 35 g/l, it decreases downward along the section from 4.0 g/l to 0.5-0.7 g/l.

In this manner, the highly mineralized solutions, having formed from the cryogenous concentration in the upper layer of permafrost, migrate downward along the section of the entire stratum of granular frozen deposits under the influence of thermal gradient and gravitation. Beyond the limits of the occurrence of cryopegs (deeper than 4 m), calcium and magnesium hydrocarbonates prevail in the pore water.

The cryopegs revealed by boreholes in the kimberlite pipe in western Yakutia at a depth from 178 m to 190 m have a mineralization ranging from 76 g/l to 97 g/l[15] and of calcium-magnesium chloride composition, which was formed during the soil freezing. The chemical composition of ice taken in the superadjacent layer of permafrost illustrates the stages in the transition of salts to precipitates during the freezing of water-bearing soils: at a depth of 100 m--calcium-magnesium hydrocarbonate; farther down--sodium-calcium sulfate at a mineralization of 4 g/l.

The lenses of interpremafrost saline negative-temperature water (cryopegs) with mineralization of 13-40 g/l of sodium, magnesium, calcium chloride composition, formed on freezing of fresh water in the deposits in the Lower Ordovician, were studied by an expedition from Moscow University in 1965-1966.

In this manner, in the soils varying in age and composition, in connection with the increase in mineralization and with the variations in salt composition of water during freezing, water can ultimately form that is similar in composition and mineralization. The changes in the salt composition of freezing water-bearing deposits determine the chemical composition of subsurface water formed during its subsequent thawing. Moreover, the varying solubility of salts at temperatures close to zero is of significance.

In the thawing soils from which the easily soluble salts were largely forced out during the freezing process, there form fresh or slightly brackish waters of sodium hydrocarbonate or magnesium-sodium sulfate-hydrocarbonate composition; in the soils into which the readily soluble salts have migrated, the highly-mineralized water with salts of sodium, magnesium, and calcium chloride is forming.

In the taliks occurring in frozen soils deeper than 20 m and in the subpermafrost horizons where the temperature changes of soils occur over centuries and milleniums, the effect of cryogenous processes on metamorphization of the water's chemical composition is most gradual.

The various stages in the freezing of water-bearing soils and thawing determine to a considerable extent the diversity in the chemical composition of subsurface water in the regions where permafrost occurs.

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ROLE OF VALLEY TALIKS IN FORMATION OF UNDERGROUND DRAINAGE IN NORTHEASTERN YAKUTIA

V. YE. AFANASENKO Moscow Lomonosov State University

The region under study is coordinated to the Yana-Indigirka interfluve and encompasses areas differing sharply orographically. We include here the southern part of the Yana-Indigirka lowland plain, the Pripolousnenskaya piedmont polygenetic plain, the Polousnyy Ridge, having a middleheight relief and its spurs, such as the Selennyakhskiy, Nemkuchanskiy, Dzhapkuchan-Yuryuyete, Kaglynskiy, and Esteriktyakh-Tas ranges, the Deputatskiy, Argaa-Emnekenskiy, and Sak'yakanskiy mountain masses, and finally the Uyandinskaya intermountain area.

Against the geologic background, there is a series of complex geostructural elements: the Selennyakhskiy fringing anticlinorium of the Kolymskiy median massif formed by carbonaceous rocks of the Paleozoic era; the Yana-Kolymskaya Zone of Mesozoic folding composed of terrestrial rocks of Verkhoyansk complex with intrusions of granitoid composition; a grabenshaped superimposed structure of the Uyandinskaya Depression formed by terrestrial Neogene-Quaternary rocks; and a Piedmont synclinal zone converting in the north to the vast area of Cenozoic subsidences.

The climate in the region is extremely severe and varies from arctic maritime on the north to distinctly continental in the internal parts of the Yana-Indigirka interfluve. The average annual air temperature fluctuates from -12-14°C to -15-17°C, and the amplitude in fluctuations of air temperature ranges from 42°C to 55-57°C. The amount of precipitation comprises 250-350 mm per year, of which more than a half falls during the summer.

Within the limits of the entire area, continuous low-temperature permafrost is widespread.

The extremely scarce and disparate hydrogeologic information on the area under study could not serve as a serious basis for broad generalizations on the conditions involved in the movement of underground water, its circulation with the surface water, the features of drainage, chemical composition, etc. The differentiation of hydrogeologic structures conducted on a number of charts and maps in recent years,[3,6] on the strength of what has been said, was not always substantiated and failed to provide the opportunity of evaluating correctly a number of questions involved in the overall hydrogeologic situation, especially the conditions relating to the formation of subsurface drainage.

In the area under study, we segregated three major hydrogeologic structures[3,6] of second order: an artesian basin in the arctic belt framing the Laptev Sea, Ol'dzhoyskiy adartesian [*sic*]* basin, and the Polousnenskiy hydrogeologic massif. A detailed study of the territory and the compilation of hydrogeologic charts to a scale of 1:500,000 permitted us to refine considerably the boundaries of the enumerated structures, differing by nature of underground drainage. As a result, their total number was increased to five. This is associated with the fact that within the confines of the so-called Polousnenskiy hydrogeologic massif, we differentiated several independent structures, differing significantly among themselves in fault tectonics, types of taliks, and, as a result of this, in nature of water circulation. These two hydrogeologic massifs are the Polousnensko-Tuostakhskiy and Selennyakhskiy. The first of them in its basic outlines coincides with the Polousnensko-Tuostakhskiy anticlinorium, represented by an intensively dislocated thick terrestrial stratum of Triassic-Jurassic period; the second corresponds to the Selennyakhskiy horst-anticlinorium formed by Paleozoic carbonate rocks. In the orographic background, both structures are medium- and low-height, markedly disjointed structures in the Polousnyy Ridge and its spurs.

*The word "adartesian" is undefined and is not used in the USSR nor in North America.

In addition to the two structures specified, it appears correct to distinguish an independent Uyandinskiy postartesian [*sic*]† basin, coordinated to the superimposed Meso-Cenozoic structure with the Mesozoic and Paleozoic folded base, mantled by unconsolidated frozen deposits of varying thickness.

The indicated structure is fairly complex. For the southern half of the basin, we typically find a close hydraulic relationship with the subsurface water in the Selennyakhskiy hydrogeologic massif; however, the northern half is isolated from the Polousnensko-Tuostakhskaya structure framing it.

The processes involved in perennial freezing have been reflected on the basic modification of the paleohydrogeologic conditions of all the hydrogeologic structures that are being differentiated; this permits us to identify them as cryogenous.

These modifications led to the formation, from the surface, of a regional water barrier and to the exclusion from the water circulation cycle of a number of previously existing water-bearing channels. Specific conditions of water circulation and surface water manifestations were created. Runoff was localized exclusively in the taliks in the valley network. Prerequisites for this included the warming influence of the surface and subsurface water that, along with other factors in the natural environment, acted most effectively in the valleys. Most of the existing valleys in the region under study usually have zones of discontinuous tectonic dislocations, and this circumstance is enhanced by the fact that the delivery and discharge of subsurface water of deep circulation through the continuous taliks (infiltration and pressure-infiltration) also occurred and is occurring in the valleys. The processes in the rearrangement of the hydrogeologic conditions under study began during the formation of thick permafrost, most likely in the lower Pleistocene.

The second stage in the variations in the permafrost-hydrogeologic conditions is continuing at the present time. It reduces to modifications in the nature of taliks and their role in water circulation within the limits of the same valleys with which the subsurface drainage spatially coincides. This finds a reflection in the morphological rearrangement of the floors of a number of valleys and in variations in the general geologic conditions: freezing, thawing of deposits, mechanical undermining, silt-deposition, etc., and also in changes in the types of taliks and in the naleds associated with them.[1,2]

For example, the Uyandinskaya superimposed depression formed by supes, sand, and gravel beds prior to the time of freezing represented an intermontain artesian basin with pore and stratal-

†The term is unfortunate since the suffix "post" has a time connotation, whereas the freezing of the mantle to the artesian structure leads to its modification. It would be better to employ the term "cryogeologic basin"--Editor's Remark.

pore water. The process of directed perennial freezing led to the refreezing of unconsolidated water-bearing soils, and at present the concentrated movement of subsurface water is accomplished at the lower boundary of the permafrost layer in the bottom parts of the valleys along the fissures, caves, and karst-type cavities of carbonaceous rocks in the Paleozoic foundation of the basin.

In the areas of the discharge of water via the hydrogeogenic through-taliks, the rocks have a pore-fissure permeability, although prior to the perennial freezing, they constituted relative water barriers. Quite an important role in such a unique transformation of these rocks by individual linear sectors was played by the coordination of the water-bearing through-taliks and the zones of the most active tectonic movements.

We should also discuss the fact that the processes of prolonged freezing of the lithosphere have led to major transformations in the hydrogeologic masses and in the overall conditions involved in the existence and movement of subsurface water in them. In particular, they caused the refreezing of interstitial flooded zones in the upper part of the section and the concentration of subsurface water at the lower boundary in the permafrost layer.

Against the background of what has been said, the term "cryogenous" is used by us in a somewhat different sense from what it is utilized by a number of authors.[4]

The differentiation of the Ol'dzhoyskiy adartesian basin on a number of charts and maps[3,6] appears incorrect to us. The fact is that the boundaries of the structure on these maps correspond to the outlines of Polousnyy Ridge, Nemkuchanskiy Range, and the Argaa-Emneke mountain massif. This is a typical hydrogeologic mass, called the Polousnensko-Tuostakhskiy by us. However, the hydrogeologic structure of an adartesian type should be differentiated on the northern slopes of Polousnyy Ridge and the adjoining polygenetic plain.

As a result of the severe permafrost conditions in northeastern Yakutia, as already indicated, the subsurface drainage is localized and coincides with the valley drainage system, wherein we note a definite connection in the coordination of specific types of taliks to the various hydrogeologic structures.

The inflow taliks through which the supply of subpermafrost water is accomplished occur most widely in the Selennyakhskiy hydrogeologic massif. Such taliks are confined to the riverbed and floodplain parts of the valleys, chiefly within the limits of mountain structures, in regions of intensive dislocations and tectonic disruptions. The areas of these taliks are usually small and their dimensions in cross section vary from several tens to 200-300 m.

In these taliks we observe in winter an abrupt drop in the levels of subsurface water and the development of large zones of drying (up to 200-300 m). We note a partial freezing of talik-type zones in the upper part; basically, however, the fall in levels exceeds the rate of seasonal freezing.

The sizes of drying zones were established during systematic year-round observations of the discharges from springs in the zones of pressure in filtration and their comparison with the areas of supply to the Selennyakhskaya and Uyandinskaya hydrogeologic structures. The summertime outflows from the springs usually decrease by 3-4 times the winter critical period. The principal movement of subsurface water coincides with the direction of valleys and occurs in the form of subpermafrost flow. Since the cross section of this flow and the permeability of soils remain practically unchanged throughout the year, it becomes understandable that the main cause of seasonal variations in discharges is the decrease in the hydraulic gradients. The extent of these variations will comprise 200-300 m.

In spring during the thawing of snow and the occurrence of surface runoff, the natural supplies of subsurface water are rapidly restored in the course of a few days. The movement of subsurface water along the inflow type taliks in the supply areas and farther along in the subpermafrost fissured part of the section toward the discharging areas is fairly intensive. The velocity of flow attains 50-100 m/day, while the overall cycle of water circulation averages from 2 to 15 or 20 yr.

We should stress that the type of infiltration taliks that we are discussing in the heating areas is developed in the peripheral horst parts of the Kolymskiy median mass with distinct brokenness of relief (Selennyakhskiy Range), while the pressure-infiltration type of taliks is developed in the grabenlike superimposed structures (Uyandinskaya Vpadina [Depression]).

A distinguishing feature of the enumerated structural elements is their high tectonic activity over the extent of the entire Cenozoic period. However, within the limits of regions (relatively stable from a tectonic standpoint) formed of the same carbonaceous rocks in the Paleozoic period, for example, in the Esteriktyakh-Tas Range, the development of talik zones decreases sharply and in general the formation of large naleds does not occur. In this manner just in the sectors of active tectonic shoves where the carbonaceous rocks occur most widely, notwithstanding the intensive freezing of the lithosphere, the through-talik zones were able to remain preserved and to exist until the present day.

To a lesser degree the infiltration-type taliks are distributed in the Polousnensko-Tuostakhskaya and Pripolousnenskaya structures. This is associated primarily with the different lithologic composition of rocks, the extent of their fracturing during tectonic shoves, extent of exposure to karst processing, with the water-conducting properties, etc.

Here the overall hydrogeologic situation is different. Specifically, the through-taliks similar to those mentioned above are detected only in isolated instances. In these structures the processes of directed freezing have led to the exclusion, from the water circulation cycle, of a number of water-bearing channels, to the freezing of supply areas and to the development of areas of intensified pressures and to the disconnection of hydrogeologic systems having previously existed.

A confirmation of what has been said is the hydraulic separateness of the Polousnansko-Tuostakhskiy hydrogeologic mass from the northern half of the Uyandinskaya Vpadina and the formation of a varying type of subsurface water under the different geochemical conditions. At the present time, the mountainous part of the Deputatskiy mountain mass and of the Nemkuchanskiy Range is isolated from the northern part (connected hydraulically with them previously, prior to the time of perennial freezing) of the Uyandinskaya superimposed basin. A result of the freezing processes was the uninterrupted (without through-taliks) distribution of frozen layers from the surface in the mountainous part of the territory and the stagnant nature of the subpermafrost water.

The subsurface water in this region is characterized by the limited supplies and is brackish (its mineralization ranges from 2 g/l to 4 g/l of calcium-sulfate type. However in the northern part of the Uyandinskaya Vpadina, there occurs fresh water of sodium-hydrocarbonate type, typified by a fairly intensive water circulation. The supply of this water is achieved along the peripheral part of the basins following the rejuvenated fault zones in the Mesozoic folded basement and in the Cenozoic mantle.

It appears that the development of areas of intensified pressures of subsurface water within the limits of the Deputatskiy mountain mass, the significant qualitative difference from the water in the northern half of the Uyandinskaya Vpadina, and the difference in mineralization and temperatures confirm the opinion concerning the isolated state of these structures.

Of great scientific interest is the lack of conformity that we observe in the dimensions of the naled fields in contrast to the naleds currently forming. Many of these naleds are gigantic. In particular, the volume of ice accumulating along the Sakyndzha River exceeds 15 million m^3; along the Taryng-Yurekh River, the volume exceeds 4 million m^3. The existing nonconformities in areas of developing naleds and of naled fields[1,2] confirm the presently-occurring reconstruction of the overall hydrogeologic conditions in the process of complex interaction of subsurface water and frozen layers.

Also undergoing changes are the spatial positions of taliks.[1,2] Thus within the confines of the talik along Sakyndzha River, which we have mentioned, there occurred a change in the configuration of the area of pressure-infiltration, its localization to 300-400 m, and a simultaneous shifting by 1,000-1,200 m toward the right-hand side of the valley.

In the northern foothills part of the Polousnyy Ridge, the valleys of a number of rivers at the present time have a general naled appearance. Obviously, in a quite recent geologic period, there occurred here the discharging of subsurface water of the structure via the pressure-flow taliks. Such through-taliks were disclosed during the work on a number of river valleys. The naleds currently forming here are insignificantly small as compared with the existing naled fields and do not exert any perceptible influence on the overall appearance of the valleys. According to the volume of accumulating ice, they are classified chiefly as small. The rate of naled formation normally does not exceed thousandths of a liter per second per square kilometer.

The facts cited bear evidence that the overall cooling of the lithosphere during the Quaternary period excluded from the general water circulation cycle a series of infiltration and inflow taliks in the highest (according to absolute elevation marks) parts of the valleys. Naturally this condition was reflected in a change in the hydraulic pressures and in final analysis led to a reduction or complete cessation of pressure-infiltration in a number of taliks in the foothills and mountainous part of Polousnyy Ridge, where large naleds had become formed in the recent geologic past.

In conformity with what has been said, of considerable interest is the through-talik in the valley of Nuchchi River with a width of 100 m and length of 400 m. During the year, this talik "window" operates under varying conditions: in summer as a pressure-infiltration talik and in winter as an infiltration-type talik.

Of undoubted scientific and practical interest are the hydrogenic underground and the floodplain gravity-infiltration taliks participating in feeding the large structures of the Pripolousnenskiy adartesian basin type. Such taliks were first detected and studied within the limits of the polygenetic plain adjoining the Polousnyy Ridge on the north. The thickness of such taliks reaches 30-50 m, increasing to 80-100 m in the sectors of superimposed grabenlike structures. The width ranges from 80-100 m to 700-1,000 m. In such taliks, a powerful underground runoff, varying slightly during the year, is accomplished. For example, in the middle reaches of the Tenkela River, the right-hand tributary of Khroma River, the quantity of subsurface water percolating chiefly into the alluvial soils reached 7.6 m^3/yr. This value corresponds to an underground runoff of 0.40 $l/s \cdot km^2$.

By the studies conducted, it was proved that a large share of this quantity of water is expended in supplying water for the subpermafrost circulation through the continuous infiltration-type taliks that are coordinated to a series of regionally traced discontinuous tectonic zones within the limits of the Pripolousnenskiy adartesian basin.

The role of condensation in the formation of the surface runoff and, accordingly, of the near-surface runoff in Polousnyy Ridge and its spurs is considerable. The value for the condensation component determined on the basis of results from systematic observations is estimated for the various valleys to be 200-500 mm/yr.

Within the limits of the hydrogeologic structures that we have demarcated, we have made a quantitative estimation of the natural resources of subsurface water. For a number of structures (Selennyakhskiy hydrogeologic mass, southern half of the Uyandinskiy postartesian basin), such an estimation was conducted based on the naled supply rate; for the other structures (Pripolousnenskiy adartesian basin and others), the assessment was based on the drainage in the subsurface and floodplain taliks.

The information obtained permits us to refine our concepts concerning the distribution, movement, and chemistry of subsurface water; as a result of this, the natural resources of subsurface water calculated for a number of hydrogeologic structures differ appreciably from those indicated in the published reports.[3,5,6]

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BASIC FEATURES IN HYDROGEOLOGIC STRUCTURES OF THE PERMAFROST AREA IN THE USSR

YE. A. BASKOV AND I. K. ZAYTSEV All-Union Order of Lenin Scientific-Research Geologic Institute

Permafrost underlies about 60 percent of the entire area of the USSR. It occupies the Pechorskaya, western Siberian (except the southern fringe), eastern Siberian, and Yana-Indigirka artesian oblasts; Taymyrsko-Severozemel'skaya, Severoural'sko-Novozemel'skaya, Verkhoyano-Kolymskaya, Chukotskaya, Patomo-Vitimskaya, Baykalo-Charskaya, Daurskaya, and other hydrogeologic folded regions.[3] Moreover, it includes the large hydrogeologic structures (chiefly artesian) developed within the limits of the Arctic Ocean basin.[7]

The contemporary hydrogeologic conditions in the various hydrogeologic structures of the permafrost region became formed from the interaction of the complex aggregation of endogenic and exogenic factors during the extended geologic history of their development. In the Quaternary period, in the process of the formation of these hydrogeologic structures, there were superimposed the important factors of cryogenesis. This led to the freezing of rocks to a depth from several tens of meters to the first thousands of meters; to the transfer of a significant amount of water from the liquid phase to the solid; to a variation in the conditions and nature of rocks' permeability, conditions of supply and discharge of subsurface water, and variation in hydrochemical and hydrothermal zonings; and to the appearance of cryogenous pressures and of unique chemical types of subsurface water having formed under the effect of cryogenous processes.[8]

Moreover, the effects of permafrost on the overall hydrogeologic conditions prove to be so significant that they could be used as the basis for a subdivision of types of hydrogeologic structures into subtypes,[3,7,9] specifically artesian (and similarly) basins are divided into subtypes:

1. The permafrost is insular.
2. Continuous permafrost occurs with a thickness less than that of the zone of fresh water.
3. The thickness of permafrost equals, or is greater than, that of the freshwater zone; saline water occurs directly below the permafrost.
4. Thickness of permafrost equals, or is

greater than, the thickness of the zone of fresh and saline water; brines occur directly beneath the permafrost.

5. Thickness of permafrost is greater than that of the mantle of the artesian basin. In the mantle of such basins, in liquid phase we can encounter only saline or brine water, the freezing temperature of which is higher [*sic*] than the temperature of frozen soils.

6. As a special subtype, we can probably identify the basins within the limits of which there is two-layered frozen ground separated by thawed soils of considerable thickness having a regional significance.

The hydrogeologic masses (and structures similar to them, i.e., admasses, supermasses, etc.), depending on the distribution of permafrost, are also divided into subtypes:

1. The insular permafrost occurs chiefly on the mountain peaks; these are the "peak-frozen" hydrogeologic masses.
2. Thickness of permafrost is less than that of the zone of fissured rocks.
3. Thickness of permafrost is greater than that of the zone of fissured rocks; only the sectors of coarse and extensive tectonic fracturing of the rocks can prove to be water-bearing.

In chemical composition and extent of mineralization, the subsurface water in the permafrost areas is extremely diversified. Along with the fresh water of hydrocarbonate content, we find occurring widely here the chloride saline and brine water (including brines with mineralization up to 410-415 g/kg) and others. As a result of the compilation from 1955 to 1970 in the All-Union Scientific Research Institute of Geology of a series of survey hydrochemical USSR maps,[4,5] we revealed the basic features in the hydrochemical zonality of subsurface water in the USSR, including the permafrost areas. For this region, there is clearly shown in section and in plan the zonality of phase state, mineralization, and chemical composition of subsurface water. The hydrochemical zones differentiated by the indications of the extent of mineralization and chemical composition of subsurface water, permafrost areas, are also subdivided into two groups: the zones in which the subsurface water exists in liquid phase and the zones within which the subsurface water exists in liquid or solid phases.

Within the permafrost area, we established several types of hydrochemical sections differing in nature of alternation in section, of subsurface water, having various phase states, degrees of mineralization, and chemical compositions. It has been found that in the area, in the form of hydrochemical belts, the basic features in the structure of the artesian basins (and in structures similar to them) are closely linked with the lithologic-facies features in the water-bearing complexes. Thus within the limits of the western Siberian artesian area where the terrestrial sandy-clayey layers of Mesozoic age are mainly distributed, the hydrochemical belts are characterized by the presence of fresh water (up to 1 g/kg) in the upper parts of the section, of saline water (up to 35 g/kg), and less often of weak brines (to 70-80 g/l) in the lower horizons. In the northern regions of this area, the fresh water is often completely frozen.

In the eastern Siberian artesian area in the regions where halogenic-carbonate layers of Paleozoic age occur, in the structure of hydrochemical belts, a major role is played by the zones of strong and very strong (up to 450 g/kg and more) brines; the thickness of these zones reaches 3,000-4,000 m. In those regions where the Paleozoic halogenic-carbonate layers daylight, the total thickness of zones of fresh and saline water in them often does not exceed 200-400 m. In the northern and central parts of the platform where halogenic-carbonate layers of Paleozoic age daylight, the fresh and saline water is usually frozen and, beneath the frozen layers, the brines are revealed at once.

In this manner, the hydrochemical zoning of the artesian basins (and structures similar to them) in the permafrost region (just as of the basins situated outside of it) is determined chiefly by the conditions of their geologic development (by the nature of the component formations, tectonic regime, extent of erosion, etc.). The significant differences in the hydrochemical zoning concern only the upper parts of the section of each basin, where, in the permafrost areas, the main body of gravitational water has been frozen and the specific permafrost processes developed; these processes were the concentration of subsurface water and the variation in its salt composition as a result of precipitation of less soluble compounds from the solution at low temperatures and the accumulation of the more soluble compounds in the solution. In particular the processes of freezing and thawing lead to the accumulation of sodium hydrocarbonates[1] in the suprapermafrost and intrapermafrost water. Possibly these processes cause the accumulation, in the supercooled suprapermafrost water, of magnesium and calcium chlorides (southern part of Olenekskiy basin).

The zonal variation in the thickness and extent of discontinuity in the permafrost (generally from south to north) determines the same variation in the extent of mineralization and chemical composition of the upper horizons of subpermafrost water in the artesian basins. In the insular permafrost area, their upper subpermafrost water is chiefly fresh. In the area of continuous permafrost, it is chiefly saline and brine.

In the hydrogeologic masses, superbasins, and structures similar to them, fresh water usually occurs under the permafrost; this is caused by its highly leached-out state and the great thickness of the freshwater zone, which developed prior to the formation of permafrost. In the deeply frozen hydrogeologic masses in the USSR Northeast, we find brackish water of sulfate composition under the permafrost. The saline and brine chloride subpermafrost water is known in the slightly elevated hydrogeologic masses near the seacoasts having existed under subaqueous conditions[5,7] during the Quaternary period.

The subsurface water temperatures in the permafrost areas in the USSR vary from -5° to -6°C to +100° to +150°C and more. Within those areas, we can differentiate several hydrothermal zones:

1. Supercooled water with temperature less than 0°C.

2. Cold water, with temperature ranging from 0° to 20°C: (a) very cold, i.e., 0°-4°C, (b) moderately cold, i.e., 4°-20°C.

3. Heated water, with temperature from 20°-100°C: (a) warm, i.e., from 20° to 35°C, (b) hot, i.e., from 35° to 100°C.

4. Superheated water with temperature higher than 100°C.

In the various regions of the permafrost areas, we note in the section various combinations of the enumerated zones, forming different hydrothermal belts.[5]

The hydrothermal belt of supercooled (temperature less than 0°C) and very cold (0-4°C) water is located in a thick permafrost zone in which the fresh water exists chiefly in solid phase, while, within the permafrost layer and beneath it, there occur saline and brine waters, the freezing temperature of which lies below 0°C and depends on the degree of mineralization. This belt occurs widely in the northern part of the eastern Siberian artesian oblast, particulary on the slopes of the Anabarskiy massif, where the thickness of supercooled brine water reaches 1,400-1,500 m.[6]

The hydrothermal belt of extremely cold and moderately cold water (0°-4°C and 4°-2°C) occupies extensive areas in the hydrogeologic folded areas of the USSR Northeast, Taymyr, northern Urals, etc., and also fairly often in the peripheral parts (contiguous to the fault areas) of eastern Siberian, western Siberian, and Pechorskaya artesian regions. In the internal parts of the artesian regions, we usually find the hydrothermal belts; in the lower parts of their section, there appear zones of warm, hot and even superheated subsurface water. The superheated water occurs extensively in the northern half of the western Siberian artesian regions. We can theorize that the hot and superheated water also occurs at a great depth under the permafrost in the deep basins of the eastern Siberian artesian regions.

The permafrost in the upper parts of the section of the hydrogeological structures exerts a significant influence on the conditions of the supply and discharge of subsurface water and on hydrodynamic zonality of structures.

In the artesian structures in the regions of continuous permafrost, the zone of free exchange comprises only the suprapermafrost water. The conditions of supply of subpermafrost water from atmospheric precipitation in these regions are extremely complicated (northern parts of the Pechorskaya; western Siberian artesian regions; Khatangskiy, Kotuyskiy, Olenekskiy, and northern part of the Yakutsk basins of the eastern Siberian artesian regions, and others). Attention is drawn to the fairly frequent low (up to 100-500 m below the downcutting of river valleys) positions of piezometric levels of subpermafrost water in these regions; this is most likely caused by the discharge along the fault zones of intensively gas-saturated subsurface water (with the formation of natural gas lifts) in the absence of the supply of subpermafrost water from atmospheric precipitation.[2]

In the regions of the talik, and particularly of the insular, permafrost, the zone of free-water circulation encompasses the upper horizons of subpermafrost water up to approximately the same depths as in the unfrozen regions (with similar hydrogeologic conditions). This is confirmed by hydrogeologic investigations conducted, e.g., in the southern part of the Pechorskaya artesian region.

In the hydrogeologic fault regions with extremely broad distribution of permafrost (northeastern USSR, etc.), the hydrodynamic conditions are determined to a considerable extent by the most recent tectonic movements accompanied by the appearance of new faults or the rejuvenation of old faults. The effect of the most recent tectonic movements on the distribution of naleds is particularly noticeable.[10,11]

Further study and detailed mapping of these features are necessary for firming up the trends and techniques utilized in the exploratory activities for subsurface water suitable for the various requirements of the national economy.

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HYDRODYNAMIC AND HYDROCHEMICAL PROCESSES IN THE AREA OF THE COOLING OF EARTH'S CRUST

G. D. GINSBURG AND YA. V. NEIZVESTNOV Scientific-Research Institute for Arctic Geology

The concept concerning the area of earth's cooling has been introduced by V. I. Vernadskiy;[3] this area includes the upper layers of rock strata that form the land and underlie the seas and oceans.

Geothermal investigations launched in recent years and accompanying the deep drilling that has advanced considerably into the permafrost regions have confirmed the opinion of V. I. Vernadskiy concerning the considerable extent of the scale of cooling. Although the determination of the depths to which during the late Cenozoic period the temperature decrease extended and the establishment of the temperature ranges at specific depths still represent problems for the future, even now we can assert with certainty that in the arctic and subarctic regions the area of cooling extends at least to depths of the first few kilometers. Thus, in Yakutia ASSR in the Markha River basin, the depth of negative temperatures reaches about 1,500 m.[10] In the basins of the lower reach of the Yenisey River and the middle reaches of the Pyasina River, judging by the comparison of the contemporary and paleotemperatures established on the basis of the extent of breakdown of fine organic material of soils and according to a number of other indirect indications, the cooling reached depths of not less than 2 km, while the drop in temperature amounted to 15°-25°C.[5] The permafrost thickness in this region attains 560 m.

From a hydrogeothermal standpoint, the area of the cooling of Earth's crust includes the frozen zone, the zone of negative-temperature water (cryopegs*) and also the upper part in the zone of positive-temperature water ("pegs"* and hot springs), itself having experienced the effect of a temperature drop at the Earth's surface.

It is known that the freezing of the Earth's crust can cause considerable freezing pressures in the roof of the subpermafrost water-bearing horizons; degradation of the frozen zone from below is capable of leading to an abrupt lowering of the piezometric levels.[12] Similar phenomena occur at change in temperature at water-bearing soils produced by the irregular expansion or compression of water and soil, and as a result

*There are no English equivalents to "peg" and "cryopeg," but they mean a body of fresh water above 0°C, and a body of unfrozen salt water at a negative temperature, respectively.

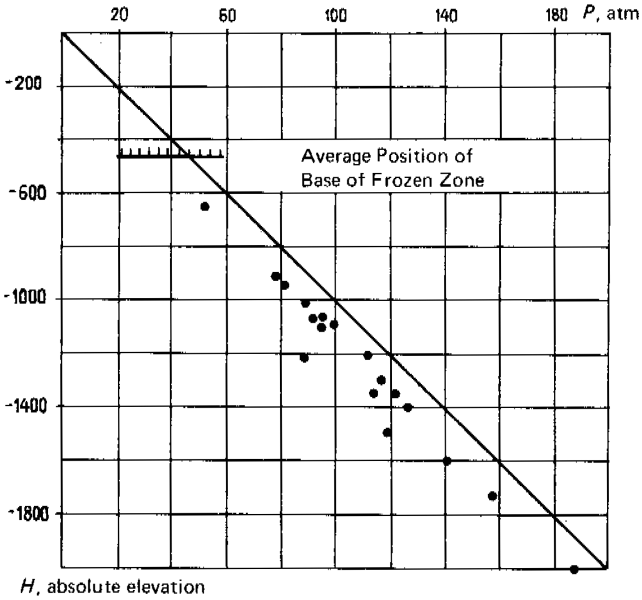

FIGURE 1 Distribution by depth of stratal [water] pressures in Jurassic and Cretaceous deposits in the area of Rassokhinskiy Bank (Yenisei-Khatangskiy regional downwarp, basin of Pyasina River in comparison with the tentative hydrostatic pressure, figured from sea level, diagonal line).

of the considerable difference in their coefficients of thermal expansion. An inspection of a model of saturated porous soil indicates that the cooling of a water-bearing horizon (in the range of temperatures from the point of maximum density and above) will always lead to a reduction in stratal pressures, while heating will lead to an increase.[6] The combined effect of freezing and cooling of the Earth's interior during aggradation of the frozen zone, thawing, and warming during degradation can cause both descending and ascending movements of water. The hydrodynamic situation becomes especially complicated at simultaneous but opposite changes in temperature at different depths. In a number of cases, the processes listed lead to the formation of "anomalous" stratal [water] pressures expressed in negative values of the adduced levels of water-bearing horizons, below 300 m.

The area of cryogenous pressures and of "anomalous" low pressures in the water-bearing horizons is very broad; they are known on the islands of the Soviet Arctic in the basins of the Pechora, Ob', Yenisei, Pyasina, Lena, Kolyma, and Anadyr' rivers. In Figure 1, as an example, we have shown the pressure distribution in the water-bearing horizons in the Jurassic and Cretaceous beds in the area of Rassokhinskiy Bank. The occurrences under study are not always susceptible to unambiguous explanation; however, their relationship with the hydrogeothermal phenomena is undoubted. A major role in the preservation of the "anomalous" pressures belongs to the actual frozen zone, supporting the greater or lesser relative closed state of the hydrodynamic systems.

The features in the distribution of stratal pressures in the cooling area are determined chiefly by the variation in water density. However, density is not the only property of water uniquely dependent on temperature; among such properties we also include viscosity, surface (interphase) tension, and adsorptivity. It is know that, with a drop in temperature, the numerical values of these properties rise, particularly near the melting point. From the standpoint of contemporary physicochemical concepts, these properties of the water are determined by the presence of hydrogen bonds and by the content (in liquid water) of icelike molecular structures;[2] Figure 2 shows their dependence on temperature. The decrease in mobility and in migration capacity of water at a reduction in temperature causes a number of interesting and important consequences: It obstructs the "flushing" of hydrogeologic structures, complicates the entry of oil and gas into, and their departure from, the traps; it predetermines the chiefly flexible regime of the development of these mineral deposits. An example of such specific deposits is provided by the Messoyakhskaya gas bed

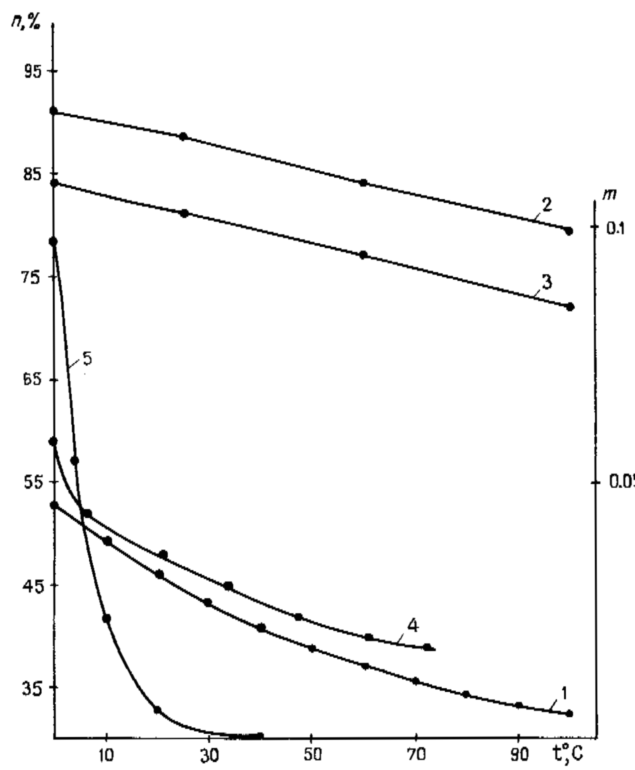

FIGURE 2 Variation in molecular structure of water as a function of temperature. The number of water molecules with preserved hydrogen bonds (n) based on data 1--Nemethy, Scherage (1962); 2--Haggis, Hastad, and Buchanan (1952); 3--Grant (1957); 4--Buijs and Choppin (1963); 5--molar concentration, in water, of clusters of ice type I (m) according to John, Grosh, Ree, and Eyring (1966). The graph was constructed from the tables in the book by A. M. Blokh (1969, Tables 7-9, 12).

on the west of the Yenisey-Khatangskiy downwarp, in which the stratal temperature is about 10°C.

An increase in the viscosity of water as a result of reduction in temperature over a significant area of the water-bearing horizon is tantamount to an increase in the hydraulic resistance and should evoke the retention of subsurface water upward along the flow. It is possible that this factor caused to a considerable extent the distribution of static levels in western Siberia for the southern part of which the excess pressures are typical.

The specific features in the hydrodynamic situation in the cooling area are reflected in the intensity of underground chemical flow, and in the rate and nature of underground chemical erosion. In the areas of artesian basins, where the deep water-bearing horizons containing highly mineralized chloride water become cooled, the transport of chlorides into the rivers diminishes; in the case of the heating of water-bearing soils (e.g., after degradation of frozen zone), the discharge of chlorides increases sharply. For example, in the Norilka River basin the ionic rate of chlorine discharge is 1.5-1.9 tons from 1 km^2 per year, while the content of chlorine-ion in the water of rivers does not reveal any dependence either on the discharge of water or on the season of year. In the basins of the lower Tunguska (in the lower reaches) and Sukhaya Tunguska rivers, this rate reaches 7.3-15.4, at a distinct increase in concentration of chlorides with decrease in water discharge during the winter low-water period; this testifies to the subsurface nature of the runoff.

The intensive expulsion of salts into the deep horizons and the surrounding basins is accomplished by the highly mineralized cryopegs, moving both under the influence of the gradients of cryogenous pressures as well as under the effect of convection. These processes are extensively developed on the island and continental coasts of the arctic seas.

The uniqueness of the hydrochemical processes in the Earth's crustal cooling is clearly shown in the freezing of soils saturated by mineralized water. The freezing causes the division of subsurface water into fresh or slightly saline ice in the frozen zone and into highly mineralized subpermafrost water. In the solid phase, along with the ice, there are released the most difficultly soluble salts, normally precipitating in the form of crystal hydrates. The mineralization of liquid phase increases owing to the accumulation of readily soluble salts.

The freezing of soils saturated by water of marine composition, occurring on the coasts of arctic seas, in the range of temperatures from -1.9° to -7.8° through -8.2° occurs with the separation into the solid phase only of pure ice and calcium carbonate. The mineralization of the solution existing in equilibrium with the solid phase increases at this time to 120 g/kg.[7] At lower temperatures, there also occurs the precipitation of mirabilite [$Na_2SO_4 \cdot 10\ H_2O$) and, in spite of this, a further increase in mineralization. On the islands of the Soviet Arctic beneath the frozen zone, the boreholes have disclosed water with mineralization ranging from 37 g/l to-191 g/l[11] or 36 g/kg to-166 g/kg, the composition of which is almost identical to the composition of water obtained during a test freezing of seawater by K. E. Gitterman[7] in the temperature range from -2°C to -15°C. By analyzing aqueous extracts of soils in the frozen zone, it was eatablished that the readily soluble salts in them contain 70 percent of sodium sulfate. The separation of mirabilite could also be seen.

Thus the freezing of soils saturated with seawater is accompanied on the one hand by their sulfatization and to a lesser extent by carbonatization; on the other hand, by a variation in chemical composition of water remaining in the liquid phase. The variations in the chemistry of seawater primarily concern their anionic composition. The hydrocarbonate-ion is almost entirely removed from the water, while the content of sulfate-ion decreases from 10 to 1 percent equivalent. The relative content of cations in the freezing seawater usually changes very little. Only the presence, in the water-enclosing soils, of readily soluble salts can exert a significant influence on the change of the cationic composition of water. As a result of the freezing and ionic exchange, under the near-surface conditions (according to V. A. Sulin), the sulfate-free calcium chloride water forms, similar to the water in a series of petroleum deposits occurring at depth greater than 5,000 m. On Franz-Josef Land, in the basalts with calcitic amygdules under the frozen zone, water of the following composition is revealed:

$$M_{89.4}\ \frac{Cl\ 98\ SO_4\ 2}{(Na + K)65\ Ca\ 18\ Mg\ 17}\ .$$

The subpermafrost water in the solid basalts is characterized by a cationic composition normal for seawater.[11]

In the complete freezing of seawater, the content of microcomponents increases. In the strongest underground brines, the bromine content reaches 360 mg/l and that of strontium reaches 75 mg/l.

During the freezing of soils saturated by mineralized water, differing in its chemical composition from seawater, the change in the liquid phase of the freezing layer occurs according to the same pattern. Mineralization of the freezing-out water increases owing to the increase in the content of the salts that dissolve most readily. At repeated freezing and thawing, in a number of cases the sulfate brines develop with mineralization ranging from 149 g/l to 212 g/l, containing more than 40 g/l of iron and 0.02-1.0 g/l of copper, arsenic, and zinc.[9]

The increased density of subpermafrost water concentrating during cryogenesis causes its convective descent, the depth of which and the vertical hydrochemical zonality forming at this time are determined by hydrogeologic conditions. On the islands in the Arctic, in the range from several tens to the first hundreds of meters, we note an increase in mineralization from 40-90

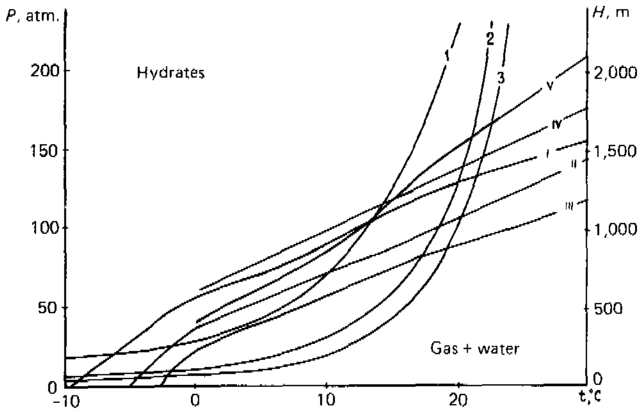

FIGURE 3 Determination of range of possible location of natural gases as hydrates.[4] 1-3--equilibrium curves reflecting formation of hydrates of natural gases with specific weight of 0.555; 0.6 and 0.7 in coordinates of pressure (*P*)-temperature (*T*). I-V--temperature distribution by depth (*H*) in areas of Yenisey-Khatangskiy downwarp with varying thickness of frozen zone.

g/l to 160-190 g/l; at great depths beneath the strong brines, water of appreciably lower mineralization is found.

The reverse hydrochemical zonality, forming in this way and connected with the presence of brines in the cooling zone, is typical for the islands and for a number of sectors along the continental shore of the arctic seas. In particular, V. Ye. Glotov[8] established such a zonality in the USSR Northeast. The hydrochemical situation becomes complicated still more by the processes of the expulsion of brines from under the frozen zone by the fresh water seeping through the talik "windows." On certain islands in the Arctic, the formation of taliks is possible beneath the thick glacier caps, in the near-bottom part of which the temperatures equal 0°. N. I. Obidin[11] links the penetration of fresh water with the formation in Spitzbergen of a zone of fresh and saline water above the brine water, below which apparently the saline waters again occur. In the cooling area, an extensive zone of fresh water can also form during the repeated aggradation-degradation of the frozen zone as a result of the cryogenous separation of subsurface water, the expulsion of freezing mineralized water, and the subsequent thawing of fresh water in the soils. Possibly this explains the existence of fresh water at depth of about 1,000 m in the Lena-Vilyuy and Pechora artesian basins.

The process of freezing of fresh subsurface water studied by N. P. Anisimova[1] in the example of the underlake taliks on Yakutia is accompanied by an increase in mineralization of the liquid phase to several grams per liter, by a partial capture by the soil of the sodium, and by precipitation of calcium carbonate and the accumulation, in the solution, of hydrocarbonates, chlorides, and, in a number of cases, of sulfates and magnesium and sodium carbonates.

The combination of all the enumerated hydrochemical processes causes the cryogenous leaching of the earth's crust, intensifying as a result of the mechanical breakdown of soils on freezing of water in their pores, fissures, cavities, and other voids.

The cooling of the Earth's crust leads to the stabilization of matter. In addition to the already-mentioned precipitation of salts during the process of freezing and exclusion from the circulation of vast water masses, it is shown in the formation of compounds of natural gases with water, namely of solid crystal hydrates. This process is also possible at positive temperatures at depths greater than 1 km. Judging by the data from drilling in the ocean, the gas hydrates have been detected at 600 m below the bottom, at oceanic depth of 3,600 m. As is evident in Figure 3, in the permafrost areas, the depth of the range in which the occurrence of gases in hydrate phase is possible, under actual geologic conditions, is determined by the thickness of the frozen zone. The formation of gas hydrates as a result of bonding of water leads to a reduction in stratal water pressure and to an increase in mineralization of the water coming in contact with the hydrates. The manifestation of the various hydrodynamic and hydrochemical processes leads to a great diversity in the actual hydrogeological conditions in the cooling zone. On the one hand, we encounter varying stages of the identical process, and, on the other hand, we are dealing with various processes and conditions: aggradation or degradation of frozen zone, subaqueous or subaerial stage in development of a region, an arbitrary hydrogeologic structure (artesian basin with alternation of water-conducting and waterblocking horizons, or a hydrogeologic mass with various fissured reservoirs), and so on.

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EFFECT OF MANTLE GLACIATIONS ON PERMAFROST-HYDROGEOLOGIC CONDITIONS IN PERIGLACIAL ZONE OF EUROPE

G. KLATKOVA (Institute of Geography at Lodz University, Poland)
AND N. N. ROMANOVSKIY (Moscow Lomonosov State University, USSR)

The effect of continental glaciations, which covered the north of Europe, western Siberia, and North America in the Pleistocene, on the variation in the permafrost-hydrogeologic conditions in these regions had not been investigated until recently. At the same time, the ice sheets and permafrost that formed in the periglacial zone doubtless must have led to a modification in the drainage system of subsurface water and its manifestation at the surface. Among the paleogeographic factors, the paleohydrogeologic aspect proved to be overlooked. The present paper is devoted to certain results of studies in this field.

In the periglacial topography of the regions adjoining the Baltic Sea from the south, including Poland and East Germany, and possibly also the USSR Baltic republics, we detect traces of active naled-forming and other processes associated with subsurface water. Obviously the sphere of their activity includes the preglacial spaces in certain stages of the Rissian and chiefly of Würmian glaciations. The consequences from these processes are detected both in the relief and in the deposits of the preglacial zone, primarily in the ice marginal drainage channels, the channels of small and medium-sized periglacial rivers, and partly on the outwash plains.

The ice marginal drainage-channels, along which there is drainage of surface and groundwater at various stages of glaciation, comprise one of the main macroelements in the topography of the indicated regions. The ice marginal drainage channels are extended in a sublatitudinal direction roughly parallel to the frontal scarps of the former ice sheets. We differentiate the Vratslav-Magdeburg, Barychi, Varsha-Berlin, Notec-Warta (or Torun-Eberswalde) and Maritime (or Lebskaya) ice marginal drainage channels.[9] The ice marginal drainage channels that were formed during the Würm epoch have been expressed most clearly. The maximum number of studies[7,10-13] that the authors have utilized have been devoted to these ice marginal drainage channels.

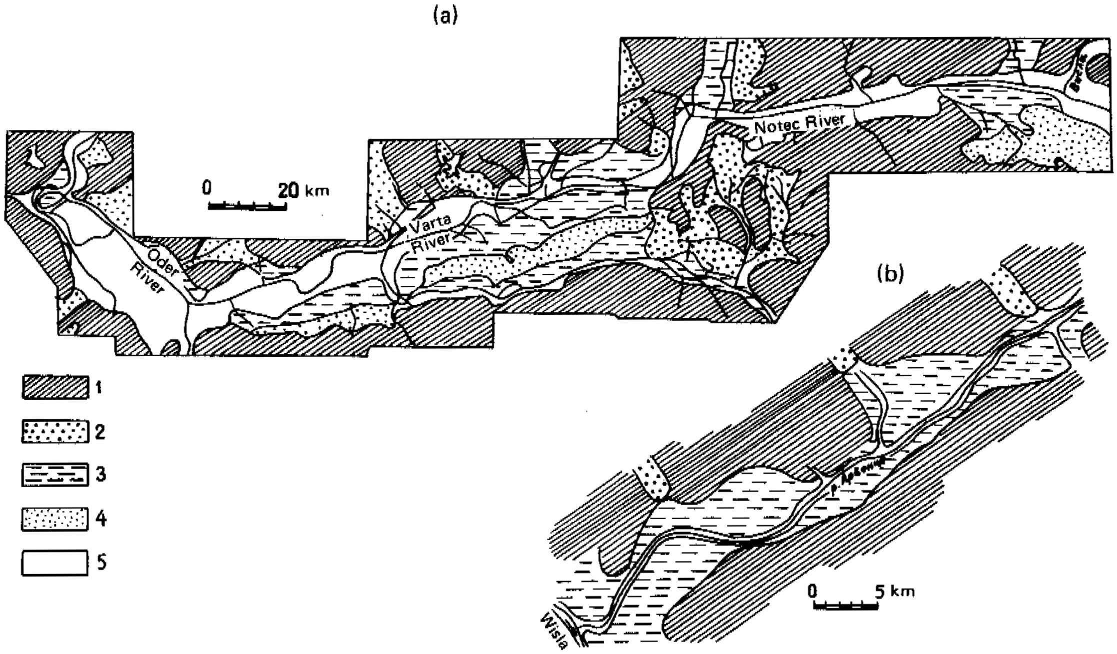

FIGURE 1 Geomorphological systems. (a) Notec-Warta ice marginal drainage channels[7] and (b) foredelta parts of Dreveca River Valley at the point of its emptying into the Torun basin:[13] 1--moraine plateau and terminal moraines; 2--old outwashes; 3--terraces of ice marginal drainage channels; 4--dunes on terraces of ice marginal drainage channels; 5--contemporary river terraces. In the systems, we distinctly note the widenings of basins and narrowings in the valley bottoms.

In plan, the ice marginal drainage channels have a quite typical form: Broadened sectors alternate with narrow ones (Figure 1a). The first have a length of several tens of kilometers and width up to 15-20 km. The narrow lengths, with widths up to 3-4 km are usually small, i.e. 10-15 km. The differences in the composition of soils of an ice marginal drainage channel in the narrow and broad sectors are not established; traces of most recent movements have not been registered.

The ice marginal drainage channels have distinct terrace formations of varying numbers of levels. For the Notec-Warta ice marginal drainage channel, R. Galov[7] identifies five levels pertaining to the upper Pleistocene and having formed under the influence of a glacier. The terraces are narrow at the contractions of the ice marginal drainage channels and quite wide in the expanded lengths. A large part of the terraces in this ice marginal drainage channel comprises chiefly levels of denudation covered by a shallow (usually up to 2-3 m) coarsely grained alluvium. In places on the terraces, there are indented sections made of alluvium with a thickness of several meters. Their position indicates the division of channels; traces of meandering are lacking. In the Notec-Warta ice marginal drainage channel, only the floodplain and low Holocene terraces are accumulative. In the other ice marginal drainage channels, the terraces that were formed during the warm and moderately cold periods of the Pleistocene are also of the accumulative type.

The valleys of a number of rivers having formed in the forefields of the glacier and having emptied into the ice marginal drainage channels have a form similar to the ice marginal drainage channels; the same can be said of certain valleys located outside the limits of direct influence of the ice sheet. An example of the first is provided by the Dreveca River Valley;[13] the form of its valley is indicated in Figure 1b. Among the latter, we include some of the river valleys in the Lodz upland.

Quite distinct and typical are the beds of periglacial alluvium corresponding most often to the stratigraphically coldest segments of the Würm epoch. They are represented chiefly by well-graded sands and coarse-grained sands with gravel and pebbles, often well-eroded. The suglinoks, supesses, and silty materials in these beds are either lacking entirely or are included in the form of thin lenses and interstratifications. The deposits of floodplain facies are lacking and only in rare instances do they have a markedly inferior development in the periglacial alluvium. Judging by the nature and composition of the bedding, the coarse-grained material found here constitutes riverbed alluvium, to be sure fairly unusual. Thus, in the alluvium of small periglacial valleys and ice marginal drainage channels, particularly on the 40-m terrace of the Notec-Warta ice marginal drainage channel,[12] we detect traces of frost-splitting. The wedge polygon structures often have traces of syngenesis with the surrounding beds. This is atypical for the riverbed formations in the contemporary permafrost area, with the exception of the alluvium in the naled fields.

Also in the periglacial alluvium, quite often we succeed in observing traces of settling and compaction of deposits established in the form of a system of small faults outlining the settling sectors. Specifically, indications of such subsidences have been described by S. Kozarski[12] for the alluvium in a 5-6-m terrace of the Golitse-Staraya Rudnitsa periglacial valley, situated in the peripheral zone of the Kostzhin basin (lower Oder). Just here on the terrace, as in many other places, we find a wide distribution of drainless sinkholes with a diameter up to several tens of meters and depth to 2-3 m. Indeed this author explains the detected subsidences by the thawing of dead glacial ice. Let us point out that the conditions for the burial of glacial ice on the river terraces are extremely unfavorable. There is little likelihood that the dead ice could have been preserved beneath the river channels during the formation of valleys and thawed only during the period of degradation of the frozen layers. Most likely the thawed ice is from the hydrolaccoliths and the structure-forming ice in the alluvium. It became formed during the freezing of the latter after the shifting of the channels to one side. In the sections we can observe that in places the periglacial alluvium loses its stratification; in this connection, over a distance of several meters there sometimes is disrupted not one but several superimposed patches, which indicates the high extent of ice saturation in the past.

All the above-indicated features in the structure of periglacial valleys and alluvium in central Europe are quite typical for the contemporary river valleys, in which active naled formation is occurring under fairly rigorous permafrost conditions. Under present conditions, naled formation is extensively known in the mountainous regions of Siberia having a severe distinctly continental climate. The conditions and causes for the origination and development of naleds, the main features of their geologic activity, and influence on the relief, composition, and structure of the deposits in their beds are known from the reports prepared by M. I. Sumgin, N. I. Tolstikhin, V. G. Petrov, P. F. Shvetsov, A. I. Kalabin, A. S. Simakov, O. N. Tolstikhin, and others. They have also been studied over a period of many years by one of the authors of the present report.

For the river valleys where naleds occur, we typically find the predominance of lateral erosion over bottom erosion and the formation of widenings of the valleys in the naled formation sectors; the erosion of deposits in the floodplain facies and surface outcrops of sand-gravel-pebble riverbed deposits; the division of channels into sectors of naled fields and the absence of their meandering; and the division of a single river into braided streams with varying speeds and directions, which provides the opportunity both for the scouring of deposits and in places the accumulation of pockets made of fine material. In the sectors of naled fields where frozen soils have developed, there occurs frost-splitting and formation of ice-wedge polygons,[5] and the development of structural forms, etc. During the formation of naleds, there often originate naled hummocks, in the body of which alluvium lenses are enclosed. Under the harsh freezing conditions, beneath and within the body of naled lenses of ice-saturated alluvium develop with a basal cryostructure and also hydrolaccoliths. The subsequent thawing of the hydrolaccoliths quite often leads to the formation of closed sinkholes at their site. The thawing of pockets of ice-rich alluvium is accompanied by its compaction, with a disruption of the original bedding of the layers, a partial loss of stratification, the origination of systems of cracks, etc. As was indicated above, all this occurs in the periglacial alluvial deposits as well.

The traces of the former naleds, the existence of hydrolaccoliths and of ice-rich "swollen" alluvium, i.e., the consequences from the activity of subsurface water under the conditions of severe climate and permafrost required an analysis of the natural conditions of the glacial period. In this context, we used the reports prepared by Ya. Dylik,[2] A. Yan,[9] S. Kozarskiy,[12] A. A. Velichko,[1] and others concerning the natural conditions of the glacial zone of Würm glaciation and also the results of the authors' personal observations.

During the periods of glaciation in central Europe, the runoff of surface water and groundwater was accomplished from the south, from the side of the mountains, and from the north, from the side of the ice sheet. Ahead of the glacier front these flows met and turned west. The subsequent shifting of the glacier front during regression of glaciation led in the Würm epoch, to the formation of three "northern" ice marginal drainage channels arranged in a subparallel direction. Considering as a whole the severe distinctly continental and dry climate during the maximal development and degradation of Würm glaciation, it should be considered that the surface runoff in winter in the periglacial zone was reduced significantly or was entirely lacking. Therefore, it might be thought that the formation of naleds was mainly owing to the alluvial waters from the ground-infiltration taliks under the river channel.

The runoff from the side of the glacier sheet was achieved by the thawing of glacial ice, both from the surface and from beneath, under the influence of geothermal heat. Most of the researchers consider that the temperature at the foot of the glacier was 0°C. An exception was at the marginal zone of the glacier, where beneath the ice, which was thin, frozen layers could have developed. If the thawing of ice from the surface had a distinctly seasonal nature and a considerable share of water with high velocities ran off during the summer, the bottom thawing was constant and uniform throughout the year. Waters having originated during bottom thawing, during the winter were able to supply the underchannel flow and be utilized in the subsequent formation of naleds. During surface thawing of the glacial ice, some of the water moved along the fissures to reach the bed of the glacier, and flowed with relatively low velocities. Their runoff extended through time included part of the winter. Obviously this water also participated in feeding the naleds.

During the maximum of Würm glaciation and in its second half, in front of the glaciers there occurred an intensive eolian activity[2] both during the summer and winter seasons. The latter condition is evidenced by the large number of polygonal systems containing sand wedges that became formed in winter on intensive deflation of material and the flowing of sand and gravel into the open frost cracks as if into traps.[6,8] The presence of sand wedges indirectly indicates the intensive blowing about of snow, which should have led to extremely irregular, in places very deep, seasonal freezing of taliks in the valleys. Undoubtedly the latter circumstance promoted the outburst of groundwater. The water outcropping to the surface under severe winter conditions froze, thus forming naleds.

The distribution of frozen layers in the middle and end of the Würm epoch was universal. This is evidenced by the extensive development of pseudomorphs along the multiple-ice-wedges, even in the sand-gravel fluvioglacial and alluvial beds.[12] The formation of wedge ice in the latter could have occurred at soil temperatures not higher than -5°C to -7°C. Under such conditions, the taliks could have been associated only with the water streams. Obviously these were chiefly underchannel and floodplain ground-infiltration taliks. The wide distribution of extensively water-permeable sand-gravel-pebble alluvial and fluvioglacial beds created favorable conditions for the development of such taliks. Possibly certain taliks were pressure-infiltration type, and through them the discharging of water from deep circulation was accomplished.

In central Europe, the position of ice marginal drainage channels that had extended along the front of glaciers at a slight distance from them created fairly unique and severe conditions, which favored naled formation. At the same time in the eastern sector of Europe, where runoff from the glacier was chiefly toward the south, the conditions were obviously less favorable for the extensive development of naleds.

Evidently the naled formation in the valleys of small and medium-sized rivers also had other causes. While in the valleys cutting through the outwashes and extending from the side of the glacier, the water sources were the same as in the ice marginal drainage channels, in the areas not situated within the region of direct influence of thawed glacial water, their sources could have been the suprapermafrost water in the masses of fluvioglacial sands. Here the water accumulation could have been accomplished by the deep seasonal thawing of sand beds and the formation of "nonmerging permafrost" and a system of ground-infiltration blind taliks. These taliks were drained by river valleys in which naled formation also proceeded. A similar pattern is observed in central Yakutia on the Bestyakhskaya terrace along the Lena River[3] and on the fringe of a mass of outwash sands in Charskaya basin.[6] Under such conditions, the intensity of naled formation was probably appreciably more in the relatively warm and longer periods of the Würm epoch. On the contrary, the maximal development of naleds, fed from the water that originated during the bottom thawing of the ice sheet, took place during the colder periods.

In the ice marginal drainage channels, the large areas of the terraces, which because of their structure could be regarded as naled fields, we are compelled to assume the migration of naleds, i.e., of a systematic variation in their position, forms, and dimensions. As a result of migration appreciably larger areas were subjected to the influence than the dimensions of the actual naleds that developed every year. Obviously the migration was associated with changes in the severity of climate and frost conditions, position of springs feeding the water to taliks in the river valleys, volume of this water, and other causes.

An analysis of the available material makes necessary the formulation of the question concerning the overall variations in the hydrogeologic conditions during the glaciation periods over wide areas of Europe representing chiefly artesian basins of an open type. The glaciers advancing from the north covered the marine basins, at the present time serving as a basis for the runoff of surface and subsurface water. Evidently they made quite difficult and at times impossible the discharge of artesian water. Beneath the ice sheets of the "warm type," owing to heat within the ground, the thawing of ice constituted in conversion to water $3\text{-}6 \cdot 10^3 m^3/yr \cdot km^2$. Often these waters existed under high pressure[4] which predetermined the possibility of their infiltration into the soils. As a result, certain sectors involved in the discharge of subsurface water could be transformed temporarily to areas of supply. The extent of flooding of a zone under the effect of a glacier increased significantly, intensifying the circulation of subsurface water; there was a change in its composition, as compared with the conditions having existed in the interglacial period. Moreover, there probably was a change in the entire nature of the subsurface water runoff. The development of frozen layers in the periglacial zone led to the localization of the discharge points and to the supply of water from deep circulation; both aspects were in effect within sectors of taliks (limited in area) gravitating toward the river valleys, toward the tectonic disruptions, or toward the areas of very permeable soils and outcropping to the surface.

Moreover, there could have been an increase in the intensity of water movement below the frozen layers; the thickness of the active circulation zone increased. Favorable conditions for this built up primarily in the regions of carbonate development where the processes of leaching intensified and the deep-seated nature of karst formation grew. An example of such regions can be provided by the Yura-Pol'skaya Territory (Krakow-Chenstokhovskaya) not having been covered by glaciers,[14] where karst is intensively developed. The valleys of many small rivers have broad flat bottoms, often being traced as far as the headwaters where they end "bluntly." Here the karst forms are usually developed. An analogy with the regions of contemporary development of karst in the area of persistently frozen soils furnishes a basis for theorizing that such forms of valleys comprise the result of a former development of naleds having existed during the cold periods in the Pleistocene and having been fed from the karst-type springs.

Based on what has been said, we can assert that in the glaciation periods in central Europe there transpired a significant modification in the permafrost-hydrogeologic conditions, which doubtless included:

1. A localization of groundwater in the ground-infiltration taliks of river valleys and obviously on the outwashed.
2. The formation of naleds of hydrolaccoliths and of ice-rich "swollen" alluvium in the river valleys chiefly owing to the water flowing from the subchannel ground-infiltration taliks.
3. The active geologic and geomorphological role of naleds that existed in the river valleys, especially in the sphere of influence of high summer floods, formed on the thawing of glacial ice.

Naled formation was obviously the most typical expression of the activity of subsurface water in the periglacial zone of the continental glaciations in Europe, linked with the appreciable reconstruction of hydrogeologic conditions under the effect of a glacier, layers of frozen soil, and a severe climate. The changes in the intensity in manifestation of the naled-formation processes at various stages of glaciations reflect the alternation of permafrost-hydrogeologic and paleogeographic conditions and will be the topic of a study for the immediate future.

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PRINCIPLES OF HYDRODYNAMIC CALCULATIONS OF SYSTEMS FOR EXTRACTING DEEP-SEATED HEAT FROM THE EARTH IN PERMAFROST REGIONS

V. M. MAKSIMOV *Leningrad Plekhanov Mining Institute*

In the permafrost regions where the permafrost is widely developed and severe climatic conditions occur, the hot and warm waters have great importance for the development of placer deposits, heating of buildings, and construction of hothouse facilities. In these regions, hot water can be obtained in various ways: from a ground-based boiler installation, stationary and mobile atomic power plants, by utilizing geothermal water, and also from the deep-seated heat from the earth using the so-called thermal-circulating systems consisting of injection and riser wells, the bottoms of which are located in an "underground boiler" that is formed by a powerful explosion in stratified or massive rocks.

Owing to the convective and conductive heat exchange, and to a considerably smaller extent to the throttling effect, the cold water that is being pumped from a relatively great depth becomes warmed and, already warm, by means of riser wells it reaches the surface, where it can be used for the development of placer deposits and for other purposes.

It was established by investigations conducted by the Leningrad Mining Institute, under the supervision of Yu. D. Dya'kin of the Ukrainian SSR Academy of Sciences, and by other scientific research institutes that the obtaining of warm water with the aid of thermal-circulatory systems is, from a thermophysical standpoint, the most economically advantageous and effective method among various techniques. The selection and justification of a type of system is largely determined by the geologic, permafrost-hydrogeologic, and geothermal conditions in the region of its location. From a geologic standpoint, in the region of constructing the system, there should not be deep tectonic faults and dislocations having a regional distribution; it is necessary to have thick seams and layers, or masses of materials suitable for underground (camouflet) explosions.

The permafrost-hydrogeologic and geothermal conditions in a region exert direct influence on the selection of the permanent water supply to a thermal system (surface or subsurface supply), depth of injection and riser wells, their design, technique of drilling and testing, and also on the method of hydrodynamic calculations. Therefore, the latter are considered separately by us for a system constructed in stratified and massive three-dimensional rocks, wherein the three-dimensional type problems are reduced to two-dimensional problems, i.e., they are solved in the planes *XY* and *XZ*. In a study of the movement of liquids in fissured rocks, we differentiate:[1]

1. The purely fissured rocks in which the primary permeability of individual blocks is quite low and the principal routes for the movement of liquids are the cracks (we can disregard the porosity of the blocks).
2. Fissured rocks in which secondary permeability in the blocks is very slight (it can be disregarded); flow is essentially through the rock pores.
3. Rocks with dual permeability, where the primary and secondary permeabilities are comparable with each other; movement of liquids occurs both through the pores and the cracks.

It is not difficult to see that, to the working conditions of the thermal-circulating systems, there correspond chiefly the first two types of fissured rocks, forming layers or masses. Disregarding primary permeability of the blocks of fissured rocks, which is quite permissible for deep-seated rocks, and considering that the fissuring of strata or masses is uniform (prior to the explosion) and the crushed-fissured medium as isotropic or anisotropic from a permeability standpoint (after the explosion), the movement of water in such media can be considered analogous to its movement in the porous rocks. In a solution to the flow problems related to the operation of thermal-circulating systems, this makes it possible to employ the equations for flow of water in a porous environment.

CALCULATIONS FOR A SYSTEM IN A ROCK STRATUM

Under natural conditions of the bedding of a water-bearing layer, the movement of water in the region of injection and riser wells is determined by Equation (1), which characterizes the superposition of flow fields forming during pumping of water into a stratum and during the simultaneous removal of water from it (Figure 1):

$$\Delta P = \Delta P_1 + \Delta P_2 = \left(\frac{Q_1}{2\pi KM} \ln\frac{R_1}{r_1} - \ln\frac{R_2}{r_2}\right), \qquad (1)$$

where ΔP = pressure drop at point *N* on superposition of fields, ΔP_1 = excess pressure at mouth of injection well, ΔP_2 = depression during removal of water from riser well, Q_1 = discharge from injection well (equals outflow from the riser well), R_1 and R_2 = radii of influence of injection and riser wells, r_1 and r_2 = distance of point *N* from injection and riser wells, *K* = permeability of layer, and *M* = thickness of layer. Solving Equation (1), at $R_1 = R_2 = R$ = radius of influence

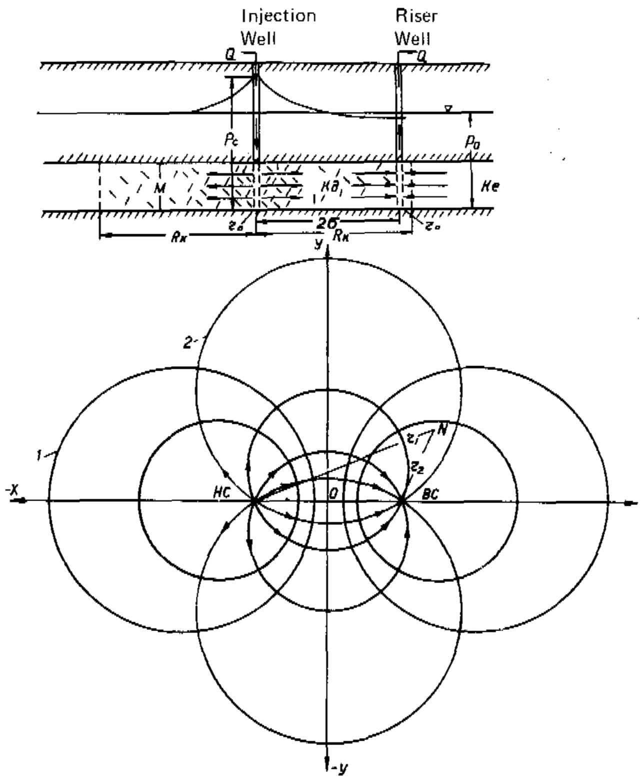

FIGURE 1 Diagram showning pumping of water into layer disrupted by camouflet explosion 1--equipotential lines; 2--flow lines. (HC--injection well; BC--discharge well).

of riser well on superposition of fields and $r_i = r_0$, $r_2 = 2_\pi$ with allowance for the factors ξ and μ, we get

$$Q_1 = \frac{2\pi KM\Delta P}{\xi\mu\ln(2_\pi/r_0)}, \qquad (2)$$

where ξ = coefficient of skin effect of the riser well (it is established experimentally: in the approximate calculations, based on experience in industrial petroleum practice, it can be assumed to equal 1.5-2.0); μ = coefficient of water viscosity (based on temperature in the layer at the point of development of the "underground boiler"); 2_π = distance between injection and riser wells; r_0 = radius of each well.

Equation (2) is applied for calculating the discharge rate from the riser well in an unlimited layer not disturbed by an explosion under conditions of steady laminar flow of the water that is being pumped. In case of a powerful camouflet explosion in a seam formed of solid rock, three "spheres" are formed: crushing, fissure formation, and elastic deformation. In the sphere of crushing and fissure formation, the rocks acquire different permeabilities. As a result of an explosion, the permeability increases appreciably near the center of the explosion and [inversely] in proportion to distance from it, gradually decreasing to the permeability [K_e] of the layer under natural conditions. Assuming that the long axis of the "sphere" of crushing and fissure formation (height of "underground boiler") equals the thickness of the layer and locating the water-lift well near the boiler's outer surface, the output from which based on Equation (2) can be computed according to the equation:

$$Q_1 = \frac{2\pi K_E M\Delta P}{\xi\mu\ln(2\sigma/r_0)}, \qquad (3)$$

where K_E = permeability of rock in the zone of influence of the explosion (determined from data from test pumpings).

Equation (3) is applicable for the "sphere" of crushing and fissure-formation of rocks having the shape of a cylinder and equalling in height the thickness of the layer, at uniform permeability of the rock. In case of nonuniform permeability of the rock in the "underground boiler," the form of which comprises an ellipsoid (Figure 2), the original equation of motion of the pumped water in plane X has the form:

$$K_1 \frac{\partial^2 P}{\partial X^2} + K_2 \frac{\partial^2 P}{\partial Z^2} = 0, \qquad (4)$$

where K_1 and K_2 = permeabilities of the crushed-fissured rocks in the direction of the X and Z axes. Solving Equation (4) through a conversion of coordinates and taking the boundary conditions into account, we get an equation for finding the outflow from the injection well at its location indicated in Figure 2, and the rate of pumping water into the dry crushed-fissured rock:

$$Q_1 = \frac{2\pi l\sqrt{K_1K_2}\cdot P_c}{\xi\mu\ln(R_K/r_0)}, \qquad (5)$$

where l = length of well screen; P_c = pressure at face of pressure well; R_K = radius of "underground boiler" ($R_K = C_p = C_p\cdot\sqrt[3]{W}$, where C_p = coefficient depending on the nature of the rock forming the layer, its density, fissured state, and lithologic-petrographic composition; and W = weight of explosive, kilotons).

In case of the pumping of water into the

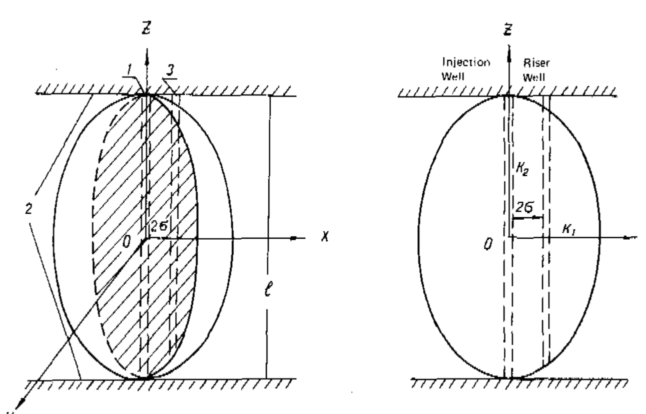

FIGURE 2 System for pumping water into "underground boiler" of ellipsoidal form. 1--injection well; 2--upper and lower impervious strata; and 3--riser well.

crushed-fissured water-bearing rock, Equation (5) will be written in the form

$$Q_1 = \frac{2\pi l\sqrt{K_1K_2}\Delta P}{\xi\mu\ln(R_K/r_0)}, \quad (6)$$

where $\Delta P = P_C - P$ where P_0 = initial stratal pressure assumed to equal the hydrostatic pressure ($P_0 = \gamma H/10$). The outflow rate from the riser well situated at distance 2σ from the injection well will be:

$$Q_1 = \frac{2\pi l\sqrt{K_1K_2}\Delta P}{\xi\mu\ln(2\sigma/r_0)}. \quad (7)$$

If we have several risers, the outflow from each of them will be:

$$Q_1 = \frac{2\pi l\sqrt{K_1K_2}\Delta P}{n\xi\mu\ln(2\sigma/r_0)}, \quad (8)$$

where n = number of these wells in the system.

It should be pointed out that the hydrodynamic calculations of a system with consideration of the anisotropy of crushed-fissured rocks reflect more reliably the actual conditions involved in the flow of pumped water in the "underground boilers," created by powerful explosions in the Earth's interior, than calculations based on an average permeability of crushed-fissured rocks entering the equations obtained by the method of superposition of flow fields. In the last case, the discharges of the injection and riser wells, as calculations of actual systems demonstrate, at identical original parameters, turn out to be exaggerated. Assuming in Equations (3) and (7) that $2\sigma = P_K$ and $\Delta P = 1$, we derive the formulas for the calculation of the coefficient of the receiving capability of the injection well:

$$q = \frac{2\pi K_E M}{\xi\mu\ln(R_K/r_0)}; \quad (9)$$

$$q = \frac{2\pi l\sqrt{K_1K_2}}{\xi\mu\ln(R_K/r_0)}. \quad (10)$$

The time for the progress of water in the pressure well to the waterlift well is computed under conditions of anisotropic permeability from the formula

$$T = \frac{4}{3}\,\frac{\pi M m\sigma^2}{\xi\mu Q}. \quad (11)$$

The total flow area of the pumped water in this case is found from the formula:

$$F = \frac{4}{3}\,\frac{\pi\sigma^2}{\xi\mu}. \quad (11a)$$

The coefficient for the fissured rocks in the zone of influence of an underground explosion entering Equation (11) can be determined approximately from the formula developed by F. I. Kotyakhov:[5]

$$m = 1.73\cdot 10^{-3}\sqrt{q_1\mu \lg(R_K/r_0)S^2}, \quad (12)$$

where q_1 = reception factor of the injection well, m^3/day · atm; μ = viscosity of water, centipoises; S = coefficient of fissures' thicknesses, cm^{-1} (is determined from photographic logging).

CALCULATIONS OF A SYSTEM IN MASSIVE ROCK

For a system constructed in massive crystalline rocks consisting of injection and riser wells, we have developed approximate hydrodynamic calculations of this system for the arrangement of the injection well in the central part of an "underground boiler." Such an arrangement assures the smallest losses of water into the outer area of the rock mass during the system's operation. During water pumping into the rock mass shattered by an explosion, the original equations relating the flow and the underground boiler of cylindrical form with two hemispheres (Figure 3) have the form:

$$Q_1 = -K(4\pi r^2 + 2\pi r l)\,\frac{dP}{dr}, \quad (13)$$

$$Q_1 = m(4\pi r^2 + 2\pi r l)\,\frac{dr}{dt}. \quad (14)$$

Solving Equations (13) and (14) under the assumptions that--the surface of the hemispheres and the cylinder's lateral surface constitute sur-

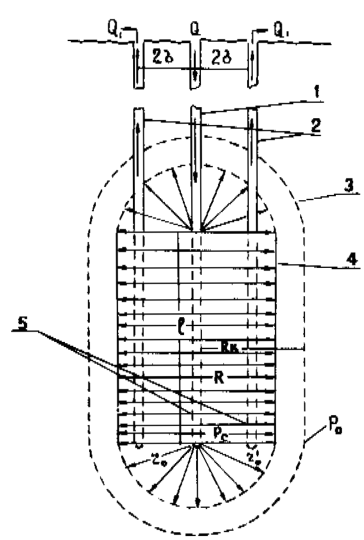

FIGURE 3 Diagram indicating pumping of water into massive rock. 1--injection well; 2--riser well; 3--"underground boiler" having the shape of a cylinder with two hemisphers; 4--diagram indicating flow in surface; and 5--screens in wells.

faces of equal pressure; the extent of saturation in the underground boiler during pumping increases gradually; flow from the well occurs along radial lines; the permeability in the boiler is uniform and constant; pressure at the boiler's external boundary is equal to initial stratal pressure; and elasticity of gases forming during explosion, as well as gravitational forces and water inertia, are disregarded--we derive equations for determining the discharge rate from the injection and riser wells.[3,4]

The yield* at installation of a screen with length in the boiler

$$Q_1{**} = \frac{2\pi K_E l(P_c - P_0)}{\xi\mu \ln\frac{R_K(2r_0 + l)}{r_0(2R_K + l)}}, \qquad (15)$$

$$Q_1{*} = \frac{2\pi K_E l(P_c - P_0)}{\xi\mu \ln\frac{R_K(r_0 + l)}{r_0(R_K + l)}}. \qquad (16)$$

The capacity of the injection well will accordingly be:

$$q_1 = \frac{2\pi K_E l}{\xi\mu \ln\frac{R_K(2r_0 + l)}{r_0(R_K + l)}}, \qquad (17)$$

$$q_1 = \frac{2\pi K_E l}{\xi\mu \ln\frac{R_K(r_0 + l)}{r_0(R_K + l)}}. \qquad (18)$$

In a system consisting of an injection well and of several risers, the yield from each riser well will be determined from the equation

$$Q_i = Q/n. \qquad (19)$$

Equations (15), (16), and (19) are valid under the following conditions:

1. Uniform permeability of rock in the boiler --in all directions from the injection well.
2. Steady flow.
3. Practical impermeability of external surface of boiler (absence, in the rock mass, of regional tectonic fissures reaching the boiler).
4. Constancy of skin effect in the influence zone of injection well.
5. Standard position of screens in the wells.

The thermal circulation system can operate at prescribed output (Q_1 = const) or constant pumping pressure (P_H = const).

In connection with this, the radius of influence of the injection well at any time of pumping will be determined from various equations obtained from Equations (13) and (14). For finding R at Q_1 = const from Equation (14), we get

$$R{**} = \sqrt{\frac{Q_1 t}{\xi\mu(l + 1.33R)\pi m}}, \qquad (20)$$

$$R{*} = \sqrt{\frac{Q_1 t}{\xi\mu(l + 0.66R)\pi m}}. \qquad (21)$$

For finding R at P_H = const, from a combined solution to Equations (13) and (14), we derive:

$$R{**} = \sqrt{\frac{K_E l(P_c - P_0)t}{\xi\mu\left(\frac{l}{2} + \frac{2}{3}R\right)m\ln\frac{R(2r_0 + l)}{r_0(2R + l)}}}, \qquad (22)$$

$$R{*} = \sqrt{\frac{K_E l(P_c - P_0)t}{\xi\mu\left(\frac{l}{2} + \frac{R}{3}\right)m\ln\frac{R(r_0 + l)}{r_0(R + l)}}}. \qquad (23)$$

Equations (20)-(23) are solved graphically (Figure 4a, b). From the curves it is obvious that at Q_1 = const the radius of influence of the pressure well increases with an increase in Q_1 and decrease in μ; at P_H = const, R_1 depends to a considerable extent on K_e and to a lesser degree on pressure P_H. In both instances, the R_1 value is influenced greatly by the duration of the pumping process. The duration of the cycle involved in replenishment of water in the boiler at Q_1 = const will be:

$$t = \frac{\pi R^2_K m(l + 1.33R_K)}{\xi\mu Q_1}. \qquad (24)$$

OPERATING REGIME OF SYSTEMS AND ESTIMATION OF WATER LEAKAGE

The operating regime of systems is established by the ratio of pressures P_c and P_0, in relation to which, three regimes can exist:

Regime I, when $P_c = P_0$. For the given regime $Q_1 = Q_2$, where Q_2 = yield from the water-discharging well (excluding the effect of the external area which supplies the seam or mass).

Regime II, when $P_c > P_0$. Under this condition some of the water that is being pumped can pass into the external area of the stratum or massive rock; for given regime, $Q_1 > Q_2$.

Regime III, when $P_c < P_0$. This condition can develop only during the cessation of water supply into the pressure well, which causes the supply

*In Equations (15), (16), and (20)-(23), one asterisk (*) indicates a screen at one hemisphere, while two such symbols (**) stand for a screen at each of two hemispheres. Such screens are provided in all the wells.

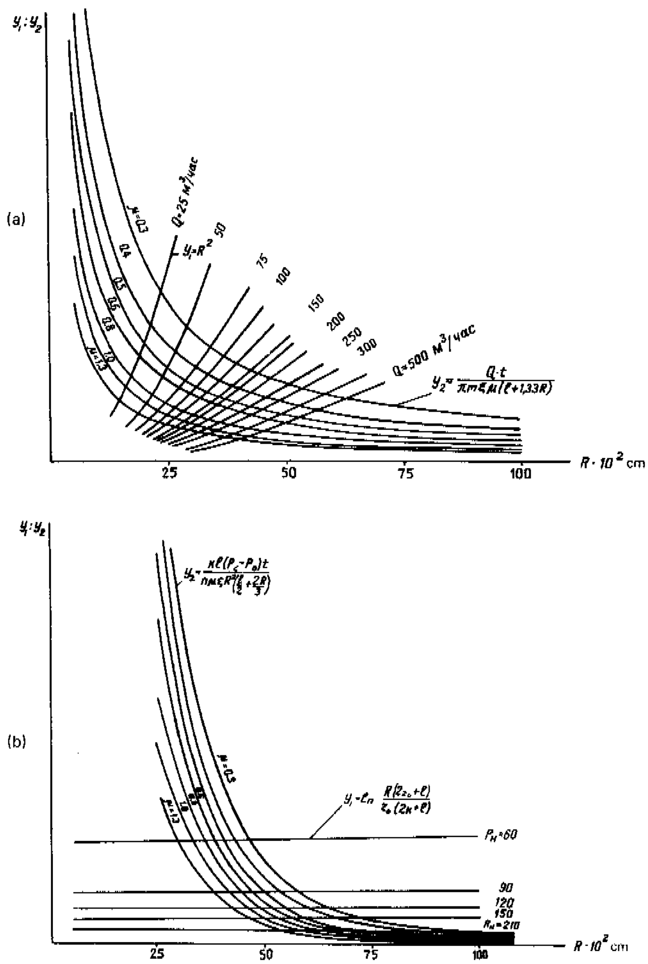

FIGURE 4 Curves for determining the radius of influence of the pressure-injection well. (a) curve $R = f(Q, \mu)$ at $\xi = 1.5$; $t = 10$ days; $l = 1$ m; m = 0.25. (b) curve $R = f(P_H, \mu)$ at $K_E = 5$ darcys; ξ 1.5; $t = 10$ days; $l = 1$ m; and m = 0.25.

of water to the water-lift well from the outer area of the layer or rock mass. Under this regime $Q_1 < Q_2$.

In this manner the leakage of water into the outer area of a stratum or rock mass can occur only under the second regime of the system. In this instance the amount of water leakage (Q_y) in a system constructed in a layer is determined from the equations

$$Q_y = \frac{2\pi K_e M \Delta P}{\xi \mu \ln (R_0/2\sigma)}, \tag{25}$$

$$Q_y = \frac{2\pi K_e M \Delta P}{\xi \mu \ln (R_t/r_0)} \tag{26}$$

where R_0 = radius of area involved in feeding the layer (it is found from the results of a geologic and hydrogeologic survey of the region involved in the construction of the system); $R_t = 1.5\sqrt{at}$, where a = permeability of the layer and t = duration of system's operation for the time span under consideration. By analogy with the stratum system, the amount of water leakage can be computed on the basis of the equation

$$Q_y = \frac{2\pi K_e l \Delta P}{\xi \mu \ln \dfrac{R_t(4\sigma + 1)}{2\sigma(2R_t + 1)}}. \tag{27}$$

Since K_e value at great depths is extremely small, usually equalling 10^{-3}-10^{-5} darcys, we see that water leakage into the outer area of the layer or rock mass comprises tenths or units of percents of the yield from the injection well.

The Hydrodynamic calculations form the basis for the hydraulic estimates of the systems, including a determination of optimal pressure at the top of the wells, surface pressure in the injection well, and losses of pressure in the wells. These calculations can be conducted on the familiar equations from hydraulics, broadly utilized in petroleum industrial and drilling practice.

In the completion of the hydrodynamic calculations of a system, we find the transformation coefficient of heat energy based on the familiar equation[4]

$$\alpha_t = \frac{H_2}{H_1} \tag{28}$$

where H_2 = amount of heat energy obtained, kcal/day [$H_2 = CQ_2(T_1 - T_2)$]; H_1 = amount of thermal energy expanded, kcal/day ($H_1 = Q_1 H/102\eta$); P = pressure developed by pump, m; η = efficiency of pumping installation; c = heat capacity of water, kcal/kg-degrees (for water c = 1); T_1 = water temperature in underground boiler, °C; T_2 = water temperature at top of riser well, °C. High values for the coefficient α_t indicate the thermophysical and economic effectiveness of the thermal-circulation setup.

In conclusion let us point out that the thermal-circulating systems will operate during a fairly extended time period (10-20 yr). Therefore, the uninterrupted supply of water to these systems in the necessary amount and proper quality is an important problem in solving the question of utilizing the deep-seated heat from earth. This problem includes two interrelated questions:

1. Selection of a reliable source of water supply, completely guaranteeing the uninterrupted operation of the thermal-circulating system in respect to water supplies.
2. Determination of the necessary quantity and quality of water for supplying the thermal-circulation setup.

For water supply to the thermal-circulation systems, we can utilize fresh surface and subsurface water. Among the latter, under the conditions of the northern regions in the USSR, of practical significance are the intrapermafrost water in the through taliks and river valleys and the subpermafrost water in the artesian basins, mountain-folded regions, and hydrogeologic masses. Naturally, the explorations and testing of underground water should be preceded by special permafrost-hydrogeologic investigations, terminated by an evaluation of the reserves and quality of water with application to the purposes of water supply in the thermal-circulating systems.

For the normal conditions in the operation of thermal-circulating systems, special attention should be directed to the quality of the water that is being pumped. Based on the experience gained in the development and investigation of pressure wells in the petroleum industry, the water for pumping must not:

1. Reduce the permeability of rocks in the zone of crushing and fissure formation.
2. Cause corrosion to the drilling pipes or the pumping equipment.
3. Deposit difficultly-soluble salts (for example, gypsum $CaSO_4.2H_2O$) in the fissures of the zones of crushing or in the rock pores.
4. Deposit calcium carbonate based on reaction $Ca(HCO_3)_2 \rightleftarrows CaCO_3 + H_2O + CO_2$.
5. Contain a large amount of iron and oxygen, since in the cracks and pores of rocks in the depths, oxidizing-reduction processes with the formation of $Fe(OH)_2$ and $Fe(OH)_3$ can occur.
6. Contain a large amount of suspended particles that could plug up the cracks.

CONCLUSIONS

1. The selection of a system of hydrodynamic calculations for the setup for extraction of deep-seated heat from the earth is determined largely by the geologic structure and permafrost-hydrogeologic conditions existing in the regions of their [proposed] utilization. Based on the geologic factor, the hydrodynamic calculations are divided into calculations of systems in layers and masses of rocks; moreover, as a result of the assumptions made, they are approximate.
2. In the calculations of systems in the layers for a determination of the yield of injection and riser wells, the capacity factor of the injection well, radius of pumping, and leakage of water, use can be made of Equations (3)-(12), (25), and (26); in a mass, Equations (15)-(18), (20)-(23), (24), and (27).
3. From Equations (5)-(8), (15), and (16), it is obvious that the productivity of the systems can be controlled by increasing or decreasing l 2σ, P_H, P_k [$R_k = f(W)$] at known values for K_e, μ, ξ, and S_0.
4. The equations reviewed have been obtained for stable conditions; the analysis of the functioning of unstable systems comprises a problem that the author is currently solving.
5. The hydrodynamic studies conducted for planning and constructing systems for extracting heat from earth in the regions of the USSR Northeast constitute the beginning of a new branch in the applied hydrogeology of the cryolithosphere.

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ZONATION AND REGIME OF NALEDS IN TRANS-BAIKAL REGION

I. M. OSOKIN Chitinskiy Pedagogical Institute

Based on the existing definitions of the naled concept[3,8,22,23] and considering the new factual information, the author refines this definition and suggests that the term "naleds" be applied to the ice bodies that have formed on the surface of ice, snow, and soil (rock) as a result of the freezing of upwelled surface and subsurface, melted snow and glacial water, and also the dumped industrial-public and other waters during the cold season of the year.

Based on the existing subdivisions of naleds by origin and area,[10,15,24,27,29] supplementing them and utilizing new criteria, the author suggests a composite classification of ground-based naleds.[14] In it, he adopts the dimensions of naled areas according to N. I. Tolstikhin[24] with the addition of the gigantic naleds, proposed by other authors. According to origin, six types of natural naleds are identified (river, ground, spring, glacial, from snow thawing in winter, mixed) and five types of artificial or man-made naleds (at dams on rivers, from the dumping of industrial and public sewage, from water overflowing from boreholes and wells, flooding of fields and meadows, heating of placer deposits). In frequency of occurrence of sizes for each of the naled types, we differentiate four degrees of probability. The classification suggested reflects more fully the diversity of naleds on the Earth's surface.

The geographic distribution of naleds in Trans-Baikal does not confirm the view of certain researchers to the effect that naleds are typical only of the regions where permafrost occurs. In the region under discussion, they occur universally, although the intensity of their development is dissimilar. Naleds occur most widely in the mountainous regions, where we find a frozen zone with taliks, tectonic faults, and a fissured condition of rocks (Stanovoye upland, Daurskoye arched uplift, etc.). Naleds are developed most weakly in the arid steppe and forest-steppe regions with seasonal freezing of soils, a sparse river network, and deep occurrence of groundwater. But the volume and percentage of man-made naleds increase here. The frequency of naleds in the river channels increases in the upper reaches with a decrease in the dimensions of rivers and of watershed areas and also in a direction from the steppes toward the mountain-taiga zone.

Distribution of naleds in Trans-Baikal is determined by the features of geologic development of individual parts of the region (tectonic faults, fissured condition, contacts of rocks, stream pattern), by the hydrogeologic structure, and by contemporary climatic conditions.

Therefore, for a description of the naleds in Trans-Baikal, we also adopted a system of zoning that takes all these factors into account. These requirements are satisfied most fully by the system of natural zoning of the Trans-Baikal region prepared by L. I. Mukhina, V. S. Preobrazhenskiy, and N. V. Fadeyeva.[17] According to conditions of formation, nature of occurrence and regime of naleds, and their dimensions and origin, in the region under study we identify seven regions: Stanovoye upland, Vitimskoye plateau, Vitimo-Olekminskoye medium-mountain and Tungiro-Olekminskoye low-mountain relief, Selenginskoye medium-mountain relief, Khentey-Chikoyskoye upland, Verkhneamurskoye medium-mountain relief, and Uldza-Toreyskaya plain. Let us point out certain specific features in the regions identified.

The Stanovoye upland is a region of widely occurring small, large, very large, and uniquely large naleds. We rarely find even gigantic naleds with an area of more than 10 km^2. There exist not only seasonal naleds, but also those that persist through the summer. Many of the naleds are located on slopes with northern exposure, in the high boundaries between the present snow line and the ends of glaciers from the epochs of maximal glaciation.[19] The relative extent of naled occurrence in the headwaters of certain rivers is more than 1 percent.

The Khentey-Chikoyskoye upland is the second region in the distribution of naleds in the Trans-Baikal area. The tectonic faults and extent of fissuring of rocks, the outcroppings of cold and hot mineral springs, the density of river network maximum for Trans-Baikal, and the deep erosional cutting of rivers cause favorable conditions for the extensive occurrence of naleds at this point.

In the Vitimskoye plateau, fewer naleds are developed than on the Stanovoye upland.

One of the features in the naleds on the Tungiro-Olekminskoye low-mountain relief is the relatively great development of naleds from groundwater, which are especially typical for the regions around the Amur River.

The Selenginskoye and Verkhneamurskoye areas of medium-mountain relief are characterized by great irregularity in the distribution of naleds and by an increased share of anthropogenic naleds.

The Uldza-Toreyskaya high plain has the least favorable hydrogeologic, geomorphological, and climatic conditions for the formation of natural naleds. Here, however, we often encounter anthropogenic naleds in the region of pastures, roads, and mining enterprises.

The absolute height of the belt of maximum development of naleds and of the upper boundary of their occurrence (see Table 1) in the mountains of the USSR Asiatic sector increases from the mountains in northeastern Siberia toward the mountains in southern Siberia and on toward the mountains in central Asia. In the same direction, there is an increase in the vertical thickness of the naled belt in the mountains. These relationships cannot be explained merely by the absolute height of mountains. In the mountains of northeastern Siberia, the upper limit of naled occurs at roughly half of their absolute height. The upper (Bald Mountain) belt having a vertical thickness up to a thousand and more meters does not have naleds, since the hydrogeologic conditions (the permafrost layer penetrates the entire zone of active fissuring of the Earth's crust) and the conditions for replenishing the subsurface water by precipitation are unfavorable. The local deviations caused by the features in hydrogeologic structure (northern slope of the Suntar-Khayata) do not alter the overall pattern of this regularity. This once again provides evidence that the climatic factor of naled formation in the scales of a vast region exerts an even greater influence on the regularity in the distribution and development of naleds than within the limits of a small region.

The relationship between the area and thickness of naleds in Trans-Baikal is not traced, but a link does exist between the area and size of naleds. For the very small, small, and medium naleds, this link is determined by the expression $W = 0.9\ S$ (at $r = 0.67$), while for the large and very large naleds, $W = 1.12\ S$ (at $r = 0.72$), where W = volume of naled, m^3, and S = area of naled, m^2.

For comparison, let us point out that for the large naleds in the USSR Northeast, O. N. Tolstikhin[26] obtained the form: $W = 1.52\ S$.

The aspects or stages in the life of naleds confined to specific time segments during the cold and warm seasons are divided quite tentatively and roughly. For the naleds in the eastern Trans-Baikal region, N. I. Tolstikhin and N. I. Obidin[25] have indicated four stages.

1. "Juvenile," i.e., October-December, when after the advent of the first heavy frosts, at the site of the future naled, the first thin and unstable ice develops. The naled gradually begins to grow in width, length and thickness.
2. "Youth stage," i.e., December-January. The naled grows rapidly; its outlines form and the naled hummocks mature.
3. "Stage of maturity," i.e., January-April. The naled has the maximum dimensions in length and width. Further growth in these directions almost stops. The naled hummocks are broken by fissures from which water flows. The growth of the naled in height continues. Thickness of ice increases. The naled hummocks are filled in from all sides with ice; as a result their relative excess in height above, the naled decreases. Sometimes the water in the hummocks' cracks freezes.
4. "Old stage" of a naled arrives with the coming of spring (April-July, sometimes March). The growth of the naled stops. There initially begins a gradual and then an ever-faster disappearance of the naled. Its surface is covered by depressions as a result of the uneven thawing of various parts of the naled. Exogenic cracks develop. Running water forms channels in the body of the naled. The hummocks settle, and funnels develop in their places. Disintegration of the naled proceeds both from the upper and from the lower surface, as well as from within along the channels and cracks. The naled gradu-

TABLE 1 Maximum Confinement of Naleds to the High-Level Belts of Mountainous Systems in the Asiatic Sector of the USSR

	Belt of Maximum Development of Naleds, Absolute Height, m	Absolute Height of Upper Limit of Naleds, m	Vertical Thickness of Naled Belt, m	Bibliography
Mountains in Eastern Yakutia				
Verkhoyanskiy Range				
Western slope	Below 500	1,100	800-1,000	
Eastern slope	600-1,000	1,200	800-1,000	
Southern Verkhoyansk region	500-11,500	1,450	1,000-1,200	13
Suntar-Khayata				
Northern slope	1,100-1,300	1,600	800	
Momskiy Range	650-1,000	1,200	800-1,200	
Cherskiy Range	500-1,000	--	800-1,200	
Mountains in Southern Yakutia				
	600-1,200	--	1,000	2
Mountains in Trans-Baikal Region				
Stanovoye upland	Up to 1,200	1,800	1,100	19
Eastern part of Stanovoye upland	Up to 1,250	1,640	1,000	11
Eastern Sayan				
Tunkinskiy and Pogranichnyy ranges	Above 1,200	--	1,200-1,500	30
Khentey-Chikoyskoye upland	--	2,100	1,400	28
Tyan'-Shan Mountains				
	2,500-4,000	4,000	1,500-2,000	6

ally breaks into separate blocks and finally thaws completely.

These stages should be regarded as classic examples of the development of a naled.

The authors of the naled's "life stages" have already observed that there are specific deviations in nature from the pattern suggested. These deviations concern primarily the periods involved in the appearance of naleds and various stages of their development. The appearance of naleds in the Trans-Baikal region extends from October to March. Even in March, as a result of deep seasonal freezing, springs sometimes appear that were not there in summer, fall, or during the first half of winter. However, in March, small naleds form from the water flowing from these springs. In the second half of winter, naleds even develop from the thawing of snow. Artificial naleds can appear at any time in winter in connection with man's economic activity.

The ablation of naleds in the Trans-Baikal region begins in March-April and even at the beginning of May. This depends on the height of terrain and natural zone in which the naleds are situated (see Table 2).

The amount of thawing of naleds in the Trans-Baikal region and other regions in the USSR Asiatic sector reckoned for 1° of increase in average diurnal air temperature is of the same magnitude as the amount of thawing of glacial ice and snow, but increases from northeast to southwest; this is associated with a decrease in cloudiness and an increase in the intensity of solar radiation in this direction. Within the limits of the same region, the differences in the amount of thawing of naleds are associated not only with the rise in diurnal air temperatures, but also with the density of ice, extent of its mineralization, and with the falling of solid and liquid precipitation during the thawing period. The ablation of naleds in the high-mountain region is retarded during the period of summer snowfalls and temporary snow covers.

The stratification of naleds is an index to the stages of their formation, i.e., a key to the interaction of the atmospheric and other factors in naled formation.

The dimensions and seasonal regime of ground naleds testifies to the extent of saturation of near-surface water-bearing horizons and to the extent of their summer-fall moistening.

The areas of large naleds, as O. N. Tolstikhin[26] has established for the USSR Northeast, constitute an index to the outflow from springs, from which the naleds are formed. This relationship is linear, with an excellent coefficient of correlation (0.75). However, there is no such relationship for naleds with an area of less than 1 km^2. He has also established that the numerical values of relative naled extent and naled runoff are practically equal. Consequently, for a rough estimate of the total naled runoff, it is adequate to determine the ratio of naled-areas to the area of the region in which they form.

Naleds are indexes to the extent of exposure of hydrogeologic structures and outcroppings of subsurface water, indexes of tectonic dislocations, and contacts of rocks having varying lithologic composition. The naleds characterize the productivity of the water-bearing horizons. The chains of naleds indicate the direction of cracks or of fault zones.

The naleds provide a concept concerning the chemical and gaseous composition of subsurface water and can be adopted as auxiliary evaluating criteria in the searches for rare elements and other minerals, especially in those cases when the naleds have formed from deep subsurface water. An interpretation of the results obtained from chemical analyses of naled ice has been given by I. A. Nekrasov and others[12] in the example of the lower Ingamakitskaya naled. The utilization of naleds as prospecting indexes in the eastern Trans-Baikal region has been demonstrated by Yu. V. Pavlenko[15] during the discovery of a fluorite deposit when the preceeding large-scale geophysical operations had failed to yield any positive results.

Naleds are an index to the direction of subsurface runoff. The naled hummocks form during

TABLE 2 Average Amount of Ablation of Naleds in Trans-Baikal as Compared with Other Regions in the Asiatic Sector of the USSR

Location of Naled	Coefficient of Thawing, mm/degree	Bibliography
Anmangyndinskiy Naled (upper reaches of Kolyma River)	1.1	7
Lower Kyrskaya Naled (basin on Indigirka River)	3.5	21
Central Yakutia	4.5	9
Dozhdevoy Creek (basin of Tetyukha River)	5.5	21
Lower Ingamakitskaya naled (Stanovoye upland)	6.5	4
Eastern Sayan (river naleds)	5.0	30
Central Asia	9.6	1

the presence of a transverse permafrost belt along the path of subsurface flow. However, the chains of hummocks run along the permafrost belt. The riverbed hummocks are indications of taliks beneath the channel in the permafrost zone.

Examples of the reconstruction of ancient glaciation in the Trans-Baikal region based on the contemporary snow cover and naleds are given by V. P. Chichagov[28] for the Khentey-Chikoyskoye upland and by V. S. Preobrazhenskiy[19] for the Stanovoye upland. Based on data furnished by V. S. Preobrazhenskiy,[18] in the eastern part of the Stanovoye upland, the total area of naleds greatly exceeds the area of contemporary glaciation.

Subsequently, this provided him[20] with a basis for expressing the concept that the naleds in northern Trans-Baikal comprise a specific form of the manifestation of contemporary late glaciation.

The dimensions of naleds around the wells for watering livestock in pastures, in populated points, and on transport routes constitute indexes of the proper or improper maintenance and utilization of wells. Based on the size of these naleds, we can make a rough estimate of inefficient utilization and losses of water. The number and dimensions of naleds that have newly appeared along railways, highways, and seasonal vehicular routes on ice provide evidence as to how successfully a route has been chosen and how completely we took into account the hydrogeologic conditions during the construction of a road.

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FORMATION OF NALEDS IN BASIN OF KHUBSUGUL LAKE (MONGOLIAN PEOPLE'S REPUBLIC)

B. I. PISARSKIY, G. M. SHPEYZER, D. BADRAKH, L. A. MINEYEVA, AND E. E. ENEBISH Soviet-Mongolian Combined Khubsugul'skaya Expedition

The region under study, with a total area of more than 5,000 km^2, encompasses the Khubsugul'skaya neotectonic basin of Baikal type[5] and its mountain framing, comprising the alpine ranges at the juncture of the folded systems of eastern Sayan, Tuva, and spurs of the Khangay system.

Important information concerning the natural conditions, geologic structure, and developmental history of the region at the contemporary level of study has been generalized in a report by N. A. Marinov[3] and the symposium: "Natural Conditions and Resources of Prikhubsugul'ye Region."[4]

At this point, we will merely point out that the basic factors influencing the nature, extent, and intrayear distribution of surface and subsurface runoff and promoting the formation of naleds are as follows:

1. Presence, in the basin's central part, of deep (up to 246 m) Khubsugul Lake with a basin area of about 2,700 km^2, receiving more than 60 small rivers, streams, and temporary streamflows. From the lake, there flows the only river, the Egin-gol, with an average yearly discharge of 18 m^3/s.

2. The great ruggedness of relief and its asymmetry (western slope--steep and high, eastern--depressed and sloping more gently), with overall lowering of absolute heights of watersheds from 3,000-3,500 m on the north and northwest to 1,800-2,000 m on the south, in the headwaters of the Egin-gol River.

3. The continental nature and harshness of climate, associated with remoteness of the region from the seas and oceans and its high location (absolute elevation of Khubsugul Lake is 1,645 m). At an annual precipitation of 250-450 mm, 80 percent falls during the warm season (June-September).

4. A predominant distribution, in the mountainous sector of the basin, of crystalline (igneous and metamorphic) rocks (gneisses, marbles, granitoids, limestones) of varying age, from pre-Cambrian to Cenozoic. The unconsolidated Quaternary beds occur only in the estuarial parts of large streams and lake-shore terraces. No information is available concerning the thickness and composition of the sedimentary layer in the Khubsugul basin. Based on general geologic assumptions, we can only theorize that here the thickness of sediments does not exceed the first hundreds of meters.

Faults are extensively developed, chiefly of northwestern and northeastern strike.

5. Continuous permafrost occurs with thickness up to 100-200 m.[2] The through-taliks that have been revealed are linked with the zones of tectonic dislocations and sectors involving the development of karsted carbonate rocks.

A main source of supply for the rivers is provided by rainfall; the snow and subsurface supply play a subordinate role for most of the channels. In October the formation of ice cover on the rivers begins and by December the rivers are completely frozen through. The melt water runoff begins in early April, and by mid-May rivers are completely ice-free.

Regional subsurface runoff is also associated with the infiltration of rainfall and is accomplished chiefly in the form of slope runoff. As we approach the beginning of November, liquid subsurface runoff stops, and the water (having been preserved within the active layer) converts to the solid phase.

Available data permit us to conclude that, in the critical period of the year, liquid runoff is accomplished in the basin only in the zones of tectonic dislocations where the springs have been identified, from which the outflow during the critical period reaches 1-5 l/s in granites, gneisses, and basalts and 20-40 l/s in the carbonate rocks.

The naleds in the region under study can be subdivided into riverbed, spring-riverbed, and spring.

The riverbed naled forms in sectors with small gradients of the bottom, at rapid freezing-through of the cross section of river's channels, owing to the circulation of water through the cracks in the ice cover or the breakthrough of channel water onto the shore and island banks. As a result, naled hummocks form with dimensions up to 4 × 30 m and height up to 4 m, extended along the channels (Khankha-gol, Noyan-gol, Toyn-gol, and others). The hummocks forming in the channels are comprised of pure ice, while the shore and island hummocks are formed from a mixture of ice and soil, pebbles, and gravel. Some of them are covered with a shallow soil-sod layer. In case of limited force of the water flow, the small hummocks are hollow. The formation of channel (riverbed) naleds is a brief process accomplished during November. The thawing of naleds begins from the second 10 days in April and is completed in the first 10 days in May.

The spring-riverbed naled forms in the sectors where the subsurface water in the zones of tectonic disruptions discharges, intersecting the river valleys, owing to the constantly flowing springs with water temperature during winter ranging from 2°C to 5°C. Below a spring, there is maintained a sector of liquid runoff with a length of 50-200 m, and then, as a rule, large naled fields up to 300-400 m in diameter form. The presence of a constant source of supply predetermines the continuity of the process of naled formation accomplished in three periods: the first--the increment in naled owing to the freezing of spring and river water (October-November); the second--increment in naled only from spring water (December-first 10 days in April); and the third--thawing of naled (from the second 10 days in April). The complete thawing of naleds is accomplished from July-August; some of them do not thaw out until the beginning of the new naled formation.

A specific feature of this type of naleds is the total expenditure of spring runoff on the formation of the naled, the result of which is the direct dependence of maximal volume of naled on the discharge of the feeding spring. The maximal volume of the naleds investigated, being fed by springs from the zones of faults in granites (Zhilkh-arshan) and in basalts (Del'ger-bulak) from 1971-1972 comprised 30,000-4,8000 m^3. The average discharge from springs necessary for creating such a volume of naled, according to calculation, should be 3-5 l/s, while actually the discharge at the end of the critical period in these springs comprised 0.5-1.0 l/s. Consequently, during the time of naled formation, the discharge of the feeding spring decreased gradually owing to the exhaustion of subsurface water supplies. This was confirmed during an investigation of these springs from July-August, when their discharge equalled 5-10 l/s.

In the carbonate rocks, the outflow of springs from the fault zones is maintained constant throughout the year,* in connection with which the level of discharge from springs necessary for creating the maximal volume of naleds amounting to 180,000 m^3 (Burgas Spring) and to 750,000 m^3 (Spring on Bayan-gol River) is 19 l/s and 40 l/s corresponded to the actual discharge rate of 20-50 l/s at the end of the critical period.

The spring type of naled occurs on the western shore of the lake near the mouth of the Chzhiglig River. The spring is located at the foot of a slope; runoff from it is accomplished along a vaguely revealed channel on the surface of the lake terrace. Water runoff in the lake was from a naled, the maximal volume of which (based on traces) reached 10,500 m^3. The average calculated discharge from the spring for the formation of such a naled should be 0.5-1 l/s. Obviously the discharge constantly decreased in winter, and probably by March-April the spring had completely stopped flowing. The thawing of the naled by analogy with others probably began in mid-April, in connection with which on the date of investigation only individual ice blocks were preserved.

Of great interest are the physicochemical processes occurring during freezing and thawing of ice. These processes are extremely complex and at the given stage of study cannot be described in detail or explained theoretically. At the same time, the results obtained from analytical and experimental studies have permitted us to reveal certain general tendencies in the variation of mineralization and composition of naled water during the phase transitions.

Among these tendencies, typical for all of the types of naleds classified, we include the following:

1. The formation of all the naleds studies began with the discharge onto the surface, of the magnesium-calcium waters with mineralization hydrocarbonates ranging from 0.1-0.25 g/l. In this connection, as systematic observations have indicated, mineralization and composition of feed water changed very little during the entire period up to the beginning of thaw.

2. The accretion of a naled and crystallization of the original water in final analysis led to the formation of an ice layer, heterogeneous in composition and degree of salt saturation in vertical section, independently of the overall naled thickness. In this connection, we found an increase in the overall mineralization and concentration of the ions HCO_3^-, CO_3^{2-}, Ca^{2+}, and Mg^{2+} upward along the section toward the surface of this naled. The concentration of salts in the near-surface layer often exceeded by several times the mineralization of the original water. Thus, in the spring-type naled near the Chzhiglig-gol River in the near-surface ice layer in a direct identification at the place of sampling, there were 1,280 mg/l HCO_3^- and 24 mg/l CO_3^{2-}, and, after complete thawing of the ice sample, from 2.5 l of solution, 8 g of $CaCO_3$ precipitated, while the mineralization of water was 230 mg/l at 150 mg/l concentration of hydrocarbonate ion.

The supersaturation of the upper ice layer by calcium carbonate was accompanied by a sharp increase in the pH to 9-9.5.

Downward along the section, there was a decrease both in the overall mineralization of ice and in the concentration of the ions listed. However the content of ions SO_4^{2-}, Cl^-, and Na^+ varied slightly within the limits of the entire section. Moreover in connection with the high concentration of Na^+ in the ice composition, its relative content increased downward along the section and the lower ice layers at very slight mineralization contained sodium-hydrocarbonate, while the medium was neutral (pH:7.5-7.7).

In this manner the results from the studies conducted on the Prikhubsugul'ye naleds confirm the regional nature of the physicochemical processes involved in crystallization of naled ice revealed by L. T. Chistotinova and O. N. Tolstikhin[6] during investigations on the Ulakhan-Taryn naled.

3. The processes occurring in the thawing of naleds led to a systematic decrease in mineralization and a modification in the ice composition. In the initial period (April-May), the ice thawed from the surface of the naled having maximal mineralization (up to 0.4 g/l), exceeding that in the original water, less often equalling it (naleds on the Noyan-gol, Turug-gol, Bayan-gol, Khankha-gol, and other rivers), a calcium-magnesium carbonate-hydrocarbonate composition, and alkaline reaction (pH 9-9.2).

Later (May-June), the middle layers of the naled section thawed where the ice mineralization decreased to its value in the original water, and below (0.1-0.15 g/l), and the medium became slightly alkaline (pH 8-8.2) containing magnesium-calcium hydrocarbonate (naleds on Chzhiglig-gol, Dayan-gol, Khankha-gol, and other rivers). Finally the residual blocks of ice (preserved to July-August and sometimes even to the end of summer) had uniquely low mineralization (30-40 mg/l), neutral to medium (pH 7.5-7.7) with sodium hydrocarbonate, less often calcium- or magnesium-sodium (naleds on Chzhiglig-gol, Bayan-gol, Khankha-gol, and other rivers).

For a confirmation of the facts disclosed, the authors conducted experimental studies relating to the changes in mineralization and composition of naled ice on thawing after artificial freezing. They subjected to freezing at -6°C water samples obtained by thawing ice taken from naleds, while, for control purposes, they took river water not having been frozen previously. The original batch of water in all samples was 0.75 l. The thawing took place at 1°-2°C at successive sampling of equal (0.25 l) batches of water.

The results from the experiment (see Table 1) indicated that the physiochemical processes involved in freezing and thawing proceed in the direction of reduction in mineralization and variation in content from calcium hydrocarbonate and magnesium-calcium to sodium hydrocarbonate, i.e., analogously to such processes under natural conditions.

*N. I. T. and O. N. T, editors.

TABLE 1 Data Concerning Chemical Composition of Ice Obtained from Naled Water During Artificial Freezing

	Mineralization of Thawed Ice, mg/l	Content, mg/l					
		Ca^{2+}	Mg^{2+}	$Na^{+} + K^{+}$	HCO_3^{-} CO_3^{2+}	SO_4^{2-}	Cl^{-}
Naled on Chzhiglig-gol River							
Water prior to freezing	43.00	9.02	0.49	4.75	21.20	5.75	1.91
Ice							
Upper layer	122.29	12.82	1.95	18.75	73.20	12.76	2.81
Middle layer	79.62	6.3	0.24	17.00	41.48	11.93	2.67
Lower layer	62.55	3.15	0.97	13.75	36.60	5.76	2.32
Naled on Bayan-gol River							
Water prior to freezing	38.63	5.81	0.31	4.75	19.83	6.99	0.89
Ice							
Upper layer	62.46	4.81	0.49	12.75	34.16	8.23	2.02
Middle layer	62.11	3.94	2.55	9.75	36.60	7.41	1.86
Lower layer	37.39	1.97	1.34	7.00	19.52	6.17	1.39
Spring-type naled near Chzhiglig River on the shore of Khubsugul Lake							
Water[a] prior to freezing	234.50	43.69	3.77	4.00	152.50	5.34	1.20
Ice[b]							
Upper layer	176.88	24.05	2.92	19.50	115.90	11.93	2.58
Middle layer	181.01	25.99	5.22	14.25	122.00	11.11	2.44
Lower layer	70.87	5.12	1.94	12.25	42.70	7.00	1.86
Khankha-gol River							
Water prior to freezing	122.21	21.47	5.47	1.75	83.88	8.63	1.01
Ice							
Upper layer	242.38	36.47	7.78	16.00	164.70	14.40	3.03
Middle layer	97.81	12.60	5.47	5.00	65.88	7.00	1.86
Lower layer	39.09	0.63	1.58	8.75	20.74	5.76	1.63

[a]For the experiment, we took a sample of 2.5 l during its thawing, and 8 g of $CaCO_3$ was precipitated.

[b]During thawing, 14.5 mg/l of $CaCO_3$ precipitate was formed.

It is necessary to point out that similar results of tests on artificial freezing of water with mineralization, calcium-magnesium hydrocarbonate 0.05 g/l, were obtained by N. P. Anisimova.[1] The data from the studies and experimental work were not confirmed by the conclusion made by L. T. Chistotinova and O. N. Tolstikhin[6] concerning the migration of salts toward the base of a naled during its thawing for the conditions existing in the Prikhubsugul'ye region.

The studies conducted demonstrated that at all stages of the naled process, water and ice are characterized by weakly oxidizing conditions of formation (E°:280-380 mV). Among the microcomponents, we have found in small quantities: silicon (0.3-5.0 mg/l) and fluorine (0.01-0.6 mg/l), at abrupt fluctuations in the value of oxidizability from 0.8 to 27 mg O/l and a stably high content of dissolved oxygen of 10-11 mg/l. The data presented reflect the results from the first stage of the study of the naled runoff in the basin of Khubsugul Lake.

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TECHNIQUE AND RESULTS OF EXPLORING FOR FRESH SUBSURFACE WATER IN THE NORIL'SK MINING-INDUSTRIAL REGIONS UNDER CONDITIONS OF THE FAR NORTH

I. I. ROGINETS   Noril'skiy Mining-Metallurgical Combine

The discoveries of large mineral deposits in extenisive areas in the Far North constituted a basis for the creation of mining-industrial enterprises and complexes and also of populated places and modern cities.

One of the important conditions for the rapid development of deposits under harsh climatic conditions is the timely supply to the enterprises and population of high-quality water in the necessary amounts. Moreover, the requirements for water are quite high, since, in the difficultly accessible regions, we as a rule primarily develop large and valuable deposits that can quickly repay for the expenses invested in the construction of the enterprises.

From a hydrogeologic standpoint, at the present time the vast regions in the USSR North have been poorly studied. Therefore, among scientists and engineers, there is no standard opinion concerning the reserves of subsurface water in these regions and the conditions of their formation.

Thus V. M. Ponomarev[1] remarks: "Numerous examples show that the fresh subsurface water here (in the area with depth of frozen soils greater than 100 m--I.R.) contrary to the existing opinion has wide distribution and considerable reserves." Along these lines, N. I. Plotnikov[2] makes the generalization: "The available hydrogeologic materials show that in the permafrost area, relatively limited subsurface water reserves are being formed."

The study of the hydrogeologic conditions in the Far North for practical purposes is at the stage of research and gathering of experience from the search for subsurface water. Only the areas in which the exploitation of and exploration for minerals are under way, a unique type of "oases" in the snow desert, have been investigated in detail. The remaining areas remain uninvestigated.

With the passage of time, in proportion to the accumulation of experience concerning the study of hydrogeologic conditions in the already discovered and operating mineral deposits, we shall develop efficient techniques for exploring the subsurface water for water supply under the conditions of the Far North. In the present report, we present a method for exploring the subsurface water in the Noril'sk mining-industrial region situated on the northwestern border of the central Siberian Plateau. The climate in the region is subarctic and severe, with a long winter (8 months) and a short summer. The average annual air temperature is -8.8°C. Average precipitation for the plains is 400 mm;

for the mountain areas, the total is 800 mm.

From an orographic standpoint, the region under discussion consists of two parts: of the eastern sector of the Yeniseie plain, elevated 50-100 m above sea level and of the northwestern sector of the central Siberian Plateau, with elevations ranging from 1,200-1,600 m, cut by river valleys 600-1,000 m deep. The main rivers in the region having year-round runoff are the Yenisei and Rybnaya-Noril'skay-Pyasina. The other, shallower, rivers freeze through in winter and have runoff only during the warm season.

From a geologic standpoint, the Noril'sk mining-industrial region belongs to the periphery of the Siberian Platform; in its structure, there participate the sedimentary, volcanic-sedimentary formations of the upper Proterozoic; carbonate-argillaceous beds in the Cambrian, Ordovician, Silurian, Devonian, and Carboniferous; terrestrial rocks in the Permian; volcanic formations of the Triassic; and terrestrial rocks of Jurassic, Cretaceous, and Quaternary beds.

The intrusions, related to a complex of Siberian traps, have penetrated into various horizons of the stratigraphic section, chiefly from the Devonian to the Triassic. The upper part of the section of rocks in the region has a negative temperature (frozen rocks).

In the mountain sections of the region from the river valleys to the mountain peaks, the thickness of the frozen rocks increases from 20-50 m to 400-450 m, at a decrease in their temperatures from 0.2-1°C to -7 or -8°C. Frozen rocks are lacking under the large lakes and rivers. The frozen rocks occupying vast areas in the region under review perform the role of regional water barriers and exert a significant influence on the conditions of supply, circulation, and discharge of the subsurface water. This effect of frozen rocks on the hydrogeologic conditions is typical for all the areas where they occur. A study of the permafrost conditions in the Far North has enabled us to establish the presence of through-taliks and the fact that the "permafrost is not conitnuous but is discontinuous."[1] The through-taliks are linked primarily with the accumulators and carriers of heat, which are formed by the rivers and large lakes that do not freeze all the way through. However, among the numerous types of taliks, of principal importance are those that are permeable for water.

An analysis of the results obtained from hydrogeologic studies conducted in the region since 1940 has permitted us to establish the degree of saturation of the rock complexes and also to make a determination of the quality of subsurface water. It was established that the youngest basalts, tuffs, and tuffites (among the bedrocks) of the Permian-Triassic chiefly form the mountain plateaus and have become frozen to depths of 300-450 m. The suprapermafrost water-bearing horizons of these rocks have low permeability and low water content. The shales in the Tungusskaya series of the Permian-Carboniferous in the entire region are practically devoid of water and act as water barriers. In the valley sectors of the region under study, we find marine sedimentary rocks with significant supplies of subsurface water; however, its increased mineralization, from 1.5-3.5 g/l to 40.0-65.5 g/l, precludes us from considering this water as an object of investigation for water-supply purposes.

Of basic interest in respect to the quests for fresh subsurface water are the Quaternary beds occurring in the ancient valleys, where the beds attain a thickness of 180-200 m.

The origin of ancient valleys is linked with the subsidence, in blocks, of the bedrocks as a result of tectonic activity and of subsequent river and glacial erosion. The lower part of the section of ancient valleys is formed by boulder-pebble beds of lower and middle Quaternary age. The thickness of the water-bearing horizon in them reaches 80 m, at average values of 40-50 m. In individual valleys, impermeable interbeddings of clays and clay-loams appear, separating the horizon into two parts.

Where valleys emerge from the mountain gorges onto the plain, the boulder-pebble beds taper out abruptly and are replaced by predominant clays of marine origin. Their thickness decreases to several meters. One obtains a unique dam of impermeable clays supporting the water-bearing reservoir. Since the ancient valleys have a considerable slope in the mountainous part, in turn predetermining the high gradient on the smoothed surface of the water-bearing horizon--up to 0.013-0.018 in places where the pebble beds thin out--conditions exist for an intensive discharge of subsurface water. Below such places, the outpouring of subsurface water from Quaternary beds to the daylight surface, naleds form that are appreciable in area and volume. The best studied naled in the Yergalakh River valley has an ice volume up to 10 million m^3; this establishes for the winter period (8 months) the expenditure of about 500 l/s of subsurface water in naled formation.

We shall discuss below the hydrogeologic conditions of one of the most thoroughly studied ancient valleys in the region, where, since 1963, a water reservoir has been functioning to utilize the subsurface water.

The length of this valley in the foothills is 6 km, with width varying from 500 m to 1,200 m. The water-bearing beds are of boulders and pebbles with a heterogeneous infilling, ranging from gravel and sands to clay-loams. The thickness of the water-bearing bed varies abruptly from the edges, where it equals from 2-10 m to 80 m in the valley center. By exploratory operations, we established the relation between the increase of permeability of beds with an increase in the water-bearing bed's thickness from 2-5 m/day at the edges to 200-260 m/day in the places of maximum thickness. The discharges from boreholes also increase in the same way from 10-18 m^3/h to 400-850 m^3/h. The individual outflows from boreholes drilled in the middle of the valley in isolated cases attain 225 m^3/h per meter. The basic factor determining the water abundance in the boulder-shingle beds is the infilling material. At the valleys' fringes, clay-loam predominates as a filler, while in the ancient valley the gravelly-sandy material is prevalent. All these factors predetermine the drilling of exploratory wells along the axis of the ancient valley.

In the ancient valleys of the Noril'sk region, relatively small natural supplies of subsurface water are formed. The static reserves amount to 60-100 million m^3; the dynamic reserves vary from 1,400 m^3/h to 2,000 m^3/h for the large valleys.

As experience gained in operating a test reservoir has indicated, the organization of large reservoirs (up to several thousand of m^3/h) is possible only if we can involve additional water supplies from surface runoff from rivers flowing over the area of the ancient valleys. However, the feeding of subsurface water by surface water in significant amounts can be achieved only where there are through-taliks having fairly high permeability. At the beginning of the flood stage, we were able to observe how in a through-talik with a length of 100 m, a flow reaching 10,000 m^3 percolates completely into the water-bearing bed and replenishes the reservoir, which has become dried up during the winter. Consequently, the hydrogeologic investigation of subsurface water should be closely linked with a simultaneous study of the hydrologic conditions in a river that would be regarded as practically the sole source of supply for the subsurface water, although we also do not exclude the replenishment of the bed through the discharge of water from the original water-bearing strata.

Let us stress here once again that, during the exploration of subsurface water in the ancient river valleys, special attention should be given to studying the permafrost conditions and to the priority discovery of the presence of through-taliks along the river channels; this is a uniquely important factor for establishing the future operational potentialities of the water-bearing horizon. In its turn, a significant tapping of water at a limited natural supply in the ancient valleys is possible under the stipulation of the consumption of the static supplies and of their subsequent augmentation by river water during summer, since during winter, with the advent of negative temperatures, the mountain rivers freeze through. In summer, as a rule the rivers have an average annual outflow from 1 m^3/s to 3-5 m^3/s; under favorable infiltration conditions, this is quite adequate for the complete replenishment of the spent reserves.

For a verification of this important factor under conditions of the ancient valley that we are discussing, experimental studies were conducted. In the first stage (from 1965 to 1969), the output from five wells was raised to 1,000 m^3/h; in the second stage (from January 1, 1970, to January 1, 1971), the output from nine wells was raised to 2,000 m^3/h; and in the final decisive stage (from January 1 to July 1, 1971), at the deposit 12 wells were functioning, with an output of 2,500 m^3/h. During this period we removed 18 million m^3 of water from the bed. The test wells were arranged in a longitudinal line. The spacing of wells in the southern sector of the series was 150-200m, and in the central sector it ranged from 250 m to 400 m. The outflow from individual wells was more than 360-450 m^3/h. In proportion to the increase in the draw off, the levels of subsurface water in the bed fell, and, during maximum outflow in 1971, the maximal drawdown in dynamic levels from the Earth's surface in the southern pressure part of the water-bearing level comprised 20 m; the drop was 25 m in the central sector and 35-40 m in the northern sector, with residual depth of water-bearing bed of about 50 m. In this connection, in the northern sector a depletion of the bed occurred. After the beginning of snow thaw (June 1, 1971), for the first 45 days the surface water completely replenished the subsurface water supplies that had been used during the winter. The levels of the water-bearing bed reached their maximal possible position, which, in each of the segments of the valley in the through-taliks' zone, is controlled by the water level in the river. As the test studies indicated, favorable replenishment condtions permitted us to calculate the additional possible depletion of subsurface water supplies during the winter and in this way to provide operating supplies of high-quality subsurface water from the ancient valley of Talnakh River, falling within high industrial categories.

After the completion of the experimental studies, an underground reservoir of nine operating wells is continuing to function; this reservoir's output comprises 1,500 m^3/h, the reserve capacities attain 750 m^3/h, rock structures are set up in the wells, and two separate electric lines have been connected. The pumps' operation was entirely automatically controlled. The gathering and main pipelines have been insulated against the effect of low temperatures. On disconnection of the pumps, to avoid freezing of the water conduits and wells, reverse circulation of water has been provided, with its passage via the pumps into the bed. The considerable savings, which amount to 500,000 rubles annually with the tapping of 1,000 m^3/h of water and also the reliable and highly productive operation of the water-bearing wells for a long period (7-8 yr), have confirmed the possibility of large-scale industrial development of subsurface water under the conditions of the Far North.

By analogy, we began exploratory operations for water in another ancient valley in the vicinity of Noril'sk. Based on the results, we clarified the hydrogeologic conditions and reserves of subsurface water; they proved to be significant and adequate to supply a large industrial city completely with high-quality water. In this ancient valley, among the Quaternary beds (the thickness of which attains 180-200 m), there occur two water-bearing horizons persisting in cross section and into which we are now drilling some operational boreholes.

In the Noril'sk region there is another series of ancient river valleys, which based on a preliminary forecast, include a considerable amount of subsurface water estimated at 4-5 m^3/s.

CONCLUSIONS

1. Hydrogeologic investigations conducted in the Noril'sk mining-industrial region have permitted us to reveal and estimate the reserves of subsurface water adequate for supplying a large mining complex and the city of Noril'sk; the studies also permitted us to establish the main trends for future exploration and development of subsurface water.

2. The main source of possible water supply to industrial enterprises and populated places under the conditions existing in the Noril'sk region consists of the ancient overdeepened valleys, including extensive subsurface water-bearing strata.

3. In the surveying and prospecting of fresh subsurface water, interdisciplinary projects should be launched in the valleys of all the rivers in the piedmont sectors. For a clarification of the conditions relating to replenishment of subsurface water along the riverbeds, geophysical and drilling operations must be conducted to study the permafrost conditons and to reveal the presence of through permeable taliks.

4. The potentiality of the surveying and exploration of significant reserves of subsurface water specifically in the ancient valleys of the Noril'sk region is confirmed by experimental and operational activities.

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PREDICTION OF NALEDS AND WAYS OF REGULATING THE NALED PROCESS

N. F. SAVKO   Omsk Branch of Soyuzdornii

Each winter over a vast area underlain by permafrost and also in many regions of deep seasonal freezing of soils, unique naled processes occur extensively.

The naleds create significant difficulties in human activity, particularly during road building. Naleds developing suddenly during the construction process are dangerous; therefore, the prediction of naled phenomena and the development, on its basis, of measures to prevent the naleds or to safeguard the structures from them has major importance.

An analysis of the accumulated information, results of many years' investigations at specially equipped stations and test areas,* show that naled formation represents a most complex physicomechanical process controlled by many factors. The most significant of them are the geologic and hydrogeologic structure of an area, the climatic conditions (air temperature, precipitation), exposure of slopes, relief of terrain, and engineering activity of man.

The complex natural conditions in which naleds develop, with the additional influence on their development from road structures and methods of their construction, cause the multiplicity of naleds near the roads. However, for controlling the naled process, the choice of the essential trends in the antinaled activity, and the assignment of antinaled measures and equipment, the classification of naleds should be simple and be based on the principal criteria determining the origin of a naled. These requirements have been met to the greatest extent by the naled classification suggested by N. I. Tolstikhin, A. I. Chekotillo, and S. M. Bol'shakov.

Based on the reports prepared by those authors and the results of personal studies, during the investigations and planning of engineering structures, including the antinaled measures and equipment, we recommend that the following classification of naleds be adopted.

Depending on origin, the naleds are divided into five main groups: spring, river (from

*In the preparation of the studies, we utilized the results from many years' observations at the Omsk branch of Soyuzdornii (State All-Union Scientific Research Institute of Roads and Highways) in the test areas of the Polovinka River and 1,420 km on the Magadan-Khandyga highway; we also used data provided by the Permafrostology Institute of the Siberian Branch at the USSR Academy of Sciences at the Ulakhan-Taryn test areas and on the Moma River, data furnished by the Chitinskoye Geologic Administration for the lower Ingamakitskaya naled, and other sources.

surface water of permanent and temporary streams), ground (suprapermafrost water), naleds from snow thawing in winter, and mixed types.

According to conditions of origin, naleds are usually natural, i.e., forming under natural conditions, or artificial, that is originating from the disruption of the water-thermal regime of soils and streams as a result of construction.

According to location, naleds are subdivided into slope types (declivity type) forming on the declivities of varying steepness, on the edges of ravines, at the crests and on the slopes of depressions, and valley type (ravine) forming in the relief depressions, i.e., on the bottom of ravines and in the valleys of creeks and rivers.

Naleds are subdivided into six categories by size.[3] According to the extent of danger, naleds are classified as not dangerous, i.e., not exerting any harmful effect on engineering structures; dangerous, causing a disturbance to the normal functioning of structures; and very dangerous, presenting a direct threat to the movement of traffic and to the stability of structures.

As a rule, the types of naleds that vary by origin have a varying mechanism of naled formation. The disclosure and mathematical description of this mechanism have been provided by interdisciplinary studies, including an analysis of the literature and data bank information on the distribution and development of naleds in Siberia; instrumental investigations of naleds at engineering structures along highways and railroads; laboratory studies on the freezing mechanism of the thin layer of water moving along an ice cover; observations of specially equipped stations and test areas dealing with the dynamics of naleds, temperature regime, nature of freezing in naled sectors, and hydrostatic pressure; observations of the functioning of experimental designs in antinaled structures; and solutions to the heat-engineering and hydraulic problems. In this paper we present only the final results obtained from the investigations in respect to a qualitative and quantitative prediction of naleds.

As a rule the naleds from suprapermafrost water (spring type) form in the bottoms of relatively small rivers and streams where the spring daylights. The diagrams reflecting the typical conditions involved in the formation of naleds from suprapermafrost water have been presented in Figure 1, based on studies conducted by V. R. Alekseyev, O. N. Tolstikhin, and others.

The sectors involving the development of spring-type naleds represent anomalies in the overall pattern of terrain conditions. In morphologic features, they differ markedly from the sites of naled formation from the suprapermafrost and mixed water. Usually the naled fields are nothing more than expanded sectors of the river valley bottoms, characterized by level, or by sinkhole-hummocky, relief. The typical wooded vegetation developing in a valley is lacking and is replaced by overgrowths of low bushes.

Frequently the naled fields are generally devoid of vegetation and represent placer deposits of sharply angled rocks or gravel. Within the limits of a naled field, the riverbed almost always separates into a series of meandering channels rapidly changing direction, which erode the naled bed intensively.

If the spring is fed by subpermafrost water (Figure 1b, c), the naled usually functions through the entire winter. The naleds formed by springs fed by suprapermafrost water (Figure 1a, d) are active only during the first half of winter owing to the depletion of the springs feeding them. Moreover, at the outcropping of a spring from the fringes of a valley (Figure 1a, b), the naled begins to form intensively immediately after the advent of icing over. For the discharge of subsurface water through the fissures along the through-subchannel-taliks (Figure 1c, d), two periods of naled formation are typical: the first, when the naled begins very gradually, and the second, indicated by the intensive development of the naled, usually coinciding with the dates of the inception of the lower air temperatures.

In addition to the duration of formation, the dimensions of a spring-fed naled are influenced quite extensively by the discharge rate, temperature and mineralization of the spring, thickness of snow cover, and topography of the naled valley. For predicting the average (H_a) and maximal (H_M) thickness, volume (V), area (F), maximal length (l_M) and distance (l_0) from the point of the spring's emergence to the body of the spring-fed naled, the following equations have been developed:

$$H_M = 5 \cdot 10^{-5}\,\mu\eta t\tau + 1.1h_\tau;$$

$$H_a = (0.5\text{–}0.65)H_M;$$

$$V = \frac{21.5 \cdot 10^3\, Q_M H_M}{\mu t};$$

$$F = Bl_M;$$

$$l_M = \frac{32 \cdot 10^3 Q_M}{\mu B t};$$

$$l_0 = 36 \cdot 10^5\,\frac{Q}{C} \cdot \ln\frac{A - t_S}{A - t_F}. \qquad (1)$$

where μ = the coefficient taking into account the properties of the snow pack [depth, h_S] and, depending on its thickness at the beginning of intensive naled formation, has the following values:

h_S, m	0.10	0.15	0.20	0.25	0.30	0.35
μ	0.90	0.70	0.55	0.45	0.375	0.325

η = coefficient equalling ratio of average (for the period of intensive naled formation) discharge from springs (Q, m^3/h) to the maximal discharge (Q_M, m^3/h) for this period; t = aver-

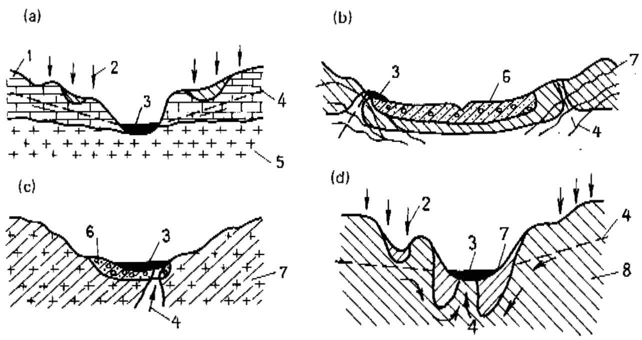

FIGURE 1 Diagrams showing typical conditions involved in the formation of spring-type naleds. 1--carbonate rocks; 2--ways of replenishing the supplies of subsurface water; 3--naleds; 4--direction of subsurface water movement along fissures; 5--crystalline rocks; 6--alluvial deposits; 7--permafrost layers; and 8--sedimentary rocks.

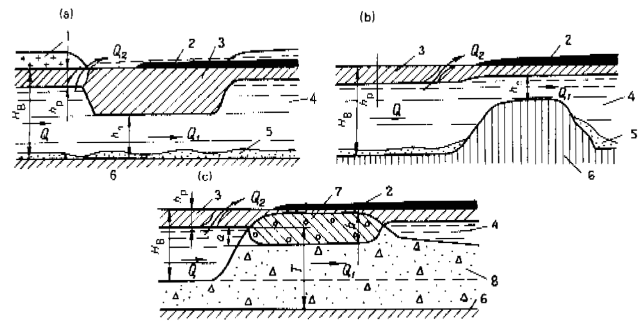

FIGURE 2 Diagrams indicating typical conditions involved in formation of a river naled. 1--snow cover; 2--naled; 3--ice cover; 4--water; 5--sand sediments; 6--water barrier; 7--frozen soil; and 8--talik.

age 10-day air temperature at the beginning of the intensive naled formation period, °C; τ = duration of period of intensive naled development, h; h_τ = quantity of precipitation, *m*, falling during period τ; B = average width of naled valley in m; t_S and t_F = respectively temperature of spring winter and temperature of its freezing, °C; and A and C = parameters that are determined with the equation

$$A = 8430\,Qi + \alpha Bt;\quad C = B(\alpha + \alpha_1),$$

where i = hydraulic gradient; α and α_1 = respectively the coefficients of heat transfer of water to air and from water to frozen ground, kcal/m^2 · h · deg.

A comparison of the design and actual data (see Table 1) permits us to form a conclusion concerning the adequately high reliability of the formulas suggested.

The mechanism of formation of river, and ground, naleds is basically one and the same. It reduces to the freezing of the water-bearing channels and to the development of pressure in individual sectors; under the influence of this pressure, water daylights and forms a naled. The diagrams of typical conditions involved in the formation of river naleds have been presented in Figure 2, while the diagrams for the ground-type naleds have been shown in Figure 3.

The difference in the mechanism of formation of river, and ground, naleds consists only in the mechanics of the movement of naled flow, which in the first case is based on the laws of the established motion of liquids in pres-

TABLE 1 Results of Calculations and Actual Observations of Naled Formation

Year of Observations	Average Thickness, m	Maximal Thickness, m	Maximal Volume, m^3	Maximal Length, m
Naled on Ulakhan-Taryn River				
1964/65	$\frac{1.15}{1.07}$	$\frac{2.30}{2.36}$	$\frac{1{,}250{,}000}{1{,}324{,}000}$	$\frac{3{,}000}{3{,}000}$
1965/66	$\frac{1.16}{1.16}$	$\frac{2.06}{2.02}$	$\frac{1{,}500{,}000}{1{,}446{,}000}$	--
1966/67	$\frac{1.56}{1.24}$	$\frac{3.13}{3.04}$	$\frac{1{,}630{,}000}{1{,}634{,}000}$	--
Naled on Anmangynda River				
1962/63	$\frac{1.68}{1.68}$	$\frac{2.94}{3.00}$	$\frac{11{,}600{,}000}{10{,}300{,}000}$	$\frac{6{,}750}{>7{,}000}$

NOTE: In the numerator we give the design values, while in the denominator, we indicate the actual[1] values for the naleds' parameters.

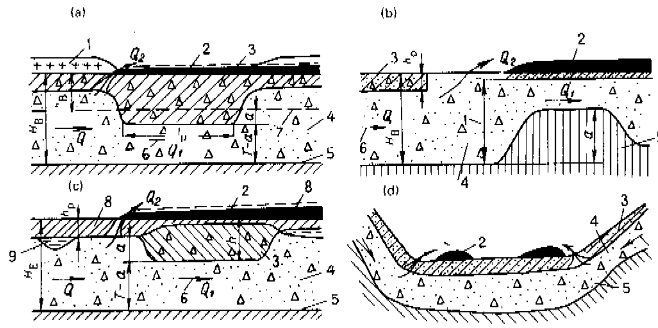

FIGURE 3 Diagrams indicating typical conditions involved in formation of ground naled. 1--snow cover; 2--naled; 3--frozen ground; 4--talik; 5--water barrier; 6--direction of ground water flow; 7--groundwater table; 8--ice cover; and 9--water.

sure-type pipes of variable section; in the second case, it is based on the laws of the unsteady flow of groundwater. Utilizing the stated laws and also taking into account the features of each of the diagrams listed in Figures 2 and 3 (for the diagrams in Figure 2a and Figure 3a and 3b, we typically have a periodicity in the pressure of the system; however, for the remaining systems, there is a constant pressure regime leading to the continuous outflow of naled water onto the surface), we obtained a series of analytical equations linking the average thickness of naleds with the basic factors of naled formation.

1. River naleds
   (a) are formed according to the system shown in Figure 2a:

$$H_a = 1.1Z = \frac{1.1\zeta_f K_1{}^2 l}{2gh_n^2}; \quad H_M = 1.5H_a, \tag{2}$$

(b) for the system in Figure 2b, H_a = 1.1 Z, where the Z value is determined according to the equation

$$1 = A\sqrt{Z} + BZ\sqrt{Z},$$

while the parameters A and B are found from the equations:

$$A = \frac{h_n}{K_1}\sqrt{\frac{2g}{\zeta_f l}}; \quad B = \frac{5}{K_1}; \tag{3}$$

(c) for the system in Figure 2c, H_a = 1.1 Z, where the Z value is determined from the equation

$$1 = A_1 Z + B_1 Z\sqrt{Z},$$

while parameters A_1 and B_1 are found from formulas

$$A_1 = \frac{K\phi}{\Sigma\zeta K_1\sqrt{I}}; \quad B_1 = \frac{5}{K_1}. \tag{4}$$

2. Ground-type naleds
   (a) are formed by the system indicated in Figure 3a

$$H_a = 1.1Z = 1.1\Sigma\zeta l^m(H_B - h_B);$$

$$H_M = (1.65 \text{ to } 1.5)H_a; \tag{5}$$

(b) for the system in Figure 3b

$$H_a = 1.1Z = 1.1\Sigma\zeta l^m\,(H_B - h_p); \tag{6}$$

(c) for the system indicated in Figure 3c, H_a = 1.1Z , where the Z value is determined from the equation:

$$1 = AZ + BZ\sqrt{Z},$$

while the A and B parameters are found from the formulas

$$A = \frac{1}{\Sigma\zeta l^m(H_B - h_p)}; \quad B = \frac{5\sqrt{I}}{K\phi l^m(H_B - h_p)}. \tag{7}$$

For facilitating the calculations, the equations for finding the values should be presented in graphic form.

In Equations (2)-(7), we have adopted the following notations: ζ_f and $\Sigma\zeta$ = total coefficients for resistance of subsurface contour;

$$\zeta_f = (1.5 \text{ to } 2.5)\,\frac{h_M}{H_B - h_p} + \left(\frac{h_n}{H_B - h_{np}}\right)^2;$$

$$\Sigma\zeta = 0.88 + \frac{3a}{T} + \frac{l_p}{T};$$

K_1 = specific rate of flow in closed contour at gradient i = 1, m^2/s.

At the design values for the coefficients of roughness of ice cover and of river bottom, $P_1 = P_2 = 0.04$, the K_1 values are as follows:

$H_B - h_p$, m	0.2	0.5	0.7	1.0	2.0	3.0	5.0
K_1, m^2/s	1.07	4.92	8.68	15.7	50.0	98.9	233

g = gravitational acceleration; K_ϕ = coefficient of soil permeability, m/s; at average coarseness of gravel 0.05-0.2 m and void ratio of soil 0.3-0.5, K_ϕ = 0.10 to 0.35 m/s; m = coefficient taking into account nature of flow; at laminar flow, m = 1, while at turbulent flow, m = 0.5.

For determining the volume, area and length of the river and ground naleds, we can utilize Equations (1), where the output Q_M is determined by the Chezy equation [$Q_M = 3{,}600\ ZBC_p\sqrt{Zi}$, C_p = Pavlov's coefficient at Z = (0.07 to 0.10)].

Along the small rivers and creeks with slightly-worked channels, where the main mass of water flows along the water-bearing riverbed deposits of the stream valley, we most often encounter naleds of mixed type; initially the naled is formed from the surface water in the water flow, then (the main volume) from the groundwater of the subchannel talik, and, toward the end of the winter and at the beginning of the spring, the naled is formed by the water from thawing snow. As observations made in test areas show, the ground-type naled accounts for about 70 percent of the total naled thickness. Thus, in the test area on the Polovinka River, where the naled is a mixed type and is formed according to the systems indicated in Figure 2c and Figure 3b (H_B = 7-8 m, h_p = 0.4 m, and, at the end of the naled formation in 1970, h = 2.5 m), the average thickness of river naleds was 0.15 m, while that of ground naleds was 0.8 m and that from thawing snow was 0.15 m. It should be noted that by calculation [Equations (4) and (6)], $H_a{}^R$ = 0.12 m and $H_a{}^G$ = 0.85 m. [$H_a{}^R$ evidently refers to a river naled and HG_a to a ground naled.]

An excellent agreement between the design and actual data was also obtained for the other test areas. Thus the average thickness (H_a) of the lower Ingamakitskaya naled for a 10-yr period formed according to the system in Figure 3c (H_B = 100 m, h_p = I = a = 100 m, and i = 0.003) was 2.1 m. By a calculation based on Equation (7), we derive $\Sigma\zeta$ = 10, A = 0.02, B = 0.4, Z = 1.85 m, and H_a = 2.04 m.

Investigations conducted according to the theory of naled formation permit us to solve a number of important practical problems:

1. To predict the sites of the most likely appearance of naleds and to pinpoint a series of necessary engineering-geologic investigations in the naled sectors.
2. To determine the basic quantitative characteristics of naleds for estimating the openings in bridges and pipes, determining the height of an earth embankment, for the planning and design of antinaled systems and structures in the naled sectors.[2]
3. To earmark measures for controlling the naled process for the purpose of its reduction or complete elimination.

An analysis of the decisions made indicates that in the regions with negative average annual air temperature the development of naleds should be anticipated:

1. In excavations, quarries, and at sites of foundation trenches and pits revealing water-bearing layers or promoting a partial or complete freezing of them.
2. Where water-bearing layers are blocked by solid foundations of structures and also by high embankments built of clayey soils.
3. In sections of streams with bars, rapids, debris cones, islands, layered gravelly beds, and banks; at a large quantity of rocks; in estuarial stretches of rivers and their tributaries.
4. Where engineering structures are built within slopes having outlets or shallow (less than the freezing depth) groundwater tables.
5. In places with a disturbed thermal regime of water flows and saturated soils--the mossy-peaty cover is removed, snow is cleared off regularly, heated buildings have been erected, etc.
6. Where there are open channels and troughs of considerable length, which are utilized during winter for the passage of subsurface water in streams and springs.

The most important factor in naled formation is the degree of obstruction of the water-bearing layer by seasonal freezing. By regulating the process of freezing (heating of channels, installation of insulation in the foundations of structures, the concentration of flows in heated channels, the laying of heating cables, etc.), we can also regulate the naled process. Among the progressive techniques for reducing the danger, lessening the growth, or the complete elimination of the naled process, we should include the draining of a locality by open and closed (drainage) canals and trenches; the deepening and straightening of riverbeds, increasing the mineralization of flows and the provision, within the limits of a structure, of increased rates of air movement; the preferential application of bridges instead of culverts; the application of lightweight designs, for example of bridges of the pile-trestle type or of suspension bridges completely spanning the entire naled valley; the excitation of the naled process at a safe distance from an engineering structure by means of the setting up of permafrost and naled belts and active antinaled banks; the application of cooling installations (for example, S. I. Gapeyev's system); and so forth. Reasonable measures for combating naleds are assigned in each specific case, with consideration of the genetic type of naled, terrain relief, and engineering-geologic conditions within the limits of the naled sector; these measures are coordinated with planning decisions made during the designs of engineering structures.

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REGIME OF NALEDS

B. L. SOKOLOV State Hydrologic Institute

One of the typical features in the water regime of the frozen zone is the formation of naleds, i. e. of masses of stratified ice forming in winter from subsurface and river water within the limits of the channel and floodplain sectors of river valleys. As a special type of seasonal glaciation, naled formation represents a phenomenon, typical chiefly of the mountain-folded regions, and is distributed most widely in the USSR Northeast. Here the areas of individual naleds attain several tens of square kilometers, while the volume reaches hundreds of millions of cubic meters. In all in this region we have found about 10,000 naleds, in which every year toward the end of winter there accumulates, while in summer as a result of thawing there enters the river network, approximately 30 km^3 of water, or 25 mm in relation to the total area of their distribution, equalling 1,210,000 km^2.

In recent times in the USSR, among specialists in various fields, interest has grown in a study of naled occurrences; this has been a reflection of the overall development of research on evaluating the natural resources in connection with implementing the plans for developing the northern and northeastern regions of the country. One of the main trends in this research includes an estimation of naleds as a factor in the formation and regulation of the reserves of surface and subsurface water. The naleds have come to be regarded as one of the most important indexes to the hydrogeothermal processes occurring in the permafrost zone and determining the principles of its heat and water exchange.

Results obtained from research in recent years indicate that, in the naleds of individual river basins in the USSR Northeast, every year by the end of winter up to 50-80 mm of water accumulates, or 25-30 percent of the annual volume of river and 60-80 percent of the subsurface drainage into the rivers. In winter, on the formation of naleds, 50-85 percent of river flow of the winter season is diverted, while in the freezing rivers the entire flow concentrates in the naleds. During the warm season, the volume of naled runoff attains 40-75 percent of the volume of river flow during the spring floodwater stage. Hence the naleds do not increase and do not decrease the amount of total water reserves, but only redistribute them from the cold to the warm season. From the quantitative description of naleds adduced, the requirement becomes quite obvious for studying the tendencies in their regime as a significant factor of the intraseason redistribution of water reserves.

The modern concepts concerning the nature of naleds and their relationship with the environment were established in the reports prepared by Russian and Soviet scientists S. Ya. Pod'yakonov, A. V. L'vov, M. I. Sumgin, V. G. Petrov, N. I. Tolstikhin, P. F. Shvetsov, B. V. Zonov, A. S. Simakov, O. N. Tolstikhin, and others. However, the existing concepts concerning the relationships in naleds' formation under natural conditions are mainly qualitative. The effort of individual researchers to express the naled-forming process in the form of mathematical equations has not found practical application. An empirical generalization of the observational information collected in recent years permits us to approach the discovery of the essentials in the naled regime.

A typical feature of the great majority of naleds includes the fact that from year to year they form at identical locations and are similar in their overall morphological structure. The formations of naled sectors in the form of an expansion of the riverbeds and floodplains of river valleys occur as a result of erosion from slopes during the passage of river flow around the naleds. In a general case, the dimensions of these sectors are a function of the outflow from the naled-forming source, the energy of river flow, and geologic structure determining the resistance to the eroding influence. The generality of the morphologic structure is expressed in the fact that the relationship between the dimensions of naleds at identical stages of their formation remains about the same. This provides the possibility of extrapolating the principles discovered to objects that have not been studied.

In a general case, for the origination and further development of a naled, we require a source of water and conditions, as a result of which the surbsurface or river flow emerges to the surface of ground or ice. In the absence of one of these causes, a naled fails to develop.

The question concerning the first cause, namely, a source of supply to the naleds, is debatable up to this time. Among all researchers, there is no doubt of the subsurface origin of naled-forming water. The differences in opinions develop in respect to participation of various categories of subsurface water in the frozen zone, namely the suprapermafrost, intrapermafrost, or subpermafrost water. Most of the researchers are convinced of the mixed supply to naleds; this does not exclude the predominance in isolated cases of water of one or another type.

The second necessary condition for the formation of a naled is the presence of a barrier obstructing the movement of flow along the customary path and leading it to the surface. These barriers develop at the points of abrupt reduction in the slope of a valley, daylighting of compact bedrocks, the freezing of water-enclosing rocks of slight thickness, reduction in permeability of rocks, total freezing of river flow, etc. The complex aggregation of the conditions mentioned determines the wide diversity in the actual causes, as a result of which any given naled develops. At the same time, the regime of naleds is subjected to general effects, which in turn are linked with the generality of natural factors of the formation of naleds.

In the dynamics involved in naleds, during the year we clearly trace two periods, namely the period of growth and period of disintegration, differing in duration. As a rule, the naleds thaw out completely; therefore, the volumes of water that were expended in winter on their formation and that came into the riverbed network of a basin during the warm season from the thawing of ice are practically equal. For conditions existing in the USSR Northeast, intensity of water runoff from naleds is roughly 1.6 times greater than intensity of naled formation; this is stipulated by the ratio of the durations of the cold and warm seasons of the year.

Most of the naleds begin to form after the stable passage of mean diurnal air temperatures through 0° and the freezing of the active layer. The largest of them (with an area greater than 0.2-0.5 km^2) accrete during the entire cold season and reach the maximal dimensions at the end of winter. The formation of the smaller naleds is finished sooner and depends on the depletion of groundwater supplies. Often the development of ground-type naleds stops, even at the beginning of winter, and then becomes renewed from March to April. Sometimes the accretion periods also occur in midwinter. The intermittent development of ground-type naleds is explained by the ratio of the freezing depth and the height of the groundwater table. If the groundwater table during usage of this water drops beneath the freezing depth, the naled process stops, since the hydrostatic pressure disappears, and vice versa.

The total duration of naled formation, other things being equal, is determined by the supplies of naled-forming water. The empirical curves in Figure 1 show that the time for naled formation for naleds with volumes greater than 4×10^6 m^3 and with areas greater than 0.2 km^2 essentially do not depend on their dimensions and on an average is about 210 days.

The rate of growth in volumes and areas of naleds varies, as is illustrated in Figure 2, where we have used data from observations of a typical naled in the Northeast in the Anmangynda River basin. The increase in naled area at the initial period of naled formation occurs at a faster rate, as compared with the variation in its volume because of the circulation of naled-forming water over the smoothed bottom of the naled sector. In the middle and at the end of winter, this ratio changes, and the increment in volume runs ahead of the increment in area. Other conditions being equal, the rate of increase in the areas of naleds is determined by

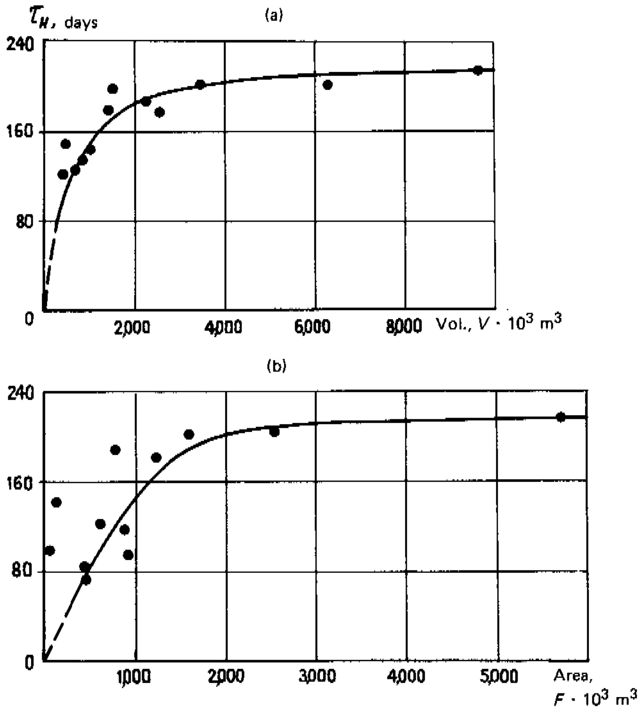

FIGURE 1 Time for naled formation τ_H versus dimensions of naleds.

the morphological structure of naled sectors, while the increase in their volume is established by the permafrost-hydrogeologic factors.

In comparing the total hydrographs of naled supply in individual river basins with the graph reflecting the variation in negative air temperatures, as expressed in fractions of a unit of the average values of these parameters for the season (modular coefficients), we showed their coincidence, not only in time but also in amplitude (Figure 3). In the given instance, air temperature determines the intensity and depth of freezing and consequently the intensity with which the naled-forming water is forced to the surface.

The decrease in volumes by 65-95 percent of naleds during the warm season is caused by the thawing of ice from their surface by radiation-advective thawing factors. A significant factor is also the mechanical breakdown of ice by river flow, which is particularly intensive during the spring flood stage and the appreciable rain-caused high waters. In individual periods of the warm season, the volume of mechanical disintegration reaches 30-35 percent and more of the total decrease in volume of the riverbed-type naleds.

The rate of decrease in volume and area of naleds during the warm season, in distinction from the rate of their growth in winter, is about the same (see Figure 2); it is confirmed by observational data for other naleds. In individual time periods, this does not hold because of specific features in the morphological structure of naled sectors.

In a general case, the rate of variation in

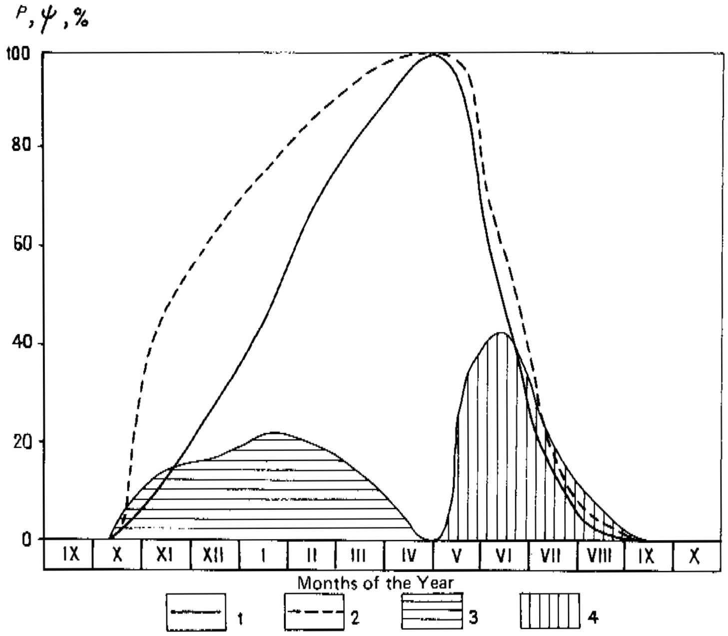

FIGURE 2 Curves of variation in volume and area expressed in percentage of their maximum values at the end of winter for Anmangyndinskaya naled in the process of its growth and disintegration through the closed cycle, 1963-1964. 1--volume of naled (ψ); 2--naled area (*P*); 3--hydrograph of naled supply; and 4--hydrograph of naled runoff.

naleds' areas and volume is given by the tangent of inclination angle or be the first time derivative at each point of the integral curves (see Figure 2). If we replace this derivative by the increment in function for equal time intervals, we can obtain an averaged parameter reflecting the variation in rate of naleds' formation. We have shown in Figure 2 the curves of monthly increments in volumes of Anmangyndinskaya naled expressed in percentages of its maximal volume at the end of winter, 1963-1964. Essentially, they comprise hydrographs of naled supply and of naled runoff; they permit us to obtain the absolute values of water volumes expended on forming a naled in winter and arriving in summer at the riverbed network of the basin. The hydrographs shown provide evidence that the maximum of naled supply arriving in January is roughly one-half as much as the maximum of naled runoff arriving in June. With minor deviations, the tendency indicated is typical for the great majority of large naleds in the USSR Northeast.

In showing the general relationships in the regime for various naleds, the asynchronicity in dates of the beginning of their formation can be eliminated, if, as one of the basic arguments, we utilize the running time ordinate, having made consistent the dates of the beginning of naled formation. Figure 4 shows curves describing the rate of increase and of decrease in volumes and areas of naleds during the warm and cold seasons. The curves have been constructed based on observational data for naleds located in various regions in the permafrost zone; in the USSR Northeast, in central and southern Yakutia, in the Trans-Baikal, in the Far East, and in the Maritime region. As a basic relationship, it is necessary to note the coincidence of seasonal curves for naleds of identical dimensions in respect to both volumes and areas, which is associated with the generality of the factors involved in their formation. In Figure 4c and 4d, the upper curves describe the dynamics of naleds with maximal values of their areas at the end of winter from 0.6 km^2 to 23 km^2 and with volumes ranging from 1.5-10 $\times$ 10^6 m^3. The curves for the smaller naleds lie lower down, since the process of their disintegration occurs more rapidly.

An analysis of the observational data permitted us to establish that the variations in average size (thickness) of naleds during the process of growth and decay do not characterize their

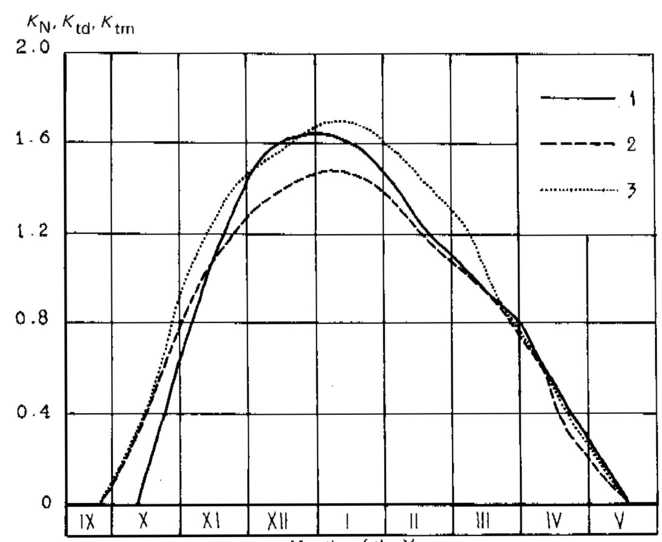

FIGURE 3 Variation in modular coefficients for naled supply (K_n), of average 10-day (K_{td}) and average monthly (K_{tm}) negative air temperatures in Indigirka River basin, as far as the Yurta settlement (based on averages of many years' data).

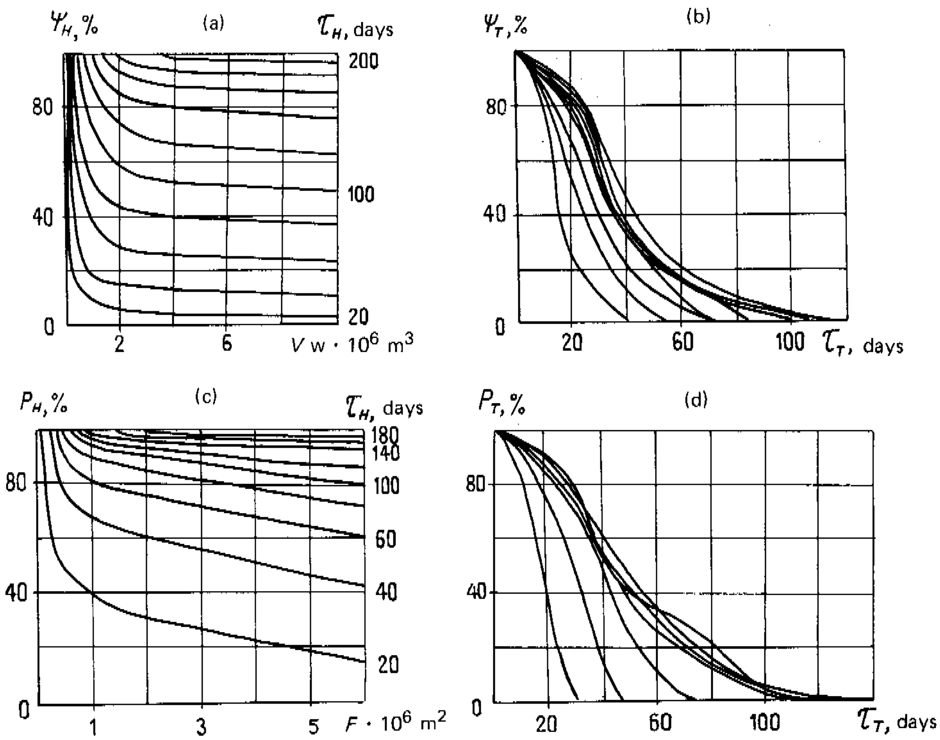

FIGURE 4 Curves showing the increase in volumes (ψ_H) (a) and areas (P_H) (b) and also decrease in volumes (ψ_T) (c) and areas (P_T) (d) expressed in percentages of their maximal values at end of winter (*V* and *F*, respectively) for naleds located in various regions of the permafrost zone, in the process of their growth and disintegration as a function of the number of days after the beginning of naled formation (τ_H) and start of thawing (τ_T).

regime. In individual periods during winter, it can decrease; in the warm season, it can increase, both as a result of the disproportionate change in areas and volumes of naleds and because of the influence of certain features in the morphological structure of naled sectors.

Based on observational data from 310 naleds located in various regions of the permafrost zone, including in Alaska, we established the dependence of their maximal volumes at the end of winter (*V*, thousands of m^3) on areas (*F*, thousands of m^2):

$$V = 0.96\ F^{1.09}. \qquad (1)$$

This relation is based on the generality of natural factors in the formation of naleds and the morphological structure of the naled sectors. To some degree, the latter is the result of the functioning of the naleds themselves.

The long-term variation in the naleds' dimensions is determined chiefly by variability in climatic conditions. The maximum naled formation usually corresponds to severe winters, and hence at this time we have the maximum dimensions of naleds. Owing to its insulation properties, the snow cover protects the ground from deep freezing; as a result, in winters with abundant snow, the naleds' dimensions decrease. A generalization of the data from many years' observations on naleds permitted us to obtain the following approximate empirical formulas for the coefficients of variation in annual volumetric values (C_{V_V}) and areas (C_{V_F}):

$$C_{V_V} = \frac{2.6}{V0.356};\ C_{V_F} = \frac{2.0}{F0.264}. \qquad (2)$$

According to these equations, the larger the naleds' dimensions, the less their variation in the many years' term, and vice versa.

In conclusion, it is necessary to point out that the results obtained, typifying the general relationships in the naleds' regime, apart from their belonging to any certain region, doubtless have scientific and practical interest in a study of the hydrogeothermal processes in the permafrost zone in connection with its industrial development. The relationships discussed have been utilized in the State Hydrologic Institute in developing a procedure for estimating the naled supply and naled ablation and also for evaluating their effect on the fluvial and subsurface flows for rivers in the USSR Northeast.

FEATURES IN NALEDS' DISTRIBUTION IN CENTRAL PART OF OLEKMA-VITIMSKAYA MOUNTAINOUS COUNTRY

K. A. CHERNYAVSKAYA All-Union Aerogeologic Trust

The basic orographic elements in the central sector of the Olekma-Vitimskaya mountainous area, namely the Kodar and Udokan ranges and Verkhne-Charskaya Depression, are caused by the nature of the most recent movements and have a northeasterly strike. The maximal relative height of the ranges' peaks above the valley floors is 2,380 m. The rivers belong to the Lena basin; many river valleys occur along the faults and fissures. The mean air temperature is negative, i.e. -4° to -11°.

The quantity of atmospheric precipitation increases with height from 600 (in the depressions) to 1,000 mm/yr (in the mountains). About 60 percent of the precipitation falls during the brief cool summer. The severity of the geocryologic conditions is expressed by the great thickness (in places more than 1,000 m), by low temperatures (to -15°) in the permafrost zone in the mountains, and by their sharp contrast with temperature and thickness of rocks in the depressions. We typically find a slight discontinuous state of the frozen zone; in the very largest depression, namely the Verkhne-Charskaya, on an average it does not exceed 5 percent.[2]

The main part of the mountain structures is formed by sedimentary-metamorphic and intrusive rocks of the pre-Paleozoic age. In the watershed of the Udokan Range, the Cenozoic volcano formations occur. In the depression, we find the carbonate, terrestial, and unconsolidated Meso-Cenozoic deposits. In the tectonics, the leading role is played by the large variously oriented faults of thrust type, having repeatedly become activated all the way to the contemporary period. We should differentiate the northeastern and sublatitudinal systems of faults. Among the first of them is included the rift structure consisting of the upper Charskaya depression, extended linearly from southwest to northeast, framed on the north by the high-throw Kodarskiy block-type upthrust and on the south by the Udokanskiy block-type upthrust.

The manifestations of most recent volcanism are localized in the southwestern part of the rift structure. At an angle to the rift-forming faults, there occurs the sublatitudinal zone of clustering of epicenters of contemporary earthquakes with intensity, K, up to 13 or 14, the maximum concentration of which has been registered in the graben-valley region along the Konda River.

In the mountain structures, we typically find a broad distribution of water in fissures and cracks, a considerable quantity of which occurs in the fault zones intersecting the igneous and ancient metamorphic rocks. In the upper Kalarskaya depression, we find a prevalence of fissured-karst water in the carbonate rocks, while in the upper Charskaya and other depressions, we find pore water in the loose deposits.

On the map showing the naleds in the central part of the Olekma-Vitimskaya mountainous country, compiled on the basis of field investigations by the author, with aerial-visual observations and the interpretation of aerial photographs, we have shown the naleds with area greater than 0.1 km^2 (Figure 1). The processing of data concerning the distribution of 537 naleds by area (35,000 km^2) and height (from 380 m to 2,000 m absolute elevation), the construction of frequency curves and relationship of naleds' dimensions on sea-level elevations based on O. N. Tolstikhin's technique,[3] and also a determination of the relative naled concentration enabled us to reveal additional characteristics for each naled region and to emphasize the close relationship in naled distribution with neotectonics, geologic structure, and permafrost-hydrogeologic conditions in the mountainous country.

To the naled areas, there correspond the rift zone and the zones of extrarift mountain formation replacing the rift zone on the northwest and southeast.

The northwestern area is divided into two regions, namely the Patomskiy and Nichatskiy. In the Patomskiy region, the naleds are distributed in the height range from 380 m to 1,600 m (one naled is higher). The distribution curve for the naleds is elongated and asymmetric; the lower branch of it is steeper than the upper brance (Figure 2a). The maximal frequency of naleds is observed at height ranging from 450 m to 850 m. Naleds with area from 0.1 km^2 to 0.2 km^2 predominate. In proportion to increase in the sea-level elevation, the average areas of naleds* decrease regularly (refer to Figure 3a). The relative naled density comprises 0.22 percent.

The complex shape of the frequency curve of naleds by heights is explained by the complex morphology of relief and the heterogeneous conditions in supply of the naleds. Obviously the increase in naled frequency at height of about 1,500 m is caused by their supply from high-mountain glaciers. At an elevation below 500 m, there is an abrupt deterioration in the naled-

*Here and in the subsequent graphs, the deviations in the average areas of naleds from the norms are slight or do not occur.

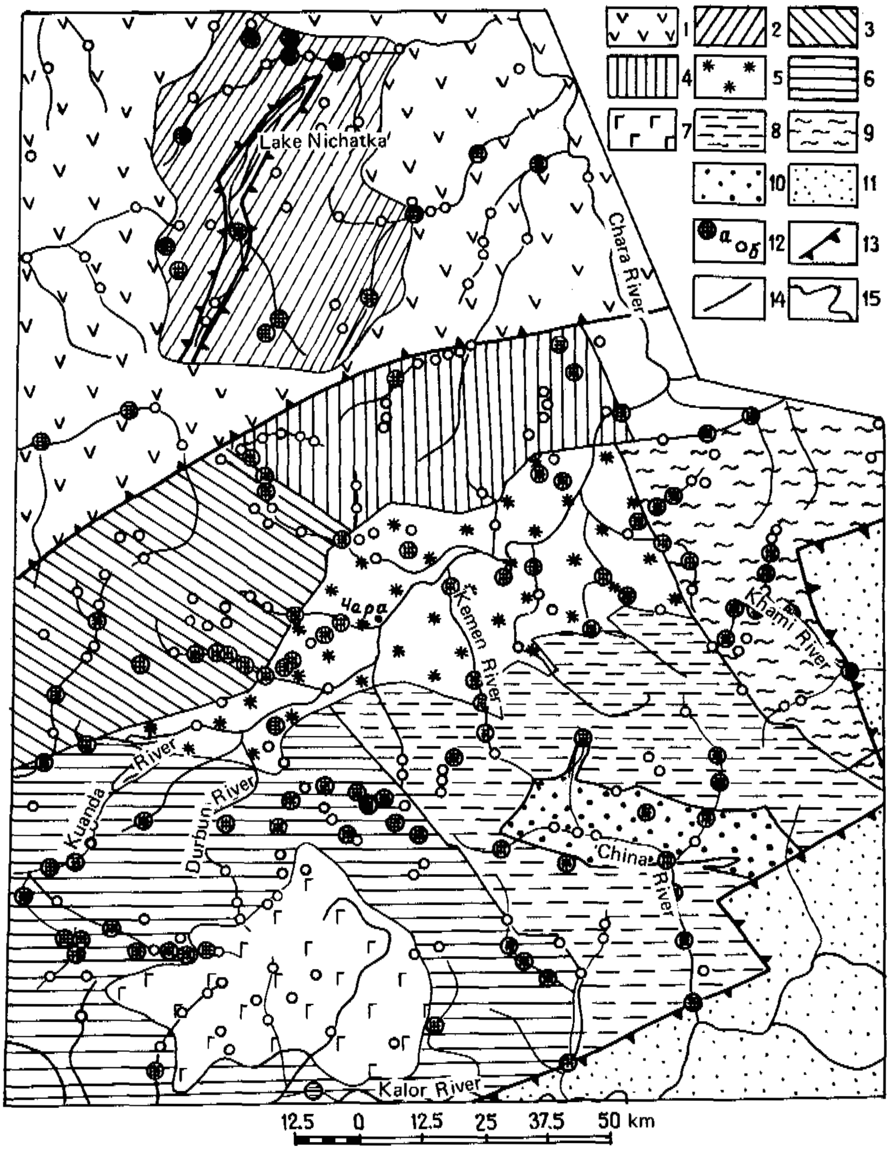

FIGURE 1 Map indicating naled regions and naleds in central part of Olekma-Vitimskaya mountainous country. *Naled regions of northwestern area:* 1--Patomskiy; 2--Nichatskiy. *Rift regions:* 3--western Kodarskiy; 4--eastern-Kodarskiy; 5--upper-Charskaya depression; 6--western Udokanskiy; 7--plateau basalts; 8--central Udokanskiy; 9--eastern-Udokanskiy; 10--Verkhene-Kalarskaya depression; 11--southeastern region; 12--naleds having an area in km^2 (a--0.1-0.5, b--0.5-5.0); 13--boundary of naled areas; 14--tectonic boundary of naled basins; 15--geomorphologic boundary of naled basins.

formation conditions owing to the existence of the nonfreezing talik beneath the Chara River channel and by the extensive development at these elevations of crystalline shales of low water content in the Udokanskaya series of the lower Proterozoic.

The Nichatskiy region includes the "newly formed" graben of Nichatka Lake and the adjoining basins of El'ger, Bogayutka, and Tora rivers. The crystalline rocks in it are separated by a denser network of radial faults, renewed during the Cenozoic time. The boundary of the naled region is drawn mainly along the watershed, the elevations of which do not exceed 1,000-1,900 m. The naleds occur at heights ranging from 500 m to 1,700 m; a large share of them occurs in the 850-1,150-m range of sea-level elevarions. The distribution curve by altitudes is less extended, with convex lower and concave upper branches (Figure 2b). The naleds with a size up to 0.02 km^2 and from 0.1 km^2 to 0.2 km^2 predominate; the curve of frequency by sizes has two distinct peaks. In proportion to the increase in absolute elevations in the naleds' distribution, we can trace a slight reduction in the average values of their area (Figure 3b). The relative naled density is 0.59 percent.

The intensively fissured state and its associated increased abundance of water in rocks within the boundaries of the Nichatskiy region has caused the differences in the shapes of the distribution curves for the naleds and relative naled density for the Nichatskiy and Patomskiy regions. At elevations below 500 m, a talik occurs near Lake Nichatka, establishing the lower limit of occurrence of naleds by altitude. The relative nature of comparative naled density of all regions is represented graphically in Table 1.

The rift area is divided into eight naled regions, which correspond basically to the neotectonic structures II, III, and higher. In the areas we have counted 405 naleds.

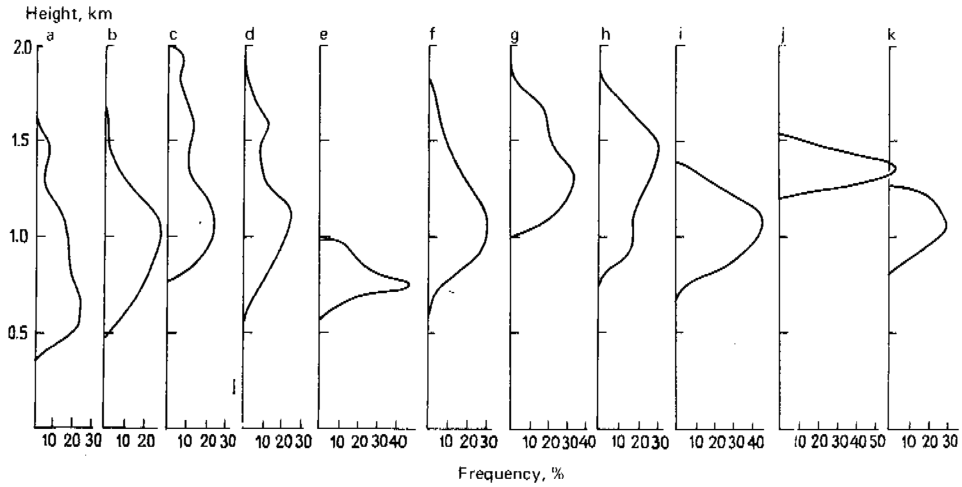

FIGURE 2 Frequency of naleds by heights. Frequency curves of naleds for regions: (a) Patomskiy, (b) Nichatskiy, (c) western Kodarskiy, (d) eastern Kodarskiy, (e) Verkhne-Charskaya depression, (f) western Udonkahnskiy, (g) plateau basalts, (h) central Udokanskiy, (i) eastern Udokanskiy, (j) Verkhne-Kalarskaya depression, and (k) southeastern area.

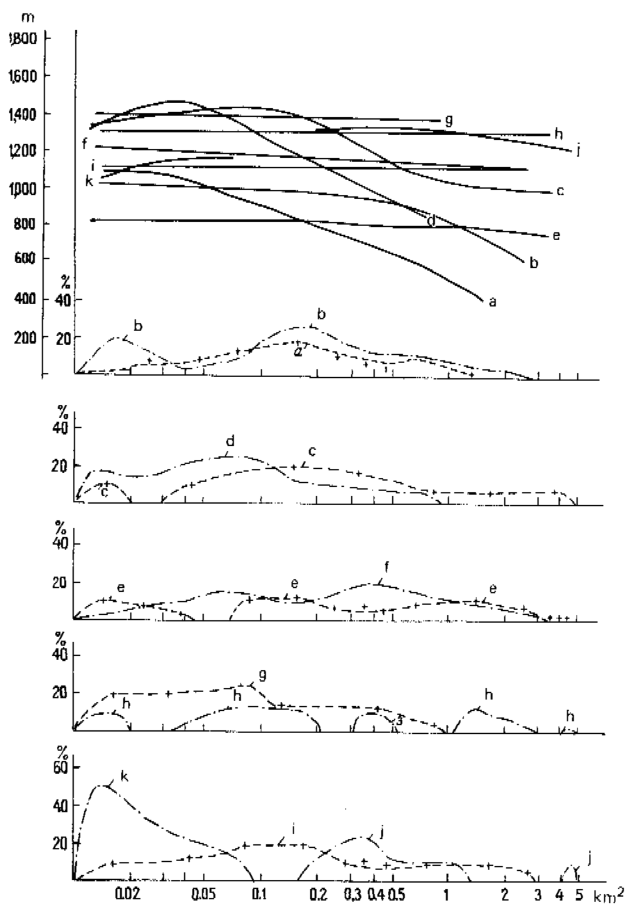

FIGURE 3 Naled areas and heights of terrain where they occur. 1--average sizes of naled (in km^2) and variation in average heights of their occurrence; 2 and 3--frequency curves for naleds by sizes. Distribution of naleds in regions: (a) Patomskiy; (b) Nichatskiy; (c) western Kodarskiy; (d) eastern Kodarskiy; (e) upper Charskaya depression; (f) western Udokanskiy; (g) plateau basalts; (h) central Udokanskiy; (i) eastern Udokanskiy; (j) upper-Kalarskaya depression; and (k) southeastern region.

In the western Kodarskiy region with absolute elevations up to 3,000 m, the naleds occur at heights of 800-1,200 m, while maximal frequency falls in the 900-1,200 m range. A certain increase in naled frequency is recorded above 1,500 m (see Figure 2c). Naled areas from 0.2 km^2 to 0.1-0.3 km^2 predominate. The distribution curve of average naleds' dimensions and average heights of their occurrence has a complex shape: Initially, with increase in elevation, we note a slight increase in average areas (up to 0.1 km^2), then an abrupt decrease in the 0.1-1.0 km^2 range; after this, the curve flattens out rapidly, and, at the 1,000-m mark, it is horizontal (see Figure 3c). Among the block-type upthrusts, this region is typified by the maximum relative naled density, i.e., 1.12 percent, at maximal amplitude of displacement in blocks and intensified seismicity.

Within the limits of the eastern Kodarskiy region with watershed elevations of 1,900-2,600 m, the naleds occur at 600-1,800 m (one naled occurs higher up). The distribution curve is asymmetrical, more elongated, and has two maximums--in the altitude range from 900-1,200 m and at 1,500 m absolute altitude (see Figure 2d). The maximal size of naleds is limited to an area of 1 km^2; the areas from 0.05 to 0.01 km^2 predominate. In proportion to increase in the elevation, we note a gradual increase in the average areas to 0.04 km^2; with height, the large dimensions of naleds diminish fairly abruptly (see Figure 3d). Relative naled frequency in the region is 0.59 percent. The overall resemblance of the distribution curves for the naled in western and eastern Kodarskiy regions emphasizes their belonging to the same neotectonic structure. The complex configuration of curves can be explained in terms of the contrasting block-type structure of Kodarskiy upthrust, characterized by the presence of deep narrow grabens. In particular, the gradually sloping right branch of the variation curve for the western Kodarskiy region corresponds to the elevation at the bottom of grabens, i.e., the main reservoirs of subsurface water in the mountainous structures.

A certain increase in average areas in proportion to increase in elevation marks and frequency of naleds above 1,500-m altitude is explained by the feeding of naleds by glaciers developed extensively in the Kodar high-mountain region.

In the upper Charskaya depression, the naleds occur over the entire floor at elevation ranging from 615 m to 1,000 m. The distribution curve of naleds by heights is almost symmetrical; their maximum number has developed in the 700-800-m range (see Figure 2e). Predominant are the areas in the 0.08-0.2 km^2 and 1-2 km^2 ranges; the variation curve reflecting naled frequency by sizes is greatly elongated and lacks a distinct maximum. The curve of average dimensions is subparallel to the *X* axis (see Figure 3e). Naleds with an area exceeding 1 km^2 comprise 24 percent; a large share of them is formed as a result of the discharge from a thick water-bearing horizon in sands.[1] Relative naled density is 1.38 percent.

The consolidated curve for frequency by heights characterizes the morphology of the depression with a level floor, somewhat elevated along the edges. The deterioration in the conditions involved in the naled formation in the Chara River channel and the lower reaches of the Kemen River is explained by the presence of nonfreezing taliks.

In the western Udokanskiy region (watershed elevation from 2,000 m to 2,400 m), the naleds occur at heights from 600 m to 1,800 m. The distribution curve is elongated, asymmetrical, with a steep rising branch (see Figure 2f). The naleds in the 850-1,250-m range of heights of grabens are developed the most. The average dimensions of naled areas occur in this same range (see Figure 3f). The areas in the 0.05-0.1 km^2 and 0.3-0.5 km^2 ranges predominate. The large number (22 percent) of small naleds (smaller than 0.05 km^2) caused a relative naled density of 0.66 percent, which is low for a seismically active region. The abrupt decrease in the number and sizes of naleds is noted at a tapering-out of the loose deposits.

On the plateau basalts, the naleds occur at the 1,000-1,800-m elevation. In the altitude range from 1,200m to 1,400 m, we find the maximal occurrence of naleds (see Figure 2g). The altitude of 1,400 m, where the medium areas of naleds occur is maximum for the Udokanskaya upthrust (see Figure 3g). The distribution curves of naled frequency by sizes testifies to the predominance of an area from 0.02 km^2 to 0.1 km^2. The relative naled density is also low, i.e., 0.33 percent. Within the limits of the volcanic cover, individual naleds are forming in the zone of post-low-Quaternary faulting with a northeastern strike. Improved conditions of supply to the naleds are noted at the contact of basalts with granitoids.

In the central Udokanskiy region with a watershed at 2,000-2,500 m, the naleds occur at heights from 800 m to 1,800 m. The distribution curve by height is asymmetrical: maximum at the range from 1,350 m to 1,550 m and the upper branch is steeper (see Figure 2h). The distribution curve of frequency and size is irregular, elongated along the entire *X* axis; the areas in the 0.05-0.2 km^2, 0.4-0.5 km^2, and 1.3-2.0 km^2 ranges predominate. Relative naled density is 0.49 percent. The naleds occur chiefly at the boundary of the high upthrust horst with other rift structures.

In the eastern Udokanskiy region, the water divide is located at the 2,200-2,480-m mark, while the naleds are found at 800-1,400 m altitude. Below 800 m, one naled has been recorded. The distribution curve is symmetrical and has a maximum in the 950-1,150-m range (see Figure 2i). Average dimensions of the naleds' area occur at 1,100 m elevation. The areas ranging in size from 0.07 km^2 to 0.2 km^2 predominate (see Figure 3i). The increase in relative naled density to 0.75 percent can be explained by the

TABLE 1 Relative Naled Density in Central Part of Olekma-Vitimskaya Mountainous Country

Region	Area, lcm^2	Number of Naleds	Area of Naleds, km^2	Relative Naled Density, %
Patomskiy	4,310	44	9.31	0.22
Nichatskiy	3,290	69	19.40	0.59
Western Kodarskiy	3,040	72	34.04	1.12
Eastern Kodarksiy	1,980	54	11.64	0.58
Upper Charskava depression	2,300	51	34.72	1.38
Western Udokanskiy	6,190	80	40.54	0.66
Plateau basalts	2,280	56	7.46	0.33
Central Udokanskiy	4,910	46	24.22	0.49
Eastern Udokanskiy	2,670	38	20.06	0.75
Upper Kalarskaya depression	790	8	8.20	1.21
Southeastern area	3,220	19	0.45	0.01
Entire region	34,980	537	207.04	0.59

contemporary activation of faults. The deterioration in the conditions of supply to naleds in a direction toward southeasterly zone of orogeny outside of the rift determines the steepness in the lower branch in the frequency curve by heights.

The configurations in the distribution curves for the naleds for the regions of the Udokanskiy block-type upthrust have much in common. The differences in details have been caused by the elevation of lowered blocks, geologic structures, and varying seismic activity of regions. The naleds do not form on the slightly separated watersheds of Udokan Range, since the subsurface water discharges into the zones of faults intersecting the deep river valleys.

In the upper Kalarskaya depressions, the elevation for the bottom comprise 1,200-1,350 m in the center and 1,600 m on the periphery. Five naleds out of eight occur at altitudes of 1,300-1,350 m (see Figure 2j). The increase in naleds' areas to 0.5 km^2 in proportion to the rise in level of elevation can be explained by the percentages of participation of snow in the formation of naleds in the high-mountain depressions. The areas from 0.3 km^2 to 0.4 km^2 (see Figure 3j) predominate. The relative naled density is 1.21 percent. Taliks have become developed in the depression's central part; hence, naleds do not form.

In the southeastern area, we have counted 19 naleds; their area does not exceed 0.08 km^2. Although certain places in the watershed rise to 2,000 m, naleds do occur below 1,250 m down to 850 m. Their maximum number develop at the 1,000-1,150-m elevation (see Figure 2k). The increase in average areas in proportion to rise in elevation (see Figure 3k) is explained by the participation of snow in feeding the naleds. The relative naled density is 0.01 percent.

CONCLUSIONS

1. The maximum number of naleds is localized in the rift zone and is manifested at the intersection points of the northeasterly rift-forming fault with the sublatitudinal ones, controlling the epicenters of contemporary earthquakes.
2. The very largest naleds occur in the deep depressions and in the limits of upthrusts, characterized by maximal amplitudes of displacements during the Cenozoic time.
3. An abrupt decrease in the number and sizes of naleds, evoked by a decrease in fissuring of rocks, occurs outside the rift zones.
4. According to the nature of distribution curves (see Figures 2 and 3), we record two types of naleds: those fed by springs and the mixed type, with an appreciable participation by glaciers and snow.

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CERTAIN CRITERIA FOR ESTIMATING THE SUBSURFACE WATER SUPPLIES IN SUB-ALASSAL TALIKS (EXEMPLIFIED IN CENTRAL YAKUTIA)

V. D. SHCHEGLOV AND YE. M. DMITRIYEV   Yakutsk Territorial Geologic Administration

Among the basic factors complicating the utilization of subsurface water for water supply in central Yakutia, we include the permafrost, the anomalously low [groundwater] levels over a significant part of the area, and the composition of subpermafrost water, which is not always favorable.

In recent years, in the process of hydrogeologic studies conducted by the Yakutsk Geologic Administration, we discovered the possibility of utilizing the water from talik zones[4] for water supplies.

The talik waters form a special group of subsurface water typical for the regions where permafrost develops. Diversified in form and size, the flooded zones of thawed materials form in a layer of frozen deposits, usually under the warming effect of the surface reservoirs and with the direct participation of the latter in the formation of water from the talik horizons.

The extensive distribution, within the limits of central Yakutia, of thermokarst and erosion-thermokarst lacustrine depressions (alasses) and the taliks associated with them, creates favorable assumptions for solving the problem of water supply to the numerous farms, which are relatively small and scattered over a wide territory. In connection with this, at all stages of the hydrogeologic investigations of a region, the question arises concerning an estimation of the natural supplies, the natural resources, and operational reserves of water-bearing horizons in the talik zones.

The essential criteria for evaluating the stocks of talik water are as follows:

1. The volumes and forms of talik zones, as a whole, and of potential horizons within them, in particular.

The volumetric parameters of taliks assume primary importance since cases are frequent when it is necessary to be oriented only toward the development of natural reserves and sometimes only in regard to a certain part of the water reserves that is suitable for utilization.

2. The permeability and capacity of the talik aquifer.

The feature of these characteristics in reference to the talik water can be described by citing the face that, in addition to the indexes for the permeability, there is an increase in the value of the reserve supply of the water-bearing horizons.

3. Boundary conditions determining the regime of the talik water.

The permafrost containing the talik zones can in no way be regarded as a simple analog of a lithologic water barrier. The close interrelationship of the frozen rocks and of the water-bearing horizons is manifested in the dynamics of the talik's outline, variation in chemical composition of talik water during the freeze-thaw cycles, the formation of cryogenous pressures in the taliks, etc. In this manner, the permafrost constitutes a factor complicating the boundary conditions. We must have a precise and unambiguous definition of this factor, otherwise an estimate of the operating reserves of talik water would be impossible.

4. Mineralization and composition of talik water.

One of the features involved in the talik water is the chaotic nature of its chemical composition, occasioned by the isolated state of the taliks from each other and by the diversity in the composition of the lake water with which they are closely interrelated.

The significance and importance of studying the stated criteria at the stages of a hydrogeologic survey and exploration for talik water are dissimilar.

The regional tendencies in the distribution of talik zones are shown by small-scale hydrogeologic surveys and by the hydrogeologic zoning based on standardizing the taliks and on clarifying the permafrost-hydrogeologic and physicogeographic conditions in a region. Since the investigation of many talik zones isolated from each other is a laborious task, primary significance at this stage of research is imparted to revealing the correlations in the taliks' parameters with the physicogeographic and permafrost-geologic indicators. The close interrelationship and interdependency of surface and talik waters serves as a cause for a number of specific features inherent only to talik water; these features must be taken into account when assessing them as a possible water-supply source. However, the hydrogeologic studies permit us to find certain parameters of the talik horizons based on results from analyzing the surface conditions: namely the sizes of lake and lake basins, intensity of water circulation in lakes, erosion-type dissection of land, etc.

As an example of the determination of such correlations, we review below the relationship between area and thickness of taliks and area of alas basins to which they are connected and the relationship between areas of alas-type basins within the limits of the morphologic area and the intensity of its erosion-type dissection.

A rough estimate of the supplies of talik water at the stage of small-scale hydrogeologic surveying is given by means of calculations of the moduli of their natural supplies based on the segregated morphogenetic sectors (regions). The calculations are made with the equation:

$$M_e = K_T \cdot m \cdot \mu \cdot 10^4, \quad (1)$$

where M_e = modulus of natural reserves, m^3/km^2; K_T = coefficient of talik density (percengate of taliks' area within the limits of a morphologic sector to the sector area); μ = water yield of rocks forming the taliks; and m = thickness of flooded part of talik, m.

The dimensions of the water-bearing horizons of talik water and the water yield at the survey stage are determined from test areas (springs) [and] according to results obtained from drilling and test operations.

The determination of the talik density coefficients was fairly uncertain until recently. The areas of taliks were tentatively equated to the areas of alas basins. This difficulty could be overcome by establishing a correlation between the taliks' area and area of alas basins to which they are coordinated. A statistical processing of data for 56 taliks within the limits of the Lena-Amginskoye interfluve (Figure 1) indicated that these values are related fairly closely (correlation coefficient = 0.93) by the equation:

$$S_T = 0.64 \cdot S_a. \quad (2)$$

where S_T = talik area, km^2; S_a = areas of alas basin, km^2.

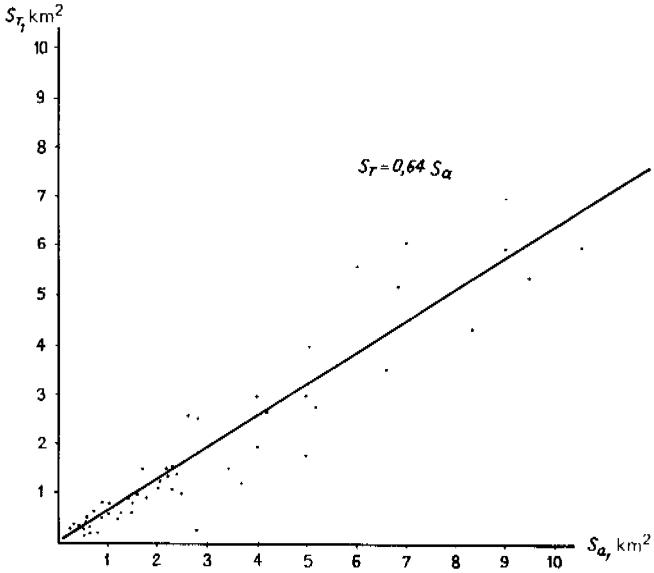

FIGURE 1 Curve showing taliks' area (S_T) versus area of alas basins (S_a).

The error [*sic*]* in the regression equation is 0.9 km^2.

In this manner the possibility appears of identifying the area of taliks within the limits of the morphologic sectors, which have been differentiated by measuring the areas of the alas basins from cartographic materials and subsequent calculations based on Equation (2). However, for a large number of alas basins, this technique appears too laborious, since it involves measuring each actual basin. In connection with this, for finding the area of the alas basins the suggestion is made to use the dependence of the alas density factor upon intensity of the area's erosion-type dissection. The alas density factor is the percentage of the area of alas basin to the area of sector. Expressing the intensity of erosion-type dissection based on existing techniques of numbers on a scale,[7] we can derive empirical equations quantitatively linking the specified values.[3,5] Within the limits of central Yakutia, for the terraces on Lena River, an equation has been derived as a result of the statistical processing of data on a computer from 272 determinations. The correlation coefficient is 0.88; this indicates a fairly close link between the coefficient of alas density and density of erosion dissection (Figure 2). The regression equation is:

$$K_a = 661/R^{1.35}, \qquad (3)$$

where K_a = alas coefficient and R = erosion dissection intensity in points on the scale.

The error [*sic*] in the regression equation amounts to 1.5 percent.

*The term used by the author is not that used for "standard deviation," which one would expect here.

In Figure 2, in addition to the curve of K_a and R, we have shown the curve of K_T and R, constructed from Equation (2). Transforming Equation (1), with the aid of Equations (2) and (3), we derive the equation for the modulus of talik water's natural reserves directly from erosion dissection intensity within the area:

$$M_e = \frac{4.23 \cdot m \cdot \mu \cdot 10^6}{R^{1.35}} .$$

Using Equations (1), (2), and (3) and the charts reflecting the intensity of erosion dissection, we have constructed a system for the distribution of the moduli of talik water's natural reserves in the alluvial beds of terraces along the Lena River and its middle course (Figure 3). The original data on thickness and water yield of Quaternary beds are listed in Table 1. In constructing the system, we assumed that in the entire area of taliks' occurrence, their thickness exceeds 100 m, i.e., the alluvial beds within the confines of the talik zones' contours exist in a thawed state and are flooded with water. Extensive data acquired from surveying and exploratory operations within the boundaries of central Yakutia permit us to make a similar assumption, since, in the determination of all the relationships presented, we adopted in the calculation the alas basins of not less than 300 m in diameter (3 mm on a map to a scale of 1:100,000), while as a rule the taliks' thicknesses comprise not less than one-third of their diameters.

Intensity of erosion dissection is given in the system in following ranges: 0-20, 20-40, 40-60, 60-80, and 80-100 points. The moduli of

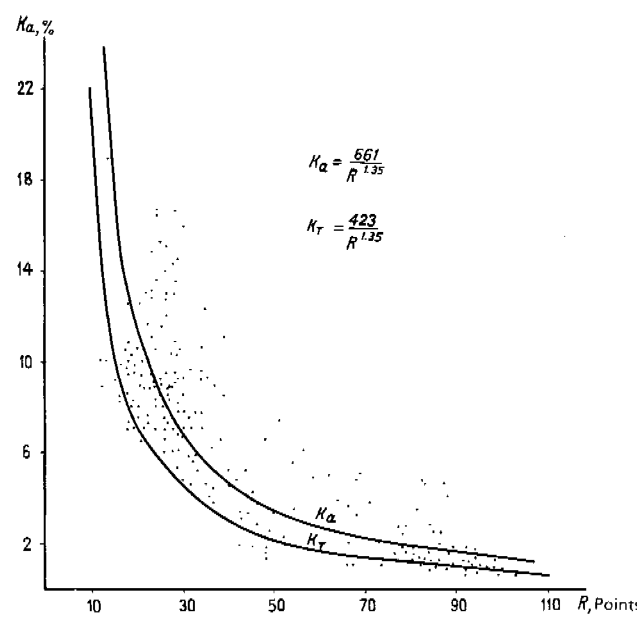

FIGURE 2 Curves indicating dependence of alas density coefficient (K_a) and talik density (K_T) on intensity of erosion dissection of terraces on Lena River (R).

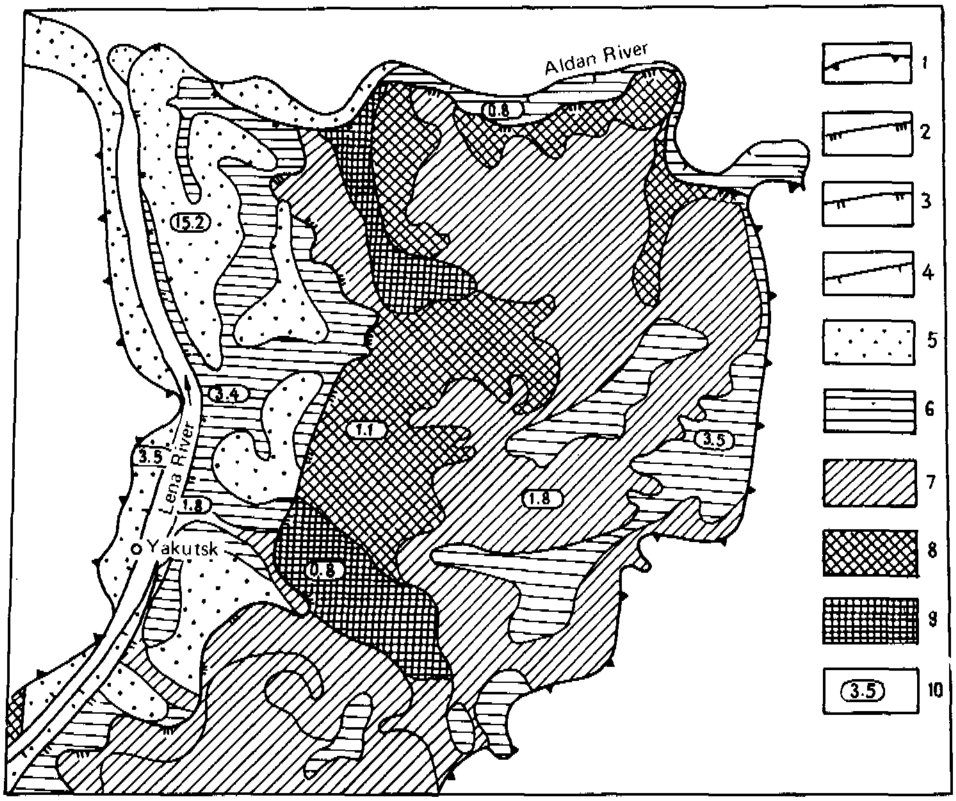

FIGURE 3 Diagram indicating intensity of erosion dissection of terraces along the Lena River, and distribution of moduli for natural reserves of talik water in the alluvial beds. *Boundaries of morphologic sectors:* 1--denudation plateau; 2--grouping of high terraces; 3--grouping of terraces of medium height; 4--grouping of low terraces. *Intensity of erosion dissection of terraces by number:* 5--less than 20; 6--20-40; 7--40-60; 8--60-80; 9--80-100; and 10--moduli reflecting natural reserves of talik water in *n* × 10^5 m^3/km^2. (Complied by V. D. Shcheglov.)

the natural supplies of talik water were calculated for sectors based on each step in the range, where the values reflecting extent of dissection *(R)* were taken along the center of each step, i.e., 10, 30, 50, 70, and 90 points. Within the limits of each morphologic sector delineated in the system (high terraces, medium-height terraces, etc.), the computed value for the modulus of talik water's natural reserves is identical for the same step in intensity of erosion dissection.

The impossibility of conducting (at the small-scale mapping stage) a broad hydrochemical testing of the talik water compels us to indicate its qualities based on results obtained from testing the surface water, with consideration of water circulation rate, stages in development of thermokarst basins, and so forth. This problem is subject to further treatment, since at present it does not have a clearcut solution.

The boundary conditions applying to the talik horizons at the surveying stage are established from analyzing the general permafrost-hydrogeologic conditions of the region with allowance for the thickness of permafrost, type of taliks (through, blind), the occurrence depths of subpermafrost water levels, presence of a lake in the thermokarst basin, etc.

In the exploratory stages, we conduct hydrogeologic studies of promising taliks disclosed by the survey. During the exploration using geophysical techniques (electric geophysical exploration), we refine the contours for the near-alas talik in plan and obtain a more precise idea of its thickness, type, lithologic-facies structure, and talik water's mineraliza-

TABLE 1 Largest Lakes in the Cryolithic Zone of the USSR

Terraces	Loams, Sandy Loams		Sands		Gravels				
	$\bar{m}$	$\mu_{weighted}$	$\bar{m}$	$\mu_{weighted}$	$\bar{m}$	$\mu_{weighted}$	$\bar{m}$	$\mu_{weighted}$	$\bar{m}_{av}$, $\mu_{weighted}$
Low	14	0.05	5	0.15	1	0.3	20	0.09	1,80
Medium	25	0.05	45	0.14	3	0.2	73	0.11	8,03
High	70	0.05	30	0.14	3	0.2	103	0.08	8,24

tion. The scales of electric profiling and vertical electric sounding are determined by the dimensions and complexity in the talik zones' structure. Examples are given in Figure 4 of the allocation of electric exploration operations depending on the talik's sizes and type. The main problem in hydrogeologic research at the exploration stage consists in refining the parameters necessary for assessing the talik water reserves, which have been roughly established at the preceding stage.

The chief task in hydrogeologic investigations at the exploration stage is to provide valid calculations of the operating reserves of talik water, an efficient type of reservoir, and the conditions for its operation. This problem is solved by a combination of tasks, including electric geophysical exploration, drilling, pumping from the boreholes, and systematic observations. The purpose of the tasks is to choose a design system for estimating the operating reserves and to obtain experimentally the reliable hydrogeologic parameters entering the calculations.[2] In connection with this, in the final stage special attention is directed to an exact determination of the boundary conditions of the talik water's horizons.

We established that, in plan, the talik rocks are bounded by an impenetrable boundary of permafrost. Certain variations in the boundary conditions are possible. In the talik's upper boundary, two types of conditions can exist, de-

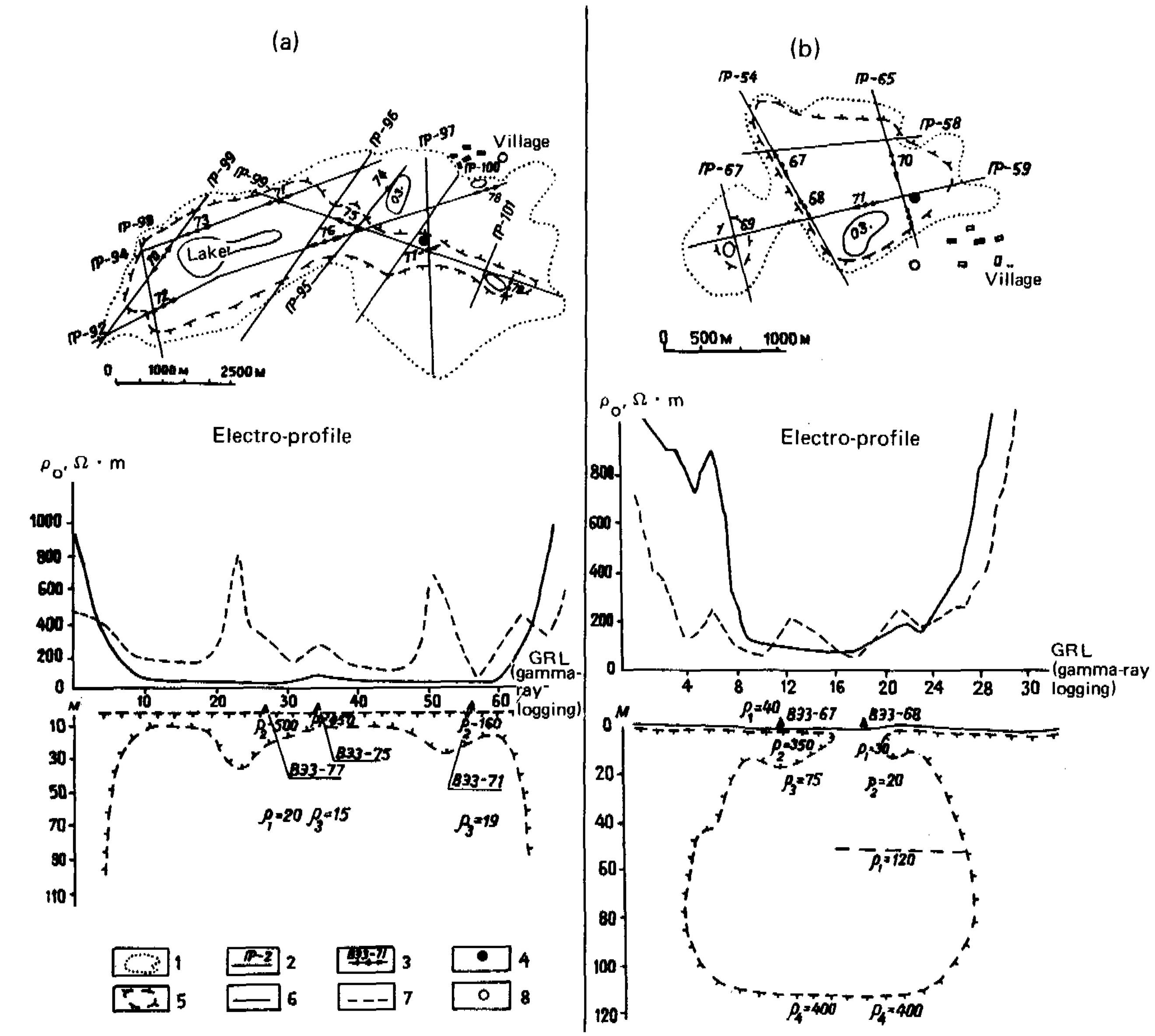

FIGURE 4 Arrangement of electro-profiles, VES (Vertical Electric Sounding) points, and drilled boreholes in exploratory operations for talik water. (A)--through-talik zone; (B) blind-talik zone. 1--contour of thermokarst basin (alas); 2--electrical geophysical prospecting profile and its number; 3--VES point; 4--location of the hydrogeologic borehole; 5--limit of thawed and frozen rock; 6--curve reflecting apparent specific resistances measured with "large" feed line AB-300m; 7--curve indicating apparent specific resistances measured with "small" feed line, AB-30m; and 8--location of the geologic borehole.

pending on the presence (in the alas basin) of a lake connected hydraulically with the talik water, or its absence. The amount of the lake water's infiltration forms the most important incoming part of the operating reserves of water from the taliks. In connection with this, at the exploration stage particular attention is diverted to studying the balance and regime of lakes in thermokarst basins that are regarded as plains where constant pressure exists. The upper limit may be without a water barrier (talik water lacks pressure) or with a water barrier formed by the frozen rocks of a "deflector" (the talik water is of the pressure type). At the talik zones' lower boundary, the conditions depend primarily on the type of talik, i.e., through or not through.

In the case of a blind talik, the nature of the conditions is determined by the impermeable contour of the permafrost. Two cases are possible for the through-taliks. In the first, the role of the impermeable contour is played by the regional lithologic water barrier (or a local one, exceeding in area the talik's contour). In the case of an established hydraulic link between the talik and subpermafrost water, the nature of the boundary conditions is determined by the ratio of their pressures. Thus for the central part of the Lena-Vilyuy artesian basin, where the piezometric surface of the subpermafrost water-bearing horizons is sometimes located below sea level, replenishment of subpermafrost water supplies occurs through overflow from the talik beds. In the basin's peripheral parts, where the subpermafrost water levels are above the depth of the taliks' water-bearing horizons, the reverse pattern is possible, namely the replenishment of talik water supplies from the subpermafrost water.

In this way a valid estimation of talik water supplies is possible only as a result of analyzing all the sources of their income and expenditure.

Exploration for talik water is accompanied by detailed hydrochemical testing. In this procedure, we consider the heterogeneity of the water's composition in plan and in section of the talik, as well as the features involved in its formation.[1] We determine simultaneously the content of harmful components in the water. With consideration of the data presented above, in each specific instance we solve the problem concerning the referral of the calculated operating supplies to any given group (balance, unbalance) and recommendations are issued concerning the feasibility of treating the water (softening it, decontaminating it, removing the iron, etc.).[6]

The results of the hydrogeologic studies conducted in central Yakutia permit us to form a conclusion about the possibility, if not of a complete, at least of a partial, solution to the problem of furnishing water from the vast plains areas in the permafrost zone, through tapping the water-bearing horizons of the taliks of the thermokarst basins.

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HYDROCHEMICAL ZONING OF SUBSURFACE WATER AS ONE OF THE INDICATORS OF PALEOPERMAFROST CONDITIONS

R. S. KONONOVA   Permafrost Institute at Siberian Branch, USSR Academy of Sciences

Until the present time, the dynamics involved in the development of the frozen zone and the range in its individual parameters have been little studied. The chief reason for this is the difficulty involved in recreating the frozen zone's developmental history, because after its disappearance it leaves considerably fewer traces, than, for example, glaciers. Its restoration is based on the paleoclimatic and morphological, i.e., external, factors of its formation and manifestation. However, the internal geologic-lithologic and especially the hydrogeologic factors as indexes of permafrost conditions in the past fail to find the proper reflection in the paleopermafrost structures.

Among the hydrogeologic factors, the most sensitive and reliable indicators of a variation in the frozen zone's parameters can be the features in the hydrochemical profiles reflecting the zonation under various geologic structural conditions caused by the cryogenous metamorphism of the water's composition.

Information at hand in the literature concerning the influence of the cryogenous factors on the natural water's chemical composition and mineralization is rather disorganized and scanty. Most of the known reports describe the chemical or physicochemical processes occurring during the annual freezing and thawing of groundwater, water in the seasonally thawing layer, and the surface, chiefly lake,[2,3,6] or talik, water.[1] Quite interesting are the observations of the Soviet and foreign researchers on the effect of freezing on the variation in composition of marine mineralized water.[4,5,7,8,10] The basic concepts ensuing from these reports can be summarized briefly as follows:

1. The freezing either of fresh or of saline seawater leads to the formation of a solid phase, i.e., of ice; its mineralization is always several times lower than that of the original water subjected to freezing; freezing also leads to an increase in the mineralization of the remaining solution owing to the expulsion of salts on freezing.
2. The amount of salts involved in the ice phase depends on their concentration in the solution, on conditions and rate of movement of the freezing liquid, and rate of ice formation. Slow gradual freezing leads to the formation of the least mineralized ice.
3. Between the ice and the freezing solution, there occurs a redistribution of the salts. Salts are drawn selectively into the ice. Simultaneously with the ice, compounds are separated in conformity with the temperatures at which total precipitation of salts from the solution occurs. It is primarily the least soluble calcium carbonates that precipitate in the fresh water, while in the more mineralized water it is mainly the sodium and calcium sulfates that precipitate. Freezing leads to an increase in mineralization, i.e., to a "cryogenous concentration" owing to the compounds that are most mobile and easily soluble at low temperatures; these include the sodium hydrocarbonates and the calcium, magnesium, and sodium chlorides.
4. On subsequent thawing, the carbonate compounds of calcium and magnesium and the sodium and calcium sulfates do not convert completely to a solution, by the same token reducing the mineralization of the thawing solution ("cryogenous freshening") and modifying its composition in relation to the composition that preceded the freezing.
5. The alternating freeze-thaw cycles of the fresh calcium hydrocarbonate water as a whole through the stage of magnesium hydrocarbonate water lead to the formation of less mineralized sodium hydrocarbonate water. The freezing of saline seawater leads to an appreciable increase in its composition of calcium chlorides owing to crystallization and precipitation of sodium sulfates, while at low temperatures (below -23°C), freezing also leads to the precipitation of sodium chlorides.

Hence as a result of the cryogenous processes (freezing and thawing), both the mineralization and composition of natural water change.

The concepts expressed and an analysis of the materials on the hydrogeologic conditions in the northeastern part of the Siberian Platform permitted us to reveal (within the territorial limits of continuous development of the frozen zone) the types of hydrochemical sections and to pinpoint their features, occasioned by the periodic fluctuations in thicknesses of the frozen zone.

The first type of profile is typical for the terrestrial continental and freshened littoral-marine beds in the Cenozoic and Mesozoic with a thickness to 2-3 km, supported by marine terrestrial beds in the Mesozoic and upper Paleozoic, and described in the central part of the Lena-Vilyuy artesian basin. The frozen zone attains 650 m; beneath this, a successive alternation of fresh, saline, and brine water is preserved. For this type of section, low piezometric levels of subpermafrost water and conditions of high enclosure are normal. From hydrochemical standpoints, a number of specific features are observed:

1. Low mineralization of subsurface water (0.3-0.4 g/l) at depths to 1,000-2,000 m; accordingly, the lower limit of the freshwater zone descends deeply.
2. A considerable increase in mineralization of subsurface water (by 3-4 times) in specific intervals of the profile at depths greater than 1,000 m (e.g., for the Manskaya borehole, in the 2,100-2,300 m range, see Figure 1a) with subsequent abrupt reduction and then a gradual increase with depth.
3. The predominance of sodic water at low contents of calcium and magnesium. For a long

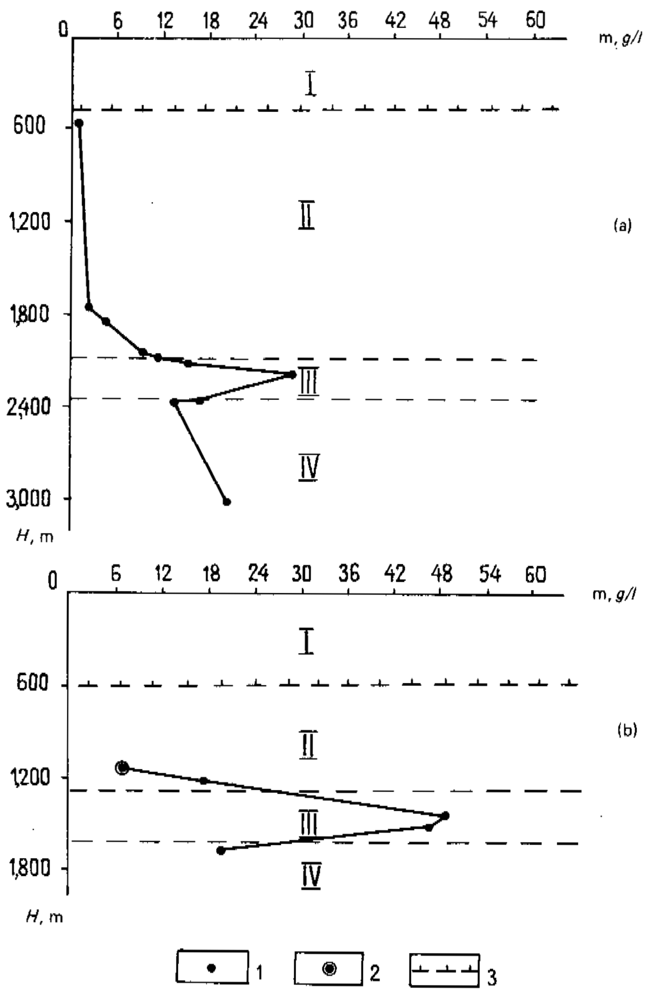

FIGURE 1 Variation in mineralization of subpermafrost water with depth. (a) In terrestrial-continental and marine beds along Namskaya reference borehole. (b) In terrestrial marine beds along Dzhardzhanskaya (1) and Ulakhan-Yuryakhskaya (2) boreholes; 1, 2-- points of water samples; 3--lower limit of frozen zone; I--present frozen zone; II--freshening zone; III--zone of concentration; and IV--zone of water unchanged by cryogenesis.

this remained unexplained, but it is becoming understandable from the standpoints of the effect of cryogenesis on the hydrochemical zonation that formed prior to freezing.

The deep primary freezing during the cold glacial epochs was accompanied, on the one hand, by the formation of slightly mineralized ice and, on the other hand, by the relative increase in overall mineralization of water at the lower freezing limit and by the formation of unique concentration zones. During freezing a large share of the calcium hydrocarbonates precipitated from the solutions, and the accumulation of the more soluble sodium hydrocarbonates occurred. The ice-thawing that then followed caused, owing to degradation of the frozen zone, an additional freshening of the thawing water-bearing horizons; this also led to the propagation, to a considerable depth, of the slightly mineralized subpermafrost water with predominance of sodium and to an extensive development of cryogenous calcitization of water-bearing rocks. As a result, in the subpermafrost horizons of the type of profile under discussion, under the cryogenous processes' effect, there formed a unique zoning. Within the limits of this zoning, zones of freshened, concentrated--and still lower down--of water unaltered by cryogenesis, are differentiated. The thickness of the freshened water zone for the Manskaya borehole reaches about 1,800 m, increasing toward the central parts of the Lena-Vilyuy basin, and indicating that the frozen zone's thickness in the glacial epochs was much greater than at the present time.

The second type of profile characterizes the marine terrestrial or carbonate beds and is described in the Khatanga and Lena-Vilyuy basins. The frozen zone's thickness runs from 300 m to 500 m, less often from 100 m to 200 m. From hydrochemical standpoints, the features of this type of profile are:

1. Absence, in the regional plan, of freshwater subpermafrost horizons.
2. Low mineralization of upper subpermafrost horizons, anomalous for water in marine deposits (up to 8-12 g/l).
3. An abrupt increase in mineralization by 4-6 times at depths greater than 1,400 m (see Figure 1b).
4. A chiefly magnesium-sodium hydrocarbonate or sulfate composition of water in the carbonate beds of the Cambrian period.

These features can also be explained by the cryogenous metamorphization of water in the profile upper part; its mechanism has been reviewed above. In this type of profile, the desalting of the water-bearing horizons can be traced to depths of around 1,400 m, becoming replaced by clearly expressed zones of cryogenous concentration.

A third type of profile has been described in the central part of the Siberian Platform on the southern slope of the Anabarskaya Anticline and is represented by salt- and gypsum-bearing rocks of the lower and middle Paleozoic, interrupted by kimberlites. Here the thickness of the frozen zone varies between 150 m and 400 m, comprising only a part of the negative temperature belt attaining 1,500 m. For this type of section we typically find: presence of weak brines just below the frozen zone; composition in the limestones is chiefly calcium chloride; in the kimberlites, it is mainly magnesium chloride; a clearly expressed zoning of secondary mineral formation expressed in the successive alteration, from above downward in the section, of calcite, dolomite, and sulfate minerals, while starting from 400-500-m depths, there are individual pockets of rocksalt crystals.

The role of cryogenous processes in the formation of this type of a hydrochemical profile is also quite significant, although their orientation and results are different from the preceding ones. Freezing led to a relative concentration of brines near the surface, to the formation of a specific composition as a result of the precipitation of mirabilite, to the cationic exchange between the brines and the enclosing rocks, and to the appearance of secondary cryogenous minerals in the enclosing rocks. Owing to the presence of weak brines at relatively slight depths, the thickness of the negative temperature zone greatly exceeds that of the frozen zone;[9] as a result, the fluctuation in the latter could scarcely be significant.

The differentiation of the three described types of hydrochemical profiles over the area of the Siberian Platform within the limits of the frozen zone, the specific features of these sections expressed in the chemical composition of the subsurface water, and the presence of desalted and concentrated solutions caused by cryogenesis provide the possibility of indirectly judging the ranges of fluctuations in the frozen zone's thickness in the Quaternary period for each of the types of sections considered. The maximal thickness of the frozen zone and its maximal fluctuations occur in the sections formed by continental deposits and by beds in the littoral-marine-freshened basins. For them, the thickness of the frozen zone greatly exceeded 1,000 m. Somewhat reduced thicknesses and smaller amplitudes are inherent to the profiles in the marine carbonate and terrestrial deposits. At a significant value for the thickness of the negative temperature belt, the least thickness of frozen zone occurs in the salt- and gypsum-bearing complexes in the Paleozoic period.

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CONCERNING CERTAIN CHEMICAL PROCESSES OCCURRING IN CLAYEY AND SANDY SOILS AT NEGATIVE TEMPERATURE

N. N. LAPINA   Scientific Research Institute of Arctic Geology

A study of the chemical processes occurring at negative temperature in clayey soils was conducted with the aid of determining the chemical composition of water extracts obtained prior to and after the freezing of clays having a varying mineral composition (kaolinitic, hydromicaceous, and montmorillonitic).

A study of the chemical processes occurring in sands with an additive of clay having a varying mineral composition was conducted by comparing the chemical composition of filtrates obtained after a determination of the coefficient of permeability of sands at negative temperature with the chemical composition of the original solution.

In a study of the water extracts obtained from clays by the conventional method (shaking together with 5 times the volume of water) at positive temperature of sample and after freezing for 24 h, we obtained the results shown in Figure 1 and in Table 1.

In the kaolinitic clay, the total of water-soluble salts after freezing the clay increases. In the composition of the cations, there is an abrupt rise in Na content, while in the composition of Cl anions and several SO_4 anions, the amount of Ca decreases. The very same pattern also occurs in a determination of the chemical composition of a water extract of montmorillonite clay. The composition of water extracts made of hydromica clay yields different results. The total of easily soluble salts in the extract decreases slightly on freezing. In the soluble part there is a significant increase in the Na and Cl content; the quantity of Ca, Mg, SO_4, and HCO_3 drops sharply.

In this manner, in the clays of varying mineral composition, at the time of freezing, chemical processes occur as a result of which the sodium chloride and in part the sulfate salts acquire a greater mobility and are attracted more easily to the water extract. This process can be traced particularly clearly in the montmorillonitic clay; hence in spring, immediately after the active layer's thawing, from the ground and soil a greater amount of sodic salts is borne out than during the summer, when chemical equilibrium has already set in between water circulation and the soil.

The form of particles of kaolinite and mont-

TABLE 1 Results of Analyzing Water Extracts of Clays (in Percent per 100 g of Dry Batch)

Components	Kaolinite Clay		Montmorillonite Clay		Hydromica Clay	
	Before Freezing	After Freezing	Before Freezing	After Freezing	Before Freezing	After Freezing
HCO_3	0.0488	0.0488	0.1035	0.3076	0.0854	0.0549
Cl	0.0071	0.0335	0.6122	0.8869	0.0124	0.0670
SO_4	0.0348	0.0576	0.3605	1.0261	0.0674	0.0134
Ca	0.0160	0.0100	0.0497	0.0141	0.0280	0.0018
Mg	0.0022	0.0022	0.0148	0.0026	0.0056	0.0003
Na	0.1083	0.0520	0.5234	1.1611	0.0298	0.0679
Total of salts	0.1275	0.2041	1.6641	3.3984	0.2286	0.2053
pH	7.5	7.5	7.5	7.5	7.5	7.5
H_2O at 105°C	0.80	0.37	11.56	7.5	1.80	1.52

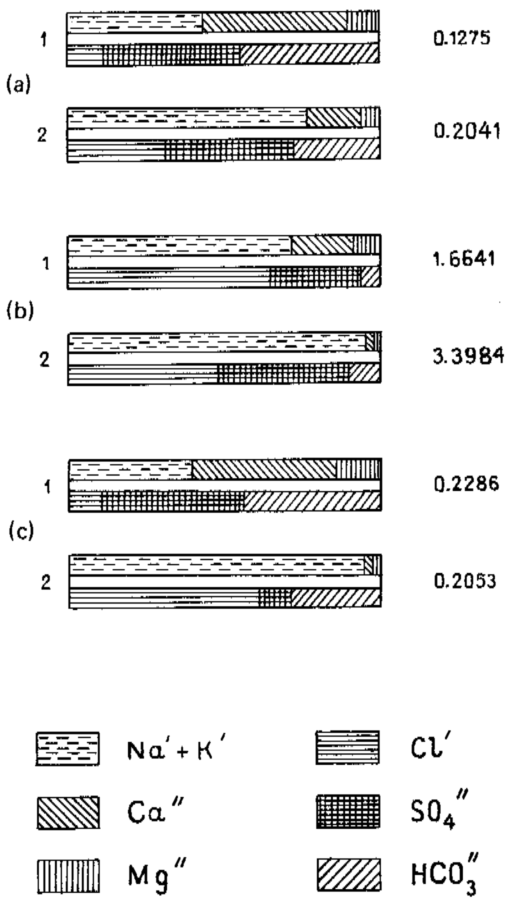

FIGURE 1 Chemical composition of water extracts of kaolinite (a), montmorillonite (b), and hydromica (c) clays: 1--before freezing; 2--after freezing. Numbers on right indicate total of salts, percent per 100 g of dry batch.

morillonite as a result of freezing does not change while the isometrically-lamellar particles of hydromica almost disappear after freezing, and the main mass of the preparation is formed by elongated, distinctly outlined particles with length of about 1 μm.

For a study of the chemical processes occurring during the solutions' filtration through the sandy soils, we took medium-grained quartz sand and added 1 percent by weight of kaolinite clay, 1 percent of hydromica clay, and 1 and 2 percent of montmorillonite clay. As filtering solutions, we took those of sodium and calcium chlorides of varying concentration. We then determined the chemical composition both of the original solutions and of the filtrates obtained at temperature of -1°C and -5°C.

During the filtration of the chloride salt solution through sand with an admixture of clay, the interaction between soil and solution also occurs at a negative temperature (Figure 2).

During filtration of the sodium chloride solution through sand with an admixture of 1 percent of kaolinite clay, the chemical composition of the solution changes. In the solution, there is an increase in the calcium content and in part of the magnesium; the dry residue increases. At filtration through sand with an addition of 2 percent of montmorillonitic clay, the dry residue increases appreciably owing to the increase in content of calcium and magnesium. With an increase in the concentration of the sodium chloride filtering solution, the orientation of the process is retained, and there passes into the solution an ever-increasing absolute amount of calcium and magnesium, especially in the sand with an admixture of montmorillonite; however, the dry residue decreases, i.e., the filtrate is somewhat "freshened."

During the filtration of calcium chloride, the pattern is retained; at slight concentrations of the solution, there occurs an increase in dry residue in the filtrate as compared with the original solution. At an increase in the original solution's concentration, the filtrate is desalted. The chemical composition of the filtrates also varies, but at this time the variations depend on the mineral composition of the clay present in the sand (see Table 1).

The data obtained provide evidence that at negative temperature on interaction of soil with water or solutions, the exchange processes proceed fairly intensively and in isolated cases, uniquely. For example, in a water extract of hydromica clay made at positive temperature, potassium is lacking, while in filtrates of quartz sand with an admixture of the same clay, at negative temperature, the potassium content comprises 648.2 mg/l.

The nature and intensity of the exchange reactions at negative temperature are linked with the phase content of water in soils, which is determined by the mineral and granulometric composition of soils and by temperature. The exchange reactions develop more intensively in the montmorillonite clays with higher exchange capacity and a larger quantity of free water, as compared with the hydromica and kaolinite clays. The effect of temperature on the nature of exchange reactions is manifested most distinctly during investigation of sand filtrates with an admixture of hydromica clay. Thus, at temperatures of -1.5°C to -2°C, potassium passes into solution, but, with a drop in the temperature to -5° to -6°, sodium appears in the solution in lieu of potassium. Below -6°C, the exchange reactions disappear.

The results obtained have practical importance in a solution to a number of questions in engineering geology, especially in identifying the properties of soils subjected to periodic freezing and thawing. However, the scale of studies conducted is quite inadequate for fully describing the specifics involved in the chemical processes occurring at negative temperature.

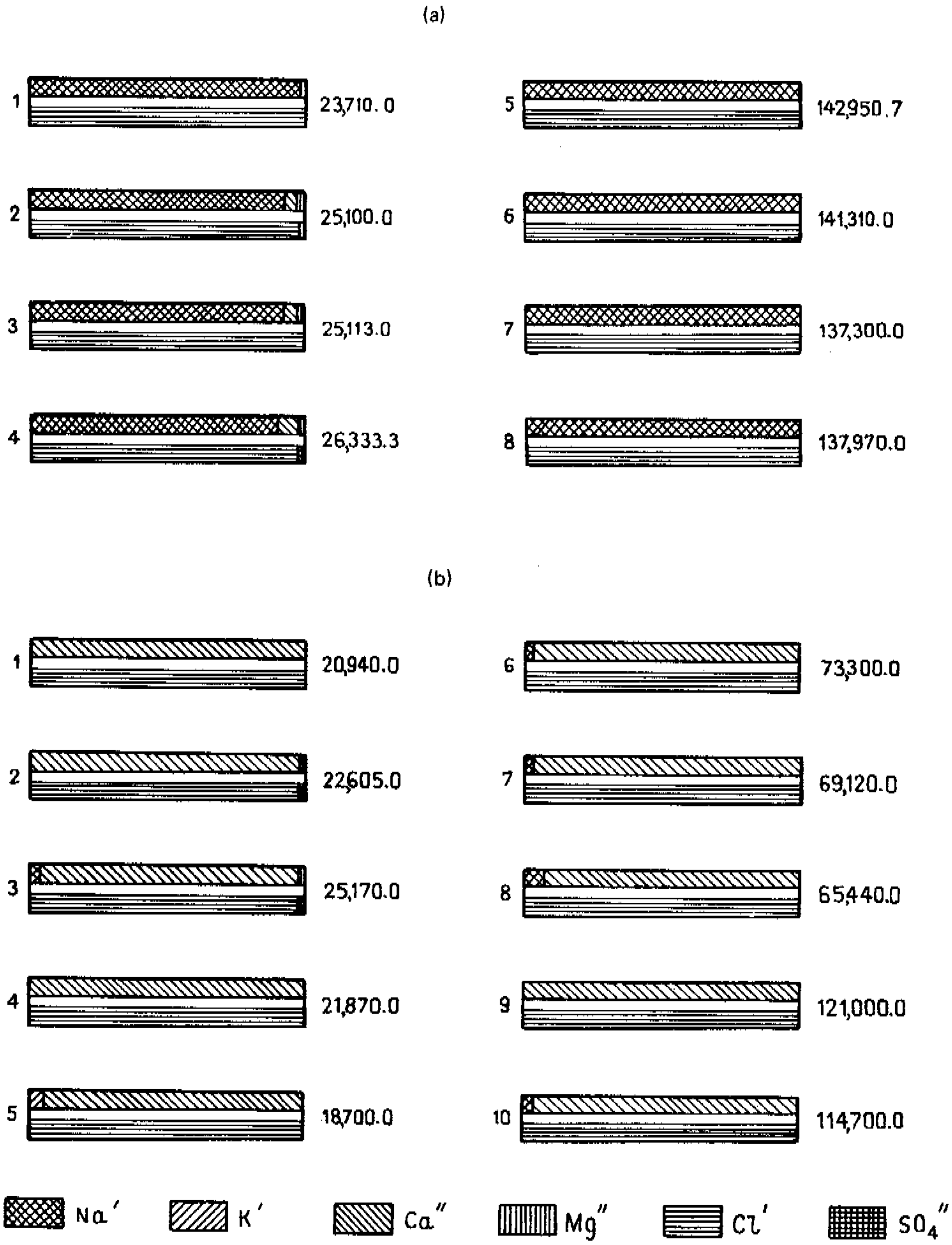

FIGURE 2 Chemical composition of filtrates having passed through sand with admixture of clay. (a) NaCl solution: 1--original solution for filtrates 2-4; 5--original solution for filtrates 6-8 (2,6--sand + 1 percent kaolinite clay; 3, 7--sand + 1 percent montmorillonite clay; 4,8--sand + 2 percent montmorillonite clay); (b) solution $CaCl_2$: 1--original solution for filtrates 2,3; 4--original solution for filtrate 5; 6--original solution for filtrates 7,8; 9--original solution for filtrate 10 (2,7--sand + 1 percent kaolinite clay; 3,8--sand + 2 percent montmorillonite clay; 5,10--sand + 1 percent hydromica clay). Numbers on right = dry residue, mg/l.

CERTAIN TYPES OF GEOELECTRIC SECTIONS OF THE NEGATIVE TEMPERATURE BELT IN THE ARCTIC AND SUBARCTIC IN CONNECTION WITH EXPLORATION FOR SUBPERMAFROST WATER

S. M. LARIN, G. P. MAROV, M. A. KHOLMYANSKIY, AND YA. V. NEIZVESTNOV Scientific-Research Institute of Arctic Geology

In the vast territory of the Arctic and Subarctic, the permafrost zone experiences different stages in its development. Depending on the history of formation during the Holocene and pre-Holocene time, the frozen zone can exist at stages of aggradation and degradation and can bear traces of one or several subterranean glaciations, alternating with periods of relative warming. This is all reflected in the features of the frozen zone's structure and in its physical properties, specifically in the electrical conductance both of the frozen rocks and of the intrapermafrost and subpermafrost horizons. Electrical resistance of the rocks in the frozen zone and of the entire cooling zone as a whole depends primarily on the content of ice in the rocks and on content of various water types: intrapermafrost, subpermafrost, and interpermafrost. The mineralization of this water also plays a major role. In conformity with the structure of the frozen zone and the features of the horizons forming it, we note the following types of geoelectric sections:

First type. Sections of Arctic islands, also typical for certain parts on the continental coast.

In respect to electrical resistances, the permafrost zone of Arctic islands composed of various rocks will be divided into three horizons. The typical VES (vertical electric sounding) curves have been shown in Figure 4.

The upper horizon includes both Quaternary beds and an upper eroded part of underlying rocks. Resistivity of the horizon's frozen rocks varies from 800 to 80,000 Ω·m. The electrical resistance of Quaternary beds as a function of the pocket's ice content is subjected to the maximum fluctuations. For the permafrozen upper Pleistocene moraine clay-loams, the established correlation between ice content and resistance (Figure 1) is expressed by the following equation:

$$i_r = a \lg (\rho/\rho_0),$$

where ρ = resistivity of frozen loams at ice content of pockets equaling i_r (proportion by weight); ρ_0 = the same, at $i_r = 0$; a = correlating coefficient equaling 0.5.

The electrical resistance of the upper part of the pre-Quaternary rocks is more constant (see Table 1).

The thickness of the first horizon varies within wide limits from units to tens and, in some instances, even higher than 100 m.

The second horizon, with resistivity from 500 to 800 Ω · m is confined to the contemporary weathering crust of pre-Quaternary formations to depths of 50-70 m. The cavities in the rocks at the given horizon are almost completely ice-filled.

The third horizon occurring at the foot of the frozen zone is typified by a resistivity from 160 to 400 Ω · m for the fissured magmatic rocks and from 120 to 200 Ω · m for the terrestrial deposits.

In the rocks' fissures in this horizon, the ice pockets are not continuous; the cracks are partly filled with mirabilite and other salts precipitating during the freezing out of the brine. The depth of the horizon's lower limit forming the base of the frozen zone will fluctuate from the first tens of meters on the shore to 200-250 m at several kilometers from the shore. The frozen zone is supported by a zone of cryopegs in which the rocks are saturated by negative-temperature brines and by saline water. In this zone the rocks' resistivity ranges from 100 to 60 Ω · m, decreasing for the highly porous terrestrial deposits to 18-12 Ω · m (see Figure 4, curve 9).

The table shows the values for resistivities of rocks at various horizons in the negative temperature belt on the Arctic islands.

If we trace the geoelectric boundary caused

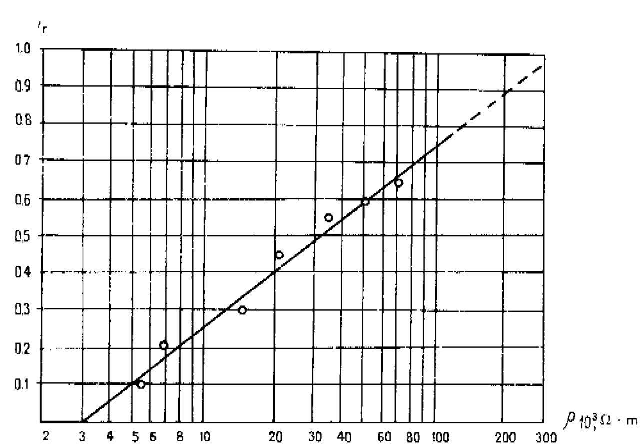

FIGURE 1 Resistivity of suglinoks versus ice content of pockets.

TABLE 1 Resistivity of Rocks

Zone	Horizon	Resistivity [Ω • m] Of Magnetic Rocks	Of Terrestrial Beds
Permafrost	Quaternary beds	--	800-80,000
	Pre-Quaternary rocks		
	First horizon	1,000-4,000	1,000-3,000
	Second horizon	500-800	
	Third horizon	160-400	120-200
Cryopegs	--	60-100	12-100

by the drop in resistivity from 400 to 160 Ω • m to 100 to 12 Ω • m, we can find the depth of cryopegs and thickness of the frozen zone. Figure 2 presents a schematic permafrosthydrogeologic section compiled from results of electric exploration and drilling operations on a number of islands. In this connection we have established an empirical dependence of the frozen zone's thickness on the time of freezing determined by the radiocarbon dating method according to the age of driftwood on the surface of marine terraces (specimens were collected by V. S. Golubkov (see Figure 3).

The second, third, and fourth types of geoelectrical sections are typical of the continental part of the Arctic and Subarctic.

Second type. The permafrost zone consists of unconsolidated Quaternary or Mesozoic carbonate beds; it can also include disintegrated horizons of more ancient depostis. The frozen zone's structure is mostly single-layered with thickness up to 50 m. From a geoelectric standpoint, the frozen zone comprises one horizon of increased resistance. The resistance values are well-differentiated in conformity with the frozen soils' lithologic structure. We present in Figure 4 the typical VES curve (curves 1-3). The lowest resistivities of 200-1,000 Ω • m correspond to the finely-grained varieties. Maximum resistivities (2,000 - 12,000 Ω • m) correspond to the sands and to the boulder-gravel beds. For absolutely all varieties of beds, reduction in their resistances (to the lower limit) is assured by increasing the content of intrapermafrost water, i.e., of the liquid phase of negative-temperature water in the frozen rocks.

The subpermafrost and interpermafrost horizons of the fresh and slightly saline water are typified by specific resistances of about 100 Ω • m. In the presence of interpermafrost water-bearing horizons, the geoelectric section is represented by two or three low-ohmic horizons and by two or three high-resistance horizons. This type of geoelectric section is found in the south of the area where the frozen zone occurs constantly.

Third type. Typical for a region where the development of a trap formation occurs widely. In this zone, there exist in a frozen state the contemporary unconsolidated formations and effusives that make up the upper horizon of high resistance (2,000-30,000 Ω • m). The volume of pores, cracks, and other cavities of effusives comprises the first percents, while the ice filling the cracks does not form continuous interstratifications. As a result of this, in

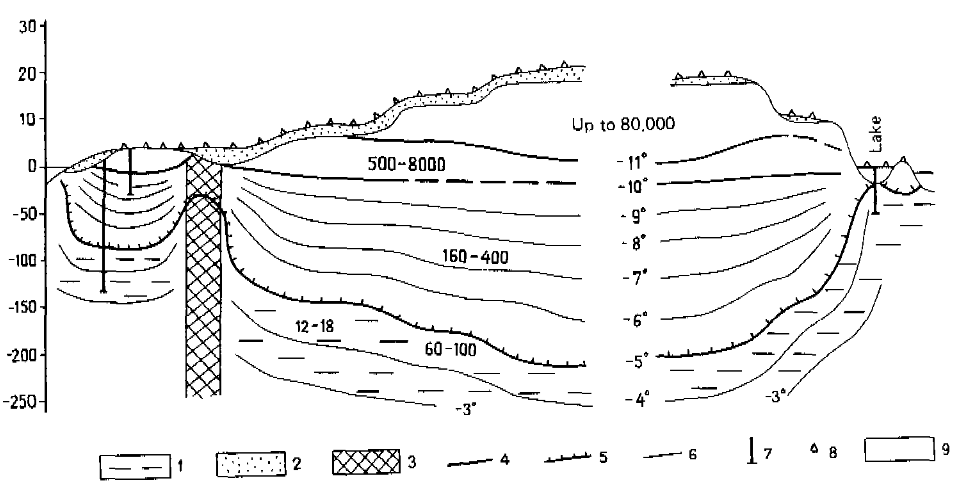

FIGURE 2 Schematic permafrost-hydrogeologic section in littoral part of one of the Arctic islands. 1--cryopegs; 2--Quaternary beds; 3--fault-zones; 4--boundaries of horizons of various electrical resistances; 5--lower boundary of permafrost; 6--isotherms; 7--boreholes; 8--points of the VES; and 9--frozen pre-Quaternary formations.

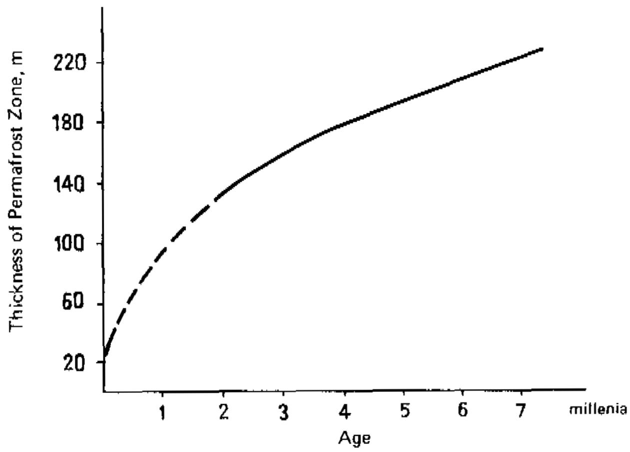

FIGURE 3 Thickness of Permafrost versus age of terraces.

frozen condition the effusives differ little in resistance from the thawed effusives. The frozen unconsolidated mantle beds are characterized by resistivities similar to those in the underlying effusives. In the region under study, the frozen zone's base occurs quite often in the effusives, and, since the contrast between frozen and thawed rocks is slight, it does not appear possible to determine the thickness of the frozen zone based on the vertical electric sounding data. For the third type of geoelectrical sections, the overall form of VES curves is the same as for the second. However, in the given case the terrestrial deposits with subpermafrost water-bearing horizons appear as the lower conducting layer. The resistivities of terrestrial deposits saturated with fresh water are 300-200 Ω • m. The presence of saline and brine water in these deposits reduces the resistances to 50 Ω • m.

The fourth type. Typical for areas where a thick Cenozoic mantle formed chiefly by highly-porous sandy-clayey beds occurs possessing in a frozen state a high ice content and characterized by resistivities ranging from the some tens of thousands to the first hundreds of thousands of Ω • m.

In this connection, the maximum resistivity values are typical for the sands and sandstones. For the frozen clayey beds, the resistances can drop to 80-200 Ω • m. The thawed siltstones and sandstones containing fresh water are characterized by about the same resistances.

The entire frozen zone can be represented on the VES curves as a unified high-resistance horizon. In these cases the interpermafrost water-bearing horizons and the relatively thin interbeddings of chiefly clayey composition are not differentiated in the VES curves; their presence in the section is manifested in the nature of the curve (shifting of the entire ρ_K-curve to the left, and a reduction in the value of the maximum). In certain sectors, the interpermafrost taliks and layers of clayey beds attain great thicknesses and occur relatively near the day surface. In these instances they complicate the shape of the VES curves and are expressed in the curves as a horizon of reduced pressure.

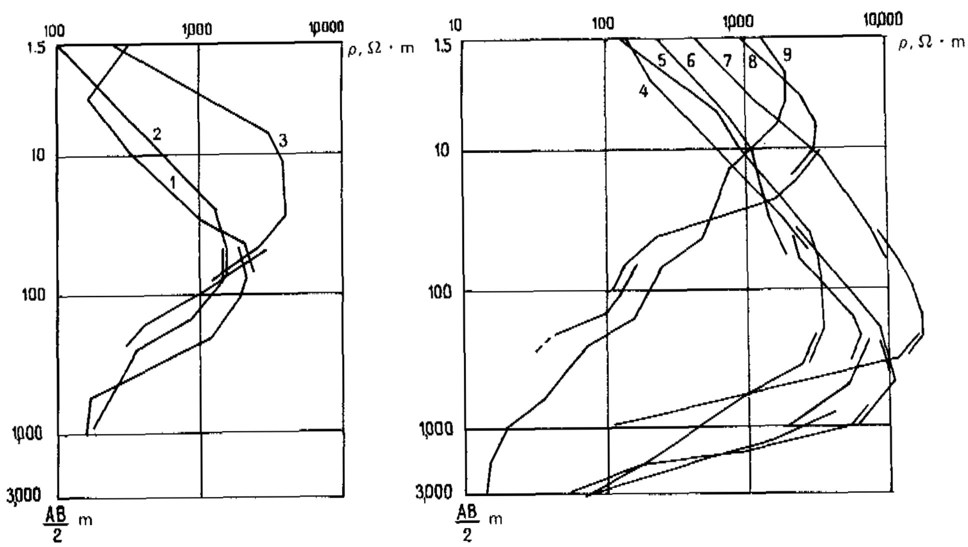

FIGURE 4 Typical VES curves for particular types of geoelectric sections. 1-3--for type II; 4-6--for type III; 7--for type IV; and 8-9--for type I.

UNDERGROUND CRYOPEGS ON SOVIET ARCTIC SHELF AND ISLANDS

Y. V. NEIZVESTNOV AND YU. P. SEMENOV (Scientific-Research Institute of Arctic Geology)

Underground cryopegs represent a liquid phase of water in the lithosphere's negative temperature belt. The negative temperature belt on the Arctic shelf is almost continuous. Constant negative temperatures of the bottom deposits are atypical only for sea sectors under the effect of warm river water and seawater of Atlantic and Pacific origin. The structure of the negative temperature belt is two-zonal. The upper zone of permafrost has a quite limited distribution. The presence of permafrost on the Arctic shelf, with thickness to 6-20 m, has been established only for certain sectors in the coastal shallow-water zones where the sea freezes to the bottom. The temperature of frozen rocks varies from -2°C to -6°C. The width of the coastal strip of permafrost on the shelf varies from several tens and hundreds of meters to several kilometers. At the shores formed of sheer rocks, with steep or medium slopes of the seafloor, the width of this strip is slight. The maximum width of the coastal strip of frozen rocks can be traced along the receding shores formed of unconsolidated Quaternary deposits.

As a result of the slight bottom gradients, the sea depth 10-15 km off such shores not exceed 3-5 m. Expansion of the frozen rock belt on the seafloor is promoted by intensive thermal erosion of the shores formed of Quaternary deposits.[3]

In the deeper regions of the basin, frozen rocks have not been found at the bottom, either by sparsely spaced boreholes or during the collecting of many bottom core samples in all the Arctic seas of the USSR. The assumptions expressed by certain researchers concerning the wide distribution of frozen rocks on the bottom of Arctic seas are based on indirect criteria, chiefly in various forms of bottom relief. In this context, we do not take into account that the conditions for the preservation of the frozen zone as a rule are lacking, even in the sectors of the seafloor where negative temperatures of the bottom water, reaching -1.5°C, or -1.8°C,[5] persist the year 'round. This is caused by the fact that usually the mineralization of subsurface water in the deposits forming the bottom of Arctic seas is much higher than mineralization of seawater (see Table 1). The freezing temperature of subsurface water coming in contact with frozen rocks usually occurs in the limits from -2°C to -10°C; in connection with this, at the higher but still negative temperatures, dissolving of ice in the frozen zone at contact with the brines occurs. In this manner, the hydrogeologic conditions promote degradation of "frozen ground" under the sea bottom, even in the belt of negative temperatures.

In addition, the rough calculations made by V. V. Baulin[1] indicate that the frozen rocks' zone with thickness of 100-120 m at ice content by weight of 15-20 percent can thaw over a period of 3,000-4,000 yr only by geothermal flow with an intensity of 30-50 $cal/cm^2 \cdot yr$. However, in case of the lower ice content values so typical of the frozen pre-Quaternary rocks, during the time of Holocene transgression, under the effect of a warm current, a frozen zone of several hundred meters' thickness could have thawed through.

The conditions for preservation of frozen rocks at the seafloor in the negative temperature belt, excluding the freezing shoals, exist only near the coasts of continents where fresh or slightly saline water has developed beneath the frozen zone.

In the Arctic, the lower zone of the negative temperature belt, formed by frozen rocks containing cryopegs, has an extensive, almost universal, occurrence. Its maximum measured thickness in the Vaygach Island region attained 60-100 m at sea depth of 5-15 m.[2] As a rule in the littoral zones of land and sea, the cryopegs revealed in the pre-Qauternary rocks were of the pressure type. According to the authors' data, this water is of marine type with mineralization from 35 g/l to 94 g/l and temperatures of -1.8°C to -4.8°C. In the deeper regions of the Arctic seas, by now cryopegs have been found represented only by pore water of bottom deposits having mineralization up to 56 g/l (see Table 1) and negative temperature to -1.8°C.

On the islands of the Soviet Arctic, the frozen zone exists almost everywhere from a depth of 0.2-2.5 m. Its thickness is 5-220 m and more. At 10-18-m depth, its temperature is 3.5°-14°C. Depending on the geologic section, the frozen zone can be subdivided into three horizons. The upper one is confined to the Quaternary mantle. Thickness runs from a few to several tens of meters. For the horizon, we typically find the presence of a considerable quantity of intrapermafrost* water, circulating intensively in the layer of seasonal temperature fluctuations; also typical is the occurrence of thick ice bodies (up to several tens of meters) of diversified form and varying origin. The subsurface ice is

*Intrapermafrost water is water in the liquid phase at negative-temperature in frozen rocks.

TABLE 1 Chemical Composition of Underground Cryopegs Taken From Rocks in Bottom of Arctic Seas

Place of Taking Sample; Enclosing Rocks	Sea Depth	Depth Range of Sample, m	Water Temp., °C	Mineralization, g/l	Content of Basic Components, Equiv %						Remarks
					Cl	SO_4	HCO_3	Na+K	Mg	Ca	
Barents Sea Basalts under layer of bottom deposits	5.9	10.9-13.9	-1.8	37.4	90.5	9.5	--	79.7	17.2	3.1	Br--73 mg/l I--3.4 mg/l F--0.4 mg/l
Basalts	1.9	8.0- 9.5	-4.8	93.6	90.9	8.8	0.3	78.2	17.8	4.0	Br--215 mg/l I--4.2 mg/l F--1.0 mg/l Sea frozen to bottom
Laptev Sea Bottom Deposits	19.0	0.0- 0.4	--	34.5	89.7	9.9	0.4	80.4	16.6	3.0	Mineralization of bottom water 24.7 g/l
	32.0	0.0- 0.2	--	37.8	89.1	10.1	0.8	81.4	16.0	2.6	Mineralization of bottom water 33.7 g/l
East Siberian Sea Bottom Deposits	13.0	0.0- 0.2	--	23.7	88.5	10.6	0.9	79.7	17.3	3.0	Mineralization of bottom water 20.1 g/l
Chukchi Sea Bottom Deposits	69.0	0.0- 0.4	--	55.6	87.9	9.7	2.4	79.6	19.2	1.2	Mineralization of bottom water 20.1 g/l
	71.0	0.0- 0.3	--	48.7	90.3	7.6	2.1	78.8	18.9	2.3	Mineralization of bottom water 32.7 g/l

from fresh water. The intrapermafrost water is either fresh or slightly saline.

The second and third horizons are formed by pre-Quaternary rocks. One of them is located in the limits of the weathering crust at a depth to 50-70 m; the other is confined to the underlying weathered or slightly weathered rocks. Intrapermafrost water in significant quantity in these horizons is inherent only to the fault zones where the rocks have been deformed and modified to complete pelitization and also to the relatively rarely encountered brown coals and other beds with high specific surface. Thickness of the ice veins confined to the fissures in the second horizon runs to several cm, while in the third horizon the thickness is several millimeters. In the third horizon, continuous ice veins are lacking. The ice is from fresh water or is slightly saline.

On the Arctic islands not occupied by glaciers, the zone of cryopegs everywhere supports the frozen zone. The depth of occurrence of this zone in the littoral part of the islands studied with width up to 3-5 km is from 5-20 to 110-220 m, with thickness ranging from 30 m to 160 m. The cryopegs are sodium-chloride brines that formed during the freezing of rocks saturated with seawater.[4] The mineralization of cryopegs bedded at contact with the frozen zone is determined by their temperature, usually running near the freezing point. Temperature of cryopegs at contact with the frozen zone runs from -10°C to 4.6°C (islands of Franz Josef Land) through -3°C to -2°C (Kolguyev, Novaya Zemlya Islands); mineralization ranges from 142-88 g/l to 60-38 g/l.

Mineralization increases with depth, attaining 160-140 g/l at the base of the cryopeg zone.

The cryopeg zone on the Arctic islands can be absent only near the largest glaciers, such as the glacial cap on Novaya Zemlya, where ice temperature in the base is close to zero.

Among the factors causing the thickness of the negative temperature belt as a whole and the thickness of the frozen zone on the islands, the most important factor is the hydrogeologic one, determining the conditions of convective heat transport in the Earth's crust. Under favorable conditions of convective descent of cryopegs forming during the freezing process, the frozen zone can attain thicknesses of several hundreds of meters, while the overall thickness of the belt can become more than 1 km. In the event of unfavorable, even the most severe, climatic conditions, the thickness of the frozen zone can be limited to several meters, while the negative-temperature belt can be restricted to 100-200 m.

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INTERACTION OF CRYOLITHOSPHERE AND SUBSURFACE WATER IN DEEP HORIZONS OF ARTESIAN BASINS

YE. V. PINNEKER Institute of Earth's Crust at Siberian Branch of USSR Academy of Sciences

The cryolithosphere exerts a great influence on the dynamics and zoning of subsurface water, chiefly of the upper water-bearing horizons. However, given certain specific conditions (continuous permafrost, rise in piezometric levels with depth, presence of neotectonic disturbances, predominance of rocks in the section, etc.), it also plays the role of a waterproof barrier for the subterranean water in the deep horizons. Then the taliks have an evolving nature, while the subpermafrost water represents accumulations of subsurface water rising from the deep parts of the geologic section.

The discharging taliks along which water flows from the deep horizons occur both in the mountainous-folded region and in the platform areas. In the first case, they are most often sulfuric and hot springs, and, in the second case, they are the outcrop of saline and brine water. Along with the discharging water, along the taliks there is the transfer of heat from the Earth's interior. This leads to a complex interaction between the cryolithosphere and the ascending subsurface water.

The effect of the permafrost barrier shield on the hydrodynamics and hydrogeochemical zoning of the deep horizons, coupled with the effect of Earth's heat in the discharged water, on the variation of thickness and continuity of the cryolithosphere can be traced fairly well in the example of the artesian basins in the north of the Siberian Platform (Tungusskiy, Yakutskiy and northern Siberian basins).* It is specifically here (Figure 1) that the permafrost is mainly continuous and with a thickness greater than 100 m.

As is known, in the deep horizons of the Siberian platform, vast resources of sodium- and calcium-chloride brines are concentrated. Owing to the block structure, renewed to a considerable extent by the most recent tectonic movements, along the "open" faults and weakened zones, the brines rise all the way to the surface, where they are discharged via the discharging taliks, in the form of springs.

A curious pattern is detected in the central part of the Tungusskiy artesian basin, where over a tremendous area the discharging of calcium-chloride brines occurs from the carbonate-halogenic Cambrian formations (depth 2,000-3,000 m). Open discharge to the surface is impeded by the poorly permeable layer (capping the geologic section) of Triassic tuff formations, which were frozen through practically the entire thickness (Figure 2). Therefore, beneath the base of the permafrost in the underlying Permain sandstones (300-500 m depth), a persistent horizon of calcium chloride brines, having migrated downward, is forming. By drilling, the subpermafrost brine-bearing horizon was traced along the lower Tungusska River in a sector with a length of 125 km (between the mouths of the Nidym and Vivi rivers). In chemical composition, the brines revealed are identical with the brines in the "original" occurrence but have a somewhat lower mineralization (200-300 g/l).

The permafrost barrier in the central part of the Tungusskiy artesian basin leads to a unique concealed discharge of brines; this has an influence, not only on the hydrodynamics, but also on the hydrogeochemical section. The latter is unique, being represented by one zone of brines. Zones of fresh and saline water are lacking. It can be theorized that a similar concealed discharge occurs also in other places

*The northern Siberian artesian basin consists of Khatatskiy basin on the east and of Ust'-Yeniseyskiy on the west.

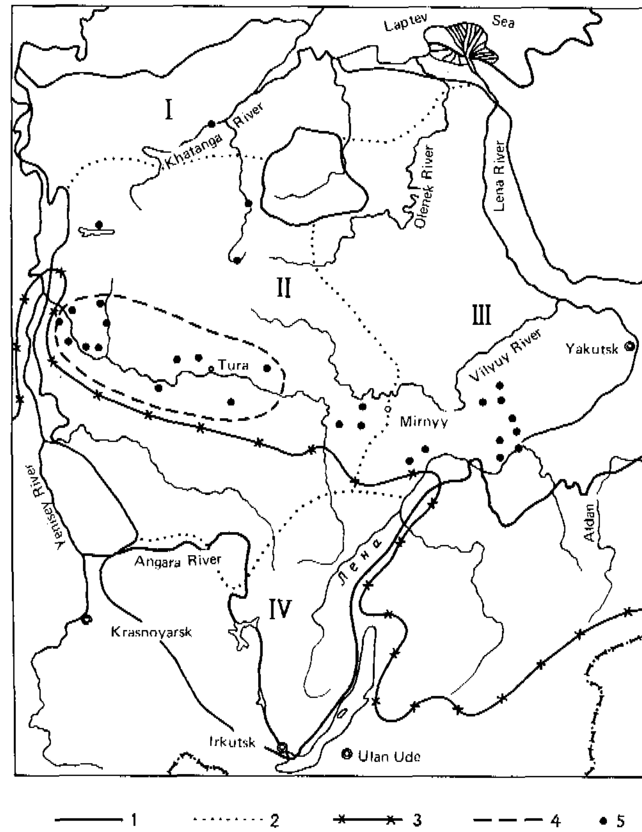

FIGURE 1 Map showing location of taliks discharging saline water and brines in the north of the Siberian Platform. 1--outline of sedimentary mantle of Siberian Platform; 2--boundary of large artesian basins (I--north Siberian, II--Tungusskiy, III--Yakutskiy, and IV--Angaro-Lenskiy); 3--southern limit of principally continuous permafrost (thickness greater than 100 m); 4--hypothetical contour for discharge of calcium-chloride brines; 5--taliks along which the surface discharge of saline water and brines from deep horizons occurs.

(e.g., along the righthand tributaries of Nizhnyaya (lower) Tunguska River).

Almost all of the taliks discharging saline water and brines to the surface in the Tungusskiy, Yakutskiy, and northern Siberian artesian basins (see Figure 1) owe their appearance to the effect of heat brought from the deep horizons. In proportion to the flow, the rising brines release a tremendous amount of heat; therefore, at the outlet to the surface, this water sometimes acquires a negative temperature.* While in the "original" bedding the brines'

*It has been shown by investigations in recent years that a negative temperature of the brines can exist, not only at points of emergence to the surface, but also in the thick stratal accumulations below the frozen zone, where they form a zone of cryopegs, at times having a thickness exceeding 1,000 m.

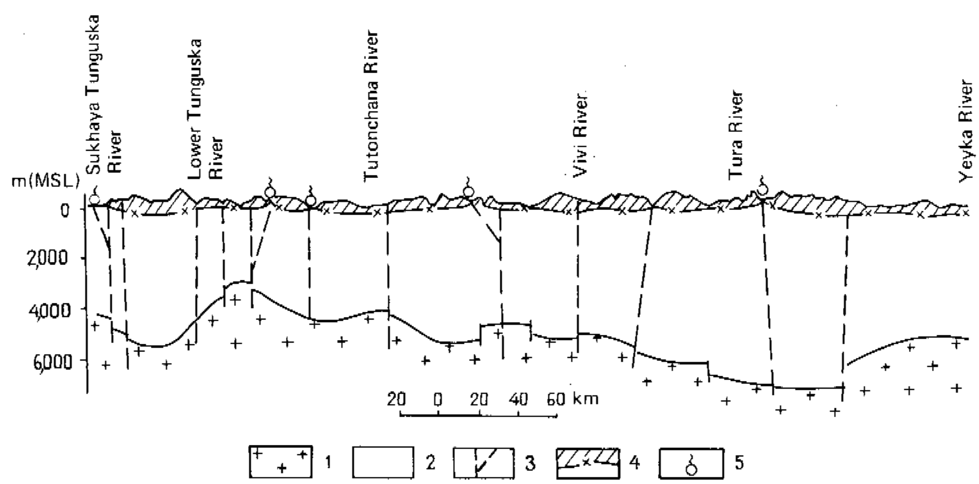

FIGURE 2 Schematic permafrost-hydrogeologic section along the lower Tunguska River (Compiled by V. N. Borisov, 1971). 1--crystalline basement; 2--thawed rocks in sedimentary mantle; 3--brine-discharging faults and weakened zones; 4--frozen rocks with lower limit of cryolithosphere; 5--outcrop of subsurface water (brines) from deep horizons.

temperature comprises 40°C, in the springs it falls to -1.5°C and even -3°C (springs on Tutonchana River, in Tabasynda River basin, etc.).

To some extent, discharge also occurs in the river valleys. It is true that under the channels of the large rivers in the north of the Siberian Platform, chiefly absorbing taliks are situated. However it would not be entirely valid to ascribe to the latter a purely exothermal origin, since the brines are also discharged into the rivers. The scale of such discharge can be judged from the saline runoff from the lower Tunguska. At Tura settlement, even in summer this river contains water composed of about 40-50 percent of sodium- and calcium-chloride, with mineralization of 0.05-0.15 g/l. During the winter-spring low-water period, the amount of chlorides increases to 70-80 percent, while mineralization of water becomes 1.2-1.3 g/l. Consequently, the formation of the through-talik near Nizhnyaya Tunguska is also due to the heat carried by the discharging brines.

Almost all the brine-discharging taliks in the river valeys in the north of the Siberian Platform have a combined exothermal and endothermal origin. In the Yakutskiy and nothern Siberian artesian basins, these taliks discharge the sodium-chloride saline water and brines. Let us point out that the surface and particularly the latent discharging of subsurface water from the deep horizons are strongly linked to the cryogenous enclosure. The Earth's heat brought from the deep horizons leads to degradation of the cryolithosphere. In other words, the information obtained testifies to the warming influence of the rising brines and of the reduction in the thickness of the permafrost. As the paleohydrogeologic reconstructions indicate, the intensive discharge of brines in the Podkamennaya Tunguska and Nepa River valleys and in the headwaters of the lower Tunguska can be one of the reasons for the northward shifting of the sourthern limit of the generally continuous permafrost.*

The interaction of the cryolithosphere and of underground water from the deep horizons is not limited to the examples cited. It has a quite different nature in the central part of the Yakutskiy artesian basin, where, below the base of the permafrost, anomalously low piezometric levels are inherent to the water-bearing horizons in the Lower Cretaceous and Upper Jurassic beds. Thus in spite of the varying approach of the researchers to this phenomenon, its satisfactory explanation should take into account the time-related interaction of the cryolithosphere with the subjacent water-bearing horizons. It is most likely that here the removal of cryogenous pressure occurred during warming and reduction in thickness of permafrost under conditions of a lack of inflow of the rising water.

*This assumption requires careful substantiation.

DEMARCATION OF CRYOLITHIC ZONE BASED ON PARAGENETIC RELATIONS BETWEEN SUBSURFACE WATER AND LAYERS OF FROZEN ROCK

S. YE. SUKHODOL'SKIY Northern Branch of Scientific Research Institute of Foundations and Underground Structures of USSR, Gosstroy

In 1927 M. I. Sumgin formulated the concept of mobile equilibrium between the quantity of subsurface water in liquid form and the nature of frozen rocks. Later on he summed up this concept in the terms: "law of mobile equilibrium between amount of water in liquid form and nature of permafrost." The gist of the rule is that "variations in climate in one direction or another are reflected on a change in the temperature conditions of the permafrost layer, while this in turn is reflected on the variation in the amount of groundwater in the liquid state.[2] An analysis of the geocryological and hydrogeological conditions in a number of regions in the cryolithic zone with varying climatic (zonal factor), with geologic and orographic (regional factors) conditions of heat-water exchange of the Earth's crust's upper layers with the atmosphere, confirms the validity of M. I. Sumgin's rule and provides the opportunity of establishing the presence of relationships, often quite distinct, among the basic elements in the subsurface water's regime (abundance, chemical and gaseous composition, intensity of exchange with surface, level regime, and so on), parameters of frozen layers (thickness, temperature, extent of discontinuity), and of the taliks separating them (dimensions, origin, predominant form of heat transfer in them).

With a certain amount of uncertainty, such relationships can be called paragenetic. In the given instance, by "paragenesis of the layers of frozen rocks and subsurface water" we understand their location in the upper layers of the lithosphere at boundary conditions (inherent to some given region or its part) of heat-water exchange in the qualitative and quantitative interrelationships, i.e., paragenetic relationships. Their nature is conditioned by the natural historical conditions involved in the formation of subsurface water and frozen rocks, determined by the climatic, orographic and geologic conditions in a region (or in a part of it), and also by the history of the development during the Pleistocene-Holocene.[3]

Regions with frozen rocks with thickness up to several tens of meters and more can be zoned according to paragenetic relationships existing between subsurface water and the permafrost layers. Such zonings can be regarded as particular forms of geocryological and hydrogeological demarcations. They first of all provide a concept of the results from the interaction of subsurface water with the frozen rocks; secondly, they favor an analysis of the processes in the interaction between them and an estimation of the factors determining them.

At the basis of the zoning, we place the separation of regions (subregions) with typical forms (subforms) of paragenetic relationships. Zonings are possible with varying extent of detail, depending on their purpose and the extent of study of the objects of zoning.

The author has compiled a system of a relatively detailed (to scale of around 1:300,000) zoning of a small part in the north of the Fore-Ural downward. The contemporary climatic and orographic conditions in the region are the same for all of its parts, and the same can be said for the history of development in the upper Pleistocene and Holocene.[1] The varieties of paragenetic relationships inherent to it and based on which three subregions have been differentiated have been caused by the features in the geologic structure of individual subregions comprising it: these are primarily thickness; lithologic composition, coupled with water-gas permeability of Cenozoic soils of sandy-clayey composition; and also composition and degree of fissuring of the underlying terrestrial and organogenic Paleozoic sediments.

The combination of paragenetic relationships distinguishing each subregion is associated with the values of "convective component" of balance of the heat-moisture field forming the subregions of rocks.[4] In each of the subregions the convective component values differing in magnitude are linked with the formation of frozen layers occurring in their limits, as well as with the development of their basic parameters and components pertaining to the subsurface water regime.

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INVESTIGATION OF RIVER NALEDS AND INCREASE IN EFFECTIVENESS OF ANTI-NALED WORK

A. A. TSVID AND A. N. KHOMICHUK Far Eastern Promstroy-niiproyekt

River naleds--a unique permafrost phenomenon--occur widely in the entire area of the Soviet Far East. The process of naled formation depends on a number of climatic and hydrologic factors: on air temperature during winter, on ground's freezing depth, quantity and time of falling of winter and summer precipitation, occurrence depth of water-impeding layer, and the river's hydrologic regime.

The dimensions of naleds vary from year to year. Depending on some given combination of the listed factors, the naleds vary in size. Often there are winters when naleds generally do not appear.

For a particular point, the geologic conditions will remain the same; only the climatic and hydrologic elements are subjected to variations. These include air temperature, amount of precipitation and its annual distribution, and amount of water flow. By the nature of the variation in the basic components influencing to the greatest extent the formation and accretion of naleds, we can also judge the actual naled process. On the other hand, having data from many years' observations of naleds, by use of analysis we can derive their quantitative characteristics; by comparing these with the change in climatic elements, we can detect certain tendencies.

At the present time, the hydrologic stations conducting observations of the rivers' regime record the days of appearance and growth of naleds in specific sectors. Data from these observations have been processed for nine winters (1959-1967) from all hydrologic stations in the Far East and the greater part of eastern Siberia.

The systematization of the data from hydrologic stations has been conducted by area of rivers' drainage system (division of all stations into five groups) and by temperature zoning, which represents the division of the Far East according to sums of mean diurnal negative air temperatures. In all we obtained 19 temperature zones.

In the report we use the following quantitative indexes of naled processes:

1. Total number *(N)* of hydrologic stations taking observations of naleds.
2. Number of stations *(n)* where naleds have been observed.
3. Probability of formation and accretion of river naleds, determined by the equation

$$j = n/N \cdot 100 \text{ percent.} \qquad (1)$$

In each temperature zone we determined the probability of the formation and accretion of river naleds on a differentiated basis by area of rivers' drainage system. We revealed a distinct dependence of probability of naled formation on the temperature factor, having been expressed in the fact that the air freezing indexes (-4,200° and -6,000° days) are critical for rivers with drainage system areas up to 1,000 and 5,000 km^2, respectively.

Their deviation in one direction or the other involves a decrease in the likelihood of naled formation. This occurs because the increase in the sum of temperatures beyond the critical values leads to a preferential freezing through of rivers with small drainage systems, without the formation of naleds on them.

The j value in rivers with drainage system areas above 5,000 km^2 is directly dependent on the temperature factor. The larger the drainage system area, the lower the probability of naled formation under the conditions of low air-freezing indexes. Thus, at freezing indexes up to -2,700°, on rivers with drainage system areas larger than 100,000 km^2, naleds do not develop. This fact testifies to the absence of critical temperatures and the lack of complete freezing of rivers under the conditions described.

The high probability of naled formation on the larger rivers at air-freezing indexes above -6,000° speaks in favor of the narrowing the active section of the channel in these rivers; however, on the small rivers (drainage system area less than 3,000 km^2) under analogous temperature conditions, complete freezing of the channel's useful cross section sets in. At this time, there is much less chance for naled formation.

The fight with naleds is by no means always successful. For this there are many reasons, but we should consider the basic ones to be the inadequate value of dynamics of naled process, imperfection of existing recommendations having a general nature and obviously containing insufficiently precise engineering calculations for determining the parameters of the antinaled techniques. Therefore, the development of methods for determining the transverse dimensions of the deepened river channel and the necessary thickness of the heater layer in the water diverting units appear quite timely.

The deepening of channels is one of the common methods of counteracting river naleds. The idea of deepening the channel is designed for the warming effect of the snow-ice cover protecting the channel from deep freezing. Successful application of the channel dredging technique largely depends on how correctly the parameters, namely the depth and width of the new channel are assigned. It has been established that the amount of deepening a channel can be computed, orienting ourselves on ice thickness on rivers in the region of interest between the average annual values and maximal values, depending on the frequency of winters with especially intensive naleds, based on the equation:

$$H_D = 1.5\ H_i, \tag{2}$$

where H_i = thickness (m) of ice on rivers of cal-*

*In the calculations, we recommend the probability be assumed 20 percent, since the likelihood of the most intensive naleds by years averages 20 percent.

culated probability. The coefficient 1.5 allows for the channel's cross section.

The width of the channel subject to deepening is computed from the condition of equality of areas of water flow section during winter, and the deepened channel, according to the equation

$$B = S/H_D, \tag{3}$$

where S = cross-sectional area of the flow in winter, m^2.

In the fight against naleds, we often utilize various water-diverting units (chutes, troughs, etc.) for redirecting the naled water. The operating quality of this technique depends basically on the extent of their warming.

The thickness of the heater layer in the water-diverting units depends not only on the material in use, but also on the climatic conditions. The identical climatic factors causing the thickness of the heater layer and thickness of ice cover on the rivers indicate the existence of a linear relationship between these parameters. Hence, using the thickness of ice on the rivers, for the water-deflecting arrangements, we can determine the thickness of the heater layer in the region of interest, from the expression

$$H_M = \lambda_M/\lambda_i \cdot H_i, \tag{4}$$

where H_M = unknown thickness of heater layer, m; λ_M = coefficient of thermal conductivity of heater material, kcal/m·h·deg; λ_i = 2.0 = coefficient of thermal conductivity of ice, kcal/m·h·deg.

The results obtained from the investigations can be recommended for application in forecasting river naleds and for planning the facilities for counteracting their accretion.

SUBSURFACE WATER REGIME IN AREAS OF OCCURRENCE OF CRYOGENOUS PROCESSES AND HYDROCHEMICAL ZONING OF FOSSIL ICE

V. G. YAS'KO   Irdutsk Polytechnical Institute

The expression of cryogenous processes determines the specific features in the regime of the subsurface water's chemical composition. The natural conditions existing in the region considered as a unit are characterized by a distinctly continental climate with mean annual negative temperatures favoring the formation of a fairly thick zone of seasonally frozen ground. At the same time, we can clearly trace from north to south the climatic zoning, expressed in a variation in a mean annual temperature from -10.8°C (Katugino settlement) to -0.8°C to -1°C (city of Kyakhta, Kaylastuy and Solov'yevsk settlements). Climatic zoning is also reflected in the length of the freezing period, attaining 79 percent of the annual cycle in the northwestern part of the Vitimskoye plateau and 60 percent in the southern part of the Trans-Baikal. From the mountain structures of Muyskiye ranges, Kodar and Udokan toward the steppe regions of Dzhida and Priargun'ye, we find a change in the mean annual total precipitation from 800-1,000 mm to 200 mm and less. In this same direction, the area of continuous permafrost with thickness greater than 100 m is supplanted by a transition area and then by an area of the insular development of permafrost with thickness less than 30 m. In the area of continuous permafrost, the rocks have low temperatures, i.e., less than -1°C to 2°C, while in the transition area and farther south, permafrost temperatures will fluctuate from -0.1°C to -0.5°C.[3,4]

The subsurface water regime had been studied at 12 points in the upper part of the free-water exchange zone in the range from 1m to 120 m; within these limits, in the Trans-Baikal territory, chiefly hydrocarbonate water is forming, with a varying ration of calcium, magnesium, and sodium, cations. At points of systematic observations, we employed springs, boreholes, and mine workings, such as shafts and adits, which made it possible for us to study:

1. Suprapermafrost water in the colluvial beds (Ikatskiy, Amazarskiy, Malkhanskiy, and Daurskiy ranges; Gusino-Udinskiy region; and Vitimskoye plateau).
2. The fissure water in the weathering zone of intrusive and metamorphic rocks (Amazarskiy Range, Erman Range, Vitimskoye plateau, and Gusino-Udinskiy region).
3. Fissure-vein water in the zones of tectonic disturbances (Amazarskiy and Nerchinsko-Kuenginskiy ranges).
4. Interstratal water in the Mesozoic basins (Malo-Amalatskaya, lower Nerchinskiye, Urulyunguyevskiy, and a number of other basins).

Based on the results obtained from processing the chemical analyses of water samples collected from the zone of seasonally frozen rocks during 1-3 yr cycles, we established a periodic frequency of abrupt increase, all the way to the predominant position of the magnesium cation, during the active layer's periods of freezing and thawing.

The periodic increase in the magnesium cation in the subsurface water's composition has been established fairly reliably, both in the suprapermafrost water undergoing phase transitions and in other types of subsurface water occurring below the depth of constant annual temperatures. However, with an increase in the depth of occurrence of subsurface water, the number and duration of periods of magnesium's prevalens in the chemical composition diminish. This is also typical for the hydrogeologic masses and for artesian basins, but the depth to which contemproary cryogenous processes are reflected on the chemical composition's regime differs for both types of hydrogeologic structures (Figure 1).

Thus, while for the suprapermafrost water, we typically have two periods, i.e., in spring (from April to May) and fall (from October and, hypothetically, until December), when the magnesium hydrocarbonates are the basic components in the water's composition, even at 15-20 m depth in the weathering zone of intrusive rocks, magnesium predominates in the subsurface water for a period of 5 months (from November to March). In the 50-70-m depth range within the limits of the hydrogeologic masses, the role of cryogenous factors in the formation of the water's composition is expressed only during the fall; freezing of the active layer proceeds, while the duration of the dominant position of magnesium in the subsurface water does not exceed 1 month (see Figure 1). The amplitude in the fluctuation of relative magnesium contents with depth also decreases significantly: from 75-100 equiv percent above the belt of constant annual temperatures to 25-40 equiv percent at depth of 50-70 m. For practical purposes, below 100m, the process of modifying the chemical composition under the effect of cryogenous factors in the hydrogeologic masses attenuates (see Figure 1).

In artesian basins, the cryogenous enrichment of magnesium has been established to a depth of 40 m (see Figure 1), although in isolated points (Malo-Amalatskaya basin) under a layer of permafrost with a thickness to 50-100 m and more, we note a sharp increase from 35-40 to 65-70 equiv percent of magnesium ion at decrease in mineralization toward the end of the spring season. A similar reconstruction of the chemical composition has also been established beyond the limits of Trans-Baikal during the prospecting of the Neryundinskoye ire-ore deposit in the north of the Angara-Lena artesian basin. Obviously, in

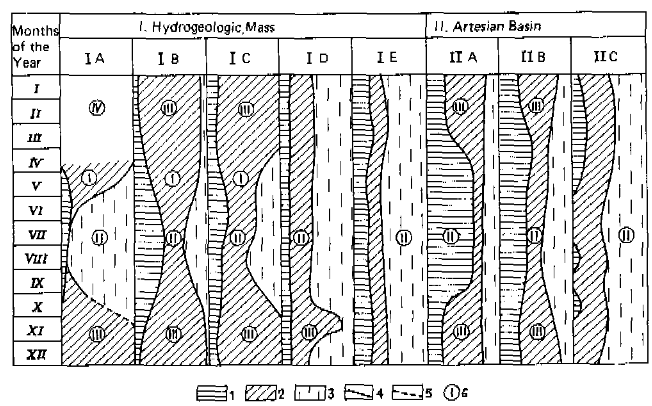

FIGURE 1 Variation in cationic composition of subsurface water in hydrogeologic masses and artesian basins under effect of cryogenous processes. Content in equiv percent: 1--Ca; 2--Mg; 3--Na + K; 4--limit of content of each element established according to analyses; 5--the same, by assumption (by analogy with lower horizons); 6--stages of formation of water composition: I--thawing of ice; II-infiltration of atmospheric precipitation; III--transformation of composition under effect of a new freezing cycle (conversion of water to ice); IV--transformation of content in zone of seasonally frozen rocks by means of transport of salts with film water. I. *Hydrogeologic mass:* IA--water in active layer (1-3 m); IB--fissure water in zone of granites' weathering (depth to 10 m); IC--the same (depth 15-20 m); ID--fissure-vein water in zones of tectonic disturbances (depths to 70 m); IE--the same (depth 100-120 m). II. *Artesian basin:* IIA--fissure-stratal water of terrestrial sedimentary beds (depth 14 m); IIB--the same (depth 20 m); and IIC--the same (depth 40 m).

an expansion of the network of systematic observations, this process will be noted for the shallowly occurring subsurface water in practically the entire occurrence territory of the seasonally freezing rocks.

The cryogenous processes leading to the specific hydrochemical regime of subsurface water in the zone of free-water exchange during a yearly period are definitely cyclic throughout geologic time, promoting a variation in the thickness of the permafrost. Under the conditions of an increase in permafrost thickness and during its degradation, changes in the chemical composition of subsurface water can occur; at the same time, the periods of prevalence of any given type of subsurface water will have a more extended time span.

Thus T. N. Yas'ko and V. G. Yas'ko[7] were the first to note the zonal magnesium hydrocarbonate water with mineralization often exceeding 1 g/l in the Mesozoic depressions in eastern Trans-Baikal within the confines of the transition area for continuous to insular development of permafrost. The fringes for the development of this water gravitate toward the islands of permafrost with an unstable temperature regime close to zero. Experimental studies have confirmed the possibility of formation of an analogous chemical type of water in the initial stage of degradation at the base of the permafrost. The hydrochemical regime of this type of subsurface water is quite constant over a 3-yr period of steady observations.

The study of the regime and determination of the role of cryogenous processes in the formation of the composition of the subsurface water involves an investigation of the ice phase. By now in the literature, information is at hand on the chemical composition of ice in lakes and naleds,[1] on marine arctic ice,[2] ice in caves,[6] and underground ice.[5] However, these data concern ice occurring either at the surface or at depth less than 10 m from the surface, whereas the chemical composition of the [ice in] frozen rocks many hundreds of meters thick has not been studied.

On the basis of the information available in the literature, we can form the conclusion that the least mineralized are the sublimation-type ice, in the composition of which magnesium hydrocarbonates predominate.[6] At the top of the permafrost where the ice has formed from the liquid phase, the total salts in the ice samples do not exceed 10-25 mg/l, while, among the cations, magnesium rarely occupies the dominant position. At depths close to the base of the permafrost, the ice has been poorly studied. New data have been obtained in testing the ice in one of the holes bored in the Mesozoic depression on Vitimskoye plateau. At a permafrost thickness up to 95 m, ice was taken from depths of 5 m, 35 m, and 47 m. Mineralization of ice increased from 21 mg/l to 159 mg/l; the composition changed correspondingly from sodium hydrocarbonate to magnesium-sodium hydrocarbonate. Sodium-magnesium hydrocarbonate water was found directly beneath the frozen rocks.

The process of enrichment by magnesium at inception of permafrost requires verification by experimental studies and by field observations. In this respect, in addition to the study of fossil ice, considerable help can be obtained from investigating the ice on the small rivers in Trans-Baikal, freezing to the bottom, often consisting of permafrost. For the first time, data on the composition of ice on the Olov River, not having runoff during the winter, were obtained by the author in 1960 during the trenching of a section of a 2-m ice layer every 0.5 m. With original water of sodium-magnesium-calcium hydrocarbonate composition with mineralization of 69 mg/l, ice layers formed with total of salts from above downward comprising 25, 83, 136, and 144 mg/l, respectively, and with relative magnesium contents ranging from zero to 72 equiv percent. Obviously, during the freezing of the water-bearing horizon occurring on the water barrier through which the forcing of water into the

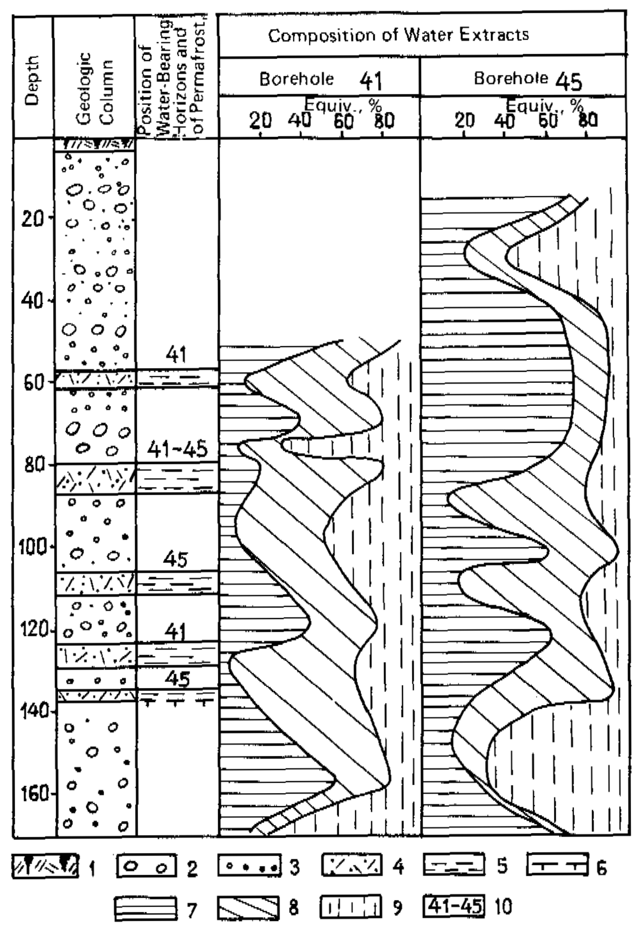

FIGURE 2 Cation composition of water extracts from permafrost Quaternary beds: 1--colluvial clay loams; Mesozoic beds: 2--conglomerates; 3--coarse sandstones; 4--interstratifications of fissured coarse sandstones and sandstones; 5--hypothetical water-bearing horizon; and 6--base of permafrost. Cations: 7--Ca; 8--Mg; 9--Na + K; and 10--numbers indicating the boreholes in which a water-bearing horizon has been found.

underlying horizons is prevented, analogous tendencies in the variation in the ice's chemical composition will occur.

As confirmation of the possibility of the formation of chemically zonal ice in the water-enclosing rocks, use can be made of water extracts from the rocks occurring in the negative-temperature zone. As the results of investigation have shown, in one of the basins on Vitimskoye plateau, the base sectors of the water-bearing horizons contain 3-10 times more mobile magnesium ions as compared with the frozen rocks which do not contain visible ice pockets (Figure 2).

In this manner, the cryogenous processes determine the features in the regime for the chemical composition of subsurface water, form a hydrochemically zonal section of ice, and are capable of enriching the permafrost with easily mobile components.

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LAKE COVERAGE IN USSR CRYOLITHIC ZONE

S. YE. MOSTAKHOV Yakutsk State University

The concentration of lakes in any territory can be described by the following indexes:[1-3] number of lakes; density of lakes (their number per 1,000 km^2 of area); total surface area of all lakes; and coefficient of lake coverage, i.e., by the ratio of total area of lake surface in each region to the total area of the region expressed as a percentage.

Let us consider these indexes for lakes within the USSR cryolithic zone. As original data, use was made of the data from an inventory of water bodies conducted by the administrations in the USSR Hydrometeorological Service from 1960 to 1966 under the supervision of the State Hydrologic Institute.[1] The calculation of the number of lakes and measurement of their surface areas were conducted on the basis of large-scale maps published in recent years.

Based on the most recent data, in the vast region of the USSR cryolithic zone (11 million km^2, there are more than 2 million lakes with a total surface area of about 300,000 km^2. Individual regions (Malozemel'skaya tundra, north Siberian, Yana-Idigirka, central Yakutian, Kolyma lowlands, and others) are distinguished by a significant number of lakes (see Table 1).

The density and lake coverage of any given region in the USSR cryolithic zone depend on a complex range of natural factors; the most important among them is permafrost. Maximum accumulations of lakes are confined to the regions with extensive development of thermokarst pro-

TABLE 1 Largest Lakes in the Cryolithic Zone of the USSR

Region	Number of Lakes: Total, thousands	Number of Lakes: per 1,000 km^2	Total Surface Area of Lakes, thousands of km^2	Lake Coverage, %	Number of Lakes with Area from 1 km^2 and Up: Total	Number of Lakes with Area from 1 km^2 and Up: per 1,000 km^2
Malozemel'skaya tundra and forest-tundra	125.0	1,008	5.1	4.1	367	3
West Siberian tundra	96.5	472	16.8	8.3	1,902	9
Trans-Baikal region	47.1	52	35.6	3.9	215	0.2
Central Siberia						
Northern part	41.1	105	3.6	0.9	369	0.9
Central part	47.6	71	2.0	0.3	201	0.3
Eastern part	137.3	241	10.4	1.8	1,505	3
North Siberian lowland	318.8	370	38.5	4.4	3,725	4
Lena River delta	58.7	2,120	3.2	12	421	15
Yana-Indigirka and Kolyma lowland	247.8	450	46.1	8.4	7,023	13
Mountainous regions						
Northeast	347.3	126	21.7	0.8	2,698	1

cesses and with complicated runoff conditions. Naturally, the smallest number of lakes is concentrated on the plateaus, uplands, and mountains.

Density of lakes in various regions of the USSR cryolithic zone will fluctuate from 2-5 to 2,000 and more. Thus, the density of lakes in the Lena River delta (area 27,000 km^2) is 2,120; on the Malozemel'skaya tundra, it is 1,008; and on the west Siberian tundra, it is 472. In the Tuyma River basin (watershed area 1,400 km^2), where 1,200 lakes have been tallied, their density equals 857. On the other hand, for the Great Chuya River basin (left hand tributary of the Lena), it equals only 2. The average density of lakes in the USSR cryolithic zone is 184.

The lake coverage of the USSR cryolithic zone in its various parts will fluctuate from 0.01 (Khoron River basin) to 12 (Lena River delta). Average extent of lake coverage is 2.7 percent.

The lakes in the cryolithic zone are quite diversified in the origin of their basins, size and shore configuration, hydrometeorological regime, chemical composition of water, etc. The lakes of thermokarst origin, usually small (up to 1-5 km^2) and shallow (up to 3-5 m) comprise the most numerous group among the lakes in the region that we are discussing. The landscapes of many lake regions in the USSR cryolithic zone are determined only by thermokarst lakes occupying up to 90 percent of the total number of lakes. It is specifically by virtue of this that we explain the relatively small area of the surface of all lakes in the USSR cryolithic zone, equalling two-thirds of the area of the Caspian Sea.

Within the boundaries of the USSR cryolithic zone, we rarely find large lakes with an area greater than 200 km^2 (Lovozero--205, Yessey--238, Lama--318, Mogotoyevo--323, and others). Only two lakes have an area exceeding 1,000 km^2 (Baikal--31,500 and Taymyr--4,560).

The lakes in the USSR cryolithic zone have not yet been adequately studied. Their detailed interdisciplinary study would have not only theoretical but also great practical importance for solving many problems in industry, construction, agriculture, etc. The time has arrived for expanding the network of lacustrine hydrometeorological stations and posts; in this connection, we should adhere to the areal principle of allocating the network, the gist of which reduces to a selective study of the most characteristic or typical lakes, for which the results of observations can be extrapolated, with known approximation, to many analogous water bodies over an extensive territory.

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PART VI

Geocryological Surveying and Prediction

V. A. KUDRYAVTSEV,
I. A. NEKRASOV, and
A. T. AKIMOV, *Editors*

FEATURES INVOLVED IN PERMAFROST ENGINEERING-GEOLOGIC INVESTIGATIONS DURING EXPLORATION FOR MINERAL DEPOSITS

G. A. GOLODKOVSKAYA, L. M. DEMIDYUK, AND L. V. SHAUMYAN
Moscow State University

The discovery and development of the richest deposits in western and eastern Siberia, a considerable share of which is situated at great depths and in the permafrost zone, makes the problem of developing a procedure for the engineering-geologic investigation of such deposits particularly urgent at the present time. In a solution to this problem, equally important are both an evaluation of the natural condition of the rock massifs, as well as a prediction of the mining geologic phenomena originating as a result of underground operations.

Many important theoretical, procedural, and practical questions in engineering geology of mineral deposits have been reviewed in the reports prepared by P. N. Panyukov, S. S. Avershin, Yu. N. Malyushitskiy, V. V. Rzhevskiy, G. L. Fisenko, and other scientists. In most of them, the problems are examined as they develop during the operation of mines. The actual technique of engineering-geologic investigations during the prospecting for deposits has been inadequately developed.

Taking into account the great scientific-procedural significance, urgency, and practical value of such investigations, the Department of Soil Sciences and Engineering Geology at Moscow University for a number of years has been conducting engineering-geologic studies on deposits of hard minerals in the Noril'sk area. Certain questions about the methodology involved in these investigations are of undoubted interest, and their answers can be utilized in conducting similar operations in other deposits in the Soviet Union.

Thus, for a series of formations in the Talnakh mining center, the extent of rock-jointing, depth of tectonic disturbance, and size of displacement can be correlated with the lithography and primary physicomechanical properties of the rocks. Here, confined to large-scale fracturing, there are extensive zones of jointing, reaching a displacement of 300 m, in a width of 200 m to 250 m. In this, zones of shattered rock, averaging 50 m to 100 m, zones of strong faulting, about 100 m, and zones of somewhat higher jointing, not more than 50 m, exist. With a reduced dislocation of 10 m to 50 m, the widths of the zones diminish correspondingly to 50 m.

The more subdued tectonic disturbances (fault displacements, fracture movements accompanying block-faulting, structural slip, and shearing) are, as a rule, associated only with zones of increased jointing 10 m to 50 m wide. The zones of shattering and intensive jointing are not more than 1 m to 5 m at the fault plane and are missing in a number of cases.

The scales and types of these operations vary at different stages of the deposits (Table 1). The quantitative data collected during the permafrost engineering-geologic studies concerning the strength, elastic parameters of rocks, their resistance to jointing, and the extent of jointing permit us to predict the nature of the bedding and the form of ore-bearing intrusions and to provide additional information in solving the problem of the deposits' origin.

TABLE 1 Engineering-Geologic Operations at Various Stages in Exploration and Development of Deposits

Stage in Exploration and Development of Deposit	Engineering-Geologic Tasks	Limits of Engineering-Geologic Studies	Cartographic Materials Compiled
Prospecting and exploration	1. Engineering-geologic reconnaissance on a scale of 1:200,000-1:50,000 2. Engineering-geologic testing of individual mapped holes and laboratory identification of classification indexes 3. Permafrost engineering-geologic interpretation of geophysical data	Coincide with limits of exploratory surveying	1. Schematic-geologic chart on scale of 1:200,000-1:50,000 2. Schematic engineering-geologic sections with indication of distribution of frozen ground
Preliminary exploration	1. Engineering-geologic surveying on a scale of 1:25,000-1:10,000 2. Engineering-geologic tests of exploratory boreholes based on guide profiles 3. Drilling of special reference permafrost engineering-geologic boreholes, their detailed testing and permafrost regime observations in them 4. Special geophysical tasks of permafrost engineering-geologic purpose 5. Laboratory determination of generalized values of classification and certain design indexes for description of the soil types	Slightly exceed the limits of deposit exploration	1. Permafrost geologic map on a scale of 1:25,000-1:10,000 2. Maps of engineering-geologic regions for underground construction on a scale of 1:25,000-1:10,000 3. Schematic chart indicating jointing of rocks on a scale of 1:25,000-1:10,000 4. Reference engineering-geologic sections with detailing of properties and condition of bedrock on a scale of 1:10,000-1:5,000 5. Reference permafrost engineering-geologic sections with detailing of properties and condition of Quaternary beds on a scale of 1:10,000-1:5,000

Detailed exploration	1. Engineering-geologic surveying on a scale of 1:10,000-and larger 2. Drilling of special permafrost engineering-geologic boreholes, permafrost systematic observations in them 3. Detailed permafrost engineering-geologic testing in special boreholes and in exploratory boreholes based on reference profiles 4. Special geophysical operations 5. Laboratory determinations of direct design indexes of the physical-mechanical properties of soils 6. Stationary observations for establishing relationship of the freeze-thaw processes with the elements of external heat exchange	Limits are determined by complexity of engineering-geologic conditions in the territory, usually exceeding the exploratory limits	1. Permafrost-geologic map to a scale of 1:10,000-1:5,000 2. Map of engineering-geologic regioning for the aboveground construction to a scale of 1:10,000-1:5,000 3. Map of engineering-geologic regioning for underground construction 1:10,000-1:5,000 4. Engineering-geologic sections to a scale of 1:10,000-1:2,000
Construction	1. Engineering-geologic studies on individual objects 2. Control workings 3. Control determinations of direct indexes of physicomechanical and thermophysical properties of soils 4. Stationary observations	Largest and most responsible objects	Special forecasting maps for standardizing and evaluating the stability of structures

NEW ELECTROMETRIC TECHNIQUE FOR STUDYING THE STRUCTURE OF PERMAFROST

YU. A. AVETIKYAN Industrial and Scientific-Research Institute of Engineering Surveys in Construction-INIIS

In the grouping of methods used in engineering geophysics during exploration in permafrost regions, an important role is relegated to direct current electrical exploration and, above all, to the techniques of vertical electrical logging (VEL), dipole sounding (DS), electrical profiling, and so forth. At the same time, in many cases for expanding the potentialities of electrical exploration, it is feasible to utilize alternating current (in a specific frequency range). Thus, in the practice of geologic-prospecting work, we use frequency sounding (FS), the effectiveness of which is most perceptible in a study of the sections containing high resistance (for example, extremely icy) horizon screens. The latter do not comprise an obstacle to the transmission of alternating current to the underlying layers. Transmission is by electromagnetic induction in distinction from the processes occurring in a medium under the effect of static fields. Incidentally, this does not exhaust the essential advantages of FS.

For a number of years, in the INIIS, research has been under way associated with the development of a slight-depth modification of frequency electromagnetic sounding. In the FS practice, we normally utilize the low frequencies (within the limits of hundreds of Hz and a maximum of tens of kHz), which determine the possibility of investigating only relatively great depths. The problem formulated regarding the study of the frozen layer and the requirement associated with it for expanding the frequency range by 2-3 orders (up to 15-17 octaves) has predetermined three directions in future operations: (1) designing and development of new equipment, (2) development of a procedure for field operations and determination of the method's potentialities, and (3) the accomplishment of a design of frequency-sounding surveyor's plane tables with consideration of the displacement currents caused by the polarization processes of the dielectrics.

The essential problem in the electromagnetic soundings includes finding the thicknesses and resistivity of the electric horizons, as a rule corresponding to the beds with a varying lithologic composition and geocryological structure. The electromagnetic field strength depends on the actual geoelectrical conditions and the necessary information can be obtained by studying the field components.

We examine below only the equatorial-frequency soundings which are conducted at constant separation r of the feeding (transmitting) and receiving systems for obtaining the dependence of the components of a stationary field on frequency.

The nature of the distribution of an electric field from any source in a homogeneous bed is determined by the Maxwell equations; their solution in the case of a stratified medium is the basis for finding the apparent specific resistance ρ_ω. Following L. L. Van'yan,[3] we assume as the basic horizontal electrical component E_x and the vertical magnetic component B_z. In the method of frequency sounding, the basic parameter controlling the penetration depth of the field is current frequency ω. The higher the frequency, the shallower the penetration depth of the field into the earth. Therefore ρ_ω satisfies the condition:

$$\lim_{\omega \to \infty} \rho_\omega = \rho_{t_1}, \tag{1}$$

where ρ_{t_1} = longitudinal resistance of the first layer. The function ρ_ω depends on the parameters of geoelectrical section, on frequency, etc. (We shall distinguish between $\rho_{\omega E}$ and $\rho_{\omega B}$, according to the measurements of the electrical and magnetic components, respectively).

In the interpretation, the experimentally obtained curves of ρ_ω and $\sqrt{T}$ are compared with theoretical curves (master curves). Here T = period of oscillation ($T = 2\pi/\omega$). The known sets of master curves are calculated on a computer for certain typical 2-4-layered sections. They are by no means always suitable for the interpretation of the results of FS for permafrost, since their calculation was based on the equations of quasistationary approximation, i.e., without considering the effect of the displacement currents. This is admissible only in a relatively low-frequency range at low values of relative dielectric permeability and specific resistance of the ground under study.

The field FS station is a relatively complicated grouping of instruments.[1] The complexity is caused by the requirement:

1. To develop in the ground fairly powerful alternating electromagnetic fields in a wide range of frequencies (field investigations have shown that usually the curves of ρ_ω typify the upper layers of the frozen bed in frequencies of several MHz).
2. To conduct amplitude measurements of the field components in the indicated frequency range under conditions of intensive electrical interferences, diversified in nature and origin and frequently exceeding the receiving signal by dozens of times.

The installation constructed is mobile and meets these requirements. It possesses a number of features that have permitted us to expand significantly the frequency characteristic of the generator and to increase the output power and efficiency. The circuit of the microvoltmeter (FS receiver) is arranged in such a manner that from the electrical circuit of the selective and amplifying stages, the resonance oscillatory systems are completely excluded. The elimination of thermosensitive oscillatory elements predetermined the stability and accuracy of measurements in a broad range of temperatures and, at the same time, the standardized sensitivity in a range of frequencies from 80 Hz to 6 MHz (linearity of frequency characteristics of the selective-amplifying channel).

The basic parameters of the FS installation are as follows:

Range of frequencies	80 Hz-6 MHz
Nominal output power of generator at load ranging from 20 to 1,000 m	100 W
Maximal sinusoidal current in emitter (coefficient of harmonic components less than 6 percent)	3 A
Sensitivity of receiver	5 μV
Input resistance	600 kΩ
Pass band (fixed)	30, 300, and 3,000 Hz
Selectivity (at tuning to frequency equalling the established value for the pass band)	not less than 40 db
Accuracy of measurements	±1.5 percent
Range of operating temperatures	from -35°C to +45°C
Weight of all units in installation together with power sources	30 kg

In the field observations, it is necessary to take into account those complications that develop in connection with using high frequencies. The basic constructions in the theory of frequency sounding pertain to the case when the radiator is the point-source type. From this we get the condition for line *AB*:

$$AB << \lambda_1, \qquad (2)$$

where λ_1 = length of wave in first layer.

In the area of the upper limit of the particular range in use, disruptions of this inequality are possible:

$$\text{Length of wave } \lambda_1 = 2\pi/b, \qquad (3)$$

where b is determined from the expression

$$b^2 = \frac{\omega^2 \varepsilon_1 \mu_0}{2}\left[\sqrt{1 + \frac{1}{\varepsilon_1 \rho_1 \omega}^2} + 1\right]. \qquad (4)$$

Here μ_0 = magnetic permeability of vacuum while ρ_1 and ε_1 = resistivity and dielectric constant of the first layer, respectively.

As a rule, the value $AB \le$ 5-10 m is acceptable, but in certain cases we need shorter dipoles *AB*. In one of the possible extreme cases, when $\rho_1 = 50\ \Omega \cdot$ m, $\varepsilon_1 = 3 \cdot 10^{-10}$ *F*/m, f = 1 MHz, we obtain $\lambda_1 \approx 21.6$ m.

In similar cases, a decrease in the length *AB* involves a reduction in the level of the receiving signal. However this can be compensated by a certain decrease in separation r. Usually the latter as a minimum exceeds by several times the maximal depth of investigation. With an increase in separation, we have an improvement in the differentiability in the right-hand part of the field curve (this can be traced reliably from the theoretical curves).

In the high-frequency part of the range (at frequencies higher than 0.1-0.5 MHz), the intervals between the measurement points are chosen to be short (less than 100-300 MHz). In this region, curve ρ_ω can have a series of closely situated extremums corresponding to the geoelectric layers of slight thickness (and differing heterogeneities). In the short-wave part, all the amplitude curves of the FS are also characterized by oscillation around an asymptotic value.

In permafrost conditions, we fairly often encounter a geoelectrical section containing a layer consisting of a series of thin alternating intercalations of ice-soil with varying longitudinal resistance. Such a layer appears on the FS curve usually as a homogeneous layer. The equivalence of the stratified bed to a homogeneous one definitely decreases the amount of detail in the study of the geoelectrical section and determines the resolving capacity of the FS. The method of frequency sounding also is not free of the effect exerted by the equivalence principle, but its limits are not as broad as for the techniques based on the application of direct current. We can readily be convinced of this if we compare the theoretical curves ρ_k of vertical electrical logging and ρ_ω of frequency sounding (based on results from measuring only one of the field components). In the frequency-sounding process, it is possible to obtain more complete data (the measurement of two and more components and the phase displacement for each of them together with the variation in separation r). With consideration of what has been said, the limits to the principle of equivalence are obviously narrowed to the minimum.

The equipment and procedure for a shallow-depth frequency sounding have been developed under various geocryological and geologic conditions (in the Bol'shezemel'skaya tundra, in the regions of eastern and western Siberia and also in the Ukraine). As an example, we review below certain results obtained from the field measurements.

In Figure 1 we have indicated the curves ρ_{ω_E} (r = 120 m) along with the ρ_k of VEL (A. I. Khlebunov) and ρ_t. The latter has been converted (A. T. Akimov) from the VEL results based on the method of subtracting the fields. To the right

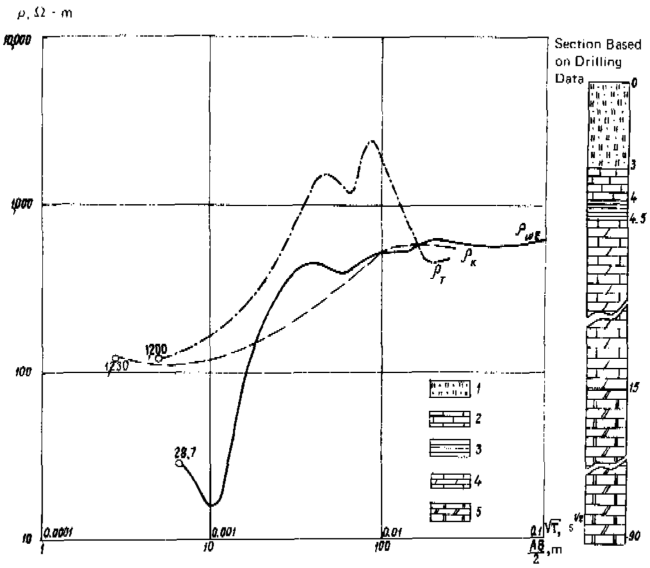

FIGURE 1 Field curves of ρ_{ω_E}, ρ_k, and ρ_t. 1--surface layer; 2--compact limestone; 3--clay; 4--shattered limestone and dolomite; and 5--competent limestone and dolomite.

in the figure we have shown a geologic section based on the drilling data. In the given sector, we have recorded frozen ground with a seasonally thawing layer of around 2 m.

Examining Figure 1, we can note the correspondence of the FS curve to the actual geologic section. Five geoelectric layers can be noted fairly distinctly in it. Curve ρ_T also reflects five layers, whereas the geoelectric section based on the results of interpreting the ρ_k of the VEL is represented [Siberian Branch of the Scientific Research Institute of Foundations and Underground Structures (CO NIIOPS)] in the form of a three-layered section with thickness of first layer h_1 = 15 m and of second h_2 = 60 m (ρ_1 = 1,200 Ω · m; ρ_2 = 10,000 Ω · m; ρ_3 = 50,000 Ω · m).

A quantitative interpretation of curve ρ_{ω_E} was conducted with the aid of four-layered sets of master curves.[5] The results of interpretation are as follows: $h_1 = 2.5$ m, $\rho_1 = 55$ Ω · m, $h_1 + h_2 = 5$ m, $\rho_2 \to \infty$, $h_1 + h_2 + h_3 = 15$ m, $\rho_3 > 220$ Ω · m, $\rho_4 \to \infty$. In the FS problem we have not included a determination of the lower limit of freezing, which was not noted by the ρ_{ω_E} curve owing to the low value of separation r.

In the given case, a higher accuracy of qualitative interpretation of the FS results can be obtained in utilization of sets of master curves with resistance ratios $\rho_2/\rho_1 = \rho_4/\rho_1 = 10^5$ and also with the use of sets of master curves calculated with consideration of the bias currents.

In Figure 2 we have shown the FS curve (ρ_{ω_E}, r = 150 m). Along with it we also present the curves of the cruciform VEL (BC TISIS, M. Okladnikov). Drilling was not accomplished at the given point, but, based on the preliminary data, a tectonic disturbance took place here.

The FS equipment permits us to conduct measurements of the methods of dipole electromagnetic

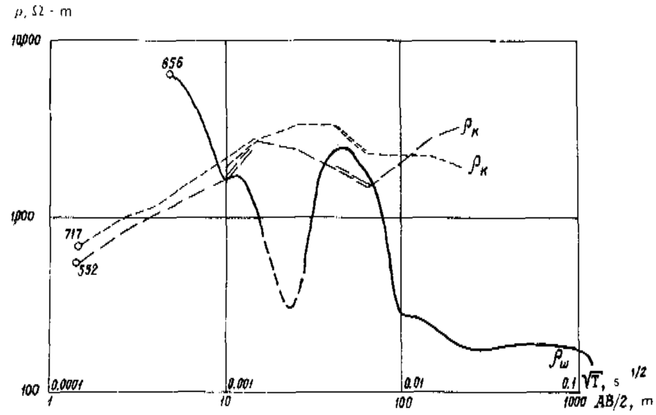

FIGURE 2 Field curves of frequency sounding and vertical electrical logging.

profiling (DEMP) at various frequencies in the limits from 80 Hz to 6 MHz. The undertaking of DEMP is feasible in the presence, in the section, of vertical or steeply dipping contacts during the mapping of faults, ore veins, ice pockets, etc. The preliminary undertaking of DEMP permits us to utilize the FS more effectively and under winter conditions, and it can yield a considerable economic effect (as compared with the direct current electric profiling), since it does not require galvanic contact with the ground.

In many cases, it is advantageous to conduct, along with the FS and DEMP, high-frequency logging of dry boreholes.[2] It can also be achieved with the aid of the FS installation equipped with

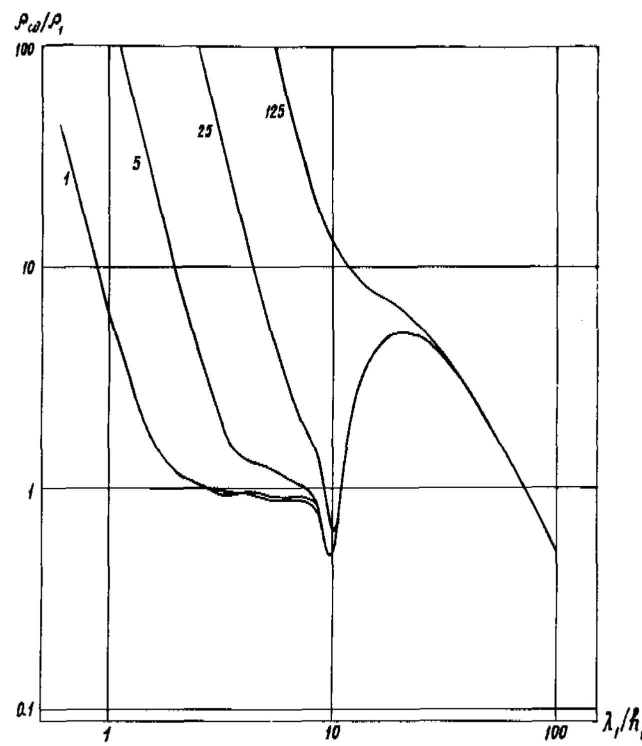

FIGURE 3 Theoretical curves B_z (μ = 9, r/h_1 = 20) for values ρ_1/h_1 = 1, 5, 25, 125 ($\eta_1 = \eta_2$ = 1) and without bias currents.

special antenna-type borehole probes.

The simpler cases when, based on the FS data, a two-layered geoelectrical section occurs have not been presented in the paper. Usually the corresponding ρ_ω curves are easily interpreted. Sometimes fairly satisfactory results are obtained with the utilization of master curves without considering the bias currents.

At the same time, it should be pointed out that the two-layered field curves ρ_ω are encountered less often than the VEL. The FS technique has a high resolving capacity and frequently permits us to register a layer not recorded by vertical electrical sounding.

The quantitative interpretation of multilayered curves so far has been encountering difficulties. In the B PNIIIC, we have derived the basic relationships[4] and for the first time a calculation has been performed (N. D. Rogulin) with theoretical data with consideration of the bias currents. However, the set of the only two-layered master curves so far available can by no means always provide a detailed solution to the problem; in connection with this, further study in this direction is under way.

In Figures 3 and 4 we have indicated the two-layered theoretical curves for B_z and E_x, respectively. Along with the generally adopted parameters ρ_2/ρ_1, r/h_1, λ_1/h_1, i.e., thickness of first layer ρ_2--resistivity of second layer, we have introduced additional parameters (in distinction from a quasistationary case): $\eta_m = \varepsilon_m/\varepsilon_0$--the equivalent dielectric constant of a layer with the number m and $\gamma = \rho_1/h_1 \sqrt{\varepsilon_0/\mu_0}$, where $\varepsilon_0 = (1/36\pi)\ 10^{-9}\ F/m$ (dielectric constant) and $\mu_0 = 4\pi \cdot 10^{-7}\ H/m$ (magnetic permeability) of

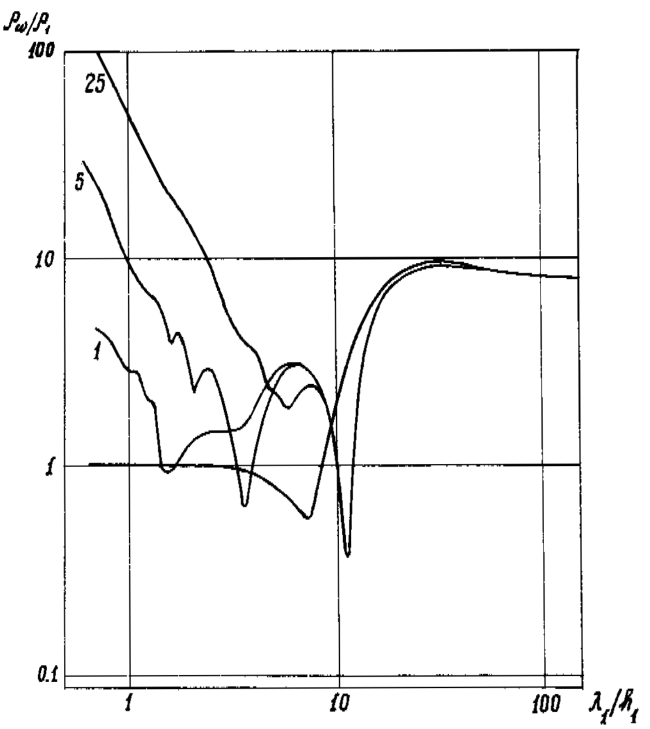

FIGURE 4 Theoretical curves E_x (μ = 81, r/h_1 = 20) for values ρ_1/h_1 = 1, 5, 25 ($\eta_1 = \eta_2$ = 1) and without bias currents.

a vacuum. In Figures 3 and 4, in place of γ we have indicated the value ρ_1/h_1, which show the numbers at the curves, while at the curve without the bias currents the numerical index is lacking. It is obvious from the illustrations shown that the ρ_1/h_1 value exerts great influence on the nature of the left-hand part of the theoretical curves. All the curves shown are calculated for the case $\eta_1 = \eta_2 = 1$. As test calculations have shown, the curves vary very little at different variations in η_1 and η_2. In this connection we ignored an investigation of the effect of frequency on dielectric constants, which obviously is very slight.

The theoretical and practical studies in the field of frequency sounding and the activities begun along the line of developing a shallow-depth modification have already yielded the material necessary for a successful utilization of the technique in solving certain geocryologic problems. Such familiar features of the method as the absence of complications in a study of the screening horizons, high resolvability, narrow limits of the effect of the equivalence principle, high productivity, and also great mobility of the shallow-depth model provide evidence of its advantages as compared with the direct current methods.

Frequency sounding can be useful in solving the following problems:

1. Large-scale geologic and geocryologic mapping.
2. Determination of the depths of upper and lower limits of the permafrost layer.
3. The classification of frozen and thawed ground in the geoelectrical section according to lithology and geocryologic criteria.
4. The outlining and prospecting of large geocryologic formations, including ice pockets.
5. The mapping of the contacts of frozen ground and thawed soil, etc.

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LOGGING IN SHALLOW DRY BOREHOLES FOR STUDYING GEOTECHNICAL AND GEODYNAMIC CHARACTERISTICS OF FROZEN SOILS

A. T. AKIMOV Industrial and Scientific Research Institute for Engineering Surveys in Construction

The existing traditional engineering-geologic methods for establishing the physical and physicomechanical properties of frozen soil in its natural bedding are laborious, do not have high instrumental accuracy, and are not adapted for the conduct of repeated measurements; in addition, their application involves a disruption in the natural bedding, water content, and pressure of the soils being tested. In connection with this, the requirement developed for improving the field techniques of study, i.e., the use of nondestructive methods by which the study of properties is accomplished without causing any disturbances to the condition of the soils. As is known, such techniques are considered to be the most precise. Among them we include the geophysical coring techniques. The initial experimental work on testing these methods has indicated that they can actually effect significant progress in the study of the geotechnical indexes of frozen ground.

We are familiar with the effectiveness of the petroleum, coal, and ore logging of boreholes. However, in the solution of prospecting problems, we usually perform well-logging of horizons occurring below the frozen layer. As a result both in the USSR and abroad, there are few boreholes investigated by well-logging in the frozen layer. In the same rare instances when standard logging is conducted within the limits of the frozen layer, for engineering-geologic purposes, as is known, they are not of great value, since they fail to yield the absolute values of the physical parameters of frozen ground and consequently cannot be used for determining the properties of soils. Therefore the results of such logging are utilized basically only for improved definition of the lower limit of the frozen layer.[6]

The specifics involved in permafrost studies require the creation of specialized well-logging differing from that being utilized in industrial geophysics. First of all, the measurements must be conducted in dry boreholes (also drilled dry) with the aid of special probes. For this purpose, probes with direct contact of the pickups with the borehole wall are applicable. The acoustic and electrical logging of shallow boreholes (20-30 m) can possibly be accomplished with the aid of brush-type electrodes, and pneumatic, spring-type, and wedge-type probes. For deep boreholes, electromechanical clamping of the pickups is best.

There are differences in the technique of measurements and in the interpretation of logging data: Most often the measurements are conducted according to a point system, while the features of interpretation include the calculation of the parameters' absolute values, the formulation of static and dynamic characteristics of frozen ground and the establishment of the relationships among them.

Among the methods without contact, we are aware of the thermometric, dielectric, and radiometric. Among the techniques without using contact, we also include frequency electromagnetic logging.[1] As a probe, we utilize here a miniaturized ferrite antenna, moving freely in the borehole. This permits us to conduct in detail continuous and automated measurements. The technique proves sensitive to changes in ice content and lithology. The improvement in the method, and particularly in the probe, permits a further increase in its resolving capacity.

The point electrical and ultrasonic logging in dry boreholes has been applied on the Bol'shezemel'skaya tundra and in western Siberia. By indicating the high values of apparent resistance (AR), the electric logging can identify the ice interbeddings, individual accumulations of ice crystals, and the ice-enriched horizons as a whole. All the AR diagrams reflect definite tendencies in the distribution of ice content by depth, differing for various types of tundra. Thus the sectors of mottled tundra are initially typified by an increase in resistance and then, starting from depth of 6.75 m, they are characterized by a decrease in resistance (Figure 1). The inversion pattern in ρ_k in the upper part of the section, explained in terms of the summer warming of the frozen soils, is lacking during the winter. In the sectors of peaty tundra, the upper part of curves reflecting ρ_k experience appreciable fluctuations, large gradients, and an appreciable divergence in the ρ_k and I values. This type of terrain is distinguished by high ρ_k values. The sectors of medium-hummocky dwarf shrub tundra containing a high temperature frozen layer are distinguished by relatively low ρ_k values and by a lower (without change in sign) gradient of ρ_k. The protective mossy-vegetative cover of this type of tundra protects the frozen layer both from summer warming, from which the branch of resistance increase is almost lacking, and from the deep winter freezing, a result of which are the relatively high negative temperatures of the frozen layer and the weak gradient of ρ_k. For each type of terrain, on the logging diagrams we can clearly observe the various thicknesses of the upper ice-enriched horizon. The heavy-gradient part in the pattern of curve ρ_k simultaneously segregates the area of significant seasonal variations in the physical characteristics of the frozen ground.

In this way, to each type of terrain differing in features of heat exchange at the surface with the environment, there exists a unique type of distribution of ice content and cryogenous texture, reflected by the appropriate type of pattern in the curve ρ_k, index, and sign of gradient content (state).

The acoustic (sonic and ultrasonic) measurements are possible in boreholes, open pits, and in core samples. In the latter case we determine the soils' coefficient of anisotropy. The possibility of measuring the longitudinal and Rayleigh (or transverse) waves permits us to calculate the dynamic modulus of elasticity, E_d (Figure 2), and the Poisson's ratio σ. As experience has shown, curves σ and E_d characterize essentially the strength component for the properties of the frozen layer, while the amplitudes and frequencies of fluctuations that are being recorded reflect to a greater extent the variation in plastic component of a complex elasto-

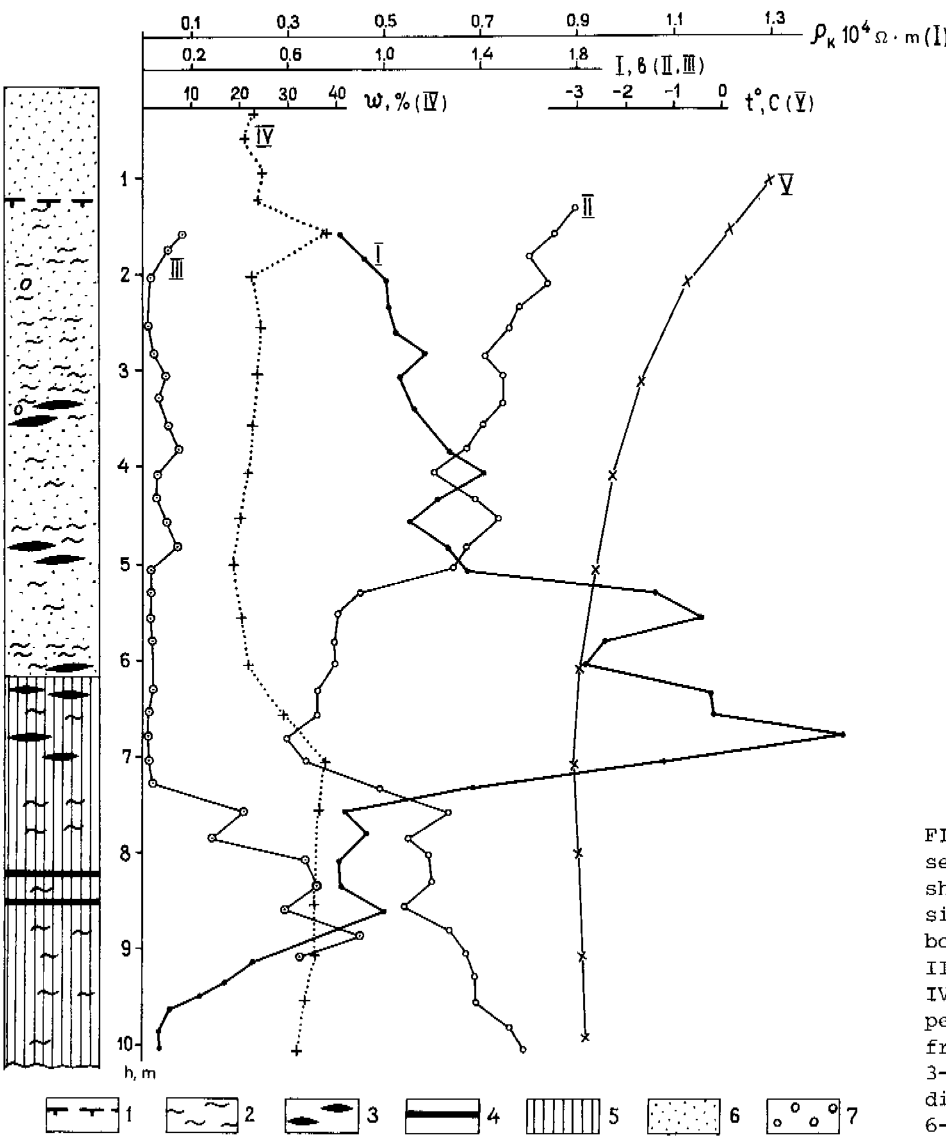

FIGURE 1 Logging diagrams of sector of mottled tundra (Bol'shezemel'skaya tundra). I--resistance; II--current in borehole filled with water; III--the same in dry borehole; IV--moisture content; V--temperature. 1--upper limit of frozen layer; 2--ice crystals; 3--ice pockets; 4--interbeddings of ice; 5--moraine clays; 6--sand; and 7--boulders.

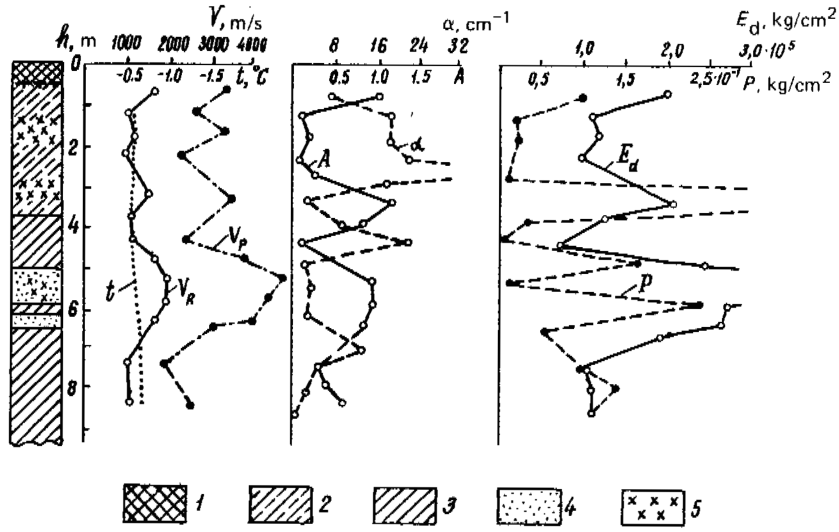

FIGURE 2 Diagrams of ultrasonic logging (Bol'shezemel'skaya tundra). v_p = propagation velocity of longitudinal waves; v_r = the same, of surface waves; t = temperature; A = amplitude of surface waves; α = attenuation coefficient; E_d = dynamic elastic modulus; and P = generalized acoustic parameter. 1--peaty cover; 2--mantle suglinoks; 3--moraine suglinoks; 4--sand; and 5--ice crystals.

viscoe-plastic body such as frozen soils represent. The generalized acoustic parameter P (refer to Figure 2) portrays the combined elastoplastic properties:

$$P = \frac{E_d f}{\alpha v_p} = \frac{E_d}{\alpha T v_p} \text{ kg/cm}^2$$

where T = periodic time of the wave being recorded, expressed in seconds; α = attenuation factor, cm^{-1}; and f = frequency, kHz.

With the aid of seismic-acoustic studies for the Bol'shezemel'skaya tundra, it was established that the propagation velocity of longitudinal waves, depending on lithology, ice content, and temperature of frozen ground, varied from 600 m/s to 4,000 m/s and more; ratio v_p/v_s varied from 1.9 to 2.4; the dynamic Young's modulus varied from $0.1 \cdot 10^4$ to $2.6 \cdot 10^5$ kg/cm², and the Poisson's ratio ranged from 0.23 to 0.5.

The logging data prove useful in three directions of scientific and industrial field investigations: geotechnical, general permafrostological, and geophysical. Let us consider them separately.

1. The logging of dry boreholes is of the maximum value in a study of the construction indexes of frozen ground. Based on the curves ρ_K, ε, v_p, v_S, and v_R, we present the qualitative characteristic of the ice content in the soils and we establish the type of variation of ice content of soils with depth. Simultaneously based on the materials from ultrasonic logging, we calculate the quantitative step-by-step approximation of the values for α and E_d. Based on the indexes for the damping of waves and on their frequency spectrum, we develop concepts concerning the frozen ground's consistency.

The gamma densimeter and neutron-type moisture gauge permit us to find density Δ and moisture content w of soils.

Based on the apparent resistance and well potential curves (potentials of natural field) and I, we refine the position of the lower limit of the thin frozen layer or the transition from the coarse to the fine soils. The values α, E_d, v, ρ, ε, Δ, and w as design parameters can be utilized in predicting the variations in the bearing capacity of frozen ground, seismic microregioning, designing of special installations, design of thermoelectro-mechanical facilities and mechanisms for the excavation of frozen ground, engineering designs of high-frequency and electric-"needle" heating of frozen ground, and finally in the adoption of solutions concerning grounding and relay installations.

It is remarkable that all the above-enumerated parameters, as compared with the widely utilized temperature characteristic (Figure 1 and 2), contain more information not only concerning the properties, but also concerning the condition of frozen ground.

A combined analysis of the data from all types of logging assures more complete, objective information concerning the qualitative and quantitative indexes for the properties of frozen soils.

2. The logging of dry boreholes can comprise large-scale determination of physical constants of frozen soils, with the aid of which we easily establish the type of relationship between ice content and depth and the nature of the distribution of cryogenous textures in plan. For practical purposes, the indexes of these relationships can be utilized in the microregioning of a territory for construction purposes, in the small- and large-scale mapping of permafrost, and for the appropriate representations on the survey and schematic maps. Simultaneously, the results from logging measurements assure fairly complete data concerning the thickness of the most icy horizon and concerning the thickness of the entire frozen layer. All these aspects taken together promote an increase in the level and quality of general permafrostological research.

3. In interpreting the data from ground and air geophysical (electrical prospecting, seismic,

and radar) studies, we require a knowledge of the values of the parameters ρ, ε, v, w, and t of individual horizons of the entire frozen massif. In addition to this, data concerning the indexes and sign of the gradient of variations in these parameters by depth and concerning the anisotropy of frozen ground prove useful. All these data provided by the logging of dry boreholes predetermine the selection of the methods of interpretation and its reliability.

What has been said above pertains to the one-time or recurrent simultaneous determinations of the physicomechanical characteristics. An equally high value is acquired by the stationary (regime) geophysical observations in boreholes. In recent years, from the results of geophysical research, it has become known that the physical parameters of a frozen layer vary within wide limits during the course of a year. Especially high are the annual variations in the resistances of the uppermost horizons of the frozen massif. According to the data compiled by V. S. Yakupov,[5] the resistance of frozen unconsolidated Quaternary deposits in the northeastern USSR for the period from August to December varies by 2.65 times. Based on the results of studies by V. A. Kirillov,[2] the resistance of frozen Quaternary soils at a depth of 2 m during the year varies from $1.3 \cdot 10^3$ to $9 \cdot 10^5$ Ω·m. The range of variation becomes narrower with depth: At a depth of 5 m, ρ varies from 10^4 to $4.5 \cdot 10^5$; at 10 m depth, it changes from $6.5 \cdot 10^4$ to $2.6 \cdot 10^5$; and at 20 m depth, it ranges from $1.9 \cdot 10^5$ to $6.4 \cdot 10^5$ Ω·m.

Based on measurements on a stationary installation of vertical electrical sounding in the middle reaches of the Yana River (V. F. Slazkin), the resistances of the frozen loose materials during the period from August to January increase by 5 times.

For the Krasnoyarsk regions characterized by deep winter freezing (3.8 m) and by the penetration of seasonal fluctuations in temperature (10-15 m), density of frozen loesses from October to January varies from 1.1 to 2.0 g/cm^2.[3]

The range of seasonal variations in temperature in the frozen layer differs in various regions. For example for the Chokurdakh region (Yakutia) near the top of the frozen layer, it falls in the limits from 0° to -18°, and for the Bol'shezemel'skaya tundra, it ranges from 0° to -4°. In the other regions of the permafrost area, it usually occupies an intermediate position. The alternation in the sign of the gradient of the summer temperature curve (point of contraflexure) establishes the depth of attenuation of seasonal variations in the physical characteristics for the upper horizon of the frozen stratum. The maximum divergences in the summer and winter temperature curves encompass the range of appreciable seasonal and many years' variations in the frozen stratum's physical parameters.

The data presented convincingly indicate that during a year, under the effect of the variations in temperature, there occur deep transformations of the frozen ground: There is a gradual transition of frozen soils from a Hooke body to a St. Venant body and vice versa and there is a change in porosity, moisture content, and the general strength and deformation-type properties. In addition to the annual climatic changes, there exist perennial (8-12 yr) changes associated with the cycle of sunspot formation. The annual variations are imposed on the many years' ones; this intensifies still more the geodynamic tendencies in the frozen series.

The significant annual and perennial variations in the physicomechanical characteristics of frozen soils lead to appreciable variations in the bearing strength of the soils in a foundation, since these parameters have a direct relation with ice content, cryogenous structure of frozen materials, variation in rheologic properties, etc.

With reference to the loesses in the Krasnoyarsk region, the depth of significant fluctuations in these parameters established by radioisotope techniques equals 2.5 m, while the depth of their complete attenuation equals 4.5 m (Figure 3). For the northeastern USSR, these depths equal, respectively, 1 m and 30 m (Figure 4).

Within the same region, these depths can vary depending on lithology, hydrogeologic conditions, and conditions of heat exchange at the surface (on types of terrain, types of structures, and the exposures of their walls). It is known that stresses in buildings develop not only from settlement during thawing, but also as a result of the heating of frozen ground under the building.[4] The stresses in buildings in such cases are caused by temporary variations in the physicomechanical constants of the frozen layer.

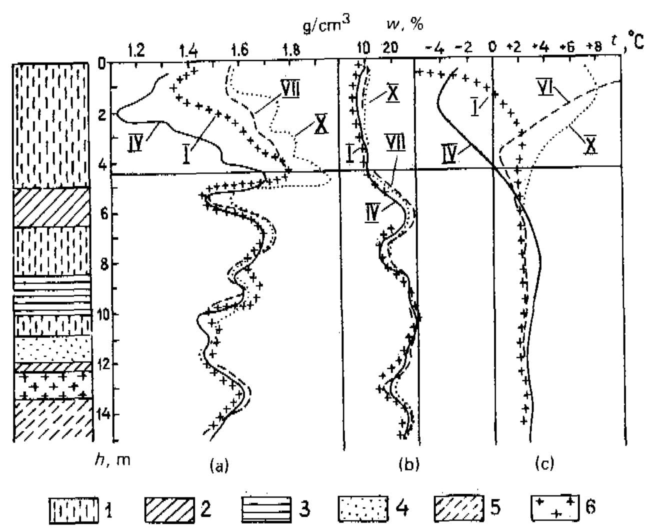

FIGURE 3 Range of significant variations in density, moisture content, and temperature of loess-like soils in the Krasnoyarsk region according to A. V. Minervin and Ye. M. Sergeyev.[3] 1--loess; 2--suglinok; 3--clay; 4--fine-grained sand; 5--coarse-grained sand; and 6--ice crystals. (a) density; (b) water content, and (c) temperature. Curves: I--January; IV--April; VII--July; X--October.

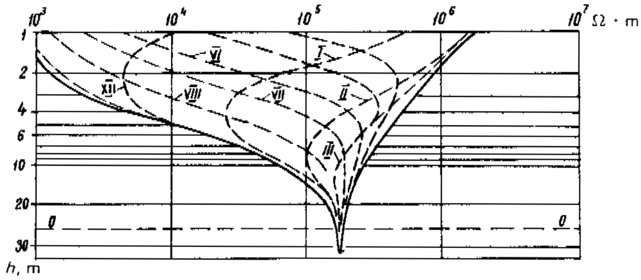

FIGURE 4 Area of significant variations in resistance and attenuation depth of fluctuations (northeastern USSR) according to V. A. Kirillov.[2] 0-0 = limit of zero annual amplitudes in temperature fluctuations. Curves: I--January; II--February; III--March; VI--third 10 days in June; VII--July; VIII--August; and XII--December.

On the other hand, the structures that prove to be stable are those that are designed with consideration of the depths and amplitudes of appreciable fluctuations in the frozen materials' physical characteristics.

In this manner, a knowledge of the limits and depths of the variations in the physicomechanical parameters for each type of terrain in the territories that are being developed and for each type of structures is important for the correct designing of structures, the provision of their long service life, and also for estimating the variations in earthquake forces in the seismic regions.

From what has been said, the practical feasibility of a spatial (by depth and in plan) and time-related study of the physicomechanical parameters of frozen ground in its natural occurrence becomes clear.

The available experience gained in geophysical measurements in dry shallow boreholes permits us to make the following conclusions:

1. The combination of the well-logging techniques in boreholes provides exhaustive, objectively precise information concerning the physicomehcanical properties of frozen soils to depths of interest in construction practice.

2. A comparison of the data from static techniques of determining the strength and deformational properties of frozen soils with the dynamic properties will permit us to correlate regional relationships between them; this will facilitate the obtaining of extensive data concerning the geotechnical characteristics of frozen soils.

3. The logging investigations in stationary boreholes provide all the necessary data concerning the annual (or perennial) dynamic tendencies in the upper horizons of the frozen layer and in the layer of deep winter freezing exerting an effect on the supporting capacity of the frozen ground in the base.

4. The most efficient combination of techniques used in geophysical observations in boreholes should include the electrical (apparent resistance, well potential and *I*), ultrasonic, dielectric, nuclear, and thermometric logging methods. The utilization of the logging of induced potentials and also telemetry logging is promising. The processing of observational data can feasibly be accomplished with the use of the statistical-stochastic methods.

5. For purposes of greater effectiveness, the study of the construction indexes of frozen soils should be conducted with combined geophysics and mechanics in the field of permafrostology.

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SEISMICITY OF THE CRYOLITHIC ZONE AND PROBLEMS OF ENGINEERING SEISMIC GEOCRYOLOGY

V. P. SOLONENKO Institute of Earth's Crust SO AN USSR

The question concerning the unique engineering-seismological conditions in the permafrost area has been raised long ago,[9] but special purposeful studies began only in 1966 on the eastern flank of the Baikal rift zone, where we find a combination of high seismic activity* and complex permafrost, the thickness of which varies within short distances (kilometers) from 0 to 1,100-1,300 m.[6,7] These and subsequent investigations conducted under the supervision of O. V. Pavlov in the Barguzin rift valley and in the Trans-Baikal region, coupled with laboratory experiments, confirmed the assumption concerning the sharp contrast in the variation of seismic degree of tremors depending on the permafrost conditions. The necessity has developed to differentiate in the seismic region a special *engineering-seismic-geocryological* area. In the Northern Hemisphere (the Antarctic is practically nonseismic), it occupies an area of more than 10 million km^2, i.e., about half the entire area of the permafrost zone, including about 3.5 million km^2 in the USSR territory (Figure 1), while the remainder is found in North America (Canada, Alaska), northeastern Greenland, and on the islands in the Arctic Ocean.[5,16-20]

The effect of seismic processes on permafrost is particularly distinct and direct in the seismogenic structures that during severe earthquakes experience displacements in vertical and horizontal directions for many meters. At this time the old fault zones become exposed and new ones develop, which provides conditions for the formation of anomalous endogenic (rising) and exogenic (descending) heat flows.

The Baikal rift zone and the central part of Alaska† are the most highly seismic in the permafrost area. The Baikal rift zone belongs entirely to the regions with earthquake intensity of 9-10 points and higher (M = 6 1/2).[11]

*Earthquakes: 27 June 1957, X-XI on the intensity scale, M:7.9; 5 Jan. 1958 and 14 Sept. 1958, intensity IX, M:6.5; 18 Jan. 1967, intensity IX-X, M:7. Note by technical editor: The magnitude scale (M and arabic numerals) is the Richter scale used in North America. The intensity scale (Roman numerals) used in the USSR is almost identical with the Modified Mercalli scale, which is used in North America

†The most severe earthquakes in Canada and Alaska (M to 8.7, Intensity XII) are linked with the Pacific Ocean seismic belt and their epicentral zones lie beyond the boundary of the permafrost area.

The cryolithic zone is also unusual here. In the rift valleys there occurs frozen ground of the "Baikal type".[9] Even near the southern boundary of the cryolithic zone, the persistently frozen soils occur to depths of hundreds of meters. There is a two-layered structure in the section of frozen ground. The upper horizon of contemporary, chiefly discontinuous, frozen ground has a depth to tens of meters, rarely more than 100 m (in the northern depressions, based on geophysical data, up to 200-250 m). The lower permafrost horizon is relict and syngenetic. It apparently was formed at the boundary of the Holocene, when the climatic conditions favored the development of permafrost, while the high mobility of the Earth's crust typified by seismic activity promoted its sinking and burial. These horizons rarely merge and as a rule are separated by a horizon of thawed soils with a thickness ranging from tens to hundred of meters.

The existence of permafrost of the Baikal type is so unusual that certain permafrostologists deny the possibility of its existence. It has not been reflected even in the most recent geocryological map of the Trans-Baikal region.[7] At the same time, its presence is confirmed by recent drilling and geophysical data.[1,4]

Frozen ground of the Baikal type can be encountered not only in the rift zone, but also in other geologic structures having experienced considerable subsidence at the end of the Pleistocene beginning of Holocene periods.

In Alaska a seismically and highly active zone with average frequency of IX-X earthquakes (M = 6 1/2-7 1/4) once every 12 yr intersects the peninsula from southeast to northwest (between Prince William [Sound] and Kotzebue bay), running along a diagonal through the alpine Alaska Range and the basin of the middle reaches of the Yukon River. Just in the southeastern third, this zone encompasses the regions of insular* permafrost or of permafrost with islands of taliks (types II and III) and in the remaining extent is located in the zone of continuous permafrost (type IV).

The dependence of the cryolithic zone on the type and nature of movement of tectonic structures has been confirmed by geothermal observa-

*The author divides discontinuous permafrost into two types: (a) having islands of permafrost in thawed ground; (b) having islands of thawed ground (talik) in frozen ground (permafrost).

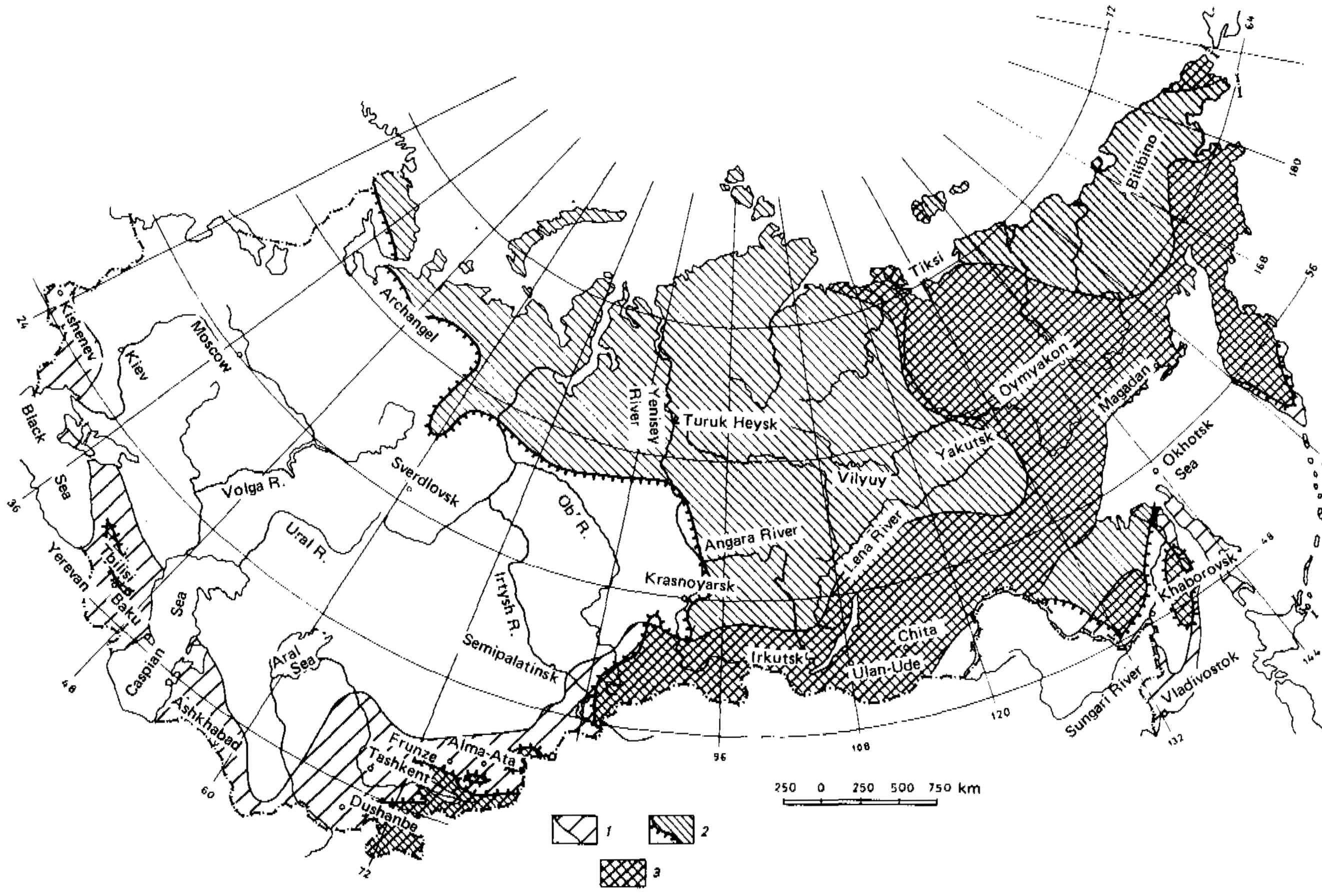

FIGURE 1 Distribution of permafrost in the seismic zone of the USSR. 1--seismic area (intensity of earthquakes greater than or equal to VI; 2--permafrost area; and 3--seismic regions in permafrost area.

tions.[2] The lowest value of heat flows (of around 300 $kcal/m^2 \cdot yr$ and maximum thickness of permafrost have been established in the negative structures. The heat flow is considerably higher, i.e., up to 4,000-26,000 $kcal/m^2 \cdot yr$ in the positive* structures and maximal, i.e., up to 160,000 $kcal/m^2 \cdot yr$ and more along the zones of tectonic disturbances. Therefore, in the seismically highly-active regions, even in the frozen layer of low-temperature (up to -10° and lower), deep (up to 1,100 m and more) permafrost, we encounter narrow but extended taliks of seismogenic lineaments. Frequently they can be identified well by the vegetation (thicker and more thermophilic as compared with the vegetation in a number of sectors located nearby). However, there can be exceptions to this general rule. The inactive faults in the pattern of seismogenic movements of the Earth's crust often open passively and into them cold water can flow; this promotes the development of permafrost and the formation of wedge-ice to considerable depths (on the Udokan ridge on the eastern flank of the Baikal rift zone, ice wedges with thickness up to tens of centimeters are encountered at depth of around 700 m).

*Below and above 0°C, respectively.

The permafrost greatly complicates seismic regioning. In the regioning, the greatest significance is acquired by the proper selection of the combination of permafrost and seismic characteristics.

In a general geographic demarcation on a map, it is impossible to take all the diversity of permafrost conditions into account. Therefore, it is inevitable that one must utilize a coarse classification geocryological system. As an initial experiment, we suggested its following modification (Figure 2): I--individual islands; II--insular; III--with islands of taliks [(a) block type, (b) cellular or mosaic]; IV--continuous; and V--Baikal type.

The seismic properties of permafrost depend not only on type but temperature, etc. These relationships are heterogeneous, and many of them have not yet been studied. In a first approximation, the frozen ground is subdivided by seismic properties into: (1) solidly frozen, (2) plastic frozen, and (3) pseudo-thawed and loosely frozen. Among the solidly frozen soils, we should include those which are similar to rocks in velocities of longitudinal seismic waves (3.4-4 km/s and more) and in amplitudes of oscillations. Their temperature is below -2° or -3°. The loosely frozen and pseudo-thawed soils have a temperature above -1°, and their seismic properties are approximately the same as in the actually thawed soils.

With the accumulation of actual material, the possibility will be presented of conducting a standardization of the seismic permafrost-lithological complexes. Their boundaries by no means always coincide with the boundaries of the engineering-geological microregions. For example, in one seismic lithologic permafrost complex, there can be plastically frozen coarse-grained beds and solidly frozen silty soils (the velocities of longitudinal waves and amplitude-frequency characteristics in them can be practically the same). However, in the transition to a pseudo-thawed state, the intensity of the first will increase by 1, and of the second by 2-3; the bearing capacity and other geotechnical properties will be different.

Each of the identified types of permafrost has its unique specific engineering-seismic geocryological features.

Type I. The thickness of permafrost in its

FIGURE 2 Diagram showing seismic zoning of USSR permafrost area. Seismic regions (1-4): 1--intensity of VI; 2--intensities of VII and VII + VIII; 3--intensity of IX and higher; 4--limits of seismic regions (marks on side--in direction of regions of decreased intensity). Permafrost regions (5-10): 5--individual islands of plastically frozen and pseudo-thawed soils; 6--insular type of permafrost with plastically frozen and pseudo-thawed soils; 7--permafrost with islands of taliks, with complexes of solidly frozen and plastically frozen soils; 8--continuous permafrost with solidly frozen soils; 9--limits of permafrost regions (name of regions and condition of soils--according to prevalence); and 10--limits of permafrost area.

individual islands is usually less than 15 m to 20 m. In the thin (few meters) pockets of permafrost on passage of the seismic waves, the observed resonance phenomena set in abruptly. In the presence of water-soaked soils below them, which are capable of mud eruptions, the permafrost lenses can receive additional low-frequency vibrations. We do not exlcude the shattering of beds and lenses of frozen ground, the settling of individual blocks, and rapid degradation. Therefore, construction on permafrost islands in seismic regions is undesirable.

Type II. The insular permafrost is usually at a high temperature, while the soils are plastically frozen and pseudo-thawed. As compared with the rocky and solid-frozen soils, intensity of earthquakes increases by 1 to 3 points on the scale. Particularly dangerous are the silty ice-rich soils with stratified and reticular texture.

With a permafrost thickness of around 20 m to 40 m or included in it, interbeddings of thawed or pseudo-thawed soils, a resonance-type phenomenon, and an increment in intensity of 2 to 3 points are likely. In the presence of taliks and pseudo-talik funnels, cumulative processes are possible, which can involve a repeated increase in the amplitudes of oscillations and the liquefaction of soils in the central parts of the taliks.

Construction on permafrost of this type is also undesirable, but it can prove to be unavoidable since it often occurs in the most convenient floodplain areas in the river valleys and intermontane areas.

Type III. Frozen ground with islands of taliks can prove most complex (especially the cellular subtype) for engineering-seismic-geocryological prospecting and zoning, as a result of the complicated combination of solidly frozen, plastically frozen, pseudo-thawed and thawed soils with sharply differing seismic qualities, and the actual morphology of the cryolithic zone favors an extensive and diversified manifestation of resonance and cumulative processes.

In the block subtype of permafrost, most seismically dangerous are the talik "lineaments" in the region of the fault, as a result of the total increase in intensity owing to the thawed ground and the advances along the dislocations with discontinuous breaks.

The variations in intensity, as compared with the standard in the regions with type III of permafrost, can fluctuate from -1 and -2 to +1 and -3 points on the intensity scale.

Type IV. Solid permafrost chiefly with hard-frozen soils supporting structures (in the case of construction with retention of permafrost in the pile-type foundations with ventilated subfloor). Under these conditions the standard intensity can be reduced by 1.

The nature of operation of piles in the two- (in winter) or three-layered medium, markedly heterogeneous from a seismic standpoint, remains unclear. Fluctuations in the foundation occur owing to the upper part of the soils, while the amplitude of vibrations rapidly decreases with depth. All these factors can lead to intensive shearing stresses in the piles. However, the latter can play the role of a flexible footing, which, given specific design features of the structures, has a positive effect on their seismic stability.

As a rule, talik-type slump holes develop under structures with hot technological processes; this abruptly deteriorates the seismic conditions.

Type V. In the regions with the Baikal type of permafrost, there develops a series of specific engineering-seismic geocryological problems. The multilayered structure of an unconsolidated layer filling the depressions of the Baikal type can abruptly alter the isoseismic field. Obviously, this explains the appreciable divergence in the intensity, not only in the closely located points, but also in individual parts of the same structure, which we repeatedly encountered during the investigation of the consequences of severe earthquakes.

Prior to the special studies of the seismic properties of the regions with the Baikal type of permafrost, it is necessary to become oriented with respect to the properties of the upper (contemporary) permafrost horizon with consideration of the fact that the subjacent thick (hundreds of meters) beds of loose and frozen sediments reduce the intensity of seismic impacts by at least 1.[10,14]

During seismic zoning of the permafrost area, it is necessary to compile either a composite permafrost-seismic map (see Figure 2) or to attach a geocryologic map of the same scale to the basic map.

It is most efficient to conduct detailed seismic zoning on the basis of a morphostructural composite engineering-geologic and geocryological demarcation.

During seismic microzoning, in accordance with the basic construction principles (with retention or elimination of permafrost), we suggest compiling two maps: for natural conditions and for thawed soils.[8,15]

During zoning for the actual construction complexes, we can compile a specialized map with consideration for the new postconstruction permafrost conditions, or temporary maps based on stages of developing the territory, taking into account that temperature, ice content, ratio of solid and liquid phases of water, and porosity and density of soils during the development of the territory and in the course of a year constantly vary both reversibly and irreversibly. The seismic qualities of rocks also vary accordingly: acoustic rigidity, amplitudes and periods in fluctuations of soils, resonance and cumulative properties, and also the nature of the interaction in the soil-structure system.

Earthquakes actively influence the pattern and especially the tempos of geocryological processes and occurrences: They accelerate solifluction and facilitate the movement of placer deposits; the large-scale sliding of the active layer disrupts the established thermal regime in the upper horizons of permafrost, causes its degradation, with all the ensuing consequences, and promotes the formation of earth avalanches. At the same time, permafrost determines to a considerable degree the nature of the seismic-gravitational phenomena, consolidating the mountainous masses on the slopes and by the same token decreasing the number of landslides, cave-ins, and earth and rock avalanches, especially during the earthquakes that occurred during the total freezing of the active layer.

Indicative along these lines was the Oymyakon earthquake on May 18, 1971 (IX, *M*-7), the epicenter of which was not far from the "Cold Pole" in the Northern Hemisphere. In the isoseismic zone in an area of 3 by 6 km along the fault zone, there occurred massive strippings of the thawed part (≈0.3 m) of the active layer. The soil-vegetative mass in the valleys of the steep water-courses formed torrentlike mudflows with a depth up to 5-6 m. On the gradually dipping slopes (up to 15°) and horizontal areas, there occurred massive liquefaction of fine detritus;[3] however, it was not accompanied by appreciable subsidences as occur in the regions away from the permafrost. As a whole, owing to the permafrost, the seismic-gravitational effects during this earthquake occupied a smaller area and had smaller dimensions than during identical earthquakes in the regions outside the permafrost zone.

Also remarkable is the distribution of shocks over the Earth's surface. In the region of continuous permafrost, the intensity of the shock weakened regularly. At a distance of 250 km to 350 km, it had decreased 4 points on the scale and continued to attenuate further; at an epicentral distance of 450 km to 500 km, on emergence into the region of insular permafrost near the Sea of Okhotsk (Figure 2), intensity of the shock increased at least by 1 to 2 points on the scale. A 4-point shock was registered on the coast of the Sea of Okhotsk at a distance of about 700 km. The relationship of the intensity of shock with type of permafrost in the given case is graphic and undisputed.

At places in the seismogenic fissured zones, very thick naleds form, ranging from 10 m to 15 m and more in the Syul'banskaya, China-Vakatskaya seismogenic structures in the rift system of the Stanovoye upland.[14] During earthquakes, there sometimes occurs an intensive "volleylike" outpouring of water; under winter conditions this causes a catastrophic formation of gigantic naleds. Thus, during the Gobi-Altay earthquake on December 4, 1957 (Intensity XII, *M*-8.6), a naled with a length of more than 10 km formed.[13]

In the sectors with extremely icy soils, the earthquakes cause or activate the thermokarst processes. This is supported either owing to the formation of the fissures along which the ground and surface water flows into the permafrost or by the downthrow of the active layer and the outcropping of the permafrost. Powerful seismic shocks are not essential for this--a series of weak earthquakes could cause the same effect. For example, in 1967 in the vicinity of Lake Leprindo on the Stanovoye upland in the epicentral area of a swarm of earthquakes (85 shocks) of moderate intensity, there suddenly formed a thermokarst-mudflow ravine with a length of 650 m and width of 10 m to 15 m with a depth to 6 m.

Complex problems have risen before engineering seismic geology in general[12] and under permafrost conditions in particular. The existing techniques of seismic zoning have been developed on the basis of materials for the seismic regions in our southern republics and foreign countries with a positive geothermal regime of soils. The engineering-geologic and instrumental seismologic observations provide parameters of the soils inherent to them during the investigation. During a further engineering-seismic-geologic interpretation, we envisage the stability of these parameters. However, under permafrost conditions, the geotectonic and seismic properties of soils in the influence zone of engineering structures vary constantly. Therefore, the existing developments in respect to a technique for seismic zoning, particularly microzoning, in the permafrost area are to a considerable extent unacceptable.

The ever-increasing rates of developing the seismic regions in the permafrost area are vitalizing a new scientific trend, namely that of engineering-seismic geocryology.

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PRINCIPLES OF COMPILING SMALL-SCALE GEOCRYOLOGIC MAPS

A. V. GAVRILOV AND K. A. KONDRAT'YEVA Moscow Lomonosov State University

The collation and generalization of information concerning frozen strata in cartographic form have great significance as a result of the capacity and descriptiveness of this technique for presenting material. At the present time, in proportion to the expansion of knowledge concerning the tendencies in the formation and development of frozen strata, the content of the general-purpose permafrost map is becoming more and more complex. The map is becoming the basic source of information. It predetermines firstly the requirement for the development of the principal bases of permafrost mapping. Secondly, the increase in the information content of a map even now is in contradiction to its readability. Therefore, the efforts toward efficient ways of reflecting the content of a map are becoming an inseparable part of the work on its compilation.

These problems are acquiring particular urgency in connection with the fact that one of the essential problems in regional permafrostology in the immediate years is the compilation of geocryological maps of the USSR. The topic of this lecture is a review of the principles and methods of depicting the content of general-purpose permafrost maps for various regions on a scale of 1:2,500,000, compiled according to a standardized legend; in their development, we took into account the requirements that could be imposed in the legends of USSR maps on this scale.

The vastness of the information and the requirement to preserve the legibility of the maps prevent us from accomplishing the portrayal of the permafrost content in complete scale on one map. Therefore, proceeding from the features involved in the formation of seasonal and perennial permafrost, it is feasible and logical to compile two maps, i.e., maps of the perennially frozen soils (permafrost maps) and maps of seasonally frozen and seasonally thawed soils. The essential items synthesizing the content of these maps would be the natural-historical complexes characterized by the affinity in the composition, structure, and basic parameters of the seasonally and perennially frozen soils, reflecting the generality of the thermodynamic and geologic-geographic conditions of their formation and history of development. Such complexes on the permafrost map are the types of perennially frozen strata,[6] while in the map of the seasonally thawed and seasonally frozen soils, such complexes are the types of seasonal thawing and freezing of soils.[2]

The basic principle in general permafrost mapping, consisting in clarifying the tendencies in the formation and development of frozen strata, assumes the portrayal of their composition, structure, and basic characteristics as a result of heat exchange under specific geologic-geographic conditions. Therefore, the chief problems that need to be solved during the development of the legend and compilation of the maps were as follows:

1. Ways of clarifying on the map the natural conditions determining the conditions of heat exchange on the surface and in the stratum of rocks.
2. Development of methods for an areal representation of the composition and structure of frozen strata.
3. Development of techniques for indicating their basic parameters.

For purposes of a cartographic representation of the *natural conditions*, we utilized their subdivision, based on considering the energy factors determining the formation of the actual natural conditions and the basic relationships in their spatial distribution (zonal, regional). In this respect, we differentiate two groups of natural conditions:

1. Exogenic or physicogeographic (climatic, geobotanic), expressing the conditions of heat exchange at the Earth's surface.
2. Endogenic or geologic (geologic-structural, hydrogeologic, and deep-seated geothermal) controlling heat exchange in the layers of rocks. Among this same group we include the geomorphological conditions determining the high-level differentiation of heat exchange at the Earth's surface.

The generality of the laws for the territorial distribution of natural conditions of each of these groups constitutes the result of the existence of close interrelationships between them and occasions the possibility for utilizing or constructing classification systems for their cartographic representation.

At the basis of the cartographic representation of *exogenic conditions*, there should lie the principles of the distribution of solar radiation and circulation of air masses, causing heat exchange at the Earth's surface with the atmosphere. Therefore, the presentation of these conditions is accomplished by us on the basis of a system of climatic zoning for the USSR developed by B. P. Alisov[1] based on the indication of the receipt and expenditure of solar heat and circulation of the atmosphere as concerning the determinative climate-forming processes and taking vegetation into account as an indirect qualitative characteristic of climate. The territorial units of the physicogeographic zoning are shown on the inset map, in the legend of which we present the region-by-region description of climatic conditions (radiation-heat balance and its components, mean annual temperature and annual amplitude of air temperature, depth of snow cover, annual amount of precipitation and type of high-altitude zonation) and basic modifications of the vegetative cover.

The cartographic representation of *geologic conditions* should be based on an indication of the tectonic conditions in a region. This is associated with the fact that, within the limits of the tectonic structures, the tendencies in geologic development and consequently the geologic-geomorphological, hydrogeologic, and deep-seated geothermal conditons are distinguished by variability. Particularly significant is the role of explaining in permafrost maps the most recent stage in geologic development, having determined the contemporary nature of the conditions listed and commensurate with the formation time of the frozen strata.

The representation of geologic conditions with the aid of notations for the tectonic structures presumes the segregation of the latter in such a way that within their limits were generally distinguished by: first, the high-level characteristics of the terrain determining the heat-exchange conditons at Earth's surface; secondly, the thickness of unconsolidated sediments, degree and nature of the fissuring of soils causing the features of heat transfer in their layer; and thirdly, the deep-seated geothermal conditons and types of hydrogeologic structures determining the transport of heat within the ground and the nature of heat exchange in the layers of rocks. The tectonic zoning for the permafrost mapping is based on the differentiation of contemporary structural elements in the Earth's crust (platforms, orogenic regions, and contemporary geosynclines),[3] separated according to the conditions of heat exchange at the surface in the rock layer. Within the limits of the differentiated contemporary structures, there have developed the complexes of formations corresponding not only to the Cenozoic stage of their development, but also reflecting the preceding stages of their tectonic history. The separation of types of formations, depending on the tectonic conditons of the rocks' formation (platform, geosynclinal), promotes a complete consideration of the material features in the contemporary structures and a clarification of the deep-seated geothermal and certain features in the geologic-geomorphological and hydrogeological conditions. The cartographic representation of contemporary structures has been based on a consideration of the relationship with them of the soils' composition and is, therefore, conducted during a description of the methods for representing the latter.

Another even more pressing problem confronting permafrost mapping on a scale of 1:2,500,000 is the question of representing the composition and structure of frozen strata. The gist of permafrost mapping envisages the presentation of such groups and modifications of soils according to composition, genesis, and age, which, being generally adopted in geology, would have taken into account the difference in the effect of soils on the pattern and results of heat exchange in their stratum. The specifics of soils as a medium in which the heat-exchange processes transpire are expressed, on the one hand, in the varying nature and mechanism of ice release in the process of soils' freezing and, on the other hand, by the varying nature and conditions of heat transfer in the massif (thermophysical properties, possibility and extent of participation of convection, etc.).

In respect to the influence on the heat-exchange processes, rocks and soils of varying composition are subdivided primarily by nature of structural bonds, including those acquired during their freezing.[4] In this respect, we differentiate rocks with rigid bonds existing independently of the aggregate condition of water in them (rock and semirock), and rock without rigid bonds, in frozen state characterized by the presence of cryogenic bonds (soils).

The differences in these two basic groups of rocks are very great; however, they have not yet found proper graphic representation in the compilation of permafrost maps. Therefore, the search for ways of graphic portrayal in the differences in soils, on the one hand, and of rocks and semirocks, on the other hand, is one of the basic problems that has to be solved in developing techniques for representing the composition of rocks.

The principles and techniques of representing the composition and cryogenic structure of soils on the maps based on the suggested legend consist of the following aspects:

1. The basis for the mapping of unconsolidated Cenozoic deposits is the differentiation of their genetic types. This is associated with the fact that, first, the genetic type serves as a basic classification unit in studying and subdividing the unconsolidated rocks in Quaternary geology. Secondly, by the genesis there is determined the composition of soils, their nature of bedding, composition, initial moisture content, i.e., those factors that are some of the decisive ones in the formation of the cryogenic structure and ice content of deposits.

The genetic types of soils are reflected on the maps by the main graphic means, i.e., by a colored background (blue-green spectrum of shades) accompanied by genetic indexes.

2. The composition of soils within the limits of the genetic types is reported by the conventional lithologic notations (primed and dotted marks) of black color inscribed in the form of short cross-hatching on a checkered grid. According to composition, on the map we show peat, clays, supesses, silty soils, sands, and coarse detrital rocks or their combinations. In the latter case, the symbols are grouped and also posted on the chart according to a grid. The representation of the uncemented and slightly cemented continental Mesozoic deposits is conducted by the same signs for the composition, but their color corresponds to that of the terrestrial formation of bedrocks. In this case the cross-hatching is indicated as a continuous covering over the area.

The differentiated lithologic variants of the loose deposits are distinguished by the nature of influence on the heat-exchange processes. This is connected with the fact that on the granulometry of soil particles and the nature of the bond between them, there depend their permeability qualities and the features of ice segregation on freezing of the soils.

3. The stratigraphic classification of soils is indicated on the map with the aid of their age index.

4. The type of freezing of the deposits is represented as follows: the epigenetically frozen strata are additionally cross-hatched; at occurrence of syngenetic strata from the surface, this cross-hatching is omitted.

5. The representation of cryogenous features of soils consists of the following factors: (a) the indication in the map legend of the ranges of predominant values of ice content in the upper horizons of soils and cryogenous textures depending on origin, composition and also age of soils, type of freezing, and generalized description of the thermodynamic conditons involved in the formation of frozen strata (their northern and southern types). (b) representation directly on the map of recurrent wedges, ice determining the macro-ice content of soils, and other cryogenous formations portrayed by large ice accumulations.

The technique of representing the cryotextures and ice content of soils as a function of the factors determining them is the only method permitting us to show them with adequate completeness.

The principles and techniques of cartographic representation of the composition and cryogenous structure of the rocky and semirocky materials are:

1. The representation of formations is the basis for mapping the rocky and semirocky materials. This is associated with the fact that, first of all, the diversity in the composition of rocks within their limits is limited and strictly determined; secondly, within the frameworks of the formations, there are differentiated the strata of rocks typified by fairly specific thermophysical properties and the state of fissuring, i.e., by factors causing the cryogenous structure of the rocks and the nature of heat transfer in their layer. In addition, to each formation there correspond definite tectonic (including neotectonic) geomorphological, hydro-

geological, and abyssal geothermal conditions. In connection with this, the formations constitute an expression of the regional conditions of heat exchange on the surface and in the layers of rocks.

2. To the extent that the composition, thermophysical properties, and fissuring of a rock are formed during the course of the entire geologic history of their existence, the basic stages of its history should be taken into account in differentiating the formations reflected on the permafrost map. Therefore, the formations segregated, i.e., intrusive, effusive, metamorphic, effusive-terrestrial, terrestrial, terrestrial-carbonate, halogenic, and molassic*[5] depending on tectonic conditions of the formation of the rock are subdivided into geosynclinal and platform-type formations and are considered within the limits of the basic contemporary tectonic structures.

3. The main graphical means for showing the composition of rocks and semirocks, and also the tectonic conditions, is the fine-colored cross-hatching (in the publication--lithographic grid) filling the role of a colored background. The yellow-brown-red shades of hatching correspond to the formation. The intensity of its color, comprising a means for showing the basic contemporary geostructural areas, indicates the features of composition (neotectonic and exogenic fissuring), connected indirectly or directly with the effect of neotectonics. The direction of hatching testifies to the classification of the rocks among the geosynclinal or platform-type formations and by the same token indicates the most important pre-Cenozoic structures and the basic structural elements of the platforms.

The graphical representation of the basic (from a permafrost standpoint) differences in the rocks and semirocks from the soils is achieved by the application, first of all, of various means of representation (cross-hatching, background) and secondly by the use of abruptly differing color ranges.

4. The technique adopted for representing the formations permits us to indicate the rocks and semirocks as a medium determining the nature of heat exchange in their depth. This is associated with the fact that the lithologic-petrographic composition of rocks, the tectonic conditions of formation, and their lithification and neotectonic conditions essentially constitute factors determining the heat-exchange processes, since they control the thermophysical characteristics and amount and varieties of rock voids. Such a genetic approach to the representation of the hollowness permits us (a) to relfect the types of cryogenous textures of the rocks and semirocks; (b) to judge the thickness of the zone of exogenic fissuring and to develop a concept concerning the possible depth of ice-saturated cracks; and (c) to predict the presence and nature of the thermal interaction of underground water and frozen strata and to reflect the relationships in the formation of the temperature regime and the thickness of permafrost and talik zones.

The cryogenous textures corresponding to the rocks of specific formations are listed in the legend, and therefore the conventional symbols for the formation constitute indications of the cryotextures entering the composition of the rocks.

5. The stratigraphic classification of the rocks within the formations' limits is shown by an index.

The method of showing the structure of permafrost can be described as follows:

1. By the color background and color cross-hatching, we indicate the rocks occurring at the surface or beneath a thin (mainly less than 5 m) layer of surface beds.

2. For the indication (by area) of the structure of the permafrost, we utilize an age index of the rocks, while for the soils we utilize genetic indexing. The listing of these indexes permits us to show on the map a series of beddings of various age, origin, and composition forming the permafrost.

The structure of double-tiered permafrost, i.e., formed in the upper part of soils and in the lower part by rocks and semirocks is shown on the map in the form of a fraction, the numerator of which is the indexes of unconsolidated deposits, while the denominator is the indexes of the rocks and semirocks. The structure of the single-tiered frozen strata is indicated by indexes. The last of them indicates the occurrence in these rocks of the lower boundary of permafrost.

3. With the indication of the thicknesses of rocks of varying composition, origin, and age, the section of frozen strata (within the limits of the various tectonic structures) is characterized in the columns of the reference boreholes shown directly on the map.

The conventional notations of the basic parameters of permafrost are combined in one composite symbol, i.e., shading made in bright wide strips of light blue. The nature of interruption in the shading indicates the distribution, direction indicates occurrence depth, and frequency shows the thickness of frozen strata, while the nature of filling between the shading lines gives their temperature regime. The cryogenous age of the rocks is shown by the age indexes of a violet color.

The distribution, occurrence depth, and thickness of the rocks cooled below 0° containing cryopegs* are reflected by a shading of green. The symbols outside the scale indicate the genetic subdivisions of taliks and cryogenous formations associated with the layer of permafrost. In the columns for the reference boreholes, in addition to the composition and structure of fro-

*Molassic--resembling a millstone, coarse soft sandstone, rich in feldspar and mica, of Tertiary age.

*Bodies of saline water at a negative temperature.

zen layers, there are indicated the nature of interaction of underground water and permafrost and the dynamics involved in the development of the latter (by listing temperature curves typical for the region).

On the other permafrost map, i.e., that of seasonally frozen and seasonally thawed rocks, there are indicated the types of seasonal thawing and freezing of rocks according to V. A. Kudryavtsev[2] and their basic characteristics: composition of deposits within the confines of genetic types, water content of rocks, average annual temperature and range of temperatures at the ground surface and also at depth of seasonal thawing, freezing of rocks, and the cryogenous formations associated with this layer. The indication of the characteristics for the seasonally frozen rocks is accomplished by methods assuring the comparability and mutual supplementing of both permafrost maps. This permits us to consider them as a unified general permafrost (geocryological) map.

In this way, the representation of the relationships in the formation and development of the seasonally and perennially frozen strata on the general permafrost map of the USSR on a scale of 1:2,500,000 dictates the necessity for indicating the most extensive information. The capacity of the methods listed for its reporting permits us to present the content of the map in the indicated volume, thereby having retained its excellent legibility.

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PERMAFROST PREDICTION AND METHODS OF ITS FORMULATION

V. A. KUDRYAVTSEV, L. S. GARAGULYA, S. A. ZAMOLOTCHIKOVA, V. G. MELAMED, AND N. F. POLTEV   Moscow State University

The problems associated with the prediction of permafrost conditions varying during the development of an area are currently solved by numerous organizations. However, until now there has been no standardized approach to a solution of the complex forecasting problem. Some researchers place primary emphasis on the heat-engineering calculations, devoting inadequate attention to and sometimes completely ignoring the variations in the geologic-geographic factors of the natural environment associated either with their natural dynamics or with construction measures. As a result of the complexity of the heat-engineering calculations, others simplify the problems to qualitative evaluations. The effect of the overlapping of the individual engineering influences and the formation of the average integral variations within the limits of the structural complex are quite rarely taken into account. It seems possible to us to combine the efforts of all organizations in solving the problems of predicting the changes in permafrost conditions during the construction and operation of buildings, on the basis of the following fundamental concepts.

The permafrost conditions in an area are determined by a combination of the natural environment. In a definite manner, each of the geologic-geographic factors influences the formation of the permafrost condition. Therefore, in the process of a permafrost survey, we conduct a study of the relationships in the effects of geologic-geographic factors on the formation and dynamics of the seasonally and perennially frozen rocks. In this context, we establish the general and specific regularities involved in the formation of composition, structure, thermal-moisture regime, conditions of bedding, distribution thickness of seasonally and perennially frozen rocks, their mechanical and thermophysical properties, and also the tendencies involved in the formation and development of permafrost processes and phenomena.

The indicated tendencies are determined on the one hand by the thermodynamic and thermophysical laws and on the other hand by the geologic features existing in the region under study. Therefore, in establishing the quantitative bilateral relationship between individual elements in the natural environment and the characteristics of the seasonally and perennially frozen rock, the input parameters should reflect in the design systems, the geologic-geographic conditions in such a manner that they could be measured directly in the field and tied in with the features in the natural conditions in the process of a permafrost survey. Only on the basis of such a procedure can we obtain the necessary original data for predicting the changes in the permafrost conditions in conjunction with the economic development of a territory.

In formulating a prediction of the variation in permafrost conditions on the basis of the data from a permafrost survey, use can be made of various calculation methods. The changes in the geologic-geographic situation as a result of construction are reflected primarily on the radiation-heat balance at the surface and on its structure. The latter determines the heat condition of rocks, their temperature regime, and the dynamics involved in freezing and thawing. Therefore, at the basis of the compilation of a prediction of the variation in permafrost conditions, there should be placed an analysis of the variation in radiation-heat balance of the surface in connection with the construction. Specifically, it is necessary to analyze how there is reflected on each of the components the removal of the snow or vegetative cover, the variation in exposure of terrain and steepness of slopes, draining of soils, percolation of warm summer precipitation, and so forth.

The most significant variation in the radiation-heat balance occurs owing to the variation in water content of soils; moreover, this is reflected not only on evaporation from the surface, but also on the values of albedo, convective heat exchange, and heat conduction in the soils as well as on the effective radiation values.

In this manner an analysis of the variation in each of the components of the radiation-heat balance should be reinforced by a calculation of the radiation-heat balance under the altered conditions as a whole. Obviously, it is impossible to make such a calculation by a laboratory approach. For this purpose it is necessary to discover, under field conditions in the process of a permafrost survey, the role and significance of each natural factor in the formation of the radiation-heat balance at the surface in the investigation region for each type of landscape.

The study should be conducted in such a manner as to identify the dependence of the variation in the components of the radiation-heat balance in connection with the modification in the temperature regime of the ground. In this context, the indicated components are represented as functions of the temperature conditions of the rocks at the Earth's surface. On the basis of the principles disclosed, the formulation of the prediction is conducted in conformity with the systems developed according to the generally known equations for determining the temperature field and depth of seasonal freezing and thawing. In those cases when the natural moisture content of soils is relatively low (does not exceed the value of maximal molecular water capacity), we utilize the systems in the calculation of conductive heat transfer. These calculations are the simplest and the best developed. At the present time, a series of approximate formulas has been suggested for estimating the depths of seasonal and perennial freezing and thawing under conductive conditions, yielding results that are fairly good for forecasting purposes. Among them we should point out the formula by V. A. Kudryavtsev, permitting us to compute the seasonal and perennial freezing in the periodically established regime at the closest correlation with the actual geologic-geographic situation.

The estimations of the seasonal and perennial freezing and thawing of saturated rocks are much more complex. This is primarily associated with the fact that in the given case, along with the conductive heat exchange, it is also necessary to consider mass transport, determined basically by infiltration of water into coarse soils and by the migration of moisture toward the freezing front in the fine soils. In case of appreciable moisture content, even slight variations in it are reflected appreciably on evaporation, heat exchanges, and temperatures of soils. As a result, at this time there is an abrupt change in the value of effective radiation.

In this manner the consideration of moisture transport in soils greatly complicates the solution of the questions in permafrost forecasting, since small errors in calculating the individual components of radiation-heat balance could lead to an unbalance. At the same time, the interrelationship of the processes of heat and moisture exchange during freezing and thawing in the saturated materials abruptly complicates the mathematical investigation of the questions in permafrost conditions. The calculation techniques being utilized for these purposes are quite approximate and are solved chiefly by a graphical technique. However, the methods of numerical integration with the aid of a computer are quite awkward. Therefore, as initial conditions, it is necessary to take the natural data that are established during the permafrost surveying process.

The variation in the components of radiation-heat balance leads to a corresponding change in the depths of seasonal and perennial freezing and thawing, temperatures of soils, and consequently to a modification in the bedding conditions of frozen strata, their composition, and cryogenous structure. Moreover, in addition to the questions indicated above, there also develops the problem involved in the dynamics of the upper boundary of perennially frozen rocks, the formation of taliks, the extent of frost-heaving or thaw-subsidence, with allowance for the nature of cryogenous textures and their variation in connection with freezing.

A natural result of the dynamics in the permafrost processes through time is the change in the properties of frozen and thawing soils. The qualities of frozen soils are determined substantially by their composition, cryogenous structure, and temperature regime. In connection with this, the problem develops of forecasting the variation in these basic parameters and the properties of frozen soil that are being determined by them. Since the temperature regime and cryogenous textures depend considerably on and are determined by the nature of radiation-heat balance at the surface, the latter obviously also determines the change in the qualities of the frozen and thawing soils.

At the same time the composition, cryogenous structure, and temperature regime of rocks also determine the nature of cryogenous processes and occurrences. The prediction of the development of such processes as heaving of soils, settling on thawing, heat erosion, solifluction, etc., constitutes one of the basic and complex problems in engineering-geologic investigations. Its solution also requires the application of various calculation methods in conformity with the formulation of the problems.

Prediction of frost-heave can be divided into two basic problems. The first combines the calculations of the variation in temperature fields and depths of the seasonal (perennially) freezing. The second is solved with the aid of predicting the cryogenous structure of the freezing soils and the magnitude of heaving under actual permafrost-geologic conditions. The technique of solving these problems has been developed in the department of permafrostology and can be widely used in the practice of engineering-geologic investigations. The data obtained during the solution should be plotted on a special map of the extent of soil-heave that is compiled during the permafrost-surveying process.

The prediction of the processes of thaw-settlement of soils of thermokarst and thermoerosion is associated with the solution of problems on forecasting the dynamics of temperature fields and the depths of seasonal (perennial) thawing of soils under actual permafrost-geologic conditions. The quantitative evaluation of the effects of these processes, given the known dynamics of temperature regime, can be conducted with the aid of approximate formulas developed by a number of authors. In predicting the indicated processes, it is extremely important to take into account the nature of the proposed earth operations. The basis for the compilation of a forecasting chart of the development of the processes of thermokarst, settlements during thawing, and thermoerosion should be a special chart reflecting the composition of ice content and cryogenous structure of the various geologic-genetic types of soils found in the territory under study.

In the evaluation of the development of solifluction processes, we predict the dynamics for the depths of seasonal thawing, the formation of ice content in the layer of seasonal thawing depending on the nature of freezing and temperature regime of soils; we determine the summer temperature regime in the layer of seasonal thawing for appraising the possibility of restoring the cohesive bonds in the thawing wet soils.

All the indicated problems that are formulated for obtaining an actual permafrost prediction are currently being solved with the aid of analytical calculation techniques, the methods of simulating the permafrost processes and the procedures of calculating on a computer. On the basis of precise solutions, we have developed rapid field methods for computing the permafrost parameters with the aid of approximate formulas and nomograms.

Let us discuss the features in the permafrost prediction at various stages of investigation. In small-scale surveying (1:1,000,000-1:200,000 stage of planning and technical-economic justification), a forecast is made of the variation in permafrost conditions in a very general way. Above all we should have in mind here the variation in permafrost conditions in connection with the influence exerted by the various factors in the geologic-geographic environment. We include here the effect of snow cover, vegetative and water covers, composition and moistness of soils, steepness and exposure of slopes, infiltration of summer atmospheric precipitation, surface and subsurface water, swampiness, etc. The effect of these factors should be considered from the standpoints of the formation of the radiation-heat balance of the surface and with consideration of the natural pattern of climatic variations. In this manner, in small-scale surveying, we formulate first of all a prediction of the variation in permafrost conditions in connection with the natural pattern in variation of the geologic-geographic environment. Simultaneously with this, the indicated permafrost prediction during the small-scale surveying serves as a basis for developing a general forecast, both in connection with the various types of construction and man's industrial activity. Moreover, the general forecast comprises a basis for designing and planning research activities for the ensuing stages in planning and design.

During the medium- and large-scale surveys (1:100,000 and larger, the stage of technical designing and working drawings), we develop a prediction of the variation in permafrost conditions for individual sectors and construction sites, depending on the actual features of the structures and industrial development of a territory. The basis for this prognosis is the general forecast examined above, the data in which are refined for actual conditions. Moreover, we take into account the man-made transformations in natural conditions, such as the landscape planning, the cutting and filling of ground, the erection of earthworks, draining and irrigation of soil, forest planting and removal of trees, planting of lawns, erection of artificial surfaces, and regulation of the surface and subsurface runoff.

Obviously the forecasting during the medium- and large-scale investigations should be accomplished by stages. The first stage of forecasting can be called the approximate forecast. Such a prediction is conducted simultaneously with the development of a plan and program for the research operations and sets itself the task of imparting a definite trend to the subsequent permafrost-engineering-geologic investigations. The series of rough forecasting calculations conducted at the preliminary stage of surveying simultaneously with an exhaustive study of the materials from projects having been conducted previously, i.e., general geologic, engineering-geologic, hydrogeological, and permafrost, is accomplished based on the existing approximate formulas, while for the more important buildings calculations are also made on simulated models. In the calculations we take into account the general relationships in the existence and development of frozen strata established during the small-scale studies of a region, we utilize the data excerpted from reference literature; and we make a thorough use of construction experience available in the region. In this manner, we can identify the thaw bowls beneath buildings and reservoirs, zones of active soil consolidation beneath heavy structures, thaw-aurioles under pipelines, depth of thawing in dredging areas, etc.

The basic purpose of the rough prediction is that, in a first approximation, we can reveal the investigation area and determine qualitatively the most likely developmental trends in the engineering-geologic processes and hence assign objectively a specific group of subsequent studies, determine their proportions in the combination of methods, sequence of performance, and scopes.

The subsequent stage in the formulation of the prediction of the variation in permafrost conditions begins during the field investigatory process. At this stage the basic content of the prediction is the quantitative determination of changes in the essential parameters of the perennially and seasonally frozen soils in each sector in connection with the natural dynamics of various natural factors and their variation during construction. We also establish the nature of the structure's interaction with the frozen strata under actual permafrost-engineering-geologic conditions. The data collected, reflected in an engineering-geologic chart and sections with it, should serve as a justification for the planning.

The forecast conducted on the basis of information from the permafrost surveying has the following unique feature. Since in it the basic attention is relegated to the variations in the cryogenous structure, thickness, and thermo-moisture regime of the frozen and thawing soils, the forecasting system can be considered common for any type of construction. The specifics of construction and the structure operational regime determine chiefly the range of problems that are to be solved during the ensuing stage of prediction.

In the next stage of developing the prediction, in the calculations we take into account the parameters of a structure and the nature of its interaction (thermal and mechanical) with the frozen strata in an actual sector. The data obtained comprise a basis for compiling working sketches and also programs of the measures for controlling the freezing-thawing processes during operation. The forecasting calculations quite obviously should be conducted with allowance for the data obtained during the second forecasting stage. Moreover, for the complexes of structures, we also establish the overall variations in the parameters of the frozen layers.

The results obtained are analyzed, and we establish the limits in the variation of the basic indexes typifying the permafrost-engineering-geologic conditions. In this context, we segregate those cases when there occurs the overlapping of variations in natural environment from the natural dynamics of geologic-geographic factors and from the influence of the structure.

A typical feature of the technique under discussion is the fact that the general principles, clarified during the permafrost survey, of the formation of seasonally and perennially frozen soils permit us not only to forecast the variation in permafrost conditions but also to establish the principles and methods of controlling the permafrost process for construction purposes in order to attain the optimal conditions of the functioning of the structure.

SCIENTIFIC-PROCEDURAL BASES FOR PREDICTING THE ENGINEERING-GEOCRYOLOGICAL OCCURRENCES IN SECTORS UNDER DEVELOPMENT

P. F. SHVETSOV AND N. G. BOBOV All-Union Scientific-Research Institute of Hydrogeology and Engineering Geology

The industrial development of a permafrost area is conducted at separate points. The complete development of all this region is not intended even over an extended period. In the development centers, there occurs a variation in the geocryologic situation (occurrence depths, thickness, extent of discontinuity, temperature, ice content of the permafrost and of seasonal thawing layer) as a result of the complex effects of human activities. These changes in turn exert an appreciable effect on the engineering facilities having caused them. The entire history of developing a permafrost area testifies to the great practical importance of such modifications.[6,8] The effect of these changes increases through time. The requirement for an engineering-geocryological prediction is obvious.[9]

An engineering-geocryological prediction signifies the forecast of the trend, intensity and results of changes in geocryologic conditions under the effect of future industrial and municipal-economic activity of man in all, or even in separate large parts, of an area subject to development. The engineering-geocryological variations are determined, on the one hand, by the natural geocryologic situation, so to speak, as a natural background and by the extent of disruption to the conditions of heat and water exchange, and, on the other hand, by the form and measure of influence of the industrial activity's results on this situation.

The original engineering-geocryological conditions in a sector earmarked for development are clarified as a result of a special survey that is completed by the compilation of a geocryologic map.[12] The accuracy of the latter is determined by the requirements of the forthcoming type developing a locality.

The results of the industrial activity are characterized by a definite form of influence on the geocryological conditions (see Table 1). The point, linear, and local two-dimensional form of the future effect are subjected to analytical calculations with an accuracy that is adequate

TABLE 1 Compilation of Types of Influence of Industrial Activity on Natural Geocryological Conditions, Established During Surveys (According to P. F. Shvetsov)

Geometric Form of Influence of Results from Industrial Activity on Geocryological Conditions	Type	Position in Vertical Section of Lithosphere-Soil-Atmosphere or Relative to the Earth's Surface
Point	Point (foundations of buildings, bridge supports, trestles, boreholes)	Surface Near-surface Underground
Linear	1. Vertical (drilled borehole, mine shaft)	Surface Near-surface
	2. Horizontal (pipeline, tunnel)	Underground
Two-dimensional (areal)	1. Local (building, waste pile, road, airfield runway, excavation)	Surface Near-surface Underground
	2. Local (village, above-shaft facility, city, open-cut mine, worked-out mine field)	Surface Near-surface Underground-surface
	3. Regional (Vorkuta region in Bol'shezemel'skaya tundra, Noril'sk region on central Siberian upland)	Surface

for the purpose. For this purpose, many equations were long ago developed and validated. The remaining versions of a two-dimensional form of an unforeseen industrial effect on the geocryological conditions have been studied much less thoroughly and until now have practically been overlooked.

In connection with the focal nature of the development of the territory under discussion, of special scientific and technical interest is the local two-dimensional form of the influence of industrial activity on the natural geocryologic conditions. This form of influence is composed of many diversified elements. Therefore, it proves extremely complicated to determine the extent of its effect. Our paper is devoted to the scientific-procedural bases of specifically predicting this form.

The disturbance to the combination of processes and conditions in heat and water exchange is subdivided into direct and indirect.[11]

The *direct* influence of engineering projects includes their direct heat release into the ground or heat absorption from the soil. This form of influence is manifested chiefly under each structure and fairly intensively, usually during the first 3 yr. It consists of local and smaller forms of influence.

The overall direct influence of many objects on a sector that is being developed is estimated from the density of development (intensity factor)

$$\sum_{i=1}^{n} = \frac{e_i a_i \tau_{av}}{A}$$

where e = temperature of heat contact (intensity factor); a = area of heat contact (extensity factor); τ = relaxation time; and A = area of entire sector under development.

The *indirect* influence includes the disruption of natural factors of heat exchange in the lithosphere-soil-atmosphere system: The construction and use of buildings are accompanied by a change in the illumination of the surface by the sun, moisture content, and density of soils; by the accumulation of a "man-made" layer; by a disruption in the wind regime, soil-vegetative cover, height, and compactness of snow deposits; and by releasing smoke into the atmosphere, polluting the surface, etc. This effect spreads to the geocryologic conditions of the entire area under development and is manifested more gradually over a period of 10-15 yr. It has been studied much less thoroughly and until now has not been considered during the development of a locality. The study of such a form of influence constituted the basic content of the investigations conducted.

The engineering-geocryologic prediction permits us to take into account the indirect disruption of natural heat-exchange factors. It is important to stress that the trend in the influence exerted by direct and indirect factors on the geocryologic conditions is by no means the same everywhere.[7] The complexity of this prediction determines its probabilistic nature.

Man exerts the maximum influence on individual elements in the landscape. Among them, A. I. Voeykov[3] included the following: vegetative cover; free-flowing loose soils--sandy and silty; internal surface basins and waterways, including the permafrost area and also the seasonally thawing layer of silty supersaturated ground; snow cover; layer of rocks with annual heat exchanges; and ice content exceeding the total water capacity.

The most unbalanced geocryological formation includes the subarctic and northern taiga alluvial lowlands (Yana-Indigirskaya, Kolyma-Alazeyskaya, and others) and deltas of rivers in which, along with the thin interbeddings, there also are large ice deposits. The thin layer of seasonal thawing and low temperature in the layer with annual heat exchanges cause large geotemperature gradients. The results from the thawing of large underground ice masses in Chokurdakh settlement and, in the broad valleys of the Verkhoyansk area, Khandyga settlement, can serve as examples of significant lack of equilibrium.

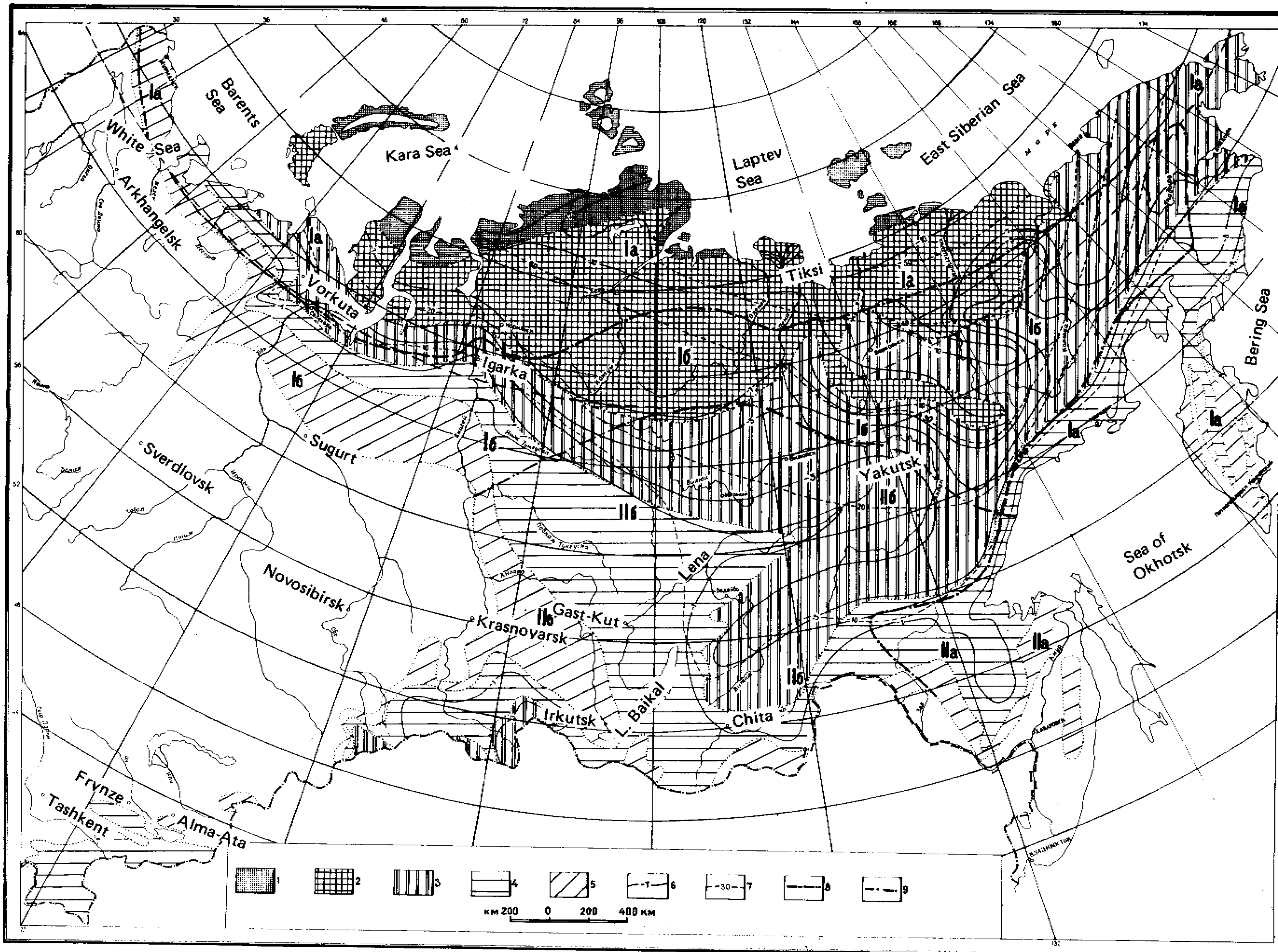
The actual diversity of microlandscapes and of their corresponding geocryologic facies determining the mobility of the soil-ground complex on thawing is extremely great. These factors have not yet been systematized. The natural factors causing the geocryologic conditions can be found proceeding from the physicogeographic axioms offered by Dokuchayev-Yachevskiy;[10] according to this, the condition of the rock stratum and the manifestation of engineering-geocryologic processes are determined by the following three groups of factors for heat and water exchange:

1. Petrographic-lithological--i.e., by the composition, structure and properties of rocks and soils; permeability is particularly important.
2. Morphological, i.e., by the position of the sector, under study, in the terrain and on a continent, i.e., by the height (absolute and relative), by angle and orientation of surface, by connection with the basin.
3. The zonal-belt energy balance of the active layer and the climate.

The remaining natural conditions existing in heat and water exchange vary according to these most important groups.

The engineering-geocryologic prediction was applied for the first time during the prospecting of deposits in the Bureinskiy coal basin. As a result of the studies performed, a conclusion was drawn that the draining of swampy sectors and the breakup of the peat-moss cover during the development of the locality, accompanied by a reduction in evaporation of moisture, would cause a rise in temperature and a thawing of frozen ground.[1] This was confirmed subsequently.[5]

A generalization of the empirical data concerning the changes in geocryologic conditions in sectors already developed has permitted us to clarify the basic trends of its development and, on this retrospective basis, for forecasting purposes to subdivide the permafrost area[2] into large territorial units (see Figure 1). The



Barents Sea
White Sea
Arkhangelsk
Kara Sea
Laptev Sea
East Siberian Sea
Bering Sea
Sea of Okhotsk
Vorkuta
Igarka
Tiksi
Yakutsk
Sugurt
Sverdlovsk
Novosibirsk
Krasnovarsk
Gast-Kut
Lena
Irkutsk
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FIGURE 1 Engineering-geocryologic division of developmental area of permafrost within the boundaries of the USSR (after N. G. Bobov). Natural geocryological conditions: 1-5--zonal thickness (m) and discontinuity: 1--500-600 m, rare taliks; more frequent; 3--100-300 m, permafrost often in only half of the sectors' area; 4--25-100 m, the permafrost occurs more often "in islands" in the smaller areas; 5--25 m, rare islands of permafrost; 6--temperature of ground (°C); 7--boundary of zones of maximum thickness (m) of beds of underground ice. Engineering-geocryological zones: 8-9--boundaries of zones (Roman numbers) and subzones (letter indexes). I. Northern zone: maximum thicknesses and minimum temperatures of layer of permafrost are inherent ot the watersheds; the taliks are coordinated to the relief depressions; the variations of development in a locality can reach maximal values in accordance with the original parameters. Ia--maritime subzone; development leads more often to the reduction of the extent of the frozen layer, to a rise in its temperature and to an increase in the depth of seasonal thawing. Ib--continental subzone: development involves an increase in the distribution, a decrease in temperature of the permafrost and a reduction in the seasonal thawing layer. II. Southern subzone: relatively maximum thickensses and minimal temperatures gravitate toward the depressions in relief; taliks predominate in the watersheds with absolute elevations below 1,200 m; the engineering-geocryologic variations are less than in the northern zones. IIa--maritime subzone: the development of a locality essentially causes the thawing of the permafrost, an increase in temperature and in the extent of seasonal thawing; IIb--continental subzone; development leads more often to the new formation of permafrost, a rise in temperature, and a decrease in the seasonal thawing. Engineering-Geocryologic Division of Developmental Area of Perennially-Frozen Ground Within USSR Boundaries (after N. G. Bobov)

main types in the developing of a locality were differentiated on the basis of mining operations and aboveground construction. In this context against the background of undisturbed geocryologic conditions, there are indicated the trends in its variation evoked by the indirect influence of the existing development of a locality.

The permafrost area is divided into sublatitudinal engineering-geocryologic zones, i.e., northern and southern. In the northern zone, the relatively most continuous and low-temperature permafrost is inherent to the elevated relief. Among the combination of the external factors of heat exchange determining this, it is necessary to indicate the low values of radiation balance (20-25 kcal/cm^2), more than half of which is expended on evaporation of moisture and one-fourth is expended on turbulence in the air masses. The climatic conditions are characterized by low annual air temperature and slight evaporability of moisture. The latter constitutes less than half of the falling precipitation; this leads to an excessive wetting of the predominant area of the surface, swampiness of all the terrain elements, and also to a constant and intensive water runoff. According to the dip of slopes on elevations, infiltration of surface waters increases. An especially large volume of water percolates to the valley floors; this is favored by the high permeability of their deposits, i.e., sands, gravels, and pebbles. Strong winds are typical throughout the entire zone. The disruption of the conditions of heat and water exchange during development can cause the maximum engineering-geocryologic changes there in conformity with the original properties of these conditions.

In the southern engineering-geocryologic zone, the permafrost is more discontinuous and has a higher temperature. In contrast to the northern zone, it is located chiefly in the relatively depressed relief elements, to the absolute elevation of 700-1,200 m and on the mountain peaks exceeding these elevation marks. The taliks are confined more often to the local elevation, with levels not greater than those indicated. Among the features of the external heat-exchange factors in the southern zone, it is necessary to point out the increased radiation balance (30-40 kcal/cm^2), about half of which is expended on evaporation of moisture and a third on turbulent, convective heat exchange. The zone's climate is characterized by higher air temperatures and an appreciable evaporability of moisture (in the large enclosed basins more than the annual amount of precipitation) which provides the basis for swampiness only in the relief depressions that, often are formed at the top by suglinoks. Therefore, in summer evaporation occurs chiefly from such sectors. As a whole, we record a low water runoff and weaker winds. Temperature inversions of air masses in the basins in winter cause the low mobility of air downward and an increase in wind force with height. Therefore, the elevated parts of slopes are exposed to air masses of ever-increasing temperature. The disruption of geocryologic conditions as a result of development, at relatively slight thermal influence, can prove to be greater here than in the northern zone. In the case of intensive influence, the engineering-geocryologic variations are limited by the narrower range of natural geocryologic indicators.

The trend in the predominant development of engineering-geocryological phenomena is also strongly dependent on location of territory on the continent; on the one hand, in the maritime regions, and, on the other hand, in the continental regions. Therefore each zone is subdivided into two subzones. In the northern zone their boundaries are formed by the limit of forest-tundra and taiga and in the southern zone by the limit of the maritime dense taiga and continental suppressed taiga, alternating with steppe sectors.

In the sectors of maritime subzones under development, we note the preferential rise in temperature of ground as compared with the natural state, and also its thawing. Owing to the predominance of a cyclonic weather regime, strong winds prevail there. Their average annual speed reaches 5-7 m/s.* Maximum wind speeds exceed 20-40 m/s. Such a wind regime predetermines the intensive transport and (what is most important) accumulation and pollution of the snow. In summer this involves an increased water runoff, specifically in the cities, villages, and open-cut mines. In winter the large snow masses reduce the heat losses from the ground. In spring the contaminated snow thaws more quickly, freeing the surface for solar heating. The same condition favors the organization of the melt-water runoff, disruption in the peat-moss cover, and in particular the artificial reduction or even compaction of the seasonally thawed layer.

In the continental subzones, in the villages and cities, there occurs more often a reduction in ground temperature and an additional freezing of the thawed sectors. In the open-cut mines, there occurs a gradual rise in temperature and thawing of the permafrost. Such disturbances are caused by the prevalence of the anticyclonic weather regime where the weaker winds prove typical (3-5 m/s for the year), less blowing, and pileup of snow. Contamination of the snow favors its intensive encrustation at negative temperature. The compaction and intensified crusting of the snow cover determine its decreased insulating capabilities in the populated point, as compared with the areas away from these points. Owing to this the reduced runoff of melt water leads to a reduction in their infiltration and by the same token reduces the heat transport into the ground. Moreover, the reduced runoff in the villages and cities promotes the sanilization of groundwater. Owing to the prolonged sunshine, the shading of surface acquires great importance. All this leads to a cooling of the ground.

*The first digit in all additional double indexes typifies the conditions in the southern subzones, while the second digit refers to the same, for the northern subzones.

A lowering in the level of subsurface water in the minefields promotes a more intensive infiltration of surface water, which leads to a growth of the taliks. In the northern zone, the taliks expand from the bottom of river valleys and lake basins, while in the southern zone they extend chiefly from the watershed elevations.

The waste heaps, dumps, and road embankments, in conformity with the natural conditions, freeze more often in the northern zone, while in the southern zone they experience a rise in temperature and thawing of underlying ground.

In conclusion, it should be stressed that a consideration of the essential zonal and regional differences existing in the variation of geocryologic conditions, in combination with the disruptions in natural conditions of heat exchange generating them, will permit us to predict more reliably the effects of engineering-geocryologic processes in the sectors subject to mining-industrial and public development of a locality within the boundaries of the territorial units that have been differentiated.

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PREDICTING THE VARIATION IN SEISMIC PROPERTIES OF PERMAFROST

N. YE. ZARUBIN AND O. V. PAVLOV *Institute of Earth's Crust at the Siberian Branch of the USSR Academy of Sciences*

More than half the area of all seismically active territories in the USSR falls within the permafrost region. In the Baikal rift zone, famous for its high seismic activity, we differentiate regions with varying permafrost conditions (Figure 1).

The most complex permafrost-soil conditions are typical of the northeastern and northern regions of Trans-Baikal. In the Stanovoye upland, where earthquakes up to Intensity X have occurred and are possible, the permafrost is continuous. The thickness of the frozen layer in the intermontane areas amounts to 10-100 m; in the depressions of the Baikal type, from 200 m to 600 m, and in the mountain structures from 1,100 m to 1,300 m. The temperature of frozen soils varies from -5° in the unconsolidated deposits of depressions to -8° to -11° in the rocks on mountain ranges.[6] The frozen soils of various types from a cryotextural standpoint often contain large volumes of underground ice of varying origin, with a thickness to 10-15 m and more. Many permafrost processes and phenomena typical of these regions complicate the natural grouping of the territory significantly. Of special importance for estimating the seismic danger in actual sectors are the slope processes (rock trains, talus, and solifluction).

In the southern and western regions around Lake Baikal, where earthquakes with Intensity VI to IX[9] are possible, the permafrost occurs chiefly in the form of large masses, islands, and pockets. In a major part of the region, the thickness of the frozen layer equals 15-80 m; in places it reaches 100-300 m.[1-4] The temperature of the frozen soils is -0°, 1°, and 2°. The extent of ice saturation will fluctuate within broad limits and can reach 100 percent and more of the weight of dry soil.

The cryogenous and postcryogenous physicogeologic processes and phenomena introduce significant complications into the permafrost-geologic conditions in these regions. Differentiated by particularly complex permafrost-geologic structure are the depressions of the Baikal type: Verkhne-Angarskaya, Barguzinskaya, and Tunkinskaya, situated in the IX to X Intensity seismic regions.

In the bottoms of depressions, the upper limit of permafrost in the swampy sectors lies at depths of 0.5-1.0 m. In the dry sandy massifs, it descends for a depth of 100 m and more. Thickness of the frozen layer increases from the marginal (several meters) to the central--axial--parts of the depressions (up to 200-300 m and more).[4,8] The islands and pockets of permafrost in the depressions often come in contact with the subsurface water, causing a high water-saturation of the thawed loose deposits; this aggravates the seismic conditions in the territory. As a rule permafrost is absent from the zones of seismically active faults.

Just in recent years in the regions where permafrost occurs, several severe earthquakes have taken place: Muyskoye (1957)--X, Olekminskoye and Nyukzhinskoye (1958)--IX, Tas-Yuryakhskoye (1967)--X, and Oymyakonskoye (1971)--X on the Intensity scale.

Every year in the Baikal seismic station, we record large numbers of weak earthquakes [I-VII-(11)], testifying to the high seismic activity in these regions. According to the data from seismologic studies, in the Baikal rift zone during 1,000 yr, on an average we can expect: 126 earthquakes with an Intensity of VIII, 38 up to IX, 17 up to X, and 6 with Intensity of about XI.[5]

All these facts indicate that during construction in seismically active regions under permafrost conditions, among the complications caused by the features in the permafrost-soil conditions, such an unfavorable factor as seismicity is also added.

The studies conducted by the Institute of Earth's Crust in the Siberian Branch of the USSR Academy of Sciences in recent years in the territory of the Baikal rift zone (Trans-Baikal North, Barguzinskaya, and Yeravninskaya depressions) have shown that under natural conditions, depending on seasons, along with variation in the temperature and moisture regime of soils, their seismic properties also change. Such an objective index of the seismic stability of soils as their acoustic velocity can change by several times in the layer of annual temperature fluctuations. Accordingly the difference in the increments in intensity for the granular frozen soils can comprise 1 point and more in the course of a year.

The special field investigations of seismic danger in the soils of taliks of limited dimensions have shown that the increments in seismic Intensity calculated in relation to "rock" can attain 2-3 points. This means that during a VII earthquake, which is being experienced by "rock," its Intensity in talik can reach X. It was also established that permafrost with temperature above -2°, typical for a thin layer (5-20 m), approaches, in its seismic properties, the soils in taliks.

The values for the seismic parameters of frozen soils were confirmed by the results of a laboratory study of their seismic properties. Thus, if we compare the propagation velocities of longitudinal seismic waves in gravels of the same composition, we can observe that in a gravel of temperature of -4° to -5°, the velocity of longitudinal waves is from 3,500 to 4,000 m/s, while in gravel with a temperature of around -1°, it decreases to 800-1,000 m/s.

With the aid of field seismic stations, we also studied the seismic effect of permafrost on the surface. The results from recordings of vibrations from earthquakes and experimental explosions indicated an extremely complex pat-

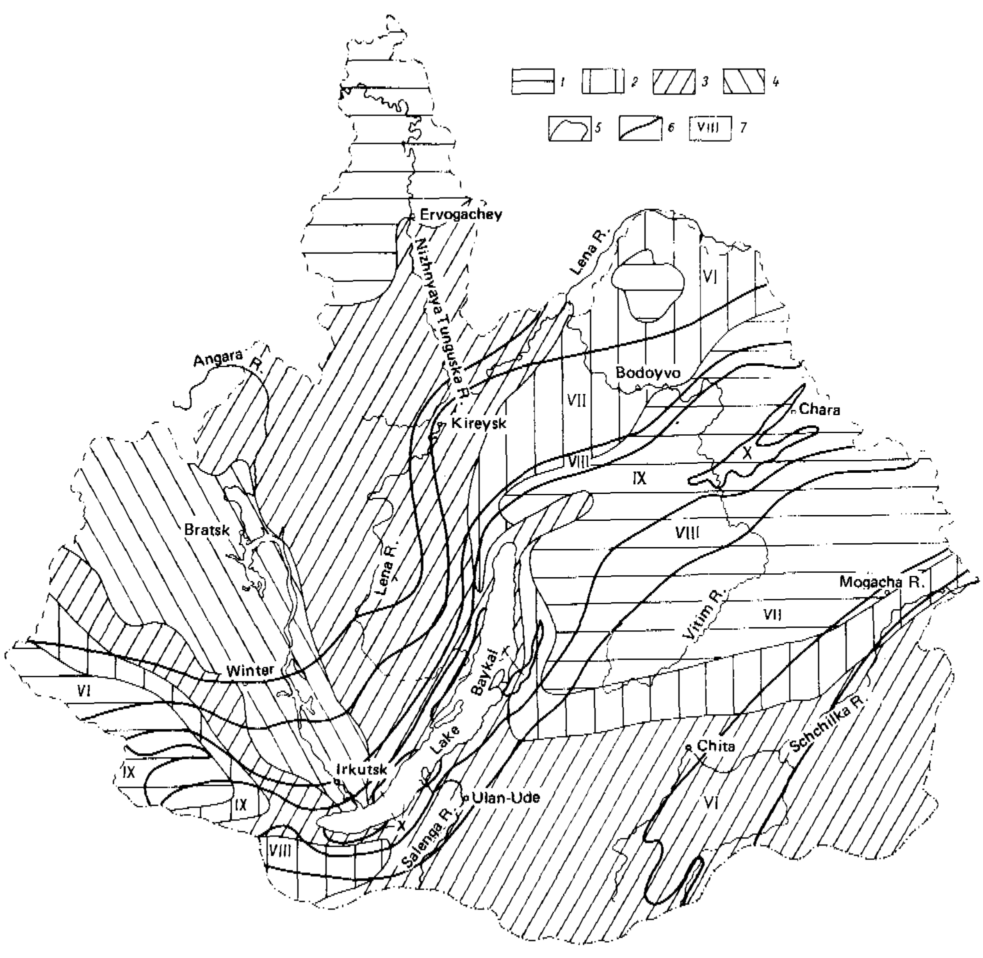

FIGURE 1 Superimposed diagram of permafrost in seismic zoning in eastern Siberia (compiled by N. Ye. Zarubin, based on data developed by L. N. Orlova, F. N. Leshchikov, V. P. Solonenko, and others). 1--regions of continuous permafrost; 2--regions with large masses of frozen ground; 3--regions of permafrost islands (rare islands and pockets of frozen ground); 4--boundaries of regions with permafrost with varying discontinuity; 5--boundaries of seismic zones; and 6--Intensity of earthquakes expressed in points on the scale.

tern of seismic behavior of soils, depending on their physical state and elastic properties. The amplitudes of oscillation in taliks and in thin permafrost layers with temperature from 0° to -1° exceed by several times the amplitudes in the rocky soils and permafrozen granular deposits with temperatures less than -2°.

Data concerning the permafrost engineering-geologic conditions and regarding the seismic properties of frozen soils obtained under natural conditions and in laboratory experiments permit us to conduct the seismic microzoning of construction sites.

However, the economic development and settlement of permafrost areas introduce considerable modifications into the permafrost conditions. In connection with this, the seismic microzoning of construction sites must be accompanied by a prediction of the variation in the permafrost and seismic conditions. The results of forecasting can be reflected on special maps, which, along with the maps of seismic microzoning for the natural condition, will take into account the changes in the seismic conditions of sites caused by the specific features of construction on the permafrost.

Even at the very outset of development of areas intended for construction, there will be sensed considerable variations in the permafrost conditions in connection with the preparation of the area for the conduct of construction operations (removal of vegetative cover, removal or packing of snow, leveling and drying of ground). This will be reflected first of all in an increase in the thickness of the seasonally thawing layer, as a result of which there will occur an increase in the seismic danger to the sites, particularly in the sectors formed of sandy silts and fine sands, having the tendency during thawing to convert to a quicksandlike and thixotropic state. Simultaneously, the intensity of thawing of subsurface ice occurring near the surface of frozen ground may increase. The slopes with thawing icy fine-grained soils will prove to be particularly dangerous. All these factors lead to a deterioration of the seismic properties of soils in the near-surface horizons.

A significant worsening of the seismic conditions of construction can take place: (a) during a decrease in thickness of the frozen layer to the point when the frozen soils during earthquakes will behave in a manner similar to the soils in the taliks; (b) in case of the through-thawing of frozen layers of slight thickness, with the formation of through taliks and also at thawing of small islands and pockets of permafrost, with their simultaneous flooding; and (c) at an increase in the temperature of permafrost, in connection with an increase in the influx of heat to the Earth's surface and with a reduction in their acoustic rigidity associated with this.

The freezing processes and occurrences during the development of a region cause unique changes. As a result of the thawing of ice in them, the frost-heave hummocks become converted to subsiding thermokarst forms, often water-filled; moreover, the subsidences can be caused by earthquakes. The development of slopes lead to an increase in the intensity of solifluction and to its rebirth. The underground jolts can activate the processes of slippage of large soil masses from the slopes. The large-scale thawing of the ice-wedge polygons, with the formation of a bed of thawed supersaturated soil, leads to a deterioration in seismic conditions in the construction area. The removal of loose accumulations from the slopes (hillside waste, talus) in some cases can increase the seismic stability of the slopes when beneath the loose material there occurs a stable slightly fractured rock mass. In other cases, when the loose mantle covers fissured ice-saturated rocks, the seismic conditions become worse.

The examples indicated do not include the entire scope of possible variations in the cryogenous and postcryogenous processes and phenomena, and their influence on the seismic properties of soils.

In order to provide an evaluation of the seismic danger of soils within the limits of actual construction areas and under individual structures, it is necessary to undertake a combination of special permafrost-type engineering-geologic and engineering-seismological studies. Having data at our disposal on the temperature regime and ice content of the permafrost and knowing the seismic parameters, we can make a prediction of the variation in seismic properties of soils, with consideration of the expected changes in natural conditions caused by the thermal interaction of soils with the buildings and structures.

For one of the areas located in Trans-Baikal North, the following predictions were made on the basis of the examination of three cases:

1. The variation in seismic properties of the permafrost under the influence of natural and artificial factors, under the conditions of an open surface.
2. Their variation under a building constructed with retention of the permafrost.
3. A variation in the seismic qualities of soils beneath a building constructed without retention of the permafrost.

In the first instance the soil consists of sand; its ice content is 20 percent. The distribution of average annual temperature and amplitudes of its fluctuation by depth are shown in Figure 2. Depth of seasonal thawing of soil under natural conditions was 1.8 m. After the development of an area, its increase to 3 m is expected. We have noted the hypothetical variation in amplitudes of soil temperature with depth. In the seasonally thawing layer, infiltrating water is present, promoting capillary moistening of the soil.

Utilizing the experimental data regarding velocities of longitudinal seismic waves in the frozen and thawed ground, obtained in the Institute of Earth's Crust at the Siberian Branch of the USSR Academy of Sciences, it appeared possible to predict their variation within the limits of the layer of annual temperature fluctuations (see Figure 2).

The most abrupt variations in seismic veloc-

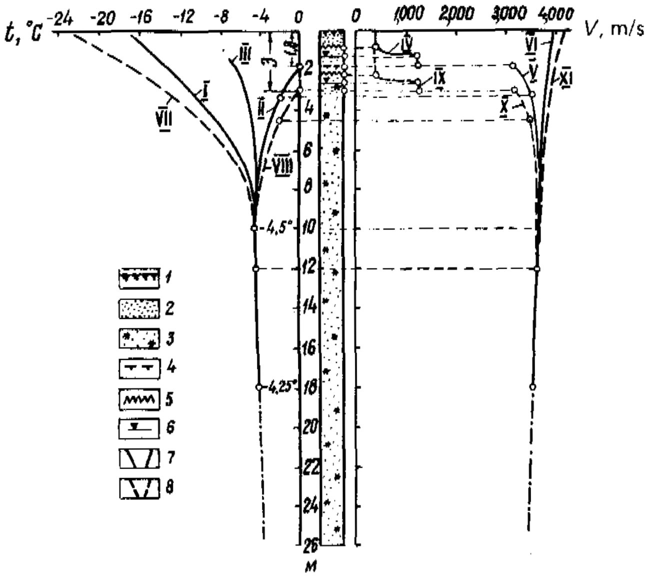

FIGURE 2 Variations in soil temperature and velocities of seismic waves with depth under conditions of an exposed surface: 1--soil-vegetative layer; 2--thawed sand; 3--frozen sand; 4--permafrost table; 5--limit of capillary rise of moisture; 6--groundwater level; 7--curves reflecting variation in soil temperature and in velocities of seismic waves by depth under undisturbed natural conditions. I--soil temperature at end of winter; II--the same, at end of summer; III--average annual temperature; IV--velocities of seismic waves in thawed ground; V--the same, in frozen ground at end of summer; VI--the same in frozen ground at the end of winter; 8--curves reflecting variation in soil temperature and velocities of seismic waves under distrubed natural conditions; VII--soil temperature at end of winter; VIII--the same, at end of summer; IX--velocities of seismic waves in thawed ground; X and XI--the same, in frozen ground at end of summer and end of winter, respectively.

ities occur in the soil of the seasonally thawing-freezing layer: from 400 m/s to 1,200 m/s in summer to 4,000 m/s in winter. The increase in the amplitudes of fluctuation in soil temperature leads to an increase in the range of variation in the seismic velocities. In winter there occurs a certain increase in velocities by depth; this leads to an improvement in the seismic conditions. Toward the end of summer, the seismic velocities decrease, thereby deteriorating the seismic conditions in the area.

In order to establish the possible variations in the seismic properties of permafrost under a building constructed with retention of the permafrost, we were required to calculate the temperature field that will form during the life of the building. The calculation was conducted according to the method developed by G. V. Porkhaev[7] with the following data: Soil conditions were taken from the preceding example; a building 15 m by 45 m with an inside temperature of 18° and an outside temperature of -7°; we took into account heat conduction under the building; temperature of air in the subfloor in winter was -5°; we assumed the corresponding thermophysical parameters of soils and of construction designs.

According to calculation, the depth of seasonal thawing of soil beneath the building will decrease to 0.83 m (Figure 3). The average annual temperature in the layer of its annual fluctuations will decrease as compared with the natural conditions, by 10 to 40 percent. There will occur a shifting of the extreme positions of temperature (for autumn and spring) in the direction of its reduction.

In conformity with the new temperature regime (see top of Figure 3), the acoustic rigidity of soils will increase correspondingly with the increase in the velocities of longitudinal seismic waves, and the seismic conditions beneath the building will improve as compared with the natural situation. In this case, in an evaluation of the seismic danger to the construction area that is being covered by buildings with retention of the permafrost, a decrease in intensity of earthquakes by 1 Intensity point is possible.

If we assume that, under these same permafrost-soil conditions, a building will be erected with thawing of the permafrost permitted (Figure 4), the thaw bowl will have a calculated depth of 12.7 m. An abrupt variation in the temperature

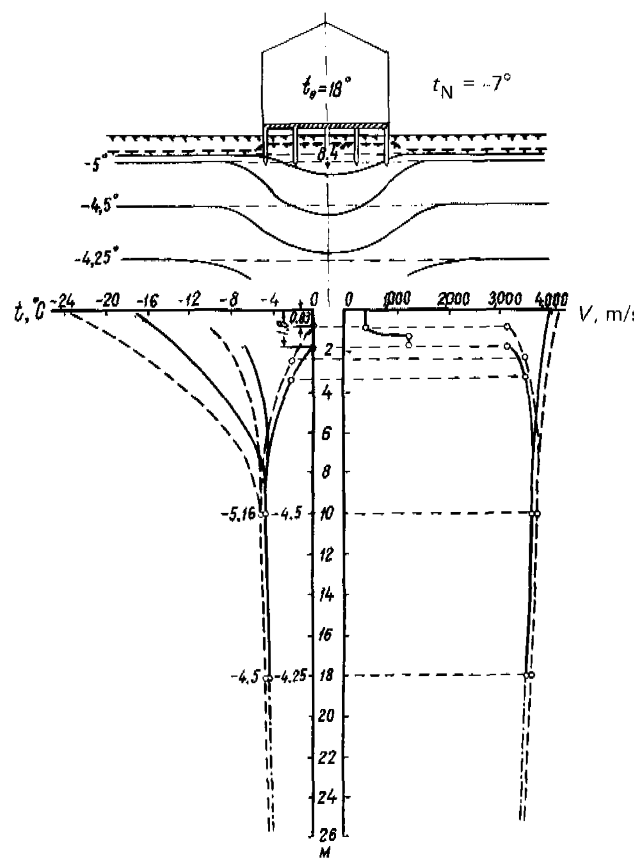

FIGURE 3 Temperature variations in soils and velocities of seismic waves by depth under a building constructed with retention of frozen base. See Figure 2 for the conventional symbols.

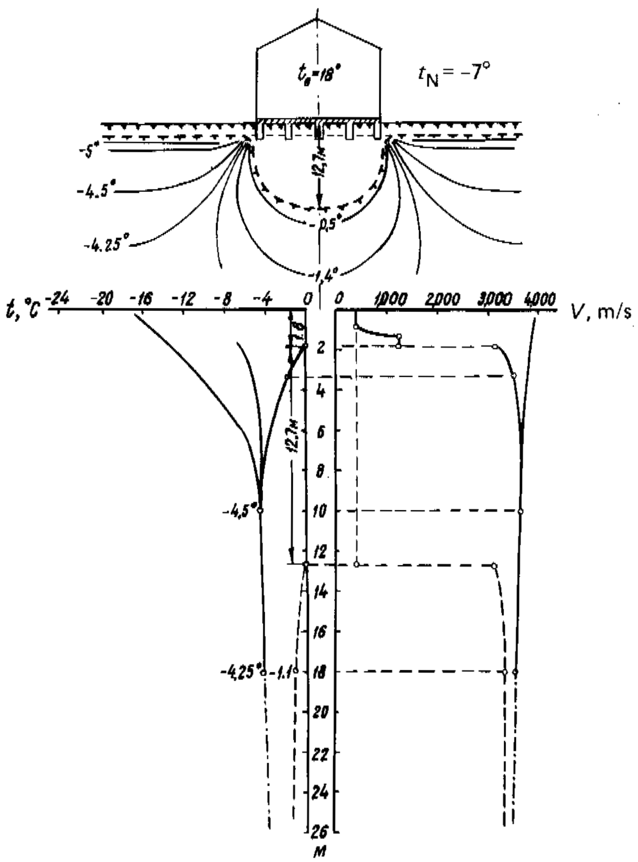

FIGURE 4 Variation in temperature of soils and in velocities of seismic waves by depth under a structure built without retention of permafrost. See Figure 2 for conventional symbols.

field in the soil mass will take place. In connection with this, in the thawing zone the velocities of seismic waves will decrease to 400 m/s to 1,200 m/s. This will lead to an increase in the seismic danger to the foundation, particularly increasing in those cases when thaw-settlement also takes place.

In an evaluation of the seismicity of a construction site where plans call for the erection of buildings and structures on thawing and thawed ground, the intensity of earthquakes can increase up to 2 points on the Intensity scale. Therefore, the buildings with considerable heat emission should be located on rocky ground where the temperature variation has little effect on the seismic properties. The thawing of fine-grained ice-rich soil in the weathering zone can lead to a seismic catastrophe.

In this manner, during the development of seismically active regions containing permafrost, it is necessary to predict the variation both in the permafrost and in the seismic conditions.

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PROBLEMS IN INVESTIGATION OF FROZEN GROUND AND SUBSURFACE WATER AS A SELF-REGULATING SYSTEM

A. B. CHIZHOV Moscow Lomonosov State University

We suggest that frozen ground and subsurface water form a complex self-regulating system. This conclusion is based on existing concepts concerning the dynamics of frozen rocks and subsurface water, the nature of their interrelationships with each other and the objects of external environment, and the capacities to reach a state of dynamic equilibrium with the varying conditions of their existence. The system under discussion consists of elements (subsystems), such as frozen ground and subsurface water. In its turn, it (the system itself) constitutes a part of more complex systems, which combines the presence of frozen ground or ice as a specific system-developing element. The natural systems, including the solid phase of water, will be referred to as cryosystems.

The very high level in the organization of the frozen ground-subsurface water system makes it necessary to differentiate in it three interrelated aspects: structure, function, and history. By structure of a system, we understand both the arrangement of the parts relative to each other and the aggregation of interrelationships between them. In our case the specific system-forming link is the thermal interaction between the frozen rocks and the subsurface water. The capacity of functioning, i.e., of specific reactions to a change in conditions of the environment is the second most important property of the system. On the basis of studying the structure and function of a system, we confront the possibility of understanding its history and predicting its behavior.

A study of systems envisages the application of a combination of mathematical procedures,[1] among which are those in geocryology and hydrogeology; at the forefront there emerge such typical cybernetic techniques as mathematical and analog modeling. In this context, it should be pointed out that the more complex the system, the more the elements of which it consists, the greater the difficulties that develop during its simulation. In connection with this, of particular interest is the application of approximation mathematical (symbolic) models. Such models are graphic and provide the possibility of an approximate quantitative evaluation of the functional relationships of the system with the environment and make it possible to estimate the interactions within the system.

The solution to the problem concerning the thermal effects of the infiltration of atmospheric water[4,8] and the prediction of the area of distribution of the infiltration-type watershed taliks[6] is connected specifically with the utilization of this method. The analysis of the effect of infiltration on depth of seasonal thawing, conducted with the use of a computer[5] also represents a very interesting experiment. Finally, it became possible to estimate to a first approximation the effect of surface water on the depth of perennial freezing.[9] It was demonstrated that a very slight rate of increase of percolation (6×10^{-3} cm/h)* is adequate to lead to a 50 percent decrease in the thickness of frozen zone.

Based on this method, the author attempted to review certain questions associated with the conditions involved in the existence of the hydrogeogenic taliks. For example, the thermal model of a pressure-infiltration talik of cylindrical form in a first approximation can be represented by the following equations:

$$q = cv\Delta t = q_1 + q_2 + q_3, \quad (1)$$

$$q_1 = cv\Delta t_1 = \frac{2}{r}\beta H, \quad (2)$$

$$q_2 = cv\Delta t_2 = \frac{t}{z}\lambda, \quad (3)$$

$$q_3 = cv\Delta t_3 = \frac{1}{\tau}Q, \quad (4)$$

where q, q_1, q_2, and q_3 are heat losses by subsurface water referred to a unit of area of talik's cross section; q = total; q_1 = for thermal interaction with perennially frozen ground surrounding the talik; q_2 = for increase in average annual ground temperature up to 0°; q_3 = excluding the seasonal freezing of thawed zone; Δt_1, Δt_2, and Δt_3 = corresponding values indicating temperature changes in subsurface water; Δt = difference in subsurface water temperatures at inlet and outlet of thawed zone; $\Delta t = \Delta t_1 + \Delta t_2 + \Delta t_3$; v = rate of filtration; β = amount of heat absorbed per unit of area in lateral surface of talik per time unit (can be calculated with a Carslaw-Jaeger formula;[3] Q = heat expended for seasonal freezing of talik; τ = duration of cold period; H = thickness of perennially frozen ground; r = radius of talik; c = specific heat capacity of water; λ = thermal conductivity of ground in the layer of annual heat exchanges; t = absolute value of average annual temperature at soil surface; and z = thickness of layer of annual heat exchanges.

An analysis of Equation (2) indicates in particular that the stability of talik increases through time T that has elapsed from the time of its formation, since the value of functions $\beta = \beta(I)$ tends asymptotically to zero with increase

*At a water temperature of about 0.5°.

in T; hence, $\Delta t_1 \to 0$ and $q_1 \to 0$. If the formation of the talik occurs through the freezing of the ground surrounding the discharge center, the role of lateral cooling of the talik will also be slight owing to the low rates of perennial freezing. From Equation (2) the effect of the r value on the talik's stability is also obvious, especially since β depends on r in the same manner as on T. From the standpoint of the questions of the system's functioning, it is interesting to note that the consequences, unfavorable for the system, of reduction in the talik's radius can be compensated within definite limits by an increase in the infiltration rate. The latter is determined by the possible elevation of the groundwater level in the area of supply and an increase in the hydraulic gradient.

We should call attention to the fact that the difference in water temperature at the inlet to the talik and at the foot of the layer of annual heat exchanges and the value $\beta(T)$ calculated on this basis provides the possibility of finding the T value. The interpretation of the T values should be conducted with aid of the available paleogeographic and paleohydrogeologic data. In certain instances, they can indicate directly the age of the thawed zone; in most other cases, they can identify the perturbation having occurred T years ago, having disturbed the system for a state of equilibrium. For example, the approximate heat calculations, as well as a number of indirect indications, testify to the extremely short age of the discharge center of subsurface water in the region of the Ulakhan-Sis Range investigated by the author, V. E. Afanasenko, and N. N. Romanovskiy in the summer of 1971.

As is obvious, the possibility of the freezing of a talik is associated chiefly with the heat losses in the layer of annual heat exchanges, which can be derived from Equatons (3) and (4). In the northern regions within the limits of the upper 20 m to 30 m of the thawed zone, we often observe the presence of islands of permafrost. The outcroppings of subsurface water are confined to the individual "windows." The area of one such "window" in the discharge center of fissured-karst subpermafrost water in the upper reaches of Seymchan River (basin of Uyandina R.) did not exceed 1.5 m^2, while the discharge reached 200 m^3/h. While the average amount of convective heat flow for this relaxation center comprised 600 kcal/h, in the "windows" it was greater by 2 to 3 orders of magnitude. Such a concentration of heat energy supports the year-round functioning of the water-bearing talik under extremely severe freezing conditions.

Another protective reaction of the system is at the place where subsurface water flows out of a reservoir; this water sharply reduces the amounts of heat ($q_2 + q_3$) required for the preservation of the talik. At reservoir depth greater than 1.5 m, they fall to practically zero.

The formulas shown permit us to estimate the effect of zonal conditions on the taliks' parameters. Thus, while for the existence of a pressure-infiltration talik with a radius of 10 m under the conditons of southern Yakutia it is sufficient to have a subsurface water flow with an input of 1 m^3/h, in the northeast of Yakutia not less than 8 m^3/h to 10 m^3/h is necessary for this. The data obtained confirm the familiar concepts concerning the features of the discharge of subsurface water in the frozen zone.[7] In conclusion, we should particularly stress that the data from permafrost-hydrogeologic studies serve as a basis for the application of the mathematical methods and a criterion for the validity of the results obtained.

Speaking of a systems approach, we also have in mind a definite viewpoint about frozen soils and subsurface water that will permit us to utilize more effectively the information accumulated concerning them. In particular, of considerable interest is a comparative analysis of the local permafrost-hydrogeologic systems existing under various zonal conditions. In this manner we can derive data concerning the mechanism of adaptation and a number of other qualities of the system, as yet little understood.

Let us limit ourselves to one example demonstrating well, in our view, the capacity of a system to adapt itself to the most severe conditions. A number of researchers noted the relationship of the valley taliks in the Northeast with the warming effect of the suprapermafrost water.[2,16] The activities of the expedition from the Department of Permafrostology from the Moscow State University in the vicinity of the Polousnyy Range indicated that the temperatures of suprapermafrost water in the seasonally thawed layer in the slope deposits do not exceed 0.2° to 0.5° and that, where they discharge directly into the valley floor, the talik is absent. The accumulation of heat necessary for the development of a talik occurs in the runoff belts and "mochazhinas" [land permanently wet from the outflow of underground water] formed by suprapermafrost water on the lower, flattened, parts of the slope. Here the water is warmed to 10° to 15° and with such a relatively high temperature that it can easily infiltrate into the deposits of the valley bottom. In the more southerly regions in the USSR Northeast, the heat energy of the actual suprapermafrost water is usually adequate for this.

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INVESTIGATION OF DEPENDENCE OF SPECIFIC RESISTANCE OF UNCONSOLIDATED FROZEN DEPOSITS IN ORIGINAL BEDDING ON CRYOGENOUS TEXTURES AND TEMPERATURES

A. N. BOGOLYUBOV Industrial and Scientific-Research Institute for Engineering Surveys in Construction

Specific electrical resistance is one of the basic parameters utilized in the permafrost-geologic interpretation of electric prospecting data and also in calculations for the "electric-needle" technique of thawing frozen ground. In a general case, specific resistance depends on several factors. While its dependence on composition, structure, texture, and ice content of frozen ground has been established fairly reliably, no unified opinion exists in respect to the influence of temperature on the variation in the specific resistance of frozen ground. Based on the results of field and laboratory observations, most of the researchers recognize the effect of temperature on specific resistance in the unconsolidated frozen ground.[1-6] However in recent years, the hypothesis has been advanced[9,10] holding that the specific resistance of absolutely all deposits in their original bedding is practically unaffected by temperature. As basic proof, the summer and winter curves of apparent resistance of vertical electric loggings obtained in various regions in the USSR Northeast for the rocks with extremely low mean annual temperatures are shown. A distinguishing feature of these VEL curves is the constancy in the values of apparent resistance above the frozen layers having a high temperature gradient.

The solution to the question concerning the extent of the temperature effect on the specific resistance of frozen rocks and soils and also the determination of the causes of this effect has great practical significance, since, along with this, we answer the question concerning the possibility of studying the cryogenous textures and temperature variations in the underlying medium with the aid of electric exploration and concerning the accuracy of interpreting the data obtained above the frozen strata.

In the process of field operations conducted in various regions of western Siberia, we obtained data that permitted us to reply fairly precisely to the question concerning the temperature effect on the specific resistances of frozen rocks in their original bedding. At their bases, there lie the results of a quantitative interpretation of the curves of apparent resistance of the parametric VEL conducted near boreholes with a known geocryologic section and temperature data. The interpretation was conducted according to the generally accepted procedure, with utilization of sets of master curves for the horizontally stratified media. In this connection, all the VEL curves, distorted in one way or another by the effect of the abruptly inclining contacts were excluded during a preliminary processing of the observations. The possibility of conducting such an operation was supported by the fact that all the VEL were conducted according to the method of two components by the AMN and MNB installations. The relatively large number of interpreted VEL curves (around 40) and valid data for the boreholes permit us to hope for the reliability of the results obtained.

An analysis of the field data indicates that the specific resistance of the main group of soils in the region (silty sands, supesses, suglinoks, and their interstratification) in a thawed condition varies within narrow limits (from 40 to 100 Ω · m). The slight spread in the values of specific resistance testifies to the uniformity of the geoelectric section of thawed soils in the region studied; this constitutes a positive conditon, since we can link all the variations in specific resistance with the variation of any factors other than lithology.

In the soils in a region, the specific resistance increases sharply on freezing. The amount of resistance jump is closely associated with the type of cryogenous texture and with the temperature of the soils.

In the case of a massive cryogenous texture, specific resistance of supesses and suglinoks increased by 5 to 8 times (Figure 1a), while that of sands increases by 10 to 15 times (see Figure 1b) in the range of temperatures from 0° to -0.5°. At a further decrease in temperature of -0.5° to -6° or -7°, the specific resistance of supesses and suglinoks increases gradually by 1.5 to 2 times. The specific resistance of sand increases much more rapidly in this range of temperatures.

In the case of the streaky cryogenous textures, specific resistance of supesses and suglinoks increases jumpwise by 15 to 20 times at the transition of temperature through 0° (see Figure 1c). At a further decrease in temperature, the specific resistance of these same soils increases gradually from 900 to 1100 Ω · m at -0.5° to 10,000 to 15,000 Ω · m at -6° or -7°, i.e., approximately by 10 to 15 times.

The results obtained in the field for soils in their original bedding agree well with the data from the laboratory determinations of the specific resistance of Salekhard sands, supesses, and suglinoks, conducted on blocks with undisturbed texture.[3] In composition and origin, these soils are similar to those in the regions which have been studied by us.

An analysis of laboratory data reveals (Figure 2) that the specific resistance of thawed Salekhard sands, supesses, and suglinoks varies from 35 to 60 Ω · m, depending on their composition and moisture content.

The specific resistance of frozen supesses with massive cryogenous texture (see Figure 2a) increases jumpwise by 2 to 3 times on variation in temperature from 0° to 0.5° and then increases gradually by another 1.5 to 2 times at drop in temperature to -10°. The variations in specific resistance noted tie in well with the decrease in the content of unfrozen water in the soil in the same range of temperatures. Practically the same pattern is noted in the sands having a massive cryogenous texture, the specific resistance of which increases at the same rate at which the amount of unfrozen water decreases in them (see Figure 2b).

A substantially different pattern is obtained during measurements on blocks of supesses and suglinoks with ice lenses of varying thickness and orientation. The jump in specific resistance on freezing of these soils is slightly greater (see Figure 2c) than in the supesses with a massive cryogenous texture. At a drop in temperature from -0.5° to -10°, the specific resistance of supesses and suglinoks with the streaky cryogenous textures increases addition-

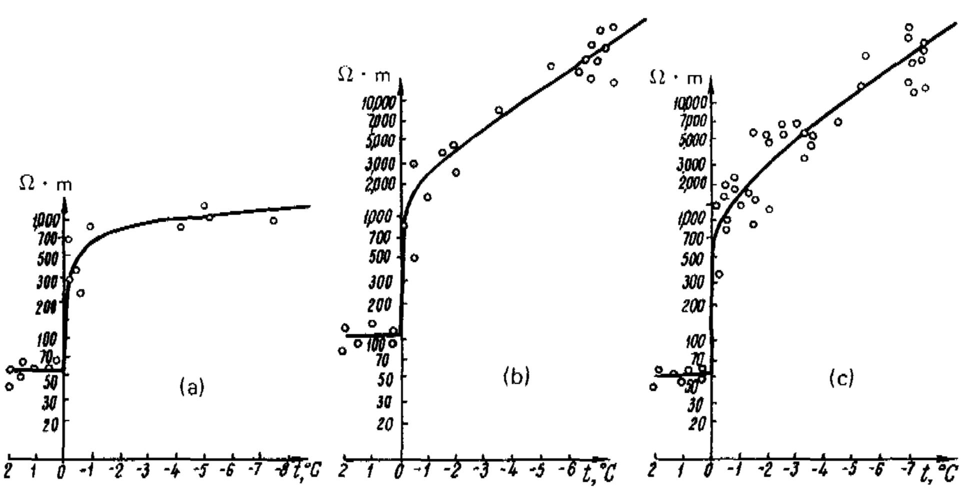

FIGURE 1 Curves of specific resistance vs. temperature of frozen soils of varying composition and cryogenous structure. a--supesses and suglinoks with massive cryogenous texture; b--silty sands with massive cryogenous texture; and c--supesses and suglinoks with streaky cryogenous textures.

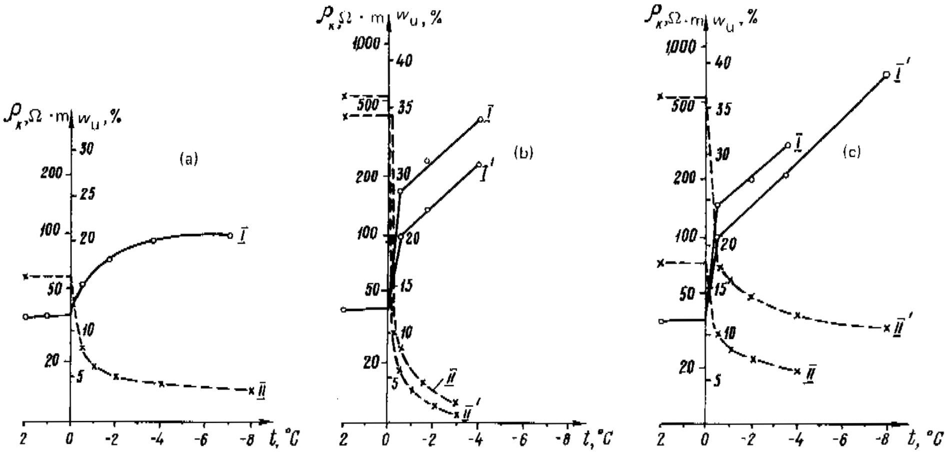

FIGURE 2 Curves of specific resistance vs. temperature and content of unfrozen water (w_u) in frozen soils of varying composition and cryogenous structure, based on laboratory data. a--supesses and suglinoks with massive cryogenous texture; b--silty sands with massive cryogenous texture; c--supesses and suglinoks with streaky cryogenous textures; I--curves indicating variation in specific resistance; and II--curves indicating variation in amount of unfrozen water.

ally by 8 to 10 times, although the content of unfrozen water in them changes in the same way as in the samples with massive cryogenous texture, i.e., by 1.5 to 2 times. At the same time, the increase noted in the specific resistance of frozen soils having streaky cryogenous textures correlates well with the variation in the specific resistance of ice at a drop in temperature.[7]

An analysis of the results obtained from the field and laboratory studies has permitted us to reach the following conclusions:

1. The variation in the specific resistance of frozen soils with massive cryogenous texture in their original bedding is closely linked with the modification of the amount of unfrozen water in them. It is likely that the dimensions and nature of arrangement of ice crystals in such soils have relatively little effect on the amount of specific resistance in the moistness (iciness) range typical of these soils.[8]

2. In the frozen soils with streaky cryogenous textures, the specific resistance in the original bedding is not determined by the quantity of unfrozen water in the mineral interstratifications, but chiefly by the type of cryogenous texture, thickness of ice lenses, and temperature dependence on the specific resistance of ice.

3. In the fine-grained soils (supesses and suglinoks), the variations in specific resistance linked with the drop in temperature are slight and in most practical cases they can be disregarded. In the soils with streaky cryogenous textures and also in sands with a massive cryogenous texture, temperature-dependence is very great and can be revealed and estimated with the aid of the appropriate electric prospecting techniques.

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EXPERIMENTAL INVESTIGATIONS OF SEISMIC PROPERTIES OF FROZEN SOILS

V. DZHURIK AND F. N. LESHCHIKOV Institute of Earth's Crust of Siberian Branch in USSR Academy of Sciences

Many experimental studies[1-4] have been devoted to an investigation of the seismic properties of frozen soils with the aid of ultrasonics. In most cases, however, the studies were conducted on the artificially prepared or randomly chosen samples of rocks. Based on these studies, we cannot say anything about the nature of variation of seismic properties of the rocks as a whole for any given region having complex permafrost and soil conditions.

For the territory in the Baikal region characterized by deep seasonal freezing, by the presence of permafrost and high seismicity, a knowledge of the elastic properties of the thawed and frozen soils has great importance in seismic microzoning.

Under laboratory conditions, we studied the seismic properties of samples of rocks (granites, marbles, limestones, dolomites, siltstones, and sandstones) selected from the weathering zone and of unconsolidated sediments (residual-diluvial silty suglinoks and supesses, residual sands, and gravels).

The velocity of seismic waves in the rock samples was measured by the IPA-65 ultrasonic device. The soil samples were frozen in the TKSI-70 refrigerating chamber. The velocity of ultrasonics was established in a broad range of temperatures (from +16° to -32°) and water content (from an air-dry state to complete saturation with water).

The velocities of seismic waves of ground in the weathered zone (to 10-15 m) for the same lithological variants are determined by the extent of weathering and by degree of fissuring, while for the frozen soils these velocities are

TABLE 1 Variation in Ultrasonic Velocity As A Function of Rock Conditions

Type of Rock or Soil	Density, g/cm³	Moisture Content %: Air-dry	Moisture Content %: Moistened	Ultrasonic Velocity, m/s: Unfrozen Air-dry	Ultrasonic Velocity, m/s: Unfrozen Moistened	Ultrasonic Velocity, m/s: Frozen at -30°
Fine-grained sandstone	2.6	0.09	9.6	1,220	1,040	3,400
Medium-grained sandstone	2.5	0.03	7.2	1,980	1,600	4,240
Dolomitic limestone	2.4	0.02	12.3	2,130	1,700	4,220
Calcareous sandstone	2.5	0.06	4.2	2,200	1,800	5,300
Fissured granite	2.9	0.03	3.8	2,800	2,800	5.100
Limestone	2.4	0.04	6.7	3,200	3,000	5,500
Dolomite-Anhydrite	2.9	0.03	0.03	5,300	5,300	5,400
Marble limestone	2.5	0.01	0.01	5,980	5,950	5,970

determined by their temperature and iciness (moistness).

Depending on their composition and properties, in the rocks, in the Baikal region, the velocity of ultrasonics varies from 1,000 m/s to 6,000 m/s (Figure 1). In the weathered sandstones, argillites, and siltstones, the velocity of ultrasonics is from 1,000 m/s to 2,000 m/s. In the metamorphosed rocks (limestones, dolomites, marbles, etc.) having high resistance to the weathering processes, the velocity of ultrasonics usually exceeds 4,000 m/s.

At a variation in positive temperatures, the increase in the ultrasonic velocity for one type of rock in air-dry state does not exceed 5 percent.
This circumstance is explained by the decrease in the effect of the soils' lithologic composition on the ultrasonic velocity in the eroded zone.

In dry rocks at decrease in negative temperatures to -32°, there does not occur a perceptible increase in ultrasonic velocity as compared with velocity in the thawed rocks. Under these conditions, the ultrasonic velocity increases by not more than 6-10 percent.

With the moistening of rocks there occurs a quite different pattern in the distribution of ultrasonic velocities. In most samples of these materials, with an increase in moisture content to 4-12 percent, velocity decreases (see Table 1).

In the metamorphosed and carbonate rocks (dolomites, anhydrites, marbles, and limestones), the velocity in the moistened rocks as compared with the air-dry ones had practically not changed.

In the frozen samples of rocks, the ultrasonic velocity as compared with their unfrozen condition increases by 2-3 times (see Figure 1). Moreover, their velocity is usually not less than 3,200 m/s. This circumstance indicates that the frozen moist competent rocks, independently of their composition and degree of erosion, in respect to velocities of seismic waves are comparable with the unweathered rocks.

The maximum increase in ultrasonic velocity in the moist rocks occurs at temperatures from 0° to -2°. At a further decrease in temperature in the rock, the velocity increases slightly.

In the unconsolidated sediments, the ultrasonic velocity at positive temperatures depends entirely on the extent of their moistening (Figure 2). Moreover, for the soils of identical lithological composition and moistness in the entire range of positive temperatures, ultrasonic velocity will remain constant. In the noncohesive soils (gravels, sands, and supesses) with an increase in their water content, the absolute velocity values increase from 500 m/s to 1,500 m/s.

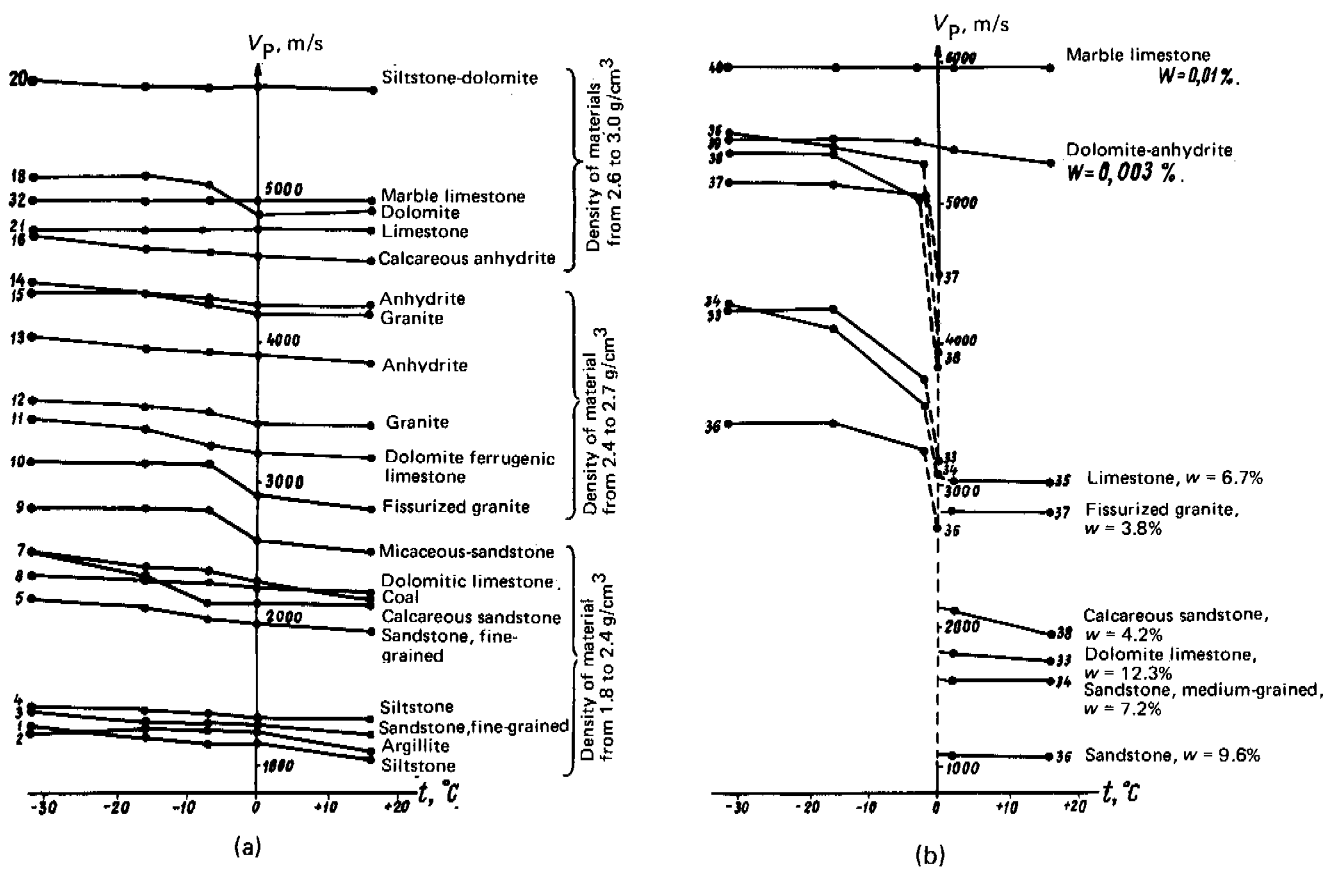

FIGURE 1 Velocity of ultransonics vs. temperature and composition of rocks and soils in Baikal region. (a) in air-dry state; (b) in moistened state.

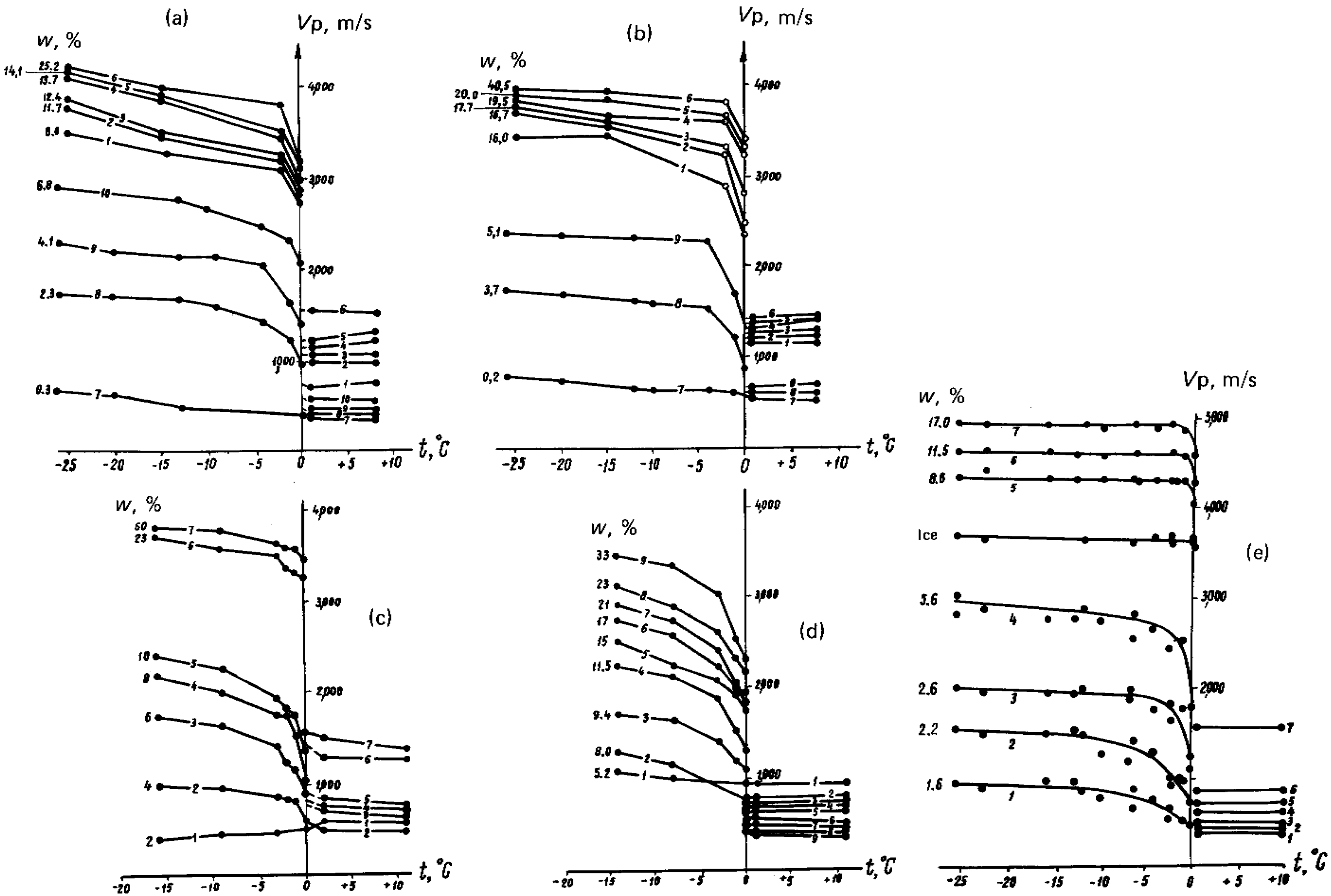

FIGURE 2 Ultrasonic velocity vs. temperature and water content in noncohesive deposits in the Baikal region. (a) in coarse-grained sand; (b) in fine-grained sand; (c) in supes; (d) in suglinok; (e) in gravel (pebbles).

Under natural conditions, in noncohesive soils during their total saturation (quicksands), the ultrasonic velocity is commensurable with the velocity in water.

In the cohesive soils (suglinoks), we note a different kind of variation of ultrasonic velocity in the range of positive temperatures. The absolute values for ultrasonic velocity vary from 300 m/s to 1,000 m/s. Moreover, in the slightly-moistened suglinoks, the velocity values are a maximum. With an increase in the moistness, the velocity of longitudinal waves decreases. A decrease in the ultrasonic velocity in the entire range of moistness investigated occurs relatively uniformly.

At negative temperatures in the unconsolidated deposits, the velocity of elastic waves increases abruptly and will fluctuate within the limits of 500 m/s to 4,000 m/s. In the slightly moistened noncohesive soil, ultrasonic velocity is comparable with that in the thawed ground. In the noncohesive soils during their freezing, a jump in velocity is noted in the temperature range from 0° to -2°. In the cohesive soils during their freezing to -16°, there occurs a gradual increase in the velocity of elastic waves. The area of phase transitions in which there transpires a more intensive increase in velocity is greater and lies in the limits from 0° to -5° through -8°.

An important feature in the sandy and clayey soils for the seismic microzoning is the abrupt variation in their elastic properties at temperature of freezing and thawing of pore water, the value of which is close to 0°. In this range of temperatures, the maximum variation in the acoustic rigidity of soil takes place.

Based on experimental data, we have obtained empirical relationships reflecting the variation in velocity of elastic waves on temperature and moisture content for each type of soil.

The material concerning the seismic properties of rocks in Baikal region can be used in evaluating the seismic effect of earthquakes, depending on the composition, temperature, and moisture regime of rocks, and in evaluating the prediction of the variation in seismic danger of soils at disruption in their natural state as a result of construction.

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RADAR PROBING OF ICE AND POSSIBILITY OF THE SOUNDING OF PERMAFROST

V. V. BOGORODSKIY, G. V. TREPOV, B. A. FEDEROV, AND G. P. KHOKHLOV Arctic-Antarctic Scientifice-Research Institute

The success achieved in a new branch of geophysics, i.e., radar probing of media, has permitted us to formulate anew the question of investigating permafrost soils.

The radar probing of media began with the sounding of glaciers, which was conducted for the first time by us in 1964 in the Antarctic and revealed the high effectiveness of the technique. Currently radar probing is being utilized for measuring the thickness of fresh ice from 0.3 m (on freshwater lakes) to 4,000 m (antarctic glaciers). In addition to the measurement of thickness during probing, it proves possible to measure the location of heterogeneous layers in a glacier, the average effective temperature of the glacier layer based on specific absorption of radar signals in the ice and average ice density based on the propagation velocity (measured by the radar technique) of electromagnetic waves into the ice.[1]

Measurements of the thickness of a glacier are conducted equally successfully from surface vehicles and from an aircraft; moreover, important advantages of these measurements are the high accuracy, clarity, and simplicity of interpreting the results obtained. In Figure 1a, we have presented the profile of a glacier obtained by the authors in the Antarctic in 1971 during the probing from an airplane.

Although the glaciers are the most suitable objects for the radar technique (low absorption in the meter range and high degree of uniformity), other objects, namely glaciers with a moraine, and freshwater basins are also successfully probed.[2] In Figure 1b, we have indicated a section of Petrozavodskaya Bay on Lake Ladoga obtained by the authors in October 1971.

For evaluating the possibility of radar probing of permafrost, we used a calculated overall attenuation of signals for the model of the medium in the form of a plane layer. The evaluations corresponding to such a model have been verified experimentally during the probing of glaciers and freshwater basins.

The total attenuation N_ε on probing a medium from the air is determined by the expression

$$N_\varepsilon = N_d + N_o + N_f + N_r + N_a + N_p$$

in which $N_d = -20 \lg G\lambda + 20h + 10 \lg 64\pi^2$ = geometric losses associated with the divergence of the wave front in space, wherein G = gain factor of the radar antenna, λ = wave-length of radar; $N_O = -20 \lg k_f$ = losses during reflection from lower limit of layer determined by the Fresnel coefficient, k_f; $N_f = -10 \lg \eta_1 \eta_2$ = focusing factor, i.e., the losses caused by refraction of rays on passage of wave from air into the medium (η_1) and back (η_2). The expressions for the determination of $\eta_1 \eta_2$ are presented in Brekhovskikh.[5] N_r = losses caused by reflection from boundary--air-medium and intermediate boundaries--and also from the volumetric heterogenities; for a homogeneous medium $N_r = -20 \lg (1-k^2 f_1)$ where $k f_1$ = Fresnel coefficient for air--medium boundary; $N_a = N_u \cdot h$ = losses to absorption in medium determined by electric properties of medium (N_u) and thickness (h) of layer; N_p = losses owing to noncoincidence of polarization of reflected signal and polarization of receiving antenna developing on probing of media with expressed crystal structure (glaciers).

The overall weakening is determined mainly by thickness of probing layer and by specific absorption of medium (Figure 2). The remaining components depend little on properties of medium and boundaries, i.e., on type of permafrost.

Since prior to the start of work on clarifying the possibility of radar probing of permafrost materials their properties in the radio band were inadequately explained, we took measurements of electrical parameters of frozen material both in artificially prepared mixtures (rock-ice system) and in soil samples taken from the permafrost zone.[3,4] The frequency relationships of specific absorption of electromagnetic waves in various types of permafrost are shown in Figure 3; for the sake of comparison, in the same place we have indicated the specific absorptions of freshwater ice, fresh water, dry and wet sand (thickened lines).

A combined study of Figures 2 and 3 shows that the radar equipment of average capacity with receiving channel sensitivity of around 160 db relative to the power being transmitted is capable of providing the radar probing of frozen materials in the worst case (curve 1, Figure 3) for a depth of around 15 m. In more favorable cases, the frozen materials and the rock-ice system can be probed for a depth of 100 m and more. It is interesting to know that the presence, in the permafrost, of thawed layers of the moist sand type (curve 5, Figure 3) does not constitute an obstacle for the conduct of probing, since the losses in the sand do not exceed those in water (curve 2), although of course from such a layer the signal will be received as a result of contrast in respect to relative dielectric constants (ε^1).

The important factors restricting the possibilities of radar probing of materials are interferences, among which we include:

1. Scattering at upper boundary on probing from air.
2. Scattering by infravolume heterogeneities in the media.
3. Interference in the ether, i.e, radio broadcasting, communication, TV, and industrial static.
4. Direct passage of the transmitter signal into the receiver input.

An estimation of the levels of interference and optimization of the equipment parameters in

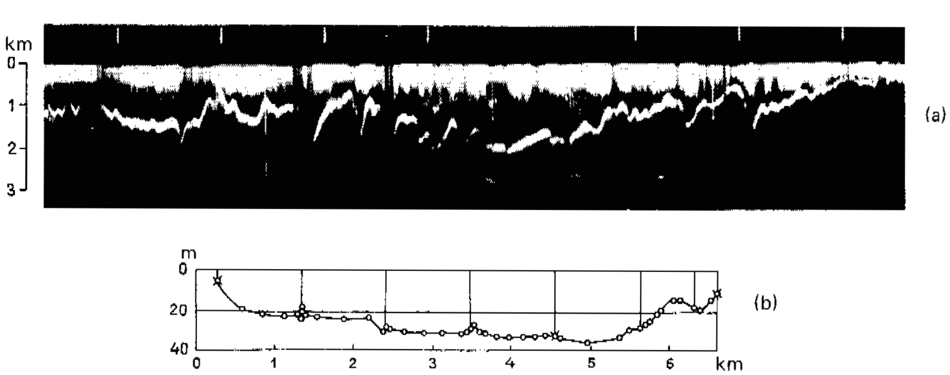

FIGURE 1 Profiles obtained by radar probing. (a) Antarctic glacier in vicinity of Molodezhnaya Station; length of course about 160 km; (b) Petrozavodskaya Bay on Lake Ladoga.

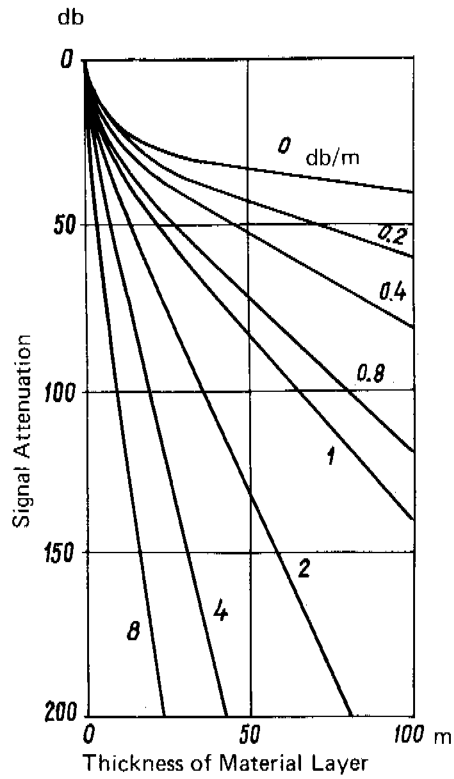

FIGURE 2 Overall attenuation of radio signal vs. thickness of material layer on variation of specific absorption from 0 db/m to 8 db/m.

order that it would provide a probing of permafrost within the prescribed limits of the soils' thicknesses comprises an independent problem, the solution of which goes beyond the scope of the present report. However, we should point out that the application, during probing, of specific procedural and technical methods should guarantee the measurement of both thin and thick layers. Among such methods, we should include the utilization of high-capacity stations and nanosecond technology, the slant probing of the medium, and the loading of antennas in the medium during sounding.

A special question is the interpretation of the signal received and the calculation of the medium's parameters based on those of the signal.

The interpretation of the signals received is quite simple only during the probing of uniform materials (permafrost of the frozen sand type) having the same reflecting boundary. Certain methods of interpretation of complex reflected signals have been reviewed in the report devoted to the probing of ice and glaciers, i.e., the processing of reflected signals based on lag and attenuation. As one of these methods, facilitating the segregation of the signal from the interface against the background of reflections from heterogeneities, we can utilize the technique of oblique probing at various distances between the transmitting and receiving antennas.

For a comparison of the material's parameters (depths of boundaries) with the signal parameters (time lag), we require a knowledge of the velocity of propagation of the electromagnetic waves in the medium. For an evaluation of the velocity, in a first approximation we can utilize the data based on relative dielectric constants (ε^1) and tangent of the angle of electric losses ($\tan \delta$) obtained during measurements of samples. The refinement of the value for the velocity can be achieved by comparing the data from radar probing with the results obtained by control drilling or we can conduct a measurement of the velocity by the oblique sounding technique.

In the performance of sounding in limited sectors, the sampling drilling obviously can feasibly be performed if only for providing an interpretation of the reflected signals at the drilling point in order that the results for the reference point could be extrapolated to the entire sector under investigation.

The concepts expounded, including the analysis of the overall attenuation of the radar signal during probing of permafrost and the electrical properties of frozen materials, moist sand,

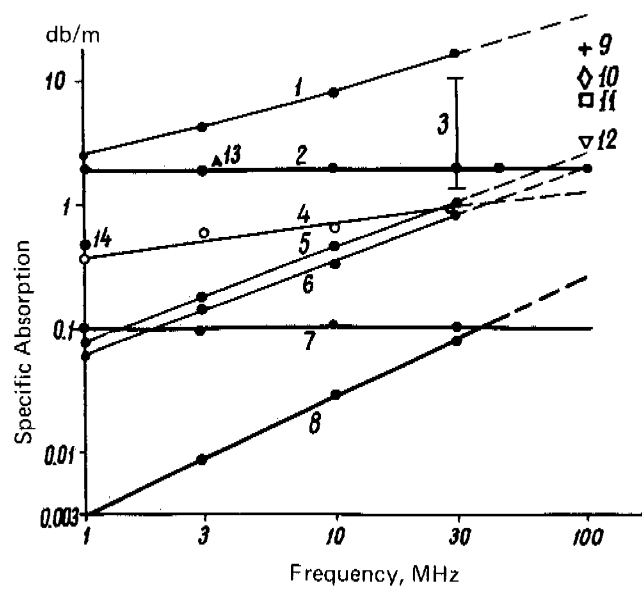

FIGURE 3 Frequency dependencies of specific absorption in various materials. 1--Arctic-tundra surface-gleyey sodded topsoil (Vaygach), 30-40 cm horizon-transition layer with soil-forming materials, -5°C;[3] 2--lake water, +10° (measurements of authors); 3--gleyish podzolic soil (Vorkuta), tundra gleyish retinized [serpentinized] soil (Taymyr), Arctic-tundra surface--gleyish soddy soil (Vaygach), -15°C;[3] 4--moss cover on ice (Vaygach), -5°C;[3] 5--sand, water concentration by volume 0.1, +18°C (measurements by authors); 6--rock-ice system, concentration of mineral material by weight 0.75, -10°C;[4] 7--freshwater ice, -1°C;[7] 8--dry sand, +18°C (authors' measurements); 9--clay with high Fe content, -35°C;[6] 10--wet clay, -35°C;[6] 11--brown soil, -35°C;[6] 12--clay with high Al content, -35°C;[6] 13--partly frozen ground in Alaska;[7] 14--frozen ground in the Antarctic, -20°C.[7]

fresh water, and ice demonstrate the basic possibility of the radar probing of permafrost to depths of practical interest. The availability of the equipment in the meter band developed for the sounding of glaciers and fresh water and in a number of cases suitable for application on permafrost objects permits us to postulate the accomplishment of the initial practical steps in the immediate future.

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STUDY OF UNDERGROUND ICE BY THE METHOD OF SEISMIC EXPLORATION

O. K. VORONKOV AND G. V. MIKHAYLOVSKIY All-Union Scientific Research Institute of Hydrogeology and Engineering Geology

The determination of the forms and dimensions of underground ice in the foundations of engineering structures is an important problem in engineering geology and permafrostology. A common type of large accumulations of underground ice is representated by the multiple-wedge ice, i.e., the product of repeated ice formation in the frost-fissures of fine-grained soils (suglinoks, supesses, etc.).[4]

For the study of underground ice, at the present time extensive use is made of drilling and sinking, while among the geophysical methods we utilize electric profiling.[2] A new approach to a quantitative study of underground ice involves the use of the seismic prospecting for shallow depths. The physical prerequisites of applying this technique for solving the given problem include the dependence of velocity of elastic waves on volumetric ice content and distinct velocity differentiation of ice-poor fine-grained deposits from ice. The investigation of underground ice by the technique of seismic profiling and seismic well-logging have been conducted by the authors in sectors of hydraulic construction during the summer-fall period (July-September) in the valleys of the Irgichan River (basin of Indigirka River; mean annual air temperature, $t_a = -15°$, temperature of soils at depths of 15-20 m, $t_s = -6°$ to $-8°$) and of the Tatta River (Lena-Amgin interfluve, $t_a = -11.6°$, $t_s = -4°$ to $-6°$).

The generalized seismic-geologic section is represented in three layers. Layer one includes the seasonal thawing of suglinoks and supesses with a thickness from 0 m to 2 m to 3 m. According to velocity V_P of longitudinal waves, this is a homogeneous or gradient layer. In it the average V_P equals 400-800 m/s. Layer two includes the frozen suglinoks and supesses containing ice of various forms, dimensions, and origin. The slightly and moderately icy soils in this layer have V_P = 2.4-3.2 km/s, while the ice will have a value V_P = 3.3-4.0 km/s. Layer three includes the frozen bedrocks of various compositions of which V_P = 4.0-5.0 km/s. Such a structure of the section's upper part is also typical for other regions where underground ice occurs.

The basic technique of observations was the continuous longitudinal seismic profiling (correlation method of refracted waves) along lines separated from each other by a distance of 5-7 m. We utilized the seismic station SS = 24 with increased accuracy of reading the time on the seismogram (markers every 5 ms at their width of 4-5 mm). The spacing of the seismic receivers was 2.5 m. The number of explosion points (6-7 for each installation of receivers) assured completely interlinked systems of hodographs [travel-time curves] of refracted waves corresponding to the surfaces of layers two and three.

In the course of the investigations, we recorded longitudinal direct and refracted waves in layer one, longitudinal refracted waves corresponding to the surface of layers two and three and also surface (Rayleigh) waves. The interpretation of the seismic prospecting data permitted us to solve a number of problems pertaining to the investigation of underground ice.

The mapping of the surface of underground ice was conducted based on the V_P values corresponding to the surface of layer two. To the sectors of ice occurrence, there corresponded V_P = 3.2-3.3 km/s. An additional criterion of ice was the decrease above it of the t_0 values corresponding to the surface of layer two, since the seasonal thawing above the ice wedges is less than in the supporting rock.[1] By seismic exploration mapping in the basin of the Irgichan River, it was established that ice ($V_P > 3.2$ km/s) occupies a large part of the area of the above-flood plain terrace (Figure 1a). By mapping in the basin of the Tatta River, we also established the predominance of ice in the upper part of the section of the interalassial remnants (Figure 1b). We surveyed the ice-rich sectors with V_P = 3.3-4.0 km/s (ice content by volume i^1 = 80 to 100 percent, wherein the ice content by volume in ice pockets i^1 = 75-100), moderately icy sectors with V_P = 3.0-3.2 km/s (t^1 = 60-75 percent, i^1 = 50-70 percent), and relatively ice-poor sectors with V_P = 2.4-3.0 km/s (i^1 = 20-60 percent, and i^1 = 0.50 percent).

Determination of height h of ice wedges. For the purpose of solving this problem radioscopy from beneath is applicable (on refracted waves corresponding to the surface of layer three). For the calculations of h at each point of the receiving stations, a formula has been suggested, obtained by analyzing the report by F. M. Lyakhovitskiy:[3]

$$h = \left[h_2 - \left(t_{0_2} - t_{0_1}\frac{\cos i_{13}}{\cos i_{12}}\right)\frac{V_C}{2\cos i_{23}}\right] : \left(1 - \frac{V_C}{V_I}\right),$$

Where h_2 = thickness of layer two; t_{0_1}, t_{0_2} = vertical time corresponding to the bases of layers one and two, respectively; i_{12} = arc sin V_1/V_2; i_{13} = arc sin V_1/V_3; i_{23} = arc sin V_2/V_3; V_1, V_2, V_3 are values for wave velocities in layers 1, 2, and 3; V_I and V_C are values for waves in ice and in ice-poor fine-grained soils

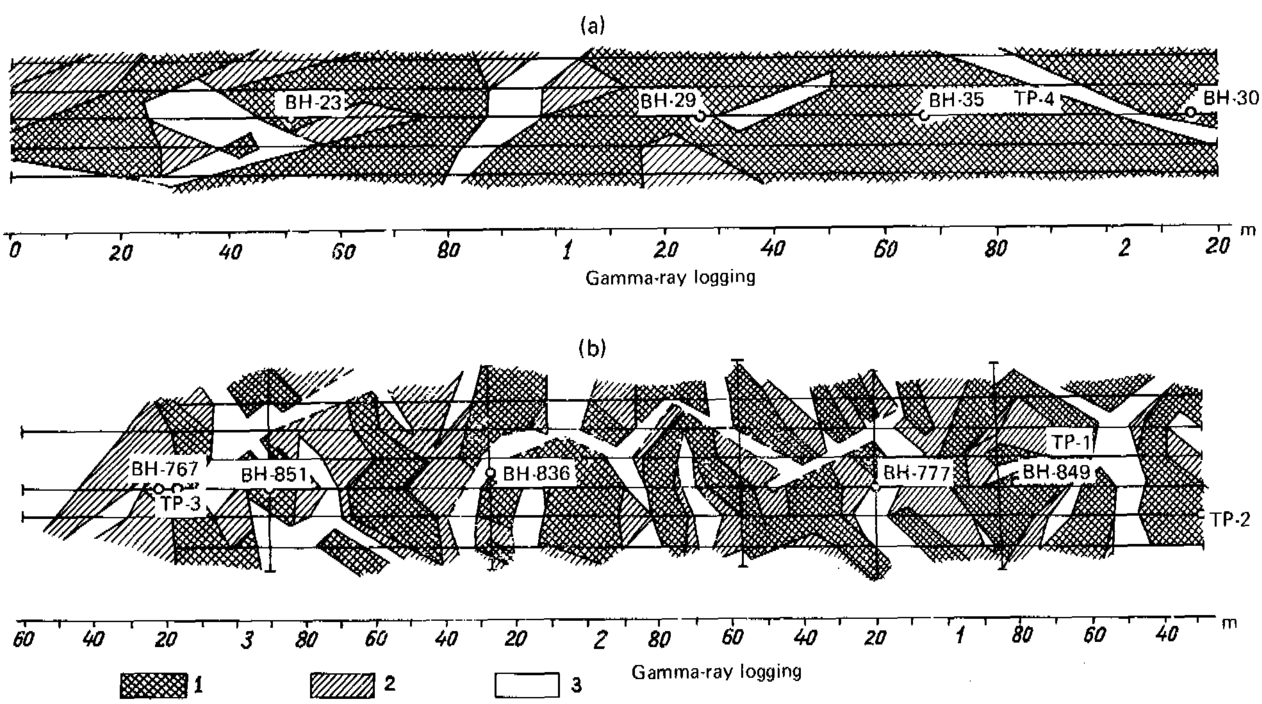

FIGURE 1 Diagrams showing distribution of underground ice in upper part of section of perennially frozen suglinoks and supesses. (a) sector of above-flood plain terrace in Irgichan River valley (basin of Indigirka River); (b) sector of interalasses in the Tatta River valley (Lena-Amgin interfluve). 1--sectors with V_P km/s--ice wedges (ice content by volume i^1 = 80-100 percent; 2--sectors with V_P 3.0-3.2 km/s (i^1 = 60-75 percent); and 3--sectors with V_P = 2.4-3.0 km/s (i^1 = 20-60 percent). BH = borehole; TP = testpit.

(supesses, suglinoks) [respectively] in layer two. In the suglinoks (basin of Tatta River), V_C = 2.55 km/s; for the interbedding of suglinoks and supesses (basin of Indigirka River), V_C = 2.4 km/s. According to borehole logging data, we assume a velocity value of 3.8 km/s for ice.

The *h* values plotted on the seismic geologic sections (Figure 2) permit us to estimate the forms and dimensions of ice bodies and also the depth to their surface. The vertical thicknesses of the repetitive ice-wedges determined from seismic exploration and drilling (see Table 1) agree satisfactorily with each other; this testifies to the possibility of an objective estimation of *h* by the seismic exploration technique. In solving the problem of determining *h*, it is desirable to utilize seismic recording equipment permitting us to determine the times of wave-arrival with precision of ±0.0001 s.

The estimation of ice content by volume i^1 is based on V_l measured under natural conditions. A comparison of V_P *in situ* based on profiles and in boreholes with laboratory determination of i^1 for soils from the same sectors permitted us to establish, for the suglinoks in the Tatta River basin, the relationship:

$$i^1 = \left(2.38 - \frac{5.24}{V_P}\right) \cdot 100\%,$$

where V_P is in km/s (at V_P 3.8 km/s, i^1 is assumed = 100 percent). Calculations based on this equation were utilized for the characteristics of sectors with varying V_P values derived from seismic prospecting during mapping. The data from the laboratory study of the interrelationship of i^1 and V_P in the suglinoks and supesses proved not to be comparable with the results in situ.

Determination of volumes of underground ice. The vertical sections and horizontal plans based on the seismic prospecting data permit us to calculate the ice volumes and to estimate their distribution on the horizontal and vertical.

We recommend using the seismic exploration method in preliminary surveys and detailed studies of underground ice. During the preliminary survey, seismic prospecting, in addition to the study of ice, solves a range of problems in engineering-geologic mapping.

In detailed studies of underground ice for refining the *h* and V_C, seismic prospecting should be combined with drilling and electric profiling.

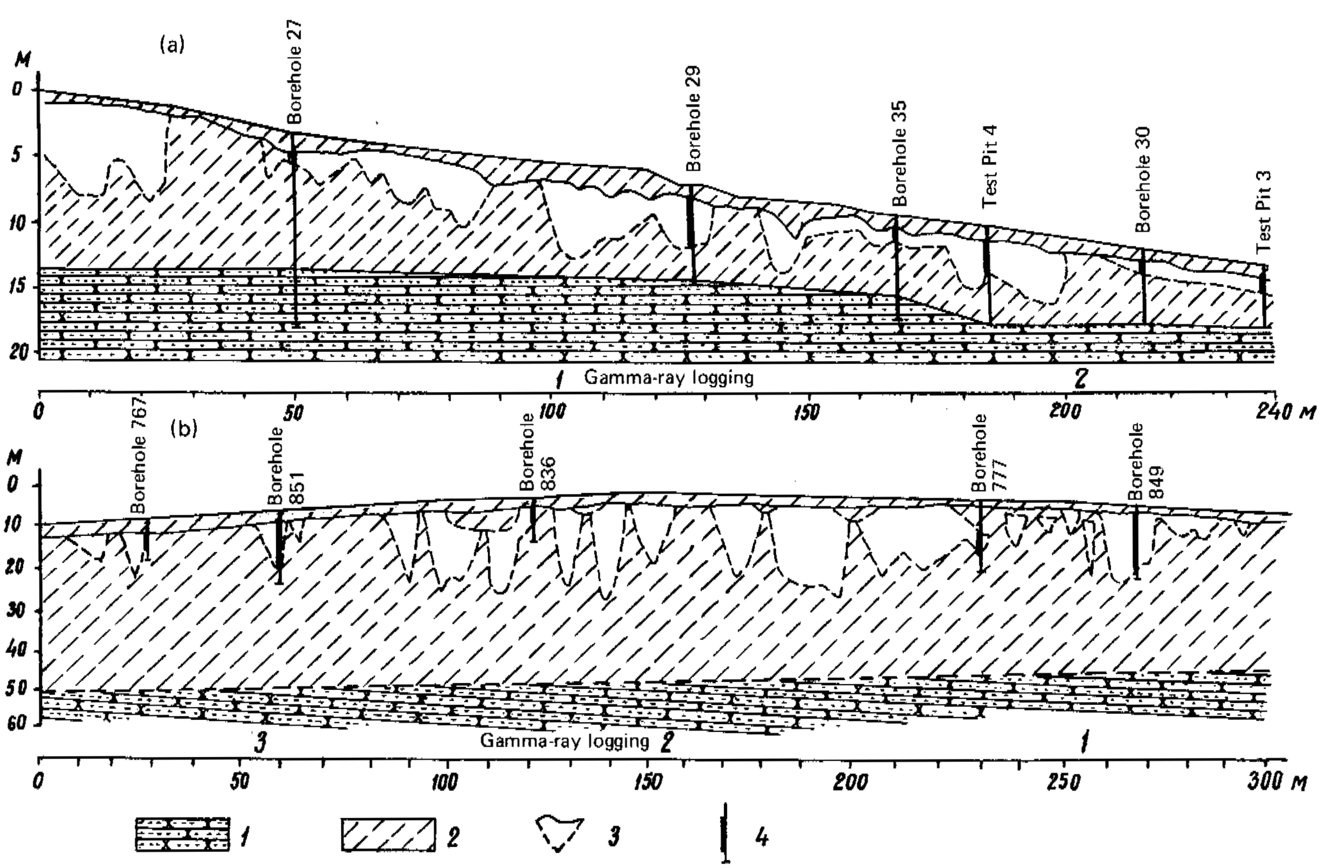

FIGURE 2 Seismic geologic sections in sectors of investigations. (a) in Irgichan River valley (b) in Tatta River valley. 1--sandstone; 2--suglinok; 3--ice wedge; 4--ice uncovered by mine workings.

TABLE 1 Distribution of Vertical Thicknesses of Underground Ice Based on Data from Seismic Exploration and Drilling in the Valleys of the Irgichan and Tatta Rivers (in % of Total Number of Determinations by the Appropriate Method)

	Irgichan River Valley		Tatta River Valley	
h,m	Seismic exploration (76 determinations), %	Drilling (49 determinations), %	Seismic (362 determinations), %	Drilling (52 determinations), %
1	42.0	31.2	10.2	11.5
2	9.0	12.5	8.2	4.0
3	12.5	8.0	11.0	7.7
4	20.5	18.3	8.2	11.5
5	8.0	6.0	7.7	13.5
6	3.0	--	9.0	15.5
7	5.0	10.0	7.0	7.5
8		4.0	5.0	5.7
9		6.0	5.5	6.0
10		4.0	4.7	5.7
11			2.5	3.7
12			3.5	--
13			2.5	5.7
14			2.5	--
15			3.5	--
16			2.0	--
17			1.5	--
18			2.0	2.0
19			1.5	--
20			1.5	--
21			0.5	--

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POTENTIALITIES OF SEISMIC-ACOUSTIC TECHNIQUES IN ENGINEERING-GEOLOGIC INVESTIGATIONS FOR CONSTRUCTION OF PERMAFROST

YU. D. ZYKOV AND YU. I. BAULIN Industrial and Scientific-Research Institute for Engineering Surveys in Construction

The considerably expanded volume of construction of industrial and residential buildings in the northern regions and the necessity of increasing the rates of this construction require a reduction in the periods of permafrost engineering-geologic prospecting and an increase in its quality. In the accomplishment of this task, a significant role is relegated to the geophysical techniques.

Nevertheless, owing to the inadequate number of special studies, both on the adoption of new methods and on the disclosure of the resolving capacity of the traditional methods, the best possibilities of geophysics are not being used to the full. This also pertains to seismic-acoustics.

The results of laboratory studies[2-4,6,7] constitute a prerequisite for the successful utilization of seismic-acoustic techniques for permafrost engineering-geologic exploration. The investigations conducted recently have established an entire series of relationships in the variation in the propagation rates of elastic waves on variation in temperature and composition of frozen soils.[5] The increase in moisture content at constant porosity of soil, i.e., an increase in the extent of saturation (q) causes an increase in the propagation velocity of elastic waves according to a law that is almost linear:

$$V \approx kq + a,$$

where k and a are coefficients depending on porosity, lithology, and temperature. In case of total saturation ($q = 1$), the effect of water on the velocity is qualitatively identical for soils of varying grain size and composition. An increase in the water content of frozen soil within a fixed range of negative temperatures leads to an increased velocity, while at lower temperatures it leads to a decrease. The extent of the indicated range in temperatures depends on the lithological composition of the material.

In order to disclose the actual possibilities of seismic acoustics during permafrost engineering-geologic investigations, for the most proper formulation of these studies and an interpretation of the data obtained, we conducted field operations of an experimental-procedural nature in the regions of the polar Urals and in the western Siberian plain. At this stage of operations, special attention was given to the possibility of the most complete engineering-geologic description of the areas under study.

In the regions of the polar Urals, the main problem of seismic exploration includes mapping the top of the bedrock and determining the thickness of the weathered crust. With the aid of a 12-channel seismic installation with impact excitation of signal, we obtained hodographs of the first arrivals of longitudinal waves. Based on these data, in the section we differentiated four velocity boundaries. The first with a velocity of 4,000 m/s to 7,000 m/s corresponds to the surface of the bedrock. The depth of this boundary will fluctuate chiefly from 5-8 m to 15-20 m. The second, i.e., 500 m/s to 3,600 m/s, corresponds to the surface of the strongly eroded bedrock represented by incompetent rock with coarse rock waste and loamy deposits. This boundary is traced everywhere at depth ranging from 3-5 m to 10-12 m. The two boundaries characterized by velocities ranging from 2,000-2,200 m/s and 1,000-1,800 m/s occur only in individual sectors and are deposited at depths ranging from 1-3 m to 5-7 m. These boundaries are caused chiefly by the alternation in lithology and iciness of unconsolidated rocks (soils) representated in the upper part of the section. For studying the potentialities of the seismic-acoustic techniques in the regions where sandy-clayey frozen soils occur, operations were conducted in a test area in the lower reaches of the Ob' River. The combination of geophysical activities included seismic profiling, seismic-acoustic observations in boreholes and mine workings, and also laboratory ultrasonic measurements.

The purpose of the laboratory investigations included the establishment of the dependence of the propagation velocity of longitudinal waves on temperature of frozen soils represented in the region under study (Figure 1). The water content and degree of saturation of artificially prepared soil samples in these experiments corresponded to the natural specimens. The results obtained from laboratory measurements agreed well with the field findings. This indicates the possibility of obtaining data not only for interpretating the field materials, but also for establishing correlations between velocities of elastic waves in frozen soils and their mechanical properties also under analysis in the laboratory.

During the ground-based seismic observations in an area formed of sandy-clayey soils, we can trace well the refracted and reflected waves of the PSP type associated with the permafrost surfaces. The first of them are characterized by

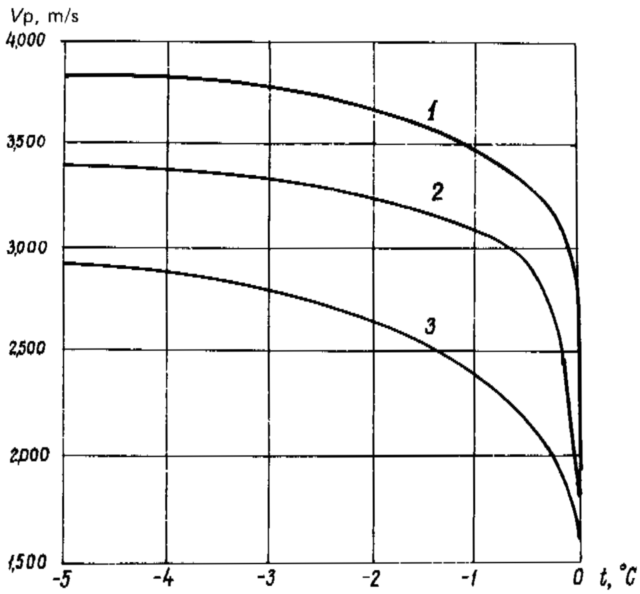

FIGURE 1 Velocity of longitudinal waves V_p vs. temperature of frozen soils. 1--sand, water content 27 percent; 2--suglinok, water content 31 percent; 3--clay, water content 42 percent.

velocity values of 2,000-2,600 m/s and in rare instances from 3,000 to 3,500 m/s. An increase in velocity V_P in certain sectors is mainly linked with the increase in the clays' iciness. Velocity of the *S* wave V_S will vary from 1,000 m/s to 1,500 m/s. The thawed soils in the area under study have a velocity of longitudinal waves of 250-350 m/s. Similar velocity values have also been obtained during ultrasonic measurements in pits. The data from seismic profiling permitted us to determine the thickness of the seasonally thawed layer varying within the limits of 1.5 m to 2.5 m and also to judge the lithologic continuity of the section and the iciness in the upper part of the frozen ground.

The investigations in the dry boreholes included ultrasonic logging (UL),[1] which was accomplished by three-point probing with recording of the longitudinal and Rayleigh waves and nonlongitudinal vertical seismic profiling (NVSP) with a three-component borehole probe. The results obtained in one of the boreholes have been shown in Figure 2.

The comparison of the UL data (Figure 2; V,3,4) with such parameters as total moisture (II) and temperature (III) shows their good agreement. The increase in moisture and decrease in temperature leads to an increase in V_P and V_R. The local maximums in velocity are coordinated to fairly thick (about 20 cm) interstratifications of ice in the clays.

Based on the results of the NVSP, we obtained a series of vertical hodographs at varying distances of the impact point (IP) from the borehole mouth. The observed vertical hodographs of the longitudinal wave (VI,6) agree well with the hodographs calculated (VI,7) for the case of stepped-linear velocity section chosen based on the UL data (V,5). The average velocities obtained also tie in well with these data.

In a review of the seismograms obtained at varying depths, attention is drawn to the phase inversion (VII), coordinated to specific depths. For the system ZZ (VII,8) (vertical impact, vertical seismic receiver) this depth is about 7 m; this is linked with the variation in the sign of velocity gradient as a result of variation in soil iciness and increase in temperature. The reversals of phases in the horizontal seismic

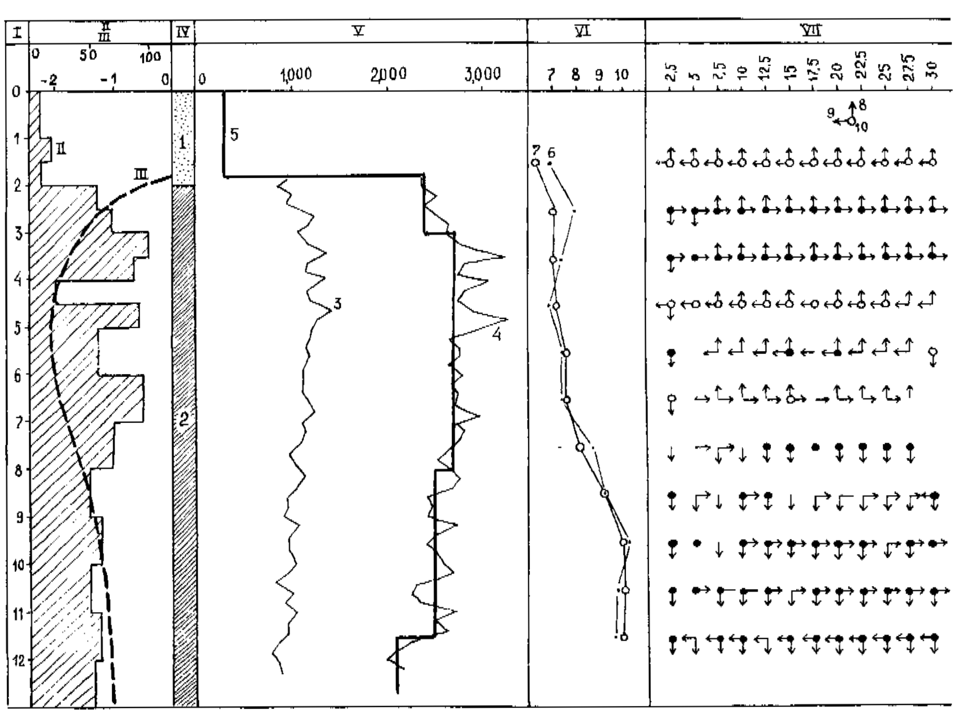

FIGURE 2 Results of borehole seismic-acoustic observations. I--depth, m; II--overall water content, percent; III-temperature, °C; IV--geologic section: 1--sand, 2--clay; V--curves of ultrasonic logging, m/s; 3--velocity of Rayleigh waves, 4--velocity of longitudinal waves, 5--averaged velocity section; VI--vertical hodographs of first arrivals of a longitudinal wave, ms: 6--observed; 7--calculated; VII--diagram showing variation in phase of observed waves for systems; 8--ZZ, 9--ZX, 10--ZY at various distances of impact point from top of borehole, m.

receivers (ZX, ZY) (VII,9,10) are recorded at depths of roughly 2, 4, 6, and 11 m. In this case the phase inversion can be explained qualitatively by the influence exerted by the variation in velocity anisotropy by depth as a result of the change in thickness and in orientation of the ice lenses.

In this way the studies conducted have demonstrated that in the case of an increase in velocities in frozen soil with depth, base rock and weathered crust, ground-based seismic prospecting yields reliable results. For studying the sandy-clayey frozen soils, it is necessary to utilize ground-based seismic, borehole seismic-acoustic, and laboratory ultrasonic research. Such a combination will permit us to study the structure of an area in plan and in section and also to obtain the extensive necessary data concerning the physicomechanical properties of frozen soils.

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CONCERNING THE NATURE OF AN INDUCED POLARIZED FIELD IN CONNECTION WITH THE SURFACE POLARIZATION AT THE BOUNDARY BETWEEN THAWED AND FROZEN SOILS

V. P. MEL'NIKOV   Permafrost Institute of USSR Academy of Sciences

The investigations of induced polarization (IP) of ion-conducting soils in samples and under field conditions permitted us to determine the general principles of its development and disappearance through time.

The overall approach to an estimation of the IP of ion-conducting soils is based on the fact that this is a volume polarization originating in each elementary volume of soils and depending on the dispersed state of the medium and a number of other factors that create an electric field coinciding in sign with the polarizing field in the process of charging and discharging. The copious data permitted us to develop a model of an ion-conducting medium and to theoretically substantiate the IP phenomenon on the basis of electrochemical processes occurring in the heterogeneous media under the effect of electric currents. There are statements[1,3,7] concerning the existence of processes not embodied in the general concepts concerning the induced polarization of ion-conducting soils. This pertains primarily to the hypothesis concerning surface polarization. Some authors[3] have attempted to explain from these standpoints the induced polarization of clays; others[1] based on theoretical calculations, have attempted hypothetically to connect the possibility of the origin of the surface polarization effect at the interfaces of soils with varying permeability. This same author[1] draws attention to one of the principle tenets differentiating surface polarization from volume polarization, namely that the time-variation in emf developing at the boundary of soils with varying structure of pore space is determined chiefly by the passage time of the exciting current and not by the structure of the pores. This scant information concerning the surface polarization stimulated interest in the problem of segregating the surface polarization effect, although the extensive data obtained during the experiments testified to the predominant influence of the volume polarization of ion-conducting soils on the variation in the IP potentials.

The investigation of the IP processes in deposits in frozen condition[4-6] opens new possibilities for establishing the nature of polarizability of heterogeneous media. The uniqueness of frozen soils, associated with the structure of pore space and with the presence of distinct boundaries (frozen soil-thawed soil) determines the necessity of considering a new model in a description of the effects which are taking place. The purpose of the present report is to draw the attention of researchers to the surface polarization of cryogenous boundaries and, in connection with this phenomenon, to the establishment of the IP potentials in time. It should be pointed out that during the observation of the induced polarization at the Earth's surface according to the normal procedure of vertical soundings, the disruption of the time-dependent relationships took place earlier, chiefly in the sectors of the development of polygonal-wedge ice, i.e., where the vertical heterogeneities occurring near the surface could lead to a shielding effect; an effort was made based on them to explain the fact of the acceleration in the drop of the IP and the rare cases of the occurrence of an IP field of negative sign.

With the aid of the following experiment, we were able to establish the fact of the existence of surface polarization at the boundary of permafrost and of the seasonally thawing layer. Utilizing the standard technique of vertical electric logging of the induced potential and the VPS-63 equipment, we recorded the total attenuation of the induced polarization field after a two-minute passage of direct current into the earth for each spacing of the supply line *AB*. Each subsequent sounding at the identical *AB* was conducted for a new position of the measuring line which by way of a layer-by-layer removal of the ground under it was lowered more and more until the permafrost surface was reached.* As is evident from Figure 1, the decay curve of induced polarization was transformed from the normal during the measurements at the surface and near the surface of the earth to the opposite in sign but close to the normal in nature of variation through time (for the recent periods) during measurements near the surface of permafrost and directly on this surface.

We were able to determine the boundary of the thawed and frozen soils by the degree of sharp-

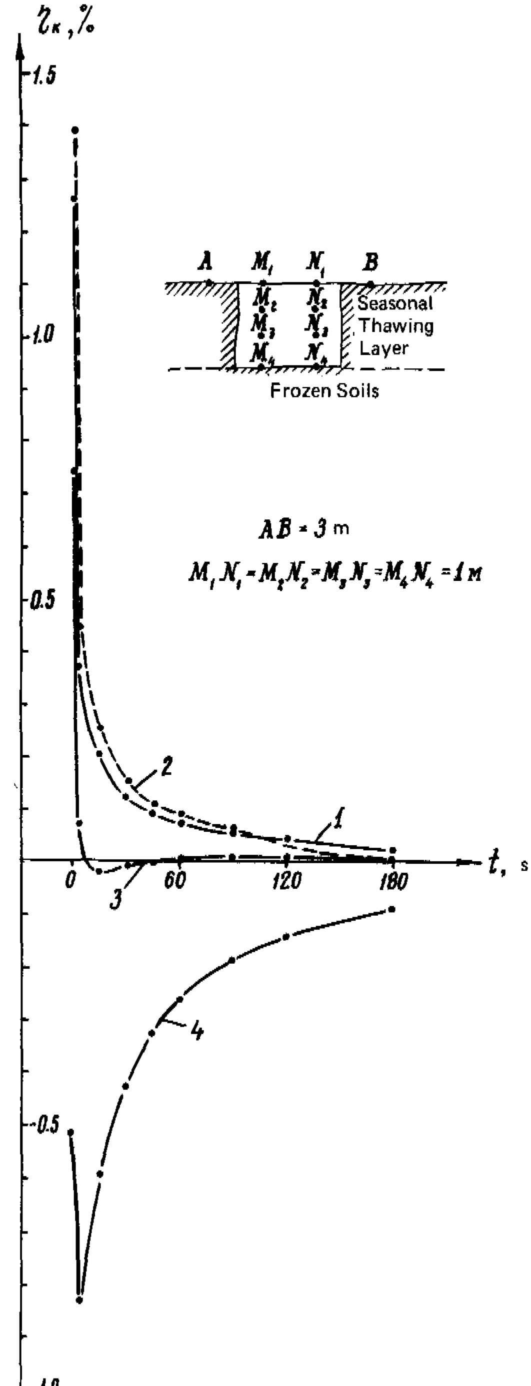

FIGURE 1 Nature of secondary field of induced polarization (IP) at varying distance of measuring electrodes from permafrost boundary. 1--*MN* on surface of Earth; 2--*MN* at depth of 0.5 m; 3--*MN* at depth of 1 m; 4--*MN* at permafrost surface.

*The thickness of the seasonally thawing layer at the given point was 1 m to 65 cm. By means of a borehole, we revealed, to a depth of 65 m, Quaternary sands with thin interbeddings of supes. The boundary between the layer of seasonal thawing and permafrost runs through sand.

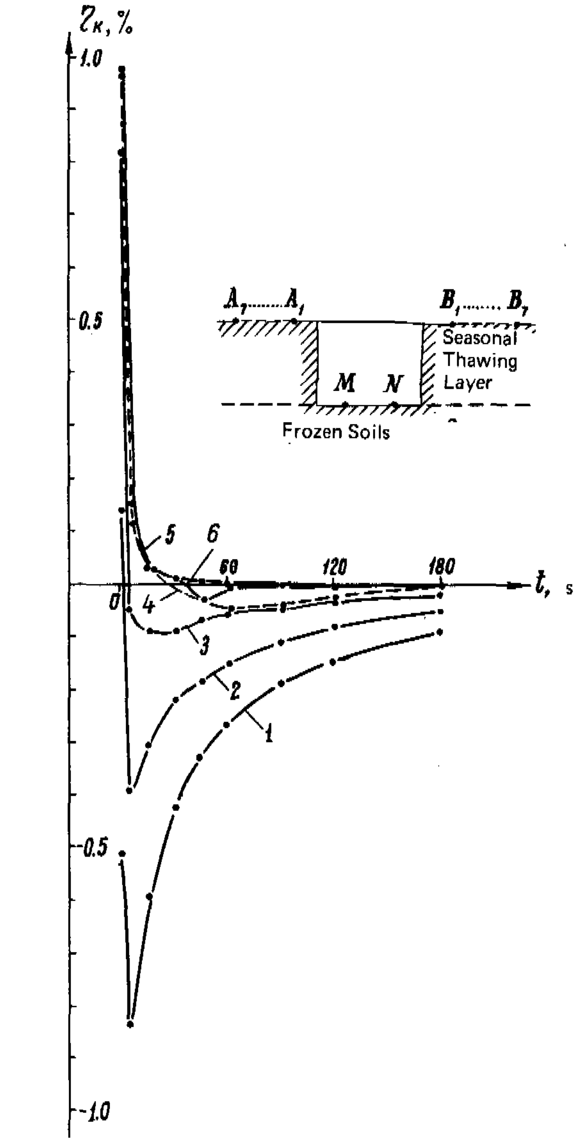

FIGURE 2 Nature of secondary induced polarization field at various spacings of supply line *AB*. 1--*AB* = 3 m; 2--*AB* = 6 m; 3--*AB* = 18 m; 4--*AB* = 30 m; 5--*AB* = 50 m; 6--*AB* = 80 m.

ness in the ρ_K curve, having a slope of 45° and testifying to the great difference in the specific resistances of soils in the layer of seasonal thawing and permafrost.

At an increase in the spacing of the supply line and, accordingly, with a decrease in the density of current near the permafrost boundary and with an increase in the volume of volumetrically polarizing soils, the emf value of surface polarization decreases (Figure 2) and the decay curve of induced polarization will tend toward the normally determined volume polarization of frozen soils. Similar results were obtained during measurements at the wall of a borehole with a dipole installation at the approach to the permafrost surface (Figure 3).

In this way the observed induced polarization fields in the regions where permafrost occurs are determined by the interaction of two fields induced by the passage of a polarizing current, i.e., by the field of volume polarization of masses of thawed and frozen rocks and by the field of surface polarization at the boundary of the thawed and frozen soils. Owing to the fact that these fields are directed oppositely while their intensity is determined by various causes, the falling off of the overall induced polariza-

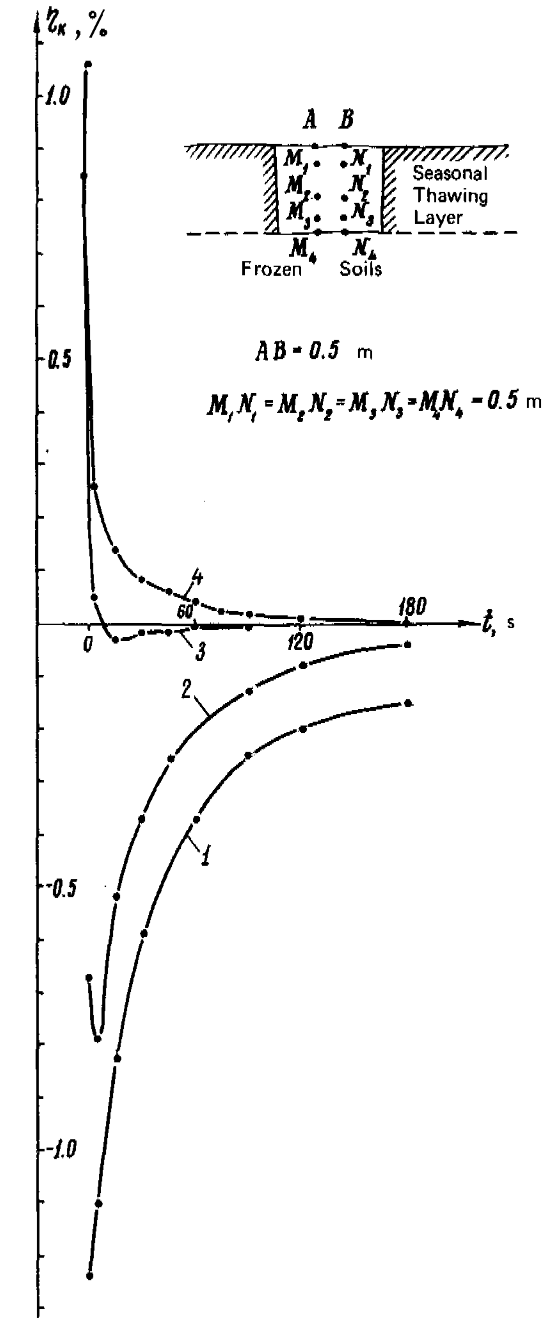

FIGURE 3 Nature of secondary induced polarization field at measurements with dipole installation on wall of borehole. 1--*MN* on permafrost surface; 2--*MN* at 0.2 m from permafrost surface; 3--*MN* at 0.6 m from permafrost; 4--*MN* at distance of 1.25 m from permafrost.

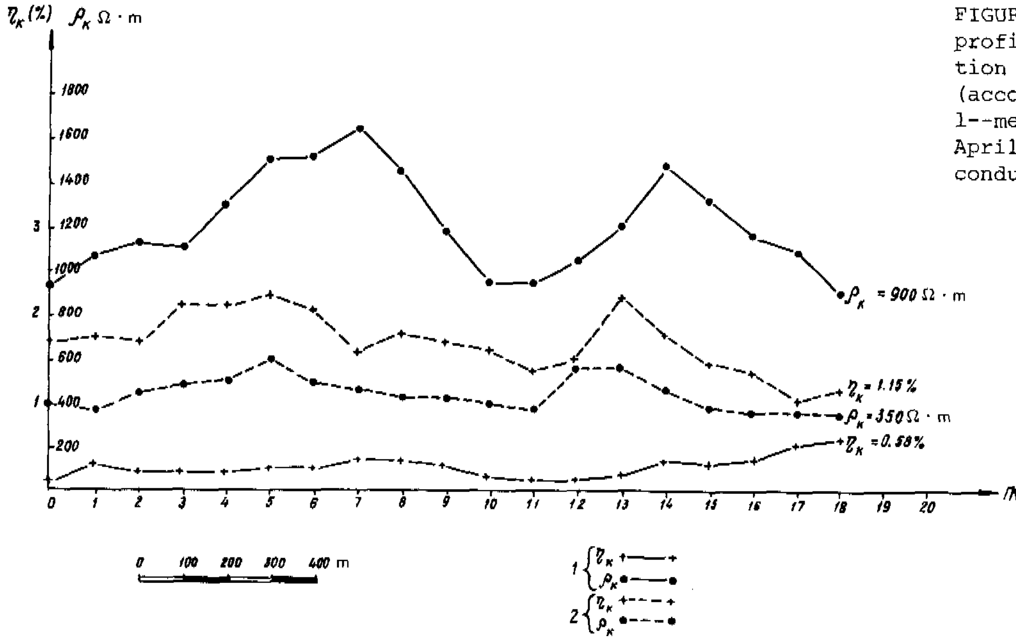

FIGURE 4 Example of symmetric profiling of induced polarization at different seasons (according to A. M. Snegirev). 1--measurements conducted in April 1971; 2--measurements conducted in August 1971.

tion field shows a complex transition process not corresponding to any one of the approximate expressions previously suggested for describing the tentative features of induced polarization.[2]

The establishment of surface polarization at contacts between the frozen and thawed soils provides an explanation to certain earlier unclear facts regarding the variation in polarizability, depending on the installations utilized, measurement techniques, and the season of year in which the observation was conducted. An example of such a dependence can be provided by Figure 4, in which we illustrate the results from observations of induced polarizations in spring and autumn, i.e., at varying thickness of seasonal thawing layer and consequently at varying distance of source of surface polarization from the measuring dipole. In spring, when the permafrost boundary is located near the Earth's surface, the source (permafrost boundary) of negative polarization exerts a more significant influence on the overall field, and the polarizability η_K proved to be much less than during the measurements at the same point in August when we are dealing with greater depth of thaw, i.e., with a more significant separation of the source of surface polarization. In this manner the phenomenon of surface polarization opens a route to a solution of a number of problems in geocryology associated with the disclosure and study of the migration of cryogenous boundaries and compels us to make a new approach to the fundamentals of the induced polarization method in a solution, with its aid, of the traditional problems in the regions where permafrost is developing.

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PRINCIPLES INVOLVED IN THE MAPPING OF PERMAFROST ACCORDING TO ITS CRYOGENOUS STRUCTURE

B. I. VTYURIN Pacific Ocean Institute, Far Eastern Science Center, USSR Academy of Sciences

The special structure of soils determined by the presence and arrangement of underground ice in it is said to be *cryogenous*. A knowledge of the latter is a necessary condition for a proper engineering-geologic evaluation of permafrost and any paleogeocryologic and consequently paleogeographic construction. From this there develops the urgency of the question concerning the techniques and methods for a cartographic representation of the relationships in the distribution of permafrost with varying cryogenous structure.

Until recently efforts had been made to depict on the general survey geocryologic maps the distribution of permafrost with varying types of cryogenous structure. Since at the present time most of the permafrostologists have arrived at a conclusion concerning the feasibility of the division of permafrost into only two genetic types, namely epigenetic and syngenetic, on the maps we have differentiated permafrost layers of the appropriate types. Following I. Ya. Baranov, certain researchers have distinguished a third type of layer, namely polygenetic, characterized by alternation in the section of permafrost of epigenetic and syngenetic structures. However, as we shall indicate below, this is already a different principle of classifying the perennially frozen soils by cryogenous *structure*--i.e., by the stratification of horizons of permafrost with varying types of cryogenous structure.

In 1971 an initial attempt was made to compile a special chart showing the distribution of permafrost of the most frequently occurring types, subtypes, and forms of cryogenous structure in the USSR (Figure 1). As the basis we classified permafrost according to cryogenous structure, which, in simplified version, can be represented as follows: Since the texture-forming ice comprises the basic mass of underground ice and plays a decisive role in the cryogenous structure, it determines the subtype of cryogenous structure of the epigenetic and syngenetic types--massive, when in the section there predominate soils with massive cryogenous textures, and streaky, when there prevail soils with streaky cryogenous textures (segregated lenses in the granular cohesive soils). The deposit-formed ice (multiple wedge ice, segregated and injected layers, primary-surface buried) only complicate the cryogenous structure. The presence or absence of them in the section determines the complex or simple form, respectively, of the subtype's cryogenous structure.

Since the most common are the multiple-wedge and sheet deposits of primary interbedded ice, they alone were taken into account in compiling the chart. Moreover, their role in the cryogenous structure is not uniform. Therefore, for practical purposes, we adopted the conventional values below which the deposit-forming ice was not taken into consideration. For the multiple wedge ice, we assumed the polygonal state of the region as 20 percent. Among the regions for the occurrence of syngenetic permafrost of streaky subtype of complex form with multiple wedge ice we included: the north of western Siberia, northern Siberia, Lena-Vilyuy, Yana-Kolymov, Anadyr, and other lowlands where 20 to 50 percent and more of the permafrost area is occupied by a system of multiple wedge ice. It proved to be more complicated to establish the limits of the sheet deposits of epigenetic layers of the complex type of permafrost. So far we have not developed either the techniques of a small-scale mapping or the methods of determining the macro-ice content; therefore, the sectors of the occurrence of epigenetic permafrost of the streaky subtype of complex form, i.e., sheet deposits in the south of the Kola Peninsula and western Siberia, in Trans-Baikal and in the Amur region, do not have distinct outlines, so they have been differentiated very tentatively.

Also segregated tentatively are the sectors, from the permafrost aspect, of both types of massive subtypes of simple forms. For the epigenetic type these are sectors of the intensive ablation of material on the denudation slopes in the mountains. It is probable that such sectors occur quite rarely, since, even under the conditions of intensive ablation of materials, the frost-cracking of soils and the formation of wedge ice precede this and cause the formation of the streaky subtype. For the syngenetic type, these are sectors of intensive accumulation of deposits (chiefly peats) in the zone of rare (insular) occurrence of fine-grained frozen soils with temperatures above -0.5°C.

Thus it is revealed that the maximal distribution is of permafrost of streaky subtypes, of simple and complex forms; this fact is extremely important in an engineering-geologic evaluation of a permafrost region.

In 1971, for the first time we made an attempt at a cartographic representation of the relationships in the distribution of permafrost layers with varying structure, i.e., with interbedding. For an engineering-geologic evaluation of permafrost and for the paleogeographic reconstructions, it is important to know not only the features in the cryogenous structure of

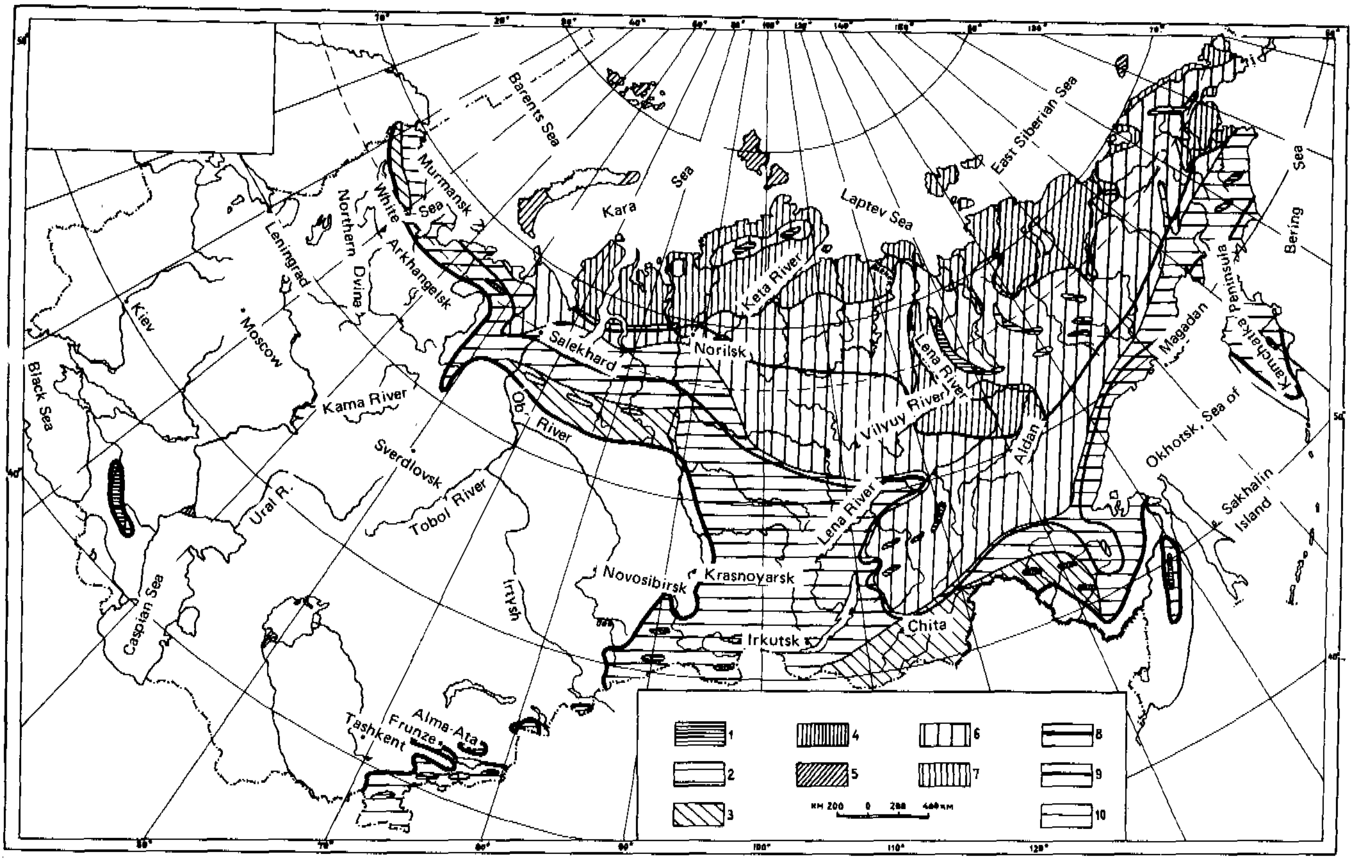

FIGURE 1 Schematic map showing distribution of permafrost with most-frequently occurring forms of cryogenous structure in the USSR. Region of soils of *epigenetic type*: 1--massive subtype of simple form; 2--streaky subtype of simple form; 3--complex form--with thin shallow layers of interbedded ice. Region of soils of *syngenetic type*: 4--massive subtype of simple form; 5--complex form with multiple wedge ice; 6--streaky subtype of simple form; 7--complex form with multiwedge ice; 8--boundary distribution of permafrost; 9--of contemporary lithocryologic zones; and 10--of areas.

soils from the surface, but also the sequence of stratification of horizons with their varying structure. For a cartographic representation of perennially frozen rocks with varying structure, so far relatively few data have been accumulated. However, the data are adequate to attempt to compile a schematic small-scale map of the permafrost soils with varying structure in the USSR (Figure 2). As the basis of the mapping, we developed the following classification: All the permafrost layers are divided into monogenetic and polygenetic types. Accordingly, the first ones can be either epigenetic or syngenetic, while the second can be two-four, etc., horizontal types.

It was explained that in the USSR there occur fairly widely monogenetic-epigenetic permafrost layers, and the syngenetic layers are only postulated. The latter are possible in the sectors with syngenetic permafrost of massive subtype having a simple form, i.e., in the south of western Siberia and the Amur region (see Figure 1). Among the polygenetic layers, most widely distributed are the two-horizon types, i.e., the syngenetic-epigenetic ones. Since the thickness of the upper syngenetic layer, depending on actual conditions, varies within wide limits, on the map it is feasible to distinguish sectors with their varying thickness. Thus, for the USSR territory we suggest the separation of sectors with thickness of syngenetic layer up to 3 m, 3 m to 10 m, 10 m to 40 m, and greater than 40 m. Thickness of the epigenetic layer is quite irregular and increases from south to north from several meters to hundreds of meters. In this manner, subsequently the researchers have the possibility of extending the system of mapping based on this principle by taking the thickness of the epigenetic layer into consideration.

The four-and-more layer polygenetic deposits of permafrost require unique conditions for their formation and occur only in isolated sectors in the region of polygenetic permafrost. At the present time there are descriptions of the four-layer deposits in western and central Siberia. We can theorize that in the regions of marine Quaternary deposits with unstable tectonic conditions where, in addition, the process of formation was complicated by the advance and retreat of glaciers, the multilayer deposits are not a rarity. Among such sectors, we have included certain regions of the northern Siberian lowland and the coastal sectors in the Chukotka lowlands (see Figure 2).

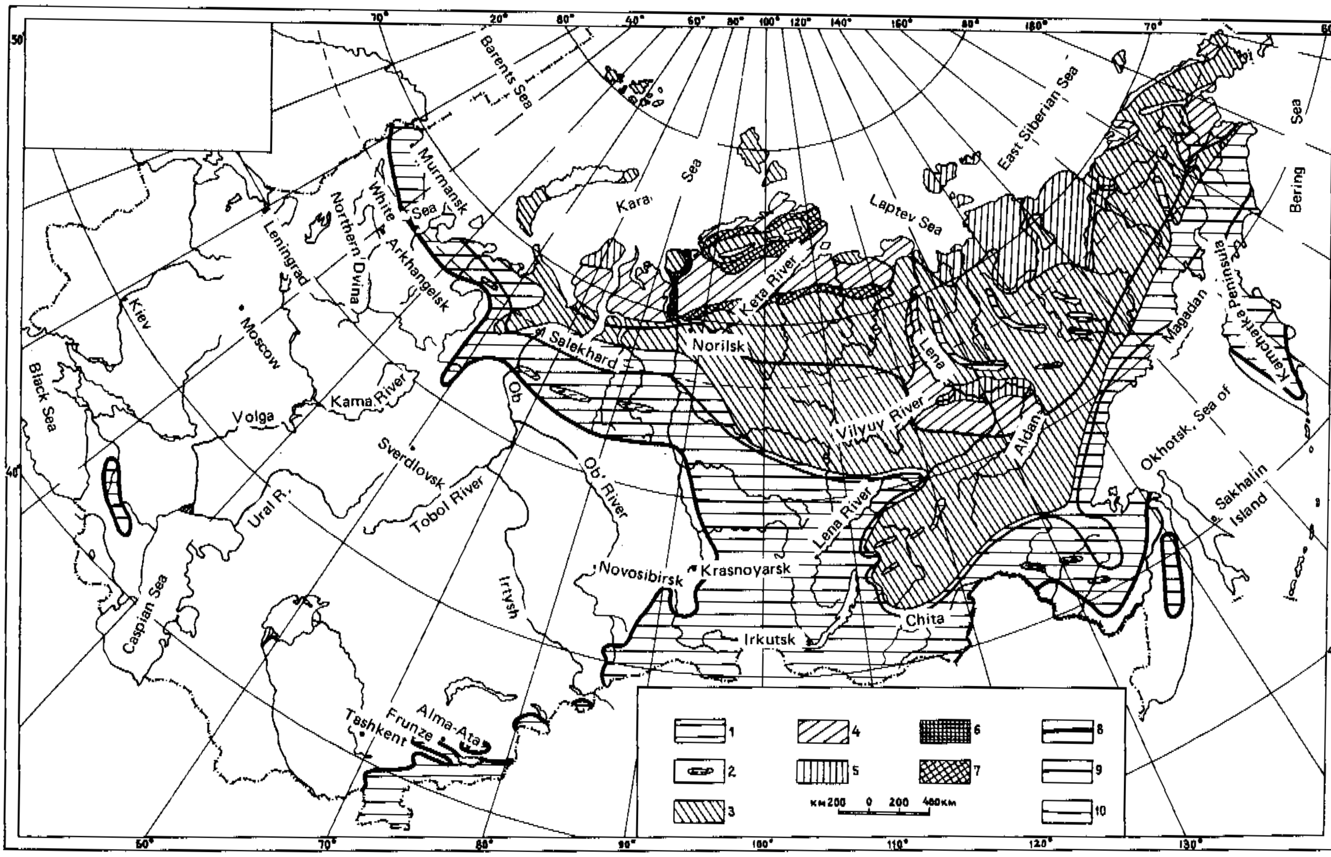

FIGURE 2 Schematic map showing distribution of permafrost with varying macrostructure in USSR. 1--monogenetic epigenetic type; 2--monogenetic syngenetic type (identified tentatively); 3--polygenetic two-horizon with a thin syngenetic layer toward the top (to 3 m); 4--the same, with thick syngenetic layer toward the top (3-10 m and more); 5--the same with very thick syngenetic layer (10-40 m and more); 6--polygenetic multilayer, established by direct observations; 7--the same, tentative; 8--boundary of frozen soils; 9--boundaries of contemporary lithocryological zones; and 10--boundaries of areas (geographic ranges).

POSSIBILITIES OF SEISMIC EXPLORATION BY REFLECTED WAVES FOR MAPPING AND ZONING OF FROZEN SOILS

YU. K. AGAFONOV, V. A. GERSHANIK, A. D. MEDVEDEV, V. K. MONASTYREV, AND L. G. TSIBULIN Glavtyumen'-geologiya. West Siberian Scientific-Research Gas-Petroleum Institute

The possibilities of the seismic technique of studying permafrost and the evaluation of certain parameters of it are embodied in the physical nature of the methods and are based on the manifestations of specific conditions with which it is necessary to deal during the exploratory operations for petroleum and gas in the regions where permafrost occurs. These conditions in the petroleum-gas regions in the north of the western Siberian lowlands are manifested primarily in the level of geologic effectiveness of seismic operations. The lack of information concerning the areal distribution of permafrost and the spatial distribution of their parameters leads to a reduced (from an information standpoint) quality of seismic observations and to significant errors in the structural deductions.

Theory permits us to indicate the following formula for the propagation velocity of a longitudinal seismic wave in terms of the elastic constants of the medium:

$$v = \sqrt{\frac{E}{\rho} \cdot \frac{1 - \sigma}{(1 + \sigma)(1 + 2\sigma)}}, \qquad (1)$$

where E = Young's modulus, σ = Poisson's ratio, and ρ = density.

In a general case, the v value depends on moisture content (iciness) and on other physical properties of rocks. It is obvious that, if in a geologic section of various sectors, the frozen soil layers are differentiated by values of elastic constants and thicknesses and, in addition, there is a difference in the number of layers, this must inevitably lead to a corresponding variation in the kinematic and dynamic parameters of the reflected waves penetrating the permafrost.

Let us consider the capabilities of seismic exploration using the reflected wave method (RWM) for a quantitative estimation of the parameters for the frozen soils within the framework of the kinematic aspects of the phenomenon.

In a multilayered medium with horizontal boundaries consisting of n layers being characterized by thicknesses h and the layer velocities v_i, the equation for the hodograph of a reflected wave has the following form:

$$t = 2\sum_{i=1}^{n} \frac{h_i}{v_i} + \frac{x^2}{4}\left(\sum_{i=1}^{n} h_i v_i\right)^{-1} - \frac{x^4}{64}\sum_{i=1}^{n}\left(h_i v_i^3\right)\left(\sum_{i=1}^{n} h_i v_i\right)^{-4} + \ldots \qquad (2)$$

It is obvious that (2) is also valid for the case when the layer of permafrost soils is horizontally stratified with parameters h_{mj} and v_{mj} such that $v_{mj} \neq v_j$ and $h_{mj} \leq h_j$ $(j = 1,2,\ldots, k; k \leq n)$.

In this case, restricting ourselves to the two first terms in series (2), the equation of the hodograph will assume the form

$$t^k = 2\left[\sum_{i=1}^{n} \frac{h_i}{v_i} + \sum_{i=1}^{n} h_{mj}\left(\frac{1}{v_{mj}} - \frac{1}{v_j}\right)\right] + \frac{x^2}{4}\left[\sum_{i=1}^{n} h_i v_i + \sum_{j=1}^{n} h_{mj} \times \left(v_{mj} - v_j\right)\right]^{-1}. \qquad (2_a)$$

The equation for the hodograph t^l for the sector L, differing from the previous one by parameters h_{ms} and v_{ms} $(s = 1,2,\ldots, l\ ;\ l \neq k)$ will have a form similar to (2^9).

For estimating the effect of a variation in the parameters in the layer of permafrost soils on the hodograph parameters, let us consider the difference in hodographs $\Delta t = t^k - t^l$.

After simple algebraic operations and simplifications, we get:

$$\Delta t = 2\left[\sum_{j=1}^{k} h_{mj}\left(\frac{1}{v_{mj}} - \frac{1}{v_j}\right) - \sum_{s=1}^{l} h_{ms}\left(\frac{1}{v_{ms}} - \frac{1}{v_s}\right)\right] - \frac{x^2}{4}\left(\sum_{i=1}^{k} h_i v_i\right)^{-2} \times \left[\sum_{j=1}^{k} h_{mj}\left(v_{mj} - v_j\right) - \sum_{s=1}^{l} h_{ms}\left(v_{ms} - v_s\right)\right]. \qquad (3)$$

The introduction of certain limiting conditions in accordance with the actual seismic-geologic situation permits us to reduce (3) to a form convenient for analysis and practical use. Let us assume

$$v_j = v_s = v_l \text{ and } v_{mj} = v_{ms} = v_m. \qquad (4)$$

The nature of the limitations (4) includes the assumption that the bed of frozen soils will be contained in the first layer (the enclosing rocks with parameters h_1, v_1) and is typified by constant velocity v_m.

Proceeding from (3) and (4), we derive

$$\Delta t = \Delta h_m\left[2\left(\frac{1}{v_m} - \frac{1}{v_i}\right) - \frac{x^2}{4}\left(\sum_{i=1}^{n} h_i v_i\right)^{-2}\left(v_m - v_l\right)\right], \qquad (5)$$

where $\Delta h_m = \sum_{i=1}^{k} h_{mj} - \sum_{s=1}^{l} \cdot h_{ms}$ = difference in overall thicknesses of frozen layers of the sectors that are being compared.

In this manner the Δt values that are being determined on the basis of the hodographs of reflected waves permit us to estimate only the relative variation in overall thickness of frozen soils.

We can derive directly from (5)

$$\Delta h_m = -\frac{\Delta t}{\left[2\left(v_i - v_m\right)^{-1} + \frac{x^2}{4}\left(\sum_{i=1}^{n} h_i v_i\right)^{-2}\right]\left(v_m - v_i\right)}. \qquad (6)$$

Since for the sandy-clayey beds $v_m > v_1$, from (6) it then follows that (other conditions being equal) Δh_m is directly proportional to the [absolute] value $|\Delta t|$ and is opposite to it in sign.

Equation (6) was utilized for estimating Δh_m under the conditions existing in the north of Tyumenskaya Oblast based on the data from areal seismic exploration operations and seismic logging of deep boreholes.[1,2] The results from this permit us to consider the possibility of utilizing the seismic data for estimating the Δh_m values of permafrost as completely realistic.

In one of the petroleum-gas-producing areas, favorable factors have appeared for estimating the effectiveness of utilizing Equation (6).

Such factors proved to include the thick network of reflected wave method observations and the relatively large number of exploration boreholes, particularly those in which an estimation was made of the overall thickness of permafrost based on the data from industrial-geophysical techniques and also the fact of a significant divergence in the structural inferences based on data obtained from drilling and seismic exploration. It was established that these differences are caused by ignoring the fact of the variability in the thickness of frozen soils during the quantitative processing of the RWM (reflected wave method) observations.

Under these conditions, the determination of the Δt values based on the known distortions in structural formulations did not present any difficulty. The estimation of Δh was conducted based on the equation

$$\Delta h_m = \frac{\Delta t}{2\left(\frac{1}{v_m} - \frac{1}{v_1}\right)}, \qquad (6a)$$

which follows from (6) assuming in it that $x = 0$. This simplification is associated with the fact that, in ascertaining the depths of reference reflecting horizons, we did not utilize all the periods of the hodograph but only one, i.e., $t_0 = t_{x=0}$.

The Δh_m values computed according to (6a) were compared with the $\Delta h^1{}_m$ values based on borehole observations. In this context, it turned out that $\delta h_m = \Delta h_m - \Delta h^1{}_m$ in 11 cases out of 13 does not exceed $\pm$ 30 m at variations in overall thickness reaching 300 m and more.

As a rule the upper part of the section containing the permafrost, for a number of reasons, was not studied by the seismic-logging operations. However, even in this case, it appears possible to utilize the data from seismic logging observations for evaluating the parameters of the permafrost soils.

In the models (2), (2a), and (4), the vertical hodograph has the form

$$t = \sum_{i=1}^{n} \frac{h_i}{v_i} + h_m\left(\frac{1}{v_m} - \frac{1}{v_1}\right). \qquad (7)$$

It follows from (7) that in a medium with varying $h_m = \sum_{i=1}^{h} h_{mj}$, the curves reflecting the vertical hodographs in the coordinates t and H represent a combination of the curves of parameter h_m, in which each of the curves is shifted relative to the adjacent one parallel to itself along the axis t by the amount

$$\Delta t = \Delta h_m \left(\frac{1}{v_m} - \frac{1}{v_1}\right). \qquad (7a)$$

In this manner the displacement of the vertical hodographs constitutes not only an indicator of the variation in thickness of permafrost but also permits us to estimate quantitatively the relative increment in overall thickness Δh_m.

Experience has demonstrated that for practical calculations of Δh_m, under the conditions of the region under discussion, we can recommend the following estimations of the velocity parameters of the section's upper part: v_1 = 1.65 km/s and v_m = 3.5 km/s.

Even more extensive capabilities of seismic exploration in the study of frozen soils are revealed in a consideration of a possibly more complete number of criteria in the diversity of features of seismic recordings of the kinematic and dynamic nature. In this diversity, the most significant role belongs to the data concerning the areal zoning of the propagation velocities of direct waves, of medium or effective velocities, variation in amplitude, frequency and energy characteristics of seismic fluctuations, etc. The actual diversity in the criteria has been caused by the variability in the parameters for the permafrost and in a certain respect it reflects its structure.

The appreciable effect of surface conditions on seismic recordings, where the permafrost factor is most important, permits us (depending on the degree of study of the area by the combination of drilling techniques, seismics, and aerial photography) to obtain a concept concerning the areal distribution of frozen soils.

The variation in frozen soils in plan based on their manifestations directly at the surface is clearly identified in the recordings of direct waves. To the sectors formed of frozen soils all the way from the surface, there correspond the velocities of about 3.5 km/s; where permafrost is absent, the velocities are 1.7 km/s, while in the sectors having a mosaic structure, the velocity value is intermediate and is all the higher, the larger the proportion of frozen soils (Figure 1). The abrupt natural boundaries between the frozen and unfrozen soils are distinctly recorded by breaks in the co-phasal axes. The sectors differeing from a geocryologic standpoint are also reflected in the dynamics of the wave recording: For the sectors not completely frozen from the surface, we typically find a relatively low-frequency recording, while for the sectors of mosaic structure, we note a high level for the background of interference and the frequent absence of the reflected-wave recordings.

The grouping of criteria of the quantitative and qualitative characteristics of seismic recordings in connection with the variation in permafrost parameters permits us not only to improve significantly the reliability of geocryologic interpretation of the aerial photographs, but also, as the investigations at the Glavtyumen'geologiya (Main Administration for Geology at Tyumen) and the ZapSibNIGINI to realize in practice the geocryologic zoning of an area in such a way that it would reflect the zoning by degree of favorableness of the surface seismic-geologic conditions.

The basis for such a zoning will be provided by the data from seismic exploration and the terrain interpretation of the aerial photographs. The main system for the data flow (Figure 2) is

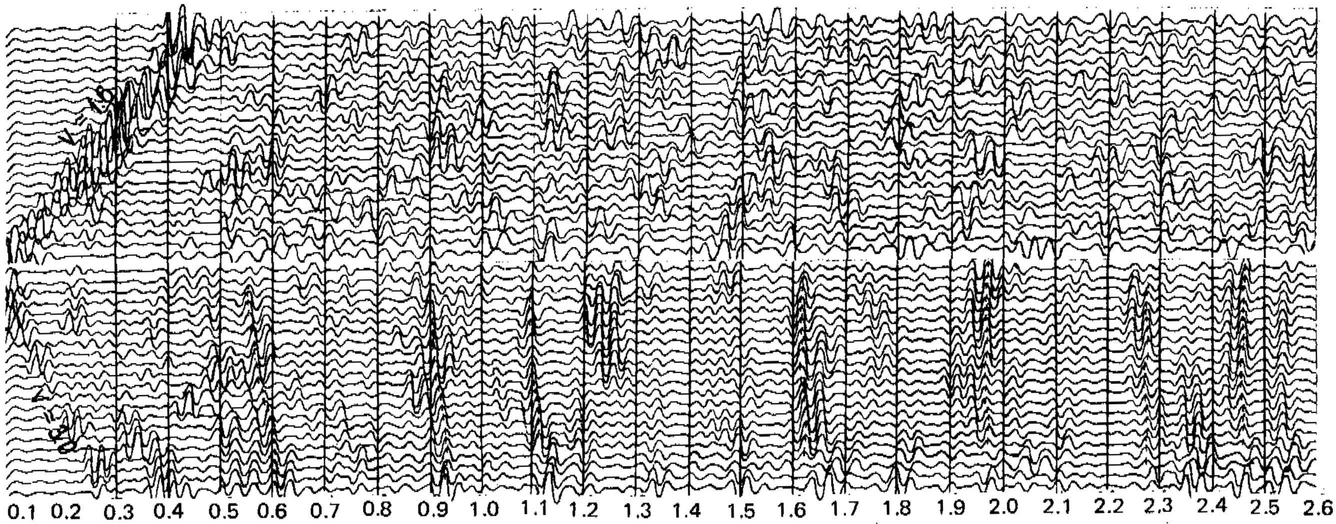

FIGURE 1 Photograph of seismogram obtained from excitation source located at boundary of two sectors with varying permafrost characteristics of surfaces. Upper part--warmed sector (velocity of direct wave, $v_{dir} \simeq 1.6$ km/s) with grassy-mossy cover and open woodland over heaved hummocky formations; the lower part is a water divide surface (chiefly frozen) ($v_{dir} = 3.0$ km/s) with small thermokarst sinkholes.

taken into account in the classification, diagnosis, and prediction of the properties of actual sectors of a locality according to degree of favorableness of seismic geologic conditions at the surface. On this basis we have developed a technique permitting us to accomplish the selection of the optimum allocation of seismic profiles based on data from aerial photographic interpretation.

It has been established that there are typified by preferentially favorable seismic-geologic conditions the sectors of a locality having a light

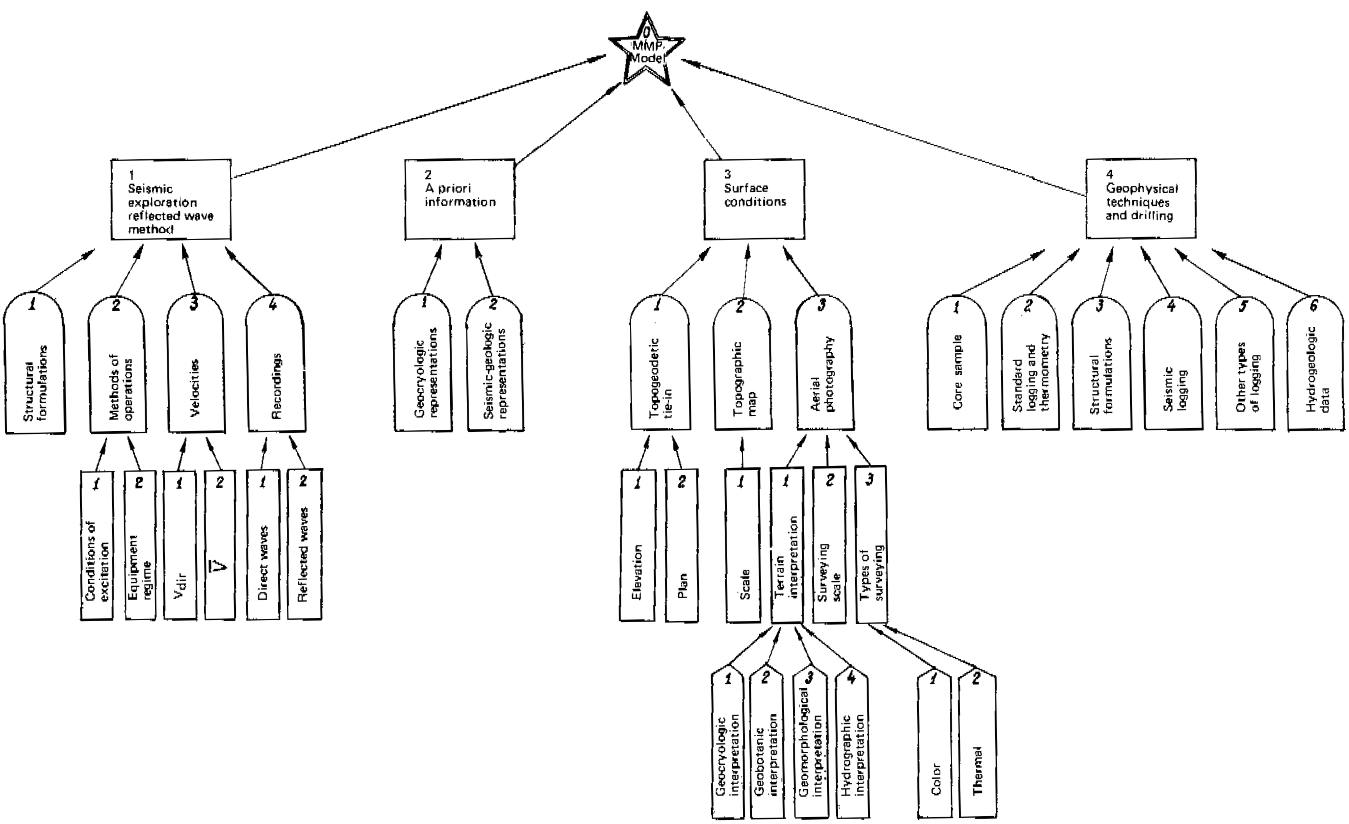

FIGURE 2 Main diagram showing data flow, used for geocryologic areal zoning.

photographic tone with development of thermokarst saucer-shaped depressions, orthogonality of erosion networks, string bogs, etc. Unfavorable seismic-geologic conditions are typical for the sectors with a dark shade of photographic image, by the development of fine-hummocky and polygonal relief forms, thickly branched deeply indented erosion networks, etc.

In this manner, what has been discussed permits us to consider as completely realistic the possibilities of seismic exploration for evaluating certain parameters with reference to the permafrost both on the vertical and horizontal. The combined utilization of the data from seismics, aerial photography, and drilling permits us to solve geocryologic problems with prospects for their practical application to the tasks involved in enhancing the geologic effectiveness of seismic exploration for petroleum and gas.

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MAPPING OF PERMAFROST WITH THE AID OF A DIPOLE ELECTROMAGNETIC PROFILING TECHNIQUE

A. G. KRASOVSKIY Industrial and Scientific-Research Institute for Engineering Surveys in Construction

The abrupt growth in the rate of development in the Far North requires the introduction of accelerated techniques in the practice of engineering-geologic investigations in the regions where permafrost occurs. One of such methods is dipole electromagnetic profiling based on the recording, along the line of observations, of the variations in the elements of an electromagnetic field excited by a special mobile oscillator. The dipole electromagnetic profiling method serves for revealing and mapping the rapidly falling heterogeneities in the upper part of the geoelectric section for a depth up to 5-20 m. As compared with electric profiling using direct current, this technique has a number of advantages: high-resistance layers (for example, the ice layers) not comprising shields, there is no requirement for grounding. This permits us to conduct, without any particular difficulties, investigations from the ice or during a thick snow cover. In addition, the dipole electromagnetic profiling method has a higher output, mobility, and economy as compared with direct-current electric profiling.

In operations using the dipole electromagnetic profiling method in permafrost areas, a number of specific features develop, caused by the high specific-resistances of frozen soils. From theory it follows that in the range of high resistances, there is a decrease in the resolving capacity of the method; moreover, this is all the more appreciable, the lower the frequency of the electromagnetic field that is being used. In the utilization of high-frequency fields in the high-resistant sections, it is necessary for the theoretical calculations on which the interpretation of the dipole electromagnetic profiling results is based, to take into consideration not only the conductivity currents, as this has been done in most of the existing calculations,[4] but also the displacement currents, with allowance for the frequency dispersion of the soils' electric properties. At the present time, owing to the inadequacy of experimental data, it is impossible to determine by calculation the effect of these factors on the results from dipole electromagnetic profiling.

For clarifying the potentialities of the dipole electromagnetic profiling during the mapping of permafrost and the development of a technique for its application during the engineering-geologic investigations, in the ISRIESC for a number of years (beginning with 1968), we conducted field experiments in various regions in the northern part of Siberia with the series-produced DEMP-3 installation operating on frequencies of 32, 128, 512, and 2048 kHz.

An analysis of the results obtained permits us to consider feasible the measurements of the vertical H_z and horizontal H_r components of magnetic field strength excited by an oscillator with an antenna in the form of a vertical magnetic dipole. The frequency of the electromagnetic field and the distance between the oscillator and receiver are chosen according to the properties of the geoelectric section that is under investigation. The measurements are conducted by individual profiles or based on a network of profiles. The measurement technique corresponds to that generally accepted for the dipole electromagnetic profiling method.[1,4] The results of measurements along the profile can be represented in the form of curves for the ratios H_r/H_z or can be converted to the effective resistance ρ_{eff} of the section, determining the latter based on the curve of H_r/H_z and frequency, spacing and specific resistance of section accepted for a homogeneous half-space.[1,4] During the area operations, the dipole electromagnetic profiling results can be expressed in the form of correlation systems of profiles or in the form of maps reflecting the ρ_{eff} values.

In the studies of the permafrost, a significant effect on the results of dipole electromagnetic profiling is exerted by the variations occurring in the layer of seasonal thawing. This

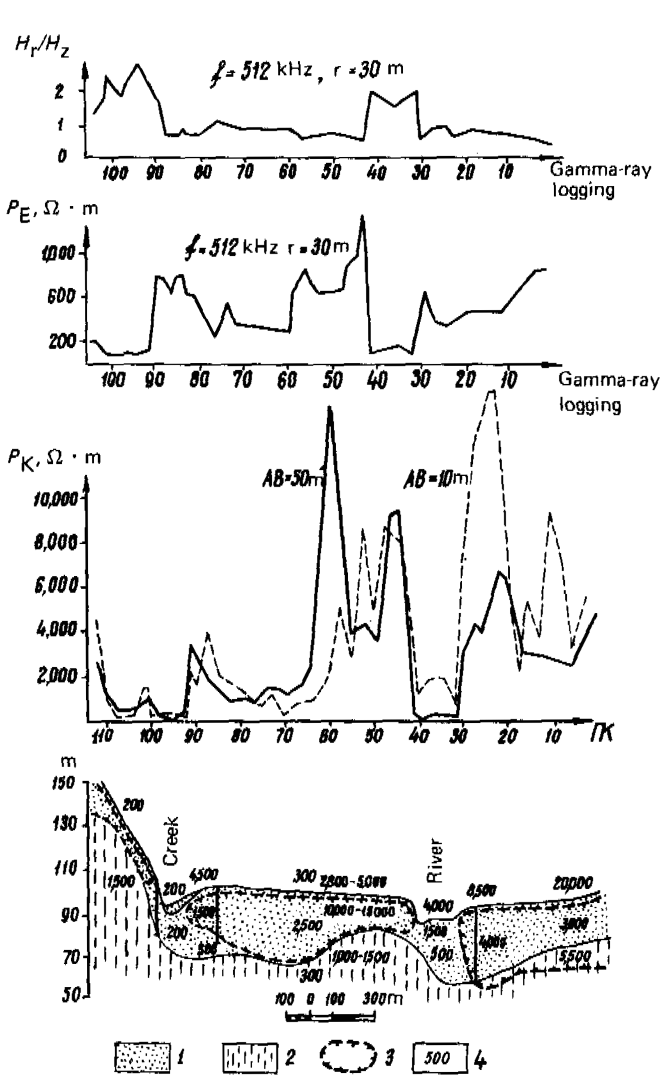

FIGURE 1 Comparison of results from DEMP and electric profiling using direct current, based on one of the profiles. 1--granular sediments; 2--original deposits; 3--boundaries of permafrost soils; and 4--values of specific resistances, ohms•m.

question requires special attention, since the penetration depth of the dipole electromagnetic profiling is all the less, the lower the resistance of the upper layer, the greater its thickness and the higher the resistance of the subjacent layers. Therefore, a fairly thick thawing layer with quite low resistances can shield the heterogeneities within a frozen layer. Nevertheless, calculations testify that, in most cases, the penetration depth of the dipole electromagnetic profiling exceeds the maximum depth of seasonal thawing, while numerous repeated observations at the beginning and end of the season in various regions show that, as a rule, the variation in the thawing depth has practically no effect on the form of the curves in the dipole electromagnetic profiling, although the ρ_{eff} value does change.[3]

For dividing the anomalies into those near the surface and those caused by dissimilarities of the deeper horizons, it is useful to conduct dipole electromagnetic profiling with different installations, altering the frequency and spacing. In this case, those anomalies that are manifested more distinctly at a higher frequency and a shorter spacing correspond to the variations in the upper layer.

As an example, we present the curves of dipole electromagnetic profiling and of symmetric electric profiling for one of the ranges, the geoelectric section of which was studied in detail by electric prospecting using direct current and drilling (see Figure 1). Here the granular deposits are represented by gravels, sandy loams, and clayey loams, while the original deposits are represented by effusive rocks. Based on the curve of dipole electromagnetic profiling at a frequency of 512 kHz with spacing between the oscillator and receiver of 30 m, in the floodplain of a river and in the channel of a creek, the taliks are clearly differentiated by the low ρ_{eff} values. In the frozen layer, by the highest ρ_{eff} values among them, there are characterized the ice-rich coarse-grained deposits. As a whole, the ρ_{eff} curve agrees well in form with the curve of electric profiling and with the geoelectrical sections. However, the ρ_{eff} values are much lower than the ρ_k values and the actual specific resistances of frozen soils. This is explained by the fact that the conversion of the dipole electromagnetic profiling results to the ρ_{eff} values was done without considering the displacement currents. Such a nonconformity prevents us from determining, based on the dipole electromagnetic profiling curves, the true values of specific resistances, but does not present an obstacle in the geocryologic interpretation of the dipole electromagnetic profiling data.

The application of the dipole electromagnetic profiling technique for mapping the permafrost was tested in regions with varying types of geocryological conditions. In the regions with deep continuous permafrost, based on the dipole electromagnetic profiling results, we succeeded in segregating the hidden taliks and the zones of elevated resistances caused by an increase in the ice content or in the amount of granular material in the soil. Wherever we encounter the frozen and thawed soils, they are differentiated clearly by the dipole electromagnetic profiling curves. In the zones where the discontinuous permafrost occurs, based on the dipole electromagnetic profiling curves, we clearly identify the sectors of shallow (up to 10 m) and thick (above 10 m) permafrost in the sandy-clayey sections.

In favorable cases when the ice-rich frozen soils occur beneath the thawed loams or clays with a known low specific resistance, based on the DEMP data we can determine the thickness of the icy horizon.[2] (See Figure 2.)

The experience gained in the studies performed reveals that the DEMP technique can be applied with success in a combination of geophysical operations during the mapping of perennially frozen soils.

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DIFFERENTIATION OF FROZEN SOILS AND FLUIDS OCCURRING WITH THEM IN THE WEST SIBERIAN LOWLAND BY THE TECHNIQUES OF INDUSTRIAL-GEOPHYSICS

N. A. IRBE West Siberian Scientific-Research Gas-Petroleum Institute

Industrial-geophysical investigations of boreholes yield the most extensive information about the beds revealed by drilling and the degree of saturation of the thawed reservoirs. Based on their data, in the sections of boreholes we differentiate the layers of monolithic permafrost, taliks in the frozen layer, and buried permafrost in the mantle of thawed soils.[3-7]

By physical properties, the cooled soils are divided into types:

1. Frozen soils where the mineral particles are cemented by monolithic ice (icy frozen soils).
2. Frozen soils where the water is mixed with ice (frozen soils containing interrupted ice).
3. Thawed soils.

A combined interpretation of the geophysical data solves the problem of the segregation of these soils. Thermal well-logging at an established heat regime permits us to separate the frozen soils ($t \lesssim 0°C$) from the thawed soils ($t > 0°C$). When the heat regime is unsteady, this problem cannot always be solved. We can obtain satisfactory information about the intervals of frozen soils by utilizing temperature measurements in a cased borehole, conducted by expelling a cement ring 3 days after the pouring of the cement.[4] In an interpretation of the ECR (expulsion of cement ring), diagrams measured after a shorter period of borehole nondisturbance, the ice pockets disrupted by drilling and filled with warm cement can be assumed to be pockets of taliks (Figure 1).

The problem of the identification of the above-mentioned types of frozen and thawed soils in western Siberia is solved by electrometry when the subpermafrost reservoirs are saturated with saline water.[3,4] If the subpermafrost water is fresh (not saline), the frozen sands are identified with the aid of standard logging, by the apparent resistances exceeding 750 Ω · m, while the water-bearing sands are identified by resistances less than 50 Ω · m. The question concerning the condition of soils at intermediate values of apparent resistances in the pockets and at any resistances in the clays remains open.[4]

Since the cohesion of particles in frozen soils is higher than in the thawed soils, we can identify the intervals containing icy soils from the rate of cutting during drilling (mechanical logging).

The method of pocket measurements yields information about the undisturbed and disturbed parts of frozen and thawed soils.

The method of natural radioactivity (NR) permits us to classify the aquifers by their degree of silting with mud, i.e., to determine the location of the pockets where water is most abundant.

Acoustic logging is capable of separating the icy soils from the thawed ones and from the frozen soils with interrupted ice content; the technique relies on the varying velocity of sound propagation.

As a rule, the method of natural field potentials (NFP) in a frozen layer does not convey information.

We have shown in Figure 1 the results obtained from an interpretation of industrial-geophysical investigations in a frozen layer in one of the exploratory boreholes in the Far North of Tyumenskaya Oblast.

The diagram of NFP has been measured qualitatively only below the conductor's contact; therefore, the lithologic column has been compiled from the electrical and gamma-logging data. Thermometry for the purpose of ECR was conducted a day after the cementing of the pipe and re-

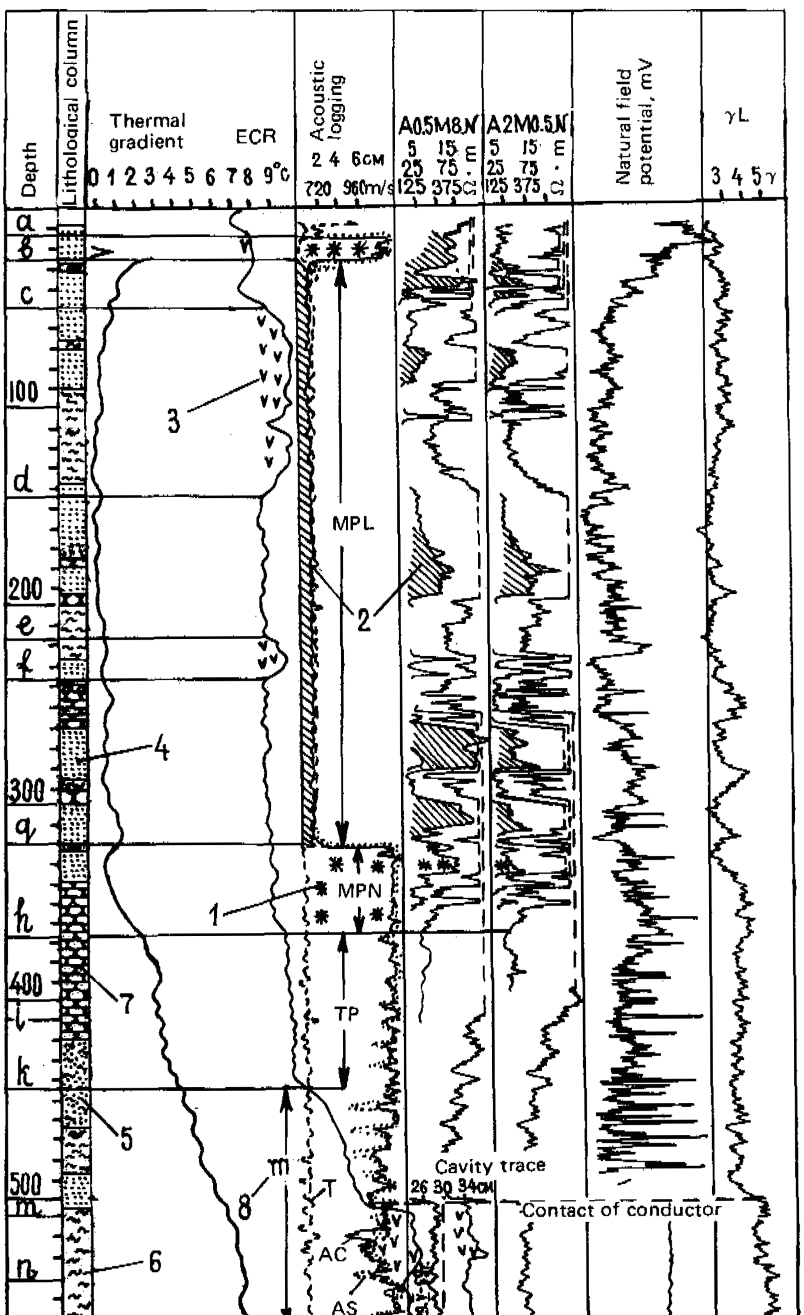

FIGURE 1 Geophysical characteristics of frozen and thawed sediments: 1--mixture of icy mass with water (MPN = frozen soils with interrupted ice content; 2--frozen icy soils (MPL); 3--gaps warmed through filling the cavities with cement; 4--sand; 5--loams; 6--clays; 7--siltstones; 8--thawed soils (TP).

peatedly every 12 days. In both thermograms the frozen layer from the surface to depth *h* was identified by the relative temperature drop. Anomalous deviations toward a temperature rise against a background of the overall slope of the ECR thermogram (intervals *c*--*d*, *e*--*f*, *m*--*n*) constitute a reflection of the enlarged pockets filled with a greater amount of warm cement than adjoining sections. On the thermogram taken later, these thermal pockets had disappeared.

The standard logging curves indicate the limits of permafrost. The apparent resistances measured by gradient-probe A2MO and 5N above 750 Ω · m separate the frozen layer from the surface to *h*, while resistances less than 25 Ω · m demarcate the lower boundary of the frozen stratum. The combined interpretation of the data from electrometry and thermometry provides a basis for considering that in testing the *h*--*i* interval, we will find a greater inflow of fresh water, while on testing the *i*--*k* interval, the inflow will be less.

Measurement with an acoustic cement meter was conducted in the casing string after 11 months. The extent of the cement's adhesion with the casing pipe is shown by the curve of wave-attenuation for the pipe-cement (A_c), while the cement's adhesion with the soil is reflected by the curve of wave damping for cement-soil (A_s). Absence of cement beyond the pipe is identified by maximum A_c and A_s values, while its presence is indicated by the minimum values. Cement is lacking between the conductor and the casing pipe, while the free space is filled by a column of drilling fluid. Ice in the pipe system rigidly cements the pipe with the soil from the surface to depth a and in the *b*--*q* interval. The above-lying soil layer *a*--*b* and soil layer *q*--*h* below were not cemented with the pipe. This is obviously a zone where the continuity of the ice has been interrupted and the pieces of ice are mixed with water.

The freshwater reservoirs within the limits of the last two types of cooled soils can serve as water-supply sources.

The airtight adhesion of the casing pipe with the icy permafrost permits us to make a statement about the capability of the icy permafrost soils to serve as a gas barrier. Toward the northeast of the center of the western Siberian lowland, there occurs a loss in blanketing and an upwardflow of gas along the section. We postulate the presence of gas deposits under the continuous layer of frozen soils in a number of areas in the Ust'-Yenisey depression,[5] where the stratal water is saline and the high resistances can be tied in with the extent of contained gas (Figure 2).

Utilizing the results from the theoretical and experimental studies conducted by A. A. Trofimuk, N. V. Cherskiy, F. A. Trebin, V. G. Vasil'yev, Yu. F. Makogon, and V. P. Tsarev,[8] we assume that the gas, detected with the geophysical data, in the cooled reservoirs exists in solid [*sic*] form. The basic raw data for calculating the equilibrium curves of hydrate formation have been taken on the basis of geophysical data: temperature based on thermograms recorded for determining the geothermal gradient and pressure proportional to the depth of the object, on the electrometry diagrams. The calculation of the equilibrium depths above which gas can prove to be in a hydrate state and below which it can prove to be in a free state has been conducted in three modifications: for upper Cretaceous beds with specific gas weight Δ = 0.58, lower Cretaceous Δ = 0.6, and Jurassic Δ = 0.7.[2,6]

The level of contained gas in hydrate beds will fluctuate from 30 m at the Tampeyskaya

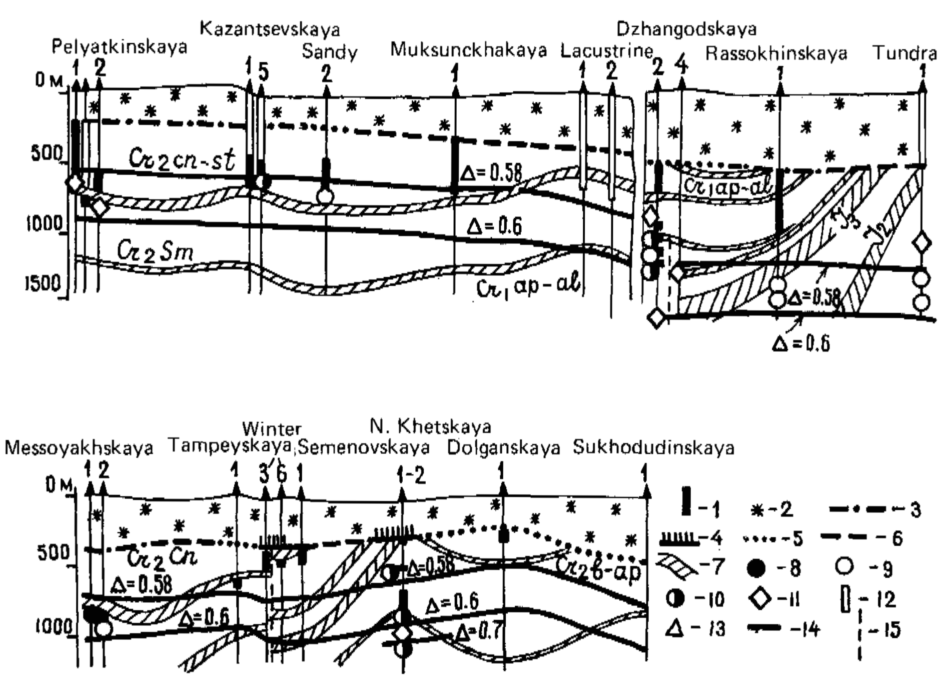

FIGURE 2 Distribution of gas-hydrate deposits in sections of Yenisey North. Compilers: N. A. Irbe, V. M. Starilova, and V. T. Raykova. 1--intervals of increased resistance attributable to hypothetical gas occurrence; 2--permafrost soils; base of frozen soils according to: 3--thermometry; 4--electrometry; 5--combination of thermometry and electrometry; 6--interpolation; 7--regional clayey caps above layers of aquifers, based on testing; 8--gas; 9--water; 10--signs of gas; 11--absence of inflow; 12--intervals not investigated by industrial-geophysical techniques; 13--relative specific proportion of gas; 14--equilibrium boundary of hydrate formation; and 15--dislocations with break in continuity.

area to 380 m at Muksunikhskaya and 440 m at Rassokhinskaya. On sampling hydrated gas at intervals not driving with a controller, we obtained either low-yield flows or the wells proved to be "dry" (see Figure 2). Industrial reserves of free CH_4 and of hydrated gas in respect to heavy hydrocarbons have so far been investigated only at the Messoyakhskoye and Dzhangodskoye deposits.[1,2]

Thus the combination of industrial-geophysical investigations of boreholes permits us to differentiate the frozen and thawed soils in a section. For a procedurally correct disclosure of the freshwater aquifers coordinated to the frozen soils, we recommend investigating the boreholes by the techniques used in standard electrometry, gamma-ray logging, interstitial measurement, acoustic logging, and thermometry. In the oil-prospecting boreholes, we recommend that the pertinent temperature measurements be conducted only after a period of 3 days following the pumping of cement into the pipe setup.

The geophysical data testify to the capacity of permafrost to serve as a cap for gas deposits. In the low-temperature section of Yenisey North, we have disclosed gas-hydrate deposits at a level ranging from 30 to 440 m.

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PART VII

Principles of the Control of Geocryological Conditions for Construction in Permafrost Regions

S. S. VYALOV and
R. M. KAMENSKIY, *Editors*

ENGINEERING-GEOCRYOLOGICAL CONDITIONS AND PRINCIPLES OF CONSTRUCTION IN MAGADAN OBLAST

V. G. GOL'DTMAN VNII-I

In Magadan oblast several engineering-geological formations have been isolated of which the most widespread are the formations of pre-Paleozoic crystalline rocks and late Mesozoic granitoid intrusions, Mesozoic shales and sandstones, Cretaceous extrusive rock, ice-rich alluvial-lacustrine deposits, and the azonal formation of alluvial gravel.

With respect to the nature of the distribution of the frozen layers and temperature of the rock, four geocryological regions are distinguished in the investigated region. The first is characterized by permafrost islands and temperatures of 0°C to -1.5°C at the depth of the annual zero amplitudes; the second is characterized by discontinuous permafrost and temperatures to -3.5°C; the third is the region of continuous permafrost 150-500-m thick, with rare continuous taliks under large bodies of water and at points of emergence of groundwater, and a temperature of -4°C to -9°C; the fourth is distinguished from the third by annual heat cycles of smaller magnitude (by two or three times). It includes the coasts of the Chukotskoye and Bering seas.

The engineering-geological formation of pre-Paleozoic crystalline rocks and late Mesozoic granitoid intrusions is found within all four geocryological regions. The rock massifs are complicated by a set of faults of different age, along which the rock is intensely metamorphosed. Frequently, in such metamorphosed zones with permafrost, the cohesion between the rock-forming minerals is weakened; during thawing of the rock there is noticeable settlement, and under the mechanical effects the rock breaks down to talus and rock debris.

Within the third geocryological region in the granite massif, there are horizontal layers of ice 20-40-cm thick confined to the fissures. The layers extend tens of meters and on a slope go to a depth of up to 30 m. The thawing of the ice in the fissures would lead to the settlement of the entire overlying massif. The ice that fills the cracks of various orientations in the granite several centimeters in width is detected in various places during mining and engineering-geological exploration. Thus, even the bedrock of the investigated formation is always far from a faultless foundation.

The residual gravel of crystalline rocks and granites is highly permeable. If it is frost-heaved, then during thawing it is immediately compacted. Under conditions of region I, the granitoids and products of their residue-colluvial porcessing are not subject to frost heave and serve as good natural foundations for building.

The engineering-geological formation of shales and sandstones of the Mesozoic folded structures exist in the mountainous zone. With gently sloping undisturbed bedding the bedrock beds of shales and sandstones in any of the geocryological regions can be used as stable foundations not resulting in settlement and thawing. The coefficient of volume change [m_V] of this rock determined during thawing under load at a depth of 3-6 m are 0.0005 to 0.0001 cm^2/kg. This permits construction with the assumption of uncontrolled thawing of the foundation material.

In places in the zones of intense tectonic faulting with increased fracturing near the large faults, the rock is converted into gravel with supes or suglinok. Here the ice sometimes takes up half of the total volume of the rock. In such zones of increased energy consumption in phase transition during freezing extending from above, closed taliks were probably formed that swelled under hydrostatic pressure, later freezing in the frozen rock. This explains the significant settlement of the rock on thawing. The coefficient of volume change of such shales and sandstones at a depth of 20 m during thawing under pressure are within the limits of 0.01 cm^2/kg to 0.002 cm^2/kg. At a pressure of 3 kg/cm^2 the thawing was accompanied by a settlement of 4 cm/m. It is probable that the prolonged effect of shearing stresses on such ice-saturated rock in the frozen state would be creep.

The residual colluvial materials of shales and sandstones, represented by gouge suglinoks or supesses in the natural frozen state, usually contain an excess of ice in the form of lenses, and on thawing they settle about 7 cm/m. The mechanical properties of this type of soil depend on the granulometric composition, and the mineralization of the porewater, which is increased; they also vary with temperature.[1]

The samples of gouge soils with mineralization of the porewater of 1.5 g/l are characterized by long-term cohesion at a shear rate of 0.05 mm/h and a temperature of -4.5°C at 7 kg/cm^2; and at -1°C, 3.5 kg/cm^2; the short-term cohesion with a shear rate of 5 mm/h and more turns out to be about 9 kg/cm^2 at the same temperatures. The angle of internal friction at a temperature of -1°C to -4.5°C during slow shear is 24°-29°, and, during fast shear, 30°-32°. With a drop in temperature from -5°C to -10°C, the shear strength of the suglinok almost doubles.

The noted strength characteristics along with the low permeability enable such ground to be used as a foundation according to principle no. 1. In some cases preconstruction thawing is used. Satisfactory compaction of the shale-sandstone-gouge suglinok under the weight of the overburden 6 m thick and more is observed.[2]

The extrusive rock and the residual-colluvial products as a type of engineering geological formation were investigated within the third geocryological region. A study was made of the upper Cretaceous andesites and andesite-basalts. In the thawed state this rock is quite permeable along the faults. After the prolonged effect of groundwater during the Neogene this rock experienced the disrupting effect of permafrost several times during the Quaternary.

During drilling and mining operations in the zones of very strong andesites or andesite-basalts, sections of rock are found where the cohesion between the mineral grains has been weakened, and the plagioclases have been completely replaced by saponite. Often a strong core of frozen andesite from a depth of 20 m to 80 m breaks down into gravel a few days after thawing. In individual samples of weathered andesites, the ice content is up to 150 kg/m^3. Even at a depth of about 30 m, horizontal interstitial layers of ice 5-7 mm thick are detected in the andesite gravel every 2-3 cm.

The fact that such frost heave is characteristic of permafrozen andesites over a large area is indicated by observations made with hydraulic thawing to a depth of 25-30 m. The benchmarks cemented at a depth of 5 m (in this way the settlement of the upper 5-m layer is excluded from the investigation) recorded a settlement of the thawing layer 15-20 thick of 130-200 mm. Judging by the intermediate benchmarks attached at different depths, the relative settlement decreases somewhat with depth, from 0.005 to 0.007. The settlement takes place simultaneously with thawing and decreases over a period of 2 or 3 months.[3]

Observations of the foundation of a concrete spillway at a dam erected on similar permafrozen andesites and andesite-basalts indicate thawing to a depth of 80 m (during 13 yr of operations) and a settlement of the thawed rock by 16 cm in connection with this.[2]

Analyzing the causes of frost heave of permafrozen fractured rock, in particular, andesites at great depth, we begin with the concept that during the cooling epoch the lower boundary of the frozen strata at a depth of tens and hundreds of meters was displaced at a low rate, on the order of 1-5 cm/yr with temperature gradients of 0.02-0.1 deg/m. In these conditions, the temperature field was complicated, as was noted above, by nonsimultaneity of freezing of the zones of variously fractured saturated rocks. In addition, even in the more or less uniformly fractured massif, the cause for heaving of the rock could be different widths of the fractures in the different planes, for example, the greater width of the vertical cracks by comparison with the horizontal ones, and so on. Obviously, in the wider fissures the freezing of the water ended later when the fine cracks were already filled with ice; then efflux and squeezing out of the water was difficult. The hydrostatic pressure created by the growing ice crystals in these joints could exceed the overburden pressure and cause heaving, predominantly in the vertical direction, inasmuch as there was freedom for displacement upward. In the case of multiple freezing over many years, irreversible deformations of the massif accumulated, which increased the total mean widths of the fractures and the ice content of the rock. Depending on the paleohydrogeological conditions, more precisely, the level of the nearest erosion base and the groundwater level, the permafrost occurred during complete saturation of the fissures with water or in the free joint zone. In the latter case the fissures were usually filled by sublimation ice, which did not have the splitting effect.

The alluvial gravel in the valleys of mountain rivers and their thicker deposits in the tectonic depressions extend within the entire mountainous region. Although there are significant lithologic differences between the gravels because of the composition of the bedrock, the hydrogeological conditions, and so on, in engineering-geocryological respects the deposits have many common features.

Within the geocryological regions II, III, and IV, the permafrozen gravel is heaved with ice in both syngenetic and epigenetic freezing. This gravel compacts on thawing as much as 4 cm/m. In the thawed state it differs from fissured rock in its higher permeability.

The frozen ice-saturated gravel at a temperature of -5°C with shear at a constant rate of creep has a cohesion of 4.5 kg/cm^2 and an angle of internal friction of 22°, and in the case of rapid shear the figures are 8 kg/cm^2 with an angle of 29°. In the case of uniaxial compression with a load of 13 kg/cm^2, steady-state creep was observed with a deformation rate of less than 0.0002/h, and at a stress of 35 kg/cm^2 with a rate of about 0.01/h.

The high strength of frozen gravel would permit of its use everywhere for foundations according to the principle of maintaining the frozen state; however, under the conditions of mountainous topography, it is not always possible to exclude unforeseen thawing of the gravel as a result of heat transfer by the groundwater. Cases of gradual deepening of unforeseen thawed zones of the foundation, accompanied by settlement, are known during the 15 yr of operation.

Gravel with a coefficient of permeability within the limits of 20-400 m/day at a flow temperature of 10°C and a hydraulic gradient of 0.01-0.1 can thaw in 1 yr, to a depth on the order of 7-12 m and cause settlement of the foundation by 0.5 m. Accordingly, construction on permafrozen settling gravel in the corresponding cases must be based on approximate calculations of the probable causes of water percolation. Here, procedural developments with respect to determining the heat transfer of the horizontal flows and the thawing rate proposed by V. V. Znamenskiy[5,6] are applicable for planning and designing drainage during thawing.

The city of Magadan, located on gentle slopes on the seacoast, is in geocryological region I. Most of the structures of this city were erected on Quaternary alluvial and colluvial gravel or sandy deposits under which, at a depth of 5-15 m, there are poorly cemented Neogenic conglomerates, sandstone, or granite. Under these conditions foundations in the form of piles with sole plates have found application.[7] The transfer of load through the footing to the rock at a depth of 6-10 m excludes any effect of settlement, and the frost heave developed during insular occurence of the permafrost and deep seasonal freezing of the taliks. In addition, the effects of nonuniform reaction of the ground with extreme variability of the foundation soil are excluded.

In the coastal lowlands in the Anadyr' River basin and on the shores of the Artic Seas, ice-rich heavy supesses with ice streaks and lenses are widespread. When building on such soil, the builders are guided by the principle of keeping the foundation in the frozen state. They use frozen-in vibrated piles. They use fills up to 1.3 m high erected in the winter immediately to the full height. Only in individual cases in hydraulic engineering construction can artificial construction thawing be used for complete removal of the underground ice.

From the briefly described characteristics of the geocryological conditions of the Magadan oblast, it is obvious that in the mountainous regions with high permeability of the soil the principle of preliminary thawing is nearly always less expedient than keeping the soil in the frozen state. In practice, both principles find application here, although the soil is gravelly and cases of unpredicted thawing of the foundation can hardly be entirely excluded.

When selecting the principle of the use of frozen ground for the foundation for a building in a given case, it is important to consider not only the achievement of minimum construction costs, but also the probable operating costs, which are minimal with preconstruction thawing. One must also not ignore the possibility of local or regional variations in climate as a result of which the temperature of the permafrost may rise during the period of operation.

Under investigated conditions, the application of construction thawing will obviously insure definite advantages; however, it must be stated that, in specific cases of planning and design, problems in using this method arise from inadequate study.

Further study of permafrost differing with respect to composition and cryogenous texture are required to determine the behavior during thawing. We have in mind the speed and completeness of settlement under the effect of load and also the efficiency of the specially applied techniques on the thawing rock, for example, washing out the vertical underground columns for drainage of the consolidated layers with the well-point method of hydraulic thawing, vibration of the well-points, temporary drainage of the thawed zone by deep lowering of the groundwater level, and other methods. The results of the research permit more precise selection of the thawing technique (hydraulic, steam, electric, and so on).

The problem of the minimum depth of preconstruction thawing in the simplest cases is solved by excluding the possibility of settlement of the ground during building operation, with an inadmissible rate or nonuniformity as a result of an uncontrolled increase in the thawed zone.[4] The solution to this problem is complicated if rocks lie under the artificially thawed zone and give significant settlement during thawing, and also if it is necessary to admit the possibility of defined variations in the temperature field of the ground in the future under the effect of external factors, including the percolation of groundwater. Therefore, the development of research in the nature of frost heave and the peculiarities of this process during slow deep freezing of fractured rock is indicated.

The problem of establishing the role of the geochemical variations of permafrost, having as their consequence weakening of the rock after thawing and possibly the exothermal physico-chemical processes, demands attention.

It is not always possible with sufficient grounds to answer the question of whether it is admissible for thawed and consolidated ground to be frozen back during the operation and life of the building and what requirements must be imposed in this case on regulating the groundwater level and excluding the causes of temperature variations.

In addition, a number of problems remain with respect to improving the technological process of preconstruction thawing.

In this report attention has been given to the problems of using rocky ground and coarsely broken ground as foundations for buildings. Of course, the problems of using fine-grained permafrost, while retaining the frozen state and with artificial lowering of temperature, are also urgent in the Magadan oblast.

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EXPERIENCE IN CONSTRUCTION ON PERMAFROST IN THE VICINITY OF NORIL'SK

V. N. KOLYADA, L. I. ANISIMOV, V. YE. POLUEKTOV AND A. P. ZAVENYAGIN Noril'sk Mining and Metallurgical Combine

The creation of the Noril'sk industrial complex and the city of Noril'sk beyond the sixty-ninth parallel under severe climatic, geological, and geocryological conditions was connected with solving a number of the most complicated problems in various fields of science and engineering. The basic problem is the construction of multistory civic buildings and large industrial structures, including metallurgical shops, large capacity electric power plants, and the installation of engineering utilities and roads under permafrost conditions.

GEOLOGICAL AND GEOCRYOLOGICAL CONDITIONS

The area of industrial and residential construction in Noril'sk is distinguished by complicated engineering-geological conditions. The thickness of the permafrost varies from 0 m to 400 m with a temperature at the depth of zero annual amplitude of 0°C to -4°C. In spite of the significant thickness of the permafrost, its discontinuity is observed both in plan and in vertical section. In the zone of the basic industrial construction talik islands occupy up to 30 percent of the area and have complicated configurations. The temperature gradient varies between 15 m/deg and 42 m/deg.

Bedrock lies at a depth of from 0 m to 80 m and consists of gabbrodiabases, dolerites, porphyrites, marls, limestones, gypsum, shales, dolomites, sandstones, argillites, and aleurolites. The structure and depth of the bedrock frequently is different even for the foundation of a single building.

The Quaternary deposits represented by fine-grained supesses and suglinoks with layered and reticular textures, fine sand and gravel of massive texture, clay, and peat are no less varied with respect to composition and cryogenous structure. The degree of ice-saturation frequently reaches 40 percent. The thaw-settlement of this rock without load varies between 0.5 and 40 percent.

CONSTRUCTION OF FOUNDATIONS

The problems of foundation construction under the given geological and geocryological conditions were solved in steps depending on the nature of the construction, in four phases, considering two principles:

1. With significant depth of bedrock the foundations of structures not having industrial heat generation and providing for the maintenance of the frozen state of the ground by their structure for their entire life are laid on granular permafrost deposits.
2. The foundations of structures, the design or industrial use of which excludes the possibility of preserving the frozen state of the ground, are erected on bedrock or nonsettling frozen ground.

In the first phase of the development of the region (1935-1945), the construction of temporary structures, experimental shops, and two-story residential buildings on wooden seats with antiheave anchors at their base and on so-called "pulsating posts" prevailed. The latter were set at a depth of 20-30 cm from the surface, and the "seats" were buried to 1 m below the active layer. The preservation of the frozen state of the ground in the foundation of the structures were insured by constructing air spaces. An example of the application of foundations of the type described is the construction in 1949 of the wooden trestle that is still used today for the Noril'sk water line 1,300 mm in diameter and 9 km long. The trestle, constructed on wooden "seats" and crib supports, has received only minor repairs in more than 20 yr.

The experience in using "pulsating" posts for light buildings has proved their efficiency only in the case of building on coarse granular soils with insignificant ice content. On fine grained ice-rich soils, they are subject to inadmissible settlement and heave.

The calculation of the bearing capacity of a foundation was made in the first step using empirical data on the resistance of permafrost to compression without considering adfreezing at the lateral surface of the foundation.

The second phase, which began in 1946, took into account the construction of major industrial and five-story civic buildings.

The industrial buildings were erected on concrete posts and reinforced concrete pads on rock to a depth of 20 m.

In the case of deep bedrock, the civic buildings were first built on rubble strip footings with a depth to 3 m and then as the structural properties of permafrost were studied and the height of the buildings increased, on concrete posts and reinforced concrete pads laid below the active layer and supported on the granular permafrost deposits with a temperature no higher than -1.5°C. The bearing capacity of the foundations was also considered, taking into account the resistance of the permafrost only to compression. The concrete pads were placed at any time of year.

The construction of post-on-pad foundations required the manual excavation of a significant volume of earth using only pneumatic drills. The construction of deep foundations on a rock base using rock excavation procedures entailed much labor and increased the cost and delayed the construction. It must be noted that the execution of these foundations in the winter was more efficient than in the summer, when excavation of pits in the ice-rich ground demanded support of the walls and was accompanied by a heavy inflow of water.

Joint experimental search by the researchers and builders of the Noril'sk Mining and Metallurgical Combine led to the third phase of improving the construction of foundations by converting to pile foundations, which were first used in 1959 in the civic construction in Noril'sk after a number of experiments determining the construction process. Here principle 1 for using permafrost as a foundation was taken as the basic principle, and the bearing capacity ϕ of the piles began to be determined according to Construction Norms and Regulations (SNiP II-B, 6-66) from a calculation of the total resistance of the permafrost under the end of the pile--the bearing ϕ_c--and along its lateral surface--the shear ϕ_s

$$\phi = \phi_c + \phi_s \qquad (1)$$

The design temperatures along the operating length of the pile and at the level of its point are determined by ground-temperature measurement data during the period of maximum temperatures--the month of October.

In the absence of these data, the design temperatures are taken as a function of the ground temperature--T_0--at the level of zero annual amplitude under the condition that T_0 is not higher than -2°C (in the vicinity of Noril'sk for construction purposes the value of T_0 is taken by the measurements at a depth of 14 m). The design temperature, T_1, of the layer in which the piles are set in the permafrost is determined by Formula 2 and at the level of the point of the pile, T_2, by Formula 3:

$$T_1 = K_{t_1} T_0, \qquad (2)$$

$$T_2 = K_{t_2} T_0, \qquad (3)$$

where T_0 is the ground temperature at a depth of 14 m; Kt_1 and Kt_2 are the coefficients, the values of which are given in Table 1.

The piles were set by cranes in predrilled holes larger in diameter than the maximum cross section of the piles after which the holes were filled with slurry. Such piles were called "drill-set frozen piles." Primarily piles 32 × 32 cm in cross section from 8 m to 12 m long were used with intensified reinforcing of the upper section and designed for a load of up to 100 tons. When building on plastic frozen ground or with significant design loads, piles up to 16 m long with the same cross section and

TABLE 1

Depth of Piles in Permafrost, m								
	2.0	3.0	4.0	5.0	6.0	7.0	8.0	9.0
K_{t_1}	0.47	0.68	0.85	0.97	1.0	1.06	1.03	1.0
K_{t_2}	0.24	0.36	0.46	0.55	0.62	0.69	0.74	0.76

combination piles with a thermally expanded base made from grouted, or from cast-in-place concrete, are used.

The holes 450-500 mm in diameter are drilled using percussion drilling rigs and the ABU thermomechanical drilling unit that excavates and blows the frozen ground out of the hole under the simultaneous effect of a mechanical drilling tool and a high-temperature gas jet with supersonic velocity. The mean output capacity of the unit is 10 lineal m/h. The unit has a hoist to assist in placing the pile.

The boreholes are back-filled with sandy slurry with the addition of a lime paste or clay as a plasticizing agent. Observations have shown that the adfreezing of the slurry to the piles and the restoration of the natural ground temperature occur in 5 to 16 days depending on the type, time of year, and method of drilling. The load is transfered from the building of piles by poured reinforced concrete grade beams.

In order to maintain the frozen state of the ground in the foundations of buildings, ventilated air spaces 1.0-1.5 m high are constructed

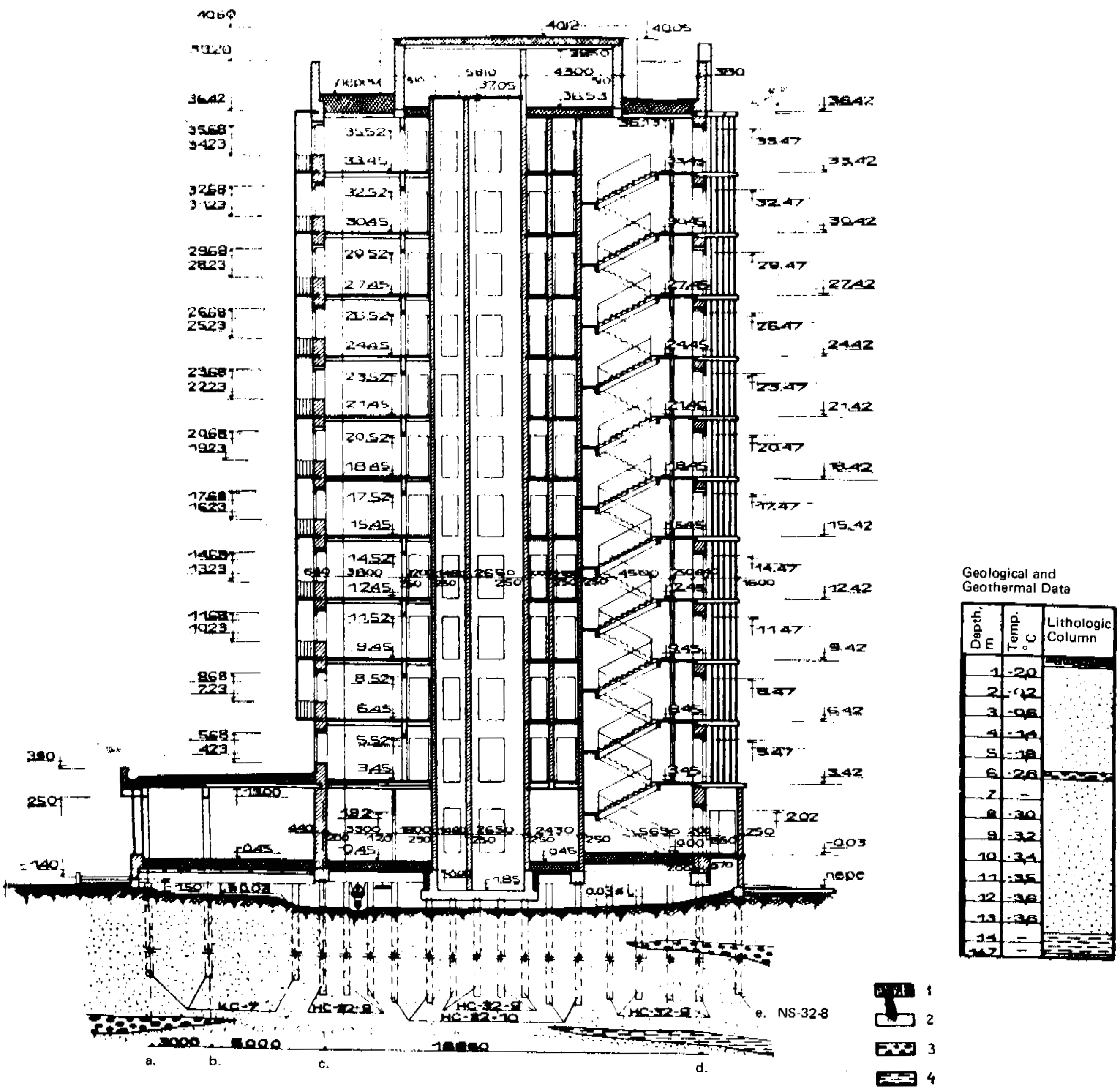

FIGURE 1 Transverse cross section of a typical 12-story residential building constructed on permafrost in Noril'sk: 1--fill--gravel, pebbles, slag with sand to 30 percent; 2--sand of various grain sizes with an ice content to 20 percent, water content when thawed; 3--gravel-pebbles with sandy supes; water content to 30 percent; 4--supes with inclusions of pebbles and gravel to 20 percent, layered cryogenous structure with ice content to 20 percent.

with dwarf walls to keep out drifting snow and with ventilation holes to provide the required air exchange. In order to drain off the surface water and accidental water, the air spaces are provided with properly planned and constructed reinforced concrete drainage channels under the engineering utility lines. The transverse cross section of a 12-story building constructed in Noril'sk on pile foundations with a ventilated air space is illustrated in Figure 1.

The temperature conditions of the ground enclosing the piles are recorded by means of temperature wells installed with the piles under various parts of the buildings.

Thirty years of experience in constructing major buildings on permafrost in Noril'sk, including 13 yr using pile foundations, has demonstrated the following:

1. The cost of the reinforced concrete pile foundation is 2.4-2.9 times lower than the cost of the concrete post foundations.
2. With proper execution of all the structural designs of the ventilated air spaces and observation of the operating and maintenance rules, the ground temperature at the base of the piles drops 0.5°-3.0°C in 2-4 yr. This is explained by the insulation of the ground under the buildings from solar radiation during the summer and the absence of snow cover during the winter. The data on the temperature variation of the ground at the base of the piles are presented in Table 2. Not one case of deformation of such buildings has been noted.
3. Even a reduction in temperature of the plastically frozen ground in the foundations of the structures having a natural temperature of 0°-0.2° was observed. It was noted in the foundations of a number of five-story residential buildings constructed on piles and having an open (but snow-protected) air space 1.0-1.5 m high 2-3 yr after the beginning of the operation of the buildings. The settlement of such buildings has not exceeded 2 m.
4. In five-story buildings having partly solidly frozen and plastically frozen ground in the foundation with a temperature of about 0°C constructed on piles with widened tips, no deformations have been noted in 2-4 yr of operation and maintenance.
5. When permafrost was thawed during operations at the construction sites, the restoration of the frozen state of the ground was achieved by blowing out the holes before installing the piles using shaft fans through canvas ducts 200 mm in diameter for 4-6 days with an outside air temperature of no more than -20°C.

Pile foundations began to be used for industrial construction in 1963 in sections with comparatively shallow bedrock. The piles were set on rock and were point-bearing. Thawing of the ground was permitted in this case.

The first experimental construction demonstrated that the application of point-bearing piles results in a reduction in the cost of the foundations by 2.4-3.0 times and a reduction in labor by 3-4 times. Since 1965 all of the industrial buildings over bedrock more than 5 m deep have been erected on such pile foundations. The piles are installed as the frozen-in piles were, but in contrast to them they are recessed into the bedrock 1.5-2.0 m in so-called "rock sockets."

Shell piles 400-600 mm in diameter up to 24 m long and square reinforced concrete piles 32 × 32 cm in cross section from 6 m to 16 m long type NS-32 were first used as the piles.

Various methods were used to install the piles and make them monolithic in the boreholes during the initial period:

1. Separate pouring of concrete by filling the spaces (and the cores for shell piles) with dry coarse aggregate and subsequent grouting with cement-grout fed to the base of the hole through an injection pipe using a pump under a pressure of 3 to 5 atmospheres [*sic*].
2. Vibratory placement of the piles in plastic concrete (type M-300 kg/cm^2 with a slump of 16 cm) prepoured in the hole.
3. Setting the pile with a crane under only its own weight in a cement-sand mix (type M-250 kg/cm^2) previously poured into the hole.

In all three cases wire electrodes inside the piles were used to heat the concrete or mortar electrically.

An analysis of the experience in constructing more than 800 piles demonstrated the following: In view of the great labor and nontechnological nature of the grouted aggregate technique, this method can be recommended only when the pile support is of shattered bedrock. The method of vibratory placement of the piles in concrete is more economical, but it requires the application of additional equipment--the vibrator--which greatly complicates the construction.

The most efficient method turns out to be backfilling the hole with cement grout. This method was subsequently used as the primary method but with some improvements. The cement-sand mix was pured into the hole after pile-placement (in deep holes vertically movable tubes were used for this purpose); the piles were made monolithic using this mix to a height of not less than 4 m; the rest of the annulus was filled with drilling mud, which subsequently froze to the piles.

Simultaneously, in order to increase the bearing capacity of the piles, their cross section was increased from 32 × 32 [square] to 40 × 40 cm [octagonal]. The chamfers permitted installation of the piles in holes of the same diameter (650 mm). This type of pile came to be called NSF-40. Along with these piles, shell piles 600 mm in diameter [100 mm WT] continued to be used.

The bearing capacity of the piles is determined in accordance with the construction Norms and guidelines (SNiP II-B.5-67 and SNiP II-V.1-62) from the condition of the ground on which the pile is supported and from the strength of the pile material. In the latter case, the longitudinal bending of the pile in reducing the load capacity is taken into account in the calculation.

TABLE 2 Temperature Variation in the Pile Foundation of Multistoried Buildings Constructed on Principle No. 1 in Various Blocks of the City of Noril'sk (According to Data of the Noril'sk Permafrost Laboratory)

Construction No.; No. of the Building	Soil Characteristics	Depth of Measurement, m	Date of Survey	Temperature during Survey, °C	Date of Acceptance of the Building	Temperature on Acceptance of Building, °C	Indexes of Subsequent Temperature Measurements, °C					
							Date of Measurement	Temperature	Date of Measurement	Temperature	Date of Measurement	Temperature
192-I	Sand with interstitial	7	12 May	-0.8	14 Oct.	-5.0	25 Sep.	-3.6	8 Oct.	-3.8	13 Oct.	-4.5
	layers suglinok	8	1963	-0.8	1964	-2.0	1965	-3.4	1968	-3.8	1970	-4.4
310-II	Supes with pebbles with	7	29 Jun.	-1.2	20 Dec.	-0.5	20 Dec.	-1.4	2 Dec.	-3.2	21 Dec.	-4.2
	5 m of compact moraine	9	1964	-1.4	1964	-0.9	1965	-1.6	1968	-3.4	1970	-4.4
74	Fine-grained sand, supes,	8	15 Oct.	-0.2	23 Dec.	-1.1	2 Dec.	-1.0	11 Nov.	-2.2	5 May	-2.0
	and suglinok	9	1965	-0.4	1967	-1.1	1968	-1.1	1969	-1.5	1970	-1.8
									5 Nov			
									1968			
192-II	Sand with lenses and	7	9 May	-0.3	14 Sep.	-1.2	22 Sep.	-3.8	13 Sep.	-4.0	30 Sep.	-4.7
	interlayers, supes, and	9	1963	-0.5	1964	-1.2	1965	-3.6	1968	-4.1	1970	-4.9
	suglinok of fluid con-											
	sistency on thawing											
193-II	The same	7	24 May	-0.5	14 Oct.	-1.9	25 Oct.	-3.3	8 Oct	-3.6	14 Oct.	-3.4
			1963		1964		1965		1968		1970	
614	Ice-saturated sand with	10	29 Nov.	-0.2	10 Jan.	-0.5	13 Jan.	-0.7	9 Jan.	-1.2	21 Jan.	-1.8
	inclusions of pebbles	12	1966	-0.5	1968	-0.3	1969	-0.7	1970	-1.2	1971	-1.3
376-I	Gravel-pebbles with	8	14 Oct.	-1.2	6 Oct.	-1.0	29 Oct.	-2.0	23 Oct.	-3.1	1 Oct.	-4.9
	supes filler	9	1965	-1.2	1966	-1.0	1967	-1.8	1968	-3.4	1969	-4.4
516	Sand with sparse gravel	8	12 Oct.	-1.9	30 May	-0.2	16 May	-4.0	13 May	-3.2	--	--
	inclusion	11	1965	-1.7	1968	-0.4	1969	-1.8	1970	-2.7		
520-II	Gravel-pebbles with sand	8	18 Aug.	-2.0	10 Dec.	-3.0	5 Dec.	-3.6	--	--	--	--
	filler	10	1965	-1.8	1966	-3.0	1970	-3.7				
517	Sand with sparse gravel	7	14 Jun.	-3.0	26 Sep.	-0.9	12 Sep.	-4.4	30 Sep.	-3.4	--	--
	inclusion	9	1967	-2.3	1968	-0.9	1969	-4.2	1970	-3.8		

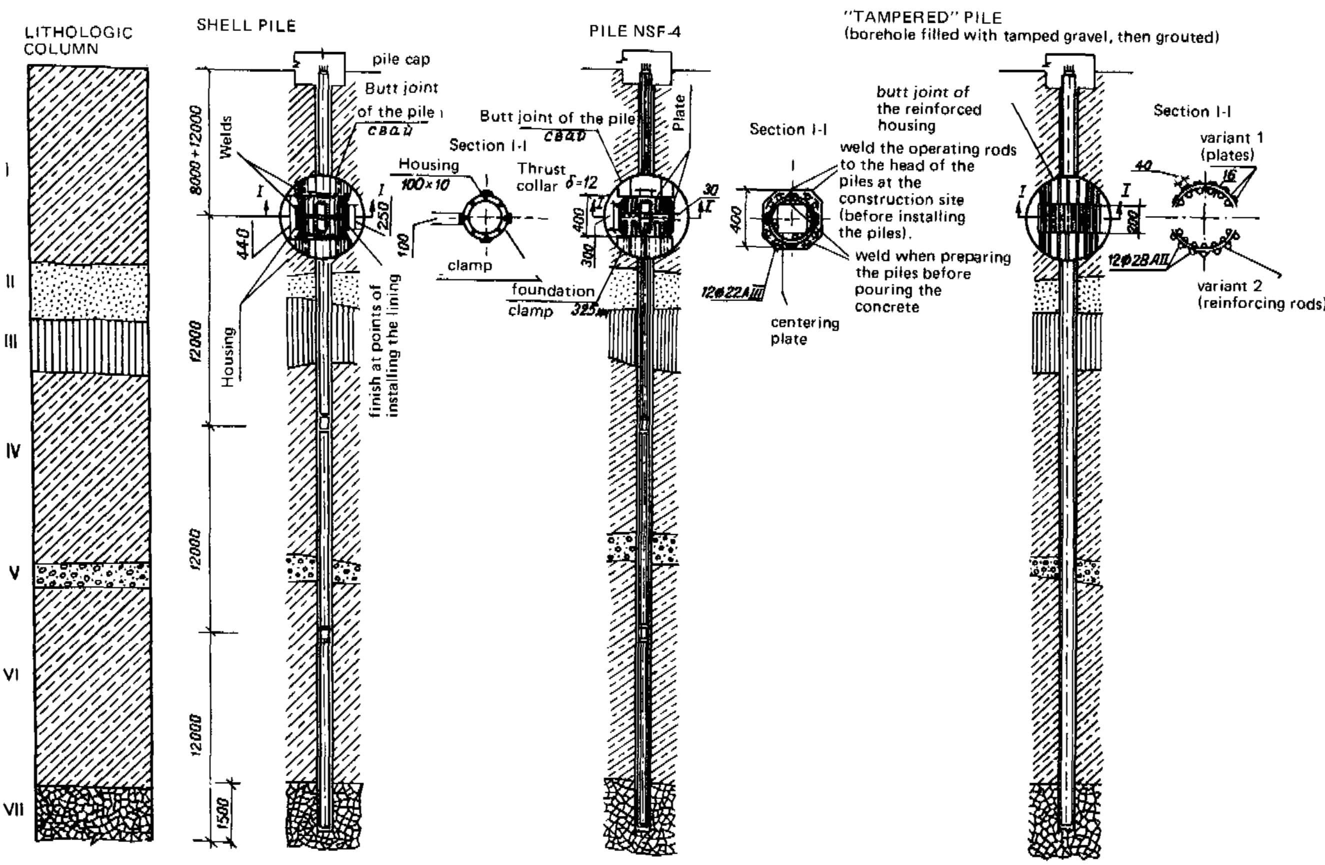

FIGURE 2 Scheme of the construction of point-bearing piles set to a depth of 48 m used in constructing the foundation complex of the Komsomolskiy mine: I--suglinok with pebbles and gravel to 25 percent by volume, ice to 30 percent; when thawed, soft plastic and fluid consistency; II--sand with various grain sizes; wet when thawed; III--supes with interlayers of sand and suglinok; plastic consistency when thawed; IV, VI--suglinok with pebbles and gravel to 25 percent [remainder of title to this figure illegible in Russian text]; V--sandy gravel; VII--must be bedrock.

According to SNiP, the free length of the pile, taking into account its socketing into rock and in a reinforced concrete mat, can be taken as one-half of the total length. The experience of installing a large number of piles (more than 10,000) in the city of Noril'sk, the results of testing 15 piles by static loads, and the data from 6 yr of observations on settlement buildings constructed on these piles offer the possibility of refining the indicated design procedure. Thus, since 1969, when designing point-bearing piles with respect to material, longitudinal bending is taken into account for the length equal to the distance from the base of the foundation mat to the permafrost table plus the length of the pile in the ground.

This variation of the design procedure made possible to significantly increase the design-bearing capacity of the piles and erecting major industrial buildings on a rock foundation at a depth to 50 m.

In order to study the operating conditions of the types of piles described and to estimate their varying capacity in 1970-1971, an experimental installation of 138 shell piles 600 mm in diameter and 22 piles of the precast NSF-40 sections was undertaken. The depth of the piles reached 48 m. The shell piles were placed in boreholes 800 mm in diameter and the whole made monolithic by backfilling with M-250 cement-sand mix, and the SNF piles were buried in holes 650 mm in diameter with the above-described method of making them monolithic. In addition, 24 "tamped"* piles 800 mm in diameter were installed. A structural sketch of the experimental piles is presented in Figure 2.

The work on the pile foundations was performed over the entire year at an outside air temperature down to -35°C. The hardening of the cement-sand mix and concrete and the holes in all types described was provided for by adding Na_2NO_3 or CaCl SLW (sulfide-liquor waste) to them in the amount of 2, 0.5, and 2 percent of the cement by weight, respectively.

The results of the study demonstrated the following:

* [A "tamped" pile consists of a drilled hole filled with tamped layers of graded aggregate and grouted to form concrete of low $W/_C$ ratio and high strength.]

TABLE 3 Technico-Economic Factors in the Construction of Pile Foundations of Different Design with a Depth of 48 m (for 100 Tons of Bearing Capacity)

Factors	Units of Measure	Type of Pile: Shell Pile 600 mm in Diameter	NSF-40	Tamped Pile 800 mm in Diameter
Cost	rubles	5834	5454	2053
Labor	man-hours	242	227	73
Cement consumption	tonnes	2.95	2.66	--
Consumption of precast reinforced concrete	m^3	2.0	2.35	--
Consumption of cast-in-place concrete	m^3	--	--	3.33

1. The bearing capacity of the NSF piles turned out to be equal to 200-360 tons.

2. Out of the two shell piles tested by static compressive loads, one withstood 700 tons and the other withstood 900 tons without significant settlement. A further increase in the load was halted in view of the beginning of deformation of the pile caps serving as reactions for the hydraulic jacks.

3. The minimum strength of the concrete of the cores of two tamped piles 110 days after their construction turned out to be 132 kg/cm^2 (with a mean strength of 253 kg/cm^2), which provide the basis for taking the minimum design load for the pile (with respect to material) as a function of its reinforcement as 600-700 tons.

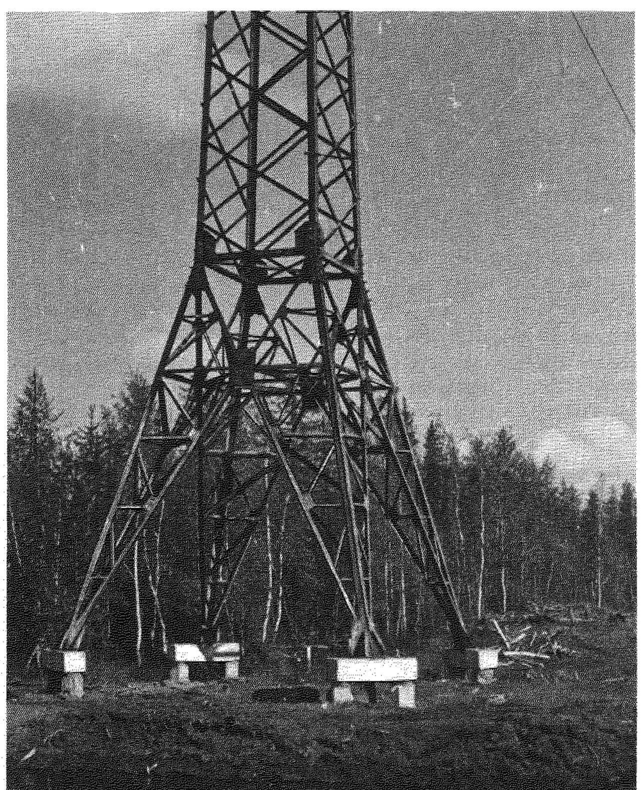

FIGURE 3 Standard pile foundation for an intermediate support of a high-voltage electrical transmission line.

4. The concrete and mortar [made] with antifreeze chemical additives, in contact with permafrost reached 70-75 percent of the brand strength in 90 days.

5. A technico-economic evaluation was made simultaneously of the alternatives for constructing the piles (Table 3). The predominance of the tamped piles is explained by the more complete utilization of the hole cross section, simplicity of the technological process, significantly lower cost of the monolithic concrete, and the exclusion of complex operations connected with the joint sections of the precast piles.

Under the conditions at Noril'sk, transmission-tower foundations are in the form of piles. Under the intermediate supports, 8 piles 7-8 m long are installed, 2 piles with a cast cap at each corner (Figure 3). Under the corner and anchor supports, pile foundations are constructed, usually 12 piles with a poured reinforced concrete mat. The pile foundations are designed just as for the thawed ground of a foundation for the effect of collapse, uplift, and horizontal forces arising under various operational and emergency conditions, except that if frost heave is likely the piles are designed for uplift forces by the formula that is presented in SNiP II-B.6-66 [Norm] but modified:

$$N_v \le km \sum_{i=1}^{n} R^n_{si} F^N_{si} - n\tau^N F_v,$$

where k, m, i, R_{si}^N, F_{si}^N have their generally accepted meanings;* N_v is the design uplift; n

*[km is a coefficient usually taken as 0.9; i is the number of a particular layer (up to n layers); R_{si}^N is the Norm value of the adfreeze stress, pile to permafrost, for the i-th layer; F_{si} is the surface area of the i-th layer embedded in permafrost. From SNiPs.]

is the load factor for heave forces taken equal to 1.2; τ^N is the Norm value of the tangential heave stress (in the absence of experimental data it is taken equal to 0.8 kg/cm^2); F_V is the area of the lateral surface of the pile within the active layer.

LAYING ENGINEERING UTILITY LINES

In order to maintain the frozen state of the ground in a built-up area when building by principle 1, the selection of the optimal plan for laying out the engineering utility lines (water lines, heating ducts, sewers) has essential significance. In Noril'sk, various methods of laying them were tested: directly in the ground, in underground channels not through, and through, utilidors, along the surface of the ground and on trestles. Many years of experience in operating the sanitary-engineering utilities provides a basis for the following conclusions:

1. The laying of heat generating engineering utilities directly in the ground leads to frequent emergencies as a result of their nonuniform settlement during thawing of the ground. It complicates their operation and maintenance and promotes degredation of the permafrost in the construction zone.

2. The laying of water lines and heat ducts along trestlework complicates servicing; it obstructs the building area and does not solve the problem of sewerage.

3. The ordinary through ducts also greatly complicate the operation and maintenance of the utility lines and promote deterioration of the permafrost in the adjacent zone.

4. With insignificant density of construction and beyond its limits ground laying of all the heat-generating lines of the engineering above-utilities is entirely justified. In this case the heat lines with corresponding thermal insulation are laid on sliding supports, the base for which is wooden cribs no less than 0.5 m high (Figure 4A) or prisms of nonsettling materials (gravel, mettallurgical slag). The dead [anchor] supports consist of two piles. This method of laying the engineering lines combines economy, minimum labor, convenience of operation, and the possibility of maintaining the frozen state of the ground in the built-up area.

5. The concentration of all the engineering utilities including the electric power supply and

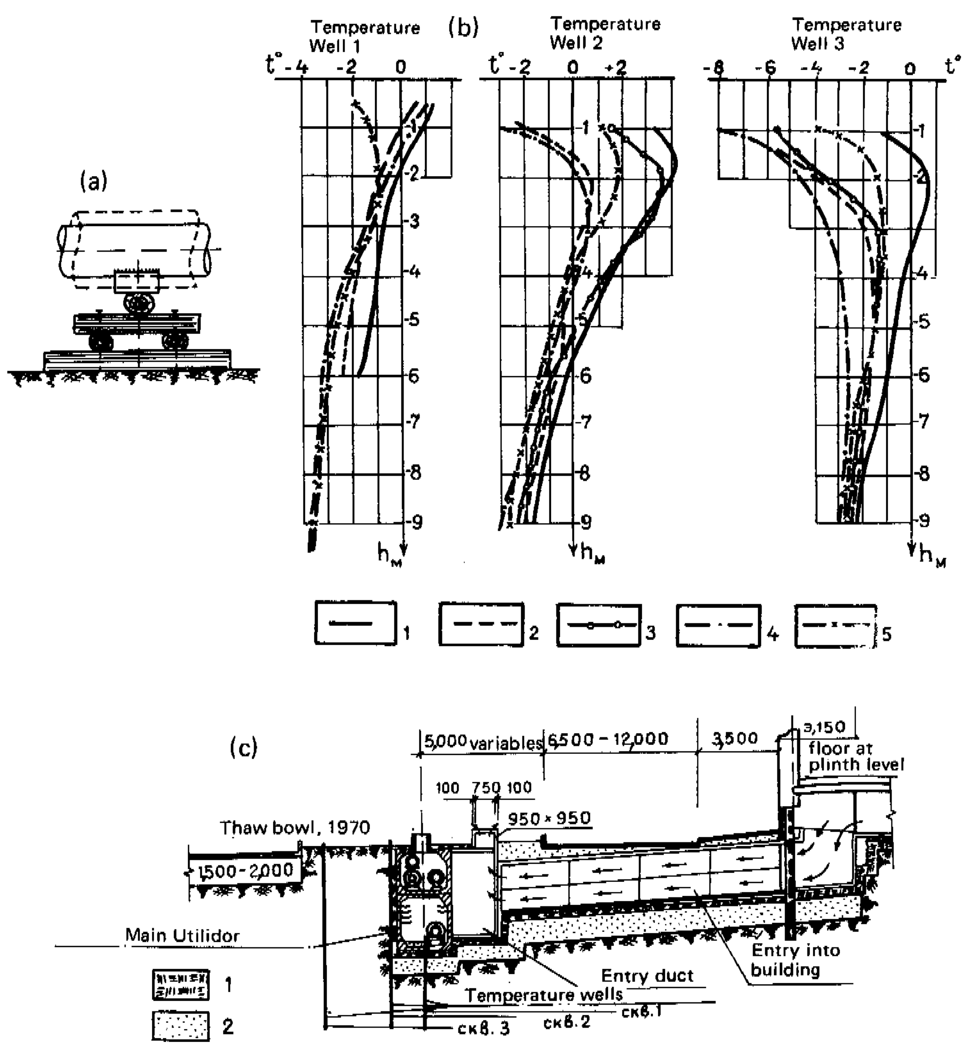

FIGURE 4 (A) Wood cribbing for on-ground laying of engineering utilities lines; (B) ground temperature variation in the zone of two-stage utilidors during their operation. The depth of [temperature] well 1 is from the bottom of the utilidor, and the depth of wells 2 and 3, from the ground surface. 1--1964, 2--1966, 3--1967, 4--1969, 5--1970. (C) standard section of a street with a main utilidor and input to a building. 1--stabilized soil; 2--backfill.

communications in the underground two-stage utilidors has been entirely justified for heavy multistory buildings where the above-ground laying of lines is excluded.

The high cost of two-stage utilidors built from prefabricated reinforced concrete sections is compensated for by convenience of operation and maintenance of the utilities and maintenance of the frozen state of the ground in the foundations of adjacent buildings. The main two-stage utilidors are built along the center-lines of streets with a slope allowing for drainage of accidental water out of the built-up area (Figure 4C).

The entries of the engineering utility lines into the buildings are made through single-stage "utilidettes." Within the buildings all the utility lines are suspended in air-cooled air spaces against the plinth wall of the floor. In order to shorten the overall length of the underground utilidors and the number of entries into the building, the latter are blocked in "chains" of four to six buildings with one common utilidette input.

Observations over many years by the Noril'sk Permafrost Laboratory of the temperature regime of the soil in the vicinity of the basic two-stage municipal utilidors have demonstrated significant reductions in their mean annual temperature (Figure 4B), which is explained by the adequate air exchange in the ducts, constant snow removal in the adjacent streets, and the prevention of penetration of surface water into the ground.

PERMAFROST CONDITIONS AND EXPERIENCE IN CONSTRUCTING FOUNDATIONS IN VORKUTA

V. N. YEROSHENKO Northern Division of the Scientific Research Institute of Foundations and Underground Structures, Vorkuta

The city of Vorkuta is located in far northeastern Europe beyond the Arctic Circle in the southern part of the permafrost region. Therefore, the permafrost conditions are distinguished by a highly spotty nature and complexity. This pertains primarily to the nature of permafrost distribution. The permafrost table has a complex configuration; its depth varies from 0.5-2.0 m to 10 m and more. As a result of this, in some sections the permafrost merges with the seasonally thawing layer with freezing of it; in other places it is quite deep (to 10 m and more). In a number of cases, stratified permafrost is found, and it mixes with the thawed interlayers according to depth.

The distribution of permafrost in plan is simple. Through-taliks are observed under river channels and large lakes and also within the built-up areas. A significant area is occupied by sections with a thick layer of frozen ground where the upper boundary of the permafrost occurs at great depths.

Within the city, such thawed ground occupies up to 80 percent of the area; in the unbuilt-up part of the Vorkuta region, the thawed ground is much less widespread (about 50 percent).

A characteristic feature of the permafrost in Vorkuta is the high negative temperature at the depth of the zero annual amplitude, causing a plastically-frozen state of the clay soil. Here, the range of variation of the temperature is quite large--from 0°C to -1.5°C, and in some cases even to -2.5°C. It is possible to consider -0.2° to -1.0° characteristic temperatures. The low temperatures (-0.6°C to -1.5°C) are characterized by permafrost, which merges with the seasonal-thawing layer during its freezing; the high temperatures (0°C to -0.3°C) are characteristic of deep-seated permafrost.

The depth of seasonal freezing and thawing of the ground in Vorkuta varies from 0.5 m to 2-3 m. In the places with deep permafrost, the thickness of the seasonal frozen layer is especially great--2-3 m, and in a severe winter, to 4-5 m. As a result, in such locations pereletoks 5-7 m thick are frequently formed.

For the upper part of the geological section of the Vorkuta region, the most characteristic are the clay soils and the least characteristic are the silty fine sands that are frequently underlain with gravelly-pebbly deposits. The soil in this region has high specific gravity: The thawed soil is on the average 2.0 to 2.4 tonnes/m^3; the permafrost runs from 1.2 to 2.3 tonnes/m^3, depending on the ice content. It is characterized by a high content of coarse inclusions (up to 40 percent) and also the presence of boulders up to 1-2 m across. The thawed clay soil is, as a rule, thixotropic.

The bedrock in the Vorkuta region is represented by shattered rock (frequently limestone) covered in the majority of cases by a layer of residual soil up to several meters thick. The residual soil contains a large amount of silt sizes.

The joint existence of solidly frozen, plastically frozen, and thawed ground sometimes in the same building site creates especially complicated and unfavorable conditions for construction.

The first stage was wood construction. Already by the 1940's builders began to convert to stone construction, and in 1960 the erection of large-panel multistory buildings began. Now primarily 5-story large-panel buildings are built, and preparations are under way to convert to 9- and 12-story residential construction.

The industrial buildings are built primarily of brick, although large-panel structural elements are used along with them.

Considering the complexity and spottiness of the permafrost conditions, as the foundations of the buildings and structures in Vorkuta it is necessary to use both thawed ground and permafrost (plastically and solidly frozen). If local islands of permafrost or pereletok are encountered in built-up sections with thawed ground, then preconstruction electric thawing of the frozen ground is used. This type of thawing is also carried out when it is impossible to insure the frozen state of the soil in the foundation during its operation (for example, under buildings with large heat release, with wet processes, with danger of thermal effects from surrounding objects, and so on).

In sections with shallow (10-12 m) bedrock, foundations of the buildings are supported on rock or the residual soil of this rock. For the most important structures, the rock is used as the foundation and with deeper bedrock. For example, the especially important industrial projects that do not permit deformations (such as the towers of mine shafts) are supported on rock, even if it occurs at a great depth (60-70 m).

For residential (especially large-panel) buildings, thawed or artificially thawed ground is used as the foundation. Sociocultural and domestic projects are constructed both on thawed or artificially thawed ground and on permafrost with maintenance of the frozen state during operation. The foundation for surface complexes of mines are permafrost with maintenance of the frozen state during operation, and in rare cases artificially thawed ground is used.

The industrial construction projects and also other projects for industrial purposes with large heat release and with wet processes are erected on thawed or artificially thawed ground.

For buildings and structures, foundations of various types are used: piles, columns, strip-footings, slabs, and deep foundations, and so on. The construction of one and two-story buildings is frequently carried out on fills. The pile foundations are considered the most promising and are finding broader and broader application. In Vorkuta they have been used since the 1940's. Until 1960, wooden posts were used with a low pile cap resting on a pad that was located at a depth of 2.5-4.0 m, that is, below the layer of seasonal freezing and thawing. Such foundations were constructed both in thawed and permafrozen soils.

Pile foundations in permafrost were used primarily in the places with deep permafrost (to 10-12 m) with the points of the piles in the permafrost to a depth of 0.5-1.0 m. In this case the soil of the foundation was kept in the frozen state during the operation of the structure.

Since 1961 reinforced-concrete piles have become widespread in Vorkuta. They are driven into the ground and founded on rock or weathered bedrock (with a depth to 10-12 m). The capping of these piles is in the form of grade beams at surface-level or somewhat lower. Granular deposits covering the bedrock can be both in the thawed and frozen states; in the latter case, thawing of them is permitted during the operation of the building. This type of foundation is used primarily for residential and public buildings.

An experimental check has been successfully passed and the new methods of driving the piles into the frozen ground (driving and drilling) and preconstruction thawing of the frozen ground of the foundations (electric thawing) developed by the Scientific Research Institute of Foundations and Underground Structures and its northern division (Vorkuta) have been used for the first time on an industrial scale.

The new methods of driving piles were used later also when building in permafrost in Salekhard and Labytnang, in Trans-Baikal and other areas where plastically frozen ground is widespread.

Recently, foundations made of driven friction piles in artificially thawed ground have been used in industrial construction.

In conclusion, let us note that the principle of using the supporting ground and the type of foundation have been selected in Vorkuta as a function of the specific permafrost conditions of the area, the type of building or structure, the nature of the load, and the required load capacity of the foundation. Here, it is necessary to consider the following:

1. In the case of shallow rock (to 10-12 m), piles are used founded on the rock or thawed residual soil. If the bearing bed of the ground is made up of frozen residual soil, then local (under each pile) thawing of it to the rock is carried out in advance. In the presence of a pereletok layer, the layer of frozen ground at the place where the pile is driven either is passed through by a leader hole not less than the size of the pile in diameter or it is thawed.

2. In the case of deep (more than 10-15 m) permafrost or in the presence of through-taliks, all forms of foundations are used for which thawed ground can serve. In the presence of pereletok, pile foundations made by driving friction piles are used that cut through the pereletok layer and fetch up in the thawed ground. Here, the pereletok layers are either penetrated in advance by leader holes or they are locally thawed.

3. With shallow (to 7-9 m) permafrost, either pile foundations made of driven or drilled piles are used or columnar foundations based on permafrost. In this case the supporting soil is kept in the frozen state during the entire operation of the building.

4. If within the limits of a single construction site, the permafrost table occurs at different depths, then leveling of it by electric thawing of the frozen ground to a depth to 15 m is used. The bearing soil for the foundations is artificially thawed and naturally thawed; foundations of various types are used.

PERMAFROST CONDITIONS AND THE CONSTRUCTION OF FOUNDATIONS IN THE VICINITY OF MIRNYY

I. YE. GUR'YANOV Scientific Research Institute of Foundations and Underground Structures

The city of Mirnyy is located in western Yakutia in the transition zone from low temperature to high temperature permafrost. The climate of the region is sharply continental; the mean annual air temperature is, according to the data from many years of observations, -7.1°. The annual amount of precipitation is on the average 330 mm. The height of the snow cover does not exceed 50 cm.

The lithologic section of the site of the study in general features is the following. Quaternary deposits--suglinoks of average ice-content and sandy-gravel soil--are encountered primarily within the 1.5-3.0 m layer of seasonal thawing. The rocks of the lower Paleozoic, represented by strata of limestones, dolomites, sandstones, marls, and clays, occur deeper. The bedrock deposits are nonuniform and the facies are not persistent. The thickness of the individual beds does not exceed 1-2 m. The rocks--limestones, dolomites, sandstones--are broken into individual blocks along the weathered joints after thawing. The clays and weathered marls have a layered-lattice cryogenous texture and with a moisture content of more than 13 percent during thaw-settlement. The total volume of the clay interlayers is less than half the total thickness of the section with a depth of more than 10 m. The ice content of the soil with respect to weight varies from 10 to 40 percent. In the northern part of the municipal site, Jurassic deposits represented by sand and supes with gravelly-pebbly material have been found.

A general analysis of the permafrost conditions of the site of the city was first carried out by A. I. Yefimov.[1] According to his data, in the area of Mirnyy it is possible to identify three sections: the low-temperature section of floodplains and the northern slope of the Ierelyakh River valley with a temperature at a depth of 15 m of -2.7° to -3.5°; the medium-temperature section, which is the basic area of the city, the slopes of the southeastern and southwestern exposures and a dry watershed with a rock temperature of -1.5° to -2.5°, and the high-temperature section, which is the northern part of the city and the watershed with a ground temperature of -1.3° and higher.

A. I. Yefimov and Ye. B. Belopukhova[2] have established that for the vicinity of Mirnyy the aspect of the slopes and the lithologic composition of the rock to a depth of 15 m have little effect on the magnitude of the mean annual temperature. At the same time, the conditions of surface runoff and the moisture regime of the layers of ground closest to the surface change the temperature at a depth of 15 m by 0.5°-1.5°.

A. I. Yefimov indicated the possibility of significant variations of the permafrost conditions on the site of the city under the effect of the economic activity of man. This was confirmed during the process of subsequent development of the area.

Since 1956, the Mirnyy region has been a constant construction site. Considering that during thawing of the Mirnyy ground under load significant nonuniform settlement is possible, the principle of maintaining the frozen condition of the foundation soil has been taken as the basic one in civil and industrial construction.

Civil construction is represented by class II buildings with respect to duration up to five stories high with the supporting walls erected from light concrete blocks. Under the buildings air spaces were constructed from 0.5 m to 2 m high. Other methods of thermal insulation of the foundations of the buildings used in Mirnyy have not been widely used.

During the first years of construction, the major buildings were erected on pad-and-post foundations with an allowable pressure at the pad corresponding to the temperature and nature of the ground. A significant drawback of foundations of this type is the labor and high cost of digging the foundation pit in the winter, thawing the ground when doing summer work and also the possibility of frost heaving of the foundation in the case of untimely loading. With the mastery of the technological process of cable percussion drilling by the construction organizations, post foundations were completely replaced by piling. The reinforced concrete piles up to 14 m long have been installed in predrilled holes backfilled with sand-clay slurry of given composition. The capping-beams and the frames of civic buildings are of poured reinforced concrete.

The general-purpose production buildings have been constructed on piles with an air space. One of the ore-dressing plants has been erected on post-and-pile foundations with the construction of such an air space. Garages and buildings with heavy loads are erected with the assumption of partial thawing of the foundation ground. These buildings are associated with places having soils of low compressibility.

A technological peculiarity of foundation construction in Mirnyy is the preliminary construction of a gravel fill to a height of 1.0-1.5 m, where the construction plant will operate. On completion of the pile operations, the floor of the future air space sloping toward the outside of the building is concreted. The construction of Mirnyy has advanced to the northeast. The permafrost regime also changes in the same direction.

By studies made in 1957, talik sections to a depth of 5 m in the central part of the present city between Lenin and Kosomol'skaya streets are associated with sandy-gravel deposits. In 1958 the area of the taliks decreased significantly, and in 1959, with the opening up of the area, complete freezing of them was noted.

In 1963 the talik sections in the northern part of the city were mapped. The Jurassic sand in the depth range of 4-8 m turned out to be in a thawed-saturated state. The southern end of the talik located lower down with respect to the topography was interrupted by a thick layer of clay soil, which disrupted the percolation of the groundwater in the talik.

The cause of the formation of the taliks was apparently a forest fire and lumbering that had destroyed the forest and the plant cover and caused intense heating of the ground during the warm period and also the presence of highly permeable sandy soil in which surface water accumulated in the absence of runoff.

During the 4 yr of development of the region, the boundary of the taliks advanced in the southeastern direction 300-700 m. In the southern part of the talik zone after its boundary shifted, in the vicinity of the present gorispolkom [city executive committee] building, the insular talik that froze at the end of 1967 remained for 2 yr. Then this section was cooled so intensely that it became possible to increase the height of the existing building without changing the foundations.

With organization of the surface drainage, the abundance of water in the taliks decreased, and their accumulating capacity decreased. After freezing of the taliks, rapid cooling of the ground began and the permafrost conditions on the site changed completely.

Observation of ground temperature in Mirnyy over a number of years indicates the general cooling of the high-temperature zone. For example, the complete destruction of the vegetation and partially the soil, the delay in construction by more than a year after installing the pile foundations--all led to significant cooling of the ground in a section about 500 m long along Leningradskaya Street. In a period of 3 yr, the temperature drop at the depth of the zero annual amplitude was 0.5°.

In the northern part of the city, the talik zone cooled from the west and east under the effect of sections of the finished construction. The displacement of the isotherms at the depth of the zero annual amplitude is on the average about 50 m a year. The most intense cooling of the soil was encountered at the site of the hospital complex where the shift of the isotherms in the easterly direction exceeded 100 m a year. The mean magnitude of the horizontal temperature gradients at a depth of 15 m was 0.5° per 100 m, and the maximum was near the intersection of Lenin and Leningradskaya streets; it reached 1° per 100 m.

The development of the area accompanied by the removal of the soil-vegetation layer activated the processes of heat exchange at the surface; the amplitudes of the seasonal temperature variations of the ground at a depth from 1 m to 4 m increased by comparison with the sections of undisturbed natural cover by up to 10 times. The increase in annual temperature amplitudes and cooling of the ground correspond to a sharp decrease in the magnitude and a change in sign of the vertical geothermal gradient.

In the sections with retained or recently removed soil-plant cover, the geothermal gradient of the permafrost layer reaches 0.5° to +0.7° in the first 10 m (Figure 1, holes 17,23). This section, located primarily north of Lenin Street, is the zone of prospective cooling of the soil. The zero geothermal gradient characterizes the sections with stabilized thermal regime where the temperature at a depth of 15 m is -1.3°C (hole 29) or the sections with variable conditions. The sections of active cooling of the ground in various stages are characterized by the negative geothermal gradient reaching -2.0° per 10 m (holes 2,47).

The depth range of 15-25 m is the zone of inflection of the gradient lines. Below this zone, the magnitude of the geothermal gradient decreases, but its sign is retained. At a depth of 70 m to 90 m, the gradient lines of different signs merge, undergoing a second inflection, and the geothermal gradient becomes negative for the entire area.

Let us consider the temperature regime of the permafrost using the concept of the geothermal level of heat exchange introduced by P. F. Shvetsov[3] as the "index of dynamic equilibrium in the thermodynamic interaction of the Earth's crust with the atmosphere expressed, in particular, by the ground temperature at the bottom of the layer with its annual fluctuations."

The superposition of the seasonal thermal waves differing according to the surface conditions changes the geothermal level of the annual heat exchange correspondingly. The axes of the annual temperature fluctuations with respect to depth (the mean annual temperature lines) shift continuously, as a result of which the vertical

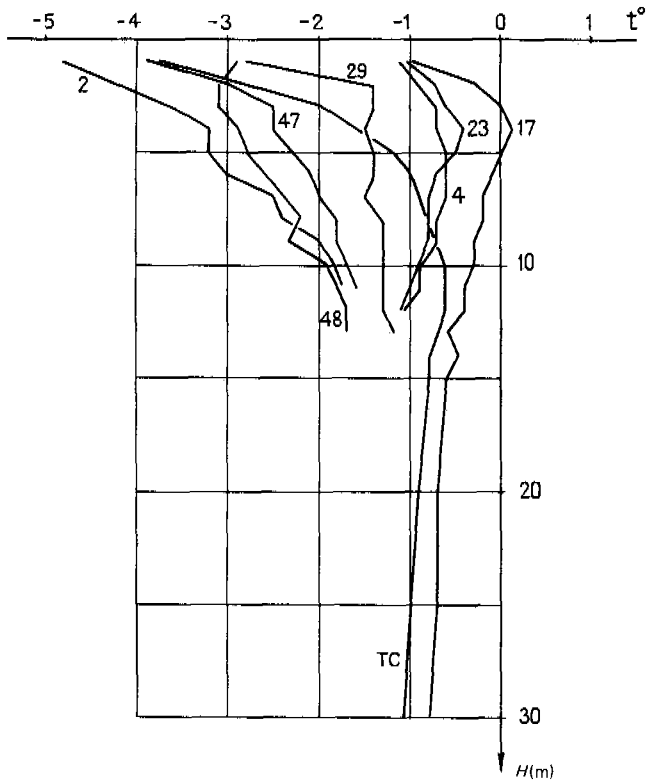

FIGURE 1 Variation of the mean annual ground temperature with respect to depth.

geothermal gradient within the depth of the seasonal temperature fluctuations varies. A characteristic feature of the site is the asymptotic approximation with depth of all of the gradient lines independently of their slope to some common position characterizing the mean geothermal level of heat exchange over many years for the given site.

The thermally active zone located below the depth of the seasonal temperature fluctuations serves as the area of accumulation of the energy of the seasonal thermal waves. This process lasts many years and with respect to duration exceeds any normative operating and maintenance periods of the structures. Therefore, the mean geothermal level of heat exchange over many years characterized by the ground temperature below the depth of the zero annual amplitudes can serve as one of the boundary conditions when determining the direction of variation of the ground temperature.

The given characteristics of the permafrost conditions of the site of Mirnyy, according to the classification of G. V. Porkhayev and V. K. Shchelokov[4] must be considered to be specific inasmuch as they are caused by the hydrogeological conditions on the watershed. This is also indicated by the distribution curves of the mean annual ground temperature with respect to depth, which in individual sections exhibit anomalous deviations from the geothermal level of heat exchange over many years caused by local conditions.

Considering the dynamics of the permafrost conditions in the city, when determining the bearing capacity of pile foundations, the temperatures corresponding to stable conditions of the temperature regime of the soil or equality of the annual geothermal levels of heat exchange and that over many years with a zero mean annual temperature gradient to a depth of 15 m are taken as the calculated values. The stabilization of the annual thermal cycles in the ground correspond to the defined law of seasonal temperature variations with depth connected uniquely with the ground temperature at the depth of the zero annual amplitudes. Thus, the ground temperature at a depth of 15 m will become the initial characteristic of the calculated temperature distribution of the foundation.

S. S. Vyalov[5] established the following relation for the adfreeze stress as a function of temperature:

$$R = R_0 + a\sqrt{|t|} \qquad (1)$$

indicating that under the given conditions [type of soil and its water content] characterized by the parameter a, the adfreeze stress R is a direct function of the temperature t [R_0 is the adfreeze stress at a particular temperature]. The seasonal temperature variations of the ground along the piles affect the adfreeze stress in accordance with the law of (1), varying the bearing capacity of the pile. The calculated bearing capacity defined by the minimum area of the adfreeze force diagram along the length of the piles must correspond to the minimum total negative temperatures along the operating length of the pile--the minimum area of the temperature diagram.

The seasonal temperature variations of the ground are defined according to Fourier theory by the function:

$$t_z = t + te^{-bz}\cos(c\tau - bz), \qquad (2)$$

where z is the depth reckoned from the bottom of the layer of maximum seasonal thawing, m; t_z is the ground temperature at the depth z, °C; $b = \sqrt{\pi/\alpha_M \tau_Y}$; $c = 2\pi/\tau_Y$; α_M is the coefficient of thermal diffusivity of the permafrost, m^2/h; τ is the time from reaching the maximum depth of seasonal thawing, h; $\tau_Y = 8{,}760$ h ≈ 1 yr [t is the mean annual temperature at the depth of zero annual amplitude].

Beginning with the investigation of the points of the piles located in the zone of greatest stable temperatures, let us integrate Equation (2) with respect to the operating length of the pile l, and then let us differentiate with respect to time in order to find the period of the minimum area of the temperature diagram.[6]

We obtain the equation:

$$\tau = \frac{\tau_Y}{2\pi}\left(\arctan\frac{e^{bl} - \cos bl}{\sin bl} - \frac{\pi}{4}\right), \qquad (3)$$

giving the relationship between the time of minimum bearing capacity of the pile on its operating length.

Calculations made by Formula (3) with the mean coefficient of thermal diffusivity of the Mirnyy soil $\alpha = 0.0043\ m^2/h$ agree acceptably with the times of the minima found by the actual temperatures (Figure 2). The period of minimum bearing capacity of the pile lasts about a month from the time of complete freezing of the seasonal layer. The time interval from the middle of November to the end of December encompasses all the "minimums" in the different sections and, in view of the small variations in bearing capacity of the pile inside this period, it is calculated for the entire area of Mirnyy.

The calculated bearing capacity of the piles during the actual minimum is 20-40 percent less than the carrying capacity determined at the time of the beginning of freezing of the seasonally thawed layer. From Formula (3), in particular, it follows that the inequality $e^{bl} - \cos bl \geq \sin bl$ becomes an equality when $l = 0$, and in the remaining cases it is strict, characterizing the lag of the minimum bearing capacity behind the time of the maximum depth of seasonal thawing and the beginning of freezing of the seasonally thawed layer.

For the known law of annual temperature fluctuations and the fixed time of determining the bearing capacity, the temperatures at the depths of the zero annual amplitudes are the initial values uniquely determining the bearing capacity of the piles. The values of the calculated bearing capacity of the lateral surface of

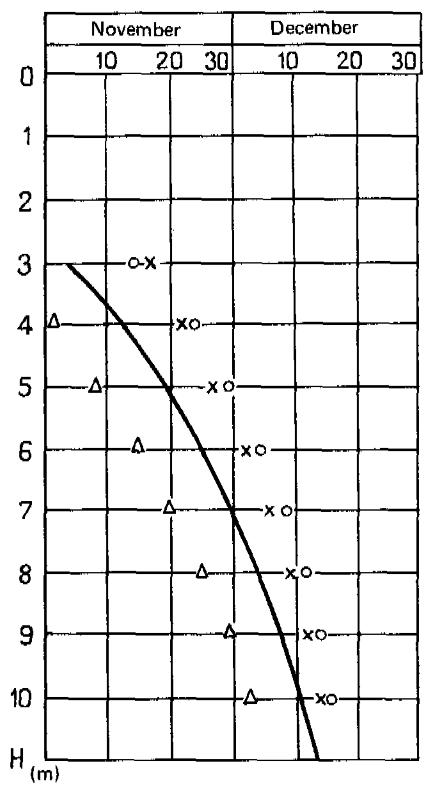

FIGURE 2 Time of the minimum bearing capacity of the piles as a function of depth: curve--theoretical for $\alpha_M = 0.0043\ m^2/h$; points according to actual temperatures: ∇--hole 37, x-hole 35, O--hole 1.

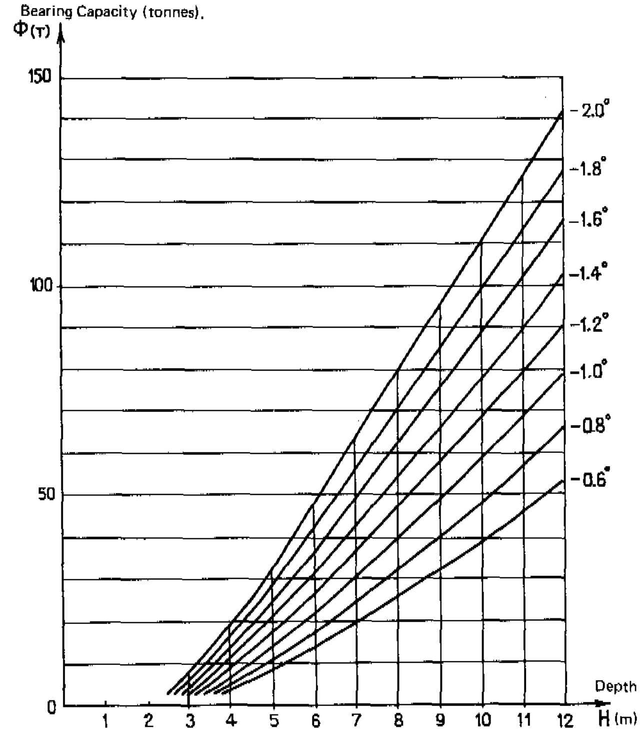

FIGURE 3 Calculated bearing capacity ϕ of the lateral surface of pile with a perimeter of 1 m as a function of depth and the temperature at the depth of zero annual amplitude, *H*.

a pile with a 1-m perimeter if used in Mirnyy with a [soil of] uniformity coefficient of unity appear in Figure 3.

In the calculations of the bearing capacity of pile foundations in the Mirnyy area, the reserve [capacity] connected with the general cooling of the high-temperature zone is neglected. It is true that the additional cooling of the ground by comparison with the perennial geothermal level is still insignificant. There are also cases of a *rise* in ground temperature. The general direction of the dynamics of permafrost conditions is connected with equalization of the ground temperature regime.

Nevertheless, the fact that the high-temperature ground in Mirnyy is cooling offers the possibility to the builders of further acceleration of the process of lowering the temperature of footings of future buildings by advance arrangements.

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EVALUATION OF THE VARIATIONS OF THE TEMPERATURE REGIME OF PERMAFROST IN A BUILT-UP AREA

G. V. PORKHAYEV AND V. K. SHCHELOKOV Scientific Research Institute of Foundations and Underground Structures

The disturbance of natural conditions when building up an area causes changes in the temperature regime of the permafrost. These variations are divided into general and local ones: The former are caused by providing the area with public services and amenities and the latter by the local sources of heat and cooling.

The planning and provision of public services and amenities in an area are accompanied by the removal and packing and, in some cases, the accumulation of snow, a change in the vegetative cover, the properties of the underlying surface, the soil moisture and the groundwater regime, and the thermophysical characteristics of the soil. As a result, the ground temperature and the depth of seasonal thawing or freezing vary, and in individual cases perennial thawing of the frozen series or perennial freezing of the ground in the talik sections occurs. All of this is characterized by variations in the mean annual ground temperature.

The mean annual ground temperature permits determination of the values of the temperature at any point in time in the zone of seasonal temperature fluctuations and evaluation of the directions of the thermal processes in the ground and also the nature of occurrence and distribution of the permafrost before and after construction. Under other equal conditions, it also characterizes the depth of seasonal freezing-and-thawing of the ground.

If under natural conditions the mean annual temperature on the ground $t_0^n < 0$ (merging permafrost), and after provision of services and amenities it has changed $t_0^c < 0$, then the permafrozen state of the ground is retained. If $t_0^c < t_0^n$, then the ground temperature drops and the depth of seasonal thawing decreases; if $t_0^c > t_0^n$, then the direction of the process is the opposite. If $t_0^c > 0$, then perennial thawing of the ground takes place; instead of seasonal thawing, seasonal freezing is set up.

In the case where under natural conditions $t_0^n \geq 0$ (the talik sections or sections with nonmerging permafrost) and after provision of services and amenities to the territory $t_0^c > 0$, then either the thawed state of the ground or the nonmerging permafrost are retained. If $t_0^c > t_0^n$, then the ground temperature increases and the depth of seasonal freezing decreases. If $t_0^c < t_0^n$, then the ground temperature drops, and the depths of seasonal freezing increases. If $t_0^c < 0$, then permafrost occurs; instead of seasonal freezing, seasonal thawing occurs.

The mean annual ground temperature varies as a result of the heat-exchange variations both during the winter and the summer. Here, in certain parts of the permafrost region, the variations in heat exchange in the winter have predominant significance and, in others, in the summer. In the northern and central zones, the snow cover has a great effect on the heat exchange. Building up of the area leading to compaction and removal of the snow cover promotes a significant decrease in the ground temperature. The destruction of the plant life, including the moss-peat cover has little effect, and the variation of the heat-balance components as a result of the construction of artificial coverings and variation of the ground composition with vertical leveling has an insignificant effect in general.

The effect of the snow cover on the heat-exchange conditions in the various regions is felt differently. In the northern regions adjacent to the arctic seacoast, as a result of strong winds and high relative humidity of the air, under natural conditions the snow cover has high density, and, as a rule, low height. Therefore, the removal and compaction of the snow cover do not cause a great reduction in the mean annual ground temperature. As we move to the south and center of the continent, the mean winter wind velocity and air relative humidity drop. The snow density decreases and its height increases. The northwestern part of Siberia, where the greatest depth of the snow cover of medium density is observed, constitutes an exception. The removal and compaction of snow here leads to great reduction in the mean annual ground temperature. Somewhat greater reduction is observed only in the northeastern part of central Yakutia (the region of Verkhoyansk, Lymyakon), where the snow cover is less deep, but it is looser and there are very low mean annual air temperatures.

In the southern regions with an insignificant snow cover, the variation in heat exchange during the summer predominates. The destruction of the moss-peat cover and the construction of artificial coverings and vertical leveling that changes the conditions of percolation of moisture into the ground can in some regions lead to an increase in the mean annual ground temperature, in spite of the removal or compaction of the snow in the winter.

As the total positive air temperatures change, and with respect to other attributes, the entire permafrost region is divided into five sectors (Table 1). The thawing index increased from sectors I to V. The freezing index is distributed nonuniformly. Thus, for sectors I and IV, the freezing indexes are approximately equal; they differ little for sectors II and V. At the same time, the ratio of the indexes decreases uniformly from sectors I to V (these ratios are not included in Table 1).

For average conditions in the sectors by our procedure,[1] the values of the mean annual ground temperatures were determined under natural conditions. A comparison of these data with the natural observation data demonstrated good convergence of the results. Thus, for the vicinity of Tiksi, we have by calculation a mean annual ground temperature of -11.4°C, and by the observation results, -11.6°C; for Anadyr', we have -4.7°C and -4.5°C, respectively; for Yakutsk, we have -3.1°C and -3.0°C; for Vorkuta, -0.2°C and -0.2°C; for Igarka, -0.7°C and -0.8°C; and for Skovorodino, -0.5°C and -0.6°C. The mean annual ground temperatures were also determined by this procedure in the absence of a snow cover, compaction of the snow, distruction of the moss-peat cover, the construction of artificial cover, and variation in ground composition. As a result the mean annual ground temperature for the changed conditions will be determined by the formula:

$$t_0^c = t_0^n + \Delta t_c^r + \Delta t_c^c + \Delta t_p + \Delta t_{sat} + \Delta t_g, \quad (1)$$

where Δt_c^r is the temperature correction for removal of the snow cover; Δt_c^c is the same as a result of compaction of the snow cover; Δt_p is the same as a result of destroying the moss-peat cover; Δt_{sat} is the same for the construction of artificial cover with respect to the saturated ground layer; and Δt_g is the same as a result of changing the ground composition with leveling.

The values of the temperature corrections are presented in Table 2. They are given for the plains regions and the average values of the humidity that will occur in the area supplied with amenities and services.[2] In Table 1 we have the mean thickness of the snow cover measured by the meteorological stations without considering the snow accumulation in the troughs or its removal from the highlands.

When determining Δt_c^r, systematic removal of the snow as it fell during the entire winter was assumed. Inasmuch as complete snow removal does not actually occur, the values of the given correction have been reduced somewhat for greater approximation to the actual conditions.

When determining Δt_c^c, the compaction of the snow during the winter by transport vehicles and pedestrians was taken into account. There are no data on the snow density in the built-up area. Therefore, the data for the density of snow compacted for highway and airport construction were used.[3]

The destruction of the grass cover at the site, as numerous observations indicate, changes the temperature regime of the soil in one direction or another very insignificantly. The situation is different with respect to moss-peat cover. On destruction of the moss-peat cover, there is a significant increase in the ground temperature. In the calculations, the moss-peat cover is taken into account as a layer of thermal insulation on the ground surface.

When determining Δt_{sat}, the artificial cover with respect to the saturated ground layer was taken into account just as thermal insulation at the ground surface. The calculations were performed for two limiting values of the thermal resistance of the thermal insulation $R_{sat} = 0.5$ $m^2 \cdot h \cdot deg/kcal$ and $R_{sat} = 1.5$ $m^2 \cdot h \cdot deg/kcal$. The values of the corrections are given only for three sectors, since for sectors I and II the magnitude of Δt is insignificant and can be dropped from consideration.

As is known, in sandy soil, under other equal conditions, the mean annual ground temperature is higher than in clayey soil. Therefore, with vertical leveling of the site, cutting away the upper clayey layer of ground with exposure of the sandy layer or filling to the surface of the sandy layer (without constructing an impervious artificial covering) leads to an increase in the mean annual temperature. This increase in temperature--Δt_g--as the calculations show does not exceed 1°C for the extreme values of the ratios of the thermal conductivity factors of thawed and frozen ground and variation of the humidity within broad limits. Therefore, for practical calculations it is possible to assume $\Delta t_g = 1°C$ with some reserve.[4]

Let us consider the local variations of the temperature regime of permafrost as a result of building up the area. As the observations show, in regions with low permafrost temperatures (for example, in central Yakutia, in Noril'sk, and in the populated areas of Magadan oblast), as a result of construction the general ground temperature reduction and local sources of heat in practice have no effect on the temperature regime of the built-up area as a whole.[5,6] Thawing or a reduction in ground temperature may take place under the buildings and structures; however, these variations almost never extend beyond the outlines of the buildings and structures.

In regions with high permafrost temperatures

TABLE 1

Climate Belts	Sectors	Thawing Index, deg. month	Mean Thickness of the Snow Cover, m	Mean Wind Velocity in the Winter, m/s	Mean Relative Air Humidity, %
Arctic	I Arctic Seacoast. Stations: Dickson Island, Kotel'nyy Island, Tiksi Base, Chelyuskin Cape	10	0.1-0.2	5-7	83-90
Subarctic	II Northeastern part of the USSR bounded by the watershed of the Kolyma and and Omoloy Rivers. Stations: Anadyr', Iliney, Provideniye Island	25	0.2-0.3	3-7	75-82
	III Bol'shezemel'skaya tundra and northwestern part of Siberia (Ob'-yenisey Interfluve and Yenisey River basin with the exception of the seacoast). Igarka, Tarko-Sale, Kochumden, Vorkuta	40	0.2-0.4	3-4	75-77
Temperate	IV Yakut ASSR with the exception of the seacoast regions. Stations: Yakutsk, Oymyakon, Vilyuysk, Aldan	50	0.2-0.3	1-3	68-73
	V Southern parts of the permafrost region. Stations: Chita, Skovorodino, Chegdomyn	60	0.1	1-4	65-75

(above -2°C), a more complicated picture is observed. Thus, in the vicinity of Chita, where the permafrost of great thickness has a temperature of -1°C and higher and is interrupted with talik sections, a drop in ground temperature also takes place in the built-up area, although thawing zones are formed under the buildings constructed without maintenance of the frozen state of the foundations.[7]

On the other hand, in the vicinity of Vorkuta, where high-temperature permafrost also occurs, as a result of the hydrogeological peculiarities and inefficient laying of the sanitary engineering lines, another picture of the formation of the permafrost conditions in the built-up area is seen. The groundwater heated in the taliks under the buildings and near the utility lines laid on the ground without ducts change the ground temperature over a large area. This was clearly demonstrated by observations of the industrial site of mine shaft 27 where in 2 yr the thawing zones around the utility lines promoted the formation of new runoff "channels" of the groundwater and caused lowering of the permafrost table over the entire site. When district heating networks were brought to the surface, restoration of the previous permafrost conditions began.

An analogous situation occurred in the first stage of the construction of the city of Noril'sk when the heating system in the city was laid in the ground in reinforced concrete unventilated ducts. This led to the formation of thawed zones near the heat ducts and disturbance of the stability of both the utility lines and nearby buildings.

At the present time in Noril'sk, tens of kilometers of utility lines are being successfully operated and maintained in buried ventilated ducts (utilidors) without disturbance of the temperature regime of the foundations of the buildings located nearby. Under buildings with ventilated air spaces, the ground temperature drops.

The examples presented indicate that if we permit convective heat transfer by the groundwater in the construction area, then independently of the permafrost temperature, its thickness or distribution in plan, intense thawing takes place, which leads to inadmissible deformations of the buildings and structures. With correct development of the area and observation of the requirements of the construction Norms, the thawing of the ground near the heat-generating structures is localized.

Fluctuations in the snow cover (Table 2) have the sharpest effect on the temperature regime of permafrost. Therefore, building up an area leading to compaction and removal of the snow cover promotes significant reduction in the permafrost temperature. In the sharply snow-drifted tundra regions, the total amount of snow in the populated areas with unorganized snow removal can turn out to be greater than in the adjacent area; however, the snow is distributed over the built-up area highly nonuniformly. In the streets, squares, and yards, it is compacted by pedestrians and transport vehicles or removed, and in the gardens and greenbelts it accumulates, forming snow drifts. Here it is possible to consider snow accumulations as local sources of heat.

Thus, the built-up area is a combination of different types of heat sources and sinks, as a rule, with narrowly localized thermal effect of each of them. Along with the formation of thawing zones under the buildings and structures transferring the heat directly to the ground, in other sections (under the streets, yards, squares, and so on) a ground temperature drop and the formation of permafrost can occur if taliks were present before construction. The latter occurs for any construction density. These variations of the temperature regime in individual sections must be considered when planning and designing the foundations of buildings and also for sanitary engineering systems, roads, and so on. For example, if the formation of thawing zones is permitted under buildings during their operation and maintenance, the fills and other road pavements and also the ground in which the pipelines are laid can be frozen. In this case, in the sections outside the outlines of the buildings (under the streets, yards, and so on), the mean annual heat fluxes that are 250-300 kcal/m^2•deg under natural conditions vary by several times.

In the southern part of the permafrost region, there are areas where building up the area leads to an increase in temperature and thawing of the ground in spite of the measures taken to keep the ground frozen. In these regions the thickness of the snow cover is insignificant. Its removal or compaction as a result of construction does not compensate for the increase in temperature caused by the destruction of the moss-peat cover. In addition, these regions are characterized usually by complicated hydrogeological conditions and even a relatively small increase in depth of thawing and percolation of precipitation into the soil as a result of the removal of vegetation or a change in the soil composition with leveling leads to intense convective heat-transfer processes in the soil. Thus, in the vicinity of the Bureynskaya depression, the permafrost is thawing in the developed area. The destruction of the plant cover and an increase in depth of thawing cause sharp changes in the hydrogeological conditions, and the movement of the groundwater leads to a decrease in thickness and an increase in temperature of the permafrost. Also, although conditions for maintaining the permafrost are creating under buildings with cold air spaces, thawing of the frozen footings can occur as a result of groundwater.

The entire permafrost region is divided into a number of sectors characterized by climatic peculiarities and the peculiarities of heat exchange in the underlying surface before and after construction. In order to determine the mean annual ground temperature that has changed as a result of construction, it is possible to calculate the temperature corrections for each sector resulting from the removal and compaction of snow, the destruction of the moss-peat cover, the construction of an artificial cover, and the change in soil composition during leveling.

In the majority of regions, as a result of construction, there is a general drop in ground temperature, and the local sources of heat do not in practice affect the built-up area as a whole if convective heat transfer by the groundwater is

TABLE 2 Values of the Temperature Corrections for Variation of the Snow Cover and Destruction of the Moss-Peat Cover and When Constructing an Artificial Cover with Respect to the Saturated Ground Layer

Sector	Mean Annual Air Temperature, °C				Δt_p, °C (Numerator); for Different Depth of Moss-Peat Cover under Natural Conditions, m (Denominator)	t_{sat}, °C for R_{sat} = 0.5 (Numerator) and for R_{sat} = 1.5 (Denominator)	
						Sandy Soil in All Versions	Clayey Soil including Silt
I	-10	0.1	-1.8	-0.6			
		0.2	-2.7	-0.9	1.0 - 1.3	--	--
	-14	0.1	-2.1	-1.0	0.15- 0.2		
		0.2	-3.4	-1.4			
II		0.2	-3.3	-1.0			
	-7	0.3	-4.2	-2.1	1.0 - 1.4	--	--
		0.2	-4.7	-2.0	0.15- 0.25		
	-13	0.3	-6.2	-4.1			
III		0.2	-3.5	-1.1			
	-5	0.4	-5.1	-2.9	1.3 - 1.6	0.2	0.2
		0.2	-4.6	-2.2	0.2 - 0.3	0.4	0.4
	-10	0.4	-7.4	-5.1			
IV		0.2	-4.2	-1.5			
	-6	0.3	-5.6	-2.3			
		0.2	-6.1	-2.3	1.5 - 2.0	0.3	0.2
	-10	0.3	-7.3	-4.2	0.3 - 0.5	0.5	0.4
		0.2	-7.3	-3.1			
	-16	0.3	-8.3	-4.5			
V	-2	0.1	-2.0	-1.5	2.0 - 3.0	0.3	0.2
	-5	0.1	-5.0	-3.0	0.3 - 0.5	0.6	0.5

not permitted. In some of the southern regions, building up of the area leads to a rise in temperature of the permafrost and thawing of the ground.

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GROUND ICE AS THE BEARING STRATUM FOR CONSTRUCTION

S. S. VYALOV Scientific Research Institute of Foundations and Underground Structures
V. V. DOKUCHAYEV AND D. P. SHEYNKMAN, LENZNIIEP
YE. I. GAYDAYENKO Geocryology Institute of the Siberian Department of the USSR Academy of Sciences
YU. M. GONCHAROV Branch of the Sevastopol' Instrument Making Institute

In permafrost areas sections are frequently encountered where the ground is surrounded by thick lenses, wedges, and integral masses of ground ice. The necessity for building on such sites requires the development of methods of designing the foundations under the special conditions. As is known, in contrast to permafrost, ground ice does not have a limit of strength but deforms continuously under the effect of additional pressure transmitted by the foundation. Therefore, the design of the foundations with respect to deformations on such bearing surfaces is determinable.

LAWS OF DEFORMATION OF ICE

The nature of deformation of ice depends on its temperature, structure, and magnitude of the load.[1] A true viscous shear flow occurs with low stresses, but only with single ice crystals if the load is applied parallel to the basal planes; the rate of the process is greatest in this case. If the load is applied perpendicular to the basal planes, the rate of the process is greatest. If the load is applied perpendicular to the basal planes, then bending and breaking of the crystal layers takes place; the deformations have an elastoplastic nature and take place with the least velocity. The deformation of polycrystalline ice of random orientation is intermediate with respect to these two extreme cases.

In the case of small loads, simultaneously with breakdown of the crystals, recrystallization takes place--a growth of less-stressed crystals. The process is steady state, and deformation takes place at an approximately constant rate. For comparatively large loads, the process of crystal breakdown prevails over recrystallization. A new structure is formed with linear orientation of the crystals along the direction of shear. Correspondingly, the deformation rate grows continuously. With large loads, this process is accompanied by the appearance of intercrystalline and intracrystalline microcracks which lower the resistance of the ice to loads.

The true yield point of the ice is absent. The ice begins to flow under any load; however, it is possible to isolate the stresses under which the nature of the deformations varies. The first of them is a stress τ_f, under which the flow rate is negligibly small from the point of view of engineering problems; the second is the stress separating the flow with constant ($\tau < \tau_s$) and continuously increasing ($\tau > \tau_s$) rates.

The flow laws of the ice can be described on the basis of the molecular-kinetic theory of Frenkel-Eyring, according to which the following relation exists between the minimum flow rate $\dot{\gamma}$, the shear stress τ, and the absolute temperature T (°K):

$$\dot{\gamma} = \dot{\gamma}^* \sinh \frac{\tau}{\tau^*} , \tag{1}$$

Where

$$\dot{\gamma}^* = 2\frac{kT}{h} X \exp(E_0/kT); \quad \tau^* = 2\frac{kT}{V} ,$$

where E_0 is the activation energy [of dislocation]; k is the Boltzmann constant; h is the Planck constant; V is the molecular volume; X is a molecule concentration function.

The correspondence of Formula (1) with experiment was demonstrated by H. Dillon and O.

Andersland.[2] Under comparatively large loads ($\tau/\tau^* \geq 1$), Formula (1) assumes the form $\dot{\gamma} = \dot{\gamma}^* \exp(\tau/\tau^*)$. For small loads it converts to the Newtonian coefficient of viscosity $\dot{\gamma} = \tau/\eta$, where $\eta = \dot{\gamma}^*/\tau^*$. It is possible to assume that this change in the process occurs when $\tau = \tau_s$.

For engineering calculations, it is admissible to use the empirical formula[3]

$$\dot{\gamma}^* = \frac{k_I}{1 + \Theta}\tau^n, \tag{2}$$

where Θ is the temperature in °C without considering the sign; k_I and n are the deformability parameters of the ice.

For the complex stress state, it is necessary to substitute the shear deformation rate intensity $\dot{\varepsilon}_i$ and the tangential stress intensity σ_i in place of $\dot{\gamma}$ and τ in Equation (2):

$$\varepsilon_i = \frac{k_I}{1 + \Theta}\sigma_i^{\,n} = \frac{\sigma_i}{\eta}, \tag{3}$$

Where

$$\eta = \frac{1 + \Theta}{k_I \sigma_i^{\,n - 1}}. \tag{4}$$

When approximating the experimental data by Formula (2) or (3), the curve for $\dot{\gamma} - \tau$ has an inflection point, which separates the regions of low, and relatively high, stresses. At this point, which can be considered as τ_s, the exponent n changes value.

In Figure 1 we have the data from compression testing of the polycrystalline ice with step loading ($\Theta = -8$°C)[1] and in Figure 2, the test data of a number of authors (for Θ = 1.5, -4, -10°C) reduced to a single graph by Dillon and Andersland.[2] From the figures, and also according to the data of other authors,[4] it follows that the parameter n has a value of 1.2-2.0 for small stresses ($\tau < \tau_s$) and 3-4 for large stresses ($\tau > \tau_s$).

If we begin with Formula (1), it is admissible to assume that the change in the process takes place when $\tau/\tau^* = 1$, that is, $\tau = \tau_s$. According to the date of Dillon and Andersland[2], when Θ = -1.5, -4, and -10°C, the value of τ^* is equal to 0.85, 1.2, and 1.7 kg/cm^2 (with corresponding conversions). When determining τ_s by varying the parameter n from Formula (2), we obtain somewhat larger values. From Figure 2 it follows that when Θ is from -1.5°C to -10°C, the value of τ_s is 1.5-4 kg/cm^2. According to the data of K. F. Voytkovskiy,[3] the values of τ_s when Θ = -1.2, -1.8, and -4°C are 1.6, 2, and 3 kg/cm^2, respectively.

It must be considered that these values are obtained for a certain period of time of operation of the load and that they correspond to sufficiently high relative values of the flow—on the order of 10^{-4}/min (Figure 2).

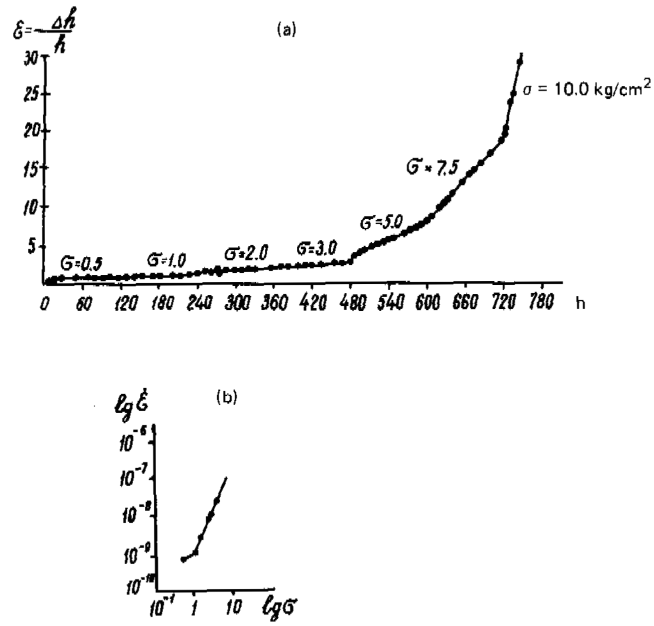


FIGURE 1 Testing polycrystalline ice in uniaxial compression ($\Theta = -8$°C). (a) creep curves; (b) curve of deformation rate (ε) as a function of stress (σ)

DEFORMATIONS OF PIER FOUNDATIONS ON GROUND ICE

When calculating the settlement of foundations, it is necessary to take into account the density and various forms of ground ice.

The relative compression of the ice caused by a change in volume of the pores as a result of an increase in pressure is defined as the decrease in its porosity.[5] Accordingly, the compaction of the ice comprising m layers (each Δh_i thick) is

$$s_c = \sum_{i=1}^{m} \frac{\Delta h_i n_i \bar{p}_i}{p_a + p_{bi} + p_i}, \tag{5}$$

where n is the porosity (the volume of the pores per unit volume of ice); p_a is the atmospheric pressure; p_{bi} is the pressure in the i-th layer of ice; $\bar{p}_i$ is the mean pressure from the structure in the same layer.

The settlement due to change of form is taken into account in the form of the plastoviscous flow of the ice at a constant rate under the effect of tangetial stresses. The progressive flow with a continuously increasing rate is not permitted by restricting the load on the foundation so that $\tau_{max} < \tau_s$. From relation (3), the relative deformation rate is determined,

$$\dot{\varepsilon}_z = \frac{k_I(\sigma_z - \bar{\sigma})\sigma_i^{n-1}}{2(1 + \Theta)}. \tag{6}$$

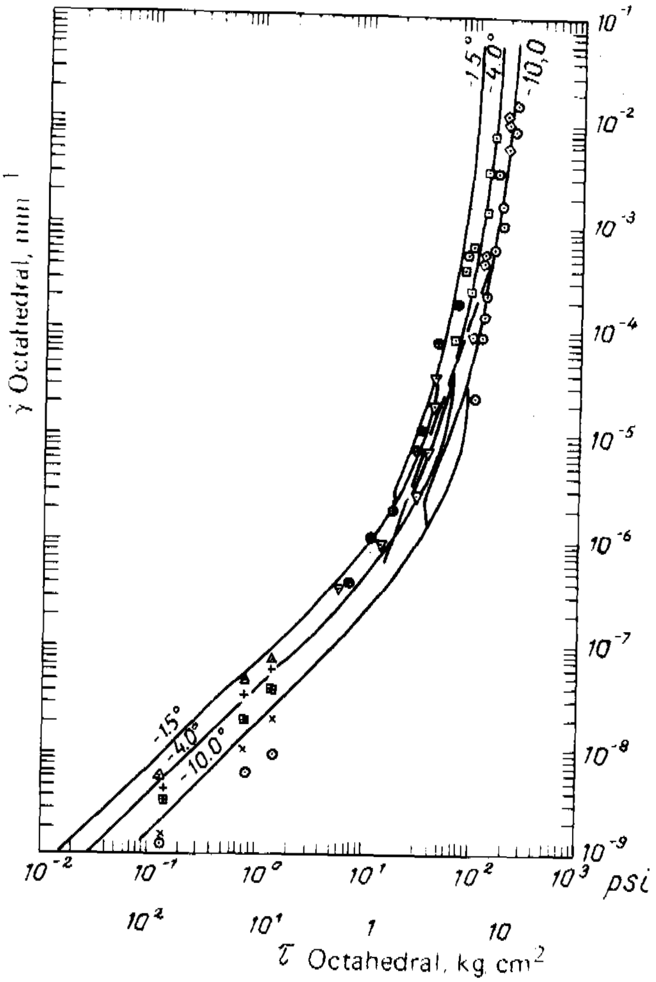


FIGURE 2 Relation between the flow rate and the stress according to the test data of Butkevich and Landauer, Glen, Khigash and Khelbruk, and Dillon (θ = -1.5, -4, -10°C).

This settlement of the layer of ice z thick in the calculated time t_p considering the temperature variation θ during the year t_Y is

$$s_f = \frac{1}{2} k_I \frac{t_p}{t_Y} \int_0^z \int_0^{t_Y} \frac{(\sigma_z - \bar{\sigma})\sigma_i^{n-1}}{1 + \theta(t)} dz dt. \tag{7}$$

Introducing the notation

$$\omega = \tfrac{1}{2} \int_0^{z/b} \frac{\sigma_z - \bar{\sigma}}{p} \left(\frac{\sigma_i}{p}\right)^{n-1} d\frac{z}{b}; \tag{8}$$

$$k'_\theta = \int_0^{t_Y} \frac{dt}{1 + \theta(t)} = k_\theta t_Y , \tag{9}$$

where p is the mean pressure at the base of the footing, and b is the width of the footing. Then we obtain

$$S_f = k_I k'_\theta t_p b^n \omega. \tag{10}$$

The calculation of the settlement reduces to determining the deformations of the rigid footings in the nonlinearly deformed bearing ground. From the solutions of Arutyunyan and Zaretskiy[6,7] obtained for punches in an infinite halfplane, it follows that the expression for the settlement of the uniform base is analogous to Equation (10) for θ = const, and it differs only by the method of determining ω. The numerical values of ω_f (for the middle of a flexible footing) and ω_r (for rigid) permit a sufficiently simple determination of the settlement.[8] However, this method has limited application, since only a punch is considered, the settlement is determined for a layer of infinite thickness, and the temperature of the ice is assumed constant.

For engineering calculations it is necessary to use the approximate procedure for which the settlement is calculated by Formula (10), that is, the nonlinear relation is maintained between the settlement and the load, but the values of ω are calculated (by Formula 8) for stresses of σ_z, $\bar{\sigma}$, σ_i calculated by the methods of the linear theory of elasticity; the values of k_θ are determined considering the temperature variation of the ice during the year.[5,8] The numerical values of ω are shown in Figure 3. A comparison of the values of ω for a flat punch with stresses determined by the methods of linear and nonlinear elasticity theory shows that the graphs in the Figure 3 give sufficient accuracy when calculating the settlement of a footing for engineering purposes.[8]

The pressure at the base of a footing is limited by the condition:

$$\tau_{max} < k\tau_s, \tag{11}$$

where τ_{max} is the maximum tangential stress in the ice layer; k is the factor of safety taking into account the conditions of operation and uniformity; τ_s is the tangential stress for which, in the given time period, progressive flow does not occur.

The determination of τ_{max} when installing a footing directly on ice presents no difficulties. Thus, the value of τ_{max} at the center of a circular flexible plate at a depth of 0.29 b is 0.288 p' and at the center of a flat base it is 0.245 p at a depth of 0.5 b; for the circular, absolutely rigid, punch it is 0.188 p at h = 0.71 b, correspondingly.[5,9]

To a depth approximately equal to the halfwidth of the footing, the maximum tangential stress increases, which implies the most intense increment of the settlement within this depth (see Figure 3). Therefore, between the footing and the ice it is expedient to place (or retain) an interlayer of soil. In this case the calcu-

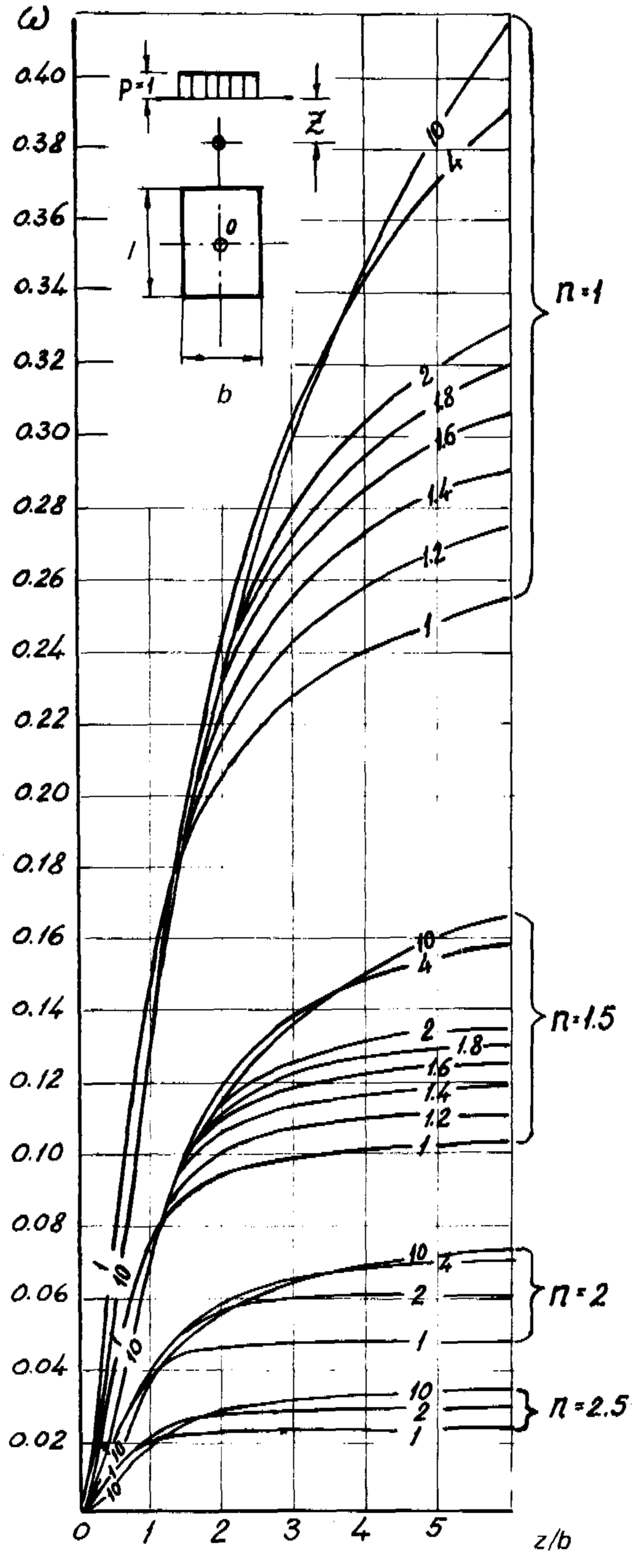


FIGURE 3 Values of the coefficients ω (the numbers on the curves denote the ratios *l/b*). [*n* is the parameter in Equation (2).]

lation of τ_{max} must be made by the principle of the two-layered base.

The analysis of the numerical values of τ_{max}[9] shows that in the presence of tangential stresses at the contact of the soil with the ice the largest values of τ_{max} occur in the ice at the contact of the layers; in the absence of tangential stresses at the contact, the maximum τ_{max} occurs in the ice and then at different depths depending on the thickness of the soil layer h_s. The values of τ_{max} for both cases are presented in Table 1; the numbers in the numerator refer to the first case; the numbers in the denominator refer to the second case.

From Table 1 it is obvious that the values of τ_{max} in the ice are the greatest when at its contact with the soil the tangential stresses are taken into account and also that in the presence of an intermediate ground layer under the footing the pressure [on the ice] can be significantly increased.

EXPERIMENTAL STUDIES OF THE PERFORMANCE OF THE ICE UNDER LOAD

The earliest experiments with respect to pressing punches into the ice were performed in Antarctica in order to study the laws of settlement for different loads and for variation of the ice structure.[1] The latest experiments (Yakutsk) were used to investigate the effect on the settlement of the ice of its structure and temperature, the size and depth of bearing the punch, and also the load.[10]

The experiments with circular rigid punches buried at various depths in the ice (for Θ = -2.3°C) demonstrated (Figure 4) that the law of deformation of the ice is identical in all cases; only the values of the steady-state deformation rates are different; they decrease as the punch becomes deeper (Table 2). From Figures 1 and 4, it is obvious that the law of deformation on pressing in the punches has the same nature as for uniaxial compression.

The experiments with ciruclar punches of different diameters [*d*] installed on the ice surface in holes with ice frozen in layers demonstrated that the rate of settlement of the ice is approximately proportional to the diameter of the punch (Table 3).

The studies of settlement under the load of multiple-veined ground ice and artificial ice frozen in layers revealed the effect of the ice structure on its deformation rate (Table 4).

The studies of the structural variations of the ice and the magnitude of the active deformation zone under the round punches demonstrated (Figure 5) that in the stressed zone there is breakdown and refinement of the crystals and the formation of a kind of kernel under the punch.[1] The size of the active deformation zone in which distortion of the layering of the ice was visually observed[10] was third in the vertical direc-

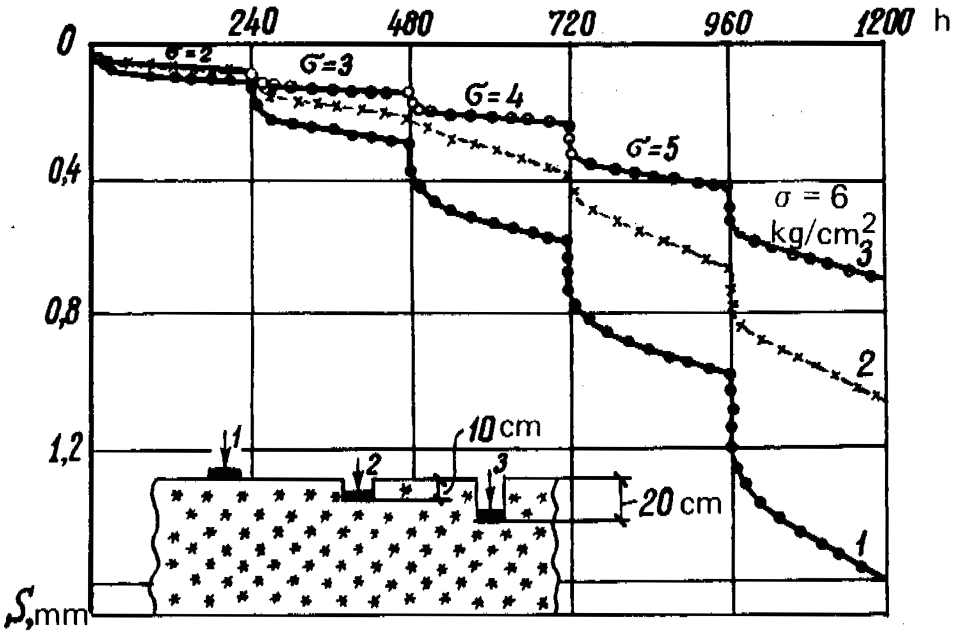


FIGURE 4 Ice deformation from loaded punches at different depths (θ = $-2.3°C$; d = 16 cm).

tion. This corresponds approximately to the zone of intensity formations under the round punch determined by calculation.[5] The initial uniform density, which is 0.910 g/cm^3, became sharply nonuniform after testing. In the layers under the punch, it increased somewhat and reached 0.924 g/cm^3, and in the zones beyond the kernel it dropped to 0.668 g/cm^3. The redistribution of the density was accompanied by a decrease in the dimensions of the air bubbles in the stressed zone and their concentration beyond it.

An analysis of the results of these studies and also the experimental data obtained by other authors shows that the settlement of the punches is given quite satisfactorily by Formula (10).

PILE FOUNDATIONS IN GROUND ICE

The study of the operation of piles frozen in the ice was first carried out in Igarka[11] using small models (d = 3.5 cm; l = 10 cm). Some results of these operations are presented in Figure 6a (here and hereafter the stress τ is the [adfreeze bond stress] on the surface area of the frozen pile). Experiments with large-scale

TABLE 1 Values of τ_{max}/p in the Ice under a Circular Footing with a Radius of r (for ν_I = 0.5 and ν_5 = 0.3)

	Values of τ_{max}/p in the ice for E_S/E_I							
	At the center				At the edge			
r/h_s	1	5	20	100	1	5	20	100
1	0.217	0.138	0.065	0.020	0.167	0.108	0.052	0.017
	0.184	0.109	0.054	0.019	0.141	0.087	0.046	0.017
0.5	0.115	0.062	0.025	0.007	0.093	0.052	0.022	0.006
	0.077	0.039	0.017	0.006	0.066	0.034	0.015	0.005
0.33	0.061	0.032	0.013	0.003	0.054	0.029	0.012	0.003
	0.038	0.019	0.008	0.002	0.035	0.018	0.007	0.002
0.25	0.037	0.019	0.008	0.002	0.034	0.018	0.007	0.002
	0.023	0.011	0.005	0.001	0.021	0.011	0.004	0.001

NOTE: ν is Poisson's ratio; E is Young's modulus, Subscript I refer to ice and S to soil; h_s is the thickness of the soil layer over the ice.

TABLE 2 Values of the Steady-State Velocities *ds/dt* for Pushing in Circular Punches (d = 16 cm; Θ = -2.3°C)

Depth of Burial of the Punch, cm	Values of *ds/dt* (mm/day) for a Load (kg/cm^2)				
	2	3	4	5	6
0	0.007	0.014	0.023	0.038	0.054
10	0.006	0.009	0.018	0.023	0.029
20	0.001	0.003	0.006	0.015	0.018

models of reinforced concrete piles (d = 16 cm, l = 32 m) were performed in Yakutsk.[10] Their results are presented in Figure 6b.

As is obvious, the adfreeze strength of the ice is low; even when τ = 0.5-1.0 kg/cm^2 (Θ = -2.3°C) and τ = 0.4 kg/cm^2 (Θ = -0.4°C), deformations develop with increasing rate. Moreover, one of the models of the piles remained under a pulling load of τ = 0.3 kg/cm^2 (Θ = -0.4°C) for 1,200 h without noticeable increase of the deformations, but then it suddenly pulled out.[11] What has been stated confirms that the true yield point for the ice is zero.

It is possible to recommend values of τ_f and p_f, considering them as the "norm" resistances R_s^N and R^N for designing piles for which the settlement of the piles will be negligibly small (Table 5).

In view of the small values of R_s^N and R^N, the freezing of the piles directly into the ice is inexpedient. It is recommended that holes be drilled in the ice of somewhat greater diameter than the pile cross section and that the piles be buried in these holes, filled in advance with soil slurry. In this case the bearing capacity of the piles increases, inasmuch as on the one hand the adfreeze stress of the piles to the ground is greater than to the ice, and on the other hand the calculated area of the lateral surface in contact with the ice increases. This is determined by the size of the hole.

The possibility of constructing pile foundations in ground containing ground ice was experimentally tested at Mokhsogollokh in the Yakut, ASSR. Six experimental piles 15 × 15 cm in cross section and 600 cm long were installed in drilled holes. They were tested for 3 yr (1968-1970). Three piles were installed in holes 395 mm in diameter and three in holes 250 mm in diameter. In addition, for some of the piles the holes were drilled 1 m below their ends, so that the effect on the settlement of the piles of a footing under the end made of packed rock debris could be evaluated. The holes were filled with various slurries made of water and fine-grained sand and sand-clay.

The results of testing all of the piles demonstrated that in the first stages of loading the settlement is insignificant and damps quickly. Thus, for example, the total magnitude of the settlement of the pile frozen into the ice after two stages of loading (15.5 and 20.5 tons) did not exceed 5 mm and was stabilized for 6-8 days. With an increase in loading to 25.5 tons, the settlement rate increased noticeably (to 0.01 mm/day), and it became nondamping. A further increase in the load to 30.5 tons led to a still more significant increase in the settlement rate (0.02 mm/day).

The field tests performed on the piles frozen in the ice confirmed that loading can always be selected that causes settlements of the piles in the design service life less than the admissible maximum.

The possibility of building on soil containing masses of ground ice is confirmed by construction practice. The buildings constructed on such

TABLE 3 Values of the Steady-State Deformation Rate *ds/dt* on Pressing in Circular Punches (Θ = -2.3°C)

Step Increasing Load, kg/cm^2	Values of *ds/dt* (mm/day) for a Punch Diameter, cm		
	3.2	16	32
2	0.003	0.007	0.016
3	0.005	0.014	0.032
4	0.006	0.023	0.054
5	0.008	0.038	--
6	0.016	--	--

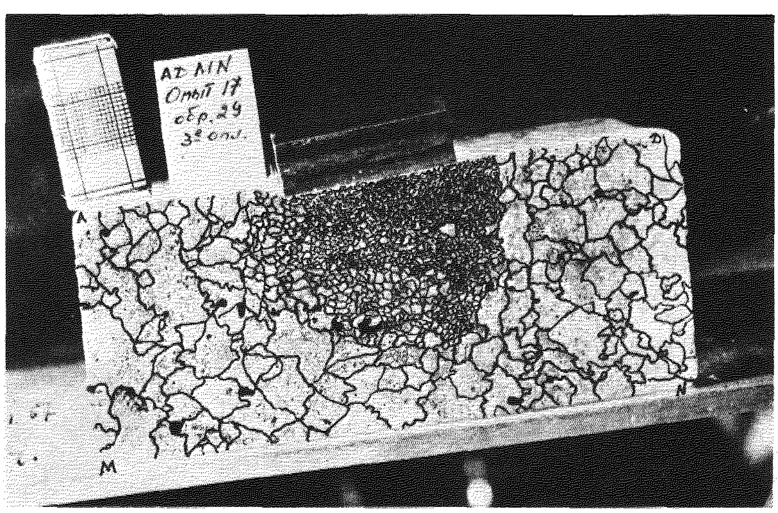
TABLE 4 Values of the Steady-State Deformation Rates *ds/dt* When Pressing in a Circular Punch Buried in the Ice to a Depth of 20 cm (d = 16 cm; Θ = -2.3°C)

Ice Characteristic	Values of *ds/dt* (mm/day) for a Load (kg/cm^2)								
	2	3	4	5	6	7	8	9	10
Layered	0.001	0.003	0.006	0.012	0.018	0.022	0.028	0.038	--
Multiple-veined	0.002	0.005	0.007	0.015	0.019	0.027	0.031	0.040	0.044

TABLE 5 "Norm" Resistances of the Ice to Shear along the Lateral Surfaces of the Pile R_s^N and Normal Pressure R^N

Form of Resistance, kg/cm^2	Temperature, °C					
	-1.0	-1.5	-2.0	-2.5	-3.0	-3.5
R_s^N	0.25	0.35	0.40	0.50	0.55	0.65
R^N	0.4	0.9	1.3	1.8	2.3	2.6

(a)



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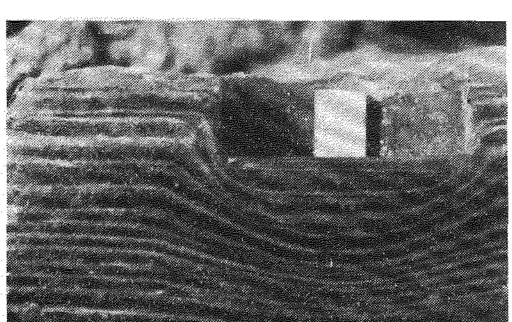

FIGURE 5 Formation of the "kernel" under the punch. (a) breakdown of the ice crystals and the formation of the packed zone [kernel] (Θ = -8°C; p = 15 kg/cm^2); δ is the deformation of the ground layers (Θ = -2.3°C; p = 25 kg/cm^2).

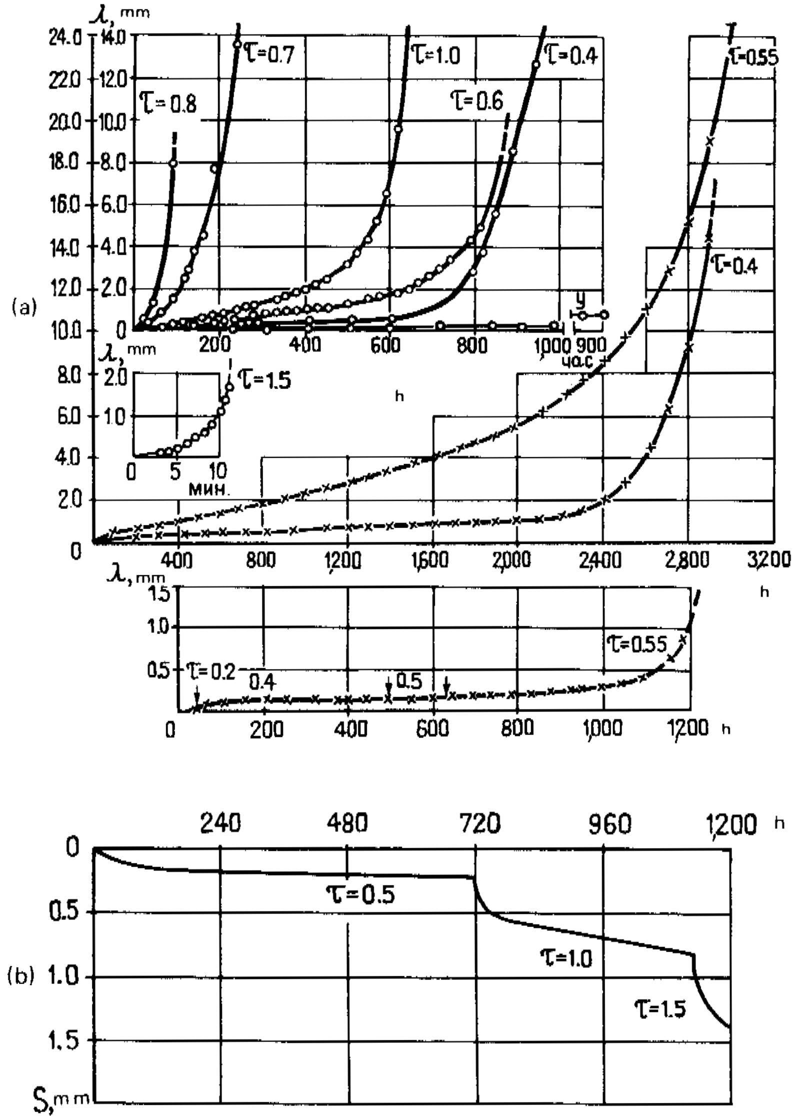

FIGURE 6 Test of model piles for extraction under the load τ, kg/cm^2. (a) wood piles 3.5 cm in diameter for Θ = -0.4°C; (b) reinforces concrete piles 16 cm in diameter at Θ = -2.3°C. [λ is pile movement.]

sites--the airport at Tiksi; residential buildings in Amderma, Cherskiy, and other settlements on the Arctic Sea Coast; the school and industrial buildings in Pokrovsk in the Yakut, ASSR, and so on--have performed normally for many years.

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SALINE PERMAFROST AS BEARING GROUND FOR CONSTRUCTION

YU. YA. VELLI, LENZNIIEP, A. A. KARPUNINA Yakutsk Branch of the Krasnoyarsk Institute for Design of Industrial Construction

The salinity of permafrost characterized by the presence of brine and salt crystals in it has a significant effect on its mechanical properties. The presence in frozen saline ground of ice crystals even visible to the eye does not always make it correct to consider it in the category of solidly frozen. Such ground can be in the plastically frozen state at a temperature appreciably below 0°C.

Saline permafrost is widespread on the arctic coast in the regions of the Primorskaya lowland and also in the territory of Yakut, ASSR--in the valleys of the Lena, Aldan, and Vilyuy rivers. The increased salinity of the soil arose as a result of multiple transgressions (regressions) of the sea and bringing of salt in by the rivers. The saline supesses and suglinoks with a silt content from 30 to 65 percent having average and high colloidal activity and also the saturated silty sands are most widespread. The salinity of the ground is distributed nonuniformly with respect to depth: In the surface layer it is 0.2-0.4 percent; then at a depth of 3-5 m it increases in Yakutsk to 0.6 percent; in the coastal arctic territories it increases to 1.5 percent of the weight of the dry soil. Sodium, magnesium, and calcium cations predominate in the salt. The salinity of the soil is frequently accompanied by increased ice content. The ice in the soil is in the form of thin seams 0.1-1.0 cm thick, sometimes in the form of lenses.

In order to determine the suitability of the saline permafrost as the bearing ground for buildings in Amderma and Yakutsk, many years of experiments have been performed with models under laboratory conditions (in caves) and with piles in construction test areas. Studies were made of physical and mechanical characteristics of the soil, including its freezing temperature, the equivalent cohesion, compressibility, and shear strength with respect to the lateral surfaces of the foundations.

The freezing temperature of saline soil was determined by the standard procedure: thermometric and thermoelectric methods with subsequent construction of graphs of the soil temperature variation with time during its cooling. The highest value in the section of the cooling curve with relatively stable temperatures was taken as the freezing point (t_f) of the soil.

The freezing point of saline soil depends primarily on the soil moisture content and salinity. The experimental data (Table 1) indicate that the granulometric composition of the soil with moisture content above 25 percent does not in practice affect the freezing temperature. With an increase in soil salinity and a decrease in its freezing point, the greatest intensity of reduction of t_f was observed in the less wet soil. Thus, with a soil moisture content of 20 percent with variation of the salinity from 0.4 percent to 1.2 percent, the freezing point dropped by 2.2°, while with the same variation in salinity of the soil but with 40 percent moisture content, the freezing point dropped only 1.2°.

Let us note that the nature and intensity of variation of the freezing point of saline soil as a function of moisture content and salinity were identical in the experiments performed at Amderma and at Yakutsk. In addition, the freezing point of the soil in Yakutsk was lower than the freezing point of the soil in Amderma by 0.5°-0.6°. This difference was retained in all the experiments. It is explained by differences in the qualitative composition of the salt in the soil in Yakutsk and the coastal deposits of Amderma. The predominance of potassium salts distinguished by a higher melting point of the eutectic mixture in the Amderma soil leads to an increased freezing point by comparison with the soil of Yakutsk.

The compressibility of saline permafrost is determined by the presence of water-soluble salts in its pores, which cause creep of the soil and its relaxation properties under a prolonged load. With an increase in salinity, the modulus of settlement of the soil increases significantly. For example, for nonsaline suglinok with a load of 1 kg/cm^2 and a temperature of -1.4°C, the modulus of settlement, according to the data of S. S. Vyalov, is 10 mm/m; under analogous conditions, but with a soil salinity of 0.45 percent, its magnitude is 90-100 mm/m.

Observations of the settlement of buildings in Amderma confirm the significant compressibility of the bearing ground. Thus, the settlement of the west wall of house No. 21 during one season (1961) amounted to 9 cm at a ground temperature under the footing of the foundation of -2.4°C and with a salinity from 0.3 to 1.0 percent. Under analogous conditions, but with a bearing ground temperature of -1.7°C, the children's building was in a state of emergency.

The equivalent cohesion [c_e] according to S. S. Vyalov, takes into account both the characteristic cohesion and the internal friction of the soil[1] and is taken as the primary criterion of strength of saline permafrost. The magnitude of it was determined by the ball test method.[2]

Experimental data on the magnitude of the equivalent cohesion of saline permafrost indicate that it depends on the same factors as that of nonsaline soil, that is, the ground temperature, its composition and moisture, the duration of effect of the load, and, the main thing, the degree of salinity of the ground.

Under the effect of a constant load, there is a reduction in magnitude of the equivalent cohesion with time where is is most intense and discontinuous during the initial very short period (several minutes). On the whole, saline permafrost is characterized by a slower reduction in equivalent cohesion during 8 h of testing than in experiments with nonsaline soil. A significant reduction in c_e is observed during 24-h and longer tests for sandy soils and supesses. The results of the tests indicate that the values of c_e for clay and suglinok with sufficient accuracy for practical purposes can be taken according to the 8-h test data. For supes and sand (especially silty), the long-term values of c_e [$=c_{eL}$] must be determined by the results of tests lasting not less than 24 h by the formula

$$c_{eL} = 0.8\ c_{24}.$$

The dependence of the equivalent cohesion on the degree of salinity of the soil is represented by the graphs (Figure 1) according to the data of tests in Amderma.[3] The results obtained in Yakutsk with respect to determining the value of c_e of suglinoks (when w = 30 percent) and sands (when w = 20 percent) with a degree of salinity from 0.10 to 0.80 percent and at ground temperatures of -2.3° and -4.5° agree satisfactorily with the data presented in the graphs.

Even a comparatively small amount of salt contained in the ground of natural composition has a very noticeable effect on the magnitude of c_e. Thus, for a temperature of -1°C and a degree of salinity (z)of the soil of 0.25 percent, the value of c_e is 0.7 kg/cm^2, and for nonsaline soil under the same conditions it is 1.5 kg/cm^2. With an increase in salinity to 1 percent, the value of c_e dropped to 0.2 kg/cm^2. The nature of variation of c_e from the degree of salinity of the soil, as is obvious from the graphs, remained identical also in other temperature intervals. The most intense reduction of c_e is observed for low salinity of the soil (0-0.3 percent). This pertains especially to sandy soil. With a reduc-

TABLE 1 Summary Data on the Freezing Point of Saline Soil with Disturbed Structure (by Experiments at Amderma and Yakutsk)

Soil Salinity, %	Freezing point [t_f] (°C) with Soil Moisture Content, %						
	15	20	25	30	35	40	45
to 0.25		-0.6[a]	-0.4[a]		-0.2[a]		
0.40	-2.0	-1.4	-1.2	-0.9		-0.8	
0.50		-1.0[a]	-0.8[a]		-0.5[a]		-0.3[a]
0.55	-2.6	-1.8	-1.5	-1.2		-1.0	
0.80		-2.1[a]	-1.6[a]		-1.1[a]		-0.7[a]
0.85	-3.8	-2.8	-2.3	-1.8		-1.4	
1.10		-2.8[a]	-2.2[a]		-1.7[a]		-1.2[a]
1.20	-4.8	-3.6	-3.1	-2.6		-2.0	
1.50		-4.0[a]	-3.2[a]		-2.6[a]		-2.0[a]

[a]Data from the Amderma Laboratory.

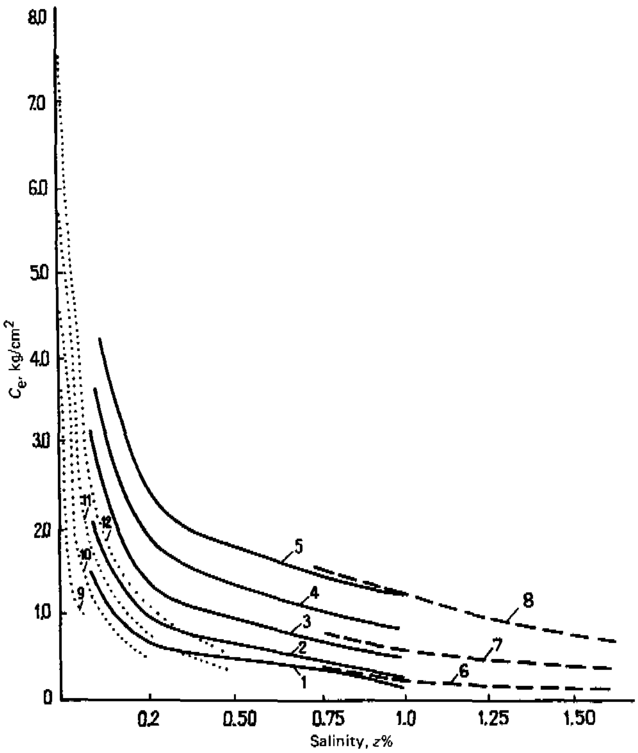

FIGURE 1 Long-term limiting value of the equivalent cohesion C_e as a function of the salinity of the soil (Z):

1--suglinok	t = 1°	7--silts	t = -3°
2--the same	2°	8--silts	-4.5-5.0°
3--the same	3°	9--sands	-1°
4--the same	4°	10--the same	-2°
5--the same	5°	11--the same	-3°
6--silt	-2°	12--the same	-4°

tion in temperature, an increase in the equivalent cohesion takes place with greater intensity than in the nonsaline soil. With variation of the soil temperature from -2°C to -4°C, the value of c_e increases from 0.3-1.0 kg/cm^2 to 0.6-2.0 kg/cm^2, respectively; that is, it doubles. In the same temperature ranges for nonsaline soil, the value varied from 2.1-2.8 kg/cm^2 to 3.4-4.6 kg/cm^2.

The frozen silt with salinity of 0.7°-1.5° differs little with respect to strength characteristics from the saline suglinok permafrost. The strength characteristics of the silt as a function of salinity and temperature are analogous to the indicated relations for saline frozen suglinok.

The sandy soils occurring naturally (z = 0.05-0.10 percent) are characterized by a reduced value of c_e with respect to the ideally nonsaline (z = 0 percent) by 1.5-4.0 times. With an increase in salinity of the sand to 0.5 percent, its bearing capacity drops, and the value of c_e drops at a temperature of -3°C to -4°C from 2.3-3.5 kg/cm^2 to 0.4-0.6 kg/cm^2. With respect to the suglinoks, the values of c_e for sand is 1.5-2.0 times lower at z = 0.2-0.5 percent. In our opinion, this difference is explained by the processes of ion exchange in the clay soil causing complex formation, strengthening of the bonds of the absorbed ions with the surfaces of the particles, and increased affinity for water. In sandy soil not having an absorption capacity, an increase in the salinity to 0.6-0.8 percent leads only to an increase in the amount of unfrozen water, which reduces the strength of the sand.

The effect of moisture on the magnitude of the equivalent cohesion of the saline permafrost is less significant than the ground temperature and soil salinity. With an increase in moisture content from 25 to 55 percent, the magnitude of c_e in the suglinoks dropped by 25-30 percent. For silt with high moisture content (60-90 percent), an increase in c_e was observed, which is apparently explainable by their water-colloid activity.

The resistance of the soil to shear with respect to the lateral surfaces of the foundation, as one of the basic characteristics used in designing foundations in saline ground, was determined experimentally during long-term testing with different types of soil, under different temperatures, with different salinity and moisture, and in soils with undisturbed and disturbed structure. Simultaneously with the laboratory experiments, determinations of the forces of freezing to piles of natural dimensions were made using single-shear and double-shear apparatus and models of the piles. At Amderma, a test was made on a metal pile 32.5 cm in diameter installed in suglinok with salinity of 0.37 percent with an average temperature with respect to pile length of -2.2°C and a wooden pile 20 cm in diameter also installed in the suglinok with a salinity of 0.45 percent at a temperature of -1.9°C. In Yakutsk experiments were run with three piles installed in sandy soil with ice inclusions at a soil temperature of -0.8°C and with a degree of salinity of 0.59, 0.51, and 0.40 percent, respectively.

The salinity of the soil, as the experimental results demonstrated (Figure 2), is one of the decisive factors determining the magnitude of the shear strength of the soil at the lateral surfaces of the pile. Thus, in silty suglinok an increase in the salt content in the ground to 0.5 percent at a temperature of -4.5° led to almost halving the value of R_s with respect to the nonsaline suglinoks (from 2.7-2.5 kg/cm^2 to 1.4-1.3 kg/cm^2). With an increase in salinity to 1.1 percent, the magnitude of R_s was 0.8 kg/cm^2, and with z above 1.5 percent the freezing of the ground to the pile was absent in practice, and the value of R_s was determined only by the shear strength of the unfrozen soil at the lateral surface of the pile. An analogous picture was observed also at other ground temperatures. For the experimental curves, a gradual variation of the curvature is characteristic with a trend of the transition to the line parallel to the x-axis corresponding to the constant value of R_s equal to the shear strength of the unfrozen ground along the lateral surface of the pile. The greatest reduction in the value of R_s takes place for a low salinity of the ground.

With a reduction in temperature, the magnitude

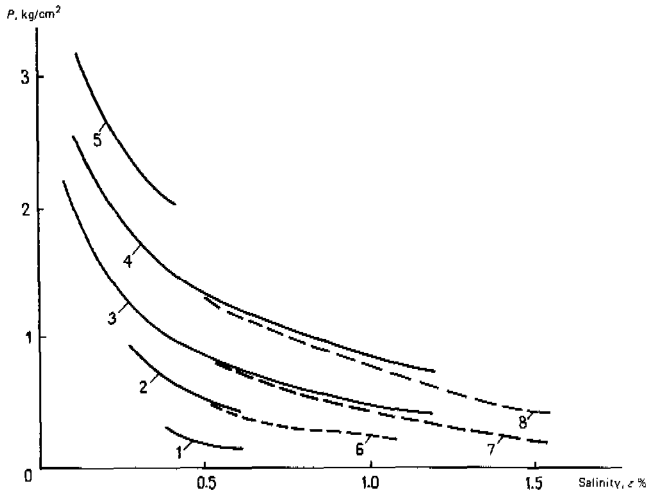

FIGURE 2 Variation of the shear strength of the soil with respect to the lateral surfaces of foundations as a function of its salinity at various temperatures (according to the Amderma test data):

1--suglinoks	$t = -1°$	5--suglinoks	$t = -5.5°$
2--the same	$-2°$	6--silts	$-2°$
3--the same	$-3°$	7--the same	$-3°$
4--the same	$-4.5°$	8--the same	$-4.5°$

of R_S rises significantly. A somewhat smaller (with respect to nonsaline soil) intensity of variation of R_S in the high temperature range is explained obviously by the stress relief of the crystal lattice of the saline ice with an increase in temperature.

The reduction in magnitude of R_S with an increase in salinity in the frozen silt is of the same nature. Unfortunately, there are few experiments with nonsaline sand, and it is premature to judge the magnitude of R_S for them. Nevertheless, attention has been attracted by the experiments in testing piles of natural sizes in Yakutsk in sandy soils. In both cases values of R_S were obtained which were somewhat higher than in suglinoks (R_S = 0.7 kg/cm^2 for t = 1.5° and z = 0.51 percent; R_S = 0.4 kg/cm^2 for t = -0.8° and z = 0.4 percent).

The reduction in magnitude of the shear strength of the permafrost along the lateral surfaces of the piles with an increase in soil salinity is unconditionally connected with a change in the amount of unfrozen water in the soil; however, here the relation is more complicated, since obviously simultaneously there is a variation of the strength characteristics of the ice remaining in the pores and its resistance to long-term loads.

Some of the variations of the long-term limiting values of R_S for suglinoks of different moisture content, as the experiments demonstrated, occur only for values of the latter below 22 percent, that is, below $w_P + 0.5I_P$. This is connected with a decrease in the amount of ice and worsening of the ground contact with the foundation. In the moisture range from $w_P + 0.5I_P$ to $w_L + 0.5I_P$, that is, most widespread under natural conditions, no noticeable variation in the value of R_S takes place.

In the practice of designing pile foundations, it is not necessary to consider the effect of the pile diameter on the magnitude of R_S. As has been established,[4] this factor applies to a pile diameter of less than 20 cm, which is primarily significant when doing research work.

The value of R_S varies significantly as a function of the state of the pile (foundation) surface. Comparable experiments with metal and concrete surfaces performed both under laboratory conditions with the models and in the test areas convincingly demonstrated the reduction in the value of R_S at the metal surfaces in comparison with concrete by 20-30 percent.[4] These divergences were caused not by the pile material, but by the state of their surfaces. This conclusion is confirmed by the results of experiments with wood slabs, the surfaces of which were planed smooth. The magnitude of R_S turned out to be half the value established on the basis of experiments with the unmachined surface. An analogous picture was observed in the experiments with metal piles with different degrees of machining of the surfaces. For piles with a smooth surface, the values of R_S turned out to be 25 percent lower than for piles with a rough surface.[4]

The technological process of embedding the piles in the frozen ground shows up in the

TABLE 2 Norm Resistance of Saline Permafrost to Normal Pressure

Type of Ground	Salinity, %	Values of R^N(kg/cm^2) at a Ground Temperature, °C				
		-1.0	-2.0	-3.0	-4.0	-5.0
Suglinoks	0.50	2.3	4.3	6.0	8.8	12.5
	1.00	1.7	2.4	3.4	4.9	7.0
Silts	1.50	*a*	1.7	2.4	3.4	4.9
Silty Sands	0.10	3	5.0	9.0	13.0	
	0.20	*a*	2.5	5.5	6.5	
	0.50	*a*	*a*	2.0	3.0	

[a]Soil can be in the nonfrozen state.

TABLE 3 Norm Shear of Frozen Saline Ground at the Surfaces of Freezing to Concrete Foundations

Type of Ground	Salinity, %	Values of R_s^N (kg/cm^2) at a Ground Temperature, °C				
		-1	-2	-3	-4	-5
Clays and silts	0.5	0.35	0.65	0.90	1.25	1.90
	1.0	*a*	0.32	0.45	0.65	1.20
	1.5	*a*	*a*	*a*	0.37	0.80
Suglinoks	0.5	0.40	0.70	1.00	1.30	1.90
	1.0	*a*	0.35	0.50	0.70	1.20
	1.5	*a*	*a*	*a*	0.40	0.80

[a] Soil can be in the nonfrozen state.

magnitude of R_s. A series of comparative experiments was performed with metal piles emplaced by various procedures in suglinok of the same granulometric composition, salinity of 0.3 percent, moisture content of 42 percent, and at a temperature of -4.5°C. Results similar to the Norm values were obtained on placing the piles in previously drilled holes of large diameter with slurry filling. When placing by steaming, R_s turned out to be approximately 10 percent less. When placing by forcing into a hole of smaller diameter than the pile itself, the value of R_s turned out to be almost 30 percent higher than the Norm value. Obviously, the reaction of the soil played a positive role here.

A noticeable increase in R_s with the forcing of the pile in indicates the progressive nature of this method of driving piles.

The Norm values of the resistance of saline permafrost to the normal pressure were established by the magnitude of the limiting load calculated in accordance with the proposals by N. A. Tsytovich and S. S. Vyalov by the Prandtl formula with substitution in it of a complex characteristic--the long-term values of the equivalent cohesion established experimentally.

The calculation results are presented in Table 2.

Comparing the data of Table 2 with the values of R^N for the nonsaline ground,[5] let us note that the Norm values of the resistances to the normal pressure for saline ground are on the average 2-4 times less than for nonsaline ground.

The Norm shear strength of the saline permafrost at the lateral surfaces of the piles obtained on the basis of the experimental data are presented in Table 3. Here there is no information about the sand, hence there are insufficient experimental data for developing recommendations.

For a soil salinity less than 0.25 percent, the values of R_s^N are taken just as for the nonsaline ground according to Table 14 of SNiP II-B.6-66, and for intermediate values of z they are determined by interpolation.

The values of R_s^N are presented in Table 3 for concrete foundations poured in forms made of ordinary boards (rough surfaces). With the use of metal piles made of ordinary rolled products stored in the air, the values of R_s^N must be taken with a factor of 0.75. For different conditions of the concrete, metal, or wooden surfaces of the foundations, or with special treatment of them, the values of R_s^N are determined experimentally.

For pile foundations the given values of R_s^N can be taken when driving the piles into previously drilled holes, the diameter of which exceeds the greatest cross section of the pile. When installing the piles by the method of steam thawing, the values of R_s^N must be taken with a factor of 0.85. When forcing piles into a hole of smaller diameter than the pile, which is efficient in plastically frozen ground with a high degree of salinity, the value of R_s^N can be used with a factor of 1.30.

The design of pile and pier foundations on saline permafrost with respect to bearing capacity (strength) is made using Formulas (9) and (9b) of SNiP II-B.6-66, taking the values of R^N and R_s^N determined considering the salinity of the soil on the basis of the above recommendations. Here, the physical characteristics of the saline permafrost are determined taking salinity into account.

The thermophysical characteristics of the saline soil are determined considering the increased content of nonfreezing water at a negative temperature. For soil in the plastically frozen state as a result of increased salinity, it is necessary to design footings with respect to deformations.

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ARTIFICIAL THAWING OF PERMAFROST BEFORE CONSTRUCTION

V. F. ZHUKOV AND V. D. PONOMAREV Scientific Research Institute of Foundations and Underground Structures

The idea of the expediency of preconstruction thawing of permafrost used to support buildings and structures arose when the great thaw of permafrost, which led to large, nonuniform settlement of structures became known in construction practice. The thaw-consolidation of the permafrost depends on the amount of ice contained in the permafrost and the type of cryogenous texture.[1]

The purpose of preconstruction thawing of permafrost is to advance consolidation of the ground (and in some cases additional fill) in order to decrease or even eliminate the thermal effects of the building on the ground and at the same time insure its stable state during its operation and maintenance. This type of thawing of the permafrost under buildings and structures is expedient and sometimes even unavoidable when:

1. The foundation is on ice-rich permafrost (the water content of the ground is greater than the liquid limit) and especially if the soil in the thawed state has a high permeability (sand, coarse soils).
2. The permafrost table is located at different depths below ground surface, and therefore the thawing of the ground under one part of the building after it is in operation begins much sooner than under another. In this case, the settlement of parts of the buildings will be especially significantly different.
3. The natural temperature of the permafrost is close to 0° (from 0° to -1°), and it is impossible to insure maintenance of the frozen state of the ground under a building after it is put into operation. This is especially true of the southern permafrost regions.
4. As a result of the technological purpose of the building (for example, in the case of wet or hot production processes) or large plan dimensions of the building, it is impossible to insure that the frozen state of the soil will be maintained or to determine whether the latter is economically disadvantageous.

The preconstruction thawing of permafrost under buildings and structures is used in various engineering-geocryologic conditions.

In Magadan oblast preconstruction thawing is especially used when building on coarse soils. Thawing of the permafrost is carried out using water pumped into the ground and moving through the pores of the thawing ground.

In northern Europe preconstruction thawing was used in suglinoks and was by means of electricity. When thawing the permafrozen suglinok to a depth of 10 m, the settlement was from 40 to 100 cm, and during the maintenance period it was significantly less and damped in general with time (Figure 1).

In recent years the properties and characteristics of thawing and thawed ground have been studied in more detail in order to refine the engineering calculations and for better organization of the production processes. The results of scientific research have served as a basis for finding answers to the problems presented in practice.

The thawing of the permafrost is accompanied by the destruction of the cryogenous texture and the forcing together of the mineral aggregates.[2] In order to destroy the texture, it is necessary to have an initial pressure of not less than 1 kg/cm^2 to overcome the resistance of the aggregates to displacement at the contacts between grains. In certain cases the aggregates do not change with respect to volume and density, but in others they swell or decompose into individual particles.[3,4]

The degree of destruction of the cryogenous

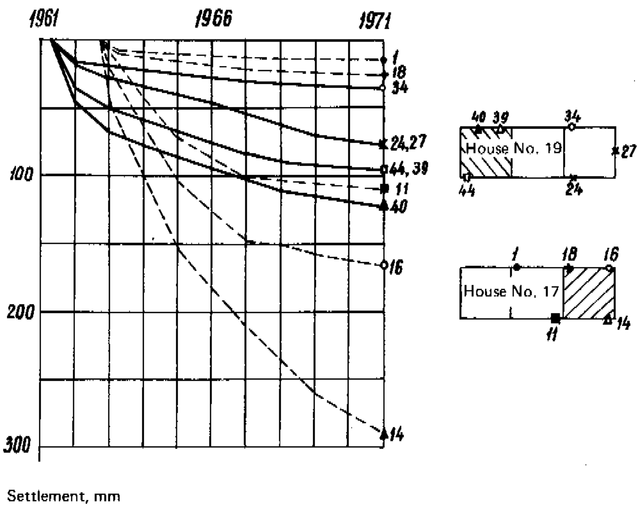

FIGURE 1 Settlement of buildings 17 and 19 along Mir Street in Vorkuta during the period from 1961 to 1970. The figures on the plan view of the buildings indicate the numbers of the benchmarks. The cross-hatching indicates the parts of the building under which the ground was thawed before construction.

structure of the thawed ground (η) can be determined by the formula:

$$\eta = \frac{w - w_P}{w_z - w_P},$$

where w is the water content of the thawed soil, w_P is its plastic limit, and w_z is the water content determined from the graph relating the cohesion of the soil in its broken-down structural state and its water content with cohesion of the soil corresponding to the thawed condition.[5]

The cryogenous texture in the thawed ground is partially retained for 2 to 6 months.[6] During this period, the deformation and strength characteristics of the thawed ground gradually change and approach the values that are noted for ordinary unfrozen ground.

The studies of the structure of the thawing bearing surfaces under loaded punches established[7] that after thawing the ground layer initally is a nonuniform body with values of the physical and mechanical characteristics of the ground, which vary with depth; the porosity of the ground increases with depth, and its variation is of an exponential nature; the strength characteristics and the deformation modulus vary in the same way. This can be explained by the fact that the time of existence of the thawed ground in its various layers with respect to depth is noticeably different (with time this difference smooths out) and also by the fact that the stress in it caused by the load decreases with depth. The soil composition has special significance. In coarse soils that are not aggregated during the freezing process, a more uniform structure is formed after thawing.

In slow disappearance of the cryogenous texture in the thawed clay soil, there is swelling of the mineral aggregates. The swelling of the thawed soil has been noted long ago. In addition to swelling during the process of soil consolidation during thawing, some of the macropores are retained because they have been converted into closed isolated cavities. Therefore, the degree of consolidation of the thawing ground turns out to be less than if it is determined by assuming the disappearance of the volume of the macropores in it.[2] In practice, this is taken into account when calculating the settlement of the soil by introducing a correction factor into the formula that characterizes the cryogenous texture (SNiP II-B.6-66).

The settlement(s) of the foundation of the building erected on the thawing permafrost is made up of three parts: settlement from destruction of the cryogenous texture of the entire thawing layer (h), settlement from the weight (P_o) of the thawed overburden, and settlement from the pressure of the structure ($\bar{P}$) transmitted to the base of the footing to a depth of h_p

$$s = \Sigma^{h} h_i A_{oi} + \Sigma^{h} h_i a_{oi} P_o - \Sigma^{hp} h_i a_{oi} \bar{P}_i .$$

When thawing the ground to a depth h before construction, the settlement of the ground will be determined only by the third term of the equation. However, in the case of incomplete consolidation after thawing (that is, if the building is erected before the cryogenous texture has disappeared), settlement takes place characterized by the second term of the equation but with a correction factor taking into account the period of time between the construction of the foundation and the preconstruction thawing.

If the preconstruction thawing is carried out to a depth of $h_{pt} < h$, then the settlement of the building will in this case increase as a result of the thawing and consolidation of the soil layer below the depth h_{pt}. The minimum required depth of thawing must be such that the calculated settlement of the bearing soils will not exceed the allowable value for the given structural design of the building. In Figure 2 we show how h_{pt} is calculated for one point of the building in plan. During the planning and design process, it is necessary to calculate the settlement for a number of points, taking into account its development so that one may determine the magnitude of the differential settlements of the individual parts of the erected building.

An entire series of methods of preconstruction thawing of the permafrost has been developed. The commonest procedures of the best technological nature are hydraulic thawing and electric thawing.

The hydraulic method of thawing is used for coarse soils.[8] In this case the heat of river, or production (discharge), water is used, which is injected into the ground.

Electric thawing is used primarily for clay soils. Two methods of thawing permafrost by electricity have been developed:

1. A current is passed directly through the

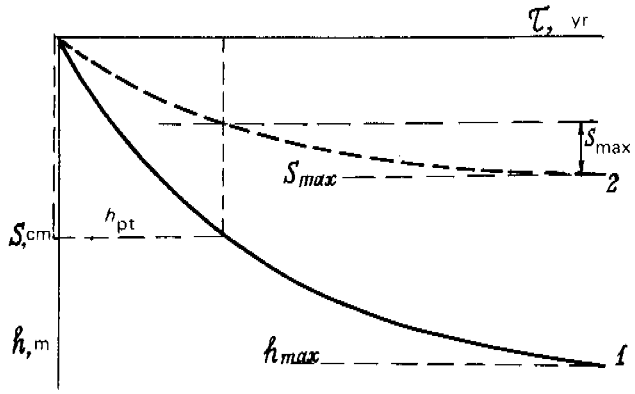

FIGURE 2 Calculation of the depth of preconstruction thawing of the ground with a given value of the limiting admissible settlement. 1--curve of thawing of the bearing surface; 2--settlement of the bearing surface during thawing.

ground. In this case the resistance of the permafrost falls at an increase in temperature and the ground warms up and thaws. This method is effective only in clay soil.

2. Various types of electric heaters are buried in the ground (for example, wire coils, etc., are placed in boreholes). This method of thawing is applicable both in coarse and fine-grained soils.

The time required to thaw the soil in the base is on the average 1.5-2.5 months. The course and degree of its consolidation depend on the method used in thawing the ground. The most complete consolidation of the thawing ground takes place with direct heating of it by an electric current. In this case the thawing begins throughout the entire area in layers with the least ohmic resistance, and it gradually spreads from top to bottom. As the ground thaws, it consolidates under the weight of the overburden. The pressure on the thawing ground decreases somewhat as a result of friction against the electrode pipes. This reduces the consolidation effect. In addition, the pipes act as drains to drain the water formed from thawing of the ice.

In the case of application of vertical heat sources, the soil melts radially around them. The melting masses of soil of cylindrical shape are located inside the frozen mass. During compression and displacement of the soil downward, friction occurs at the contact between the thawed and frozen ground. In a way it hangs, and its consolidation under its own weight becomes especially incomplete and less than in the former case.

After considering the two-dimensional problem of the limiting state of the thawed ground (Figure 3) at the contact with the frozen ground, it is possible to obtain the equation of the curve of load reduction:[9]

$$x = \frac{c}{\gamma} (In\ h_t - In\ z) + K(h_t - z) = \tan\phi,$$

where c is the cohesion at the interface between the thawed and frozen ground; γ is the unit weight of the thawed ground; K is the coefficient of earth pressure; ϕ is the angle of internal friction; h_t is the depth of thawing equal approximately to the depth to which the heat source is buried; x and z are the coordinates of the curve.

For the case of a vertical heat source, the weight of the thawed ground at a distance $x = 2\ b$ can be taken completely by the vertical walls of the frozen mass if $2\ b/h_t < K\tan\phi$.

If, for example, K - 0.6, $\phi = 15°$, and h_t = 10 m, then the settlement of the thawed ground around the vertical heat source will not take place until the thawing extends to > 0.8 m. Even with further thawing of the ground, the consolidation near the thawing plane will be incomplete.

In clay soils, the retardation of the thawing rate assists the development of heave processes and destruction of the mineral aggregates. As a result, the thawing ground becomes more dense. Thus, it is expedient to do the preconstruction thawing as fast as possible at sites with dimensions appreciably greater than the depth of thawing, and with the use of drains.

A highly laborious step in working with thawing permafrost is the installation of the electrodes, especially to a depth of more than 8 m or 10 m. Whereas previously the electrodes were buried in previously drilled holes and a great deal of time was spent on this operaton, at the present time the process is carried out much more quickly. The pipes for hydraulic and electric thawing are installed using vibrators and air hammers.[10,11] The rate at which the pipe, 73-108 mm in diameter, is buried, when driving with a vibratory hammer with two motors, each of 1 kW, is 0.3-1.0 m/min.

In recent years, on the basis of the achievements of engineering geocryology, the methods of planning and designing foundations with preliminary thawing of the permafrost have been improved significantly. The detailed study of the behavior of the ground during thawing and the forma-

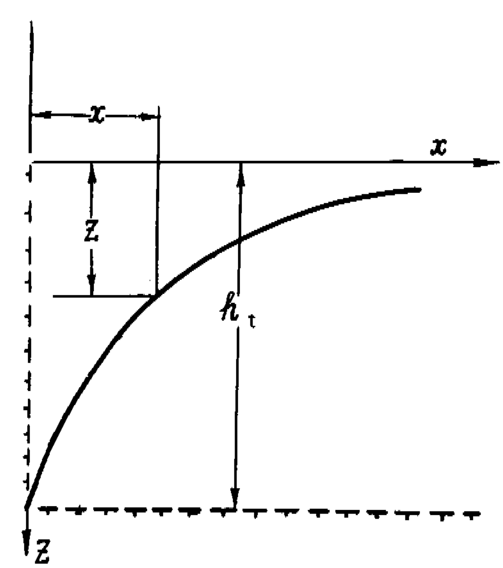

FIGURE 3 Plot of the load-relief equation.

tion of the thawed ground permits an increase in the efficiency of application of this method of insuring the stability of buildings and structures and an increase in its economy on the basis of using modern technological procedures.

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METHODS OF COOLING PLASTICALLY FROZEN GROUND

G. N. MAKSIMOV AND S. I. ZANYATIN Scientific Research Institute of Foundations and Underground Structures

A. A. KONOVALOV Krasnoyarsk Promstroyniiproyekt Institute

In the southern permafrost regions there are ground temperatures above -1°C, which complicate construction conditions in these areas.

Plastically frozen ground with a temperature close to 0°C is especially unfavorable for construction.[1,2] In order to insure stable construction and in order to increase bearing capacity, it is recommended that the ground be cooled (*Stroitel'nyye Normy i Pravila* [*Construction Norms and Guidelines*], SNiP II-B.6-66).

When solving this problem, it is necessary to establish whether it is possible to retain the cooled state of the permafrost in the footing of the buildings equipped with ventilated air spaces and to discover the most efficient methods of cooling them. Thus, there are individual data indicating that under some wooden buildings in Amderma the mean annual ground temperatures rise.[3] These data were also obtained for the first stone buildings in Noril'sk with low air spaces, which were ventilated insufficiently (Figure 1b, I, II). At the same time, for the majority of buildings with ventilated air spaces, if their height is sufficient and ventilation is provided, the negative temperature of the underlying ground is retained, and the temperature is even lowered. This is observed in Noril'sk after increasing the requirements on ventilating the air spaces and achieving a ratio of the ventilation inlet area to the building area which is taken as the ventilation modulus ($M = 0.0025$).

In order to observe a unified construction principle, it was decided to lay the heating pipes so as to retain the frozen state of the ground supporting them and to eliminate separate sewers. All of the house utilidors began to be designed in the air spaces of the buildings suspended in insulated pipes. For convenience of operation and maintenance of these systems, it was necessary to raise the height of these air spaces to 1.0–1.2 m. Many years of observations have demonstrated that in the foundations of buildings constructed observing these solutions there is a drop in the ground temperature by comparison with the temperatures existing before building up the area.[4]

At the present time, under the residential buildings in Noril'sk, the mean annual temperature at the depth of the seasonal thawing $-t'_0$ has dropped to −6.5°C to −7°C at the same time as this temperature was −3°C to −4°C before construction. An analysis of the temperature observations in Yakutsk[5] demonstrated that, in the foundations of modern buildings with ventilated air spaces, a cooling zone is also formed (Figure 1c, I).

Thus, under the buildings constructed by principle I, the mean annual negative temperatures at the boundary of the seasonally thawing layer can vary within broad limits. When

$$t_3 \geq t'_0 \geq t_0, \qquad (1)$$

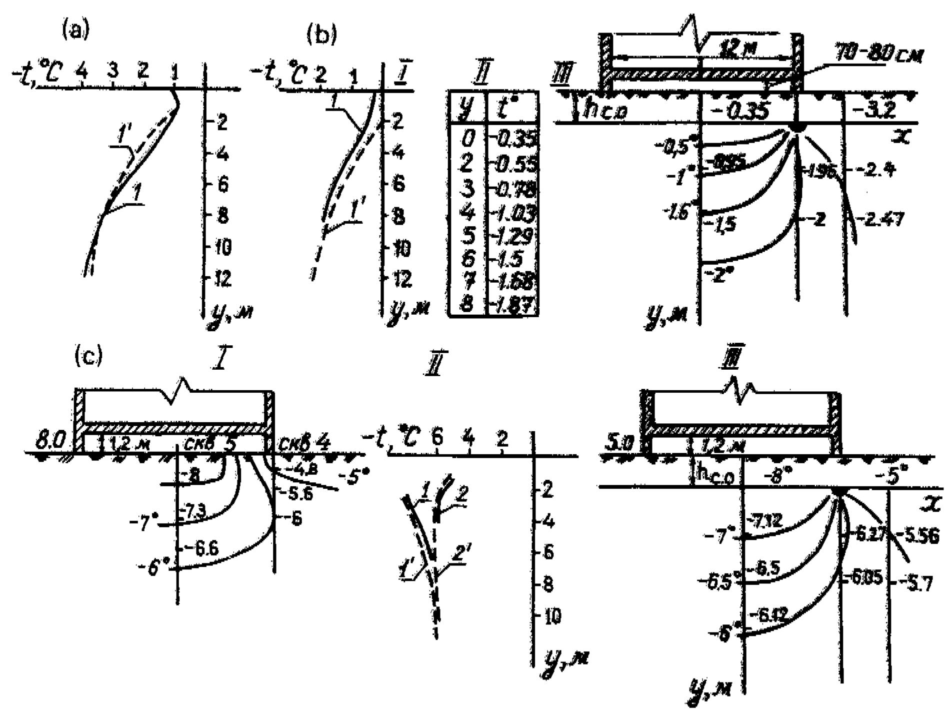

FIGURE 1 (a) Ground temperature under the center of a wooden building with air space in Amderma; (b) the same in Noril'sk under a stone building on Zavodskaya Street with unorganized ventilation of the air space. I--comparison of the natural and calculated data; II--values of the measured mean annual temperatures; III--calculated temperature field with heating zone under the building. (c) Temperature field with cooling zone under a school on Yaroslavakaya Street in Yakutsk. 1--by the processed data of G. O. Lukin. II--comparison of the natural and calculated data; 1--measured mean annual temperatures under the center of the building; 1'--the same determined by calculation; 2--measured mean annual temperatures under the edge of the building; 2'--the same under the edge of the building determined by calculation; III--calculated temperature field with cooling zone.

$$t_0 \geq t'_0 \geq t_a, \qquad (2)$$

where t_0 is the mean annual ground temperature at a depth of 10 m before constructing the building; t_3 is the freezing point of the ground ($t_3 \approx 0$); t_a is the mean annual air temperature in the given region. In order to preserve the cooled state of the ground in the foundations of the buildings during their operation and maintenance, condition (1) is inadmissible; condition (2) must be observed under the buildings.

The reduction in mean annual ground temperatures in the ground under buildings depends on the climatic indexes of the region and the structural solutions of the ventilated air spaces. On the basis of analysis and generalization of many years of observations of the air temperature in the air spaces of the buildings (the laboratory of the Noril'sk combine) and also the data of the authors,[6,7] it has been established that a significant difference, Θ_1, is observed between the mean winter air temperature in the air spaces, t_{wc}, and the mean winter temperature of the outside air, t_{ow}. For example, for buildings with weak and unorganized ventilation of the air spaces $\Theta_1 \geq 12°C$; for the buildings of Noril'sk and Yakutsk, $\Theta_1 = 6°C$ to 8°C; and only for air spaces with decorative shutters removable for the winter, $\Theta_1 = 1°-2°C$. The difference (Θ_2) between the mean summer air temperatures in the air spaces, t_{sc}, and the mean summer temperatures of the outside air, t_{so}, depends less on the structural details of the air spaces and is usually $\Theta_2 = 1°-2°C$ with a negative sign. If we introduce values of Θ_1 and Θ_2 into the formula for calculating the mean annual temperatures of the permafrost at the seasonal thawing layer boundary,[1] then it is possible to determine the mean annual temperature, t'_0, under the buildings with air spaces in different parts of the Soviet Union depending on the climatic data.

$$t'_0 = \frac{(t_{ow} + \Theta_1)\tau_w + \dfrac{\lambda_T (t_{so} + \Theta_2)\tau_s}{\lambda_M}}{365}, \qquad (3)$$

where in addition to the notation indicated above, τ_w is the duration of the period with $t < °C$ in days; τ_s is the same period with $t > 0°C$ in days; λ_T is the coefficient of thermal conductivity of the thawed ground of the seasonally thawing layer; λ_M is the same for the underlying permafrost.

By the known t'_0 it is possible to determine the mean annual temperatures at any point of the halfspace under a cold infinite strip for steady state

$$t_{x,y} = t_0 + (t'_0 - t_0)\beta_\infty, \qquad (4)$$

where the coefficient β_∞ is defined by the formula[8]

$$\beta_\infty = \frac{1}{\pi}\left(\arctan\frac{0.5B + x}{y} + \arctan\frac{0.5B - x}{y}\right), \qquad (5)$$

where B is the band width; x is the distance from its center; y is the depth reckoned from the foot of the seasonally thawing layer (h_a).

There is an analogous solution also for a strip of finite length.[9] The calculation by Formula (4) gives good comparison with the observations in the bearing ground of buildings after 4-5 yr of their operation and maintenance, which is obvious from the comparative graphs (Figure 1b, II-III, and 1c II, III).

The maximum calculated ground temperatures below the seasonally thawing layer are defined by the formulas:

$$t_{max} = t_{x,y} - (t_3 - t'_0)\beta_2 \qquad (6)$$

$$\beta_2 = \exp\left(-\frac{y}{52.83}\sqrt{\frac{c_{eff}}{\lambda_M}}\right) \qquad (7)$$

The natural cooling of the ground under the buildings takes place slowly. In Figure 2 the observed and calculated data on the temperature drop in the ground under a standard Noril'sk municipal building are compared. The methods of calculating the transient temperature fields during cooling of the ground from the surface were developed recently using analog and computer engineering.[10] The design data from the machines are approximated by the following formulas:

$$\kappa = \beta/\beta_\infty = \exp\left[-\frac{0.15}{F_0}\left(1 - \beta_1\right)\right] \quad \text{for } 1 > \beta_1 > 0.5, \qquad (8)$$

$$\kappa = \beta/\beta_\infty = \exp\left[-\frac{0.75}{F_0}\left(0.6 - \beta_1\right)\right] \quad \text{for } 0.5 > \beta_1 > 0, \qquad (9)$$

where F_0 is the Fourier number and κ is a coefficient taking time into account.

The temperature $t_{x,y,\tau}$ for the time τ at any point of the halfspace under an infinite strip is defined as follows:

$$t_{x,y,\tau} = t_0 + (t'_0 - t_0)\kappa\beta_\infty. \qquad (10)$$

If the building has a form of a short rectangle, then

$$t_{x,y,\tau} = t_0 + (t'_0 - t_0)\kappa\kappa_1\beta_\infty, \qquad (11)$$

where κ_1 is defined by the Formula (8), setting $F_0 = \alpha_{eff}\tau/L^2$; L is the length of the building.

For the building shown in Figure 2, calculations were made for two cases $\alpha_{eff} = 0.001\ m^2/h$ (plastically frozen suglinoks) and $\alpha_{eff} = 0.003\ m^2/h$ (sand). During the first year, the temperature drop under the building takes place more slowly than calculated. (This is connected with the fact that in the theoretical calculation no heating of the ground is taken into account when installing the pile foundations.) Thereafter,

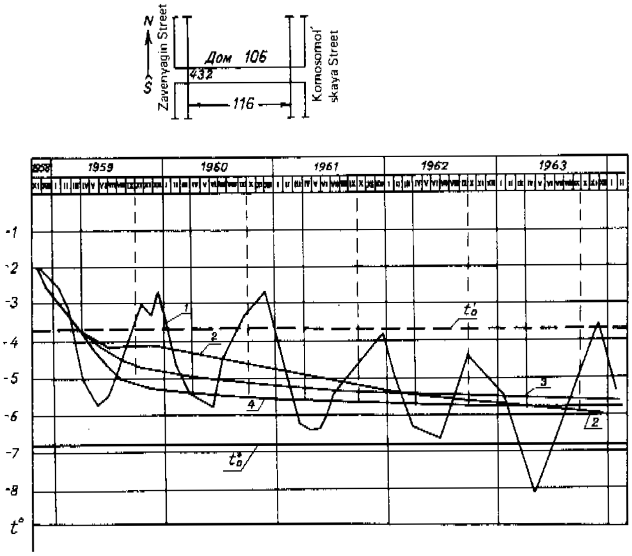

FIGURE 2 Formation of the cooled zone under a building in Noril'sk. 1--temperature variations at a depth of 5 m by month; 2--variation of the mean annual temperature according to observations; 3--calculated curve for the variation of the mean annual temperature by Formula (8) for suglinoks; 4--the same for sand; t_0--mean annual natural temperature; t'_0--mean annual temperature formed at the boundary of the seasonally thawing layer under a building during operation and maintenance.

a temperature drop takes place even more rapidly than follows from the calculations. This can be explained by an increase in the values of α_{eff} after the general soil temperature drop in the ground.

In plastically frozen ground at $t_0 \approx 0°C$, the coefficients κ in Formulas (8) and (9) characterize the relative temperature drop with time by comparison with the limiting possible steady state. For the point $x = 0$ and $y = 3.5$ m (experimental data are presented for it in Figure 2), the coefficients κ are indicated in Table 1.

From Table 1, it follows that when using the natural surface cooling method under the center of the building 80 percent of the steady-state cooled state is achieved in the sandy ground during the second year, and in the clay soil, after 4 yr. Less cooling in the southern zone of permafrost is not effective, since, with an increase in temperature in the cooling zone by the end of the year, the soil in the ground under the building will convert annually to an unstable plastically frozen state.

The Noril'sk builders took measures to accelerate the formation of the cooled zone under the buildings. For this purpose, the construction of the building was delayed. In the summer in the air space and near it around the perimeter of the building insulation made of sawdust was laid, which was taken up on the onset of winter. In the winter, snow was cleaned away periodically from the building to a width of not less than 2 m. In spite of such complex operations, the cooled zone under the building was created for only 2 yr.[11]

In 1964 it was proposed that the plastically frozen ground be cooled in the ground under the buildings through holes drilled for pile foundations.[4]

This method of cooling does not depend on the time of year. In the winter it is expediently realized by artificial ventilation of the drill holes; in the summer it is possible to use solid carbon dioxide in the same holes.

When cooling through the holes, the process of the formation of a cooled zone in the ground under the building is made up of three stages: in the first, the ground is frozen and cools significantly in a certain volume around the hole; in the second, a natural general increase in size of the cooled zone takes place with equalization and an increase in the temperature in it at depths where the primary cooling took place; in the third stage, and during the operating period, the size of the artificially cooled zone increases with the later formation of a temperature field corresponding to the steady state of heat exchange.

Consequently, when cooling the ground through holes, it is important to designate the optimal dimensions (radius) of the primary cooling zone. As applied to the five-story residential buildings widely used in construction in the North, it was proposed that it is sufficient to lower the temperature within the limits of the mean annual steady temperatures in the block of ground of dimensions L equal to the spacing of the

TABLE 1 Variation of the Coefficient κ with Respect to Years in the Case of Surface Cooling

	Suglinoks, $\alpha_{eff} = 0.001$, m^2/h		Sand, $\alpha_{eff} = 0.003$ m^2/h	
Years	F_0	$\kappa = \beta\beta_\infty$	F_0	κ $\beta\beta_\infty$
1	0.061	0.44	0.183	0.76
2	0.122	0.66	0.366	0.86
3	0.183	0.76	0.55	5.91
4	0.244	0.81	0.73	0.93
5	0.305	0.83	0.92	0.95

foundation with respect to the length of the building and in the radius $R = 0.5\ B$ below the layer of seasonal thawing.

After compiling the heat balance equation it is possible to determine the required radius R_1 of the primary cooling zone near the hole:

$$R_1 = \sqrt{\frac{r_z^{2}(t_0 - mt)n - \frac{K_1L_1R}{h_m}(t_{av}^{I} - t_0)}{n(t_0 - mt)}}, \quad (12)$$

where t is the temperature of the coolant used; n is the number of piles in the transverse cross section of the building; h_m is the depth of the piles. For plastically frozen ground it is recommended that $h_m \geq 5$ m. The methods of calculating the cooling radii--r_z^2 --are presented in Maksimov;[11] t_{av}^{I} is determined for buildings with $B \leq 12$ m as the mean between t_0' and t_n at the limit with the radius $R = 0.5\ B$:

$$t_n = t_0 - 0.5(t_0' - t_0). \quad (13)$$

For t_{av}^{I} for wider buildings, it is recommended that 6-8 points of the cooled module be determined by Formula (4).

The coefficient m takes into account the temperature drop between the wall of the hole and the coolant. When cooling the wells with outside air, $m = 0.65\text{-}0.75$.

In general form the formula by which the time (τ) of formation of the cooled cylinder of radius R_1 is determined when cooling through holes or hollow piles buried in the ground has the form of (12):

$$\tau = \frac{K_2K_3qr_0^{2}}{2\lambda_M(t_0 - t)}\left\{\left(\frac{\lambda_M}{\alpha r_0} + \frac{\lambda_M}{\lambda_1}\ \text{In}\ \frac{r_1}{r_0}\right)\left(\frac{R_1^{2}}{r_0^{2}} - 1\right) + \left[\frac{R_1^{2}}{r_0^{2}}\left(\text{In}\ \frac{R_1}{r_1} - \frac{1}{2}\right) + \text{In}\ \frac{r_1}{r_0} + \frac{1}{2}\right]\right\}, \quad (14)$$

where r_1 is the outside radius of the hollow pile; r_0 is the inside radius of the hollow pile; λ_1 is the coefficient of thermal conductivity of the pile material. When cooling through holes, $r_1 = r_0$ and expression (14) is simplified. K_2 and K_3 are coefficients taking into account the nonuniformity of the soil and the acceleration of freezing of the plastically frozen ground with layered texture. [λ_M is the coefficient of thermal conductivity of the frozen soil; a is the coefficient of surface heat transfer.]

Good coincidence of the calculations with observed data was extablished when w_{av}^{N} is determined for $t_{av}^{II} = 0.5\ (mt - t_0)$; w_o^{N} corresponds to an unfrozen water content at a natural temperature of $-t_0$. There are other methods for calculating the duration of cooling of the ground near a drill hole.[12]

Usually in plastically frozen suglinoks, to achieve the design levels, it is necessary to ventilate the wells when $t = -20°C$, $\tau = 22$ days; when $t = -30°C$, $\tau = 8$ days; when $t = -50°C$, $\tau = 5$ days. Air cooling is especially efficient when the air temperatures are below -25°C.

In 1967, summer cooling of the plastically frozen ground by solid carbon dioxide was used. During "dry-ice" cooling, as the experiments on models and the natural experiments demonstrated, it is necessary to take $m = 0.5$ and $t = t_c$ is the sublimation point of dry ice. Dry-ice cooling in the summer turned out to be more economical and simpler than the use of refrigerators for these purposes.

The piles were buried to a depth of 7 m. The dry ice was put into the piles to a height of $h = 3.5$ m. No cooling was provided for higher up, since it was considered that the ground would be cooled naturally from the surface at the beginning of winter. From a depth of 3.5 m to 7 m, no provision was made for contact of the cooled zones near the piles arranged with spacing of $L_1 = 4.0$ m to 4.3 m, since when $R_1 / r_0 \geq 10$ the cooling has low efficiency. Three tons of crushed dry ice were put into 21 of the piles in two sets with intervals of 3 days. The amount of dry ice (P) that must be put in the holes depends on the time the work is done, and it is defined by the formula

$$P = \frac{K_\tau qVn}{1{,}000A\eta}, \quad (15)$$

where n is the number of piles; A is the heat of sublimation of the dry ice (140 kcal/kg); η is the loss factor during sublimation with the exhaust gases; $\eta = 0.9$; V is the volume of the cooled zone with R_1 defined by Formula (12); K_τ is the coefficient determining the time the operations are performed on June 1; $K_\tau = 0.4$ when the operations are performed on September 1; K_τ is determined by interpolation between the indicated dates.

In Figure 3 we have the results of cooling the ground under the building that was done in the middle of August 1967.

The calculations and observed data (Figures 1 and 3) show that with cooling zones in the ground under the buildings the maximum and mean annual temperatures at the edge of the building are higher than under its center.

In the first approximation it is possible to assume that the mean annual temperatures at the edge of the building (when $x = 1$) are constant for a depth of < 5 m. They are calculated by Formula (13). Considering that $t_0 \approx 0$, it is found that for the indicated depths $t_{1,y} \cong 0.5t'_0$ or $t'_0 \cong 2t_{1,y}$. If the transition point of the plastically frozen to solidly frozen ground is taken as $t_T \geq -1.5°C$ (SNiP II-B.66-66), it is obvious that the maximum t'_0 for which it is still possible to retain the solidly frozen state of the ground under the building $t_0' = -3°C$. On the basis of the calculations by Formula (3), zones were outlined on the map of the USSR where $t_0' = -3°C$ can be achieved for buildings with different structural arrangements of the air spaces (Figure 4a). Line 1 defines the southern boundary of the one-time cooling of the bearing ground for buildings with air spaces equipped with ventilation holes with a cooling modulus, $M = 0.0025$. The equipment of

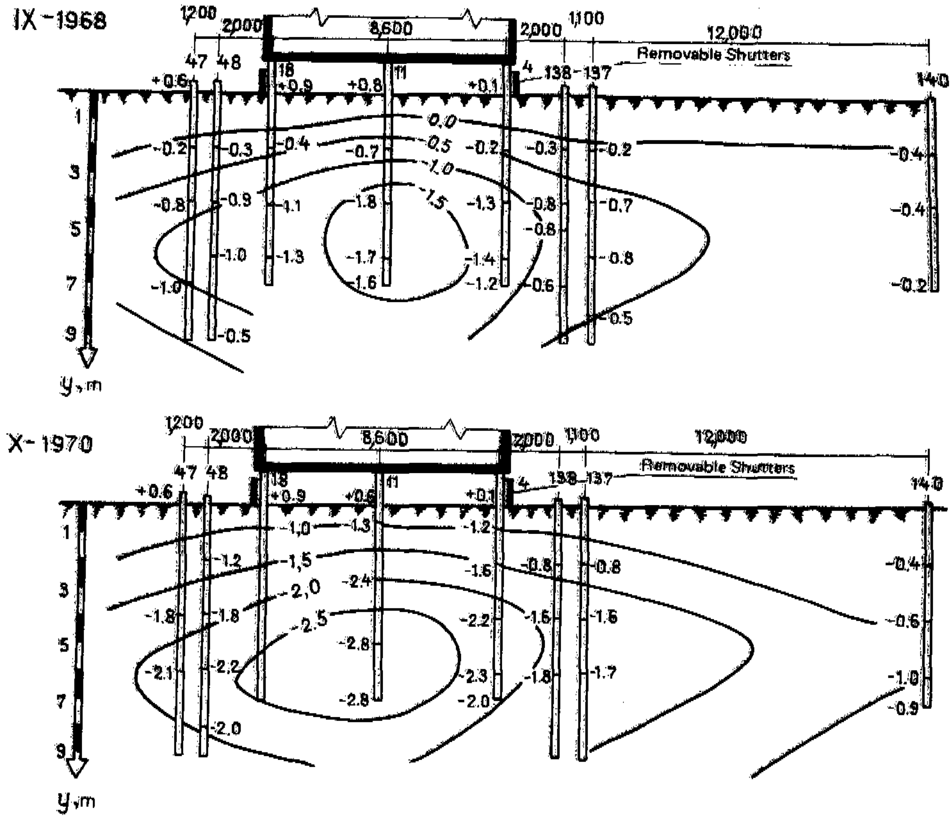

FIGURE 3 Temperature fields under a building, the supporting ground of which was cooled with dry ice.

the air spaces of the buildings with decorative or winter-removable shutters permits limitation of one-time cooling of the ground in the more southerly regions to a limit of II. Between the limits II and III, mean annual temperatures of $t'_0 > -3°C$ are formed under the building. In these regions, in addition to the mandatory primary cooling, it is recommended that seasonally active convection or gas-condensate refrigerators be installed or sealed hollow piles be used for the possibility of repeating the air cooling during operation and maintenance of the buildings.

Very good results were obtained when cooling plastically-frozen ground near pile clusters. In this case, the primarily cooled zone with a radius r must be cooled enough to ensure that in the first year after cooling the temperatures along the outer edge of the piles will not exceed t_T for the given soil. The rate of increase in the temperature in the cooled zone of cylindrical shape depends on $F_0 = \alpha\tau/r^2$, the relative radius of the pile cluster $\rho = r_0/r$, the relative temperatures $\Theta = (t_0 - t_r)/(t_0 - t_H)$ and the Biot number--Bi.

When developing the calculation method of determining the increase in the ground temperature within the limits of cylindrical zones in the first approximation, it was assumed that the heat transfer in the process of raising the temperatures is constant. This follows from the known assumption of Kh. R. Khakimov concerning the constancy of the relative thermal effect, A, in the ground outside the limits of the frozen zone. If we make this assumption, then the Biot number $Bi = 1/ln\ A \doteq 1$. As calculations proved, the deviations are 10-15 percent with a variation of r within broad limits. Later the variation in the temperature gradient at the boundary of the solidly frozen zone was analyzed on the basis of the exact solution:

$$\frac{dt}{d\tau} = \frac{2(t_0 - t_r)}{r} \int_0^{\infty} \exp\left(\frac{-\alpha\tau}{r^2} v\right) \frac{y_1(V)N_0(V) - y_0(V)N_1(V)}{y_0^2(V) + N_0^2(V)}\, dv = \frac{C_0(t_0 - t_\tau)}{r}. \quad (16)$$

TABLE 2

F_0	0.05	0.1	0.5	1	2	5	10
C_0	3.05	2.3	1.25	1	0.8	0.6	0.55

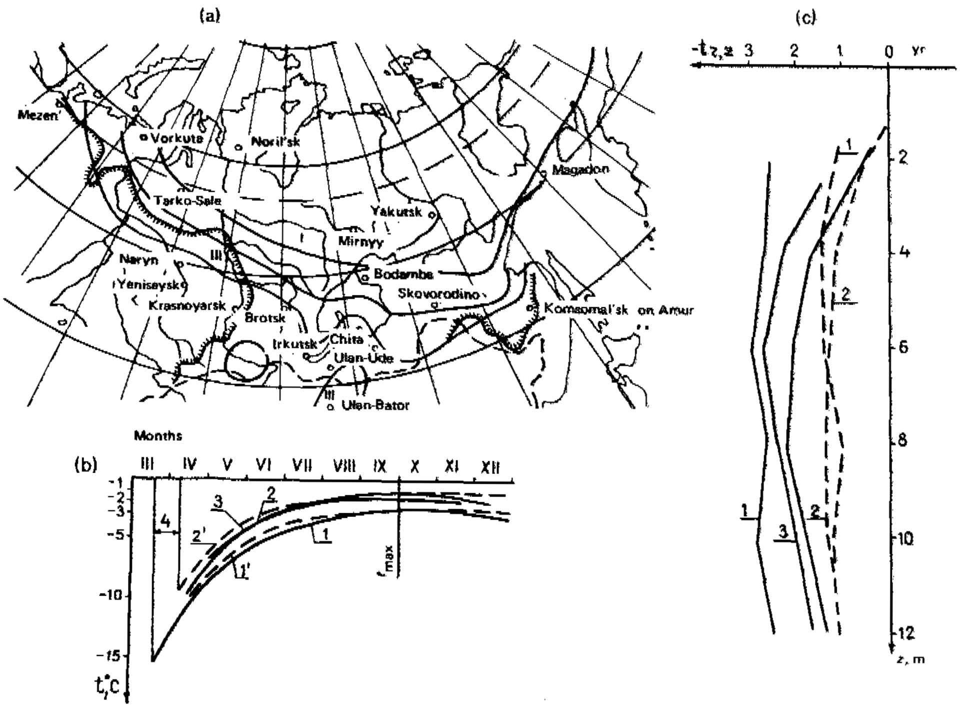

FIGURE 4 (a) Calculated map of the applicability of one-time cooling for various structural arrangements of the air spaces; (b) increase in temperature at a depth z = 6 m in the adfreeze zone of the piles of a cluster after cessation of artificial cooling. 1--observed data for a foundation, the piles of which were ventilated for 24 days, 1'--the same data from the developed calculation; 2 and 3--observed data for a foundation, the piles of which were ventilated for 12 days; 2'--calculated data. (c) Maximum temperatures in the adfreeze zone of the piles of the cluster: continuous line--cooled 24 days; dotted line--the same for 12 days: 1--1966; 2--1968; 3--1970.

TABLE 3

Temperature-Rise Processes	Possible variations of the parameters			F_0	C_0	C_{av}
	α	r,m	τ, months			
Beginning	0.0025	2	2	0.90	0.04	1.16
End	0.0015	4	7	0.47	1.28	

The values obtained for C_0 as a function of F_0 are given in Table 2.

For pile clusters, the problem of variations of F_0 during the period of an increase in temperature in the cooled zone and the values of C_0 are presented in Table 3. Here, it is taken into account that the radius of the cooled zone--r--almost doubles during the process of the temperature rise.

Thus, the assumption of steady-state condition of heat transfer can be made. This greatly facilitates the calculations. It is better to do the calculations by the step method, since it is necessary to consider the increase in the radius of the cooled zone--r_T. Tables and graphs were compiled for facilitation of the calculations. In Figure 4b the observations of the increase in temperatures were compared with the calculated values. The comparison is entirely satisfactory. In one case the primary cooling is carried out in a radius of 2 m, and in the other case, 3 m. In Figure 4c we have the maximum temperatures observed in the zone of pile cluster for 5 yr of operation and maintenance of the project. The ground temperatures at the site fluctuated from -0.2°C to -0.4°C. After cooling in the zone with a radius of r_T = 3 m, t_0 = -1.5°C to -2.0°C, and from r_T = 2 m, t_0 = -1.3°C. For the ground at this site, t_T = -1.3°; therefore, the cooling in the radius r_T = 3 m must be assumed to be excessive.

A further increase in the radius of the cooling zone by 1 m required twice as much time. Cooling of the ground to the radius of r_T = 2 m was attained in 12 days.

The pile cluster in the cooled plastically frozen ground takes loads up to 500 tons. An approximate scheme is presented[6] for their application in a modern 16-story residential building.

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[Note: "Osnovaniya, Fundamenty i Mekhanika Gruntov" is available in the United States as: "Soil Mechanics and Foundations," published by Consultants' Bureau.]

BEARING CAPACITY OF PILES IN PERMAFROST

K. F. MARKIN LenZNIIEP Institute, YU. O. TARGULYAN Scientific Research Institute of Foundations and Underground Structures

The studies of the bearing capacity of piles invesitaged in this report were made at six construction sites located in different parts of the Arctic. Some of the experimental piles were made of reinforced concrete and did not differ from those used as the foundations of the buildings. The others were made of steel pipe ϕ 320 × 10 equipped with sensors permitting stress measurements in the walls.

The readings of the sensors were recorded on the tape of the EPP-09 potentiometer. The piles designed for testing under horizontal load were also equipped with a special measuring stand on which deflectometers (0.01 mm) were installed with wires attached to the pile walls at different heights with a spacing of 300 mm.

PERFORMANCE OF A PILE UNDER HORIZONTAL LOAD

The interaction of a horizontally loaded pile with the ground was investigated under the assumption of a relation between the horizontal displacement of the pile axis y to the pressure q on its lateral surface in the form of the well-known differential equation:

$$EI\frac{d^4y}{dz^4} = q, \qquad q = ky, \tag{1}$$

where EI is the rigidity of the pile; z is the depth from the surface; k is the modulus of lateral subgrade reaction, which varies with respect to depth.

The solutions of Equation (1) give values of the transverse force, the bending moment, and the angle of rotation of the pile

$$EI\frac{d^3y}{dz^3} = -Q$$

$$EI\frac{d^2y}{dz^2} = M \qquad (2)$$

$$\frac{dy}{dz} = \phi$$

From Equations (1) and (2), it follows that in determining the only unknown parameter k it is sufficient to obtain the distribution with respect to depth of any derivative of y in the experiment. In the experiments performed, not one but two values were measured for control--the displacements and the bending moments $M = \sigma.w$. By double graphical differentiation of the curve $M(z)$, the graphs were constructed for the distribution of the reaction pressure with respect to length of the pile, and then, the parameter $k = q/y$ (Figure 1).

Similar results of testing other experimental piles made it possible to discover the relation of k to depth for a mean annual ground temperature of $t_0 = -4°C$ (Figure 2). Using the data on the temperature distribution with respect to depth, it is possible to determine the dependence of k for clay soil on the temperature:

t, °C	-0.5	-1.0	-2.0
k, kg/cm^2	800	2600	4000

The test results for k permit analysis of the interaction of the piles with permafrost and the drawing of the following conclusions:

Under the action of a horizontal static load the shape of the bent axis of the pile and the bending moment diagram had the forms of Figure 1 in all experiments. Immediately after loading, the maximum moment M_{max} was at a depth of 1.0-1.5 diameters below the boundary of the thawing layer. Then as the stresses relaxed in the upper zone of the permafrost, the moment diagram smoothed off with a decrease in the value of M_{max} and its displacement downward.

In sandy and coarse soils with significant thickness of the seasonally thawed layer and increased modulus k of the thawed ground or with less rigidity of the pile, the shape of the bent axis can have a complete wave within the limits of the thawed layer, and in this case M_{max} will be located above the boundary of the permafrost. With sign-variable and recurrent loading, the increase in deformation takes place with deceleration. Just as with constant loading, the relaxation processes are observed for several days; however, in practice the perceptible increments in the deflections and stresses are noted for the first 8-12 h.

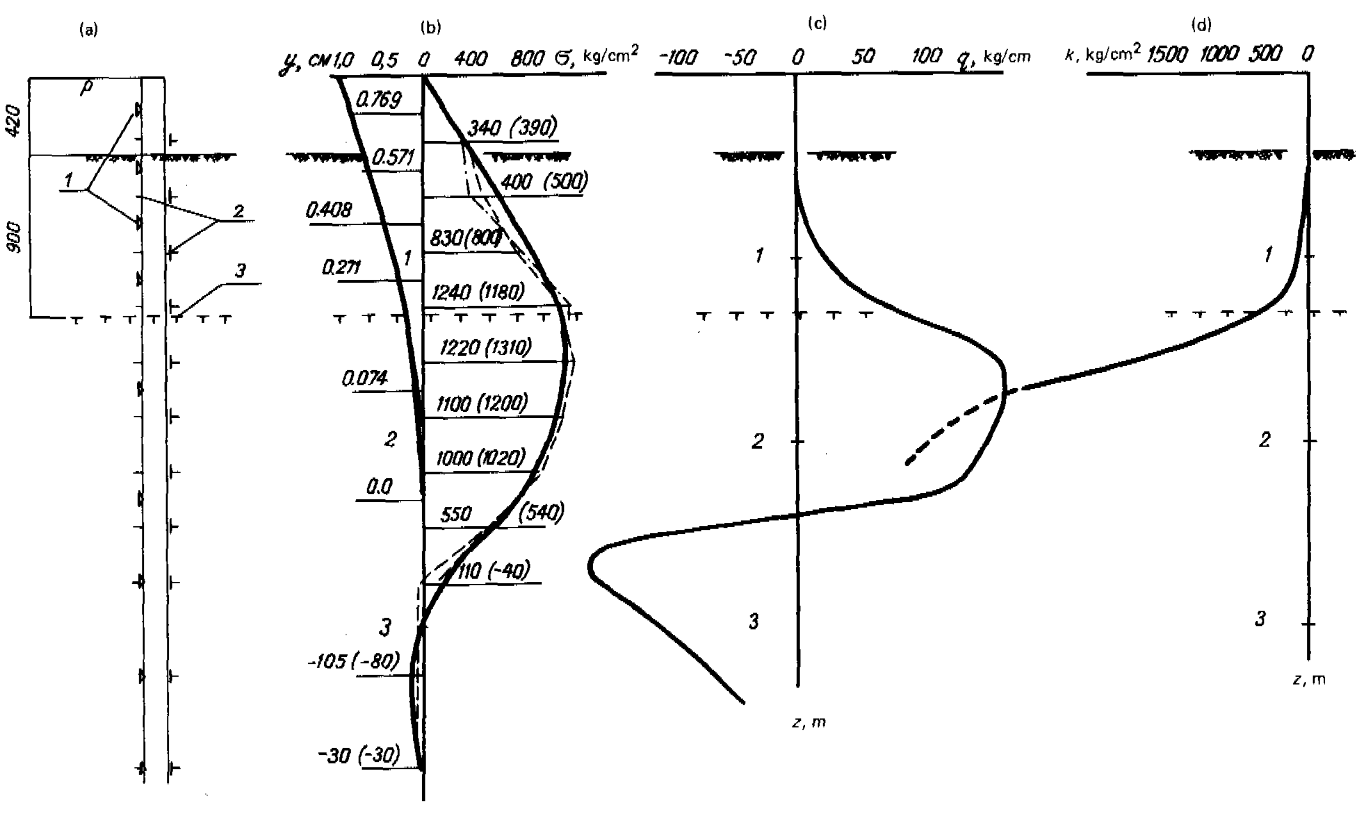

FIGURE 1 Results of testing piles under horizontal loads.

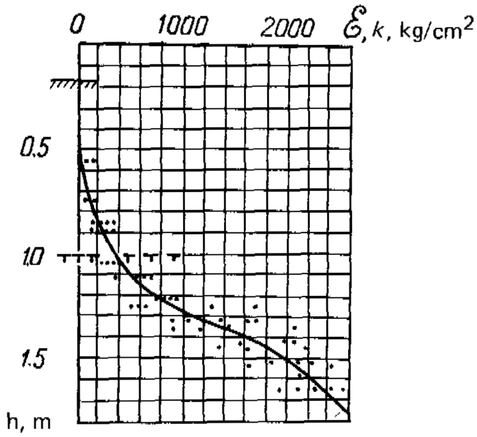

FIGURE 2 Distribution of modulus of lateral subgrade reaction with respect to depth.

Significant overloading (to 9 tonnes) caused displacements reaching 18-20 mm at the level of the ground surface. Here, the stresses in the pile walls did not exceed the design stress (2,100 kg/cm^2).

Simultaneous loading of the piles with a spacing of 60 and 100 cm in the direction of the effective force causes an increase in deflection with an increase in length of the deformed section by comparison with the individual piles; however, the maximum values of the bending moments do not increase in this case. Inasmuch as the quantitative difference in the stresses is small, there is no necessity for making any corrections to the calculations when planning and designing the pile caps.

The studies that were performed offered the possibility of developing a procedure for simplified design of the piles for horizontal loading in which the ground was represented by a two-layer medium: the seasonally-thawed layer and the permafrost layer--with a deformation modulus k constant within the limits of each layer. The divergence of the calculation results by this procedure and the actual data does not permit the divergence admissible within engineering practice. The procedure used for these calculations, plus the auxiliary graphs and tables, are presented in Dokuchaev and Markin.[1]

STUDIES OF THE OPERATION OF PILES UNDER AXIAL LOAD

The field tests under vertical loads were performed on "frozen-in" piles (installed in drill holes and slurried in). The average temperature with respect to depth at the individual sites was -2.2°C to -5°C. In all cases it was established that the processes of deformation of the ground are subject to laws characteristic of plasto-viscous bodies. The settlement of the piles is depicted by creep curves with clearly expressed stages of damped creep, steady-state flow, and progressive deformation (Figure 3). This nature of the deformation permits description of it using the laws used to describe the process in the simplest stressed states.[2] The relation between the settlement and the loading of the piles is defined by the expression

$$\sigma = A_\tau \varepsilon^m, \tag{3}$$

where σ is the nominal stress equal to the load N divided by the area of the lateral surface of the pile, kg/cm^2; A_τ is a parameter that is variable in time and reflects the deformability of the pile and ground system, kg/cm^2; ε is the settlement of the pile S referenced to its perimeter; m is the nondimensional hardening coefficient ($m < 1$) taken as independent of time.

The dependence of the parameter A_τ on the time τ and temperature is expressed by the formula

$$A_\tau = \varepsilon \tau^{-\alpha} \tag{4}$$

where ε is a parameter that depends on the permafrost temperature averaged with respect to depth t_a, kg·h $^\alpha$/cm^2; α, < 1, is a dimensionless parameter,

$$\varepsilon = \omega(t_a + 1)^\kappa, \tag{5}$$

ω is a parameter with dimensionality kg·h$^\alpha$/cm^2·deg$^\kappa$; κ is a dimensionless parameter.

The parameters A_τ, m are defined by the results of testing the experimental pile by step loading, and the parameters ε and α by the creep curve for the second pile located on the same pile site [but] under constant load. The procedure used in this analysis is analogous to that used when processing the results of studying soil samples in the simplest stressed state.[2] The bearing capacity of the experimental piles was established with respect to the intersection of the rheologic curve (the settlement-rate and load) with the x axis and also by the inflection point of the settlement-load curve constructed with logarithmic coordinates.

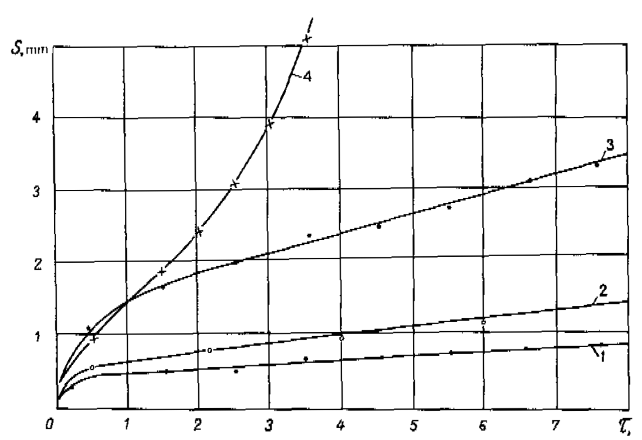

FIGURE 3 Settlement of experimental piles with time.

The deviation of the temperature of the ground measured at the time of running the tests from the maximum is taken into account when determining the calculated bearing capacity of the pile (ϕ_{calc})

$$\phi_c = \phi_0 \frac{k_i m_i \Sigma R_{s.i}{}^N F_{s.i} + k_2 m_2 R^N F}{\Sigma R_{s.i} F_{s.i} + RF} \quad (6)$$

where ϕ_0 is the bearing capacity of the pile during the tests; k_1, k_2, m_1, m_2 are the homogeneity coefficients and the conditions of the soil used according to the existing Norms;[3] $R_{s.i}{}^N$ is the Norm shear strength of the soil along the lateral surface of the pile in kg/cm^2 for the middle of the i-th layer taken according to the Norms for the calculated temperature of the bearing ground established by the data over many years; F_{si} is the shear surface area of the i-th layer, cm^2; R^N is the Norm resistance of the frozen ground to normal pressure in kg/cm^2 under the tip of the pile for the calculated temperature; $R_{s.i}$, R are the same as $R_{s.i}{}^N$, R^N but for temperatures at the time of taking the tests; F is the area of the transverse cross section of the pile, cm^2.

The parameters obtained during testing of the piles were as follows:

	t,°C	T,h	A_τ,kg/cm^2	m
Metal piles Ø325 × 10 (step loading with "rest" between steps)	-2.9	0.5	7.0	0.258
	-2.9	1.0	6.3	0.218
	-2.9	2.0	5.7	0.215
	-2.9	4.0	5.5	0.218
	-2.9	8.0	4.9	0.214
Reinforced concrete piles 320 × 320 (step loading)	-2.2	0	7.6	0.237
	-2.2	12	9.0	0.286
	-2.2	72	9.0	0.308

The values of the parameters ε and α for the first group of piles are equal to 9.5 kg·h$^\alpha$/cm^2 and 0.75, respectively; for the second group they are 7.5 kg·h$^\alpha$/cm^2 and 0.082, respectively.

Using the generalized parameters determined when taking the tests by the described procedure, it is possible to calculate the settlement of the pile foundations as a function of the duration of the load. In the case of short-term loading, the duration of which does not exceed the duration of one step of the load during the tests, the calculation is performed by the formula

$$S = U\left(\frac{N}{uh_m} \cdot \frac{1}{A_\tau}\right)^{\frac{1}{m}} . \quad (7)$$

If the load is active for a longer time, but does not exceed the period during which the temperature of the bearing ground does not change significantly, the expression for the settlement is obtained from Formula (4),

$$S = U\left(\frac{N}{uh_m} \cdot \frac{\tau^\alpha}{\xi}\right)^{\frac{1}{m}} . \quad (8)$$

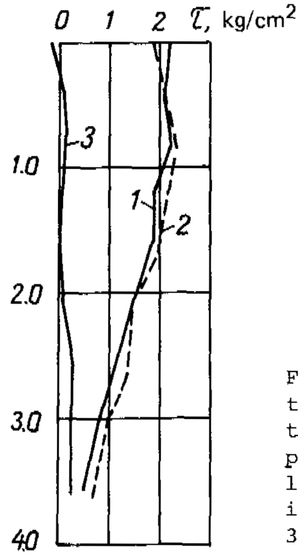

FIGURE 4 Distribution of tangential stresses along the lateral surface of the pile loaded under an axial load. 1--5 min after loading; 2--the same after 8 h; 3--5 min after unloading.

The settlement of the pile under constant loads can be defined by the formula obtained from Expression (5),

$$S = U\left(\frac{N}{uh_m} \cdot \frac{\alpha}{\omega} \cdot I_\tau\right)^{\frac{1}{m}} , \quad (9)$$

where in addition to the previously indicated notation, I_τ is the integral of the creep function taking into account the seasonal fluctuations of the temperature averaged with respect to length of the pile in h$^\alpha$/deg$^\kappa$. The values of the parameter I_τ are given in the form of graphs in Dokuchaev and Markin.[1]

The observations of the stress distribution with respect to length of the pile under axial loads permitted establishment of a number of principles of interest for the theory of pile performance and construction practice. Thus, it is possible to consider the proposition confirmed that the stage of non-steady-state creep is caused primarily by the formation of the stressed kernel under the point of the pile, inasmuch as the distribution of the tangential stresses with respect to its lateral surface takes place in practice simultaneously with the increase in load (Figure 4). On pulling the pile, the first stage--the non-steady-state creep stage--also does not appear.

When unloading the piles, the decrease in deformations takes place synchronously with reducing the load and does not exceed the values corresponding to the elastic deformations of the pile. This indicates irreversibility of the deformations of the bearing ground.

One of the important results in practice obtained during the studies is the recovery of the bearing capacity of the bearing ground lost after removing the pile. Eleven months after extracting the pile without load almost complete recovery took place both of the strength of the ground under the point and adfreezing to the lateral surface of the pile.

The noted phenomenon can be explained by the fact that during the expired period temperature fluctuations took place in the ground accompanied by variation of the equilibrium state of the liquid phase in the ground and the formation of new structural bonds.

Let us note that the results of the research performed have found reflection in the Norm documents,[3] and they have served as a basis for developing a method of production testing of piles in the construction of a number of civil buildings in the Arctic. The results of the tests also permitted the introduction of significant corrections in the plans and designs for buildings under construction.

SOME RECOMMENDATIONS WITH RESPECT TO CONSTRUCTING PILE FOUNDATIONS

When placing piles in large-diameter drill holes, the freeze-back time of the piles and the bearing capacity depend on the amount, composition, and moisture of the slurry filling the spaces between the walls of the drill hole and the pile.

Usually when constructing pile foundations, a drilling mud is used; however, the freezing point of sandy soil is higher than that of clay soil (0°C, -0.3°C, and -0.4°C), and the moisture content of the sandy soil is much lower than that of the clay soil (15-20 percent and 30-35 percent). Therefore, when using sand slurry instead of drilling mud, the time required to back-freeze the pile is reduced.

An increase in the bearing capacity of the piles in clay soil can be achieved by using a sand slurry, the adfreezing strength of which to the pile, when poured with vibration, is 1.5 (and sometimes more) times the adfreeze to the clay.

The use of drilling mud from an adjacent drilled hole instead of the special (of the required consistency) slurry for pouring in the hole (w = 30 to 35 percent) lowers the bearing capacity of the pile inadmissibly. Thus, for production testing of the piles at the site of the residential house in Anadyr made up of clay ice-rich soil where the mean water content of the mud poured in the holes reached 200-250 percent, the bearing capacity of the piles turned out to be 40 percent lower than that calculated by the Norm characteristics for the ground temperature measured during the tests. Inasmuch as these tests were performed on ordinary piles for discovery of the causes of their reduced bearing capacity, the drilling mud used was taken for analysis. The tests for freezing of it to the concrete surface performed at the Amderma Underground Laboratory of LenZNIIEP Institute at temperatures from -1°C to -4°C revealed a shear strength reaching only 25-60 percent of the Norm value.[4]

The tangential strength of the soil with respect to the lateral surface of metal piles was 20-30 percent below the Norm values for reinforced concrete and wooden piles; therefore, it is efficient to subject the metal piles to a special treatment: increase the roughness of the surface or weld on ridges (coils), coat them with a mixture of polymer materials with sand or cement, etc.

The most labor-consuming part of the operations with pile foundations is drilling the holes in permafrost. One of the means of lowering the cost and labor consumption of the operations is improvement of the methods of drilling the holes and replacing the percussion-cable drilling by more efficient percussion-rotational, thermomechanical, rotational, and leader drilling.[5]

The thawing of the ground permits relief from the labor-consuming process of mechanical breaking of the ground and facilitates the installation of the piles by many times. The most efficient is thawing with a steam needle. However, for the technology used at the present time, thawing of the ground is essentially a poorly controlled process. As a result of nonuniform thawing of the ground by steam needles, with respect to length, of the pile, its freezeback is also nonuniform. As a result of the presence of zones of thawed ground at the lower part of the pile, its bearing capacity for a long time can be less than calculated. The technological process of the operations is complicated significantly when emplacing piles in sandy soils and in ground with coarse inclusions. The improvement of the method of placing the piles with thawing of the ground consists primarily in regulating the dimensions of the thawing zone and in the application of a new method (drill driving) of placing the pile in a borehole smaller in cross section than the pile is. This "drill-driving" procedure for placing the piles is highly efficient as causing the least disturbance of the temperature regime of the ground, requiring minimum (no more than 1 day) delay after driving and improving the bearing capacity by comparison with piles installed in holes of greater diameter by 1.5-2.0 times.[6] However, of the mechanical methods of drilling holes for the "drill-driven" piles, only the rotational and leader methods which give smooth walls of the hole are suitable.

A significant increase in effectiveness of utilizing the plastically frozen soil as the bearing ground can be achieved if the ground is cooled initially, converting it to the solidly frozen state.

Depending on the time of year, the size and construction time, the purpose of the building under construction, or the structures and technical possibilities, the cooling of the frozen ground can be realized by one of the following procedures: by regulating the heat-exchange conditions at the surface, by outside air through holes prepared for installing the piles, by outside air through specially drilled holes, by "dry ice," and by refrigeration units.[7,8]

In order to cool the ground during the process of construction and subsequent maintenance, liquid and vapor-liquid refrigeration units can be used.[9]

As a result of the reduction in temperature of the permafrost during planning and design of buildings and structures, the amount of work, the cost of the pile foundations, and the construction time can be more than cut in half.

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GROUTED [PREPACKED] PILE FOUNDATIONS ON PERMAFROST

I. YA. EPSHTEIN Noril'sk Zavenyagin Mining and Metallurgical Combine

The discovery in the Far North of the USSR of the very rich deposits of minerals has brought about the construction of industrial enterprises and modern cities in these regions. In place of the previously predominant small and medium enterprises, small buildings and structures with a flexible structural design and small loads on the foundations, there are now large complex industrial complexes and multistory buildings with rigid structural designs.

The requirements on the foundations, the reliability of their performance under various conditions, the significant increase in bearing capacity of the foundations, and industrialness and economy of their construction have correspondingly increased sharply. In the reaches of the Far North, including the City of Noril'sk, in the past decade a sharp necessity has arisen for improving the methods and procedures of foundation construction.[1] Traditional methods and procedures for foundation construction were used to build the first stage of the combine and the city.

Many years of experience in construction, operation, and maintenance of buildings on such foundations have revealed a number of inherent deficiencies: the tendency toward being deformed on nonrock ground[2,3] and also on rock[1] as a result, during the operation of the building, of the thawing of interlayers of ice, which cemented the rock together, has not been excluded; there is comparatively low bearing capacity and correspondingly high specific labor consumption (per ton of load); the time required for construction and high costs are typical. In addition, the condition of density of construction, especially important in the Far North, has been violated. This pertains to the large industrial buildings for which previously sites were found with shallow depth of bedrock that predetermined the scattering of the sites even technologically connected with each other. The sites with plastically frozen ground have been left undeveloped in exactly the same way when building the city.

In the search for better solutions with regard to foundation construction in Noril'sk, in the last decade universal type "pile-injection [grouted pile] foundations" have been introduced and developed.

In view of the difference in purpose of the structures, the performance of the foundations and the principles of use of the ground as a bearing surface, two types of such foundations have been developed and introduced: (a) injected "frozen-in" point-bearing piles using the bearing soil in the thawed and thawing state (principle II) as the basic solution in industrial construction (Figure 1a); (b) injected thermally expanded piles with use of the bearing soil in the frozen state (principle I)--predominantly in civil construction (Figure 1b).[4,5]

The grouted pile can be constructed as a kind of driven pile.

Foundations have been built by the developed technological process in the following sequence:

1. The boreholes for installing the piles were drilled using the "BS-1m" percussion-cable drill without restricting the depth and the rotary thermomechanical ABU drill to a depth of 11 m. The depth is limited by the length of the drill string (12.0 m). The operating element of the ABU drill comprises a combined drilling tool (a burner with a crown or milling cutter) at the lower part of the drill string. The burner and the drilling tool can operate separately (thermal or mechanical drilling) or simultaneously (thermomechanical drilling). The removal of the drilling rod is pneumatic. The capacity of the drill in thermomechanical drilling is 50 running meters, in a mean shift of 7 h.

2. The installation of precast piles with electrodes and resistance thermometers attached to them in the drill holes. The injection pipe was lowered into the drill hole simultaneously.

3. Filling the free space in the borehole with dry sand-gravel mixture. The fill is made with a crane with the help of an inclined chute with periodic operation of a vibrator clamped to the head of the pile. When setting prepacked piles, the entire volume of the drill hole is filled with dry concrete mix after installing the grouting pipe.

4. Injection of the cement grout. In order to create a strong concrete bond tightly fastening the pile in the drill hole and also for an expanded concrete base in thermally expanded piles, sand-free grout is pumped under pressure (to 30 atm) into the well through the injection pipe. It is prepared locally in a light portable device equipped with a simple mixer, pumps, and a transformer. Injection lines 38-50 mm in diameter were used. In the lower part of the pipe there are four slots up to 10 cm high and 1-1.5 cm wide. On the upper end the manometer for monitoring the pressure is attached to the end and a grout hose is attached to the lateral connecting pipe. The grout pumped downward into the hole rises, forcing out the water and fine particles of soil; it makes the bond and expanded footing tightly monolithic and, in addition, seals the cracks and pores in the bearing ground.

In the case of sound bed rock in dry boreholes, a sand-cement grout was also used for injection or pouring into the hole (with a reduction of the design bearing capacity of the pile).

5. Electrical heating. On completion of grouting, the electrodes were connected to the transformer. The concrete temperature was measured at the point of insertion of the resistance thermometers and recorded remotely.

After 3 days of electrical heating, the concrete, as a rule, reached the design strength, which gave the possibility of loading the foundation completely. Instead of electrical heating, for technicoeconomic considerations, antifreeze additives were used.

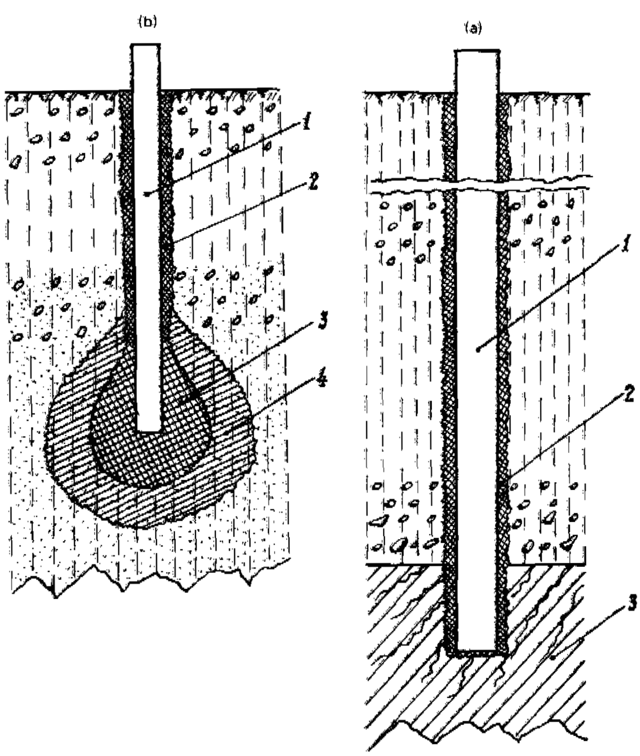

FIGURE 1 Types of grouted pile foundations. (a) Frozen-in pile; 1--reinforced concrete drilled pile; 2--grouted concrete bonding; 3--supporting bed (fissured bedrock). (b) Thermally expanded pile; 1--reinforced concrete drilled pile; 2--concrete bonding; 3--expanded base; 4--bearing ground strengthened by injection.

In the frozen-in piles in Noril'sk large mining, metallurgical, and other enterprises have been built and expanded in the last decade with hot and wet technological processes equipped with complex equipment and sensitive instruments. The enterprises include building blocks of significant area measured in hectares and up to 40 m high and more (e.g., a sintering plant) with loads on the foundation up to 2,000 tonnes and on a pile to 500 tonnes (the shop for separating the nickel matte at the nickel plant, the Komsomol'skiy mine, and so on). Most of the foundations were constructed deeply (to 25 m) and superdeeply (to 50 m) with great flexibility of the pile, with a ratio of l/d to 80 and l/b to 120. Because of the design load, reinforced concrete piles were used predominantly: with a borehole depth to 15 m, 32 × 32 cm in cross section with chamfers in holes of 45-50 cm in diameter; with a depth to 50 m, tubular piles 60-40 cm in diameter with boreholes 70-80 cm in diameter and, in particular, square 40 × 40 cm with chamfers in boreholes 50-60 cm in diameter.

Large industrial projects have been built on sites with the most varied permafrost conditions; the bearing ground includes gabbro-diabases, limestone, dolomite, gypsum, shale, etc., with temperatures from -3°C to 5°C to 43°C (the roasting-reduction ship of the nickel plant, as a result of the thermal effect of hot process water).

During construction in various years, tests have been run more than once by static loading of the frozen-in piles in different types of soil after freezing. The test data for characteristic piles are presented in Table 1 and Figure 2.

The load for an experimental pile in certain cases was a stack of piles and, in others, the weight of the above-ground part of the building. Under a load of 370 tonnes, pile no. 2 cracked. In the remaining cases the limiting load was not reached. The tests were stopped when cracks appeared in a reinforced concrete beam of the pile cap operating as a reaction beam when testing the tubular pile.[6]

The bearing capacity of the rigidly fastened piles was eventually higher (especially on fissured bedrock and semirocky soil) than that of the tamped piles, the shell piles, and the point-bearing piles put in rocky soil and designed by the Norm, in effect now and during the construction of the foundations.[5]

In the case of thermomechanical drilling and also percussion cable drilling with the application of hot water during penetration of the pile into fissured rock and coarse soils, or when using steam after filling the hole simultaneously with drilling, the ice melts out and the cracks and pores in the ground are unsealed, which improves the conditions for effective grouting. During injection these cracks and pores are cemented, and the supporting ground becomes dense and strong. The bearing capacity of the point-bearing piles with respect to conditions of resistance of the material is also higher as compared with tamped piles. This is achieved by qualitative peculiarities of the new method of foundation construction, where simultaneously with strengthening of the ground the concrete foundation is also strengthened. During the concreting by the ascending-flow method under pressure, cavities and unpacked sections in the concrete are eliminated, the water and fine particles, which are detrimental to the granulometric composition of the concrete, are forced out of the hole, and the pile is fastened tightly in the ground. The grouted piles are more reliable and more technologically satisfactory than the "drilled-driven" [pre-

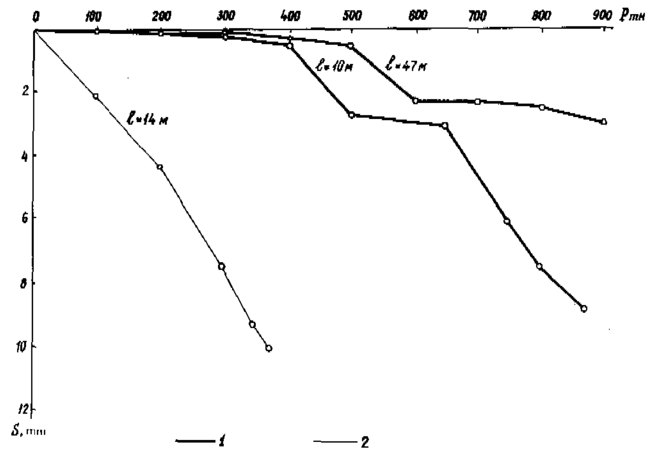

FIGURE 2 Settlement of tested operating piles as a function of load. 1--60-40-cm tubular pile; 2--32-x32-cm square pile.

packed] piles that have been used. The process of constructing these piles is continuous and can be automated.

In the design it is possible to assume that the frozen-in pile operates as a driven pile, but more reliably, and is fastened in incompressible ground to a great, previously determined depth.

The studies made and construction practice provide a basis for determining the bearing capacity with respect to the ground of slurried-in piles by the formula

$$P = kmR^{N}F, \qquad (1)$$

where km is assumed equal to 0.8; R^{N} for rocky soil is assumed equal to P_{comp}, but not less than 200 kg/cm^2, and for semirocky and coarse soil with a sandy fill R^{N} = 200 kg/cm^2; F is the area of the cross section of the lower part of the hole.

The cohesion of the bonding to the walls of the borehole and the supporting ground must provide the required resistance to inflow of the soil into the annulus surrounding the pile. In order to prevent inflow it is possible to determine the calculated depth of fastening of the pile to the supporting ground by the formula

$$h_3 = \frac{\mu D}{4} \cdot \frac{R^{N}}{R_S^{N}} \qquad (2)$$

where μ is Poisson's ratio of the supporting ground; D is the diameter of the bottom of the borehole; R^{N} is the Norm resistance of the base soil to compression; R_S^{N} is the Norm shear strength of the lateral surface of the concrete bonding with respect to the supporting ground.

In practice, it was assumed that h_3 = 2 D in rock.

The bearing capacity of the pile as a column is determined by the material. When bonding is by the grouting method for the entire depth of the borehole, the pile in ground of massive texture is designed taking into account the

TABLE 1

Test Year	Object	Soil in the Foundations	Initial Ground Temp., °C	Pile Cross Section, cm	Length of Pile, *l*, m	Maximum load			Total Settlement, mm
						Pile Flexibility *l/b*, *l/d*	Per Pile, tonnes	Per cm^2 of Pile Cross Section, kg/cm^2	
1961	Experimental	Fissured gabbro-diabase	-0	15 × 15	6.0	40	85.2	378	3.4
1964	Cement plant	Saturated limestone	-1	32 × 32	14.0	44	375	370	9.8
1965	Nickel matte separation shop	Fissured limestone	-1	Tubular 60 × 40	10.0	16	875	300	8.9
1971	Komsomol'skiy mine	Basalt, fissured at the surface		60 × 40	47.0	78	900	320	2.9

longitudinal bending only of the upper free part of it plus the magnitude of the structural settlement of the ground around it. For piles operating in ground of nonmassive texture, the design length is assumed equal to 0.5 l_0 with cross capping beams and 0.7 l_0 for single-row beams. The bearing capacity of the rigidly held piles, as practice has shown, is limited by the pile strength and not by the resistance of the bearing ground.

The comparative analyses of cost and labor performed on rigidly fastened point-bearing piles demonstrated that, for the same bearing capacity and under other equal conditions, the cost of the pier foundations exceeds that of the grouted ones by 2-5 times, and the labor consumption, by 5-10 times; and it exceeds the pile foundations freely supported on the bearing ground by 2-5 and 2-4 times, respectively.

In 1971 a special interdepartmental commission made an investigation of all of the existing basic mining, metallurgical, and other industrial enterprises of the Noril'sk Combine (96 projects) constructed in the last decade on frozen-in point-bearing piles, and it established the absence of any signs of deformation of the foundations and the structures erected on them.

In tubular construction, where it is simple to retain the frozen state of the ground of the sites with permafrost of various composition and structure, ordinary frozen-in piles have been justified in Noril'sk when building a ventilated air space not less than 1.5 m high. However, the problem of the foundation construction on plastically frozen ground, and especially icy, granular soil with coarse interlayers, has remained unsolved. In such ground nondamped creep deformations can occur. For these conditions, positive results have been obtained from injecting thermally expanded piles,[5] during the construction of which the bearing ground was simultaneously strengthened, the pile foundation was strengthened, the interaction of the bearing ground and the foundation was improved, and the reliability was improved in comparison with piers and other piles.

Foundations were constructed by the technological process discussed above. The drilling was done using the ABU unit. The borehole diameter was 45-50 cm. The borehole was drilled by thermomechanical means, and the required diameter was expanded by thermal means. Hot gases with a temperature up to 900°C heated the walls of the borehole, and a thin film was created which simultaneously strengthened the soil surface and lowered the possibility of collapse.

In the case of flame drilling of a borehole, the ice and water in the pores and cracks of the ground is converted to steam under the effect of the hot gases. As this steam is violently released, the pores in the ground expand and open up. The grouting method is used to create a strong concrete expanded base and bonding, and the bearing ground is simultaneously strengthened by cementing and removal from it of the ejected mud.

Thermally expanded piles were used in civil construction, as has been pointed out, while retaining the frozen state of the bearing ground, but these piles can also be used when it is permissible to thaw the bearing ground during the use of the building if the foundation is laid below the expected thawing limit.

Before the beginning of broad application of thermally expanded piles, the Noril'sk laboratory of bearing surfaces and foundations of the Krasnoyarsk PromstroyNIIproyekt Institute performed their test. The tests were run on the test area after thawing the ground and on house no. 221 on restoration of the initial temperature. The ground was fine-grained and silty sand, overlying gravelly pebbly and clayey varieties.

The test of 32 × 32 cm piles with expansion of the diameter to 1 m by static loads gave the results shown in Table 2 and Figure 3. The limiting load on the pile reached 300 tonnes when settling 22.98 mm. When determining the long-term compressive strength of the ground, the cohesion of the pile with the ground was excluded. The loads on the piles of house no. 221 did not reach the limit because of cracks in the capping beams serving as reactions. In a number of houses, when drilling up to 4-5 m, ore-water appeared which filled the boreholes. Practice showed that, in order to ensure reliable operation of a foundation in this case and to strengthen the disturbed ground, it is necessary in the initial stage of grouting to take the injection pipe 20 cm below the bottom of the borehole. (See Table 3.)

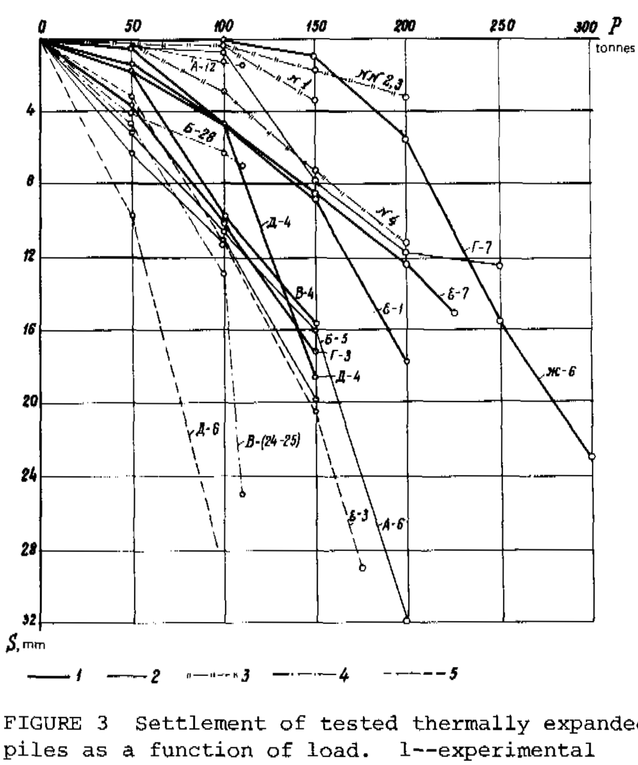

FIGURE 3 Settlement of tested thermally expanded piles as a function of load. 1--experimental injection-piles installed and tested in thawed ground; 2--also in thawed ground; 3--operating piles installed in thawed ground and tested on restoration of the permafrost (house 221); 4--the same in saturated ground; 5--experimental piles installed in concrete in thawed ground.

TABLE 2

Project	No. of Piles	Type of Thermally Expanded Pile	Thickness of Remaining Layer of Drilling Mud, cm	Bearing Ground Temperature °C		Long-term Limiting Strength of the Bearing Ground, kg/cm^2	Long-term Settlement, mm
				Initial	During the Test		
Test area	4	Injection in thawed soil	to 80	Higher than 0	Higher than 0	23.5	21.3
Test area	7	Injection in thawed plastically-frozen ground	to 40	to -0.5	Higher than 0	23.6	21.4
Test area	4	Vibration driven in concrete	to 10	to -0.5	Higher than 0	17.8	26.3
House 221, Pile 1	1	Injection in thawed ground	to 5	0	0	16.8	11.2
House 221, Pile 2	1	Injection in plastically-frozen ground	to 5	-0.4	-0.4	11.8	3.6
House 221, Pile 3	1	Injection in partly-frozen ground	0	-2.0	-2.0	16.0	2.2
House 221, Pile 4	1	Injection in thawed frozen ground	0	-2.5	-2.5	15.9	2.2

TABLE 3

Pile Number	Depth of Boreholes, m	Thickness of Layer of Remaining Drilling Mud, cm	Depth of the Pile	Load Tonnes	Stabilized Settlement, mm	Remarks
A/12	6.0	About 500	Sunk into the bottom 20 cm	120	1.33	Loads not brought to the limit as a result of insufficient strength of the capping beams serving as reactions
B/28	5.7	30	Installed on the bottom	120	7.01	
B/8	5.6	About 500	10 cm above the bottom	90	10.51	
C/9	6.6	Same	Raised above the bottom	90	12.94	
C/24-25	6.6	Same	Raised above the bottom	120	24.76	

It is possible to determine the bearing capacity of the grouted thermally expanded piles in plastically frozen, or thawed fine-grained silty, soil, including loose and saturated soil, by the formula of SNiP II-B.6-66.

$$\Phi = k_1 m_1 \sum_{i=1}^{n} R_{s \cdot i}^{N} F_{s \cdot i} + k_2 m_2 R^{N} F \qquad (3)$$

where m_1 is assumed equal to 1; $F_{s.i}$ is the shear surface area with respect to the perimeter of the borehole above the expanded base; F is the projected area of the expanded base; n is the coefficient of hardening of the soil and of the increased supporting area of the natural bearing soil.

It is possible to set nR^N in advance equal to 18 kg/cm^2. The remaining values are taken according to SNiP II-B.6-66. For the commonly used shallow foundations, the formula can be simplified by excluding the resistance due to the lateral surface,

$$\Phi = k_2 m_2 n R^{N} F.$$

It is necessary to determine more precisely the value of nR^N during field testing or take it by analogy with similar tested piles.

At a site where plastically frozen ground is mixed with solidly frozen ground, story-by-story reinforced-concrete belts at the cross-beam level along the walls of the sections on the thermally expanded piles in the buildings made of brick or blocks, and reinforced joints in the large-panel houses, were justified. It is necessary to separate the sections on the thermally expanded piles from the others by settlement joints.

The thermally expanded piles have increased resistance to being pulled out in heaving ground where their application is effective in situations of small loads on foundations such as galleries, the supports of high-voltage lines, etc.

The grouted thermally expanded piles are effective also in seismic regions. The bearing capacity of such piles is determined primarily by the resistance of the ground to compression. The shearing strength of the ground with respect to the lateral surface of the pile, significantly reduced during earthquakes, plays a secondary role, and it is possible not to take it into account in calculating the bearing capacity of the pile. Therefore, the sides of the pile can be isolated from the ground, in particular, by lubricants, which lower the direct effect of the dynamic loads on the foundation during earthquakes and improve the degree of flexibility of the pile.[7]

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INTERACTION OF FREEZING HEAVING GROUND WITH FOUNDATIONS

B. I. DALMATOV, V. D. KARLOV, I. I. TURENKO, V. M. ULITSKIY, AND V. D. KHARLAB Leningrad Engineering Construction Institute

HEAVING OF UNSATURATED SUGLINOKS

Structures erected on heaved ground are subjected to the effects of the tangential forces of frost heave, which depend on the degree of heaving of the frozen ground. This process can develop not only in saturated ground, but also in unsaturated ground.

The characteristics of the groundwater saturated moraine suglinok were the following: The degree of saturation was different so that the consistency index* varied within the limits of $0 \leq I_C \leq 0.5$; the granulometric composition was: gravel (> 2 mm): 5 percent; sand (2-0.05 mm): 30.3 percent; silt (0.05-0.002 mm): 52.1 percent; clay (> 0.002 mm), 12.6 percent; the liquid limit was 20 percent and the plastic limit, 12 percent; the unit weight was 2.75 g/cm^3. The basic mineral of the clay sections was hydromica plus, to a lesser degree, montmorillonite and iron oxide.

The laboratory studies on heave were performed with an open system, but without additonal inflow of moisture from the outside. This was achieved by freezing comparatively high suglinok samples (8.6 cm and 17.2 cm) and stopping before the desiccation zone approached the bottom of the sample. The apparatus cooled to a temperature of +3°C and the suglinok specimens were transported to the cooling chamber in which the given temperature was maintained. After the experiment, the depth of freezing was more precisely determined, the unit weights of the frozen and thawed sections of the specimen were determined, and samples were taken for moisture content at every 1.5-2.0 cm of height.

The effect of the temperature and the duration of freezing on the magnitude of the frost heave was investigated in the suglinok samples with a water content of $w \approx 15$ percent and a dry unit weight γ_d of 1.86-1.9 g/cm^3. The results of the experiments performed with multiple repetition are presented in Table 1 and Figure 1.

The magnitude of the heave and the depth of freezing of the specimens with a constant total negative degree-hours depend on the freezing temperature. The moisture content at the freezing portions of the suglinok samples with different freezing temperatures were approximately identical. Below the freezing portion there was a suction zone with a moisture content of less than the initial amount.

According to theoretical analysis,[1] the depth of suction zone at the end of the first series of experiments should be 9 mm, and the second and the third series, 13 mm and 18 mm, respectively. The experiments confirmed that the relative depth of the suction zone decreases with a reduction in the freezing temperature. The moisture gradient at the freezing limit from the direction of the thawed part of the ground was almost directly proportional to the freezing rate.

* $\left[\frac{w_L - w}{I_P}\right]$.

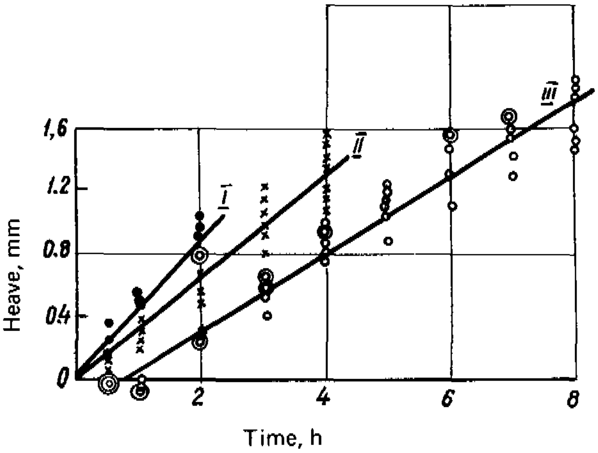

FIGURE 1 Frost heave of the suglinok samples.

Therefore, in all the experiments the moisture content of the frozen layers of suglinok and K_h were approximately identical.

In order to discover the effect of the initial density and moisture content on the magnitude of the frost heave of the unsaturated suglinok, experiments were performed with γ_d from 1.6 to 2.0 g/cm^3 and the water content of w from 11.4 to 15.7 percent, that is, for I_c from 0 to 0.5. The temperature of freezing was -5°C. The results of these experiments are presented in Table 2, from which it follows that with identical moisture content the heave increases with an increase in initial density; in the specimens with high moisture, this is exhibited to a greater degree. Moisture migration was observed in the suglinok specimens of any density with a consistency index, $I_c > 0$. When $I_c < 0$, moisture migration and heave were observed; when $\gamma_d > 1.7$ g/cm^3 and for $\gamma_d \leq 1.7$ g/cm^3, the ground experienced shrinking. A linear relation was established between K_h and the initial volumetric moisture content w_0.

The results of the experiments,[2] from the point of view of the theory of heat and moisture transfer developed by A. V. Lykov,[3] led to the conclusion that an increase in heave with an increase in density at constant moisture content w was caused by the greater degree of filling of the pores with water and intensification of the moisture migration toward the freezing front.

Field observations of the heave of this soil in fills confirmed the results obtained under laboratory conditions.

DEVELOPMENT OF TANGENTIAL HEAVE FORCES (EXPERIMENTAL STUDIES)

Inasmuch as it was established that the unsaturated suglinoks heaved, the necessity arose for investigating the effect of heave on clays, including some unsaturated soils, with freezing of the soil near the foundations and consideration of measures to decrease the tangential heave forces. The studies were performed at three sites in the Irkutsk oblast, where 25 × 25 cm experimental reinforced-concrete piles were installed to a depth of 2.8 m. As the ground froze and tangential heave forces developed, the piles were loaded so that lifting of them was prevented. The heave force was determined by the magnitude of the applied load.

The greatest heave of the soil was felt at the site made up of suglinoks containing silt fractions from 44 to 60 percent (sand: 34-50 percent; clay: 10-18 percent). The groundwater at this site was at a depth of 1.3-1.5 m before freezing, and the depth of freezing in the absence of snow cover exceeded 2 m. The heave of the piles began in the second half of November. The load applied to the pile in the winter of 1964-1965 was brought to 22 tonnes. This corresponded to a tangential heave stress of 1.1 kg/cm^2.

The many years of observations permitted experimental establishment of the magnitudes of the tangential heave stresses for three types of soil, which fluctuated from 0.3 to 1.1 kg/cm^2, and the efficiency of certain methods of reducing these stresses (Table 3).

The research performed established quite effective operation of the columnar reinforced concrete anchor piles having a supporting section located below the depth of seasonal freezing. The effectiveness of such piles depends to a significant degree on the retaining forces that develop as a result of normal pressure arising during the freezing process at the thawed-frozen ground contact and promoting and increase in the stability of the foundations. The measurements performed using an acoustic dynamometer demonstrated that the pressures at the level of the upper surface of the anchor slab increased sharply as the ground froze, and their maximum values near the base of the foundation reached 2.0-2.3 kg/cm^2. Knowing the laws of development of this pressure, it is expedient to consider it in the design when evaluating the stability of the foundations with respect to tangential heave stresses.[4]

THEORETICAL STUDY OF THE INTERACTION OF FREEZING HEAVING GROUND WITH THE VERTICAL LATERAL SURFACE OF THE FOUNDATION

In order to be able to forecast the development of the tangential heave forces, theoretical studies were made,[5-8] the basic principles of which are discussed below. The object of the study was the system made up of the frozen layer, thawed ground, and foundation (Figure 2):

1. Heaving of the ground at any point of the ground massif takes place instantaneously on passage through the given point of the freezing front.

2. Under the effect of heave forces developing in the bearing frozen layer and the reaction forces from the foundation this layer bends as a semi-infinite beam (a strip footing) or an infinite slab with a hole (a columnar foundation). The height of the beam (slab) increases with time (if the freezing front shifts).

3. The frozen ground is an elastoviscous material of the Maxwell type with Young's Modulus $E = E_0\ (1 + \alpha\Theta)$, the coefficient of viscosity

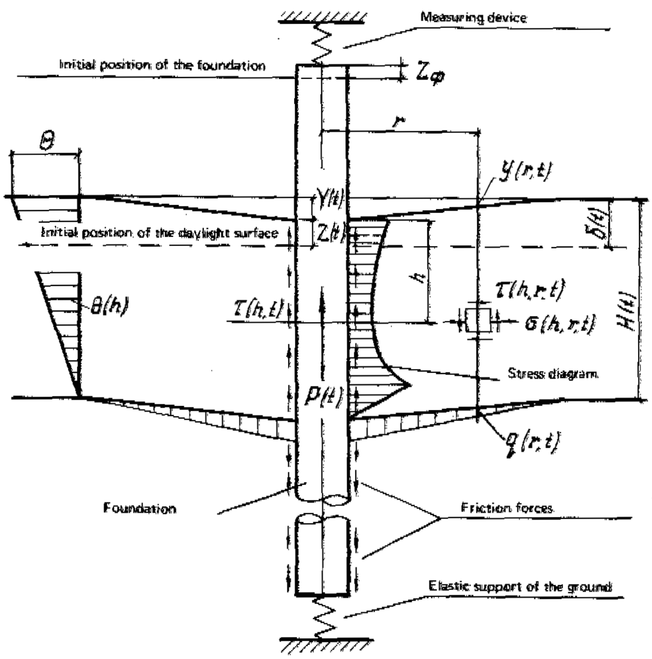

FIGURE 2 Scheme for the system made up of the frozen layer, thawed ground, and foundation.

$\lambda = \lambda_0\ (1 + \alpha\Theta)$, and the general Poisson's-Ratio ν = const, where Θ is the absolute value of the (negative) ground temperature. The parameters E_0, λ_0, and α do not depend on time and [space] coordinates.

4. The temperature Θ is distributed linearly with respect to depth of the frozen layers.

5. The thawed ground reacts to the bending of the frozen layer with a fixed value of $H(t)$ (that is, the increment in bending in an infinitesimally small time dt) like a Winkler elastic base with a constant coefficient of subgrade reaction, k.

6. The tangential stress τ at any point of the "face" of the frozen layer is connected with the displacement z of this point with respect to the foundation with the law observed in the experiments with respect to pressing the piles frozen in the ground through it[9] (Figure 3). The parameters τ_1, a, b, β, γ, ζ are considered constants. At a point of differing height on the lateral surface of the foundation, its interaction with the frozen layer is in a different stage, since a different h at the same time t corresponds to a different relative displacement.

7. The foundation may move depending on its weight, the elasticity of the foundation, and the rigidity of the supporting structures.

On the basis of assumptions 1-7, a complete system of equations of the problem was constructed, including algebraic, differential, and integral differential equations. It is used for simultaneous determination of all the most important parameters of the load-deformation conditions of the system: bending $Y(t)$ of the outside surface of the frozen layer at the foundation; bending of $y(r, t)$ at different distances from the foundation; sliding of $z(t)$ of the frozen layer along the foundation; vertical displacement of the foundation $z_\phi\ (t)$; the heave forces $P(t)$ (per unit length of the foundation perimeter); the tangential stresses $\tau(h, t)$ at all points of

TABLE 1

	Series of Experiments, Freezing Temperature, °C		
	I(-20)	II(-10)	III(-5)
Duration of the experiment, h	2	4	8
Magnitude of heave, mm	0.830	1.270	1.720
Depth of freezing, mm	38.0	51.1	70.0
Heave ratio (K_h), % [heave / depth of freezing]	2.2	2.5	2.5

TABLE 2

w,%	γ_d, g/cm^3	w_0	K_h, %	w, %	γ_d, g/cm^3	w_0	K_h, %
11.4	1.60	0.18	-0.2	13.7	1.81	0.25	1.1
11.6	1.70	0.19	-0.04	13.7	1.92	0.26	1.5
11.8	1.80	0.21	0.5	14.8	1.92	0.28	2.1
12.0	2.00	0.24	0.6	16.0	1.65	0.26	1.8
14.0	1.60	0.22	0.5	15.7	1.81	0.28	2.0
13.8	1.70	0.28	0.8				

TABLE 3

Area of Observations and Method of Reducing the Heave Forces	Years 1964-1965 Total Load, tonnes	Percent-Reduction of Total Load	1966-1967 Total Load, tonnes	Percent-reduction of Total Load
Site 1 (ground--silty suglinok)				
Filled with local soil	7.3	--	12.1	--
With an anchor slab with 60-cm bracket[a]	2.1	72	2.1	83
Salinzation of the fill with NaCl solution	2.5	66	--	--
Treatment of soil with ethyl silicate	--	--	2.5	79
Site 2 (ground--silty supes)				
Filled with local soil	7.3	--	8.5	--
With anchor slab with boom of 60 cm	2.1	72	2.1	76
Salinization of the fill with NaCl solution	6.1	17	--	--
Salinization of the fill with powdered NaCl	--	--	1.3	85
Site 3 (ground--silty suglinok saturated)				
Filled with local soil	22	--	--	--
With anchor slab with bracket 20 cm	14.7	33	--	--
The same, 40 cm	12.6	43	--	--
The same, 60 cm	6.9	69	--	--

[a]The anchor slab for the foundation was rectangular; the bracket came from the edge of the anchor.

the lateral surface of the foundation; the pressure $q(r, t)$ at the base of the frozen layer.

The initial data reqired to solve the problem are the following: the thickness of the frozen layer $H(t)$, the amount of heave far from the foundation $\delta(t)$ (or the heave factor at different depths); the experimental curve--Figure 3 (that is, its parameters τ_1, a, b, β, γ, ζ); the temperature at the surface of the frozen layer $\Theta(t)$; the characteristics of the mechanical properties of the frozen ground E_0, λ_0, α; the coefficient of subgrade reaction of the thawed ground k. By a numerical solution of the system of equations, a significant number of specific examples were fed to the computer. Some results are presented in Figure 4, and the calculations permit us to draw the following conclusions:

1. The creep of the frozen layer and, in general, its deformability are felt in the magnitude of the heave force only at shallow freezing depths (approximately to 30 cm). As a consequence, the deformability of the frozen layer turns out to be a secondary factory of the thawed base.

Thus, it is possible to neglect the bending of the frozen layer. Then, in order to determine the heave force, various simple (algebraic) formulas were obtained (the dotted line in Figure 4). There is no necessity for the four initial parameters (E_0, λ_0, α, k). In the special case where K_h = const = x and Θ = const, these formulas have the form

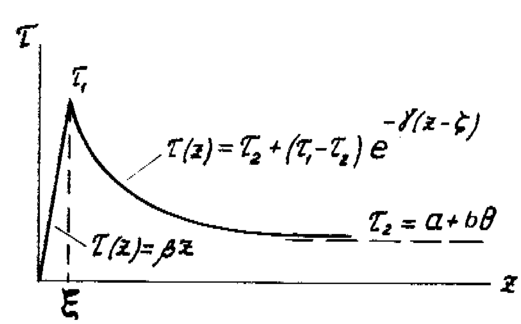

FIGURE 3 Approximation of the experimental graphs of tangential stress and displacement.

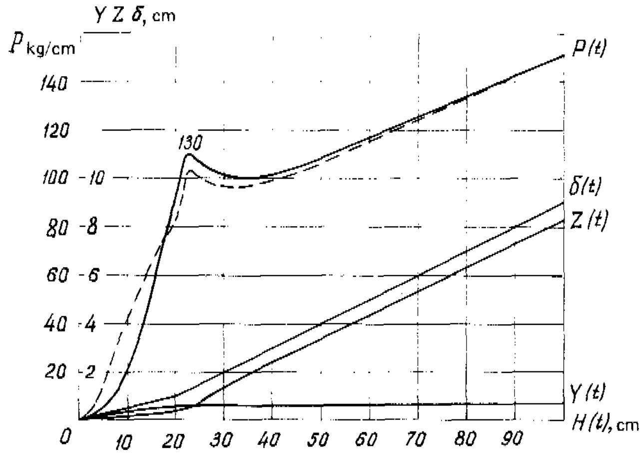

FIGURE 4 Results of the calculation.

$$P = \frac{\beta x}{2} H^2 \qquad (xH \leq \xi)$$

$$P = \left(a + \frac{b\Theta}{2}\right)H - \frac{b\Theta\xi}{x^2 H}\left(\frac{\xi}{2} + \frac{1}{\gamma}\right) + \frac{1}{x}\left(\frac{\beta\xi^2}{2} + \frac{b\Theta}{\gamma} - a\xi\right)$$

$$+\left(\frac{\tau_1 - a - b\Theta}{\gamma x} - \frac{b\Theta}{\gamma^2 x^2 H}\right)$$

$$\left[1 - e^{-\gamma(xH - \xi)}\right] \qquad (xH \geq \xi).$$

2. The experimentally observed drops in the heave forces are explained by the nonsimultaneous beginning of the process of interaction of the ground with the foundation at different points with respect to depth (from the time of arrival of the freezing front). However, the general nature of measurement of these forces can be different (in particular, its drops are not certain). The variation of the heave coefficient with depth (depending, for example, on the soil moisture) has great significance.

3. The magnitude of the heave force P depends on the heave ratio K_h. The idea that large K_h always corresponds with large P is incorrect. The dependence of P on K_h has a maximum at some K_h.

4. The distribution of the tangential stresses with height of the lateral surface of the foundation has (from some point in time) a characteristic "crest" in the lower section (see Figure 2).

5. The degree of movement of the foundation is reflected essentially in the magnitude of the heave force.

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STUDY OF THE INTERACTION OF HEAVING GROUND WITH INDIVIDUAL FOUNDATIONS

N. A. PERETRUKHIN, YU. D. DUBNOV, AND N. D. MERENKOV
TsNIIS, Ministry of Transport Construction

The studies to discover and more precisely define the laws of interaction of heaving ground with the foundations of structures have been performed at the Skovoroda permafrost station and also at the sites of production facilities using models of reinforced-concrete piles with a cross section of 10 × 10, 15 × 15, 20 × 20, and 25 × 25 cm. The models were buried to 1.0-1.5 m in an active layer of about 2.5 m, and they were also placed on the ground surface. In addition, normal foundations were used for the supports of the associated setup buried to 3.8 m.

Many years (1953-1968) of studying the process of frost heave led to the discovery that each of the elementary horizontal layers of the ground is successively at four stages of freezing, with a reduction in temperature provisionally called[1] the stages of initial freezing, active freezing, cooling, and "supercooling." The volumetric expansion of the ground and the internal stresses in the layer causing heave forces occur in the stage of active freezing. The growing layer of solidly-frozen ground in the cooling stage moves upward under the heave force.

The interaction of the foundation with the surrounding heaving ground is caused by the presence of surface action between them and also variation of the physical nature of the surface action as the state and properties of the freezing, heaving ground varies.

The results of the studies performed during 1958-1971 demonstrated that the process of interaction can provisionally be divided into three periods.[2]

The initial period is characterized by predominance of the freezing of the ground to the surface of the foundation, large magnitude of the heave stress (> 8 kg/cm^2), and insignificant total heave force (< 500 kg). The external signs of a given period of the interaction are complete immobility of the soil close to the foundation and free raising of the surface of the ground outside the contact zone.

The basic period is a continuous and irreversible upward displacement of the solidly-frozen ground, the thickness of which increases and exerts lateral pressure on the foundation, and also a constant increase in the total heave force to the maximum value under the given conditions (8 to 50 tonnes).

The concluding period begins after cessation of the displacement upward of the solidly frozen ground, and it is characterized by a steady reduction of the heave force to zero.

Out of the three periods of interaction, the basic period is the defining period. During this period, the equality of the increasing forces and reactions is maintained:

$$P = P_0, \tag{1}$$

and also the condition

$$P_0 \leq P_K, \tag{2}$$

is observed where P is the total heave force (the force of the solidly frozen ground on the stationary foundation); P_0 is the same for heave arising in the layer of ground in the stage of active freezing and causing displacement upward of the solidly frozen ground with respect to the stationary foundation; P_K is the same for the shearing strength and displacement of the solidly frozen ground along the surface of the foundation caused by the presence of contact bonds.

The total tangential heave force was measured under natural conditions using the setup shown in Figure 1. The data for many years of measurements demonstrated that the total (kg) and relative (kg/cm) heave forces during the primary period of the frost heave of the ground increase, reaching maximum values at the time of cessation

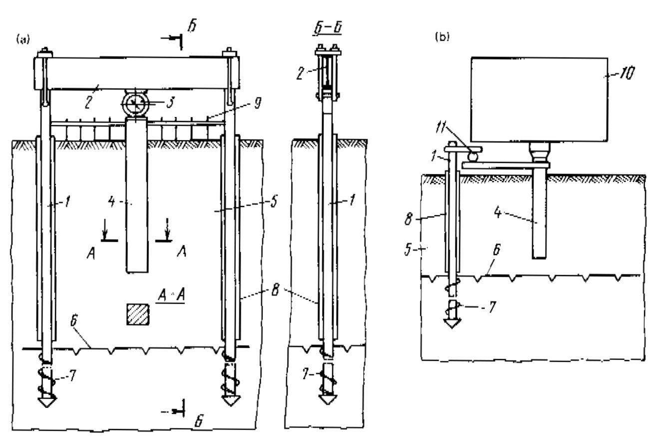

FIGURE 1 Setup for the heave-freezing tests: (a) structural designs of the TsNIIS; (b) using the method of force balancing. 1--anchor stands; 2--anchor beam; 3--measuring unit (dynamometer); 4--foundation; 5--affected active layer; 6--surface of the permafrost; 7--coil; 8--casing; 9--rods for measuring the displacement of the ground surface in the affected area; 10--box for the load balancing the heave force; 11--indicator to measure the displacement of the foundation.

of the displacement upward of the solidly frozen ground. The total heave force depends, in addition, on the transverse cross sectional dimensions and the area of the lateral surface of the foundation: With an increase in the dimensions and area, the heave force increases. The variations in the heave stress (kg/cm^2) are shown in curve (Figure 2a), which is similar to the shear strength curve.[3]

According to measurement data using duplicated setups, it was discovered that the values of the heave stress are different, not only in different years, but also in one season. Assuming that the heave force is a generalized parameter reflecting the joint effect of variable factors (the variability of the ground; its moisture, density, and temperature; the freezing conditions; the roughness of the lateral surface of the foundation, etc.), the comparable numerical values of the heave stresses obtained from the experiments were processed statistically as probability values. Here, it is established that the mean value over many years of the heave stress is 1.72 kg/cm^2 with a standard deviation of 0.22 kg/cm^2; these values were obtained for practically stationary foundations having a smooth surface.

The laboratory studies were performed by physical simulation of the basic period of the interaction during which condition (2) is observed. The similarity of the laboratory experiments to the natural process is based on the known laws of friction,[4,5] and it is ensured by the equality of the parameters characterizing the properties of the ground in nature and in the model and the conditions of this interaction.

Studies were made with a foundation during freezing on the setups of the TsNIIS[6] using the Skovoroda silty suglinok with variable values of the soil moisture (from 23 to 46 percent), the initial dry unit weight (1.5-1.7 g/cm^3), the frozen ground temperature (from -2°C to -12°C), and the displacement rate (from 0 mm/day to 20 mm/day).

The experimental results demonstrated that:

1. The adfreeze bonds are irreversible, inasmuch as they are structural-crystalline bonds and are not restored after destruction; the magnitude of the relative displacement sufficient for destruction of the adfreeze bonds does not exceed 1 mm.

2. The frictional bonds are reversible; they are formed after destruction of the adfreeze bonds following relative displacement; the displacements exist during the entire period; they do not depend on the rate and the general magnitude, and they are restored with renewal of the displacement after prolonged interruptions. The variations in the bond stresses as a function of the relative displacement (Figure 2b) are analogous to the variations of the heave stress under natural conditions(Figure 2a).

The experiments also established that the numberical values of the contact bond stress do not in practice change for different displacement rates. This characterizes the physical nature of these bonds as due to sliding friction and independent of the displacement rate.

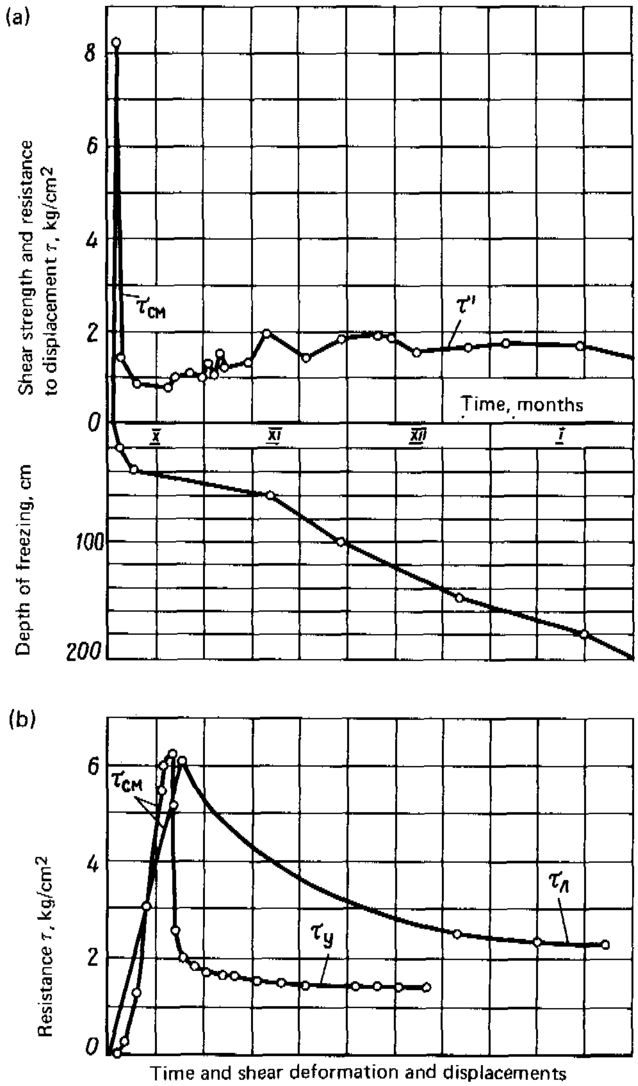

FIGURE 2 Graph of the resistance of the surface bond: (a) under natural conditions; (b) under laboratory conditions. τ_{cm}--shearing strength of the adfreeze bond; τ'--resistance of the bonds to relative displacement (tangential heave stress); τ_l--resistance of the friction bonds to relative displacement according to the data of the TsNIIS; τ_y--steady resistance to freezing.

The average resistances to displacement of the samples of solidly-frozen soils at the surface of the foundation models obtained under laboratory conditions at the temperature, initial moisture content, and density of the soil samples equal to the natural soil were compared with the average annual and mean perennial (1958-1968) values of the heave stress obtained by direct measurements under natural conditions. It turned out that the mean value obtained under laboratory conditions (1.73 kg/cm^2) is equal to the mean annual value of the heave stress (1.72 kg/cm^2). The divergences between these values and individual years do not exceed the standard deviation (0.22 kg/cm^2) calculated from statistical processing of the data over many years from measurements of the heave force under natural conditions. This comparison of the results of the laboratory tests and the measurements of the heave force under natural conditions indicates the possibility and expediency of applying the laboratory method in solving practical problems.

The interaction of the heaved ground with shallow foundations was studied under natural conditions using installations (see Figure 1a) with models of reinforced concrete foundations 10 × 10 cm and 20 × 20 cm buried to 100 cm and 150 cm or installed on the ground surface.

The research data demonstrated[7] that the occurrence and the development of the normal heave force during freezing of the ground below the base of a shallow foundation does not in practice show in the nature of the variation of the relative and total heave forces by comparison with the same characteristics during the preceding period when the foundation takes up the effect of only the tangential heave force (Figure 3).

The numerical values of the heave stress for shallow foundations were approximately determined by the formula:

$$\tau_1 = \frac{P - q}{u \cdot z_i}, \qquad (3)$$

where P is the total heave force, kg; q is the weight, kg; u is the perimeter of the foundation, cm; z_i is the thickness of the solidly-frozen layer surrounding the foundation, cm.

When the ground freezes below the base of the foundation, the total heave force P (measured by a dynamometer) was associated with the perimeter of the foundation, and it was considered as the provisional tangential heave stress τ'_G. This value was determined by Formula (3), the value z_i in which corresponded to the total thickness of the layer of solidly frozen ground, and including the layer below the base of the foundation.

The studies established that the values of τ' and τ'_G do not depend on the depth of the foundation (see Figure 3) and its perimeter. In addition, the normal heave stress referred to the area of the base of the same shallow foundations increases regularly with an increase in the thickness of solidly frozen ground. In addition, its numerical values depend on the area and transverse cross section of the foundation: With a decrease in area, the normal heave stress increases and cannot be given uniquely in the form of a Norm value.

The normal pressure of the freezing and frozen ground to the lateral surface of the foundation was measured in 1959-1963 at depths of 60 cm, 120 cm, and 180 cm. It was found that, at the time when the total heave force reaches the maximum value (the end of December to the middle of January), the normal pressure is distributed with respect to depths of the active layer in accordance with the ground temperature: In the upper part of this layer having lower temperature, the pressure turns out to be large; it decreases with depth.

The lateral pressure has an especially significant effect on the individual foundations of the supports of construction and on roadbed fills of railroads (Figure 4). In the experimental fills of silty suglinok with a moisture content of 25-

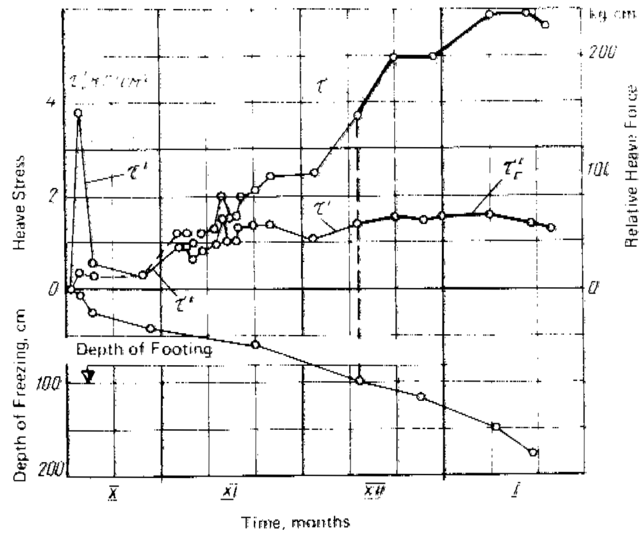

FIGURE 4 Variation with time and depth of the active layer of the tangential heave force acting on a shallow foundation. τ--relative, calculated per cm of the foundation perimeter; τ'--tangential stress during freezing of the ground above the base of the foundation; τ'_G--the same with freezing of the lower layers of the ground.

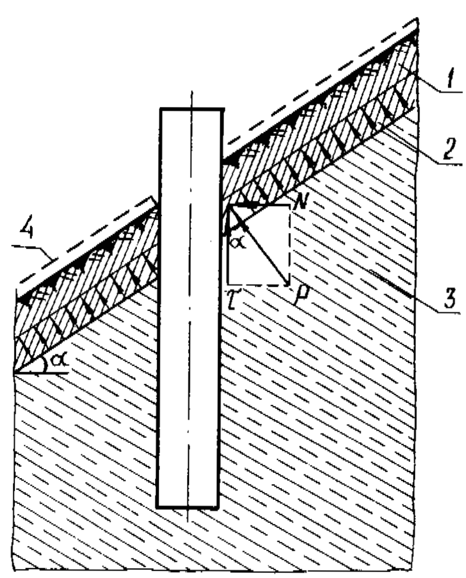

FIGURE 3 Diagram of the effect of frost heave forces on individual foundation installed on a slope. 1--layer of solidly-frozen ground; 2--layer in the stage of active freezing and heaving; 3--nonfrozen layers; 4--position of the ground surface as a result of frost heave; P--total force of frost heave; N--horizontal component of the frost heave force; τ--vertical component of the frost heave force; α--angle of inclination of the slope.

30 percent, the unit pressure on the lateral surface of the foundations and the total value of the horizontal component of the force of frost heave (*N*) were measured. The result of the measurements demonstrated that the maximum values of the unit pressure were at a depth of about 0.2 m during the period when the freezing of the ground reached about two-thirds of the thickness of the active layer. The greatest magnitude of the pressure in three seasons of observations was about 2.5 kg/cm^2, and, at a depth of 0.8 m, 1.6-2.0 kg/cm^2. It dropped to zero at a depth of about 2 m. In this case the foundations were subjected not only to heaving, but also horizontal displacement in the direction of the slope. Some displacements of the one freezing season exceeded the total magnitude of the movements permitted for the entire operating life of the supports. The large values of the total component of the heave force caused great displacements of the foundations, and the displacement rate was maintained constant in practice independently of the increase in total pressure. At the time when the depth of freezing was about two-thirds of the active layer and the horizontal component of the total heave force reached the greatest value, the displacement of the foundations was 80-90 percent.

The results of the studies show that the interaction of heaving ground with the foundations of structures is a dynamic process caused by laws of frost heave of the ground and the operating conditions of the foundations. The indexes of the interaction are generalized parameters--the heave force, the lateral pressure (squeezing), and the displacement of the solidly frozen ground surrounding the foundation from above. The numerical values of the generalized parameters of the interaction are probability values. They must be determined with a given probability according to the data of many years of measurements under natural conditions or by the results of laboratory experiments based on the use of the friction theory. When planning and designing structures on fills and steep slopes for single foundations, it is necessary to consider the horizontal components of the frost heave forces.

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PHYSICOCHEMICAL METHODS OF VARYING THE FORCES WITH WHICH EARTH FREEZES TO FOUNDATIONS

A. M. PCHELINTSEV, I. A. TYUTYUNOV, YE. A. LEVKOVICH, G. L. PRAZDNIKOVA, V. S. DESYAK, AND V. G. TENYANKO Scientific Research Institute of Foundations and Underground Structures

The necessity for varying the forces with which earth freezes to foundations arises for two reasons: in certain cases in order to decrease the tangential heave forces on foundations; in others, in order to increase the bearing capacity of the foundation in plastic permafrost or for stronger anchoring of it to the permafrost it is desirable to increase (the adfreeze forces).

Among the measures aimed at decreasing the forces of frost heave on foundations, the physicochemical procedure, consisting in lowering the forces with which the heaving ground freezes to the foundation or eliminating the heaving properties of the ground, is the most promising.

According to modern points of view, the frost heaving of foundations takes place as a result of the fact that the heaving ground, freezing to the surface of the foundation and gripping during heaving, lifts the foundation. Consequently, in order to keep the foundation from heaving, it is necessary to eliminate freezing of the ground to it and the grip of the frozen ground or elimination of the frost heave itself. The freezing of the heaving ground to the surface of the foundation takes place if the surface of the latter is wet with water absorbed by particles of the adjacent soil

$$\sigma_{sa} - \sigma_{sl} - \sigma_{la} \cdot \cos \theta > 0, \qquad (1)$$

where σ_{sa} is the surface energy of the wet surface at the interface with the air; σ_{sl} is the surface energy at the interface of the wet surface and the wetting liquid; σ_{la} is the surface energy of the liquid at the interface with the air; θ is the wetting angle.

If

$$\sigma_{sa} < \sigma_{sl} + \sigma_{la} \cdot \cos \theta, \qquad (2)$$

desorption of the water from the surface takes place, that is, the surface is not wetted by the water. It is very difficult in practice to eliminate the wetability of the surface of foundations altogether, but it is comparatively easy to reduce the degree of wetability.

As the studies performed at the Laboratory of Physical-Chemistry of Permafrost of the Scientific Research Institute of Foundations has demonstrated, by applying a thin hydrophobic film of high-molecular compounds to the foundation surface (phenol formaldehyde, melamine formaldehyde resins, epoxy compounds, siloxanes, etc.), it is possible to reduce the magnitude of the wetability and, consequently, significantly to decrease the freezing of heaving ground to the surface of foundations.

The epoxy compounds used for this purpose, lowering the forces with which the earth freezes to the foundation surface, have high adhesion to the concrete, frost resistance, and waterproofness. They form a strong, long-lasting film.

With respect to the research performed in the vicinity of Igarka and Vorkuta, the polymer film on the surface of the reinforced concrete piles as a result of cold hardening the epoxy compounds decreased the heave force from 2.5 to 8 times (Table 1).

The compressive forces of frozen ground can be reduced if a trapezoidal profile is given to the foundation. According to our experimental data, the minimum angle of taper of the faces is 1.5°. In this case the heave forces of the trapezoidal foundation are cut in half, and with an epoxy compound on the surface of the film, they are reduced by 8 times. However, the use of foundations of trapezoidal shape is possible only in areas where there are no winter thaws. The water that gets into the cracks between the foundation and the frozen ground creates large squeezing forces, and the heave forces increase significantly in this case.

According to the research of A. M. Pchelintsev,[3] a reduction in the adfreeze forces of the earth to the structural elements is possible also when using grease on which the new physicochemical method of preventing freezing of the earth to the reinforced concrete slabs and wooden sleepers on the rails of tower cranes at the construction site is based.

The surface of the reinforced concrete slabs in contact with the ground (ballast), the wooden sleeper, and the metal parts of the track under the rail are covered with a thin layer of grease that does not congeal at negative temperatures. In addition, the trapezoidal profile of the slabs and sleepers has special significance. The bevel of the faces and the ends is made at an angle of 2°-3° for the sleepers and 3°-5° for the slabs.

TABLE 1 Comparative Experimental Data for Determining the Heaving Forces on Reinforced Concrete Foundations in the Vicinities of Igarka and Vorkuta (According to the Data of A. M. Pchelintsev, O. S. Konnovoy, and A. V. Sadovskiy)

	Igarka				Vorkuta			
	Tangential Heave Stress, kg/cm^2				Tangential heave stress on the cylindrical foundation, kg/cm^2			
Form of Foundation	Without Film (Control) (*a*)	With Polymer Film of the Type (*b*)	Type of Film	Reduction Factor, $P = b/a$	Without Film (Control) (*a*)	With Polymer Film Type (*b*)	Type of Film	Reduction Factor, $P = b/a$
Prismoid	2.1	0.53	KPP	0.25	1.82	0.6	K-EPE	0.33
Cylindrical	--	--	--	--	1.82	0.7	K-EPA	0.38
Trapezoidal	--	0.25	KPP	0.12	--	--	--	--

NOTE: The reduction factor is the ratio of the heave force on the foundations covered by polymer film (*b*), to the heave force on the control (*a*).

They are laid with the narrow side on the ground, which eliminates the breaking of the frozen ground with separation of the sleepers and slabs from it, and it also decreases their jamming. On separation of the slabs and sleepers processed in this way from the frozen ground, the forces decrease by 10-15 times. Lubricants GOI-54p, TsIATIM-201 and TsIATIM-202 are recommended.

The freezing of the ground, ballast, and wet and loose materials to solid surfaces takes place as a result of the fact that at the contact between them in the freezing process ice is made and binds these solid surfaces together. The effect of the lubricant is that it eliminates the direct contact between the surfaces and the ice formed, preventing their adfreezing. With relative slipping of the solid surface and the ice or the frozen ground, the lubricant decreases the friction between them. In addition, it has an important property--it is hydrophobic.

The heaving of foundations can be eliminated by decreasing the magnitude of the heave of the surrounding ground. As is known, ground heave is connected with an increase in volume during freezing, which occurs both as a result of expansion of the water in the ground when it is freezing and as a result of the crystallization of the water that migrates to the freezing front. A generalization of the results of experimental studies of frost heave of the ground that we have performed in the Soviet Union and have been performed abroad[4,5] shows that the magnitude of frost heave basically is connected with the intensity of the moisture migration processes in the soil. Ground heave can be decreased or completely eliminated if we find an effective means of controlling the water migration process during freezing.

According to the data of I. A. Tyutyunov,[5] the motive force of water migration in the ground during freezing is the gradient of the free surface energy of the ground system. Consequently, after reducing the magnitude of the free surface energy of the ground particles, it is possible significantly to decrease or completely eliminate the migration of the water. This can be achieved by the following procedures: (1) making the ground hydrophobic; (2) replacing the multivalent cations in the exchange complex of the soil by univalent ones; (3) an irreversible decrease in magnitude of the specific surface of the soil particles by using active modifiers; and (4) joint treatment of the ground with univalent cation salts and modifiers.

Considering all of this, an entire series of chemical reagents was tested under laboratory conditions in two heaving soils--Vorkuta silty suglinok and kaolin clay. For this purpose, soil samples with different reagent concentrations were placed in a special thermostat, and they were frozen only from the top with a previously determined optimal temperature, -6°C. During the entire cycle of freezing the specimens to the lower part, an abundant supply of water was provided. After complete freezing, the specimens were thawed, and artificial seepage of "atmospheric" water took place for 2 days. The complete freezing and thawing cycle can provisionally be taken as a year. By multiple freezing and thawing, this makes it possible to establish the duration of effect of the reagents (Table 2).

The duration and the degree of reduction of heave for various reagents differ: The best effect is exhibited by reagents containing the fluorine ion and carbamide resins.

Water solutions of the reagents were used to process the silty suglinoks in experimental plots in the city of Vorkuta. The 3-yr observations indicate that the heave is completely absent in the plots.

The majority of reagents have been used to protect the reinforced concrete fence posts of the sanitation zone of the eighth hydraulic engineering complex in the city of Vorkuta from heaving where the heaving conditions are extremely favorable. Above there is a cover of silty

TABLE 2

Name of Reagents	Approximate Cost of Treatment of 1 m^3 of Soil, rubles	Mean Decrease in Heave, %	Duration of Effect of the Reagents, years
RL--1	3	100	5
RT--1	1	80	6-7
RT--2	0.8	80	7
RT--8	2.5	90	27[a]
RT--14	2.2	100	15[a]
RT--16	2.0	100	15[a]
RT--25	1.8	80	7
RL--8	2.8	100	15
RT--4	3.5	80	15
RL--12	6.3	100	15
RL--24	2.9	100	27[a]

[a]Tests are continuing.

TABLE 3 Results of Observations on the Heaving of Fence Posts around the Sanitation Zone of the Eight Hydraulic Engineering Complex (Vorkuta)

Reagents with Which the Soil was Treated	Magnitude of Heave of the Posts, cm, 1970-1971	Magnitude of Settlement of the Posts during Thawing of the Ground, cm	Magnitude of Heave of the Posts, cm, 1971-1972
RT--25	-2.5	-0.9	-0.4
RT--8a	-0.8	0.0	0.0
RT--8	-5.9	0.0	0.0
RT--4	+1.9	0.0	+2.4
RT--14	-1.4	0.0	0.0
RT--2	+0.5	-0.5	0.0
Without ground treatment (control post)			
	+36.9	0.0	+28.0
RT--25	-1.8	0.0	+0.4
RT--2	+2.8	-2.8	0.0
RT--14	-1.1	-0.5	+0.6
RT--25	1.2	-1.2	+1.5

NOTE: Plus indicates the amount the posts rose. Minus indicates the amount they settled.

suglinok, which is underlain at a depth of 2-3 m by rock. This composition favors the accumulation of groundwater, which in individual depressions emerges to the surface and forms swampy areas.

As 2 yr of observations show (Table 3), in spite of the complex natural conditions, the heaving of the fence posts is in practice absent at the same time as the posts around which the soil has not been treated are pushed up as much as 40 cm per year.

In many cases it is expedient to increase the admissible loads on the foundations without changing their structural design. This is entirely realizable if we increase the force of adfreezing of the permafrost to the foundations. This is based on corresponding theoretical research and laboratory experiments.

The formation of the contact zone between the soil and the foundation takes place in two stages. The first (prefreezing) is characterized by the wetting of the foundation surface with water adsorbed by the soil and the formation of adhesive bonds between the wet soil and the foundation. The strength of these bonds is determined by the force with which the soil adheres to the surface of the foundation. With a reduction in temperature, the force of adhesion of the bonds increases, and at negative temperature as ice forms these bonds are converted into adhesion-crystallization bonds. Their formation is the second stage of formation of the contact zone between the soil and the surface of the foundation. The strength of the bonds between the soil and the foundation sharply increases in this stage.

It has been established that, in the case of freezing at the soil-foundation contact, a layer of ice is formed and the strength of freezing of the soil to the foundation depends on its thickness. Insignificant generation of ice in the contact zone between the soil and the foundation ensures high freezing strength. The formation of a thick layer of ice in the contact zone reduces the adfreeze bond of the soil to the foundation.

The strength of the contact zone depends on the soil property, in particular the nature of the mineral particles in the skeleton and the fineness of the soil, their specific surface, the physicochemical properties and the state of the interfaces between the soil phases, the adsorption capacity of the soil, and the nature of the absorbed substances. In addition, the strength of the contact zone depends on the surface properties of the foundation--the nature of the physicochemical properties and the absorption capacity of the material, the surface of the foundation, and the porosity and the total surface of the microirregularities. The physicochemical composition of the soil at the foundation surface can be varied by treatment with chemical reagents. At a positive temperature, the chemical treatment of the soil implies variation of its fineness and adhesion properties, which at negative temperature is felt in the nature of redistribution of the water and ice generation. The chemicals that increase the fineness of the soil decrease the intensity of the migration process, preventing significant ice generation at the soil-foundation contact.

In order to study the effect of chemical treatment of the soil on the strength of adfreezing to the concrete, various chemicals were tested on Moscow suglinok.

In Table 4 we give the results of varying the forces of interaction of the soil with the concrete as a function of temperature and chemical treatment, from which it follows that the physico-

TABLE 4 Variation of the Forces of Interaction of the Soil with Concrete as a Function of Temperature and Chemical Treatment of the Soil

Reagent Injected into the Soil	Amount of Re-agent in % of the Weight of Dry Soil	Injection Procedure	Moisture Content of the Thawed Ground in % of Dry Weight	Adhesion, Kg/cm^2 Temperature 20°	Adhesion, Kg/cm^2 Temperature 0°	Moisture Content of the Frozen Ground in % of the Dry Weight	Adfreeze Forces--Soil to Foundation at -3°C, Kg/cm^2 (fast filling)
Iron oxide	1	powder	41.0	0.034	0.052	40.0	15.9
Cement-500	1	the same	41.4	0.021	0.044	40.0	20.4
GKZh-11	0.25	water solution	42.4	0.039	0.051	40.0	16.5
KO-85 lacquer	0.33	water emulsion	40.8	0.037	0.050	40.0	16.9
K-Suglinok	--	--	38.0	0.024	0.034	40.0	15.9
Suglinok	--	--	46.8	0.016	0.031	40.0	13.3
Control	--	--	42.2	0.025	0.037	40.0	13.6

chemical treatment of the soil increases the force of freezing to the concrete up to 50 percent. The best results are given by the injection of iron oxide, cement-500, the GKZh-11 waterproofing liquid, KO-85 organosilicon lacquer when these materials are injected into the soil in minimum amounts (about 1 percent). All of these materials increase the fineness of the soil, as a result of the fact that they themselves are fine-grained (cement, iron oxide) or they break down the natural aggregation of the soil (silica and organic materials). The Moscow suglinok, the absorbing complex of which is saturated with K-ions, is finer and has a greater adfreeze to concrete that the Moscow suglinok, the absorbing complex of which includes Fe-ions, which aggregate the soil particles.

It is also important to note that with a reduction in temperature from 20° to 0° the forces of adhesion of the soil to the concrete double.

The modification of the foundation surface by physicochemical means also strengthens the adfreeze bond. It can be modified by two means: treatment of the surface with hydrochloric or sulphuric acid, or by removing the upper layer of concrete by sandblasting. In Table 5 we have the primary results with respect to determining the adfreeze stresses of the Moscow suglinok (w = 40 percent) to the modified surface of the concrete, from which it follows that the adfreeze bond stress increases up to 65 percent.

TABLE 5

Type of Surface Treatment of the Concrete	Adfreeze Bond Stress, kg/cm^2
Sandblasting	26.0
Hydrochloric acid	20.1
Sulphuric acid	21.5
Control (without surface treatment)	16.1

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BASIC AREAS OF GEOCRYOLOGICAL RESEARCH IN MINING

V. P. BAKAKIN Production and Scientific-Research Institute of Engineering Research in Construction

Mining is one of the primary branches of heavy industry on the results of the activity on which natural and economic conditions have special and frequently decisive effect. Therefore, in the northern and northeastern parts of the USSR it is necessary to design special "northern" equipment and develop special technological processes. Accordingly, geocryological research has acquired great significance.

Geocryological studies of the mining engineering profile in the primary industrial centers of the North and Northeast have developed quite broadly in the last two decades, but they have encompassed individual problems of mining geocryology without special efforts at broad generalization and analysis of the results obtained.

For open-pit mines, the primary part of the research was concentrated on the methods of treating the frozen rock for excavation and the applicability of new mining techniques and equipment. When studying these methods, special attention was given to hydrothermal amelioration and methods of protecting the cover layer of rock from freezing. Broad studies were made of special methods of water amelioration in the northeastern part of the USSR, utilizing the natural percolation and artifically created heads. Hopeful results were obtained when studying film coatings that intensify natural thawing. The results in open-pit mines turned out to be successful where plastic coatings with weak electrical heating were used. Interesting results were obtained with natural snow removal and when studying the possibilities of artificial convection for retarding the process of freezing of dredge sections and other artificial bodies of water.

It is customary to consider vibration among the mechanical procedures for preparing frozen rock for excavation. By using vibro-hammers, an experimental output per shift of about 200 m^3 has been achieved with completely acceptable economic expenditures.

The results of using vibration during mechanical coring of frozen rock and the use of water, and steam, needles are equally interesting. A large role in drilling frozen coarse soil has been played by a combination vibration-impact-rotating tool, and for frozen melkozem [fine earth], the thermal method of drilling with a high-temperature flame and automated control of the process. It must be noted that in the case of mechanical exploratory and operational drilling of shallow wells in the North, there is a general trend toward converting to the use of compressed air to clean the holes and plastic casing to reinforce them.

In addition to the comparatively new developments, the northern projects are continuing to try to use intense loosening by means of heavy rippers when preparing frozen ground for removal.

In the area of creating new excavating equipment for the northern regions, the effort to increase its power and capacity is characteristic. In order to ensure the corresponding reliability of this equipment in operation, closed hydraulic control systems are now mounted on it, and antifreeze solutions and local heating are used to control freezing.

As before, preference is given to intrapit railless transportation and especially dump trucks for their maneuverability and increased capacity to go up steep snow-covered ramps. The broad application of the hydraulic method of removal, with the application of high-head monitors and elevators combined with excavators and other earth-working equipment used for preliminary loosening and removal of coarse fractions is new for open-pit mining in the northern and northeastern parts of the USSR. Surface removal of frozen rock with bulldozers with subsequent washing and hydraulic transportation is one version of the northern processes.

The new methods of underground mining have been developed appreciably less intensely in the North. In this area the primary efforts have been concentrated on the problems of surface planning and improving the methods of running out strips and cuts. The direct transfer of experience in exploration, planning, design, and preparation of the surface for large enterprises from the nonpermafrost parts of the country to the North has not been satisfactory in practice, especially with respect to ensuring stability of the structures of the complex above the mine and the preservation of engineering utilities. The scientific and procedural lack of engineering-geological exploration and lack of skill in forecasting variations in the geocryological conditions in the developed areas have had a significant effect.

It has turned out to be exceptionally difficult to select rational means of excavation in major mines in the thick Quaternary and other coarse deposits in the frozen state. Experience has demonstrated that the difficulties caused in this case by the intrapermafrost water, running, and unstable ground in the subpermafrost horizons is overcome most easily by using artificial freezing. The freezing technique has been well mastered. It corresponds to the local natural conditions and permits later use of the refrigerators to regulate the thawing aureoles in the mines. Special mine shaft "heads" have been developed for this purpose in the European North, by means of which the convective heat exchange is regulated within defined limits. Efforts there have been made more precisely to determine the nature of deformation of the timbering in the thawing zones, the flooding, and strongly developed undermining, by using bulb sensors. Positive results have been obtained with respect to improving the output capacity by running ventilation shafts and other vertical excavations with the leader borehole.

Significant difficulties frequently arise in the North when excavating beds by main inclined shafts. The frozen Quaternary deposits in the roof gradually thaw and run. The ground pressure developed in connection with this deforms the supports, including those of concrete. On the whole, the efficiency of the selected method of removal, the basis for the adopted exploitation of the mine fields in selection of the industrial sites, depends to a great extent in the North on the degree of investigation and carefulness of considering the geocryologic conditions during the planning and design stage.

With respect to resisting the ground pressure in the North, a general trend toward the maximum use of the natural bearing capacity of frozen ground in the roof, the discovery of the possibilities of using the rheologic properties of the frozen ground, and rational use of ice for strengthening the selected space is characteristic. These trends have found sufficient reflection in the development of the chamber systems of working, in broad application of excavation, by longwalls with light and basically organic support and the broad introduction of rock-bolting for preparatory and clean-up operations. The primary defining parameters of the enumerated removal systems and the methods used to control the roof were studied in detail in the mining projects of the Northeast.

Unfortunately, in this cycle of scientific research, the applicability under frozen conditions of modern systems with chamber blasting and large-tonage caving, which have great significance for the future development of underground mining capabilities in the North, was not discovered and was not analyzed. The possibility of using systems with ice remained to the same degree incompletely revealed, even though they most completely correspond to the local natural conditions.

In addition to the general problems of roof control, in the mines at many of the enterprises in the North there have been broad and detailed studies of the relation of the soil pressure to the variable temperature regime of the surrounding permafrost with the adopted methods of ventilating the mines. This research has completely confirmed the concept previously stated by the author regarding the inapplicability of the established rules for ventilating mines by hot air in regions with permafrost and dependence of maintaining the permafrost in the conditions of seasonal heat and moisture exchange. The experimental data obtained then permitted effective control of the fog formation in the subpermafrost horizons of the mines, the icing over of the fans and the mine workings, the freezing of the ore that has been removed, and many other situations caused by the convective heat and moisture transfer from the surface to the underground work specific to the North.

Interesting data were obtained on the seasonal periodicity for the best content in the air in the mine and with respect to its relation to the absolute moisture of the atmospheric air. As a result of this research, recommendations were made regarding the artificial humidification of the air in the underground mines to 8-10 g/m^3 during the winter. The recommendations are equally acceptable for northern conditions with respect to application of improved hydraulic removal to combat dust.

The defined results were achieved by introducing modern automation of individual production processes in the mines of the North and in the larger mines frequently with the use of different types of sensors including radioactive sensors.

When studying the characteristic features of the deposits themselves in the North and Northeast, interesting results were obtained on the part of more precise definition of the gas-bearing nature of the coal diamond, and part of the endogenic ore deposits. The developments that have recently been started with respect to utilizing the geothermal field anomalies in the exploration of oil and gas, ore, and other deposits in the North are promising. Interesting scientific results have been obtained from the first studies more precisely determining the effect of the local deep cooling on the processes of the formation and accumulation of oil and gas and on the formation of the zones of oxidation and weathering.

The studies of the physicomechanical properties of permafrost to improve the efficiency of mining in it have been insufficient. A consequence of this has been the known indeterminacy of the design parameters in the process of cutting permafrost, destroying it under impact and vibro-impact, the freezing factors, stability, drillability, and extractability, that is, all the parameters that mining practice especially needs.

For further improvement of mining techniques in the northern projects of the country, it is possible to recommend the following:

1. The continuation and intensification of studies of the general and deep geocryology of the most characteristic and largest deposits.
2. The need of studying the physicomechanical properties of permafrost important in mining engineering.
3. Careful study of the geocryologic processes on phenomena (including engineering-geologic) complicating the production of open-cast mines.
4. Payment of more attention to the basic technological processes of underground methods of mining in a thick bed of permafrost.
5. Improvement of the method of engineering research clarifying the conditions of the construction of mining projects in the North and forecasting changes in these conditions after the completion of construction.

EFFECT OF GEOCRYOLOGICAL CONDITIONS ON THE PRODUCTIVITY OF MINING OPERATIONS IN THE NORTHEASTERN USSR

P. D. CHABAN AND V. G. GOL'DTMAN VNII-I Institute

Almost all the mines and the workable placer deposits of gold, and also the mines and coal pits of the northeastern USSR, are located in mountainous areas under geocryological conditions characterized by the following data. The mean annual ground temperature under the seasonally thawing layer from -3° to -7° in the lower part of the slopes and to -9° in the upper part of the slopes and a permafrost thickness of 150-500 m are characteristic. The lower parts of the mountain slopes experience the effect of heat transfer from the groundwater of the seasonally thawing layer.

When mining the placer deposits under these specific natural conditions, the study of the seasonal thermal processes in the upper zone of rock to find procedures for optimal use of solar radiation and the heat of riverwater to prepare the rock for mining, primarily, for artificial thawing, has great significance.

The thawing of the ground on a large scale is necessary as a result of the fact that the bulldozers and scrapers used are designed for excavating thawed or broken frozen ground, and it is possible to rework only the thawed ground with draglines. In addition, thawing achieves the disintegration of the gold-bearing rock necessary in the concentration processes. The modern methods of thawing, based on using natural heat sources, do not always correspond to the technico-economic requirements. Accordingly, the problems of using other sources of thermal energy for thawing have been investigated.

The problem of studying, developing, and improving the method of thawing permafrost is a part of engineering geocryology. In classifying the methods of thawing, the form of heat transfer in the ground is taken as the primary attribute: conductive (the first group of procedures), convective (the second), and heat release on passage of an electric current through the ground. The first two groups are subdivided into subgroups depending on the heat source used: natural--solar radiation, atmospheric wet air, the water of the surface bodies of water or groundwater; or artificial--electric heating elements, flame or chemical heaters, hot water or steam, and so on.

When using natural heat sources, the method of layer-by-layer thawing used in the spring-summer season has become most widespread. It provides for the most frequent possible (in the ideal case, continuous) cutting of the thawing layer of ground and removal of the soil mass formed over the ice-rich permafrost. In this case there is periodic denudation of the frozen surface, on which the solar radiation is transformed into heat, the latent heat of the steam is released, and the heat of the atmospheric air is advectively assimilated. The accumulation of the layer of thawed soil through which the heat is transferred by conduction from the surface retards thawing. In an area where there is a continental climate, gravels with an ice content of 200 kg/m^3 is thawed with a daily cut on the average of 10 cm/day and suglinoks with an ice content of 300-400 kg/m^3, only 6 cm/day. Under arctic climatic conditions under the influence of the ocean, it is possible to thaw 8-4 cm/day correspondingly. The thawing effect is achieved at the end of May and the beginning of June during negative mean daily air temperature.

Under the conditions of the Chukotka seacoast, I. M. Papernov[1] established the relation for the heat flux to the periodically denuded permafrost as a function of the total solar radiation Q, the albedo A (about 0.2), the energy consumption of evaporation or condensation LE, and the air temperature t_B at a height of 2 m with a convective exchange factor a_k = 22.5 kcal/m-h. According to the mean daily data for the June-August period, the following expression is valid:

$$q = Q(0.84 - K)(1 - A) + 355t_B - 540 \quad \text{kcal/m}^2\text{-day},$$

in which

$$K = \frac{LE}{(1 - A)}.$$

Here t_B = -10°C to + 5°C, K = 0.04; when t_B = 6°-8°C, K = 0.08; when t_B = 9°-13°C, K = 0.13; and when t_B = 14°-20°C, K = 0.16. The depth of thawing is found to be $h = q/80\ G$, where G is the ice content, kg/m^3.

These calculations are performed when planning and designing open layer-by-layer stripping operations and also when controlling the process relative to variable temperature conditions.

The forcing of seasonal thawing using transparent film coverings permits excavation of the thawed layer to be started 15-20 days earlier than usual. In April and May the mean daily surface temperature of the ground under the covering is 10°-12° higher than outside it. In subsequent months in the summer, the effectiveness of the cover decreases. The covering made of polyamide or polyethylene film excludes the losses from evaporation from the energy balance of the rock surface, keeping the resultant heat influx from solar radiation to 70-80 percent. During the daylight hours, the temperature difference between the outside air and the rock surface under the vaulted film covering reaches 40°-50°. This effect was used by R. N. Novoseletskiy and A. I. Priymak[2] in "the rain-under-film" method proposed by them. Here, cold water heated in the upper layer of the ground and percolates to depth with a temperature of 3°-5°C, forming horizontal percolation to a drainage well from which the water returns. On the path to the well, the percolation thaws the permafrost over which it flows. This method is successfully used during the period from April to the end of May for drained polygons.

The known method of wellpoint thawing of permafrost has found broad application as a most effective and universal procedure. The depth of thawing varies from 4 m to 30 m. With this procedure, the ascending flow transfers heat to the cylindrical interface between the thawed and frozen zones and simultaneously has a hydromechanical effect on the thawed soil and increasing its permeability.

The results of the research of Gol'dtman *et al.*[3] permitted the development of a method of calculating the duration of effect of the wellpoint as a function of various factors: the depth to which the wellpoint is submerged and the spacing of the wellpoints, the water flow rate and temperature, the ice content and the thermal conductivity of the ground, and the permeability of the soil. For various combinations of conditions and a number of values of the water temperature from 4°C to 15°C, the following formula has been proposed.

$$T = \frac{1}{t}\, 0.895\, b^{2.234} \left(\frac{G}{W}\right)^{0.707} \left(0.11\, z + 1\right)\, l^{0.02(t - 10)},$$

where T is the duration of the effect of the wellpoint, days; t is the temperature of the pumped water, °C; b is the finite thawing radius, m; G is the ice content of the permafrost, kg/m^3; W is the mean flow rate of the water, m^3/h; z is the depth of thawing, m [l is undefined].

For further more precise definition of the method of calculation, G. Z. Perl'shteyn and V. Ye. Kapranov investigated the methods of heat transfer by percolation in sand and gravel.

The results of the engineering-geocryological studies in this area have served as a basis for the development of requirements on technical equipment of operations with respect to wellpoint thawing.

In recent years studies have been made of the processes of heat transfer by horizontal drainage of the permafrost. V. V. Znamenskiy[4] developed the method of determining the heat transfer by these flows. A study was made of the cases of percolation in a two-layer medium with different permeabilities considering the heat flux from the ground surface through the layer of soil not saturated above the free level. G. Z. Perl'shteyn investigated the problem of heat transfer with a descending flow of water from a sprinkler.

The search for methods of protecting thawed ground from winter freezing has led to the conclusion of the technicoeconomic expediency of the perennial use of special polystyrene-foam shields. Their strength, hydrophobic nature, and thermal insulation properties permit of the flooding and removal of the insulating cover independently of the weather conditions and the state of the ground surface.

The absence of surface runoff during 6 or 7 months of the year complicates the flooding of the sections in the winter, widely used in more southerly regions to protect them from freezing.

In connection with the mining of deep placers by open and underground methods, problems arise regarding the shape of the suprapermafrost and through-taliks, expedient methods of draining these taliks in the roof, and the most complex problem--insuring stability of the roof when the underground operations are performed in weak permafrost with a temperature of 0°C to -2°C where the expenditures on timbering are many times greater than in permafrost with a temperature of -4°C and lower.

From the experience in using groundwater for water supply, it is known that the preliminary drainage of the through-, and nonthrough-, taliks to a depth of 40-50 m presents no great technical difficulties. Rules for the prevention of the unexpected discovery of water-bearing taliks when running underground preparatory tunnels have also been developed.

P. N. Kalmykov performed some research on the methods of mining a flooded placer in the Sibik-Tyellakh River basin, where an initial inflow of water of 15,000 to 20,000 m^3/day is possible. These studies revealed the technical and economic preferability of using the open-pit mining procedure with preliminary drainage of the deposits using vertical drainage wells.

The effect of the geocryologic conditions on the technological process, and through it on engineering and techniques used for underground mining operations, is especially large. Thus, for the mining and beneficiation of permafrost placers, special types of drilling carriages, punches, scrapers, washing units, and so on, have been built. A machine has been built for direct mechanical crushing of a bed of very coarse deposits containing gold.

The presence of permafrost can be both a positive factor facilitating the conduct of mining operations and a negative factor complicating them. In the case of moderate natural temperature of the permafrost (-3°C to -7°C), it is characterized by quite high mechanical strength, which increases the bearing capacity of the pillars, the stability of even significant cavities in the roots, and, consequently, the possibility arising from this for conducting mining operations with less dense timbering and a reduction in the probability of injury from caving of individual pieces of rock. Under these conditions there is no inflow of water in the mines, the oxidation processes take place with low intensity, the rock is practically impervious to gas, and coal beds are weakly gas-bearing.

On the other hand, in frozen sedimentary rocks and fine soils, the bearing capacity and stability sharply fall during seasonal or progressive thawing, which unavoidably leads to deformations of the timbering and even to collapse of the workings; the icing of the mines is possible in cases of penetration of surface or subpermafrost water and freezing of the useful minerals.

The projects are experiencing difficulties with the performance of mining operations in the transition and subpermafrost horizons where the rock is unstable, fissured, flooded, and more gas-bearing and also when solving the problems of grounding underground electrical equipment as a result of the extremely low electrical conductivity of the permafrost.

The first complications are so significant that in many cases, especially when mining coal deposits and excavating resources in the transition and subpermafrost horizons, laborious and expensive work is done to find and prepare preferable new mine fields in the permafrost series. For this reason, there is up to the present time practically no experience in mining placers in the transition and subpermafrost horizons, or in the placers made up of taliks.

In addition to the factors enumerated, the negative temperatures of the rock massif and moving mine air in certain combinations create unfavorable hygienic conditions for labor, complicate dust control, and promote the development of professional diseases among miners.

The all-around study of the geological conditions for developing efficient technological procedures, rules, and norms for conducting underground mining operations and increasing the general efficiency and safety of the working of the beds in the Northeast, done in the last decade by the Leningrad Mining Institute, the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences, and the All-Union Scientific Research Institute of Gold and Rare Metals (VNII-I), has given a number of valuable scientific and practical results. In particular, a study was made of the physicomechanical properties of the bedrock and permafrozen soils and also the thermal conditions of the mines, coal mines, and placer workings in the permafrost zone.

The mechanical properties of the frozen metamorphic and volcanic bedrock, as the research results discussed in the monograph by Yu. D. Dyad'kin[5] indicate, do not, in the presence of low moisture, depend on temperature. On the contrary, the properties of the sedimentary bedrock fluctuate significantly as a function of the temperature and ice content. The established relations are used for practical purposes to evaluate and calculate rock pressures.

The study of the physicomechanical properties of very coarse-grained permafrost, which is widespread in the Northeast, was made by V. N. Taybashev. The characteristics of the permafrost obtained as a result of this research are used in designing the pillars for placer mines.

Depending on the composition and ice content, the very coarse deposits permit stable spans of the excavation chambers of 17-20 m and more. The open mining operations with an area of the roof to 1,500 m^2 to 2,000 m^2 maintain stability with a relatively low density of timbering if the rock temperature is below -4°C.

The thermal regime of the mines and shafts is shaped as a result of the complex thermal and heat-exchange processes in ventilated mines run in permafrost. The outside air entering the underground mines changes the parameters of its physical state (temperature, moisture content, heat content) as a result of heat and mass exchange with the surrounding rock and also as a result of absolute heat generation.

As a result of heat exchange with the mine air, the natural thermal state of the rock changes. The heat exchange takes place with variable intensity and with variable direction of the thermal flux: in the winter from the permafrost and the removed mass from the mine to the air; in the summer, vice versa.

Depending on the primary parameters, the mine air gives up moisture or is saturated by it; in mines with positive air temperature, the processes of evaporation or condensation take place and, with a negative temperature, the processes of sublimation or crystallization. The amount of moisture (ice) in the permafrost layers located close to the surface of the mines and in the removed rock mass also varies as a result of the heat- and moisture-exchange processes.

It is possible to list the basic features of uncontrolled thermal and moisture regime of mines and shafts:

1. Comparatively low values and sharp seasonal fluctuations of the temperature of the air in the mine.
2. High relative humidity of the air, especially in remote mines with transport.
3. The presence of seasonal temperature fluctuations in the rock surrounding the mine, which is damp on going from the rock surface into the depths of the massif and along the length of the ventilation path.
4. Participation of the latent heat of aggregate-moisture in the processes of heat exchange.
5. A negative annual heat balance of the rock massif and the formation of zones of cooled rock about the mines.
6. Negative annual moisture balance in the mines.
7. Predominant significance of the annual heat releases of the rock massif in the heat balance of the mine or shaft.

The thermal conductivity of metamorphic and volcanic permafrost in which the ore deposits are most frequently localized is higher than in the sediments; therefore, the rates of the temperature variation of the air in the mine and the damping of the amplitude of the annual fluctuations in it are comparatively high. In connection with the lower ice content (3-5 percent) of the rock in the ore deposits, their relative heating is possible with smaller heat consumption. The absence of timbering, or low density of timbering, characteristic of the mines enhance the heat-exchange processes.

The thermal regime directly or indirectly affects all the elements of mining operations, since the variation of the microclimate in the mines and the thermal condition of the rock massif is connected with hygienic factors, dust and ventilation regimes, and the conditions of the mine stability and the operation and maintenance of transportation. In order to increase the efficiency and safety of the mining operations, it is recognized as necessary to regulate the thermal regime of the mines and shafts, establishing effective limits and providing a basis for the control systems, considering the specific peculiarities of working the deposits in permafrost.

From these points of view, the inapplicability of heating the air entering the mine to +2°C, as demanded by safety rules, has become obvious. A careful analysis and investigation of the maintenance and operation requirements, as applied to the conditions of working the deposits at each enterprise, have permitted establishment of effective limits of heating the air in the mines and the coal mines within the limits of moderate negative temperatures (-5°C, -8°C), which is also realized in practice.

It has been proved that, in contrast to the mining projects that work the deposits in nonfrozen ground, heating of the air in the mines and shafts in the permafrost zone only pursues hygienic purposes. The heating of the air has permitted a reduction in the level of contraction of colds by the miners.

In one of the mines working a gold deposit in sedimentary permafrost (argillaceous and tuffaceous shales), efficient regulation of the thermal regime has required not only heating of the air to -5°C during the cold part of the year, but also cooling it to 6°C and 8°C during the warm part of the year. For miners working deposits in cold volcanic and stable metamorphic rock, cooling of the air is not required.

The studies have established the phenomenon of seasonal fluctuations of the dust content in the mine air. The primary principle of the reduced dust content of the air during the warm part of the year consists in predominant development in the mines of condensation (crystallization) processes, and the sharp increase in dust content during the cold part of the year is connected with evaporative (sublimation) processes. The evaporative processes are especially intense when the air is heated in dry heat exchangers.

In order to prevent the mines from drying out and having additional dust formation and to prevent physical weathering of the icy rock, the conclusion has been drawn that it is necessary to take directional action on the intensity and course of the mass-exchange processes in the mines by artificially humidifying the heated air. The humidification is realized to a level of relative humidity of the heated air no more than 85-90 percent in order to prevent crystallization of the water vapor in the ventilation channel and air supply shaft, the icing over of which is inadmissible. On the basis of this, an artificial system has been created and introduced in two of the mines and one coal mine of the Northeast for mine air with automated regulation and control of the process. The air is humidified by the controlled release of steam within the heating unit. As a result of the humidification, the dust-content of the air in the incoming ventilation jet is reduced from 2-10 mg/m^3 to the sanitary norm.

The regulation of the thermal regime of placer mines is also extremely necessary and expedient, but within defined limits established as a function of the purposes of the regulation and type of mine. Here, the regulation must not be identified with heating the mine air to a positive temperature, and, therefore, it must not enter into contradiction with the requirements of the safety rules.

The geocryological, mining engineering, and economic conditions of underground mining of permafrost placers require that all-around solutions be found to the ventilation and regulation of the thermal and dust regimes of the placer mines. One such solution is combined systems for ventilating the mines based on repeated use of the mine air.[6] During the recirculation process, the air is freshened and cleaned in air cleaners and frameless fabric filters. As a result of heat exchange with the rock, the air temperature in the recirculation becomes close to its natural temperature.

The geocryological conditions of working mines in permafrost have given rise to the necessity for the creation and application of specific methods and means of dust control. In the mines of the Northeast, dry dust removal and, to a lesser degree, hydraulic dust removal using sprayed chloride solutions have become widespread.

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CONSTRUCTION OF DAMS FROM LOCAL MATERIALS IN THE SOVIET NORTH

YU. E. GLUSKIN AND V. YE. ZISKOVICH   Lengidroproyekt Institute

The industrial mastery of the richest mineral reserves in the Far North has required advanced development of power. At the present time the construction of two large hydroelectric power stations is being completed (Vilyuy and Khantaya) with high-head dams primarily made from local materials.

The construction of the Vilyuy Hydroelectric Power Plant Dam, 74 m high and 600 m long at the crest, began in 1964 and was completed in 1967. The dam is on rock and was built of ungraded rock. The waterstop was of thawed detrital suglinoks.

The construction of the dam for the Khantaya Hydroelectric Power Plant, 65 m high and 420 m long at the crest, began in 1965 and is nearing completion. It is being erected on a rock base and is made of rock debris with the central core made of gravelly soil with supes.

The Lengidroproyekt Institute has designed a debris dam for the Kilyma Hydroelectric Power Plant, 124 m high and 790 m long at the crest, with a core made of local suglinoks. The dam rests on permafrost.

Along with the construction of the high-head dams, the construction of low- and medium-head dams of various types from local materials is continuing.

Some considerations are presented below regarding the peculiarities of the composition of hydraulic engineering works and the structural designs of low- and medium-head dams using frozen soils and high-head dams on frozen rock beginning with experiences in designing, erecting, operating, and maintaining dams in the Soviet North.

The natural-climatic peculiarities of the Far North predetermine to a significant degree the composition of the complex of structures, but primarily the choice of type and structural design of the dam.

The permafrost regions are characterized by the following natural-climatic conditions:

1. The mean annual air temperature in these regions is negative. The annual amplitude of the temperature fluctuations is 80°-100°C; with a minimal winter, it is -50°C to -65°C; and with a maximal summer, 30°C to 38°C.
2. As a rule, the rivers flow from south to north. Accordingly, opening them up begins with the headwaters in the southern parts of the country, and the spring floods make up 60-70 percent of the annual runoff.

In the winter and summer months the runoff of the rivers decreases sharply and they often freeze to the bottom.

3. The ice cover on the rivers is 1.5-2.0 m. The spring break-ups are accompanied by powerful ice jams, and the dimensions of individual ice floes reach a significant magnitude, which creates severe conditions for the passage of the constricted flow rates and to a significant degree affects the composition and type of the hydraulic structures, especially the dam and the spillways.
4. The poorly-developed network of highways and railroads and difficulties connected with delivering construction materials from the internal industrial regions of the Soviet Union predetermine the erection of dams in the Far North from local materials, most frequently from rock, sand, suglinok, and supes. In addition, the structural design of the body of the dam and its water tightness depend on the availability of certain local construction materials.
5. The upper beds of permafrost (as a rule, coarse Quaternary deposits) can be nonsubsiding, subsiding, or very subsiding.

The nonsubsiding soil, which gives insignificant settlement on thawing, can be used as the foundation of the dam either in frozen or thawed form.

The last two categories of soil often have inclusions in the form of thick ice lenses or a polygonal network of multiple vein ice having a depth of penetration to 25 m and more. The ice content by volume of such ground reaches 60 to 70 percent. Upon thawing, such ground goes into a fluid state and can be used as the foundation of a dam only if it is kept frozen both during construction and during operation and maintenance.

Earth dams of low- or medium-head in the Far North, as practice has demonstrated, can be erected on any ground, including that which subsides [on being thawed]. In such a case, it is necessary to construct a frozen core contiguous with the permafrost and create a single frozen mass of the dam and its foundation.

For storage dams, maintenance of the permafrost is possible only in the central section and in the lower part of the dam. Under the thawed upper part, the sharply-subsiding ground containing lenses or multiple vein ice can be replaced by quarried thawed ground or subjected to preconstruction thawing, as was done on the Kazachka River and the Melkiy Creek dams.

In the former case, the icy gravelly deposits thawed naturally during the two summer seasons after foundation excavation.

When preparing the base for the dam on the Melkiy Creek, the filled-in lake filled with ice with interlayers of peat and the sides of the valley made up of frozen ground with inclusions of multiple vein ice were subjected to thawing.

As practice has demonstrated, preconstruction thawing is complicated, and in technical respects it is a poorly effective measure requiring a long time for realization.

As a result, in the technical design for the frozen earth dam in the headwaters of the Taty River in Yakut, ASSR, designed by the Leningrad Division of the Gidroproyekt Institute in 1972, the decision was made to replace the soil in the foundation having multiple vein ice within the central part of the dam, under the upper part, and in the left bank.

The creation of the seepage stability of low- and medium-head dams in the Far North is usually achieved by using some procedure to freeze the central section and the lower part of the dam and also their foundations.

Several methods of freezing the water cutoff of the dam are known:

1. Layer-by-layer freezing of the body of the dam naturally during the construction process.
2. Freezing of the core of the dam on completion of its construction but before completing the shell, using artificial cooling. In this case, the lower part of the dam freezes, as a rule, during operation and maintenance.
3. Cooling the lower slopes from the surface in the winter, during operation and maintenance, using wooden sheds with ventilation wells and ventilation elements designed by Krylov to accelerate the natural freezing of the ground. This procedure was used when erecting the dams on the Poselkovyy and in the plan for the dam on the Yevdokimikh River to supply water to the port of Nordvik in 1945.

The second of the enumerated procedures has become most widespread. It is highly effective and reliable for dams with a head up to 20-25 m.

For earth dams more than 25 m high, it is possible to use a two-stage freezing-unit: for the base of the dam, from the Krylov unit, and for the body of the dam, from its crest.

Artificial cooling and freezing of the ground in the body and base of the dam are used during the first years of operation and maintenance. Then forced cooling is carried out in the winter once every 2 or 3 years, and, after a stationary thermal regime has been created in the dam, the freezing system can be completely shut down.

In some areas of the Far East on the Pacific Coast, where frequent winter warming and increased humidity (to 90 percent) of the air are observed on the permafrost, it is clearly expedient to erect frozen earth dams with natural freezing of the body of the dam during construction and maintenance. This solution was adopted when the Lengidroproyekt Institute did the technical design for the earth dam on the Kazachka River at Anadyr'.

The frequent warmings, severe blizzards, and snowdrifts on the shores of the Okhotsk and Bering seas exclude the possibility of using air-cooling units to freeze the dams because of the danger of the formation of ice plugs in the columns and icing of the columns themselves.

At the present time the Lengidroproyekt Institute, jointly with a number of scientific research organizations, is doing broad research, planning, and design development and is making field observations to provide a basis for the possibility of applying new effective types of cooling units to create frozen curtains in dams.

The most vulnerable things in hydroengineering with earth dams erected in the Far North remain the spillways and floodgates. The structural designs used for the floodgates in modern planning, design, and construction practice are a constant source of deformations and destruction of the dams. Therefore, in the northern regions it has become more and more common to resort to the construction of plain low-head dams using intakes of siphons and pumps.

The discharge of the floodwater is realized, as a rule, through shore spillways, going around the body of the dam, or, if there is a rocky massif adjacent on the shore, then in the form of a tunnel spillway built independently of the dam. This solution was adopted by Lengidroproyekt when preparing the technical plan for the hydroengineering complex in the headwaters of the Tatty River.

However, the optimal solutions have still not been found and are being worked on, not only when putting the spillways in subsidable icy ground, but also for rock bases.

An example of unsuccessful construction of the shore spillway can be considered to be the wooden spillway on the Kazachka River, which was destroyed as a result of thawing of the underground ice.

The structural design of high-head rockfill dams for hydroelectric power complexes in the Far North is distinguished from the low- and medium-head earth dams described above. Above all, special requirements are imposed on the foundation of the structures. The sites of a hydroengineering complex with a high-head dam is selected with the condition that the strong rock be at a relatively shallow depth from the daylight surface.

In the absence of subsidable soils in the foundations of the structures, many problems drop out that are connected with maintaining negative temperatures in the body and base of the dam.

The variants of the dams with artificially frozen body or core investigated when planning the Vilyuy and Kolyma Hydroelectric Power Plants demonstrated that such solutions offer no advantages by comparison with the thawed watertight structures.

The erection of high-head dams in the Far North has specific peculiarities connected with preparing the base of the dam, performance of the operations, passage of the construction flows, and operation and maintenance conditions.

The preparation of the base under the dam on large rivers in the permafrost regions is a complex problem. Within the limits of the low-water level of the river, the ground is in the thawed state, but on the floodplain, the permafrost is of significant thickness. Usually preference is given to the thawed variant, since there is no experience in making frozen cutoffs in the base of a high-head dam. The water cut-off in the base is made by grouting of the rock.

At the Vilyuy hydroelectric power complex, where the permeability of the rock in the base does not exceed 10 m/day, the grout curtain was made with the reservoir filled in several stages as thawing of the massif took place. This solution turned out to be technically possible, but relatively expensive.

At the Kolyma hydroelectric power complex, where the permeability of the rock exceeds 100 m/day, the possibility of grouting operations with a raised water level is doubtful. Therefore, in the plan the decision was made to do the grouting before filling the reservoir after preliminary hydraulic thawing of the rocky massif. The experimental work at the construction site proved the feasibility of hydraulic thawing, but the experiments were not complete, and it is still premature to draw final conclusions regarding the effectiveness of the given method of strengthening the base.

Until the beginning of the 1960's, there was neither foreign nor Soviet construction experience in erecting high-head dams from local materials under severe climatic conditions. According to the convictions of that time, the air temperature that was permissible to construct the shut-off curtains could not be lower than -5°C to -10°C, and, for the rock-fill, no lower than -30°C. These restrictions led to "seasonalness" of the operations and an increase in the time and cost of construction.

A large volume of winter placement of cohesive soil and the rockfill took place for the first time in world practice in the construction of the Vilyuy Hydroelectric Power Plant. This became possible as a result of the fact that a clear-cut technological scheme had been developed for the preparation, storage, transportation, and laying of the cohesive soil.

The experience in erecting the Vilyuy dam was successfully used also at the Khantaya dam with a core of gravelly suglinok soil.

For dams under the conditions of the Far North, original solutions were used when erecting the other parts of the dam. Thus, when erecting the Vilyuy and Khantaya dams, the shells of rockfill were placed year-round by a pioneering procedure in lifts from 2 m to 10 m high.

The passage of the construction flows is closely connected with the hydrologic regime of the northern rivers, the majority of which are characterized by exceptional nonuniformity of the annual runoff.

During the winter, the flow rates of the rivers drop sharply, since after freezing of the seasonally thawed layer, the groundwater supply to the river is only from the talik within the low-water bed.

This fact has been used when building the Vilyuy hydroelectric power complex: The spring and summer flow rates were passed through a special ditch cut in the rock massif on the left bank, and the winter runoff was twice accumulated completely in the head race--on drying the trench for setting up the grouting equipment for the base of the dam and when the construction trench was covered.

When building the Khantaya Hydroelectric Power Plant, a tunnel was built that was designed to pass only the winter flows. The spring and summer flows were passed over the incompleted dam.

In order to protect the fill from erosion at the foot of the low bank, a crib wall was built with dentated concrete and with a reinforced concrete slab over the fill.

This type of scheme for passing the construction flows is economical, but it has one significant disadvantage: the flooding of the excavation and the interruption of operations for the two or three summer months, which under the conditions of the Far North, is inexpedient.

The climatic conditions of the Far North have an effect on the dam and other structures of the hydraulic engineering complex, even during operation and maintenance.

The lower prism of the rockfill dams is cooled sharply. If special measures are not taken, then intense convection of the outside air through the pores of the fill occurs. As a result of condensation of moisture from the air and the penetration of precipitation, the formation of ice in the pores of the rockfill of the lower prism is possible. It is still unknown whether this has any effect on the stability of the dam. This question needs additional research.

It must be noted that, in the dams that have been built and are in operation on the Vilyuy and Khantaya, there are no direct observations confirming the formation of ice in the pores of the fill. The progress of the settlement of the Vilyuy dam also provides no basis for confirming that the pores in it are filled with ice.

As for intense convection of air through the body of the lower prism, this is completely confirmed by observations of the thermal regime of the dam at the Vilyuy Hydroelectric Power Plant for 1966-1970. In 1970-1971, along the lower slope of the dam a facing of fine gravel was laid. As a result of this measure, the intensity of the air convection dropped sharply, which temperature observations confirmed.

In order to prevent convection of the outside air, the design of the Kolyma Hydroelectric Power Plant provided for the construction of the facing along the lower slope in advance.

The design, construction, and operating experience on the dams made of local materials in the Far North is still highly limited.

However, it is now possible to make some generalizations, draw some conclusions, and make some practical recommendations with respect to the future development of dam construction in the northern regions:

1. Earth dams 5 m to 20 m high freeze under the conditions of the Far North during their operation and maintenance independently of their structural design, the method of construction, or the maintenance conditions.

2. The lower part of high-head rockfill dams is also subjected to the irreversible process of

natural freezing. This phenomenon requires the study, statement of broad natural observations, and theoretical developments.

3. The most vulnerable place in earth dams remains the contact between the dam and the floodgate or spillway. It is necessary to continue the studies of the optimal structural solutions for the floodgates and spillways.

4. The existing freezing systems for creating the frozen curtains in earth dams are complicated and expensive in operation. Accordingly, there is an urgent requirement for studying the possibility of introducing new, structurally simple, economical and effective freezing-units into hydraulic engineering construction practice.

5. The existing methods of self-compaction (i.e., excluding the traditional methods of compaction) of rockfill for high-head dams in the Far North by placing the rock in 2 m to 10 m lifts give comparatively large settlements. Accordingly, it is necessary to continue to look for optimal methods of compacting the rockfill.

6. The introduction of a method of placing the cohesive soil in the water-stops of the dams at low negative temperatures has significantly reduced the time for erection of the structures, and it has increased the effectiveness of constructing the dams from local materials under the conditions of the Far North.

7. The search for, and natural investigations of, economical cutoff structures has important significance in achieving the watertightness of permafrost bases for dams acquiring significant seepage on thawing of the rock.

Hydraulic engineering construction planned in the near future will require further study of the experience in erecting and operating hydraulic engineering structures, analysis of this experience, and the statement of broad geocryologic studies and theoretical developments with respect to solving an entire series of problems in the field of dam construction in the Far North.

EXPERIENCE IN CONSTRUCTING DAMS ON PERMAFROST IN YAKUTIA

G. F. BIYANOV Vilyuygesstroy

Of all the problems of applied geocryology and the construction of engineering structures on permafrost, the least clarified problems are those of erecting hydraulic engineering structures.

If under ordinary conditions, when filling the reservoir seepage occurs and is permitted in the body and base of the dam and when taking the corresponding measures it does not threaten the integrity of the structure, then, when construcing the dam on permafrost, it is necessary to take into account the fact that the occurrence (initially even an insignificant amount) of seepage leads to thawing of the base and then a progressive increase in seepage. The thawing of the base causes nonuniform settlement as a result of melting of the ice and self-consolidation of the ground. This can lead to loss of bearing capacity of the foundation, loss of water tightness, and stability of the structure.

The problems connected with the organization and performance of operations related to erecting structures in the North and Northeast of the USSR are no less complicated. This complexity is determined by the scattered nature of the construction sites over a broad area, the absence of transport communications, and construction materials enterprises. The high cost of bringing in construction materials dictates the broad application of local materials. The construction work is done under severe climatic conditions, where the amplitude of the daily temperature fluctuations of the outside air reaches 40°C, and the annual fluctuations, 100° with a mean annual temperature of -8° to -12° and less than 60 frost-free days per year. The outside air temperature stays below -40°, -50°, and even -60° steadily for prolonged periods.

Below let us discuss some of the interesting examples of the erection of dams. They were all constructed of local materials, but they differ significantly with respect to structural design and temperature regime.

CONSTRUCTION OF A DAM OF THE THAWED TYPE ON A FROZEN ROCK FOUNDATION

The erection of a large hydroelectric power plant of about 650,000 kW on the Vilyuy River has been completed. The first stage of this dam has been in operation since 1967.

This hydroelectric power plant was the first to be erected under the special, extremely severe, natural-climatic climatic conditions in the practice of hydroelectric power engineering construction.

The construction area is located in the zone of continuous permafrost to 300 m thick. A seasonal thawing of the ground takes place to a depth of 0.5-1.8 m. A cold winter with low temperatures lasting a prolonged period of time when the temperature reaches -65° and lower creates severe conditions for construction work. The conditions of organizing the construction are no less complicated. But this is connected with the remote location of the hydroelectric power plant in an entirely uninhabited and undeveloped area, in economic respects, and in the complete absence of communications. As a result of the rapids located downstream from the construction area, the Vilyuy River is not suitable for transport purposes.

The complex of the hydroelectric power engineering structures includes the following: a rockfill dam, a spillway with one segmental gate 40 m wide and 14 m high at the sill of the outlet, an intake structure, the hydroelectric power-plant building with penstocks, and an enclosed ZRU-220 kV distribution structure, etc. The hydroelectric power-plant building in the first stage of the semiunderground type was executed in a rock excavation on the right bank. The machine room of the second stage was open, on the left bank.

The rockfill dam with a total volume of 5 million m^3 and 75 m high creates a reservoir with perennial regulation on the Vilyuy River 36 km^3 in volume, 222,000 ha in area, about 450 km in length, and with a maximum depth of 70 m.

The construction of the rockfill dams in our country was done until recently within highly limited volumes; the greater part of the dams with low or medium head were built in the central or southern parts of the country, for which comparatively favorable natural-climatic conditions are characteristic. The most significant of these dams is the Orto-Tokoyskaya, which is 59 m high; the Irklinskaya, which is 43 m high; and the Shirokovskaya, 40 m high. Soviet experience in erecting rockfill dams with a head of more than 50 m in the regions having a mean annual negative temperature is completely absent. Beyond the border, in particular, in Sweden and Canada, the erection of such dams is limited to the rigid requirements of the technical specifications. In particular, the performance of operations with respect to placing the rockfill into the body of the dam is permitted when the outside air temperature is no lower than -30°. Still more rigid requirements are imposed on the construction of the impervious elements of the dams, which it is recommended to erect only during the warm part of the year or, as an exception, for outside air temperatures no lower than -5° to -10°.

The impervious element of the Vilyuy dam is a thin core made of rubbly suglinoks along a two-layer inverted filter made of crushed gravel in sizes of 0-4 mm and 0-150 mm (Figure 1).

The base of the dam contains diabases covered on the slopes of the shores with a layer of coarse deposits 1.5-2.0 m thick, which are the product of weathering of the diabases and comprising a gravelly suglinok with boulder inclusions. There is a through-talik under the river channel.

The contact of the suglinok core to the rock base was made in the form of a weakly reinforced concrete slab and a grouting gallery. Before filling the reservoir from the gallery [*sic*], the base of the dam was grouted within the talik under the channel. The remaining sections of the base were grouted after filling the reservoir as thawing of the rock occurred.

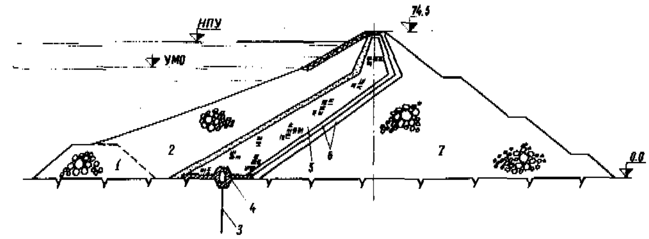

FIGURE 1 Rockfill dam with suglinok core. 1--rock toe; 2--upstream shell; 3--grout curtain; 4--grouting gallery; 5--core made of suglinok; 6--filters; 7--supporting rockfill prism.

Some Dam Quantities in 10^3 m^3

Rock excavation	5,040
Rockfill	3,955
Suglinok core	630
Filters	420
Concrete and reinforced concrete	614

The plan for organizing the operations for constructing the dam was complicated by an extremely nonuniform intrayear runoff distribution of the Vilyuy River. The basic volume of the river runoff (84 percent) goes for the spring-summer period when the maximum flow rates reach 14,000 m^3/s. At the end of the summer and the beginning of the fall, the passage of the flood is also observed with several peaks at maximum flow rates to 2,700 m^3/s. During the winter, the river flow rates drop to 4-2 m^3/s, and during the construction period, a minimum flow rate of 1.5 m^3/s was observed. For passage of flow rates with a 5 percent guarantee during the construction period (8,400 m^3/s) on the left bank rock slope, a diversion channel was excavated.

The technical specifications issued by the design organization permitted the suglinok core to be laid at an outside air temperature of not lower than -20°. However, such temperatures in the construction zone occur only from the beginning of April to the middle of October. In addition, the construction of the gorge section of the dam according to hydrologic conditions and the adopted scheme for passage of the construction flow rates is possible only in the interflood period and after the river was covered, which could be done at the end of October (October 31, 1964).

Within the limits of the construction trench, the dam could be built only after it was covered, which according to hydrologic conditions, taking into account the accumulation of the winter flow of the river in the newly formed reservoir, became possible not earlier than December (December 8, 1966).

Under these conditions, the natural solution was to perform the work during the winter. In the winter of 1964-1965, for the spring flood, the channel section of the rockfill dam was built to a height of 40 m and the suglinok core, to 25 m (300,000 m^3 of suglinok was laid). The spring flood passed through the construction trench. Before the spring flood of 1967, during five winter months, the dam was built to a height of 55 m in the construction trench section (22,000 m^3 of suglinok was placed). This made it possible to begin to fill the reservoir.

The execution of such a large amount of work during the winter became possible because an entire technological complex was developed and carried out with respect to compacting the cohesive soil into high-quality fill at low air temperatures.

The basis for the technological complex was the idea of accumulating in the suglinok during the summer period a quantity of heat from solar radiation that would suffice to keep it in the thawed state at low negative air temperatures for the entire technological cycle of compacting the soil into the core, that is, from the development of the stockpile, transportation, dumping at the site, leveling, and up to rolling--compaction.

The suglinok was prepared in the borrow pit in the following way. In the early spring the bulldozers removed the vegetative layer with removal of the tree stumps and roots and collection of them. As the soil thawed, it was moved by bulldozers into banks, and then it was stockpiled for winter storage to a height of 12-16 m and in a volume of 200,000 m^3 to 250,000 m^3. With this process of preparing and stockpiling, the sandy suglinok and scree fractions were mixed well, and the soil mixture that developed in the large stockpiles had a quite uniform grain composition and relatively uniform moisture content. The peripheral zones of the stockpiles 2.5-3.0 m thick were mixed with salt (15-25 kg of salt per m^3). Salting with calcium and sodium chlorides was used for 20 percent of the volume of the pile. Electric heating of the upper layers was done before excavation in order to facilitate the removal of the thawed ground. For this purpose, use was made of the 36/6 kV 3,150 kW mobile substations. In all, electric heating was used to thaw about 70,000 m^3 of ground at the construction site; of this 80 percent was at 1 kV. It turned out to be most expedient to do electric heating at 1 kV. The consumption of electric power to thaw 1 m^3 of soil when bringing it to a temperature of +8°C was 40 kWh.

In order to minimize the loss of heat by the suglinok during transportation, the bodies of the dump trucks were equipped with double sheathing, and they were heated by the exhaust gases of the engines. The suglinok in the truck body was covered by special "blankets." When transporting it 5.5 km, the soil temperature dropped by 4.2°C to 4.8°C. After unloading at the site, the suglinok was immediately covered with a polymer film. After accumulation of a certain volume of soil, the film was removed and the ground was leveled and compacted using weighted MAZ-525 dump trucks in 8-10 passes.

The preparation of the site and the compaction of the suglinok are two successive operations of the technological complex, the execution of which during the winter becomes a serious problem. In order to clear the concrete and rock base and also the surface of the previously laid soil to remove the ice crust, individual frozen clods, and snow, heating units in the form of discarded aircraft jet engines were successfully used. These engines, which generate an enormous quantity of gas-air mixture with a temperature up to 300°-500°C, were first used in hydraulic engineering construction when building the Vilyuy Hydroelectric Power Plant. Without the application of heating units, it would not be possible successfully to construct the impervious core of a dam under such severe climatic conditions. After heating the surface of the site, it was immediately treated with salt solution. This brought the upper layer of the soil to a plastic state, and at the same time it insured high-quality contact of the compacted layers. In order to remove large individual frozen clods of soil and boulders falling on the site, a special attachment in the form of an overhanding rake was mounted on the bulldozer blade.

The process that was developed and used successfully in construction permitted the achievement of a mean compaction of the soil in the core of a dam of 2,000 m^3/day with a high guarantee (75 percent) of the control value of the dry density of the "melkozem" (fraction less than 2 mm) of 1.6-1.65 g/cm^3. The maximum rate of compacting the suglinok was about 3,000 m^3/day, in March of 1965, and in the April of 1965 it was 3,400 m^3/day. This high rate of compaction was also promoted by the application of the radioisotopic densimeter built by the Orgenergostroy Institute, which permitted the density of the compacted soils to be obtained with high accuracy and speed. On the average, one density control sample was taken for every 75-80 m^3 of compacted soil.

In the especially severe winter of 1965-1966, an effort was made to compact the suglinok in the core of the dam by the same technological process for outside air temperatures down to -50°C to -55°C, but without success. At such low temperatures, the machinery frequently broke down, and there were interruptions in the electric power supply, and so on. For these reasons, it was impossible to insure continuity of the process. Interruptions in the work reached 3, 5, and even 10 days. With such long interruptions, the compacted soil became quite cold, as a result of which, when placing the next layer, the surface of the previously laid layer was not brought to a plastic state, and the freshly placed layer at the contact with the lower one froze before it could be compacted. As the studies demonstrated, the core of the dam became stratified, and this required partial replacement.

Obviously, an outside air temperature of -40°C (this is the limit today) is a "climatic barrier" in insuring high-quality construction of the elements of high-head structures made of cohesive soil in the winter. This limit arises not from the process used, but from the absence of machinery (cranes, excavators, vehicles, and so on) capable of operating at low air temperatures, and other reasons connected with this.

The technical specifications limited the height of the fill layers in the rockfilled prism of the dam (5 m); fines and contaminating material were also limited. Later, the restriction with respect to content of fines was removed, and so-called unsorted rock was placed into the dam. The requirements with respect to preparing the base of the dam were also eased. As for the height of the layer to which the rock was placed in the dam, it was determined by convenience of performing the operations. In different sections, zones, and stages of the work, it was different (10, 15, and 25 m), and it depended on the rate of making practicable excavations, the necessity for insuring a front for laying inverted filters and placing the suglinok, readiness of the approach routes constructed on the steep slopes of the banks for individual stages of placement, and so on.

In the winter the mean monthly rate of dam construction was brought to 11 m in height.

The compaction of the rockfill of the dam was carried out only by passes with loaded trucks. In the summer, for compacting the fill, it was sprinkled with water from hydraulic monitors with a flow rate of 4 m^3 of water per m^3 of rockfill. Subsequently, this method was discarded, since the sprinkling of water is possible only during the short summer period. In spite of the high winter rate of placing of unsorted rock without special compaction, the shrinkage of the dam in 4 yr was 4 to 6.5 percent. About 50 percent of this settlement took place in the first 2-3 months after placement.

The dam was put under the total designed head in the spring of 1967. The results of prolonged observations indicate high quality of the operations and reliability of the core. No seepage through the body of the dam was seen, and this is indirectly indicated by complete freezing of the downstream shell of the dam.

The temperature regimes of the body, and of the base, of the rockfill dam shown in Figures 2 and 3 are of great interest.

CONSTRUCTION OF FROZEN-TYPE WATER STORAGE DAMS

The problems of water tightness and general stability of these dams during their construction

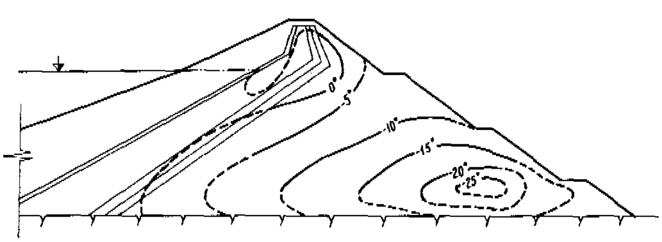

FIGURE 2 Temperature field of the body of the rockfill dam.

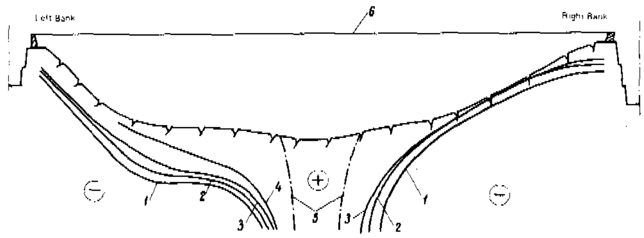

FIGURE 3 Temperature field of the base of the rockfill dam. 1-4--zero isotherms after 4, 3, 2, and 1 yr following filling of the reservoir; 5--outlines of the channel talik under natural conditions; 6--crest of the dam.

on permafrost with respect to the so-called frozen type are solved most radically. Here, the impervious curtain in the body of the dam is created by freezing of the core, which is adjacent to the permafrost in the base of the dam with respect to the entire longitudinal profile. The core, the lower wedge, and the base under these elements of the dam are continuous; during the entire period of operation and maintenance of the structure, they are kept in the frozen state.

The Vilyuygesstroy Construction Administration has constructed two such dams, 20 m and 17 m high, creating reservoirs with capacities of 12 million m^3 and 6.5 million m^3, respectively. The third dam, 22 m high, is being built to create a reservoir with a capacity of 35 million m^3.

Coarse Quaternary deposits up to 8 m thick in the permafrost condition are used as the base of the first of these dams constructed on the Iyerelyy River and remaining unique even today. The ice content is 60 percent. There are individual ice lenses 15 cm thick and more. The bedrock is represented by jointed weathered dolomitized limestones and marls. The fissures in them are filled with ice. The core of the dam penetrates the loose Quaternary deposits, and the core, made of scree-suglinok, rests on bedrock.

The dam was constructed by the thawing method, but with subsequent freezing of the core. The device for freezing the core of the dam is a system of coaxial freezing columns. These columns are installed in drilled holes with spacing up to 1.5 m. The freezing columns are grouped into individual systems of 16-30 holes each as a function of the depth of the curtain in the given section. Each such system is serviced by a fan blowing cold outside air into the holes. During circulation of the cold air through the pipes of the freezing columns, cylinders of frozen ground are formed around them, which, increasing in diameter, approximate a solid wall. In the dam on the Iyerelyy River, during the first year of operation, complete joining of the cylinders did not take place. The next winter, the thawed windows froze shut, and a continuous curtain of frozen ground up to 6 m thick was created. Three years after the beginning of operation and maintenance of the freezing sys-

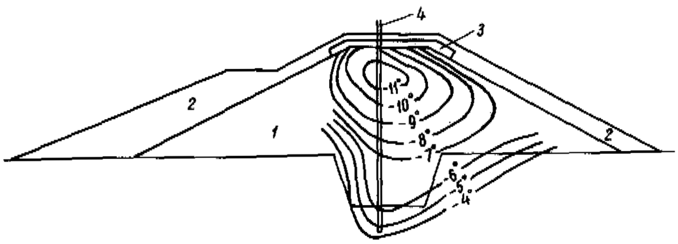

FIGURE 4 Temperature regime of the body of the frozen dam. 1--suglinok; 2--rockfill; 3--peat; 4--frozen curtain; -11° to -4°--isotherms 2 yr after the beginning of freezing.

tem, the thickness of the curtain reached 10 m, and so the operating time of the freezing system was cut in half.

During the initial period of operation, the systems were put into operation only at temperatures below -15°C, and then only at an air temperature below -20°C to -25°C.

In Figure 4 we show the temperature field in the body of the dam 2 yr after the beginning of its freezing and operation under head.

The dam was equipped with measuring and control equipment that ensured exhaustive information about the temperature regime of the body and base of the dam, and it permitted operative solution of the problems of inclusion of the entire system or part of it into operation.

In the base of the second dam of the frozen type constructed on the Oyuur-Yureg River, where it emergences from the lake system of the same name, sandstone and volcanic rock were used, and at the contacts, metamorphosed rock. The bedrock was covered from above by icy colluvial formations. The core of the dam made of supes cuts the coarse cover formations and rests on the competent bedrock. The lateral prisms are filled with rock from the excavation of the discharge channel and quarried rock. The dam took the entire head in 1972.

The spillways were made just as on the Iyerelyy River in the form of an open channel going around the dam on the left bank.

The system for freezing the core of the dam does not in practice have structural differences from the same system made for the dam of the Iyerelyy River.

When building frozen-type dams, specialists were called in for consultation from the Permafrost Institute of the Siberian Department of the USSR Academy of Sciences (P. I. Mel'nikov, K. F. Voytkovskiy, F. E. Are, R. M. Kamenskiy, and V. I. Makyrov) and the All-Union Vedeneyev Scientific Research Institute of Geology (N. N. Petrunichev and G. S. Shavrin).

CONCLUSIONS

1. The experience of building the Vilyuy Hydroelectric Power Plant proved in practice for the first time the possibility of constructing large hydroelectric power plants under the extremely complicated conditions of the northern parts of our country.

2. When building the dams of the Vilyuy Hydroelectric Power Plant, a process was developed for constructing impervious elements of large dams made of cohesive soil under severe winter conditions at air temperatures to -40°C.

3. The expediency of erecting dams in the Far North from local building materials was proved.

4. The construction of large structures under the conditions of the North requires the application of special machines, machinery, and technical devices capable of operating at temperatures to -50°C and lower.

[Interested readers who would like further important details should see: "Hydrotechnical Construction," no. 2, Feb. 1970 (a translation of Gidrotekhnicheskoe Stroitel'stvo, no. 2, pp. 17-24, Feb. 1970). "G. F. Biyanov-Construction of a rockfill dam for the Vilyuy Hydro-Electric Project." A good reference for the Vilyuy Dam (p. 170) is the Technical Trans. no. 1353 by the National Research Council of Canada: "Construction of a water-storage dam on permafrost" (Gidrotekhnicheskoe Stroitel'stvo [10]:15-23, 1965) and "Discharge and blocking up the river during construction of the Vilyuy Hydro-Electric Power Station" (Gidro-Tekh. Stroitvo [2] 1-5, 1966). Both papers are by G. F. Biyanov, translated by V. Poppe and technically reviewed by G. H. Johnston of the Division of Building Research, Ottawa.]

GEOCRYOLOGICAL CONDITIONS AND PROCEDURES FOR LAYING THE NORIL'SK-MESSOYAKHA PIPELINE

P. I. MEL'NIKOV Permafrost Institute of the Siberian Department of the USSR Academy of Sciences
F. G. BAKULIN AND YE. G. KARPOV Igarskaya Scientific Research Cryologic Station
A. A. KOLESOV Fundamentproyekt Institute

The northernmost pipeline in the world--from Noril'sk to Messoyakha--is 263 km long and 720 mm in diameter and was laid through the tundra, which is difficult of access. There are more than 60 water obstacles along its path. The largest of them is the Yenisey River, which is 47 to 64 m deep and more than 2 km wide at the crossing; the Norilka; the Bol'shaya Kheta; and the Malaya Kheta, 5 to 10 m deep.

The pipeline was planned by the Vostokgiprogaz Institute under the USSR Ministry of the Gas Industry, with the participation of the Fundamentproyekt Institute under the USSR Ministry of Installation and Special Construction. The studies of the geocryological conditions and the development of recommendations with respect to methods of laying the pipeline were participated in by the Order of the Red Banner of Labor Institute of Permafrost of the Siberian Department of the USSR Academy of Sciences, its Igarskaya Scientific Research Cryologic Station, and other organizations.

The Noril'sk-Messoyakha line is located in the polar region in an area with severe climatic and complicated geocryologic conditions.

The duration of the cold period is 245 days. The lowest temperature of -53°C was noted in January 1968, and the highest temperature (+32°), in July 1970.

The polar night with pale twilight noons (the sun below the horizon) lasts about 1.5 months (December to the first half of January). From May 20 to August 20 is the polar day (the sun does not set below the horizon).

The flat divides and the plains between the rivers are covered with bushy, spotty, polygonal swampy, and lacustrine tundra. The flat slopes of the troughs were grown up in sparse deciduous forest. The Noril'sk depression is represented by azonal northern taiga mixed with spruce-birch deciduous forest. The flat plains of the large rivers are covered with a dense growth of alder to 3.5-4.0 m high.

The pipeline routing passes through the region primarily with continuous permafrost from 100 to 200 m thick in the vicinity of the Messoyakha deposit and to 300-400 m thick on the watersheds of the left and right banks of the Yenisey.

On the watersheds, the permafrost series is basically continuous with respect to area and depth, and it is of low temperature. In the river valleys, it is frequently broken by through- and deep-taliks, developed not only under the river channels and the deep lakes, which do not freeze to the bottom, but also in the thickly forested and overgrown sections of the floodplain. The taliks were formed by the warming effect of the water bodies, the snow cover, the vegetation, and groundwater under pressure. The area of the taliks in the floodplains of the large rivers is about 30 percent.

The region through which the pipeline runs can be divided in geocryological respects into three sections: the left bank and right bank plain of the Yenisey and the Noril'sk Valley.[1] The geocryological conditions of the first two regions are similar to a great extent. The permafrost is characterized by a temperature (at the level of zero annual amplitude) of -4° to -5° in mineral soil and to -6° to -7° in peats. In the Noril'sk Valley, the permafrost has less thickness by comparison with the Yenisey plain, and it is interrupted by a large number of water-bearing through-taliks.

Under the channels of the deep rivers (the Yenisey, the Norilka, the Greater Kheta, and the Lesser Kheta), including the flows that do not freeze to the bottom 1.5-2.0 m deep, there are through-taliks. They are traced in the valleys and in the more shallow mountain rivers: the Talnakh, the Ambarnaya, and so on, the bottoms of which are made up of coarse gravelly-pebbly material with a suglinok-supes matrix.

It has been established that the separating effect of the rivers on the adjacent frozen ground is not felt within the limits of the entire valley, but directly under the deep (to a depth of more than 1.5-2.0 m) part of the channel where the through-talik is formed. On the river banks, the frozen rock is prominent under the river channels within shallow shore parts in the form of "frozen crust."

The boreholes at the shallowest places of the Norilka River have revealed relatively low-temperature, ice-saturated recent beds of frozen soil from 20 m to 30 m thick, and the minimum temperature of the frozen channel-deposits at the beginning of September 1968 at a depth of 5 m was about -3°.

On the floodplain, grown over with willow and treelike alder, where a significant layer of snow accumulates in the winter (to 3-4 m), the thickness of the frozen series by comparison with the channel sandbars is reduced sharply (to 10 m), and the ground temperature rises to -1°. Because of the warming effect of the snow cover, the frozen ground on the islands in the river formed before the brush either completely disappeared or the upper water level dropped to a depth of more than 5 m.

In the first suprafloodplain terrace of the Norilka River, which ends directly at the channel, the frozen ground is everywhere. Here, by comparison with the channel sandbars, the geocryologic conditions become milder because of the warming effect of the loose snow cover, which reaches a thickness of about 1 m in the forest. With respect to the entire depth of the frozen series (to 20 m), practically a gradientless temperature distribution of not less than -0.7° is observed. In the bluff itself, within the limits of the peat mounds, the thickness of the permafrost increases (more than 30 m), and the temperature drops to -4° to -5°.

The broad development of the permafrost is noted on the lower floodplains and the islands and sandbars of the Yenisey, Bol'shaya Kheta, and Malaya Kheta rivers.

Under the large thermokarstic lakes frozen to the bottom (Semenovskoye, Alykel', the Dolganskiye Lake group, and so on), there are obviously deep (nonthrough) taliks.

For example, let us present the geocryologic section of the thermokarstic Semenovskoye Lake 350 × 500 m and 4 m deep located in the lower course of the Bol'shaya Kheta River (Figure 1). On the floodplain of the lake grown over with brush, where the winter brings the accumulation of a significant layer of loose snow 2-3 m thick, permafrost has been found to a depth of 9.6 m. On an island 40 × 60 m and 2 m high, the thickness of the permafrost is 18.5 m, and the minimum temperature at the end of August is -4°. On the opposite, eroded, shore of the lake made up of peat and inclusions of multiple-vein ice, low-temperature permafrost is also noted.

At a reference well 25 m deep at the top of a watershed, 150 m from the lake, low-temperature frozen beds have been discovered with a minimum temperature of -8.4° (August 25, 1968) at a depth of 6 m. The temperature of the frozen ground at the foot of the layer of zero annual amplitude, at a depth of 16 m, was -5.2°.

Thus, when approaching the deep lakes not frozen to the bottom, the permafrost sharply tapers out. The lateral warming effect of the deep-lake water bodies is insiginificant and extends to the depth of the shore deposits within no more than 10 m from the water's edge. In the shallow water areas near the shore, where the lake freezes to the bottom, the permafrost is either retained or reformed.

The thickness of the seasonally frozen layer along the pipeline route fluctuates within significant limits and depends on the local conditions. The seasonal thawing of the ground begins at the end of May to the beginning of June and ends in September. The peat along the pipeline route occupies no less than 40 percent of the area (swamps, flat and hilly peat bogs). It thaws to a depth of 0.2-0.4 m in the peat bogs and 0.6-1 m in the flooded peat bogs.

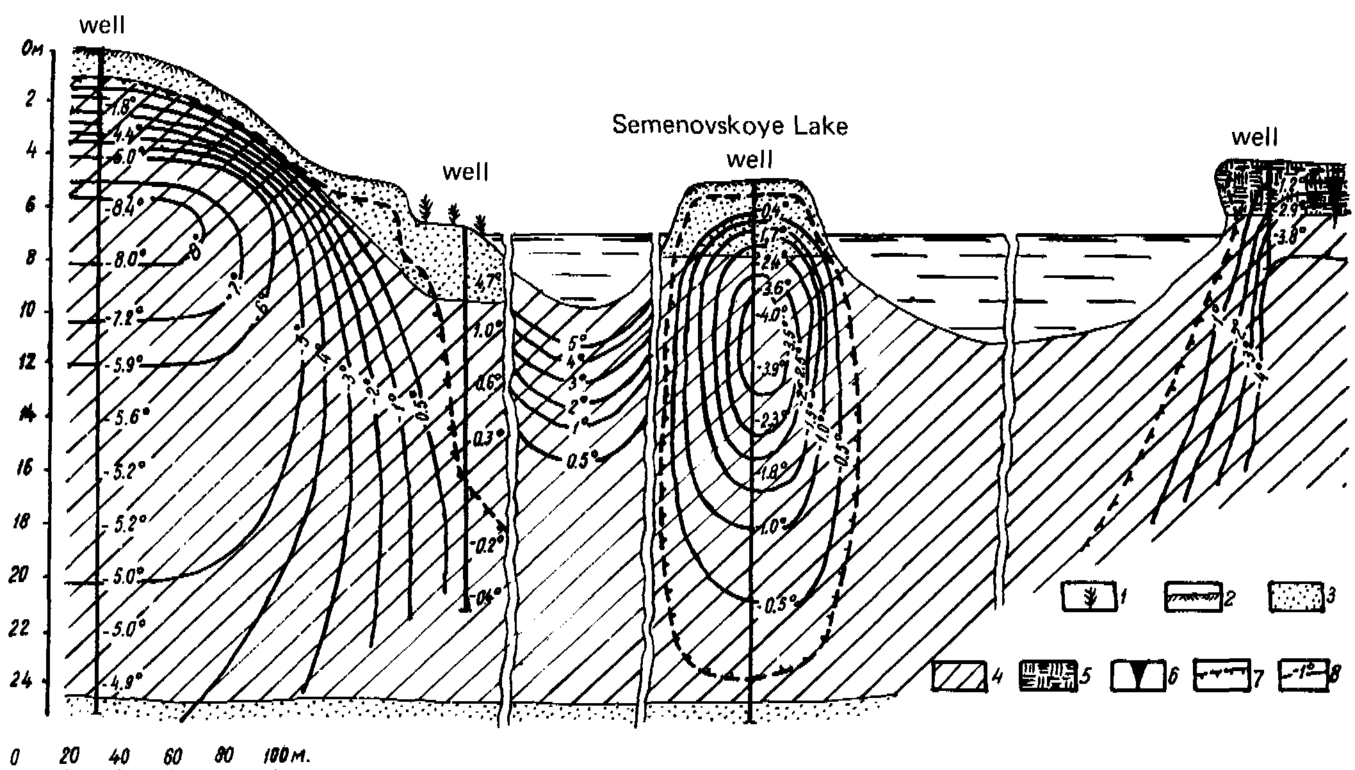

FIGURE 1 Geocryologic section through the Semenovskoye Lake (by the data of August 25, 1968). 1--brush; 2--moss-plant layer; 3--sand; 4--suglinok; 5--peat; 6--multiple-vein ice; 7--permafrost boundary; 8--isotherm.

The suglinok and supes soils of the tundra sections thaw to a depth of 0.5-1 m. The thawing depth depends on the thickness of the peat-moss layer. With a thickness of this layer of 0.2-0.4 m, the soil thaws no deeper than 0.6 m, and, where the moss cover is thin and discontinuous, the suglinoks thaw to 0.8-1.0 m and in the medallions, 1.0-1.3 m.

The supes and sandy soils on the shore sandbars of the rivers and lakes thaw to a depth of 1.2 m with a grass cover and to 1.5 m where there is none. As a result of destruction of the natural plant cover, the thawing increased by 2.0-2.5 m in the peat bogs, and by 1.3 times in the suglinok soils.

The permafrozen clayey and peaty soils are characterized by increased ice content (moisture); their upper horizons are especially ice-rich--to a depth of 5-6 m from the daylight surface. The total moisture content of the mineral soil often reaches 60-80 percent, and in the peat, several hundreds of percentages. During thawing, the ice-saturated soil gives a large degree of subsidence (up to 30 to 40 percent of the thickness of the thawed ground). The cover of supes and suglinok, which has thixotropic and heave properties, is widespread.

The underground ice along the route of the pipeline is segregated, stratal, and multiple-veined. The ice in the peat bogs and mineral soil is in the form of interlayers and pockets and also in the form of beds to 0.5-1.0 m thick. In the polygonal peat bogs, the multiple ice wedges, with a depth of 3-5 m and 1-2 m wide at the top, are well developed (Figure 2). For construction of the pipeline, the beds of underground ice that are not far (0.5 m) from the daylight surface (Figure 3) are especially dangerous. Their thickness reaches 1.0-1.5 m, and their width is from 10 m to 50 m.

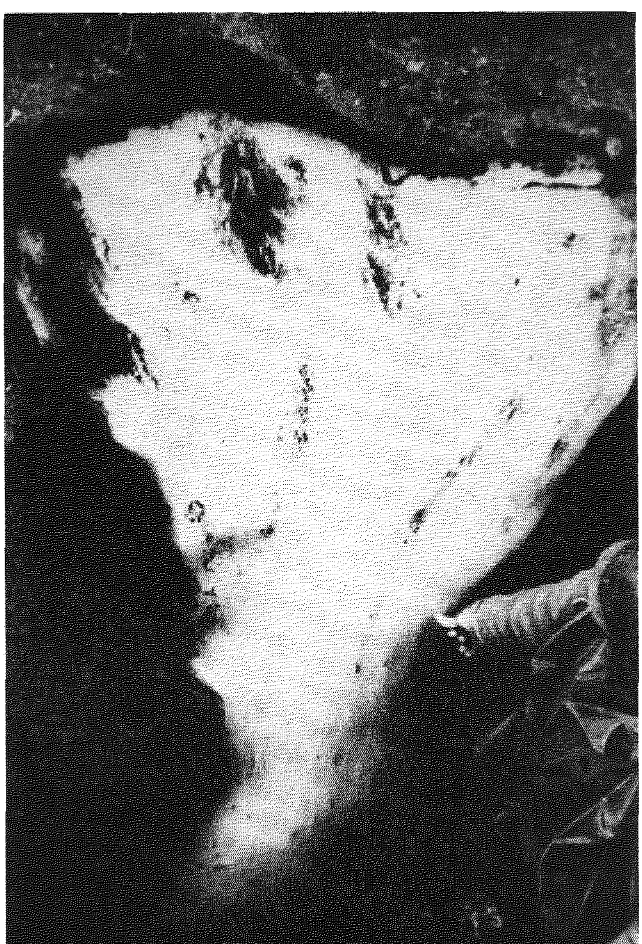

FIGURE 2 Multiple-wedge ice in the peat bogs.

FIGURE 3 Bedded ice at a depth of 0.5 m.

The cryogenous processes connected with freezing and thawing of the ground (thermokarst, heave, solifluction, frost fissuring, ice bodies, and so on) strongly occur along the pipe route. Under the condition of maintaining the natural vegetative-soil cover, the cryogenous processes are in a state of relative stability and dynamic equilibrium. The development of modern thermokarst is primarily connected with economic development of the region, and frequently it is caused by disturbance of the vegetative cover, which leads to thawing and subsidence of the shallow ice-saturated permafrost and the stratal and multiple-vein ice included in it.

In the winter, the snow is transported by the wind, and it is redeposited on the ground surface. This fact, together with the low air temperatures in the winter and high in the summer, must cause nonuniformity of thawing and freezing and different temperature conditions of the soil.

Considering the above natural factors and the thermal and mechanical effect of the pipeline on the ground and possible variations of the geocryologic conditions during operation, the primary principle of construction in the sections with permafrost, the principle of keeping the ground in the frozen state during the entire construction and operation period, is recommended.

Depending on the geocryologic conditions and the selected type of use of the ground for support, suspended, surface, and underground methods of laying the lines are used. The first two methods were recommended primarily for the linear part of the pipeline (on the watersheds, the divides, the high terraces), and the third method, at the approaches and crossings over water barriers.

In aboveground installation of the pipeline, the natural geocryologic conditions of the route are disturbed the least, and stability of the line and the possibility of convenient technical inspection of it during its operation are insured. For this type of laying, two basic types of supports are recommended: pile and surface trestles and frames. For almost all ground conditions, they provide stability of the supports and, consequently, normal operation of the pipeline.

The piles are installed: (a) in a large-diameter hole with the annulus filled with slurry, (b) driven directly into the ground or in leader holes of smaller diameter than the pile diameter, or (c) driven into thawed ground (using steam needles). The last-mentioned procedure is applicable only in sections with low-temperature ground where rapid cooling of the warmed ground and freezing of the piles to it is ensured and also when using cooling units. The bearing capacity of the piles is determined primarily as a function of the maximum temperature and composition of the soil in accordance with the existing Norm documentation (SNiP II-B 6-66) [*Construction Norms and Guidelines*, commonly called "SNiPs" in North America]. With adequate equipment drilling devices, steam units, and plant for driving and installing the piles, this method of construction becomes most economical and practical.

In the absence of the necessary construction equipment and the presence of favorable permafrost conditions where the settlement of the ground during thawing and heaving during freezing do not exceed the admissible deformations of the pipe, it is possible to use surface supports. It is necessary to note that such supports belong to the category of "pulsating" supports and are installed on the layer of seasonally thawed (or seasonally frozen) ground experiencing two types of deformations:

1. Settlement of the frozen ground during thawing in the summer.
2. Heaving of the freezing upper layer of the ground during the winter.

Along the length of the pipeline, these deformations are nonuniform; soils have different moisture content, ice content, and composition.

The pile supports were recommended in the form of trestles made up of piles (wooden, reinforced-concrete, or metal) and crossbeams (wooden or metal).

The majority of supports are made of wooden piles (larch, 22-25 cm in diameter) with mandatory protection of the structure from rotting. In most cases the tie-beams are made of metal, and wooden tie-beams are reinforced with metal plates.

In all cases the piles were designed by the bearing capacity of the pile material and the soils, for the effect of the heave forces and the effect of horizontal forces.

The bearing capacity of wooden piles 22-25 cm in diameter and 4-6 m long, depending on the type of ground and temperature, varies from 15 to 20 tonnes (see Table 1).

It is recommended that wooden piles be used in the following cases (with respect to structural requirements):

1. For fixed [anchor] supports if the depth of seasonal thawing does not exceed 0.6 m and the height from ground level to the bottom of the pipe is up to 0.4 m.
2. For middle [movable] supports, with a depth of seasonal thawing to 1.5 m and a height from ground level to the bottom of the pipe up to 1 m.

Reinforced-concrete piles 30 × 30 cm and 6-8 m long were recommended for use in the following permafrost conditions: (a) ice-rich suglinoks, supes, and clay with a total moisture content of 40 percent and more; (b) peat bogs to 4 m deep; (c) boggy sections; (d) ground with the inclusion of vein ice close to the surface; (e) slopes made up of icy-suglinok, supes, clay, and silty-sand; and (f) transitions through lowlands and small streams.

In construction of supports on peat bogs, swamps, and in sections with vein ice, it is necessary to bury the piles not less than 1.5 m

TABLE 1 Pile Dimensions Depending on the Permafrost Conditions (Excluding River Crossings)

Type of Soil	Temperature at the Level of Zero Annual Amplitude °C	Maximum Depth of Seasonal Thawing, m	Recommended Depth of the Pile in the Ground (from the Ground Surface), m
Suglinok and supes	-3 and lower	to 1.0	2.5
		to 1.5	3.5
	-2 and lower	to 1.0	3.0
		to 1.5	4.0
Sand	Independent of the ground temperature	to 2.0	3.0

NOTE: Wooden piles are used when the height of the pipeline above-ground level is no more than 1 m.

below the layer of weak soil or of ice.

Surface supports in the form of cribs were recommended for use only for the middle supports (excluding the anchor supports) and in the following types of ground: sand, coarsely fragmented, rocky, and semirocky.

The surface supports were designed by the bearing capacity of the material, the thawing layer of soil, bending from the effect of horizontal loads, and the settlement of the active layer.

The evaluation of the bearing capacity and the subsidence of the ground in the active layer has greatest significance when selecting the supporting area of the surface supports.

Ground-surface installation, maintaining the frozen state of the ground, is also realized on surface supports (frames, ground plates) and embankments in the sections with deep seasonal thawing (more than 2 m) or with deep permafrost, and also continuous fill in the sections with ice-rich soil and a thin active layer (to 1 m). The ice-saturated ground of the active layer almost completely loses its bearing capacity during thawing. Therefore, the frames and ground plates must be installed on equalizing sandy or gravel fill to redistribute the pressure over the surface of the thawing layer and to prevent the supports from nonuniform deformations during thawing and freezing.

The laying of pipelines on fill is promising for any permafrost conditions. It was recommended to be erected on sand and better, gravel or detritus, fill not less than 1.0-1.5 m high. As observations on the existing roads demonstrated, with that height of fill, the frozen state of the soil is retained, and for higher fill, even a frozen nucleus is formed in the body of the fill. With less height, the frozen ground thaws under the fill. However, as a result of the absence of the necessary quarried material, it was necessary to stop using fills.

For underground laying, the pipeline has the greatest thermal effect on the bearing ground, to say nothing of the fact that in this case, just as when placing fills, it is necessary to do a large volume of earthwork.

The thermal calculations for the thawing depth under the line 0.7 m in diameter for the case of underground laying, performed by L. P. Semenov and Yu. L. Shur, demonstrates that on variation of the gas temperature in the line in the summer from +10° to +15°, a bearing-ground temperature from -1° to -4°, and a depth of pipe from 0.5 m to 1.75 m (to the top of the pipe), the thawing aureole of the ground under the line varies within 0.4 m to 2.1 m. The calculation was performed beginning with the following assumptions: (a) the gas is transported through the line with a positive temperature only in the summer, as a result of which the thawed ground completely freezes in the winter; (b) there is no flow of the water along the trench or convective heat transfer from the line into the ground; (c) the ground temperature distribution with respect to depth and around the line is close to steady-state.

In the case of underground laying of the line in the very ice-rich (with an ice content of more than 70-80 percent by volume) ground and in the sections with stratal and vein ice, the thawing aureole significantly exceeds the above values. Here, the thawing process is irreversible. This leads to the development of thermokarst. In such sections, it is not recommended that underground and surface installation of the gas line be used.

With underground installation, unavoidably, especially in the summer time, the natural vegetative cover is disturbed to the greatest degree; the open ditch serves as a channel for runoff of the surface water promoting additional thawing of the ground. In the case of underground laying, it is necessary to do large amounts of expensive work to replace the ice-rich ground, the creation of clay dams preventing the penetration of the surface water and its movement through the trenches, and the construction of thermal insulation at the top and bottom of the trench.

In order to avoid the development of deformations from differential settlement of the frozen ground during thawing by maintaining the natural vegetative cover, basic operations must be planned with respect to constructing the line during the cold part of the year (from October to May), and for summer work it is necessary to build approaches for hauling the machinery and equipment 50-100 m from the base structures. When putting in pipelines in fill or laying them along the fill in individual sections, which are difficult of access, it is necessary to place the fill by end-tipping. When this operating condition is observed, the fill can be placed year-round. To a significant degree these measures increase the reliability of structures erected under complex geocryological conditions.

It was recommended that the line be taken around the peat massifs and the peat and mineral heave mounds with the vein and stratal ice associated with them encountered along the route, and, if it is impossible to avoid them, then to take special measures to keep the peat, and the usually ice-rich lacustrine-bog clay soils underlying it, in the frozen state.

In the sections with thawed ice, any method of laying the line is possible, but it is necessary to consider deep freezing and the large tangential and normal heave forces developing with it.

When constructing the pipeline in all sections with permafrost, the method of keeping the bearing ground in the frozen state was adopted. On the linear part of the pipeline, the most reliable aboveground procedure in this respect, basically pile supports, was used (Figure 4) at the crossings over small streams and ravines--on cribwork--and on the raised topographic elements under the conditions of broken terrain, partially on framing.

At the crossings of the major rivers (the Yenisey, Norilka, Bol'shaya and Malaya Kheta), with their wide floodplains and islands, with spring floodwaters, and ice-jams during the breakup and the sandbars with permafrost developed on them, the underground method of laying was used. Directly in these river channels, the pipeline

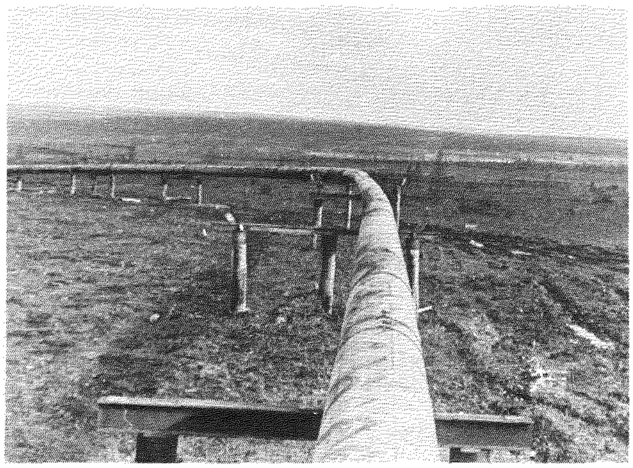

FIGURE 4 Pipeline on pile supports. Photographed by L. I. Popov.

was laid along the bottom, with necessary weighting, in addition to the natural icing on them, as an ice crust 30-40 cm thick formed around the pipe. The length of the inverted siphon along the bottom of the Yenisey reaches 2,200 m, and in the Norilka and Bol'shaya Kheta, 400 m.[2-4]

During the time of operation and maintenance of the Messoyakha-Noril'sk line, it was established that the most reliable means of laying the line was on pile supports.* In this case the natural state of the bearing ground is disturbed to the minimum. As a consequence, the majority of the surface supports were replaced by pile supports as the most stable and reliable in operation. Therefore, the second pipeline will be built only on piles, with metal tie beams and [composite] wooden piles; the upper part of each pile is replaced by a metal fitting made of pipe to the depth of seasonal thawing.

*[Extreme wind-induced resonant vibrations caused several fractures in pipe suspended from pendulum type, A-frame supports. Soviet Rossiya, 9 Oct. 1970]

A pipeline built in the polar region with severe climatic and geocryologic conditions is unique. The experience in building and maintaining it has exceptionally great significance for the future of such structures. Therefore, the Cryologic Service has organized observations of the behavior of the line, and it has equipped several typical experimental sections for accumulation and generalization of the experience of building and operating such a line on permafrost.

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[Ref. No 3 and 4 may be found in translation in: "A Survey of Soviet pipeline construction techniques and related activities in Western Siberia. Williams Brothers, Tulsa. This volume has 31 valuable translations on Pipeline Engineering in Permafrost Regions. FJS.]

PIPELINE CONSTRUCTION IN PERMAFROST REGIONS

V. V. SPIRIDONOV AND L. P. SEMENOV VNIIST Institute

B. L. KRIVOSHEIN VNIIGAZ Institute

The peculiarities of constructing pipelines in permafrost zones arise from the climatic, geocryologic, hydrogeologic, technologic, and economic factors. The northern parts of the country are characterized by a long winter with an air temperature to -65° and a short but relatively warm summer with temperatures to 30°, high winds to 40 m/s, frequent snowstorms, and significant swampiness of the region.

The presence of permafrost causes a number of phenomena that complicate the construction, operation, and maintenance of the pipelines. The silty and fine-grained ice-rich permafrost loses its bearing capacity with thawing, and as a result of the thermal effect of the pipeline, thermokarstic phenomena can develop subsidences, lakes, ravines, and landslides. The freezing of the thawed layer of ground can be accompanied by heave, the formation of frost fissures, ice bodies, and so on. The permafrost, being impervious, promotes high supersaturation of the ground of the seasonally thawing layer.

The development of pipeline transportation in various parts of the permafrost zones, the growth of pipeline diameters, and the change in the technological regimes of their operation connected with this dictate the necessity for developing new technical solutions for constructing pipelines and testing them under natural conditions.

ENGINEERING-GEOLOGICAL STUDIES

When planning pipelines in permafrost zones, a great deal of significance is attached to obtaining reliable information during the process of discovering the engineering-geologic and permafrost conditions along the route.

In the presence of permafrost, the engineering-geological studies along the route of the pipeline have the following specific nature:

1. In addition to the ordinary exploratory operations, it is required that special geocryologic studies be made, the volume of which is established as a function of the diameter of the pipeline, the proposed method of laying, and the complexity of the section of the route.
2. The width of the strip along the route in which the exploration is carried out is appreciably greater than under ordinary conditons in all stages of planning and designing the line. It depends not only on the number of loops in the gas line and the pipe diameter, but, primarily, on how complex the conditions of the region are in engineering-geologic and permafrost respects.
3. Engineering-geologic studies under the line must be made considering the possible structural solutions and the proposed method of laying, since the volume and content of the exploration required for planning, designing, and building the pipeline depend on those studies.
4. In connection with the high dynamic nature of the cryologic conditions when planning and designing the pipeline, it is necessary to begin not only with the characteristic of the frozen layers at the time of the studies, but also to predict all changes in cryologic conditions that can occur under the effect of the climate dynamics and with variation of the natural environment during the process of construction and maintenance of the pipeline.
5. Lack of knowledge of the region, difficulty of access, and the great lengths of the routes require the application of the latest methods of exploration--aerial photography, superhigh frequency [SHF] radiometry, geophysical methods, and the application of computers to select the optimal versions of the gas lines.

The studies are made in all stages of planning and designing the pipeline. In the initial stage, for the selection of the optimal version of the route, cryologic maps on a 1:100,000 scale are compiled with respect to all prospective directions of the routes. These maps are compiled on the basis of the interpretation of the aerial photographs and analysis of the cartographic materials, and they must contain the necessary information about the geocryologic conditions in all sections of the route. The selection of the optimal version of the route must be made by running versions on a computer, considering the entire set of conditions and proposed structural solutions and operating conditions of the gas line.

Thus, by the beginning of exploration, the technical design must be supported by materials characterizing the cryologic conditions of a quite broad strip in the prospective directions and possible versions of laying the line along the entire length of the route with indications of the temperature regime of the gas with respect to seasons of the year.

The studies in the technical design stage must be made only with respect to selected optimal versions of the route. In the preparatory period of the studies the following items are taken care of:

1. Approximate forecasting calculations for evaluating the variation of the cryologic conditions (mean annual temperatures and depths of seasonal freezing-thawing), depending on the variation of the conditions of heat exchange on the surface and under the effect of heat release from the pipeline, and also calculations of the thawing aureoles, the thickness of the compressed zone of soil under the structure, and so on. These problems are solved on the basis of the proposed structural plan of the gas line and the technological regime for its operation.
2. The decoding of the aerial photographs and compilation of maps of the geologic-geographic microdistricts on the basis of the topographic geologic and geomorphologic maps reflecting the propagation of defined complexes of natural conditions in the optimal directions of the routes, the basic elements of which are the geologic and geomorphologic structure of the region, the hydrogeologic conditions, climate, snow and plant cover, and so on.
3. Determination of the number and points of the characteristic supporting sections on the microdistrict map for performing field operations, the width of the investigated strip, and the volume of engineering-geocryologic studies, depending on the complexity of the permafrost conditions.

The field studies under the lines are made in two stages.

The first stage is the cryologic survey on a scale of no less than 1:10,000 for final selection of the route.

The second stage is the linear cryologic studies with respect to the noted routes.

In the first stage, within the limits of each microdistrict, a detailed study was made of the laws of propagation of the permafrost series in plan and profile, the formation of the temperature regime of the rock, the magnitude of the seasonal freezing and thawing, and the cryogenous structure of the various genetic types of rock. The data from the study are gathered by geologic, geomorphologic, geophysical, and other methods. The microdistrict map is more precisely defined simultaneously.

During the studies under the pipelines, it is necessary to consider the following: (a) the laws of propagation, the nature of occurrence, and the variation of the basic natural factors within the limits of one microdistrict, such as the radiation-thermal balance, the plant and snow cover, the relief; (b) the intensity of variation of each factor within the limits of each type of terrain during the design service life of the line; (c) the composition, cryogenous structure, physicomechanical and thermophysical properties of the rock, and the dynamics of seasonal thawing within the limits of various microdistricts; and (d) the laws of propagation and degree of expression of the cryologic phenomena in their relation to the defined set of natural conditions.

In the second stage, linear cryologic studies are made along the planned pipeline routes considering the proposed method of construction and the operating and maintenance regime. Here, a set of geologic, cryologic, engineering-geologic, climatic, and other methods of study are used.

In the vicinity of the crossings over the large river valleys in connection with the necessity for evaluating the various sections to discover the optimal version under complex cryologic-ground conditions, it is recommended that an area survey be made, the dimensions of which are determined by the cryologic conditions, the width of the valley, the nature of the stream, the structural peculiarities of the crossings, and other conditions.

During the survey, the width of the strip in individual sections can be increased in order to define more precisely the direction of the route in connection with the adopted structural plan and the operating conditions of the pipeline. Under especially unfavorable permafrost conditions, it is necessary to do the studies in several parallel sections.

During the laboratory processing of the field data from the exploration, laboratory studies are made of the properties of the soil, the thermophysical characteristics of the permafrost are calculated and determined, and a series of problems is solved with respect to determining the degree of variation of the cryologic conditions in connection with the natural dynamics of natural factors and the effect of the pipeline.

PECULIARITIES OF PLANNING AND DESIGNING PIPELINES ON PERMAFROST

The planning and design of a pipeline on permafrost is distinguished by the following: (a) the necessity for performing geocryologic studies in all stages of the planning of the pipeline, the interrelation of the volume, and results of these studies and the structural solutions and technological process for performing the operations used; (b) a complex approach to the exploratory research and the necessity for forecasting the nature of variation of the cryologic conditions over a long period of time; (c) the necessity for adopting structural solutions, methods of construction, and operating technological regimes during planning and design work that permit the course of the variation of the cryologic conditions to be influenced in the required direction to insure the reliability of the line during the entire operating and maintnenance time; (d) the dependence of the selected efficient methods of laying the pipelines on the technological regime of operation and maintenance and also on the permafrost conditions, the relief of the terrain, and the rigidity of structure of the line; (e) mandatory development of methods of doing the operations during the planning and design of the construction process, the time for fulfillment of the operations, the sequence of various types of operations, and so on; (f) the necessity for recommending methods of maintaining and procedures for restoring disturbed conditions on the surface when planning strict execution of the operations in the plan; (g) very rigid requirements on the pipe material, the machinery, and the technological equipment that must insure reliable operation of the line under the conditions of severe climate and permafrost; (h) the necessity for developing engineering-technical measures in the plan that will insure proper operation and maintenance of the gas lines in accordance with the adopted structural solutions and the construction principles.

The proposed structural solutions of the pipelines are being developed on the basis of the small-scale cryologic map using technicoeconomic comparisons of the alternatives. Hereafter, when performing field cryologic studies, these solutions can be changed. The optimal version insuring high reliability need not correspond to minimum cost.

The planning and design in the working-drawing stage, the final selection of the structural solutions, and the technological process for transporting the gas are carried out not only beginning with the permafrost conditions at the time of performing the studies, but primarily considering their variations and forecasting the development of the cryologic processes for the planned operation and maintenance.

FORECASTING THE INTERACTION OF THE PIPELINES WITH THE ENVIRONMENT

Forecasting the interaction of the pipeline with the environment includes the study of variatons of the permafrost conditions as a result of the construction, operation, and maintenance of the line; the thermal and mechanical interaction of the pipelines with the permafrost; all possible technological regimes of the operation; the structural solutions for the pipeline and various permafrost conditions of freezing the pipelines at underwater crossings over rivers and when cold gas flows through them; and heat exchange between the pipelines above the ground and the air, considering solar radiation, wind velocity, and so on.

When forecasting the variations of the permafrost conditions, qualitative and quantitative estimates are made: (a) the variation in natural conditions of the vicinity of the route as a result of building the pipeline (removal of vegetation, leveling the terrain, variations in snow accumulation, ground composition and properties, and so on), considering the proposed methods of laying and the structural solutions; (b) the thermal effect of the pipeline on the frozen ground in the vicinity of its direct effect during the operation and maintenance process and the intensity of development of heave, settlement, ice bodies, and other cryogenous processes. In addition, it is necessary to consider the possibility of aquifers, centers of discharge of groundwater, and swamps along the route.

The forecasting characteristic can be obtained during the process of developing the design assignment when, depending on the natural conditions of the sections, the structural solutions of the pipelines are selected and the operating and maintenance regimes are determined by the placement of compressor stations, cooling units, and the like along the route.

The forecast of the variation in permafrost conditions is made in two steps--approximate and detailed.

During the approximate forecast, the direction in which the cryogenous and engineering-geological processes develop under variable conditions is determined, and the volume of research is noted. The quantitative estimate of the appearance of these processes with known dynamics of the temperature regime is made using approximate formulas. The required calculated characteristics are selected on the basis of studying the background material and reference literature. The results of the calculations are taken into account when determining the volume and forms of field research, the depth of drilling and excavation of the rock, the schemes for placement of the mines, stationary sites, and so on.

A detailed forecast with quite exact quantitative estimates of the variations of the cryogenous conditions is made using the materials from field investigations and observations at the stationary field sites, and the final choice of the structural solutions for the pipeline and the method of performing the operations is made on the basis of its results.

The forecasting of the variation of the permafrost conditions during the construction, operation, and maintenance of the pipeline must not be passive. It is necessary to recommend measures providing for development of the cryogenic processes in the desired direction in order to ensure normal operation of the structures.

One of the basic goals of forecasting the variation of the permafrost conditions in the underground sections of the pipeline is forecasting the dynamics of the temperature field and establishing the boundaries of the zone of thermal effect of the pipe on the surrounding soil. The thermal effect of the pipeline is exhibited most significantly in the sections of transition from suspending above the ground to underground laying, from underwater to underground, from fill to ditch, and so on. Special attention is given to these sections during the forecast calculations.

The solution of the thermophysical properties during detailed forecasting of the variation of the permafrost conditions at the given time is found, as a rule, on analog and digital computers. The results of the calculations with respect to the forecast and the developed complex of measures with respect to controlling the cryogenous processes to insure stability of the structure must be presented in the technical plan with indications of the mandatory execution of the recommended measures.

METHODS OF LAYING THE PIPELINE

Pipelines on permafrost can be classified as a function of the type of interaction with the environment, the type of structural design, and the technological parameters.

As a result of the spottiness of the cryogenous conditions along the route and variations in relief, it is impossible, as a rule, to maintain a single or predominant method of laying along the entire length of the pipeline. It is obvious that, for pipelines more than 1,400 mm in diameter having great rigidity of the pipe, the traditional methods of laying are unsatisfactory and require modification. The high rigidity of the pipe faces the designers with the problem of selecting the method of laying, not so much from the condition of adaptation to the permafrost conditions as from the conditions of rigidity of the pipe and minimum admissible radii of curvature. As a result, in the same section it is possible for it to be necessary to use the underground method of laying on positive forms of the relief, and the suspended method on local depressions and basins.

The temperature regime of the gas in the pipe has a significant effect on selecting the structural solutions and the methods of laying the pipeline. By this attribute it is necessary to divide the sections of the gas line into "hot" and "cold." In the regions with permafrost, the "hot" pipelines include the sections in which the gas temperature in the pipe is higher than or equal to -1°C. Dividing sections must be considered "cold."

Depending on the position with respect to the ground surface, gas lines are divided into suspended, surface, and underground. At the present time, the following structural solutions for the pipelines and methods of preparing the bearing ground have become widespread: in the case of the underground procedure--rectilinear laying with slow bend sections on different types of supports--on piles, gravel pedestals, and so on; with the surface method, directly on the ground, on ground plates, along leveled ground, and on fills with thermal insulation under the pipe, and without it; for the underground procedure, laying the line within the layer of seasonal thawing or partially within the permafrost with preparation of the bearing ground, and without it, and also the semideep laying in rolling ground.

Usually in construction practice, the sections with different methods of laying alternate, depending on the permafrost conditions and the relief.

The underground laying of pipelines must be used in the "cold" sections of the gas line if the route passes through a territory with a relatively smooth relief and does not intersect a large number of streams. On the route of the underground sections of the gas lines, there must be a comparatively low groundwater level, both during the construction period and during operation and maintenance, and the soil must be sandy and supessy ground with low ice content and without inclusions of large ice lenses and wedges. During the construction of underground gas lines, it is necessary to consider the intensity of the thermal effect of these lines on the permafrost in the underground sections following the suspended sections or in the case of using equipment for air cooling of the gas and also the mechanical effect of the thawing ground on the pipeline and the possibility of erosion of the bearing ground of the line on disturbance of the natural regime of the suprapermafrost water.

It is not recommended that this laying procedure be used on subsiding and heaving ground, in very ice-rich soil, on silty soil that loses its bearing capacity on thawing, on unstable slopes, or in short sections of predominantly surface and suspended laying.

Surface laying is used in terrain with quiet relief without manifestations of thermokarst, heave mounds, landslides, and so on. Surface laying without embankments is recommended only in the "cold" sections of the route. Its application in the "hot" sections usually leads to the formation of thawing aureoles, subsidence, and slides; therefore, the straight sections of the line are recommended for laying on fill made of the soil or on the moss layer, and thermal insulation is put under the pipe. The temperature compensation sections of the surface line usually are raised on supports above the surface of the snow cover.

Laying on fill is used in sections made up of ground with adequate bearing capacity after thawing or in the "cold" sections if the ground loses bearing capacity during thawing. Laying on fill is frequently used as a transition section from aboveground to underground in order to lower the level of the stresses in the pipeline.

Aboveground support of the pipeline can be used for any gas temperature in the sections with any relief except sections on the floodplains of rivers that flood during the break-up. For this type of support, the pipeline is placed on piles, surface fill, and other supports. The type of supporting structure--roller, sliding, and so on--depends on the method of compensation of the longitudinal deformations of the pipeline arising as a result of variation of the air or gas temperature and also the inside pressure. Depending on the structural solution, the pipelines can be with compensation of longitudinal deformations or without it. The individual forms of structural solutions provide for partial or periodic compensation, for example, during the summer. In the case of aboveground support, the distance from the ground surface to the bottom of the pipe must be not less than 20 cm, and the radius of curvature in the vertical plane, 2500 D.

TECHNOLOGICAL CHARACTERISTICS OF GAS TRANSPORT

The mean annual gas temperature in the pipe is defined from the point of view of limiting the thawing aureoles of permafrost and the possibility of deformations. With a negative mean annual gas temperature, the maximum thawing aureole is equal to the aureole during the first year of operation of the pipeline. In subsequent years, the thawing aureole also will not exceed this magnitude if deformation of it and variation of the hydrogeologic regime of the zone of seasonal thawing or the thermophysical properties of the soil do not occur.

The sections of the lines located beyond the aboveground sections will have the ambient temperature in the winter, and in the summer, as a result of solar radiation, a temperature 5° to 10° above the ambient temperature. The gas temperature in the sections of the line beyond the compressor stations will depend on the operating regime of the station, the adopted cooling system, and the ambient temperature.

With an increase in the pipe diameter, the gas temperature in the pipe increases sharply, which changes the concept of the nature of the thermal interaction of the pipeline with the permafrost. When transporting large amounts of gas, the latter does not provide cooling in the section between the two adjacent compressor stations; therefore, on passage through a number of compressor stations, the gas temperature rises, and at the end of the line it can reach 60° to 90°. Accordingly, the necessity arises for regulating the thermal regime of the pipeline in order to decrease its effect on the permafrost and increase the carrying capacity. The reduction in release of heat by the pipeline is brought about by using aboveground and surface laying or cooling the gas after compression. A combination of both solutions is possible. The selection of the cooling procedure is determined by technicoeconomic calculations considering the poor frost conditions. The basic proportion of heat must be released from the pipeline as a result of using the aboveground method of laying in which the possibilities of natural cooling are used to the maximum.

Calculations have demonstrated that without artificial cooling of the gas during surface and underground laying, the depths of thawing reach 4 m and 10 m, respectively, after 3 yr of operation (with a gas temperature at the exit from the compressor station on the order of 60°C). The application of thermal insulation in the case of underground laying only retards the thawing of the permafrost, but 6 yr to 7 yr later the depth of thawing reaches 6 m to 7 m. For surface laying, the depth of thawing under analogous conditions is 3 m to 4 m. With this type of disturbance of the temperature regime of the frozen ground in the vicinity of the pipeline, it is possible to expect settlement of it to 1 m to 2 m. This can lead to collapse of the fills, erosion of the soil during the movement of the suprapermafrost water through the thawing aureole, and heave of the ground, even if it was not heaving ground before the beginning of operation. Hence, it follows that it is inadmissible to use underground and surface laying with a gas temperature above 30°-60°C.

At the present time, air-cooling devices are used to cool the gas. It has been established that only by combining air-cooling and thermal insulation under the pipe is it possible to insure conditions where the settlement of the ground will not be beyond the limits of admissible deformations of the pipeline, and the variations in the temperature regime of the ground around the pipe do not go beyond the limits of the natural fluctuations caused by fluctuations in the air temperature and are shifted only with respect to phase. Regulation of the gas temperature and the magnitude of the phase shift with respect to the air temperature is attained by selecting the material, thickness, and configuration of the thermal insulation.

However, when using the air-cooling equipment, it is necessary to install additional capacity at the compressor stations, inasmuch as the pipeline operates for a year under nonstationary thermal and hydraulic regimes. The seasonal fluctuations of the gas temperature at the exit from the booster station by 50° to 60° cause fluctuations in the output capacity by seasons reaching on the order of 20 percent. Another version of cooling is the use of refrigerators for deep cooling of the gas (to -30°C). This permits an increase in the carrying capacity of the gasline as a result of an increase in the compressibility factor of the gas at low temperatures to a greater degree when using the air-cooling equipment. The utilization of deep cooling of the gas permits the application of surface and underground methods of laying the line when building large-diameter gas pipelines. However, in this case, as a result of the formation of the frozen belt along the pipeline, development of heave processes, disturbance of the natural runoff of the suprapermafrost water, formation of ice bodies, and the formation of swamps are possible.

At the present time, the application of combined systems for cooling the gas by air-cooling apparatus and, in a number of cases, the refrigeration units is most reasonable.

SELECTION OF THE OPTIMAL ALTERNATIVE OF ROUTE OF A MAJOR GAS PIPELINE ON PERMAFROST

A. K. DERTSAKYAN, B. D. MAKUROV, AND R. E. FRIMAN   Giprospetsgaz Institute
I. YE. DUKHIN   BNIIIS Institute, YE. S. MEL'NIKOV   VSEGINGEL Institute

The selection of the optimal direction of a route is one of the primary problems of planning and designing major gas pipelines determining the national economic efficiency of the gas-transport systems.

As applied to the planning, design, and construction of systems of northern gas pipelines, the significance of selecting the route can be demonstrated by the following data. The length of the gas line from the deposits of Tyumen' oblast to the western Ukraine and the Baltic, where the end consumers of the gas are located, is more than 4,000 km, of which about 800 km are in permafrost regions. If the gas line is built from pipe 1,420 mm in diameter (which is now used), it is necessary to build a system made up of seven to eight loops. The reduction or lengthening of the route by only 1 percent is equivalent to reducing or lengthening a single pipeline by as much as 300 km, which expressed financially amounts to 60 to 80 million rubles.

This indicates the necessity for applying new methods of exploration and planning, permitting rapid determination of the optimal alternative of the pipeline route with minimum expenditures, especially in the permafrost regions, where, in addition to the economic factors, the necessity for ensuring high reliability of the structures under the most complex natural climatic and engineering-geocryologic conditions has primary significance.

As is known, the best route is determined from the following factors: (a) minimum length; (b) most favorable ground-hydrogeologic conditions; (c) advantageously distinguished hydrologic conditions of the intersected water obstacles; (d) least number of intersected artificial structures; (e) minimum removal of cultivated ground; (f) the presence of communications along the route; and (g) acceptable distance from industrial sites and populated areas, and so on.

TABLE 1 Recommended Methods of Laying a Gas Line in Permafrost Regions

Type of Section with Respect to Complexity of the Geocryologic Conditions	Terrain Classification Groups	Characteristic of the Soil with Respect to Settlement	Recommended Methods of Laying the Gas Line		
			Hot	Warm	Cold
Normal	*Group IV* River terraces, forested floodplain	*Category I* Nonsubsiding during thawing giving insignificant uniform settlement	Underground and surface	Underground and surface	Underground and surface
	Group III Swamps in interfluves, suprafloodplain terraces		Underwater with load and in-fill	Underwater with load and in-fill	Underwater with load and in-fill
Complex	*Group II* Tundra and forested tundra drained sections	*Category II* Weakly subsiding, giving uniform settlement to 10% of the thickness of the thawing layer	Aboveground on piles and ground supports	Aboveground and surface on a thermally insulated pad	Surface on a leveling fill without embankment and underground
Especially complex	*Group I* Hummocky and flat peat-bogs, solifluction slopes	*Category III* Ice-saturated, giving nonuniform settling of 10-40% of the thickness of the thawing layer	Aboveground on piles	Aboveground on piles and surface supports	Aboveground with thermally insulated pad and embankment
		Category IV Contain group inclusions of underground ice; thermokarstic cave-ins formed as a result of thawing			

The best pipeline route in the sections with permafrost is determined by a number of other significant factors in addition to those listed. In particular, it is necessary to know and consider the extent and thickness of the permafrost; the thickness of the active layer; the temperature conditions, composition, structure, and ice content of the soil; the thermophysical properties of the soil; the tendency of the soil toward heaving and settlement; and also the possibility of the occurrence of solifluction processes, frost-fissuring, ground-icing, and so on.

Even a simple list of the elements required for correct evaluation and forecasting of the conditions of laying the pipeline and the maintenance regime in the investigated sections indicates the complexity of this problem.

A detailed study of the set of parameters that would permit us to determine the presence and intensity of the majority of permafrost phenomena and processes during one-time short-term studies under the conditions of a route of significant extent is in practice impossible, considering that it is necessary to begin with the most unfavorable combinations of climatic conditions and variations that take place along the route during construction as a result of cutting down forests, removing moss, and other disturbances of the surface.

The basis for the engineering-geologic research from the technical design of the gas pipeline must be the method of landscape districting of the alternative routes and the determination of the engineering-geologic and permafrost conditions of the isolated landscape subdivisions with careful prefield development of materials on the construction zone. Usually the landscape districting map is compiled on a scale no smaller than 1:100,000. Aerial photographs on a scale of not less than 1:60,000 are used as the basis for the studies.

Each landscape subdivision must be a strip not less than 4 km wide (at 2 km to the right and left of the center line). Here, information must be obtained at the most characteristic points of the subdivision as a result of drilling, geophysical work, and landscape examination.

Forecasting the temperature regime of the soil during the operation and maintenance process is carried out for the most characteristic conditions of each engineering-geological subdivision in the zone of the thermal effect of the planned pipeline, with various methods of ensuring the stability of the lines.

The volume of research required for providing a detailed basis for one of the methods of laying the gas line--underground, on ground, or aboveground--differs. Therefore, for a scientifically based approach to selecting the optimal route in the sections that are characteristic with respect to landscape attributes, depending on the permafrost conditions and the forecast of their fluctuation, a table is prepared, the data in which provide a basis for preliminary determination of the volume of engineering studies in individual sections, taking into account the proposed method of laying the pipeline.

In addition, the consideration of an entire set of conditions affecting the selection of the optimal route, considering the design of the pipeline, can be carried out only on the basis of the broad application of mathematical methods, computers, and aerial photographic methods.

The content and volume of engineering-geological studies varies, depending on the stage of planning and design and the actual possibilities of obtaining information during the given periods, and when using a digital computer they differ significantly by comparison with the ordinary method of selecting the optimal route.[1]

A detailed study of cartographic and other materials does not exclude aerial visual and surface examinations of the sections of the routing. The data obtained are given in the form of the basic parameters in the mathematical model, and they are taken into account during calculation on a digital computer, along with the total of all other possible indexes with respect to the given section.

The broad possibilities that are opened up when using modern methods of research with the application of digital computers and aerial photography have been convincingly proved in the example of the studies and planning and design of the routes for the northern gas lines.

In the region where the permafrost is extensive, the sections with evidence of active cryogenous processes and phenomena are unfavorable; the favorable sections are those with thawed, folded rocky, semirocky, gravel-pebbles, and rubbly ground--sections with the least total moisture (iciness) of the rock and with minimum depth of seasonal thawing.

Beginning with the method of creating the digital model of the terrain, it appears possible to make a differentiated evaluation of the intersected territory with respect to engineering-geocryologic attributes. In the case of computerized optimization of the routes, this will make it possible to discover competitive routes. Moreover, the designation (with the implication Table 1 presented) of the methods of laying the gas lines in the individual sections will take into account the characteristic features of the temperature regime of the structure both during construction and during operation and maintenance.

Thus, the application of mathematical methods and digital computers ensures the possibility of an all-around evaluation of both the natural and structural factors connected with methods of construction, material expenditures, construction times, and so on. Here, maximum reduction of subjectivity in evaluating the decisions with respect to selecting the route, that is, in determining the actively optimal alternative is achieved. The most effectively indicated optimization of the route can be brought about using the set of construction engineering maps taking into account all of the primary features of the region of interest.[2]

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ENGINEERING-GEOCRYOLOGICAL SURVEY OF MAJOR PIPELINE ROUTES

M. A. MINKIN Fundamentproyekt Institute

The construction of large-diameter pipelines in the areas where permafrost is widespread is at this time imposing increased requirements on the quality and times of executing the exploration plans. A special role is being played by the problems of making the engineering-geocryologic survey, since the selection of the optimal planning solutions and, consequently, the cost and the time of constructing the pipeline and its reliability and operation can be based on a comprehensive and detailed study of the permafrost conditions along the route, taking into account their predicted variations.

In this paper a procedure is proposed for an all-around permafrost survey developed on the basis of many years of experience of the Fundamentproyekt [Foundation Planning] Institute and other organizations with respect to pipeline routings in the northern part of western Siberia.

The engineering-geocryologic survey, according to V. A. Kudryavtsev,[1] consists of field, laboratory, and camera work: field work along the route; laboratory determinations of the physicomechanical and thermophysical properties of the soil; experimental and routine work; experimental studies and continued observations on experimental sections of pipelines, which have been constructed and are being constructed; and calculations using a digital computer.

The scale of the permafrost survey, and the content and volume of the engineering-geological studies depend on the complexity of the permafrost conditions along the route, and they are determined by the problems that are posed in the stages of the planning and exploration, given below.

During the process of planning the pipeline, it is possible to distinguish three basic successive levels of development: the selection of the optimal route, the preliminary design (the technical design stage), and final design (the working-drawing stage).

The selection of the optimal pipeline centerline is considered as the primary problem. In order to optimize the selection, a number of mathematical techniques on a computer have been developed and are being used.[2-4]

In order to obtain the basic information for the first stage about the permafrost conditions along the route, it is necessary to do an area permafrost survey within the proposed zone for the pipeline. The scale of this survey is determined from the conditions of obtaining the information sufficient to select the optimal route, but with minimum amount of detail.

As has been demonstrated by Babin *et al.*,[2] a scale of 1:100,000 is sufficient for this purpose. With a further increase in detail, the divergence within the alternatives does not exceed 1 to 2 percent. The methods of such surveys with broad utilization of express methods have been developed by the geocryological department of Moscow State University and VSEGINGEO [All-Union Scientific Research Institute of Hydrogeology and Engineering Geology], and therefore they are not considered in this report. The next steps in planning and designing the pipeline are the selection of the principle of use of the ground as the bearing surface and determination of the laying procedures and types of supports. On the basis of the solutions in the technical plan, the volumes and costs of constructing the pipeline, the requirements for equipment and details, and so on must be estimated.

Experience shows that it is possible to do this only on the basis of the data from a detailed permafrost engineering-geologic survey and not as a result of medium-scale investigation. The necessity for a detailed survey in this stage of the planning and exploration arises, on the one hand, from the complexity, variety, and nonuniformity of the permafrost conditions along the route, and on the other hand, the increased requirements on the pipelines as engineering structures, their high stiffness, and increased sensitivity to nonuniform settlement.

Let us consider some procedural problems of a detailed permafrost survey.

The permafrost engineering-geologic studies of the optimal route begin with a detailed cryologic-geographic survey in a corridor 200 to 500 m wide (in mountainous areas, to 1-2 km wide). The purpose of these studies is the isolation of sections with uniform natural conditions within the limits of the strip, outlining anomalous sections with intense development of the physicogeological cryologic phenomena and division of the route into microregions on the basis of this study.

The cryologic-geographic survey is taken by a field study and point description of the terrain. The number of observation points is determined by the necessity for isolating uniform microregions on the scale of the survey. Considering this, at each point of the description, the variations in natural conditions are fixed, depending on their complexity and variability within the radius of 10 m to 50 m. Hence, the minimum distance between the observation points is usually from 20 m to 100 m.

Simultaneously with making a cryologic-geographic survey, a symmetric electric profile is made with an interval of 5-10 m along the corridor and with respect to characteristic profiles. The complex application of them permits isolation of uniform microregions, not only on the basis of the variations in surface conditions (the nature of vegetation, the interchange of geomorphological elements, surface drainage, peculiarities of the microclimate, and so on), but also with respect to changes in the nature of the geoelectric field.

By using the data from the set of detailed investigations, the interrelations of the permafrost conditions with the individual factors and groups of factors of the natural situation are evaluated quantitatively, the parameters of variability of all the basic engineering-geologic characteristics are determined, the uniform elements with respect to groups of indexes of the soil properties are isolated statistically, and, on the basis of this, engineering-geocryologic division of the routes into districts can be carried out.

In order to increase the level of the information obtained during a detailed permafrost survey, to simplify and accelerate the processing and classification of the research data, digital recording of the information on forms for the M-20 digital computer is used both for quantitative parameters and for the qualitative characteristics of the permafrost conditions.

The proposed coding procedure is advantageously distinguished from the edge-punched cards that have become widespread at the present time[5,6] by a higher degree of automation and the possibility of direct utilization of the existing standard programs for statistical processing and thermophysical calculations.

Simultaneously with exploration along the route for the pipeline in the technical design stage, continued observations are made of the temperature-moisture regime of the soil of the most typical microregions, taking into account the principal variations of the natural situation during construction of the route. The observations must be made for not less than 1 yr.

Predicting the variation of the permafrost conditions is an important component of both the medium-scale and detailed permafrost survey. A diagram of the problems and methods of predicting the variation of the permafrost conditions in various stages of the research, planning, design, and construction of the pipeline is presented in Table 1.

In order to determine the interrelations of the permafrost conditions with the factors of the geologic-geographic medium, the quantitative evaluation of the variations in the temperature regime of the ground, the trend of the permafrost processes, different procedures of laying the lines, and types of supports, the Fundamentproyekt Institute has developed and is using programs for calculations on the BESM-6 and M-20 computers.

The programs permit calculation of the non-steady-state one-dimensional and two-dimensional

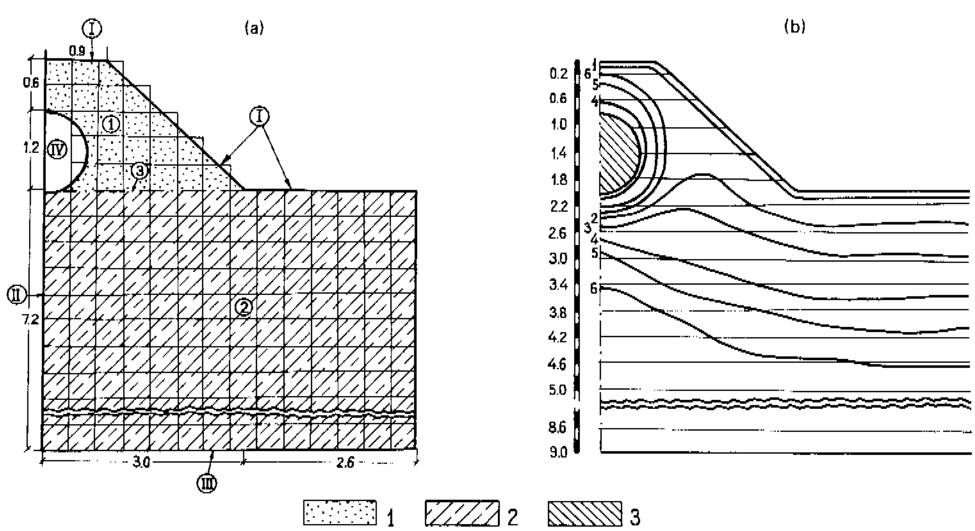
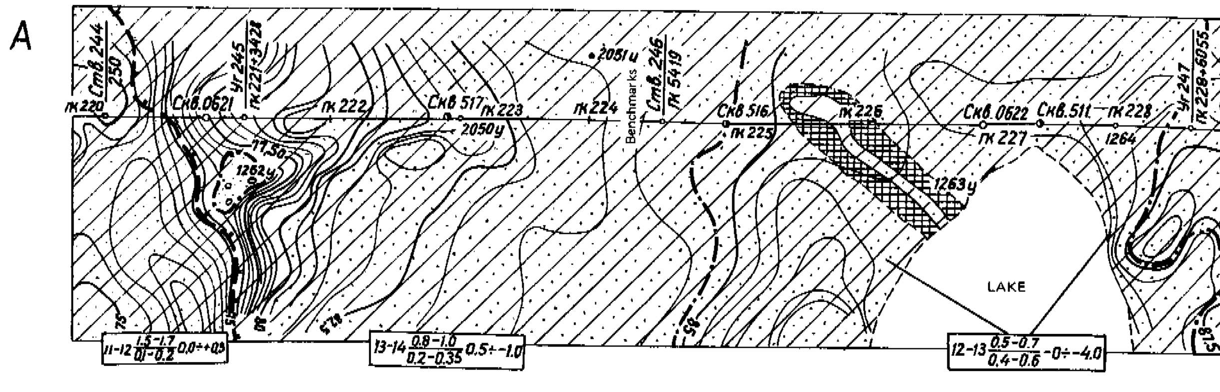


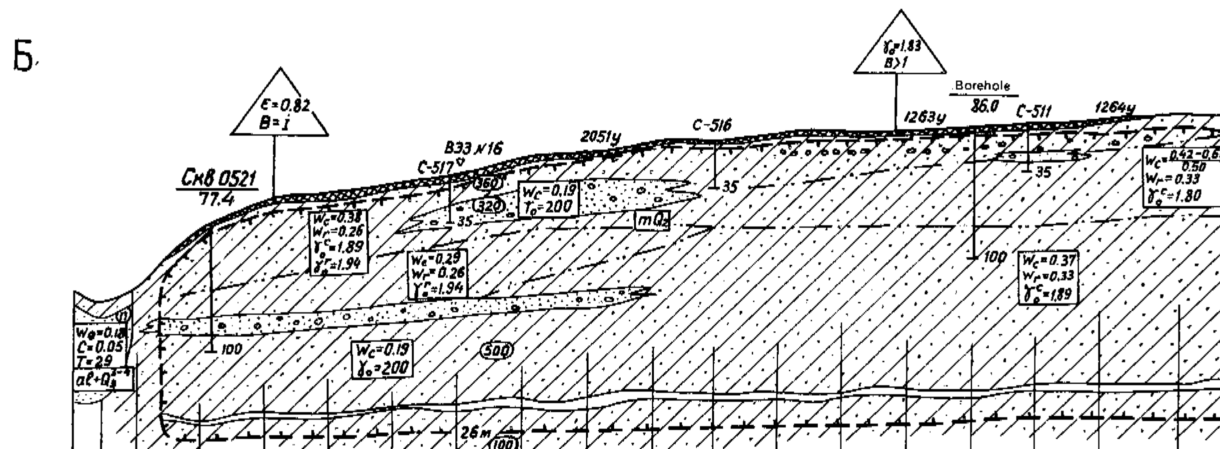
FIGURE 1 Calculating the temperature fields on the BESM-6 digital computer around a pipeline laid in an embankment. (a) Scheme of the calculation region: I, II, III, IV--numbers of the boundary condition zones; 1--fine sand; 2--suglinok; 3--pipe. (b) Position of the 0°C isotherm after a time τ following the beginning of operation of the pipeline; 1--τ = 720 h; 2--τ = 1,440 h; 3--τ = 2,160 h; 4--τ = 3,600 h; 5--τ = 5,760 h; 6--τ = 14,400 h.

TABLE 1

Planning and Design or Construction Stage	Content of the Stage	Stages of the Engineering-Geocryologic Research	Scale of the Engineering-Geocryologic Survey	Prediction Problems	Prediction Techniques
Technical plan (Preliminary design)	1. Selection of the route	Prefield period	According to the small-scale survey data	(a) Definition of the region of investigation (b) Qualitative evaluation of the trend of engineering-geocryologic processes	Calculations by approximate formulas
	2. Selection of the laying procedure	Studies to select the optimal	1:50,000 to 1:100,000	(a) Determination of the interrelations of the permafrost conditions and the factors of the geologic-geographic survey for all types of microregions	(a) Calculations by approximate formulas
	3. Selection of of the principle of utilizing the soil for bearing			(b) Quantitative evaluation of the trend of engineering-geocryologic processes	(b) Calculations on analog and digital computers
	4. Selection of of types of support	Exploration of the optimal route	1:2000 to 1:10000	(a) Determining the variations in the engineering geocryologic conditions for all types of microregions with different methods of laying and types of supports (b) Determination of unfavorable sections for construction	(a) Calculations by formulas (b) Calculations by analog and digital computers (c) Continued observations

Working drawings (Final design)	1. More precise definition and detailed outline of the design solutions adopted in the technical plan	Additional studies and individual complex sections in engineering-geocryologic respects	1:500 to 1:2000	(a) Quantitative evaluation of the variations of the engineering-geocryologic conditions of the route for the adopted methods of laying and types of supports	(a) Calculations on digital and analog computers
	2. Development of measures with respect to ensuring reliability and stability of the pipeline operation	Experimental and regime operations		(b) Quantitative evaluation of various measures with respect to controlling the permafrost conditions	(b) Experimental and routine operations in typical sections
Experimental construction	1. Development of the processes and methods of construction	Observations in experimental sections	1:500 to 1:2000	Compiling the final forecast of variations of the permafrost conditions during construction and maintenance of the pipeline	(a) Calculations on digital and analog computers
	2. Checking the correctness of the technical solutions				(b) Observations on experimental sections





Elevation stakes (absolute elevations)	72.4; 71.2; 72.6; 77.4; 78.8; 79.5; 79.8; 80.5; 81.5; 83.0; 83.3; 84.4; 84.4; 85.2; 85.3; 85.6; 86.0; 86.1; 86.5; 87.2
Spacing of stakes	25; 26; 50; 34; 16; 50; 50; ; 50; 50; 55; 45; 50; 50; 50; 50; 55; 50; 60; 40 (stakes 220, 1, 2, 3, 4, 22.5, 6, 7, 8)
Engineering-geological and frozen-ground conditions	Silty sand — Suglinok with occasional gravel L_V^* to 5 m = 0.17, L_V 5-10 m = 0.05. In the 2.0-3.6 and 7.1-8.7 range the sand is fine with gravel inclusions up to 30%. Cryotexture of the sand—massive. — Peat 0.5 m below the suglinok in the 1.8-2 m range with inclusions of gravel and pebbles to 40%. L_V to 2 m = 0.17; L_V 2.0-8.0 = 0.2-0.4; L_V 8.0-10.0 m = 0.06.
"Standard" and computed engineering-geological and frozen-ground characteristics	*L_V is the ratio of $\frac{\text{segregated ice}}{\text{total ice (pore ice + segregated ice)}}$ — Resistance to normal pressure of the bearing soil R^N = 0.5 kg/cm².
Recommended types and properties of supports	On surface and semi-buried laying of the pipe in embankment. Thickness of soil above the pipe not less than 0.8 m.
Scale ratio with the soil (see Table 1)	300; 3000; 500; 2000; 500; 200; 2000; >10000; 500; 1500; 500

PROVISIONAL

I Stratigraphy, lithography, and genesis

- al-Q_4 floodplain deposits
- h-lh$_1$-Q_{3-4} swamp and lacustrine-swamp deposits
- m-Q_5 marine deposits and marine terraces (Kazantsevskaya suite)
- m-Q_2 marine deposits and marine terraces (Salekhardskaya suite)

peat

clay

suglinok

silty suglinok

sandy suglinok

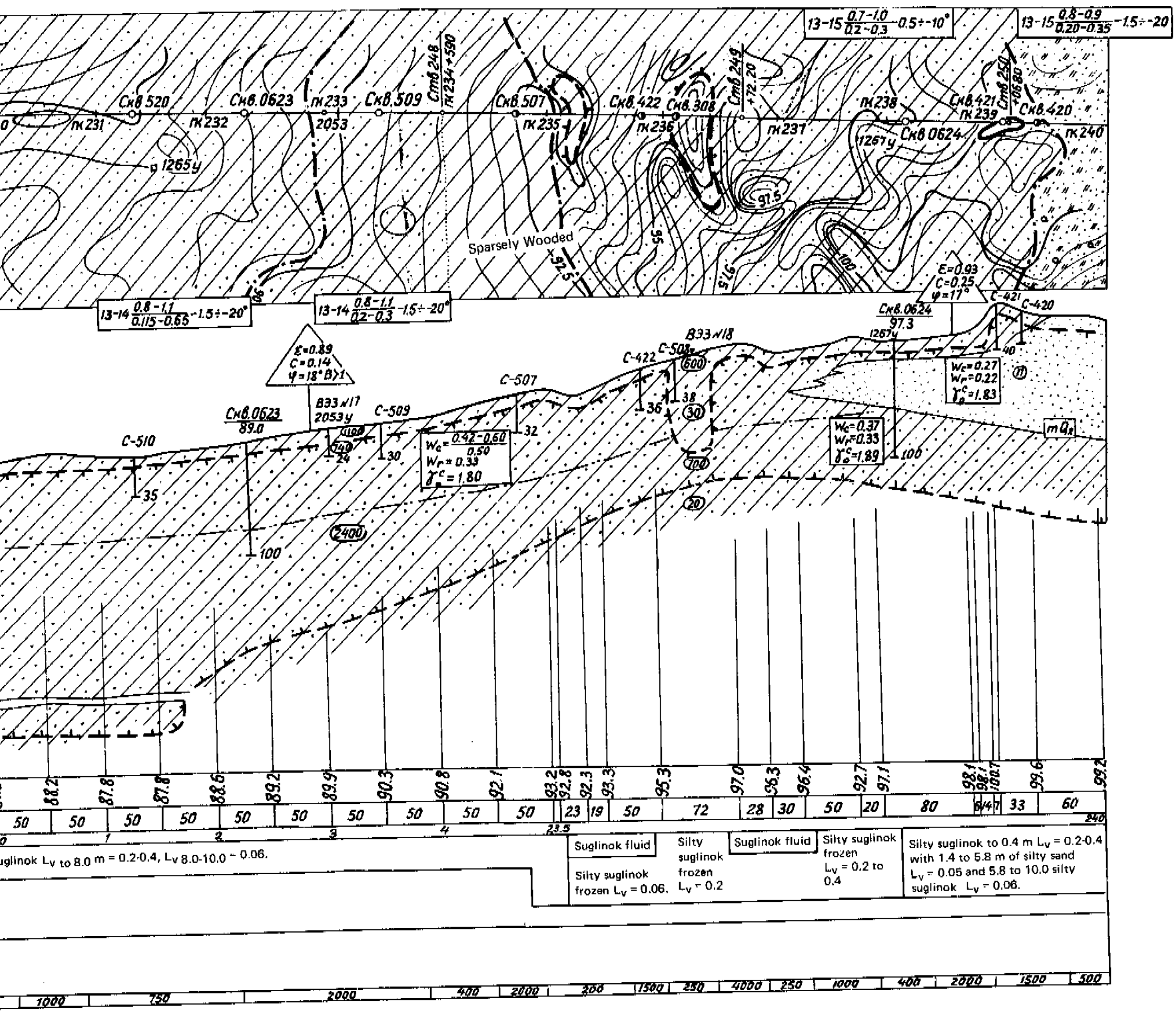
sand: m fine; p silty; sr medium coarse

II Engineering–geocryological characteristics of the soils

10-12 $\frac{0.9 \text{ to } 1.2\ (a)}{0.25 \text{ to } 0.40\ (b)}$ $\frac{-0.5 \text{ to } -1.0}{-0.3 \text{ to } -0.2}$ Before the fraction: amplitude of the temperature wave at the ground surface, °C;

Numerator: maximum depth of seasonal thawing (freezing), m;
Denominator: total water content of the layer of seasonal thawing (freezing) [as a fraction];
After the fraction: mean annual temperature of the ground, °C.
[(a) and (b) are not defined]

- w_c total moisture content of the frozen ground as a fraction
- w_e moisture content of the mineral part of the frozen ground as a fraction
- w_2 natural moisture content of the thawed ground as a fraction
- L_V ice content of the ground as a result of ice inclusions as a fraction
- B consistency of the thawed ground
- ε void ratio
- c norm* value of cohesion, kg/cm²
- φ internal angle of friction, deg.



NOTATION

γ_2 unit weight of the thawed ground, g/cm^4

γ_{d1} dry unit weight of the frozen ground, g/cm^3

γ_1 unit weight of the frozen ground, g/cm^3

R^N norm* value of the resistance of the ground to normal pressure, kg/cm^2

*'norm' refers to "SNiP'S"—"Construction Norms and Guidelines." SNiP:11-B.6-66. M., 1967. FSS.

III Boundaries

—·—·— boundary of the frozen microregions

—··—·· boundary of the ground with different moisture/ice content

stratigraphic

—┬—┬— between the permafrost and thawed ground (stroke in the direction of freezing)

IV Other provisional notations

$\frac{\text{BH \#25}}{30.42}$ engineering-geological borehole, number in the numerator; in the denominator is the benchmark level of the top of the borehole

C-282 logging hole and its number

▼ point of vertical electric logging

(300) apparent ground resistance, ohms/m

$\frac{▼50}{6\cdot x\text{-}69r.}$ In the numerator--ground level. In the denominator--date of measurement

△ in the triangle is the ground norm for the seasonal freezing layer

□ the ground normals below the seasonal freezing (thawing) layer given in the rectangle temperature fields in the freezing-thawing ground under variable boundary conditions with a nonuniform geologic section and arbitrary shape of the region of investigation.

The comparison made between the solutions obtained and the natural observation data, the calculations on the hydrointegrator, and the exact analytical solutions demonstrated that, when ensuring stability of the different circuits used, quick convergence of the results is achieved (deviation on the average of 2 to 4 percent).

In Figure 1 we have an example of calculating the depth of freezing of thawed ground around the pipeline laid in an embankment. The boundary conditions of the problem are presented in Table 2.

In working out the final design, along with the calculation techniques, a special role is played by the experimental and regime operations. The types and methods of this research differ as a function of the specific permafrost conditions, the adopted principle of using the ground as a bearing surface, the method of laying the pipeline, and the type of pipeline supports. For example, in the case of the aboveground method of laying the pipeline with pile supports, a study must be made of the expected temperature-moisture regime of the soil at the pipeline supports. The bearing capacity of the piles for a horizontal load is determined, and for plastically frozen ground, the vertical load also. The tangential heave forces and the vertical displacements of the supports are also studied.

In the case of surface, semiburied, and buried methods of laying the line, the primary problems are determining the thermal interaction of the line with the ground of the bearing surface, determination of the nonuniformity of the settlement of the ground along the length of the line with time, and the reaction pressure on the pipe.

According to the proposed procedure, in 1969-1971, permafrost studies were made on a section of the route from the North to the Center, a line more than 600 km long.

The route in the investigated section intersects three large regions (the western Siberian lowland, the polar Urals, and the northeastern European plains), differing significantly with respect to geologic structure, tectonics, and physical geography, and the laws of cryogenesis characterized by significant variety and complexity of the permafrost conditions.

The engineering-geocryological division of the routing into districts, which was performed on the basis of the data from the cryologic survey on a 1:2000 scale, made it possible to isolate 203 microregions within the investigated section, in each of which the permafrost conditions (the set including the defined temperature regime, the lithologic state of the ground properties, hydrogeological conditions, and the intensity of the physical-geological processes and phenomena, and so on) are uniform and persistent, and they differ from the permafrost conditions of the other microregions at least with respect to one of the listed factors. The extent of the microregions differs: from several tens of meters to several kilometers.

As a result of discovering general and partial laws of the formation of the engineering-geocryological conditions, the quantitative evaluation of the trend of the permafrost processes, and statistical processing of the physical-mechanical data and thermophysical properties of the soil, the isolated microregions were combined into two groups with general permafrost characteristics.

Each such group of microregions represents a defined type of permafrost conditions characterized by generality or similarity of the design parameters that essentially determine the choice of the basic planning solutions for the gas line (the principle of use of the ground as the bearing surface, the method of laying the line, the type of support).

The parameters for the sections of the route made up of series of Quaternary soil deposits (the western Siberian lowland, the northeastern European plain) are the following: the nature of propagation and the condition of the frozen and thawed ground, the mean annual temperatures and lithologic composition, the ice content (for frozen ground), the consistency and degree of saturation (for thawed ground), and the intensity of occurrence of cryogenic processes and phenomena.

For the sections of the route with shallow bedrock (the polar Urals), the primary factors are the depth of the kinds of the bedrock cover, the composition and properties of the large rubbly deposits covering it, and the presence of slope processes and phenomena.

In all, within the limits of the investigated part of the route, 23 types of permafrost condi-

TABLE 2

No. of the Boundary Condition Zone	Type of Boundary Condition	Months											
		I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
I	1	-25	-25	-20	-10	-1	+10	+17	+13	+6	-5	-18	-24
II	2	Insulation condition: $q = 0$											
III	1	t_N = const = 0.5°C											
IV	1		-5				+10				-5		

tions were isolated, as a function of which the designers developed the primary planning solutions. They adopted planning solutions that are tied to the specific permafrost conditions of the microregions by cryologic maps on a 1:2000 scale compiled for each 2 km of routing and the geocryological sections on the same scale (Figure 2).

The research performed permitted the designers to obtain the necessary data to develop the preliminary design and the final design of the gas pipeline in the planned time.

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THAWING AND THE DIFFERENTIAL SETTLEMENT OF THE GROUND AROUND OIL WELLS IN PERMAFROST

A. PALMER Cambridge University, England

The discovered oil reserves in Alaska and in the Canadian Arctic have forced attention to the problem of drilling operating wells in permafrost. In the paper by Koch[1] a survey of this problem is presented. In the northern part of Alaska, the thickness of the permafrost reaches about 600 m. The oil flowing through the well has a temperature from 60°C to 85°C. In spite of the special measures for cooling or insulation of the well casing, with time the ground around the well thaws; the radius of the thawing zone is about 15 m in 10 yr. If the thawing ground cannot withstand the load of the higher-lying layers acting on it, its consolidation will take place, and the corresponding settlement can cause significant surface shifts. However, the settlement of the ground in the vertical direction has a greater significance when the shear stress acts on the casing downward and causes compressive stresses which can be quite high and deform it. Inasmuch as the insulation and cooling of the pipe is extraordinarily expensive, and thawing of the ground is not completely excluded, it is important to establish the magnitude of the settlement and the axial forces caused by it in the case of an uninsulated well.

In the report a study was made of the factors that determine the settlement of the ground and the stress in the casing. In order to establish the basic and secondary factors, certain assumptions were made: For example, the temperature variation of the permafrost with depth was taken into account. First, there was a brief study of the problem of heat transfer; then the simple models of consolidation and deformation of the ground layer with thawing and the use of these models to study the thawing of the ground around the well were investigated.

HEAT TRANSFER

The distinguishing feature of the geometry of the system is its great length in comparison with its radius. This system is axisymmetric. The investigation of the thermal problem in the given case is necessary only to determine the thawing radius, and it will be executed in a general way. If the initial temperature, ice content, and the thermophysical properties of the permafrost layer are constant and when the oil goes into the well the temperature varies insignificantly, then the heat flux around the wells will be two-dimensional and axisymmetric, and the radial temperature distribution θ will be subject to the following equation:

$$\frac{1}{\chi} \cdot \frac{\partial \theta}{\partial t} = \frac{\partial^2 \theta}{\partial r^2} + \frac{1}{r} \cdot \frac{\partial \theta}{\partial r},$$

where r is the radius, t is time, and χ is the thermal diffusivity under the condition that convection can be neglected.

On the surface of the casing when $r = a$, the temperature θ_1 is assumed equal to the oil temperature. The initial temperature of the permafrost is θ_0. The radius of the thawing zone increases with time, and by the time t it is $R(t)$; the temperature at the thawing boundary is θ_F. If the energy required for thawing a unit volume of ground is λ, then the heat balance at the thawing front will be the following:

$$-\lambda \frac{dR}{dt} = \left[k \frac{\partial \theta}{\partial r}\right]_{r=R},$$

where k is the coefficient of thermal conductivity, and the right-hand side of the equation is the heat flux to the boundary $r = R$. If $\chi \to \infty$, a simple approximate solution is obtained, which is applied to the ground with high ice content. From the physical point of view, this means that in the temperature distribution the role of the latent heat--that is, the energy required to thaw the ground, is much greater than the heat capacity, which is the heat required to heat the ground from θ_0 to θ_F before thawing and from θ_F to θ_1 after thawing. At any point in time the temperature distribution corresponds to the quasistationary thermal state for the following radius R. It is easily possible to demonstrate that the thawing radius R and the time t from the beginning of this process are related by the equation

$$t = \frac{a^2 \lambda}{k(\theta_1 - \theta_F)} \cdot \left\{1/2 (R/a)^2 \left[\ln (R/a) - \frac{1}{2}\right] + \frac{1}{4}\right\}.$$

If we take $\theta_1 - \theta_F = 80°C$, $a = 0.2$ m, $\lambda = 110$ MJ/m^3, and $k = 2$ W/m·C (which corresponds to the values for the silty ground with an ice content of 20 percent), then according to Equation (3) the thawing radius will be 15 m after 9.4 yr. This approximate solution agrees with the numerical calculations of Koch.[1] Failure to consider the specific heat capacity leads to a somewhat increased thawing radius; this can be corrected using the parallel Pekeris and Slichter method.[2]

MODEL OF THE GROUND CONSOLIDATION PROCESS DURING THAWING

Where under natural conditions the permafrost is completely consolidated analogously to water saturation of the thawed ground there will be equilibrium with the groundwater level at the surface. Then the ice contained between the soil particles transfers the hydrostatic stress equal to the ice density multiplied by the depth, and the load remaining from the higher-lying layers is completely transferred to the skeleton of the soil, that is, it is the effective stress. Inasmuch as in this case the ice is not subject to shear, it does not exhibit a tendency toward creep and redistribution of the total stress. If the layer of this soil thaws with a constant void ratio, it does not exhibit a tendency toward deformation, since the effective stress remains identical everywhere. However, usually the soil does not reach a state of complete consolidation, and during thawing, under the effect of the initial pressure of the higher-lying layer, it has a tendency toward deformation. The effective stresses used for the initial value of the porosity factor are insufficiently large to withstand the applied load, and quickly after thawing pore pressures occur that take up part of the total stress. When the pore pressure is dispersed, the void ratio decreases and the effective stresses increase.

In order to define these ideas more precisely, let us consider the uniform permafrost sample that is subjected to load in an odometer. In Figure 1 we have the graph of the void ratio as a function of the vertical pressure; here, it is proposed that the consolidation occurs in the instance of lateral expansion. Initially the void ratio is e_1, and the total stress is q; this state can be denoted on the graph by the point 1. The curve ac expresses the dependence of the void ratio on the effective stress in the generally-accepted consolidation test. If the soil subsides quickly without variation of the void ratio, the effective stress is p, and the pore pressure equal to $q - p$ arises. This state of the soil is denoted on the graph by point 2. The disperson rate of the pore pressures is controlled by the consolidation factors; on the graph the points expressing the state of the soil are displaced along the ac line from point 2 to point 4, and the void ratio drops from e_1 to e_4. If during thawing the ground remains saturated and the void ratio decreases proportionally to the ice and water density, the state of the soil immediately after thawing is depicted on the graph by point 3. In each case the specific state is identical, and the deformation

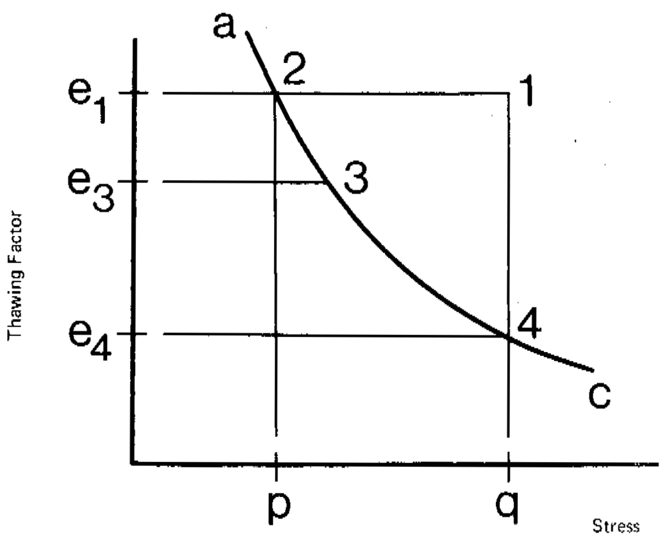

FIGURE 1 Settlement during thawing period. 1--initial state; 2--state after thawing without variation of the volume; 3--the same if the soil remains water saturated; 4--state after consolidation; *q*--total stress; *p*--effective stress in state two.

during thawing is equal to $(e_1 - e_4)$ $(1 + e_1)$. The curve *ac* can be obtained in the experiment with respect to thawing the soil sample and used to determine the pore pressures occurring during the loading process. The consolidation theories for models of this type were analyzed by Tsytovich,[3] Zaretskiy,[4] Morgenstern and Nixon,[5] and others, but a detailed comparison of the experiments was not published.

CONSOLIDATION DURING THAWING AROUND A BOREHOLE

Let us consider the application of this simple model to the two extreme cases of settlement in the thawing cylinder of ground. In one case the thawing ground is considered as an isolated cylinder of soil that does not have a support on the side of the permafrost surrounding it, but under the effect of the overburden it is compressed in the vertical direction; the effective stress increases in this case. In the other extreme case (where the thawing round contacts the permafrost), the deformation transmits the entire weight of the thawing ground to the surrounding permafrost, which cannot be compensated for by the initial vertical effective stress. None of these extreme cases reflects the actual conditions, but after considering them individually, it becomes clear that deformation determines the behavior of the central part of the thawing cylinder of ground, and consolidation determines its behavior in the edge section.

First, let us assume that the thawing zone has no support on the side of the surrounding permafrost, and the shears transmitted from it to the casing have a negligible effect on the stress in thr ground. If the ring of ground falls quickly, the complete stress aimed perpendicular to the horizontal surface is equal to the specific weight of the ground with respect to depth. The effective stress is the stress that exists in the frozen ground; the rest of the total stress is the pore-water pressure; in turn, part of the latter is hydrostatic pressure (water density multiplied by depth), and the rest is the excess pore-water pressure.

In Figure 2 there is a schematic representation of the stresses. When the soil is packed, the thawing ring compresses, and, if the displacement in the base of this ring is negligibly small, then the displacement at a depth *z* is expressed as follows:

$$\int_z^L \varepsilon_T(z)\,dz,$$

where $\varepsilon_T(z)$ is the consolidation deformation under the effective stress

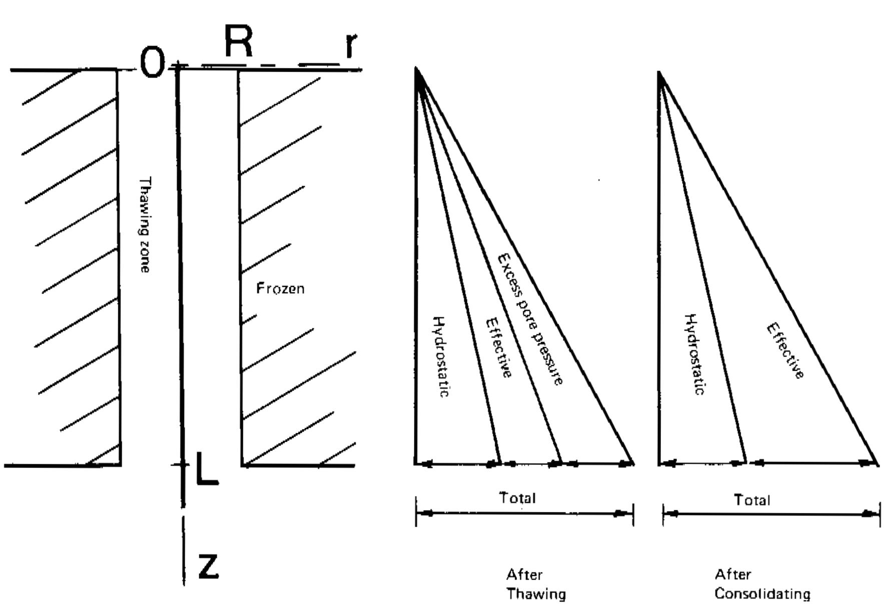

FIGURE 2 Stress distribution in the thawing zone. *r*--radius from the casing access; *R*--radius of the thawing zone; *z*--depth of the surface; *L*--depth to the permafrost footing.

$$\int_0^z [\gamma(z) - \gamma_w]dz,$$

where $\gamma(z)$ is the total specific weight of the soil at a depth z, and γ_w is the specific weight of the water. If the casing follows this displacement of the soil, then its axial deformation will be ε_T. If the soil is displaced downward relative to the casing, then the shear force aimed downward along the casing is $\mu k_0 (\gamma - \gamma_w) z$, where μ is the coefficient friction and k_0 is the ratio of the vertical stress to the horizontal stress; consequently, the axial stress in the casing τ at a depth of z is

$$\sigma_c = \frac{1}{t} \int_0^z \mu k_0 (\gamma - \gamma_w) z dz.$$

If $\mu = 0.2$, $k_0 = 0.6$, $\gamma - \gamma_w = 8 kN/m^3$ and $t =$ 20 mm, then at a depth of 100 m this stress reaches a value of 240 N/mm^2, and a great depth becomes inadmissible.

Thus, large axial stresses develop in the casing in the case where the soil has no deflection. The time during which these stresses develop is determined by the rate of dissipation of the pore pressure, and, if the thawing ground is water impervious, then it can be vary large. Inasmuch as the ground around the thawing ring is frozen, the water cannot percolate in the horizontal direction and will move up or down. The runoff path has a magnitude on the order of $L/2$. The theory of consolidation[6] shows that the time during which the settlement reaches half of its limiting value is approximately $0.2\ (L/2)^2\ c_{VC}$, where c_{VC} is the Terzaghi consolidation factor; its exact numerical value depends on the initial distribution of the pore pressure. If $L = 600$ m and c_{VC} is $2 \times 10^{-8}\ m^2/s$ (the standard value for fine silty soil), then this time is on the order of 3,000 yr. Consequently, in the case of silty or still-finer soils the settlement will develop too slowly to have a harmful effect on the borehole during its existence. For nonuniform ground it is determined by the thickness and permeability of the least permeable layers.

The presented calculation demonstrates the significance of the pore pressures and the movement of the water inside the ring of thawing ground. Harmful effects can arise if there is artesian pressure below the permafrost footing. Although it is possible to insulate the casing, it is difficult to block the entire ring of thawing ground against these pressures; the water in this ring will move upward, sharply increasing the thawing rate (by convective heat transfer) and decreasing the stability of the thawing ground. It is possible to propose that an analogous mechanism causes the formation of heave hummocks--ground ice.[7]

ARCHING EFFECT

In the second extreme case, the specific weight (in the suspended state) is compensated for partially by the effective stresses that existed before thawing and partially by the shears, which are transmitted in the horizontal direction. Let us assume that the specific weight in the suspended state is $\gamma - \gamma_w$, so that added up the total stress after subtracting the equilibrium pore pressure is $(\gamma - \gamma_w)\ z$ and that the previously existing effective normal stress is perpendicular to the horizontal plane and equal to $\lambda\ (\gamma - \gamma_w)\ z$ where λ is less than 1. The ground for which $\lambda = 1$ must be entirely consolidated in the frozen state. As we show in Figure 3, the section dz between two horizontal planes can during thawing remain in equilibrium without an increase in the effective stress perpendicular to the horizontal plane if its edges withstand shear equal to $(\pi/2)R(1 - \lambda)\ (\gamma - \gamma_w)$ with respect to the outside radius of the thawing ring; this shear increases lineally with the radius. The ratio of this maximum shear to the effective stress $\lambda\ (\gamma - \gamma_w)z$ is perpendicular to the horizontal plane and equal to

$$(\pi/2)\frac{R}{z} \cdot \frac{1 - \lambda}{\lambda}.$$

For example, if $\lambda = 0.5$ this ratio is $(\pi/2)\ R/z$, and it becomes less than 0.3 if the depth z is greater than 5.2 R. For a coarse-grained soil, this ratio of the shear to the normal effective stress can be withstood after the deformations on the order of 0.02.[8] Inasmuch as in practice the ratio L/R has a value on the order of 40, the large part of the thawing ring can be supported as a result of the arch effect without the necessity for a significant increase in the effective stress. However, in the upper part of the thawing ring itself, the deflection is not entirely effective since the previously-existing effective stresses are too small. The

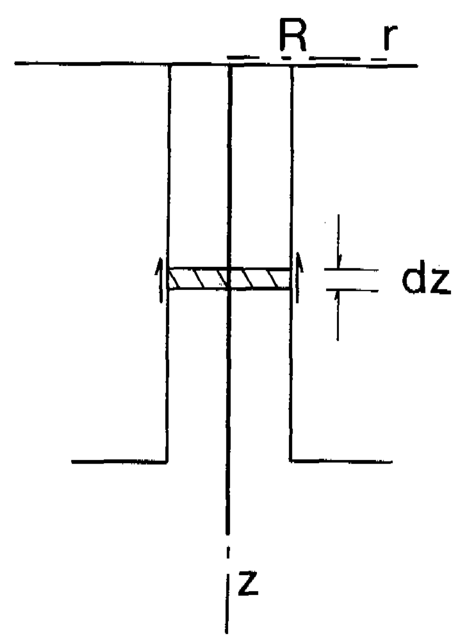

FIGURE 3 Arch effect. r--radius from the casing axis; R--radius of the thawing zone; z--depth from the surface.

above presented analysis uses the classical arch effect analysis performed by Janssen.[9]

The ground displacements that are necessary to cause deflection, are very insignificant. The displacements of the upper boundary of the ring of thawing ground are negligibly small: They are the consequence of additional loads that are transmitted to the surrounding frozen ground by the shear forces acting at the thawing front, and the corresponding stresses decrease appreciably more rapidly than r^{-1} with an increase in the thawing radius. If the vertical displacement at the enter of the ring of thawing ground is Δ and it is assumed that the shear deformations increase linearly with the shearing stress, then its deformation at the outer boundary of the ring of thawing ground is $2\ \Delta/R$. If it is 0.02, Δ is 0.01 R or 0.15 m for a thawing radius of 15 m; the settlement will reach a depth on the order of 5.2 R or 80 m. Inasmuch as the ratio of the maximum shear stress to the normal effective stress drops with depth z, the deformations required for this stress ratio also decrease with depth and at the same time the vertical movements decrease.

If the displacements of the casing are the same as the vertical displacements of the soil in the center of the thawing ring, then the axial deformation in the casing is a function of the displacement with respect to depth. Assuming that the ratio of the shear to the normal effective stress equal to 0.3 is caused by shear deformation equal to 0.02, which is proportional to this ratio, it is possible to give the following approximate expression for displacement of the center of the ring of thawing ground:

$$\Delta = 0.1\ \frac{1 - \lambda}{\lambda} \cdot \frac{R^2}{z}\ ,$$

and the axial compressive stress in the casing is approximately

$$\sigma_c = 0.1\ \frac{1 - \lambda}{\lambda}\ (R/z)^2 \cdot E,$$

where E is the Young's modulus of the casing material.

As shown above, if z is very small this solution is inapplicable. It predicts the acceptable stresses in the casing for the section where its behavior is determined by deformation.

INTERRELATION OF DEFORMATION AND CONSOLIDATION DURING THAWING

When the ground thawing around the casing of the well is not completely consolidated, part of its weight cannot withstand the existing intergranular forces and the corresponding effective stresses. It was demonstrated that the deformation effect can withstand a significant part of the ring of thawing ground and that highly insignificant displacements are sufficient for the effect to develop, but it is not effective in the upper part of the thawing zone to a depth of approximately 5 R. This has been demonstrated schematically in Figure 4; in the upper part the ground is consolidated to the greatest degree: Part of the load is taken by the deforming layer, and part is compensated for by the increasing effective stresses. At the surface, the axial deformation in the ground is on the order of the consolidation during thawing--ε_t. Near the surface, the shear forces that can be transmitted by the ground are small; they are insufficiently large to force the casing to be displaced with the ground; the ground is displaced downward with respect to the casing and the corresponding increase in the stresses of the casing with a depth is subject to Equation (4). In the section where settlement predominates, the casing can be displaced with the ground and its stress is given in Equation (6). There is also a transition zone where the pipe does not follow the displacement of the ground but the settlement is only partially effective. Here the displacement Δ decrease with depth (since the soil must be consolidated), and the axial consolidation--$d\Delta/dz$--also decreases with depth (inasmuch as the required increase in effective stress decreases with depth when the deformation becomes more important). From this it follows that in the transition zone the axial consolidation in the casing also must decrease with depth. The consolidation stress decreases analogously and always has a value less than the stress given by Equation (6). In this section the forces on the casing must be upward, and the consolidation stress in the casing reaches the maximum value at the point y--near the upper end of the transition zone and the lower end of the zone where the ground is displaced downward with respect to the casing. It is possible to set the upper boundary of the maximum stress assuming that (6) is located below the point y, and (4) above it. For simplicity we assume that $\gamma - \gamma_w$ is uniform with respect to depth, and z at the point y is equal to y. Then, inasmuch as the stress of the casing must be continuous at the point y:

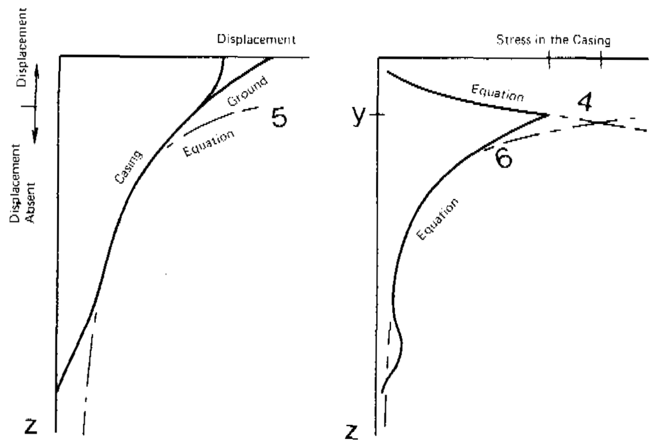

FIGURE 4 Interrelation of deformation and consolidation. y--depth below which the ground does not shirt with respect to the casing; z--depth from the surface.

$$\mu k_0(\gamma - \gamma_w)y^2/2t = 0.1\,\frac{1-\lambda}{\lambda}\,(R/y)^2E,$$

hence,

$$y/R = \left\{0.2\,\frac{1-\lambda}{\lambda}\cdot\frac{tE}{\mu k_0(\gamma-\gamma_w)R^2}\right\}^{1/4}.$$

If $t = 20$ mm, $\lambda = 0.5$; $E = 210\ kN/mm^2$, $\gamma - \gamma_w = 8\ kN/mm^2$, $R = 15$ m, $k_0 = 0.6$ m, and $\mu = 0.2$, then $y/R = 7.9$.

The corresponding stress of the casing is 340 N/mm^2 (49,000 ft/in).[2] It must be emphasized that this calculation gives the upper boundary of the maximum stress of the casing and is not typical for any real case. The numerical coefficient 0.1 in Equation (6) and in the right-hand side of Equation (7) depends on the properties of the medium; in the derivation it was assumed that for a stress ratio equal to 0.3 it is necessary to have shear deformation equal to 0.02.

Of course, these calculations are approximate, but in a number of cases they are sufficient to demonstrate that settlement has no significant effect on the casing. It is interesting to note that the stresses predicted by this calculation have the same order of magnitude as those arrived at by Koch[1] and Smith and Clegg.[10]

The forecast of the displacments calculated by Equation (5) shows that they asymptotically approach 0 when z is sufficiently large at the same time as in reality below the permafrost layer they probably do not exist in general. In the base of the thawing ring there must be another section where the sliding between the soil and the casing leads to equalization of the displacements. As was demonstrated in Figure 4 a small increase in the consoldiation stress can occur in the casing. However, it must be small, since all of the displacements are small and the surrounding medium is sufficiently deformed.

CONCLUSIONS

What field data are most important when evaluating the settlement and the stresses on the casing according to the present analysis?

It is extraordinarily important to know how far the permafrost is initially from the condition of complete consolidation, that is, to what degree the load from the overburden is taken up by the effective stresses existing before thawing. If the ground is completely consolidated ($\lambda = 1$), then the settlement during thawing will be very little. If λ is close to 1, then the settlement will be significant; large displacements will occur only in direct proximity to the surface, and the stresses in the casing will be very small. If λ is small, the deformation is appreciably less effective. The degree of initial consolidation is more a property of the construction sites than the type of ground, and it can be determined directly only by testing for thaw-settlement in samples of known composition, although important additional information comes the geology and the geomorphology of the region.

Other properties of the soil are also important, but many of them can be determined with sufficient accuracy in disturbed samples. In particular, in the laboratory it is possible to establish the relation between the stress and strain, which determines the magnitude of the vertical displacement required for deflection to occur. The stress in the soil in nature is also important; however, until recently information about the initial stresses in permafrost has been unavailable.

The above calculations show that the thawing system is quite simple in a number of cases, and it is possible to draw useful conclusions beginning with the basic principles of soil mechanics. A study was made of the case of an entirely uninsulated borehole, but the same principles can be used also for other cases: for example, such as when the thawing near a borehole decreases at the cooling surface and the casing is partially subjected to axial tension. This would be extremely valuable if the calculations were checked by comparison with field observations in the existing boreholes. Koch[1] obtained results from a valid test in which the thawing radius reached 6 m, but no significant settlement was observed. The results of more prolonged tests would be of greater interest.

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PRINCIPLES OF PLANNING, DESIGN, AND CONSTRUCTION OF HIGHWAYS ON PERMAFROST

A. A. MALYSHEV AND B. I. POPOV
Omsk Branch of the Soyuzdornii Institute

This paper deals with the principles of designing and building highways in a permafrost area, developed on the basis of many years of studies in the Omsk branch of the Soyuzdornii Institute.

The permafrost zone is characterized by various soil and natural-climatic conditions. Therefore, for the design and construction of highways in this zone, it was necessary to subdivide it into road-climate areas and develop different norms and technical solutions for them.

The factors[1,2] that have a decisive effect on the stability of road construction were taken as the districting principle: the type of soil of the active (seasonally thawing) layer and its water content, the nature of distribution of the permafrost, and the temperature. Four large regions have been isolated.

The first is the region of high water content (above the liqud limit) soils of the seasonally thawing layer. It primarily encompasses the tundra and is characterized by a complex distribution of permafrost 300-500 m thick, and more, with a temperature (at a depth of the annual zero amplitude) of -8°C to 12°C. The ground is soft silty tundra type with a predominance of silty suglinoks and clays. The depth of seasonal thawing is to 1 m.

The second region is one of moderate water content (up to the liquid limit) soil of the seasonally thawing layer. The region encompasses the taiga and is characterized by a complex distribution of permafrost 100-300 m thick with a temperature of -1.0°C to 8.0°C. The soil is acid taiga, primarily podsolic suglinoks and clays, rarely supesses and sands. The seasonal thawing depth is 1 m to 2.5 m.

The third region is a region of low water content (less than 0.8 of the liquid limit) of the soil in the seasonally thawing layer. The region encompasses the mountain taiga and is characterized by a complex distribution of permafrost 25 m to 200 m thick with a temperature of -1.0°C to -5.0°C. The ground is rock debris, pebbly gravel, and rock. The seasonal thawing depth is 2.0-3.0 m.

The fourth region is the region of high-temperature permafrost with a water content of the seasonally thawing layer from 0.7 to 1.0 of the liquid limit. The region encompasses the forested steppe terrain of the zone and is characterized both by continuous and insular distribution of the permafrost 25-50 m thick with a temperature from 0°C to -1.0°C. The soil is varied--from clay to detritus and pebbly gravel with a seasonal thawing depth to 4.0 m.

The studies[2,3] demonstrated that within the limits of the isolated zones various conditions can be encountered; therefore, the route of the road being built must be divided into sections (with respect to types of terrain) with similar permafrost and hydrogeologic conditions and relief. Three types of terrain are distinguished: dry, wet, and constantly wet. These types of terrain are provided for by the construction Norms and Guidelines (SNiP II-D.5-62); however, for the permaforst zone the proposed division is distinguished with respect to content from the generally accepted ones for the remaining zones.

The dry areas include rocky prominences, steep slopes of mud cones, sandbars made of rock, debris, sandy, and argillaceous nonsubsiding soil with a water content of less than 0.77 w_L. By the known surface runoff, the thickness of the seasonally thawing layer in these places reaches 2.5 m and more. The wet areas include gently-sloping mountains with southern exposure and flat watersheds made up of sandy and argillaceous low-subsiding soil with a water content not higher than the liquid limit. With poor surface runoff, the thickness of the seasonally thawing layer does not exceed 2.5 m. The constantly wet zones include Mari, closed, and other lowland sections with poor drainage and excess moisture. The thickness of the seasonally thawing layer does not exceed 1.3 m. The soil is argillaceous, ice saturated, and sharply subsiding with a water content above the liquid limit.

The division into districts with isolation of the types of terrain permitted the development of a method of planning and designing roadways based on selecting the defined principle of soil use (in frozen or thawed state) as the basis for the earth fill. Three planning principles were proposed:

The first entails raising the surface of the permafrost to the base of the fill and maintaining it at this level for the entire period of operation of the road.

The second principle limits the depth of the base soil under the roadway beginning with admissible deformations.

The third principle ensures maximum thawing and drainage of the permafrost under the road.

The principles of design and construction of a soil roadbed are given for each characteristic section of the route as a function of the type of terrain, the degree of subsidability of the base soil, and considering the results of technico-economic calculations (Table 1).

From Table 1 it is obvious that, when planning and designing by the first principles, the earth

TABLE 1

Type of Terrain	Subsidability of the Soil Base	Planning Principle	Type of Structural Design and Soil of the Roadbed
Constantly wet	Sharply subsiding, $e > 10\%$	First	Fills made of noncohesive coarse soils; argillaceous soils permitted in the second stage
Wet	Low-subsiding, $3\% < e < 10\%$	Second and third	Fills made of clayey and noncohesive coarse soil; cuts are permitted
Dry	Nonsubsiding, $e < 3\%$	According to SNiP II-D.5-62 for the II road-climatic zone	Fills and cuts according to the general SNiP and VSN norms.

NOTE: e--relative subsidence during thawing (%) equal to the ratio of the settlement to the depth of thaw.

roadbed will be constructed and the embankments (not using roadside borrow) made of noncohesive coarse granular materials that have been brought in. Here, it is mandatory to keep the moss-vegetative cover at the base of the fill in the undisturbed condition. In a forested area the width of the cut must be limited to the width of the fill through it. In order to protect the moss-vegetative cover from destruction, it is recommended that in the lower part of the fill provision be made for a layer of free-draining coarse soil (not bigger than 70-100 mm). In a number of cases, primarily to reduce the height of the fills, it is necessary to use insulating interlayers of local materials in them having low thermal conductivity and sufficient strength (packed moss and peat, wood, slag) or artificial materials (foamed plastic, foamed polystyrene, and so on).

When planning by the second principle, the earth roadbed is constructed in fills made of local cohesive soils with use of route materials (see Figure 1). A mandatory condition is keeping the moss-vegetative cover at the base of the fill.

It is permissible to plan cuts primarily in sections with favorable soil-hydrogeological conditions--rocky, debris, gravelly soil in the absence of ice interlayers in the foundation. Under complicated soil-hydrogeological conditions (wet areas), cuts are permitted by special plans providing for thermal insulation of the fills and embankments, replacement of the "superwet" argillaceous soil within the traffic way by sandy soil and other high-quality materials and the construction of heat-insulating layers in the base.

According to the third principle, the earth roadbed is designed for areas made up of sandy and supessy soils, with preplanned thawing and drainage.

When planning the earth roadbed a great deal of singificance is attached to the steepness of the banks of the fills and cuts. The banks must be more gently sloping than under ordinary conditions. As demonstrated by research,[4] this makes it possible to get a smooth (in transverse cross section) outline of the upper boundary of the permafrost under the roadbed, as a result of which the stability of the entire road is improved. The most efficient bank slope is from 1:3 to 1:5 (depending on the depth of the thaw and the depth of the cut) if the roadbed is made

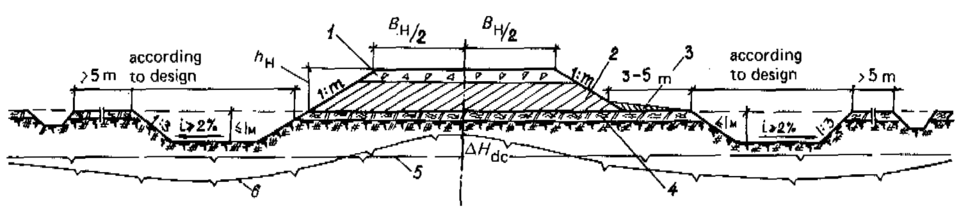

FIGURE 1 Transverse profile of a fill made of local clayey soils. 1--crushed rock or gravel designed for strength (but not less than 0.5 m); 2--clayey soil; 3--earth berm; 4--moss-vegetative cover; 5--upper boundary of the permafrost before filling; 6--upper limit of the permafrost after filling; ΔH_{dc} is the rise of the permafrost table along the centerline of the road.

of clay and from 1:1.5 to 1:2 if the roadbed is made of noncohesive crushed rock.

In doing the operations during the summer it is preferable to construct a roadbed of clay soils. This permits a reduction in cost by 3 to 4 times. In order to build the road structure (see Figure 1), clay, suglinok, and supes are used, but differentiated depending on the road-climatic districts of the zone and the type of paving used.[2] Thus, heavy clay and suglinok are permitted for the construction of a roadbed primarily in the second to fourth road-climatic zones for all types of road pavings. The heavy silty suglinok is used in all road-climatic districts if the total content of silt and clay particles in it does not exceed 60 percent; for a high content (≈70 percent), this soil can be used only for roads built with improved, light transition-types of paving. The light silty suglinoks and silty supesses are permitted if the total content of silt and clay particles does not exceed 50 percent. With high indexes it is permitted only when building roads with transition types of paving.

The locations of side stakes for the roadbed are determined by thermal calculations. The calculations are made beginning with the requirement that the settlement (S_p) of the soil in the base does not exceed the admissible (S_d) for a paving of the adopted type, that is, $S_p \leq S_d$.

On the basis of this it is proposed that the height of the fill (H) be defined by the following formulas:

For the first design principle

$$H = H_p \cdot m_t \cdot K_g,$$

for the second

$$H = H_p \cdot m_t \frac{H_p \cdot S_d}{H_p - H_{ds}} + S,$$

for the third

$$H = H_p \cdot m_t - \frac{H_p \cdot S_d}{H_{ds} \cdot e} + S,$$

where H_p is the calculated depth of thaw of the soil making up the fill, fill, defined according to SNiP II-b.6-66, m; m_t is a coefficient of thermal effect of the paving; for cement-concrete paving, it is 1.05; for blacktop, from 1.1 to 1.6, depending on the type of soil in the fill and the calculated surface temperatures; K_g is a coefficient taking into account the additional heat flux in the bearing surface as a result of development and the draingage strip of the roadbed; it is assumed equal to 1.16 to 1.22, depending on the road-climatic regions; H_{ds} is the thickness of the seasonally-thawing layer before drainage of the roadbed, determined according to SNiP II-B.6-66, ms; e is the relative settlement of the ground in the base of the fill as a fraction after thawing under a load defined according to SNiP II-B.6-66; S is the settlement of the soil in the base taking place during the construction period and requiring additional volumes of earthwork; it is taken according to the data in Table 2.

In building the roadbed from noncohesive coarse soil and clayey soil, the rules arising from the

TABLE 2

Type of Terrain	Natural Moisture Content of the Bearing Soil as a Fraction of the Liquid Limit	Coefficient of Consistency[a]	Soil in the Bearing Surface	Settlement, cm
Wet	> 1.0	> 1.0	Silty clay	30
			Silty suglinok	20
			Heavy supes	15
			Silty sand	10
Damp	0.77 to 1.0	0.5-1.0	Silty clay	20
			Silty suglinok	15
			Light silty supes	10
			Silty sand	6
Dry	< 0.7	< 0.5	Silty clay	10
			Silty suglinok	6
			Light supes	5
			Silty sand	4

NOTE: The settlement of the moss-vegetative cover is taken as 50 percent, and the settlement of the moss-peat cover is taken as 40 percent, of the initial thickness recorded during the studies.

[a] $\left[\frac{w_L - w}{I_p}\right]$

planning and designing principles must be strictly observed.

In sections of terrain where the roadbed is designed according to the first principle, maximum maintenance of the natural conditions of the terrain existing before the beginning of operations is provided for. It is not permissible to make cuts in anticipation (for operations in the next year), or to uproot stumps and disturb the moss-peat cover at a distance closer than 100 m to the edge of the cut.

In sections where the roadbed is designed by the second principle, all operations with respect to clearing the right-of-way are done in the winter, in the spring the moss-peat cover is removed from the borrow area and the drainage ditches are constructed. The work is done no later than a year before the beginning of the basic earthwork, so that the ground in the borrow pile will drain for a year. In sections where the roadbed is being constructed by the third principle, the right-of-way is prepared taking into account maximum drainage of the soil in the seasonally thawing layer and permafrost for which all of the preparatory operation, including removal of the moss-peat layer from the entire right-of-way, is done not later than a year before the basic earthwork.

The roadbed designed by the first principle is built observing the following rules:

1. It is drained in the winter to the complete height after freezing of the base to a depth of not less than 30 cm.
2. The lower layers of the fill to a height of 0.5 m are drained "naturally" and the next ones, longitudinally.
3. In the mari flooded sections, the ice cover is removed to the width of the fill base before draining the roadbed, and the trees and brush cut when clearing are laid in the form of a cover or paving.

The roadbed designed by the second principle is built as under ordinary conditions but with mandatory preservation of the moss-peat in the foundation. For this purpose, the first layer is drained "naturally." When building the roadbed from clayey soils, the road machinery is arranged at the front and the time alloted for earthwork is designated with consideration of its natural water content. With a soil moisture content in the borrow of 0.7 to -0.8 of the liquid limit the soil is worked in layers, as it thaws, at 15-20 cm. If the moisture content is less than 0.7 of the liquid limit, deep freezing if permitted, and the soil is worked in layers as it is drained from the surface.[3]

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SCIENTIFIC-PROCEDURAL PRINCIPLES OF INVESTIGATION AND THE MAIN FEATURES OF THE HYDROGEOLOGY OF NORTHERN COUNTRIES (EXEMPLIFIED BY NORTHEASTERN USSR)

A. I. KALABIN Moscow Lomonosov State University

In the mountainous part of the Verkhoyansk-Chukotsk region and in the northern and the northeastern parts of Asia as a whole, the formation of groundwater reserves is subject to broad zonality and high discontinuity. This is connected with the peculiarities of the distribution of permafrost and of tectonics. The distribution and thickness of the permafrost depend on latitude and altitude.

The vertical zonality of the permafrost zone in the northeastern USSR and in the northern parts of eastern Siberia is characterized by the following interrelated laws: (a) with increase in the sea-level height of the terrain, the thickness of the permafrost increases; (b) independent of the sea-level height of the given terrain, the thickness of the permafrost under the prominences is greater than under the depressions; (c) in mountainous areas the lower surface of the permafrost is an undulating surface repeating the shape of the Earth's surface in a somewhat smoothed form; (d) the thickness of the permafrost and its temperature, as a rule, are interrelated; (e) the least thickness and the highest temperature of the permafrost, the greatest fracturing and saturation of the rock are noted at the bottom of river valleys independent of the other conditions; (f) the depth of the regional weathering in thick permafrost is appreciably higher than under ordinary conditions (it reaches 1,000 m and more).

In the region studied, 3 types and 13 subtypes of groundwater (see Table 1) are isolated and described with respect to geostructural, lithologic, genetic, and hydrodynamic factors, and also by distribution. The subpermafrost water and suprapermafrost water of the river valleys have the greatest importance.

In general, the classifications of groundwater must be created for each of the three basic zones of permafrost: insular, discontinuous, and continuous.

The thick, sandy-gravel-shingle deposits playing the role of intermediate reservoirs are the most important in the formation of the suprapermafrost water reserves in the river valleys. A large amount of suprapermafrost (atmospheric) water accumulates in these aquifers. They can feed the subpermafrost water, form springs, polyn'ya, and naleds.

With the passage of time in the foothills and in any river valley where there are constant flows of surface and suprapermafrost water during the warm part of the year, even that of atmospheric origin, water-bearing taliks are formed, within the limits of the entire series, in highly permeable soil. Under the small flows with a watershed up to 25-30 km^2, the taliks are found only under an active channel.

The large permanent suprapermafrost taliks are an important factor in the formation of the subpermafrost water reserves, the points of their supply and discharge. Therefore, they, just as the through-taliks, must be shown on the permafrost-hydrogeological maps.

The suprapermafrost (closed) and through-taliks have dual significance in hydrogeology: as regions of accumulation and circulation of groundwater of different origin and as transition paths along which the interaction of surface and suprapermafrost water with the subpermafrost water and the supply and discharge of the latter are realized.

Deep fractures, faults, zone of fissuring and crushing, fissure veins, and dikes occurring in connection with intense folding at the end of the Mesozoic and most recent movement of the Earth's crust are clearly expressed in the geological structures of the mountainous regions of the Northeast. The modern conditions of the formation of the subpermafrost water are connected with the tectonic manifestations of the Quaternary period. Under modern conditions [water] supply comes primarily from the infiltration cycle.

The supply, runoff, and discharge of subpermafrost water take place in most cases in the same river valleys.

Within the confines of the mountainous regions where the permafrost is almost continuous, the groundwater (subpermafrost) runoff coincides in general with the direction and position of the modern hydrographic network. Here, small suprapermafrost subchannel taliks are noted in the upper course, then thick floodplain, suprapermafrost taliks and lower down, in the presence of year-round runoff, through-taliks.

The best way to tie the permafrost beds to the hydrogeological structures is the permafrost-hydrogeological map compiled on the basis of the water-bearing complexes.

The water-bearing complex, with the permafrost characteristics imposed on it, is the main element from exploratory analysis of the hydrogeological structures. The water-bearing complexes appear in a definite combination of the relief, the tectonics, and neotectonics.

The primary water-bearing factors of the hydrogeological structures must be considered to be the presence of cryogenous interpermafrost fissures and zones of tectonic fractures and also the ratio of the thickness of the permafrost zone

TABLE 1 Abbreviated Exploration Classification of Groundwater in the Northeastern USSR and Other Parts of the Earth Where there are Thick Permafrost Zones

Types (Classification) of Groundwater	Groundwater Subtypes	Characteristics of the Country Rock	Feed Sources	Water Regime and Properties	Distribution, Reserves, Possibilities of Use for Winter Water Supply and the Creation of Artificial Reserves
1	2	3	4	5	6
Supra-permafrost (pore and fissure) water	1. Water of the active layer of the watersheds and interfluvial areas of mountainous regions	Quaternary deposits and fissured rock of the weathering zone	Atmospheric precipitation and water vapor condensed in the ground on the slopes	Freezes during the first half of the winter, forms small springs and ice bodies. Fresh water, summer temperature 0°-5°	Reserves inconstant and insignificant. In the summer it can be used for water supply. An increase in the reserves is possible using snow retention and underground dams on the slopes.
	2. Water of the active layer of the plains and lowlands	The same	The same	The same, but difficult, runoff and usually ice bodies and springs are not formed	Practical utilization difficult. Artificial reserves can be created by thawing water out of the frozen ground.
	3. Water of the flood-plain taliks with surface (atmospheric) supply	Predominantly modern alluvial deposits and fissured rock of the weathering zone	Water of the active layer	Magnitude and duration of the runoff depend on the basin dimensions and the thickness of the talik. Water fresh, winter temperature less than 0.5°. Springs and medium ice bodies formed	Winter reserves continuously decrease. In the valleys of the medium-sized rivers, highly insignificant even at the end of the winter. Feasible for water supply. The creation of artificial reserves most frequently inexpedient.
	4. Water of the flood-plain taliks of mixed supply	The same	Surface and suprapermafrost water, subpermafrost springs	In the winter runs off and does not freeze. Large springs, polyn'yas and large naleds are formed. Fresh water, temperature in March-April 0.5°-5°	Widespread: in February-April reserves are significant. Reliable source of water supply. Artificial increase in reserves expedient.

	5. Water of the supra-permafrost taliks of foothill alluvial plains and alluvial fans	Very coarse soils	Meteoric, condensation, surface and subpermafrost water	Usually does not freeze entirely, sometimes has runoff the entire winter. Large springs and naleds are formed.	Reserves are large even in the winter. Reliable source of water supply
	6. Water the supra-permafrost taliks of the gentle slopes and uplands	Coarse-grained residual and colluvial soils	The same	In the presence of thick alluvial deposits does not freeze; feeds the flood-plain taliks, sometimes small springs and naleds are formed	Reserves sometimes significant; have no independent significance for winter water supply
Inter-permafrost water	7. Water of the sub-lacustrine suprapermafrost taliks	Usually sand, gravel, fine silt	Water of the lakes and active layer	Does not freeze. Fresh water, winter temperature 1°-2°	In the case of freezing lakes can serve as a source of winter water supply
	8. Inter-permafrost water of the through taliks; pore, fissure and fissure-karstic	Quaternary deposits and bed-rock	Subpermafrost, suprapermafrost, and surface water	Regime varies depending on the feed source. Water temperature 0°-3°. Creates springs and naleds of various dimensions	Reserves significant, inconstant. Used for water supply. Sometimes artificial flooding of the through talik is expedient
	9. Inter-permafrost stratal water of the river valleys	Predominantly modern alluvial deposits	Supra-permafrost and surface water	Often the taliks become waterless by the end of the winter. Average springs and naleds are formed	Where encountered; have no significance for water supply

TABLE 1 (Continued)

Types (Classification) of Groundwater	Groundwater Subtypes	Characteristics of the Country Rock	Feed Sources	Water Regime and Properties	Distribution, Reserves Possibilities of Use for Winter Water Supply and the Creation of Artificial Reserves
1	2	3	4	5	6
Subpermafrost water	10. Subpermafrost water of the loose mantle of the regional fissure weathering zone and shallow karst	(a) Tertiary and Cretaceous conglomerates, gravel, sand, clay; deposits associated with young depressions	Predominantly ground surface water	Regime stable, fresh water, temp. 1°-5°, mineralization to 150-200 mg/l. Springs with discharge from tens of liters to several cubic meters per sec; giant naleds are formed.	Reserves often signigicant. Discharge of single wells 2 to 10 m/s. Suitable for water supply. The creation of reserves is expedient by pumping surface water through the wells and also using explosions
	11. Subpermafrost water of cryogenic jointing occurring directly at the lower surface of the permafrost or near it	(a) Sandstone, sandy-argillaceous slate, hornfels with tuffite, porphyry, conglomerate and limestone interlayers (Permian, Cretaceous)	Predominantly infiltration water and also head water of the deep subpermafrost horizons	The same	Occurs in almost all hydrogeologic structures. Constitutes primary reserves of subpermafrost water. Will discharge 10-15 l/s. Has most important significance for water supply. Creation of artificial reserves expedient
		(b) Crystalline slate, gneiss and sharply metamorphosed sandstone			
		(c) Liparites, dacites, andesites, basalts, and their tuffs of Jurassic, Cretaceous, and Tertiary ages			

	(d) Granites, diorites, granodiorites of Jurassic, Cretaceous, and Tertiary age, and so on			
12. Deep subpermafrost water: fissure, fissure-vein, fissure-karstic	Predominantly in thermokarst and in the zones of Tectonic faulting	Primarily infiltration with vadose and rare internal water	Regime stable, mineralization and temperature fluctuate within broad limits. Springs and naleds rarely formed	Reserves less significant than in the preceding case. Extrusive rock full of water; granites only in the zones of Tectonic fractures. Creation of artificial reserves possible
13. Hot and warm mineral water: (a) coastal strips	Associated with faulted zones appearing in various geological structures with respect to age and composition	Primarily infiltration water of rivers, lakes, seas and the upper horizons of groundwater	Regime stable mineralization 15 g/1 and more; sodium-calcium-chloride water; temperature 20°-90°	Reserves significant; water has varied therapeutic properties. Creation of artificial reserves and artificial thermal springs expedient by flooding fractures with surface water through wells or using powerful explosions.
(b) Continental	The same	The same but without participation of seawater	Weakly mineralized sodium silicic hydrocarbonate water has a temp. to 100°.	The same.

NOTE The classification of the floodplain suprapermafrost water is given primarily as applied to the valleys of freezing mountain rivers.

to the thickness of the zone of regional jointing, weathering, and thermokarst. In the absence of thermokarst in the structures, the defining factor is the cryogenous subpermafrost fracturing.

Our research in 1944-1970[1-7] established that within the permafrost and directly under the modern lower surface of the permafrost, and somewhat lower, there are zones of active cryogenous fissuring in all the rock (see Figure 1).

Frost action in the rock at depth as a result of multiple freezing and thawing under the effect of the change in direction of the freezing pressures has led to the formation of fissured zones. The direction, intensity and dimensions of these cracks (the fissured zones) depend primarily on the lithologic features and petrographic composition of the rock, the degree and the nature of the initial jointing, and degree of saturation. In the more brittle rock under the effect of hydrothermal and thermal stresses, more intense fracturing occurs.

Directly under the modern lower surface of the permafrost, fissuring almost always exists, although sometimes its thickness does not exceed 2-5 m. In most cases its thickness is 20-50 m and, rarely, more.

The initial flooding and subsequent supply of these subpermafrost jointed zones took place and takes place primarily with fresh water from the interpermafrost and through-taliks associated with the fracture zones and other tectonic dislocations with a break in continuity and river channels and also the subpermafrost water expected from the freezing slopes (see Figure 1).

Inasmuch as the through-talik water is supplied from the surface water, it almost always has a high head (not less than the thickness of the permafrost in a given area) with respect to the subpermafrost cryogenous fissuring zone and, consequently, it easily penetrates even into the very thin cryogenous cracks.

The presence of subpermafrost fissuring in general form was noted earlier (in 1916) by A. V. L'vov, who investigated the subpermafrost groundwater and ice bodies in the Trans-Baikal, where the permafrost has an insular nature. However, A. V. L'vov and other researchers (N. Tolstikhin in 1962; O. Tolstikhin in 1968, and so on), noting in general the presence of subpermafrost fissuring and some of its significance for hydrogeology, not only failed to attach primary significance, but even regional significance to this zone. The indicated hydrogeological structures of the cryogenous fissure water were also found and studied in the northeastern part of the USSR in 1966-1968 by co-workers of the Cryogeological Department of Moscow State University (V. A. Kudryavtsev,[8] N. N. Romanovskiy, A. B. Chuzhov, N. I. Trush, V. Ye. Afanasenko, *et al.*), and in 1967-1969 they were studied by O. Tolstikhin.

The cryogenous fissure basins of the hydrogeological massifs and adjacent "negative" structures are in the overwhelming majority of cases closely connected with each other and constitute a united whole, generally a united water pressure system of subpermafrost water of cryogenous fissuring (see Figure 1). The slopes of the hydrogeological massifs are the primary region of supply (accumulation and runoff) of the fissure water, and the cryogenous fissure basins associated with the intermontane depressions, troughs, and synclinal zones; the zones of broad, deep erosion cuts (large river valleys) are the region of discharge, accumulation (sometimes supply), and regional runoff of the fissure water of the subpermafrost circulation.

The characteristics of the subpermafrost water in cryogenous fissures are:

1. It is very widespread in the northeastern USSR and similar territories of the Earth. This water appears in all hydrogeological structures

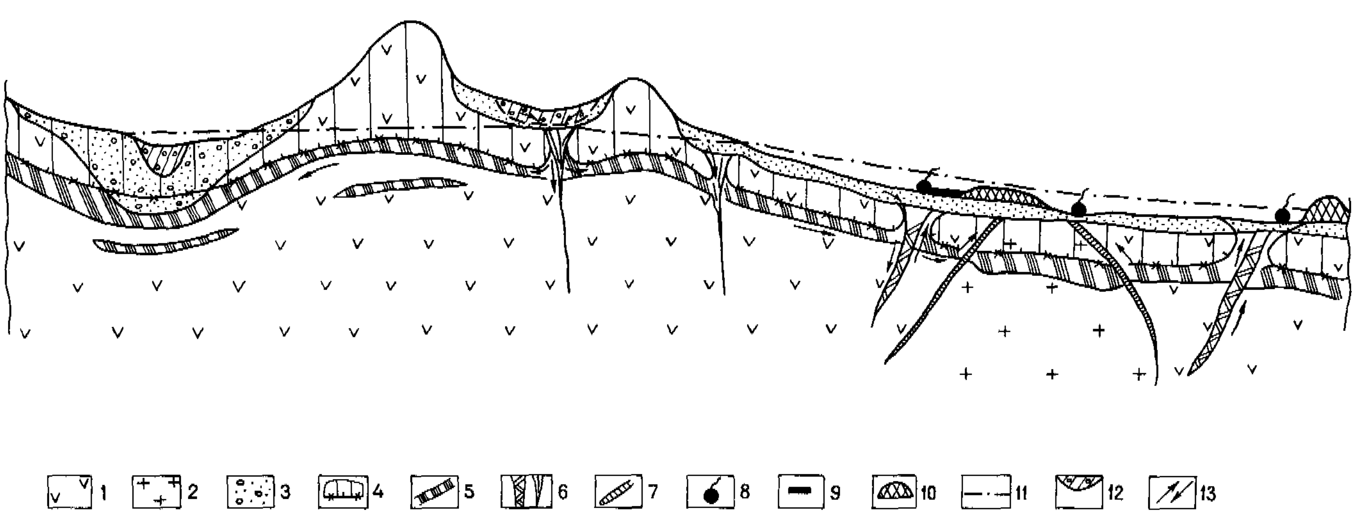

FIGURE 1 Generalized scheme of subpermafrost [water] supply, drainage, and discharge. 1--sedimentary series (sandstones, clay shales, and so on); 2--igneous rock; 3--unconsolidated formations; 4--permafrost; 5--flooded zone of the subpermafrost cryogenic fissures; 6--fractures; 7--jointed zone of intrusive contacts; 8--rising spring; 9--polyn'ya; 10--naleds; 11--piezometric level; 12--suprapermafrost talik; 13--arrows indicate the direction of movement of the water.

where the thickness of the permafrost is greater than the thickness of the unconsolidated mantle and regional fracturing of the weathering zone and where the direction of the freezing process in the past (and in the present) has changed. At this time there are pressure or nonpressure cryogenous basins (water-bearing systems) of fissure water in all hydrogeological structures.

2. In medium and high mountains, the cryogenous pressure basins are usually broken up and exist more frequently under the individual and rarely under several adjacent river basins, intermontane depressions, and other large lowlands. The areas of the subpermafrost flows connected with cryogenous fissuring begin to go beyond the limits of the individual intermontane depressions, river basins, and even the large hydrogeological massifs within the limits of small mountains, synclinal zones, and depressions in the region of stable subsidence expressed by the plains and lowlands. The cryogenous-fissure pressure water here extends within the limits of several river basins and is often controlled by the boundaries of entire geological structures or basins of the largest streams.

3. In the cryogenous fissure basins in water-bearing complexes, the basic reserves of the fresh groundwater are usually concentrated in the form of the primary subpermafrost water-bearing horizons directly connected hydraulically with the subpermafrost and surface water. This water is the most important source of industrial and potable water supply.

4. The geological and permafrost-hydrogeological interpretation of the cryogenous subpermafrost fissure zones can be a method of studying the problems of the dynamics of permafrost.

According to the existing concepts in the northeastern part of the USSR[9-10] there are four interchangeable cryogenous-hydrogeological belts: the top belt is the pre-naled high altitude belt where there is accumulation of heat from suprapermafrost and channel water; the second belt is the infiltraiton and inflow of surface water with characteristic attributes--naleds, polyn'yas, predominance of channel taliks over floodplain, with the presence of supply sections of subpermafrost water through the taliks; the third belt is a transit and accumulation belt where various floodplain taliks, the primary naleds, and polyn'yas forming the so-called naled belt are developed. The lowest belt appears at the base--the groundwater discharge belt.

In reality, such clearly expressed belts and their subsequent interchange in the northeastern USSR and eastern Siberia have not been noted. The number and sequence of arrangement of the permafrost-hydrogeological zones on the slopes of the hydrogeological massifs and other structures appear in a strictly defined combination in each specific river basin as a function of the peculiarities of the entire set of natural situations and, above all, the supply conditions and permafrost-geotectonic structure of the basin.

The belt of hydrothermal accumulation is simultaneously a region of accumulation, supply, and discharge of groundwater, and the third belt combines the functions of transit and the first two belts. The discharge, supply, and hydrothermal accumulation take place in all belts. They appear only under defined geostructural and permafrost conditions. The hypsometric position and number of them are usually connected with the permafrost zonality and tectonics.

The springs and groundwater naleds are widespread within almost the entire mountainous area of Yano-Chukotsk. The springs and large and gigantic naleds are most widespread in the central part of the northeastern USSR and Trans-Baikal.

The significance of rising springs of subpermafrost water consists in the fact that under favorable conditions they create thick water-bearing taliks in the river valleys interrelated with the suprapermafrost streams. For large and gigantic naleds the primary source is subpermafrost water.

The role of suprapermafrost water in the formation of naleds decreases in the direction from the intermountain depressions and river valleys to the divides.

In spite of existing concepts,[10] in the northeastern USSR and in other analogous places there are no so-called naled belts lying "in a strictly defined range of absolute altitudes." For example, the powerful springs and naled lines (not of the belt) are associated with the faults.

The medium and even the large and the gigantic naleds can be created by mixed river and suprapermafrost water held at the beinning of the winter in the powerful floodplain water-bearing taliks. Numerous medium and small naleds occurring as a result of the sources of suprapermafrost and subpermafrost water are created at any sea-level elevation in the low, medium, and high mountains. Their general qualititative significance is entirely comparable to the naleds of the subpermafrost water. The presence of stable taliks is not a mandatory condition for the formation of naleds. The naleds of the subpermafrost and mixed waters, on the contrary, can occur only in the absence of taliks for accumulation and runoff.

Within the Okhotsk-Kolyma and the Omolono-Yablonsk volcanic uplands, the Anadyr' plateau, and the Chukotsk volcanogenic highlands hot springs appear in the young faults. Here, more than 25 large thermal springs have been investigated. With respect to the physicochemical properties, two of the basic groups of springs are noted: the strongly mineralized springs of the coastal belt and the poorly mineralized springs of the continental belts. The latter have discharges of 0.2 m^3/s to 20 m^3/s and mineralization of 0.15 g/l to 0.7 g/l. The mineralization of the coastal chloride water is 5 to 10 times more.

In the newly occurring or "renewed" tectonic faults, the formation of thermal springs has two stages: In the first stage the heating of the fissure groundwater occurs in the nonstationary thermal field as a result of the cooling of the country rock, and the water temperature gradually drops, until the primary role in its heating begins to be the geothermal flux. From that time a second stable stage of temperature formation of the groundwater appears. The spring temperature depends on the "thermal age" of the system forming it. The number of thermal springs and their temperatures, for example, within the Okhotsko-Chukotsk volcanogenic belt decreases with an increase in age of the volcanic rock. However, these conditions of spring formation pertain to faults located in the regions with usually normal geothermal conditions. The other spatially bounded sources of heat for water in the faults are volcanic heat and radioactive heating of the rock. The springs take away a large quantity of heat, which in practice is still used very insignificantly (the springs of the Lorinskaya valley in Chukotka; the Tal'skiy hot springs in the upper course of the Kolyma River).

An increase in the discharge of these springs and also the creation of artificial thermal springs can be achieved by artificial flooding of the faulted zones with surface water in which hot springs are formed. The most favorable conditions for the appearance of hot mineral springs are found in the coastal belt, primarily within the limits of the Okhotsko-Chukotsk volcanic belt. Although the active magmatic activity of this belt also ended at the end of the Neogene encompassing partially the lower Quaternary time, tectonic phenomena in this mountainous area continue even today.

In the hydrogeologic structures, three basic stages of groundwater are noted: The upper stage includes the suprapermafrost and interpermafrost; the middle stage includes primarily the subpermafrost water of the cryogenous fissures and depressions; and the lower stage, the deep groundwater.

The reserves of fresh subpermafrost water of the middle and lower stages are primary reserves. It is possible even to draw a somewhat paradoxical conclusion: These reserves are larger than the reserves in the same intervals in similar hydrogeological structures located outside the permafrost region. This arises from the presence of cryogenous subpermafrost fissuring.

The points of discharge of the subpermafrost water within the thick permafrost zone are highly limited. This gives rise to two laws: (a) the presence of discharge of this subpermafrost water in the southern regions where the permafrost is absent (that is, the northern regions) often serve as regions of supply for groundwater in the south; (b) the runoff of individual rivers in the North is formed at the expense of the discharge of subpermafrost water formed even in individual basins.

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REGULATION OF THE HEAT EXCHANGE OF BUILDINGS AND STRUCTURES WITH VENTILATED PADS ON FROZEN GROUND

O. V. SNEZHKO AND YE. P. KABANOV
Krasnoyarsk Promstroyniiproyekt Institute

One of the possible means of keeping the soil under a building in the frozen state is the proposed method[1] for erecting buildings on porous ventilated pads (see Figure 1).

This idea is based on using fills of sorted gravel, rubble, crushed rock, and similar materials as the porous pad through which forced ventilation by outside air is arranged in the winter in order to remove the excess heat generated by the building, and to cool the foundation.

The required cooling temperature of the ground in the designed cross section (at the end of the air percolation path) can be determined by the formula obtained from solving the heat-balance equation for the cooling and the heating periods under the assumption that the soil before the beginning of the ventilation of the pad is in the frozen state:[2]

$$T_E = \frac{h_0 T_B + T_0 A}{0.4h_0 - 2K_M \frac{t_B}{t_0} - \frac{C_P H}{t_0} + \frac{\lambda_M (L + B)}{4LB} + A}; \quad (1)$$

where

$$A = \frac{1}{t_0}\left[2K_M \sqrt{t_B}\left(\sqrt{t_r} - \sqrt{t_B}\right) - \sqrt{\lambda_M C_I \pi t_0}\ \operatorname{erfc}\sqrt{\frac{\pi t_B}{t_r}}\right],$$

h_0 is the [surface] heat-transfer coefficient between floor and fill; $K_M = \sqrt{\lambda_M C_M / \pi t_B}$ is the heat-assimilation factor of the ground under continuous heating effects; λ_M is the coefficient of thermal conductivity of the frozen ground; C_M and C_P are the volumetric heat capacities of the frozen ground and pad material; H is the height of the fill; L and B are the dimensions of the building in plan view; T_B is the air temperature in the facility during the break in ventilating the fill; T_0 is the ground temperature at the depth of the zero annual amplitudes; t_B is the ventilation period (the cooling period); t_0 is the duration of the break in ventilation (the warming period); t_Y is the duration of the year.

Consideration of the lateral heat losses of the supporting ground in regions with low outside air temperatures in the winter permits, in individual cases, an increase in the required cooling temperature of the ground, that is ventilation of the fill in the fall-winter period with the fans shut down not only in the summer but also in the period with the highest degree of frost. This decreases the heat losses of the building through the floor in the winter.[3] Depending on the ventilation regime of the pad, the mean annual ground temperature in the footing can be reduced to -6°C to -10°C.

The required amount of air for cooling the soil under the pad at the end of the percolation path to the required temperature during the ventilation period is determined from the expression:[4,5]

$$W = \frac{lLn}{-\frac{C_P}{h_0 + 2K_M}\ \ln\frac{h_0 T_B - (h_0 + 2K_M)(T_E - \Delta T)}{h_0 T_B - (h_0 + 2K_M)\overline{T}_A}}, \quad (2)$$

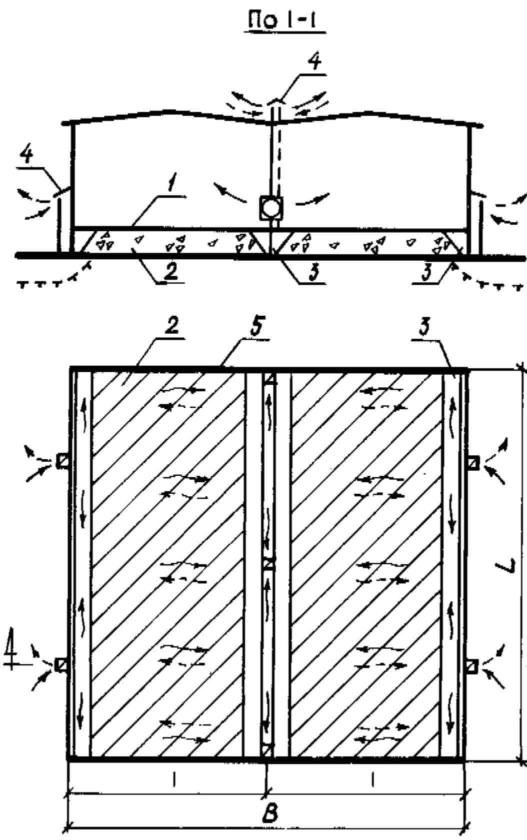

FIGURE 1 Schematic of a building with a floor over a ventilated fill [pad]. 1--floor with thermal insulation; 2--porous ventilated pad; 3--distribution channel; 4--ventilation shafts; 5--moisture barrier.

where l is the percolation path length; n is the number of ventilated sections; h_0 is the heat-transfer coefficient of the floor to the fill; ΔT is the heating of the air at the ground surface with respect to the temperature in the mid-section with respect to height of the fill; $\bar{T}_A$ is the mean air temperature entering the fill.

The values of the temperatures in Formulas (1) and (2) are taken with their own sign.

It has been experimentally established that the air permeability of the fills made of gravel, rubble, and crushed rock depends on the size and uniformity of the particles of the porous layer and increases with an increase in them. The established dependence of the aerodynamic drag on the air velocity is approximated by known theoretical formulas considering the corresponding shape factors of the layer.[6]

In order to discover the effect of the system for inflow and removal of air on the air-exchange conditions in the fill, studies were made on models 2.0 × 3.0 × 0.2 m made of gravel (10-40 mm). The studies demonstrated that the air exchange in the fill under other equal conditions depends on the number and arrangement of the ventilation shafts in the plan view. The air consumption increases with an increase in the number of shafts and with arrangement of the intake in exhaust shafts in a checkered arrangement.

For more uniform cooling of the base of a multispan building (see Figure 1), the fill in the direction of the percolation path can be divided into individual sections. The length of the ventilated section (the air flow path) is assumed on the basis of the technicoeconomic considerations.

As is demonstrated in the Figure 1, the lateral walls in the distribution channels can be banks of natural fill. The air velocity distribution with respect to height of the fill bounded by the natural slopes perpendicular to the flow path depends primarily on the ratio of the length of the bank slope to the mean flow path. With a decrease in its ratio, equalization of velocities takes place. In practice, in fills made of coarse materials with a flow path of more than 10 m, the air velocity with respect to height of the fill can be assumed constant.

Thus, the porous ventilated fills permit regulation of the heat exchange of the buildings and structures with the base and, as one of the promising means of maintaining the frozen state of the ground, they require a good deal of attention by researchers and engineers.

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EFFICIENT COOLING OF THE PERMAFROST FOUNDATION OF A MULTISPAN INDUSTRIAL BUILDING

G. F. SHISHKANOV, G. I. KUZNETSOV, AND N. B. KUTVITSKAYA
Krasnoyarsk Polytechnic Institute

In order to maintain the stability and bearing capacity of a foundation [on permafrost], local cooling of the ground is recommended for the foundations, and it is recommended that the heating effect of the building be limited by arranging a heat-insulating fill under the floor of the first story. An effective cooling system not only permits the frozen state of the soil in the footing to be maintained, but also makes it possible to improve the strength of the more heavily loaded areas of the soil by reducing its temperature. Some thawing is permitted only in the middle of the span, but in this case, in order to avoid deformations of the floor, it is recommended that thermal amelioration techniques or the replacement of the icy ground within the calculated thawing bowl be undertaken.

The cold influx into the ground is brought about with movement of the cold outside air inside the cooling system. Two cooling systems are proposed: In the first, cooling is realized by channels (in the required cases by through ducts) located along the foundations within the fill or with some bearing on the ground (Figure 1a). The ducts can be made of standard reinforced-concrete channel sections or retaining walls covered with slabs. According to the second system, the cold [air] goes into the ground into the entire depth of the reinforced zone by movement of the air through vertical or inclined freezing columns or floor piles. The air in the column (pile) is fed through horizontal ducts, the designs of which are taken the same as for surface cooling by the first system (Figure 1b). This combined system is effective for cooling plastically frozen ground in the foundation of large-span buildings or for freezing taliks and localizing the effect of the heat sources of increased intensity.

The formation of the temperature field in the footing cooled by the proposed systems is affected by the following factors: the structure of the cooling units (the arrangement, dimensions, and shape of the ducts and columns); the aerodynamic characteristics of the air flow (path, flow rate, velocity) and its temperature; the heat insulating effect of the fill and floor; the natural thermal conditions and the thermophysical properties of the soil in the base; climatic conditions; and size and shape of the building and the microclimate inside it.

The choice of the cooling system and the purpose of its parameters are made on the basis of analyzing the limiting temperature state of the footing investigated by the method of electrothermal analogs on two-dimensional models. When simulating the second system, the total cooling effect of a number of freezing columns provisionally replaced by a continuous slit was taken into account.* The boundary conditions take into account the various combinations of the characteristic dimensions of the cooling element and the building--the height of the duct or the depth of the column h, the width of the duct or the column diameter b, and the width of the span B for the range of standard spans of industrial buildings used at the present time (see Table 1).

The results of the simulation correspond to any values of h, b, or B, multiples of their values and the ratios presented in Table 1. The assignment of the boundary conditions and the construction of the temperature field in the ETA models are realized at relative temperatures u. The mean annual surface temperature of the cooling element t_k corresponds to the relative temperature $u_k = 0$; the surface temperature of the floor of the building t_f--the value of $u_f = 1$; the temperature of the phase transition t_0--the value of

$$u_0 = \frac{t_k}{t_k + \frac{\lambda_T}{\lambda_M} \cdot t_f}.$$

[λ_M is the coefficient of thermal conductivity, frozen; and λ_t, thawed condition.] The lower boundary condition is characterized by the relative temperature u_g corresponding to the natural temperature of the permafrost t_g at the depth where the thermal effect of the building and the cooling systems was not felt in practice. This depth was determined in a series of preliminary solutions in which distortions introduced into the natural temperature regime by the total thermal flux from the building and the surrounding system were analyzed.

During the course of the investigation, the following depths were found under the middle of the span for all the calculated variants (see Table 1) at various boundary temperatures. A study was made of the range of relative temperatures u_0 and u_g from zero to one with steps

*G. I. Kuznetsov and R. T. Shugayeva. Resistance of calculated and actual thermal regime of a frozen dam. *Tr. Instituta VODGEO* (Works of the All Union Scientific Research Institute of Water Supply, Sewer Systems, Hydraulic Engineering Structures and Engineering Hydrology), no. 30. Moscow, 1971.

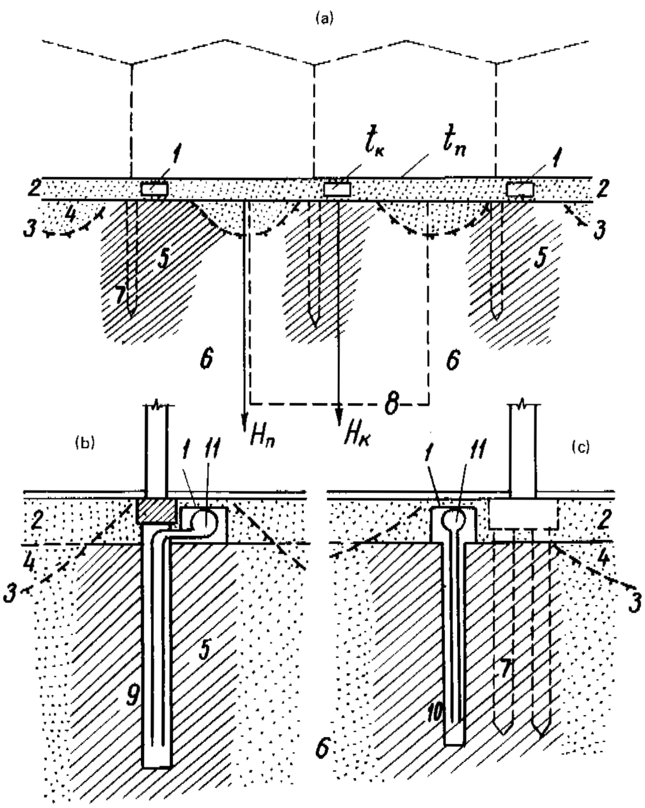

FIGURE 1 Schematic of efficient cooling of the ground: (a) ventilated cooling channel made of prefabricated reinforced-concrete elements (surface cooling); (b,c) ventilated floor piles and freezing columns (deep and combined cooling). 1--cooling channel; 2--heat insulating fill; 3--boundary of the thawing zone in the span; 4--local thawing zone; 5--zone of reduced temperature of the permafrost; 6--natural frozen ground; 7--bearing ground; 8--calculated fragment of the temperature field in the case of analog and physical simulation; 9--reinforced concrete cooled hollow pile; 10--metal freezing column; 11--air duct laid in a cooling channel 1.

of 0.1. This offers the possibility of using the results obtained for any values of t_k, t_f, and t_g. For each value of u_g, the depths of the local thawing bowls under the middle of the span H_f were determined and the graphical relations $H_f = f(u_0, u_g)$ were constructed.

The minimum admissible width of the nonbearing frozen zone separating the local thawing bowls in the adjacent spans is taken as equal to 4 m. In Figure 2a, one of the graphs is presented for variant 1 offering the possibility of determining the depth of local thawing in any middle span of a multispan building with a span width of $B = n \cdot 18$ for a ratio of dimensions of the cooling ducts of $h:b = n(1:2)$ for any boundary temperatures $u_0 = f(t_f, t_k)$ and $u_g = f(t_f, t_k, t_g)$. The developed series of graphs permits selection of the schematic and dimensions of the cooling system so as to obtain a limited thawing depth in the center of the span ensuring adequate dimensions of the cooled supported frozen zone.

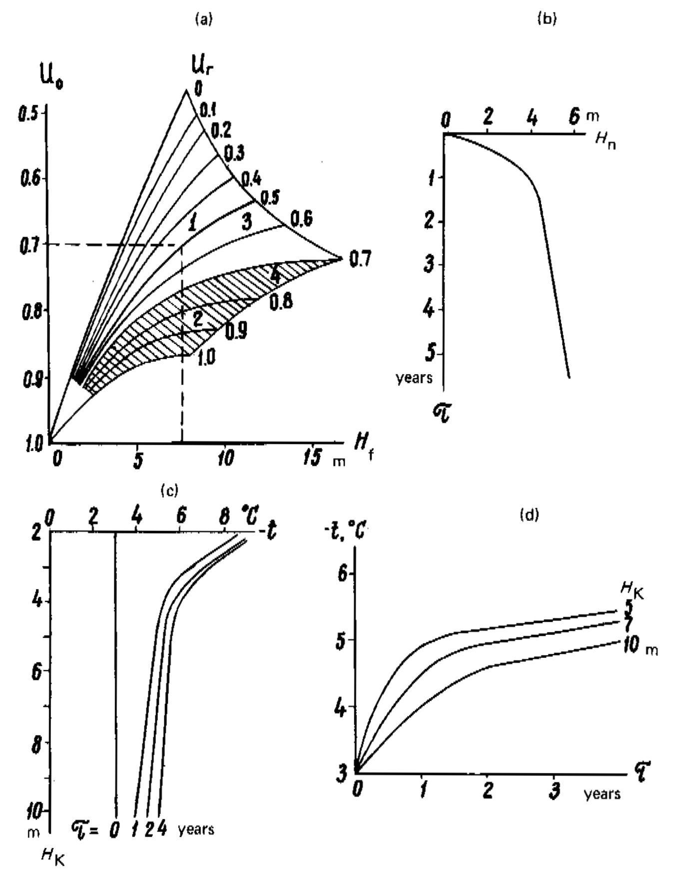

FIGURE 2 Selected results of the analog (a) and physical (b,c,d) simulation for variant I. (a) limiting thawing H_f in the center of the span; (b) thawing dynamics in the center of the span; (c,d) reduction in temperature with respect to depth H_k under the cooling channel at the end of the annual freezing cycles. 1--region of cooling of the permafrost base; 2--region of freezing of the ground; 3--line corresponding to the minimum admissible width of the permafrost zone (4 m) between the adjacent taliks in the spans; 4--line corresponding to the 4-m width of the thawed zone between two adjacent cooled frozen massifs.

TABLE 1

Variants	I	II	III	IV	V	VI	VII	VIII	IX	X
h:b:B	1:2:18	1:1:18	2:1:18	1:1:24	1:2:24	2:1:24	1:1:12	6:06:18	9:0.6:18	12:0.6:18

The development of the temperature field in the footing of the building with cooling ducts was also studied in a physical model. A study was made of the fragment of the soil base (see Figure 1a, variant 1, the table) on a 1:50 and 1:100 scale. The material of the model was medium-grain sand (moisture content 6 percent). The boundary temperatures were simulated with an accuracy of $\pm 1^{\circ}C$. Five series of experiments were performed in which a study was made of the dynamics of the development of the thawed zones and the processes of reducing the temperature in the reinforced frozen zone under the duct for the first 5 yr of operation under various thermal and aerodynamic conditions of operation of the cooling system. The development of the thawing depth in the center of the span is shown in Figure 2b ($t_f = +9.2^{\circ}C$; $t_k = -15.6$, $t_g = 3^{\circ}C$, $\lambda_T/\lambda_M = 0.7$) physical simulation confirms the efficiency of the process of reducing the ground temperature under the cooling ducts (Figure 2c), especially during the first 2 yr of operation of the system (Figure 2d). The summer warming of the ground under the duct is extended to a depth not exceeding 2-3 m.

From what has been discussed it is possible to draw the following conclusions:

1. The studies of the physical model and by the ETA method demonstrated the effectiveness of using the recommended cooling systems to insure stability and carrying capacity of the permafrost for the foundation of a multispan industrial building on the basis of the principle of local forced air cooling of the ground under the foundation with simple design of the cooling units.

2. By varying the parameters of the cooling systems and thermally insulating the fill under the floor, it is possible to control the temperature conditions of the footing in order to increase the strength and bearing capacity.

3. In especially complicated cases (plastically frozen ground, taliks, buildings with increased heat generation, and so on), the combined cooling system is most expedient.

4. The estimated cost calculations performed for the designed products in the Yakut ASSR demonstrated that the cost of constructing the foundation with controlled cooling is 20-40 percent below the cost of alternatives with traditional air spaces.

LIQUID COOLING UNITS FOR FREEZING THAWED GROUND AND COOLING PLASTICALLY FROZEN GROUND FOR CONSTRUCTION IN AREAS WITH HARSH CLIMATES

G. F. BIYANOV, VILYUYGESSTROY, V. I. MAKAROV,
VILYUYSKAYA NIMS, A. D. MOLOCHNIKOV Yakutniiproalmaz Institute

The freezing of thawed ground and cooling of plastically frozen ground is a widespread, efficient engineering procedure used for construction in permafrost areas. The high cost of cooling the ground using brine or cold air forces us to find other, more efficient, methods.

At the beginning of the century (1915, Mombeyg, France), ideas were proposed for using cooling units, the primary characteristic feature of which was natural automatic heat exchange with heat transfer from the soil to the atmosphere during the cold part of the year. It was proposed that the internal heat exchange in such units be realized by gas, liquid, and vapor-liquid media utilizing convection and the cycle of evaporation and condensation of the working media. Genuine efforts at practical implementation of such procedures began in the 1960's.

The idea of vapor-liquid cooling units was used in the designs proposed in the USSR (Tsinman, Blyier, Badyl'kes, 1945-1948) and in the United States by Long.[1] At the present time experimental studies are being made of such devices in the USSR.

S. I. Gapeyev (1957-1967) proposed simple designs of liquid cooling units and demonstrated their efficiency by clear experiments.[2] Three basic structural versions of cooling units exist: the multitube, the two-tube, and the single-tube. They have a number of advantages and are widely used in the design of the Lengiprotrans Institute. There is also the practical application of the units in road, industrial, and civil construction, which demonstrates their high economic efficiency.

At the Yakutniiproalmaz Institute (1964-1970) and also at the Geocryology Institute of the Siberian Department of the USSR Academy of Sciences (1971), versions of the liquid cooling units of coaxial design were developed.[3] In 1969 a report[4] appeared on the development of a liquid thermopile of coaxial design in the United States. The compactness and reliability of the coaxial designs create favorable prerequisites for their practical application.

Natural tests of the coaxial liquid cooling units were made in Mirnyy. Tests were made of devices with a depth of burial in the ground from 5 m to 25 m. Good results were obtained in all the experiments.

The tests of deep cooling units were made in the lower race of the Irelyakhskaya Dam in the thawed saturated ground with a groundwater level of about 6 m from the surface. Coaxial freezing columns 169 mm in diameter are inserted to a depth of 25 m. In order to dissipate the heat in the atmosphere, cast-iron ribbed radiators are used, which are installed with a slight inclination in the direction of motion of the liquid.

The cooling units were put into operation in December 1971, and after 2 months intersection of ice-soil cylinders was noted on all the devices located at a distance of 2 m from each other. In Figure 1 we have the zero isotherms with respect to the cross section of the cooling units in December (before the beginning of operation of the devices), February, and April. The data permit the proposition that under the experimental conditions the interlocking of the ice-soil cylinders could take place even with a column spacing of 3 m if the devices were put into operation from the very beginning of the winter.

The experimental results confirm the expediency of using deep cooling units with natural circulation of the liquid heat-exchange agent to create a waterproof curtain in hydraulic engineering structures in the Far North.

In addition to the units 25 m deep, tests were run in plastically frozen ground on devices with coaxial columns of 10, 8, 6, and 5 m. The latter

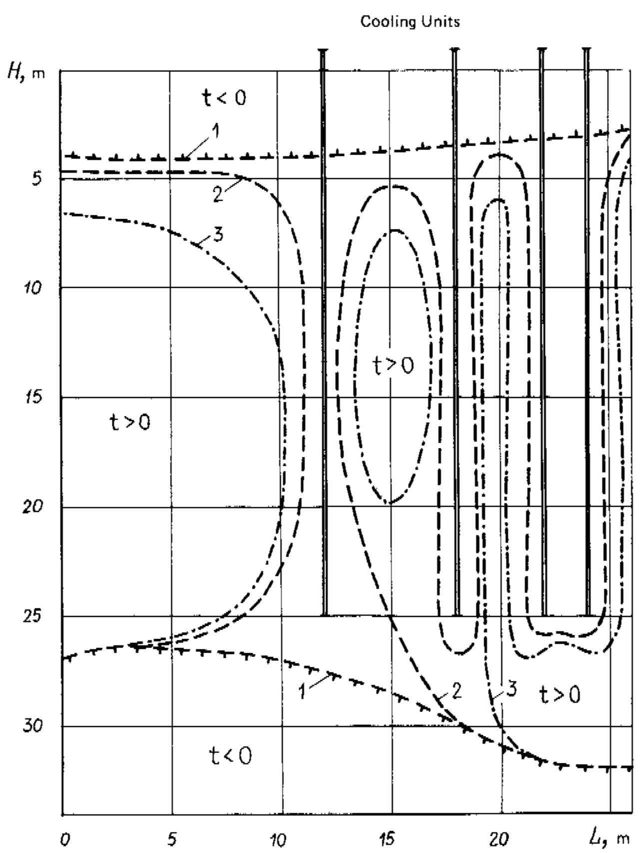

FIGURE 1 Freezing of unfrozen ground by coaxial columns 25 m deep. Zero isotherms: 1--December 1971; 2--February 1972; 3--April 1972.

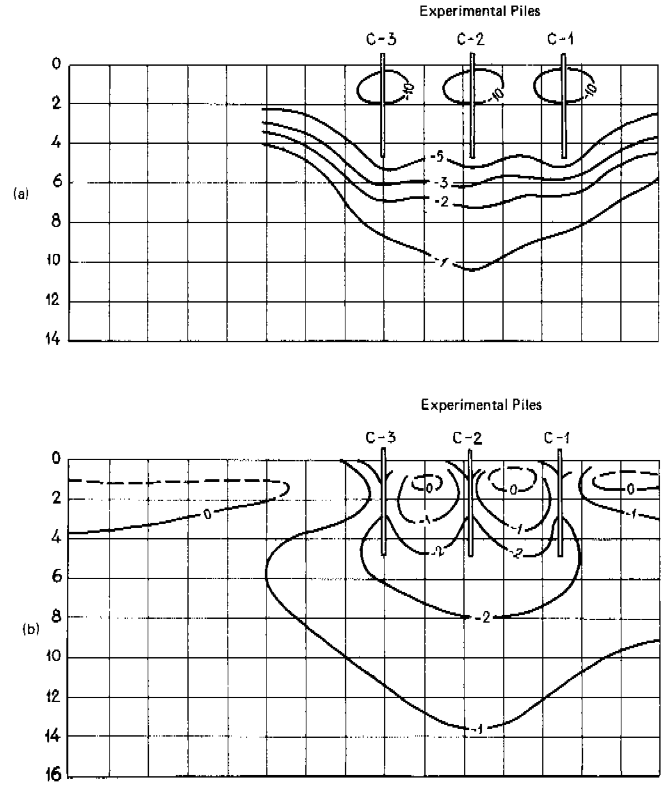

FIGURE 2 Temperature field at an experimental site with thermopiles during characteristic periods: (a) end of April 1969; (b) the same in October.

were designed for placement inside reinforced concrete piles.

The piles with installed cooling units came to be called "cold" or "self-cooling" piles in Mirnyy.

The tests were run on the "cold" piles[3] in an area with plastically frozen ground at a natural temperature of -0.2°C. The experimental units and the piles were installed at the site in accordance with the planned arrangement of the piles under a real building. In Figure 2 we have the isotherms for the experimental site for characteristic periods: at the end of the cooling season and the beginning of the next season. The observations of the dynamics of the thermal conditions of the soil surrounding the experimental piles demonstrated that during the winter the ground temperature drops to -15°C. During the spring and summer, the ground temperature rises to -2°C, after which new cooling begins.

On the basis of natural tests in Mirnyy, the application of reinforced-concrete piles with installed cooling units began.

The technology of manufacturing self-cooling piles was perfected by the Vilyuygesstroy Institute.

The calculations show that the application of liquid thermopiles in the future will permit a reduction in cost of pile foundations to 6 to 8 percent of the cost of major buildings erected on permafrost, with a simultaneous increase in the reliability of the foundations.

In order to ensure broad introduction of liquid thermopiles in major construction, it is necessary to solve a number of problems of which the most important are the following: (a) insuring maximum reliability and service life of the structures; (b) selection of effective working media, inert with respect to ice-cemented rock; (c) a development of compact and feasible structural versions of the devices with minimum consumption of metal; (d) development of thermopile designs that will exclude the possibility of circulation of the liquid during the summer, and so on.

In the most effective cooling units the heat exchange is realized by a liquid circulating in the regime identical to the forced regime. Motion of the liquid is caused by the pressure difference of two communicating liquid columns of equal height. The pressure difference is formed as a result of various liquid temperatures in two "columns" having different heat-exchange conditions with the surrounding environment. The latter is achieved by using coaxial designs, pipe of different diameters, and so on.

When the atmospheric air temperature varies, the liquid temperature in the communicating columns cannot vary synchronously because of different heat-exchange conditions. The temperature difference that arises shapes the pressure difference, which incites the movement of the liquid through the entire circulating loop. This movement occurs not only during the winter, but also in summer with sharp air temperature variations. This fact has great significance for thermopiles, since it can lead to an inadmissible increase in temperature at the contact of the pile surface with the surrounding ground. In natural experiments, significant specific thermal fluxes and temperature gradients were measured that confirmed the possibility of summer circulation. This circulation and its role in the formation of the temperature field of the ground have still not been studied in detail. In all the presently known structural designs there is the theoretical possibility of the occurrence of summer circulation with respect to the entire height of the device. Only two structural versions exist that exclude the possibility of occurrence of summer circulation: the single tube device of S. I. Gapayev and the coaxial design developed at the Geocryology Institute.

The general thermal design of the liquid cooling units and the temperature field of the surrounding ground, that is, the conjugate problem of nonstationary heat exchange, can be performed on the basis of known[5] exact analytical solutions. However, it is necessary to note that the exact solutions are extraordinarily unwieldy, and, even when using computers, they will hardly be used in engineering practice. The necessity of averaging these initial data which cannot be given in any exact form eliminates the advantages of precise analytical solutions. It is most expedient to use approximate, but simple, solutions with correction of the results if necessary by natural tests by a special simplified procedure.

The demands of practice and research data indicate that, as a result of the remarkable properties (simplicity, minimum operating expenditures, easy technological manufacture, the possibility of freezing of the ground at great depths, and so on), the liquid cooling units must with time find broad application in transport, hydraulic engineering, power engineering, and industrial-civil construction.

The individual examples of the practical utilization of liquid cooling units convincingly demonstrate their promise.

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PERFORMANCE OF FLEXIBLE FOOTINGS ON THAWING PERMAFROST

A. A. KOLESOV
Fundamentproyekt Institute

When constructing buildings and structures using permafrost in the thawing state, strips and slabs are used as the basic types of foundations. In this case, the design of the flexible footings and also flexible structures (gas pipelines, oil pipelines) must be made, taking into account nonuniform thawing of the frozen ground and nonuniformity in its mechanical properties.

In studying the interaction of flexible strip footings with a nonuniformly thawing base, the author performed special experiments which made use of models of strip footings 30 × 210 cm. The purpose of the experiment was to determine the reaction pressures at the contact between the foundation and the ground and to discover the nature of their distribution as a function of the following factors: nonuniformity of thawing of the permafrost with respect to length of the model foundation, stiffness (EI) of the foundation model, depth of thawing of the ground, and increase in load.

In measuring the reaction pressures of the ground using the experimental setup, individual rigid loading plates were used with dynamometers through which the load was transferred from the test beam (the model of the strip footing). The overall dimensions of the separate rigid plates corresponded to the overall dimensions of the model footing (Figure 1).

The individual rigid plates were installed directly on the frozen ground, and the model of the strip footing was freely supported on the stands on which the dynamometers connected to the plates were mounted.

As a result of the experiments, the relation between the settlement and the reaction pressure of the ground was checked. For uniform thawing soil (uniform distribution of the ice content, massive cryogenous texture, uniform granulometric composition), it was found that between the ground reaction and the thawing depth an inverse proportional relation is observed, that is, with a decrease in thickness of the thawing layer, its resistance increases and vice versa (Figures 2 and 3). This relation is seen between the total settlement and the resistance of the ground, that is, with an increase in the deformation of the ground with respect to the length of the footing, the ground pressure decreases and vice versa.

In order to determine the [effect of] stiffness (EI) of the footing on the distribution of the reaction pressures of the thawing ground, a series of experiments was performed with models of footings of different stiffnesses where the tests were run as follows: (a) on channels No. 30, 2.10 m long with a stiffness $EI = 68 \cdot 10^8$ kg/cm^2; (b) metal strips 30 cm wide and 2.1 m long with $EI = 1.92 \cdot 10^7$ kg/cm^2; (c) strips 2.1 m long made of individual plates with an area 30 × 30 cm not connected to each other.

It has been established that with a decrease in stiffness of the strip footing the reaction diagram of the thawing ground flattens out, and under the strip made of individual plates the reaction becomes uniformly distributed. This indicates the absence of interaction between adjacent footings.

The greatest nonuniformity in the distribution of the ground reaction was observed under the models of footings having the greatest rigidity.

In order to estimate the effect of the variation in thickness of a thawing layer of ground on the reaction distribution of the base with respect to length of the strip footing, two series of experiments were performed.

In the first series, with a general increase in the size of the thaw bowl, the identical shape of the thaw bowl was retained, that is, in different time intervals the outlines of the thawing zones were similar.

In the second, a nonuniform increase in depth of thaw was created, that is, the shape of the thaw bowl in the subsequent experiments had a different form (a steeper drop from the edge to the middle of the footing than in the preceding experiments).

The first series of experiments (see Figure 2) were with nonuniformity in depth of thaw and nonuniformity of the settlement, which were kept approximately constant. In all, four experiments were performed with the following values of the relative thickness of the thawing layer (external load 0.5 kg/cm^2): (1) h/b = 1.3; (2) h/b = 2.6; (3) h/b = 3.4; (4) h/b = 3.9. Thus, the absolute depth of the thaw bowl during the experiments tripled.

It was established that, with an increasing depth of thaw keeping the same shape of the thaw bowl and keeping the nonuniformity constant in the settlement with respect to length of the strip footing, the reaction diagrams of the ground had identical shape in all the experiments (see Figure 2). The maximum depths of thaw correspond to the minimum values of the ground resistance and vice versa, that is, the profile of the resistance diagram for uniformly thawing ground is the inverse of the profile of the thaw bowl. Consequently, with an increase in thawing depth but maintaining the nature of the outline of the thaw bowl, the shape of the curve of ground reaction does not change.

In the second series of experiments, the depth of thaw increased nonuniformly with respect to

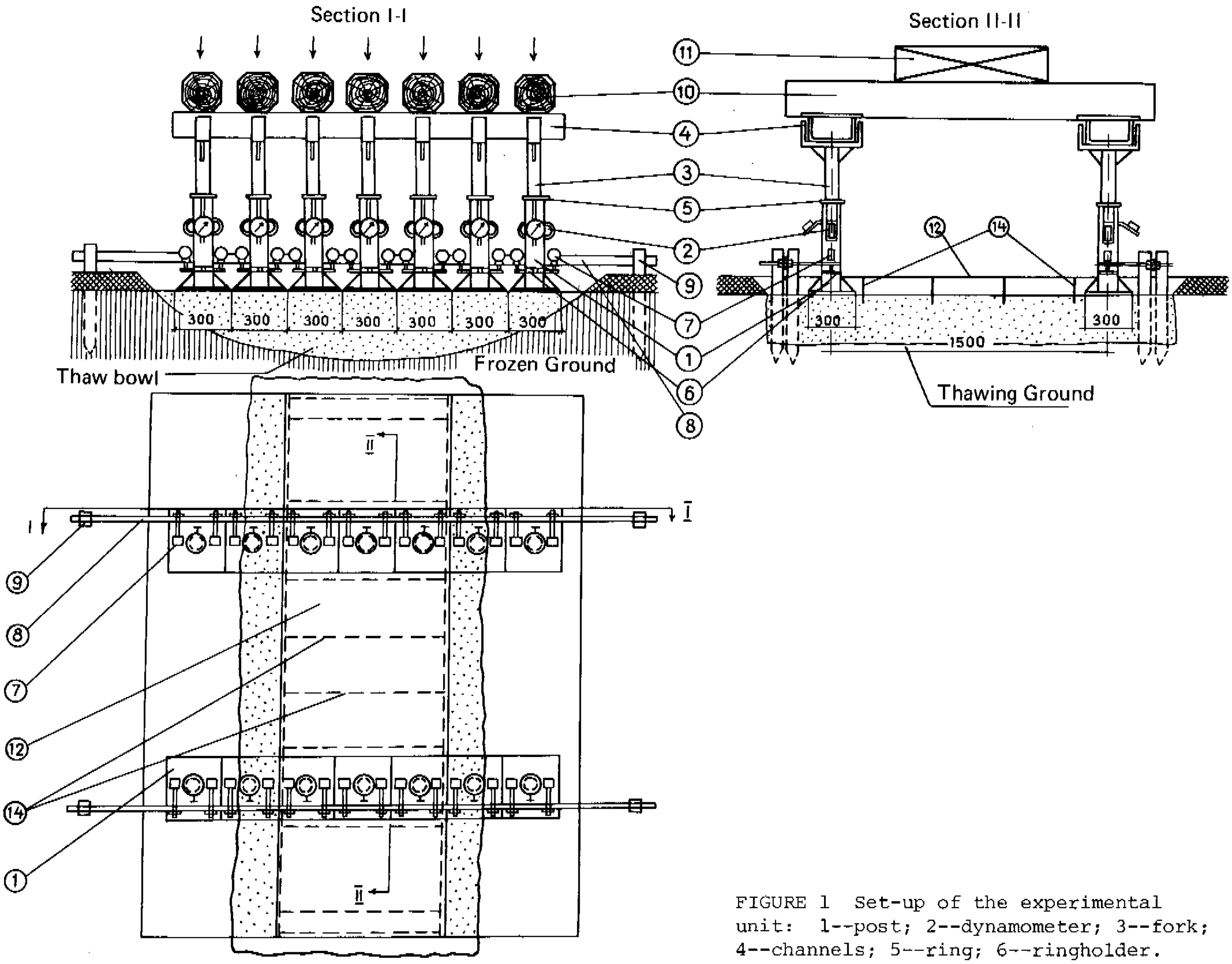

FIGURE 1 Set-up of the experimental unit: 1--post; 2--dynamometer; 3--fork; 4--channels; 5--ring; 6--ringholder.

length of the footing and the ratio of the thickness of the thawing layer under the center to that under the edge of the footing was 1.1, 1.15, and 1.2 for the first, second, and third experiments, respectively. The magnitude of the external load was the same as in the first series of experiments.

The experiments established that with an increase in the ratio h_c/h_k and with an increase in the differential settlement with respect to length of the strip footing correspondingly redistribution of the ground reactions takes place; namely, a decrease in the magnitude of the resistance in the midsection and an increase in its edge values (see Figure 3). The depths of thawing were controlled by using a variable thickness of the thermal insulation on the ground surface.

By analyzing the data of these experiments, it is possible to draw the conclusion that the nature of the distribution of the ground reaction with respect to length of the strip footing for uniformly thawing permafrost depends primarily on the nonuniformity of the thawing, and the shape of the reaction diagram does not depend on the variation in absolute thickness of the thawing layer of the ground.

In order to design the structures, the footings for which will be on nonuniformly thawing permafrost, it is possible to apply the method of Professor B. N. Zhemochkin on the basis of which a curvilinear diagram of the ground reaction is replaced by a stepped straight line, and each section is replaced by a concentrated force applied in the center of the stepped section of the diagram. As a result, we obtain a design method including the forces applied from both sides of the beam. It is possible to make this design by the methods of structural mechanics.

The unit displacements required for design by this procedure will comprise the settlement of the thawing ground at the given point and the deflection of the beam. Its settlement is

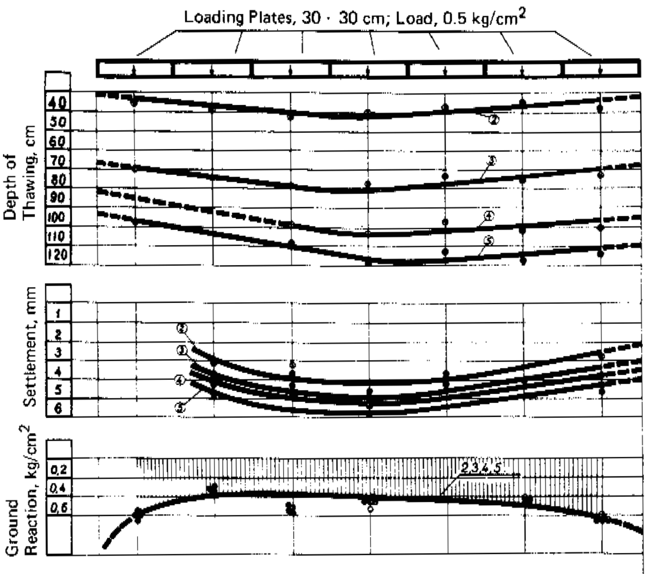

FIGURE 2 Distribution of the ground reaction with a constant h_c/h_k ratio for the thaw bowl (h_c is the depth of thaw under the middle of the strip; h_k is the same under the edge of the strip); 2, 3, 4, 5 are the experiment numbers.

possible only at the point of application of a load, and it has no effect on the settlement at adjacent points, but is propagated only by the deformation of the beam material.

Let us assume the following notation:

y_{ii}, y_{ki} are the total displacements of the beam jointly with the ground under the effect of the force x_i;

δ_{ii}, δ_{ki} are the displacements under the effect of unit forces equal to

$$P = \frac{1}{cb}; \tag{1}$$

Δ_i is the thermal settlement of the thawing layer of the permafrost h_i thick;

ω_{ii}, ω_{ki} are the unit deflections of the beam (the footing) determined by the Maxwell-Mohr method or by tables;

b is the width of the beam;

c is the distance between the columns;

e_{ot}, κ_{ot} are the thawing and yielding factors of the ground respectively.

The displacements y_{11}, y_{22}, $y_{33} \cdots y_{ii}$ are determined from the expression

$$y_{ii} = x_i \delta_{ii} + \Delta_i,$$

where

$$\delta_{ii} = \frac{h_i}{\kappa_{ot} cb} + \frac{c^3}{6EI}\,\omega_{ii} \tag{2}$$

$$\Delta_i = e_{ot} \cdot h_i \tag{3}$$

According to the adopted design method the displacements

$$y_{12}; y_{23}; y_{34} \cdots y_{ki} \tag{4}$$

are defined by

$$y_{ki} = x_i \delta_{ki} \cdots$$

$$\delta_{ki} = \frac{c^3}{6EI}\,\omega_{ki}, \tag{5}$$

For reinforced-concrete beams (footings), this expression assumes the form

$$\delta_{ki} = \frac{c^3}{5.1 \cdot E_c\, I_p}\,\omega_{ki}, \tag{6}$$

where E_c is the Young's modulus of the concrete; I_p is the given moment of inertia of the cross section.

In order to design the footing beam, the system of equations is compiled in which the ground reactions x_i are assumed known and the displacements are defined by Formulas 3 to 6.

For example, the system of equations for a symmetric loading scheme is written in the form:

$$x_1\delta_{11} + x_2\delta_{12} + x_3\delta_{13} + \Delta_1 + \Delta_{1p} + y_0 + d_1\phi_0 = 0,$$

$$x_1\delta_{12} + x_2\delta_{22} + x_3\delta_{23} + \Delta_2 + \Delta_{2p} + y_0 + d_2\phi_0 = 0,$$

$$x_1\delta_{13} + x_2\delta_{32} + x_3\delta_{33} + \Delta_3 + \Delta_{3p} + y_0 + d_3\phi_0 = 0,$$

$$x_1 + x_2 + x_3 - \Sigma_p = 0, \tag{7}$$

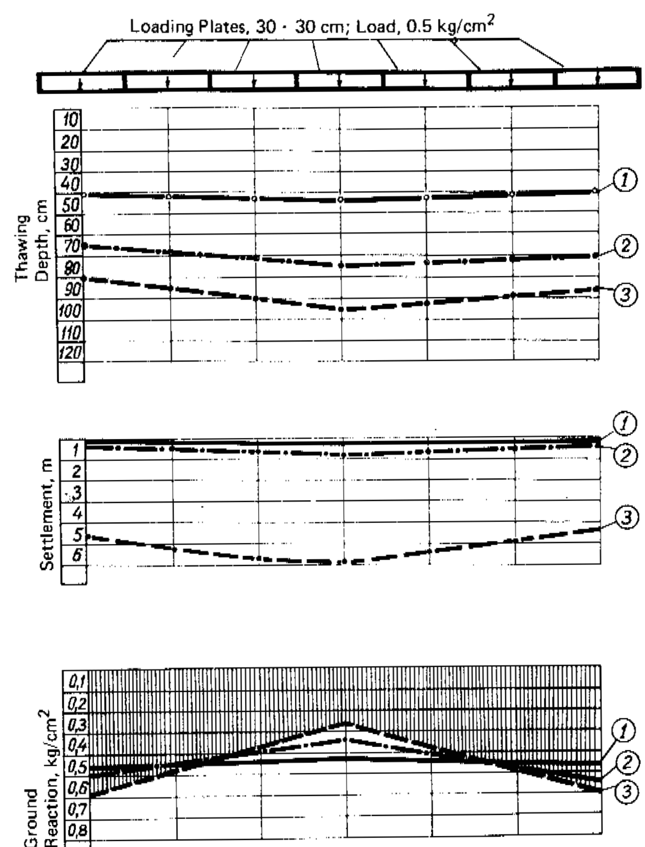

FIGURE 3 Distribution of the ground reaction with an increase in the h_c/h_k ratio for the thaw bowl 1, 2, 3--experiment number.

where x_1, x_2, x_3 are the unknown forces of the ground reaction; y_0 and ϕ_0 are the settlement and the angle of rotation at the point of the assumed positioning of the column, respectively; d_1 is the distance from the point of application of the unit load to the assumed position [of the column]; Δ_{1p} is the displacement in the direction of the external load.

From the solution of system (7), values of the unknowns are obtained after which the diagrams of ground pressure, bending moments, transverse forces and settlement of the planned structural design are defined.

CALCULATING THE SETTLEMENT OF FOUNDATIONS ON PLASTICALLY FROZEN GROUND

A. G. ZATSARNAYA Scientific Research Institute of Bearing Surfaces and Underground Structures

By the projects started in the 1950's and proposed in the following years,[1-3] it was demonstrated that in spite of previously existing ideas permafrost has the capacity to develop volumetric deformations, that is, to consolidate. This is especially clearly exhibited in the case of plastically frozen ground containing a large amount of unfrozen water.

The studies made permitted the introduction of changes into the construction Norms and Guidelines for planning and designing foundations of buildings and structures in permafrost in connection with which in the Norms published in 1966--SNiP II-B.6-66--provision was made for considering the compressibility of permafrost and calculations involving plastically frozen ground by its limiting (stabilized) deformation.

On the basis of the experimental studies of permafrost compressibility, the author discovered the laws of this phenomenon. The studies were made in the underground laboratories of Igarka, Mirnyy, and Yakutsk and in the permafrost mechanics laboratory of the Scientific Research Institute of Foundations and Underground Structures with permafrost of different composition, structure, and temperature. They consisted in tests run on a modified compression apparatus (d = 9 cm and from 1.5 to 6 cm high) using samples of frozen ground under compression without the possibility of their lateral expansion. The distinguishing feature of the tests was keeping a constant negative temperature and protection of the samples from deterioration, primarily from sublimation of the ice and also significant duration of the experiments.

As a result of the research, it was established that the relation between the strains and stresses is clearly nonlinear. The compression curves of the permafrost (with the exception of the highly compressible permafrost) have a sign-variable nature and comprise two basic sections (Figure 1)--convex *CB* and concave *BP*. The outline of the curve within the limits of the first branch indicates the increase in increments of strain with an increase in pressure. This is characteristic of the uniaxial compression diagram. At some pressure p_c the curve changes in sign (curvature) and corresponds to the volumetric deformation diagram. The relation between the strain $\varepsilon = \Delta h/h$ and the pressure p in compression of permafrost without the possibility of lateral expansion is described, as is known, by the expression $\varepsilon = ap$ where, as follows from the experiments, the compressibility factor a is a variable (see Figure 1) that can be expressed by the formulas:

$$a = \frac{B^{-\omega e^{-Bp}}}{\alpha + \beta p} \quad \text{or} \quad a = \frac{1}{\alpha + \beta p}, \qquad (1)$$

where the second of the formulas is the simplified expression of the first ($\omega = 0$). The compressibility factor also depends on the temperature of the frozen ground. This relation can be assumed to have the following form:

$$a = \frac{a_\theta}{(1 + \theta)^n}. \qquad (2)$$

The deformations of the frozen ground under compression are time-dependent. This process can be divided into two stages: the primary consolidation or the migration-viscous deformation connected with compression of the nonfrozen water and secondary consolidation or deformation caused by the creep of the ice and the soil grains ($\varepsilon = \varepsilon_c + \varepsilon_{cr}$). In the first case consolidation ε_c prevailed, and in the second case, the creep deformation ε_{cr}.

As is known, the creep deformations, that is, secondary consolidation, do not depend on the height of the compressed soil sample; therefore, in order to discover the above-indicated types of deformation, tests were run using specimens of different height. The experiments showed that the effect of height is felt up to a defined point in time, after which the relative

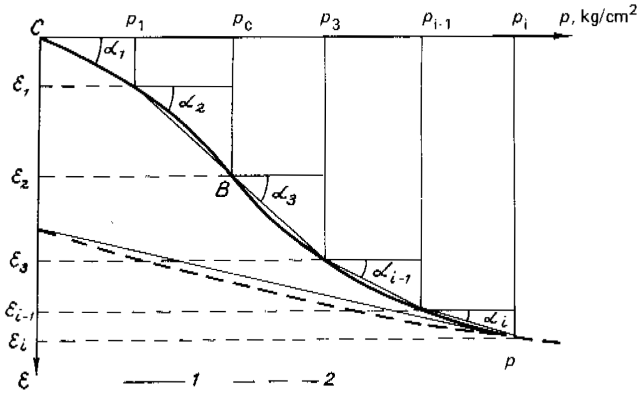

FIGURE 1 Compression curves of plastically frozen ground. 1--loading; 2--unloading.

deformation becomes independent of the height of the specimen. This means that during the initial period, the deformation process has a composite nature, that is, there is primary and secondary consolidation and then the process converts to deformation as a result of creep.

When testing argillaceous plastically frozen soil, the creep deformations during stresses not exceeding the Norm values are from 10-40 percent of the total deformation; the remaining 90-60 percent are in consolidation. The development of the creep deformation with time can be described by the power equation

$$\varepsilon = \varepsilon_H + bt^n. \quad (3)$$

The development of compressive deformations with time has the nature of damped creep. At any point in time, the total deformation ε can be considered as the sum of the elastic deformation ε_{el} and consolidation deformation ε_c; in the latter it is possible to distinguish the structurally reversible part ε_0, that is, the deformation developing with time and gradually recovering after removal of the load and the structurally irreversible ε_1--the deformation not recovered after removal of the load:

$$\varepsilon = \varepsilon_0 + \varepsilon_1. \quad (4)$$

All of the indicated forms of deformations appear during the entire consolidation process. However, in individual stages of this process various mechanisms of deformation are observed, and one form or another of deformation predominates in each of them.

The studies demonstrated that the recovery on unloading can be assumed linear (Figure 1); here, the amount of recoverable strain is 30 to 40 percent, and 70 to 60 percent is irrecoverable.

The essence of the consolidation process can be expressed in the following way. The elastic deformations of the frozen soil take place as a result of elastic compression of all of its structural elements--the mineral particles, ice, unfrozen water and air. The nature of the structurally irreversible deformation still has not been clarified. Obviously, one of the causes of this type of deformation is the irreversible phase transitions of ice into water that take place with an increase in pressure according to the principle of the equilibrium of the solid and liquid phases of the water in the soil.[2] The phase transformations of the water promote an increase in the thickness of the water-colloid film; the elastic compression of these films under load according to the statements of Professor N. Ya. Denisov also causes structurally reversible deformation of the soil.

The physical basis for the irreversible deformations can also be explained by the theory of Professor N. A. Tsytovich on the dynamic equilibrium of the solid and liquid phases of the water in the ground.

In the frozen ground, in contrast to the thawed ground, the load is completely transferred to the mineral skeleton and ice (just as to a solid mineral). Under the effect of load on the contracts of the particles cemented by the ice, excess stresses are created, the dynamic equilibrium between the ice and the film water is disturbed, and part of the ice is converted into unfrozen water. The increase in content of unfrozen water leads to a decrease in the cohesion of the soil and promotes migration of the water and its extrusion from places with significant stresses into less stressed sections.[1] The displacement of the moisture in the liquid phase is realized in the form of film migration, that is, unfrozen water moving under moisture gradients. Under the applied load, migration of ice also takes place--displacement as a result of plastic flow of the ice developing for any, even small, stresses. The compaction of it as a result of forcing out the air in it and compression of this air proceeds simultaneously. As a result of all of these causes there is volumetric compression of the soil, which is stablized after pressing the unfrozen water and air out of the frozen ground. With a further increase in load, the equilibrium between the ice and the film water is again disturbed and the entire process is repeated.

In the case of irreversible deformations, significant variations occur in the ice structure, which were proved by the crystallo-optical studies of specimens of frozen soil having ice lenses. The structure of these lenses underwent large variations: Recrystallization of the ice crystals and variation of their shape and size took place (they decreased by 6-9 times); the melting of the ice was observed at the boundary with the mineral layers.

The compressibility of frozen ground is accompanied not only by the destruction of the structural bonds, but also strengthening of them. Depending on the relation between these two processes in frozen soil, certain deformations take place. With the development of the process of destruction of the bonds, structurally-irreversible deformations prevail, and with the development of the strengthening process, elastic and structurally reversible ones.

When calculating the settlement of foundations, it is necessary to take into account the above-investigated peculiarities of the process of consolidation of permafrost: namely, the non-linearity of deformation with stress and the application of the characteristics of the deformability (a) as a function of load and temperature.

The settlement of individual footings is calculated by the well-known formula:

$$S = \sum_{i=1}^{n} \varepsilon_i h_i, \qquad (5)$$

where ε_i is the relative compression of the i-th layer of the permafrost corresponding to a temperature of θ_i and a pressure p_i in the middle of this layer; h_i is the thickness of the i-th layer of the ground.

When calculating the settlement it is necessary to begin with the fact that the stress distribution under the foundation, under the approximation proposed by Professor S. S. Vyalov, is determined from elasticity theory. The magnitude of the relative compression ε_i for each i-th layer is determined considering the non-linear relation between the deformations and pressures. For any i-th layer it is determined by summing the increments of the deformations according to the variations in magnitude of the deformation by the compression curve (see Figure 1):

$$\varepsilon_i = a_1 p_1 + a_2(p_2 - p_1) + \ldots + a_i(p_i - p_{i-1}) \qquad (6)$$

The values of a_1, $a_2 \ldots a_i$ are determined from the compression curve for the given values of the pressures p_1, $p_2 \ldots p_i$. Correspondingly, compression tests must be performed for various pressure intervals corresponding to the possible diagram of vertical pressures at the base of the structure, and for various values of the negative temperature, which must correspond to the ground temperature distribution diagram with respect to depth at the base of the structure (Figure 2). Here, the diagram of the temperature distribution with respect to depth must be constructed from the data for the maximum mean monthly temperatures. However, for soil in which the mean annual

TABLE 1

Type of Soil	Physical Indexes: Total Moisture Content, w,%	Unfrozen Water w_u, %	Unit Weight, γ, g/cm³	Soil Temperature, θ,°C	Coefficient of Volume Change [mv] a (10^{-4} cm²/kg) in the Pressure Range kg/cm²: 0-1	1-2	2-4	4-6	6-8
Medium grain sand	22	0.0	--	-4.0	1	4	5	10	9
The same	21	0.2	1.99	-0.6	12	9	6	4	3
The same	27	0.1	1.87	-4.2	17	13	10	7	5
The same	27	0.2	1.86	-0.4	32	26	14	8	5
Heavy supes, silty, massive texture	25	5.2	1.90	-3.5	6	14	18	22	23
The same	27	8.0	1.88	-0.4	24	29	26	18	14
Medium suglinok, silty massive texture	35	12.3	1.83	-4.0	8	15	26	28	24
The same	32	17.7	1.84	-0.4	36	42	37	21	14
Medium suglinok, silty, reticular structure	42	11.6	1.71	-3.8	5	10	18	42	32
The same	38	16.1	--	-0.4	56	59	39	24	16
Medium suglinok, silty layered structure	104	11.6	1.36	-3.6	54	54	59	44	34
The same	92	16.1	1.43	-0.4	191	137	74	36	18
Varved clay	36	12.9	1.84	-3.6	15	22	26	23	19
The same	34	27.0	1.87	-0.4	32	30	25	20	16

NOTE: Values of the given Coefficient of Volume Change, a, are given for undisturbed soils (with the exception of sand).

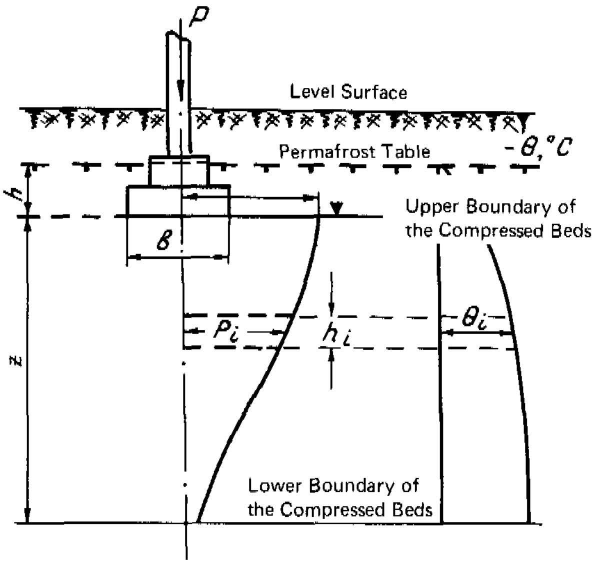

FIGURE 2 Diagram of the pressure and temperature distribution under a footing.

temperature is close to zero, it is possible to take its mean annual value as the design temperature. The experimental values of the parameter *a* for typical frozen soils are presented in the Table 1.

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ADFREEZE BETWEEN GROUND AND FOUNDATION MATERIALS

A. V. SADOVSKIY, Scientific Research Institute of Foundations and Underground Structures

The adfreeze strength is the resistance of the frozen soil to shear along a solid surface and, in particular, the foundation material. In the USSR this research was begun in the 1930's by N. A. Tsytovich[1] and I. S. Vologdina.[2] At the same time, the first data were obtained on the effect of soil composition, moisture content, and temperature on the adfreeze strength. Later, a study of these forces was continued by N. I. Bykov, M. N. Gol'dshteyn, I. D. Dergunov, S. S. Vyalov, K. Ye. Yegerev, B. I. Dalmatov, A. A. Zhigul'skiy, V. O. Orlov, A. V. Sadovskiy, and others.

At the Scientific Research Institute of Foundations and Underground Structures this study is continuing in two areas:

1. The development of methods of changing trends in adfreeze stresses.[3]
2. The study of the nature of the adfreeze stress from the point of view of frozen-soil mechanics and the development of methods of evaluating it for purposes of the most complete consideration of the strength characteristics of permafrost when designing structures. These studies are made by the procedure developed at the Frozen-Soil Mechanics Laboratory at the Scientific Research Institute of Foundations. The basic difference in this procedure is the use of single-shear apparatus permitting determination of the adfreeze force during movement of the frozen soil along the freezing surface and under a simultaneous normal pressure. The specimens are prepared in special molds in which various conditions of freezing of the ground are simulated (one-way and two-way, cooling with subflow and without subflow of water to the freezing front, and so on), which, in turn, offers the possibility of obtaining different cryogenous textures of the frozen soil at the contact with the foundation material.[4]

It is known that the most important factors determining the adfreeze strength are the temperature of the frozen soil, its water content and ice content, the granulometric composition, the freezing conditions, and the physical properties of the solid surface to which the given soil freezes.

The effect of the temperature on the adfreeze strength between ground and concrete was studied in the range from -1° to -18° for three kinds of soil: sand, supes, and suglinok. Here, the soil was saturated. On the basis of more than

400 experiments, it was established that the dependence of adfreeze strength on temperature is described by the expression:

$$\tau\Theta = a + B\ (\Theta)^{m}, \qquad (1)$$

where a, m, and B are the parameters which depend on the physical properties of the soil (the moisture content, the granulometric composition) and the foundation material (the roughness, wettability); Θ is the temperature in degrees below zero.

For the investigated soils Expression (1) assumes the following forms:

$\tau_{\Theta} = 1 + 9.2\ (\Theta)^{0.5}$, kg/cm^2--for sand of medium grain with a moisture content of 20 percent;

$\tau_{\Theta} = 1 + 7.1\ (\Theta)^{0.5}$, kg/cm^2--for supes (40 percent sand and 60 percent suglinok, with moisture content of 28 percent);

$\tau_{\Theta} = 1 + 5.5\ (\Theta)^{0.5}$, kg/cm^2--for silty suglinok with a moisture content of 35-40 percent.

The effect of moisture content on the adfreeze strength was evaluated in the example of sandy soil (moving with respect to the concrete surface). The experimental data indicate that this relation is nonlinear. With an increase in the soil moisture content to a definite limit equal to approximately the maximum moisture content, an increase in adfreeze strength is observed. Exceeding that limit leads to a decrease in strength. Nevertheless, the strength with which the supersaturated soil freezes to the concrete, as a rule, is appreciably higher than the strength with which pure ice freezes to the concrete with rapid shear. Accordingly, the discovery of the effect of the cryogenous texture of the frozen soil on the adfreeze strength with concrete is of great interest. S. T. Tsvetkova[5] and, subsequently, other authors established that on the interface between the frozen soil (bis) and the pile, an ice film is always formed, and the movement of the frozen soil along the foundation surface proceeds along that film. It has been established that the most favorable condition for the formation of the ice contact film is a heat flux directed from the ground to the foundation surface (Table 1).

In practice conditions can be encountered during winter operations when the strongly cooled piling is installed in steam-treated ground or in a borehole filled with liquid mud.

The adfreeze strength is affected not only by the ice film itself, but also by its thickness. For example, for suglinok the greatest adfreeze strength under the conditions of rapid movement corresponds to a thickness of the ice film of 0.5-0.8 mm. A decrease in thickness of the ice film to 0.1-0.2 mm or an increase in it to 2 mm leads to a reduction in the freezing strength by 10-15 percent. At the same time, the strength with which the pure ice freezes to the concrete is 30 percent less than the strength of the soil in the presence of the contact ice zone (a thickness of 1-2 mm). Obviously, a reduction in the adfreeze strength with an increase in thickness of the ice film is explained by an increase in the number of defects in the crystal lattice of the ice. In the presence of a very thin film of ice or in its absence, a decrease in the actual area of freezing of the ground to the solid surface is encountered.

In general, with rapid application of load, the ice film increases the adfreeze strength (see Table 1). However, under the prolonged effect of a load as a result of the plastic properties of the ice the film begins to play the opposite role. S. S. Vyalov[6] established experimentally that the presence of the ice film always leads to a decrease in the long-term adfreeze strength. Thus, the long-term strength of the models of the piles in heavy supes at a temperature of -3.6°C was -4.0 kg/cm^2 (piles driven, no ice film) and 2.75 kg/cm^2 (piles frozen in, ice film along the contact with the soil).

It is known[6] that the shear strength of the pile along the lateral surface essentially depends on the magnitude of the normal (to the pile surface) pressure developed when the ground surrounding the pile freezes, or arising by the lateral reaction when driving the piles into the frozen ground. However, the degree of effect of this pressure was not investigated. The test procedure developed at the Scientific Research Institute of Foundations first permitted evaluation of the freezing strength with simultaneous effect of normal and shearing stresses (Table 2).

TABLE 1 Short-term Shear Strength R_s (kg/cm^2) of the Frozen Soil at the Contact with the Concrete and the Metal with Different Direction of the Heat Flux and Freezing of the Ground (Temperature, -5.5°C)

Form of Contact	Shear Strength for the Direction of the Heat Flux	
	Toward the Material From the Ground	From the Material Toward the Ground
Suglinok-concrete	22.5 (ice film exists)	20.4 (ice film partially)
Suglinok-metal	15.2 (ice film exists)	13.4 (no ice film)

TABLE 2 Short-term Shear Strength of Frozen Soil and Ice along the Contact with Concrete and Metal, kg/cm^2 (Temperature, -5.5°C)

Type of Contact	Normal Pressure, kg/cm^2			
	0	6.25	12.5	18.7
Suglinok-concrete (through the ice film)	19.6	20.5	21.2	22.1
Suglinok-metal (through the ice film)	15.2	16.0	15.8	15.8
Suglinok-metal (without an ice film)	13.4	14.2	14.6	14.3
Ice-concrete	17.8	21.5	24.2	31.6
Ice-metal	11.1	12.8	13.1	13.2

The tests were performed at a temperature of -5.5°C under the conditions of rapid application of a shearing load. The normal load was kept constant for the entire experiment, where a study was made of the adfreeze strength of the soil (Moscow suglinok with a moisture content of 30 percent) and ice with concrete and metal. It was established that the shear strength of the frozen soil or ice along a solid surface, τ_p, is described by the expression

$$\tau_p = \tau_0 + f(\sigma)^n, \qquad (2)$$

where τ_0 is the shear strength in the absence of normal pressure, kg/cm^2; σ is the normal pressure, kg/cm^2; f and n are the parameters determined experimentally, where n depends primarily on the surface roughness.

For systems of the concrete-soil and concrete-ice type, Expression (2) has a linear nature, that is, $n = 1$. With a decrease in surface roughness, the index n becomes less than 1 (Figure 1). The general nature of the expression is retained even under the prolonged effect of a shearing load.

Interesting results were obtained when determining the coefficient of friction of the frozen soil and ice along the concrete and metal surfaces. The experiments were performed on samples for which the ice bonds in the soil or between the ice and the material were first destroyed in advance (by shear). It was found that the friction factor was approximately equal (with the exception of the concrete-ice contact) to the tangent of the angle of internal friction with adfreeze (Figure 1b).

For more complete representation of the adfreeze strength, along with the shear tests, tests were run for separation of the frozen soil from the [solid] material (Table 3). The temporary strength was determined in specimens prepared analogously to those that were tested for shear at the same soil temperature and moisture. The tests were run on test apparatus designed by the author.

Comparing the data of Tables 2 and 3, it is possible to see that the resistance to separation is 25-50 percent higher than the shear strength.

Thus, having data on the resistance of the frozen soil to shear (with and without) a normal load, and separation, it is possible to construct a diagram that represents the complex characteristic of the adfreeze strength--the envelope of the limiting circles of stress.

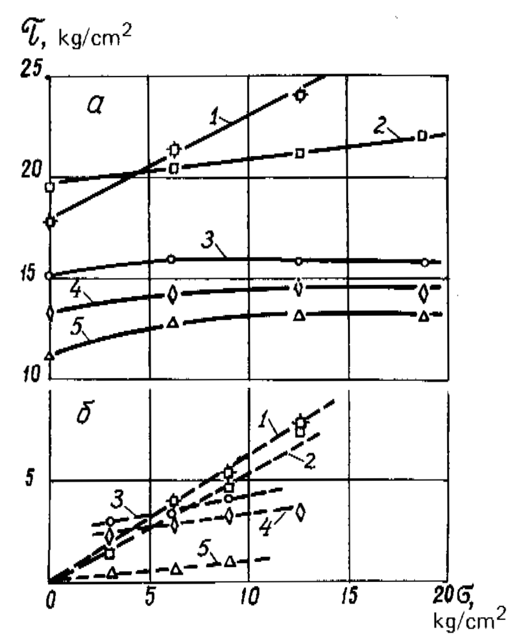

FIGURE 1 Adfreeze strength between soil and foundation material: (a) with adfreezing; (b) without adfreezing (friction). 1--ice-concrete; 2--suglinok-concrete; 3--suglinok-metal (through ice film); 4--suglinok-metal (without ice film); 5--ice-metal.

TABLE 3 Short-term Resistance of the Frozen Soil to Separation from the Concrete or Metal Surface (Temperature, -5.5°C)

Type of Contact	Separation Resistance, kg/cm^2
Suglinok-Metal	7.4
Suglinok-Concrete	14.5

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VALUES OF THE TANGENTIAL STRESSES OF GROUND HEAVE

M. F. KISELEV Scientific Research Institute of Foundations and Underground Structures

When planning the foundations of buildings and structures in heaving ground, it is necessary to design these foundations for the effect of the total force of frost heave for which, in turn, it is necessary to know the values of the tangential stress of heave.

By the total force we mean the force (kg) heaving the foundation when the ground freezes above the footings, and the stress is the heave force per cm^2 of lateral surface of the foundation, found by dividing the heave force by the contact area of freezing.

In the last 30 yr many researchers have determined the tangential forces of frost heave under natural conditions. Some have also performed the determinations by successive application of load to the foundation, preventing heave (Table 1). Others have determined the heave force by a dynamometric method (Table 2).

As is obvious, the data obtained by the two methods differ somewhat. It is possible that both the freezing conditions and the differences in the procedure are felt. However, in all cases it is possible to draw some general conclusions.

First, it is possible to note a trend toward a reduction in the [rate of] increase of heave stress with an increase in depth of freezing after achieving a limiting value of this depth. As a rule, the greatest values of the heave stress are observed when the ground freezes to a depth of 2 m. With further freezing these forces cease to increase, and sometimes they even de-

TABLE 1 Data from Determining the Tangential Heave Forces by the Method of Applying a Load to the Foundation under Field Conditions

Author of the Research	Ground	Depth of Freezing, cm	Tangential Heaving Force		Where the Experiments Were Performed
			kg/cm^2	kg/cm	
N. I. Bykov (1940)	suglinok	250	0.48	120	Skovorodino
I. D. Dergunov (1945)	supes	--	0.2	--	Chitinskaya oblast
I. D. Belokrylov (1944)	suglinok soils	--	0.5-1.5	--	Petrovsk-Zabaykal'sk
B. I. Dalmatov (1957)	suglinok	150	0.97-11.15	145-173	Cherevobets of Vologskaya oblast
V. P. Ushkalov (1963)	suglinok,	280	0.4-0.56	113-157	Petrovsk-Zabaykal'sk
	silty sand	280	0.35	60-98	
V. B. Shvets (1965)	clay soil	150	0.53	80	Sverdlovsk oblast, Urals
B. I. Kalmatov, V. S. Lastochkin, V. M. Ulitskiy (1966)	suglinok	215	1.0	220	Irkutsk oblast
M. S. Uspenskiy (1966)	suglinok	150	1.0	--	Usol'ye-Sibirskoye
B. G. Petrov (1970)	suglinok	150-158	1.0-1.2	158-159	Zagorsk station in Moscow oblast

crease. All of this is confirmed by observations both of the dynamics of the development of the heave forces and the magnitude of the heave of the ground surface during the freezing process.

Then, we consider that the magnitude of the tangential heave stress does not depend on the climatic conditions. Of course, these conditions have a significant effect on the depth of freezing of the soil, consequently, the total heave force. However, the force per unit area of freezing, that is, the specific heave stress is not directly affected by the climatic factor.

At the same time, the magnitude of the heave stress depends directly on the type of ground, its natural moisture content, and the hydrogeological conditions. The degree of heave of the ground also depends on the same factors. Therefore, it appears expedient to adopt (independently of the climatic conditions) the degree of heaving of the ground as the criterion for evaluating the tangential heave stress, determining this from engineering-geological exploration.

We propose subdividing the soil types with respect to their degree of frost heave into severe heaving, medium heaving, weakly heaving, and provisionally nonheaving. It is possible to assign the soil to one category or another with respect to type, consistency, and moisture conditions. Correspondingly, it is proposed that the value of the heave stresses be differentiated: for strongly heaving ground this value is 1 kg/cm^2; for medium heaving, 0.8 kg/cm^2; and for weakly heaving, 0.6 kg/cm^2.

The recommended values of the heave stresses are obtained on the basis of analysis and generalization of the above-presented experimental data. The author used the mean and not the extreme values.

Confirmation of the fact of a decrease in magnitude of the tangential heave stresses for medium, and weakly, heaving ground can be obtained from the experimental data of N. A. Tsytovich (1937) and I. D. Belokrylov (1944). The former performed studies on the dependence of the adfreeze strength on the type of soil and its moisture content under laboratory conditions, and the latter studied the same relations under field conditions in the permafrost region.

The classification of soil with respect to

TABLE 2 Data from Determining the Tangential Heave Forces under Field Conditions by the Dynamometric Method

Author of the Research	Ground	Depth of Freezing, cm	Tangential Heaving Force		Where the Experiments Were Performed
			kg/cm^2	kg/cm	
S. S. Vyalov and N. I. Yegorov (1958)	suglinok	150	0.5-1.24	--	Igarka
K. Ye. Yegerev (1960)	suglinok and supes	150	0.6-0.83	90-124	Yakutsk
V. O. Orlov (1962)	suglinok	150-300	1.03-1.36	146.7-168	Igarka
A. B. Patoleyev, G. S. Alayev (1965)	suglinok	150-190	0.88-0.7	132	Khabarovsk
N. A. Peretrukhin (1967)	suglinok	175-200	0.6-1.69	100.6-338	Skovorodino
A. M. Pchelintsev (1969)	suglinok	88-130	1.19-2.11	--	Igarka
O. S. Konnova and A. V. Sadovskiy (1970)	suglinok	127	1.7-1.82	--	Vorkuta

TABLE 3 Subdivision of Soils with Respect to Degree of Heave and Tangential Heave Stress, τ

Degree of Heaving of the Soils and Consistency B^a	Types of soil, position of the water table, Z					τ kg/cm²
	Fine Sand	Silty Sand	Supes	Suglinok	Clay	
Strongly heaving at $B > 0.5$	--	--	$Z \leq 0.5$	$Z \leq 1.0$	$Z \leq 1.5$	1
Medium heaving at $0.25 < B \leq 0.5$	--	$Z \leq 0.6$	$0.5 < Z \leq 1$	$1 < Z \leq 1.5$	$1.5 < Z \leq 2$	0.8
Weakly heaving $0 \leq B < 0.25$	$Z \leq 0.5$	$0.6 < Z \leq 1$	$1 < Z \leq 1.5$	$1.5 < Z \leq 2$	$2 < Z \leq 3$	0.6
Conditionally not heaving at $B < 0$	$Z \geq 1$	$Z > 1$	$Z > 1.5$	$Z > 2.5$	$Z > 3$	--

NOTES: 1. The classification of the soil with respect to degree of heave is determined by one of two indexes: B or Z; 2. B is the consistency of clay soil according to SNiP II B.1-62. It is based on the natural moisture content of the soil in the seasonally thawing layer as the weighted mean value. The moisture content of the soil at a depth from 0 to 0.5 m is not used in the calculation; 3. Z is the difference between the depth of the groundwater table and the calculated depth of freezing of the soil. It is determined by the formula:

$$Z = H_0 - H,$$

where H_0 is the distance from the water table, m; H is the calculated depth of freezing of the soil, (SNiP II-B. I-62).

[a] [B is the Liquidity Index, $I_L = \frac{W - W_p}{I_p}$]

degree of heave and the recommended values of the heave stresses are presented in Table 3.

Table 3 permits the selection of the design values of the heave stresses on the basis of the simplest physical characteristics of the soil and the hydrogeological conditions. It is reasonable that in these conditions it will be necessary to check and more precisely define the recommended data on the basis of special experimental studies.

THERMAL INTERACTION OF PERMAFROST AND UNDERGROUND STRUCTURES

A. F. ZIL'BERBORD, YU. L. SHUR VSEGINGEO

Recently more and more attention has been given to the design and construction of underground structures for various purposes. The storage of petroleum products, natural and liquid gas, food products, industrial wares and products, equipment, documents, archives, medicines and the underground location of scientific research laboratories, electric power plants and garages--this is far from a complete list of examples of the utilization of underground space.

The primary advantages of underground structures is their low cost as a result of using the ground in its natural state as construction and thermal insulation materials, simplicity of ensuring various thermal moisture regimes, protection from external effects, and the possibility of utilizing natural refrigerating resources.

With respect to thermal regime, underground structures can be divided into structures with negative and positive temperatures.

For structures with negative temperatures, two versions of the thermal regime required by the technological conditions are possible:

1. The air and temperature inside the facility must be negative without specifying the magnitude of this temperature.

When designing such structures, it is necessary to consider the heat released inside the structure and the heat entering during air exchange of the structure with the atmosphere and also the variation in heat-exchange conditions at the surface of the ground in the zone of development. For a significant part of the region with permafrost, the heat generated by the structure can be absorbed by the frozen mass without thawing.

2. The air temperature inside the facilities must be lower than the natural temperature of the frozen beds (for refrigerators -20° to -30°; for storing liquid hydrocarbon gases, below -40°).

To ensure the regime of the technological process, three procedures are possible: winter cooling of the structure and the ground by natural cold air, mechanical cooling, and a combination of natural and mechanical cooling.

In order to select the technical media providing for the given temperature regime in the underground structure, it is necessary to have at our disposal data on the temperature field of the surrounding ground for calculating the heat fluxes to the surface of the structure.

For structures with a positive temperature, the determination of the thickness of the thawing layer has special significance in connection with the necessity for calculating the settlement during thawing and taking it into account when selecting the structural design of the structure and its location.

When planning and designing underground structures with given negative air temperature, it is necessary to determine the following: the possibility of utilizing natural cold to insure the required temperature regimes, the possibility of satisfying the technological process requirements as a result of cold accumulated by the ground, and the operating regime of the refrigerators and their duration.

The problem of a natural regime for the underground structures with negative temperature has no general solution, so numerical methods and a hydro-integrator were used.

The initial ground temperature around the structure was taken on the basis of the geothermal observations. The temperature at the ground surface was given by the average monthly temperature of the ambient air. At the lower boundary of the investigated part of the ground, the temperature was assumed constant and equal to the natural temperature at the given depth. On the surface of the underground structure during the winter months (the cooling period), a temperature was taken as 5° above the temperature of the outside air. During the warm part of the year three boundary conditions were replaced by two on the ground surface and beyond the limits of the zone of thermal effect of the structure. This offers the possibility of recording the variation of the temperature field as a result of heat exchange between the structure and its surroundings. In the thermophysical calculations, the temperature of -18° was taken as the maximum admissible temperature.

The calculations demonstrated that even under the most favorable conditions (Yakutsk), characterized by 130,000 negative degree-hours, the given temperature is reached only during the second year of operation of the structure. Here, in connection with the effect of the uncooled massif and the temperature fluctuations of the atmospheric air, the temperature in the underground structure varies with time. The amplitudes of the temperature fluctuations were: for the first year of operation, 15.7°C; the second year, 12.4°C; third year, 10°C; and the fourth year, 8.4°C. The maximum temperatures in the underground structure were -14°C, -18°C, -20.5°C, and -23°C, respectively.

The cold accumulated in the permafrost can be expended on the technological requirements beginning with the fourth to the fifth year of operation of the structure. Here, it is necessary to consider the regular discontinuities in reusing the natural cold required to restore the temperature in the surface layer of the ground as a result of heat discharge to the more remote parts of the massif.

Thus, for example, in the fifth year of operation, about 40,000 kcal/m^2 of roof and floor of the structure can be expended. This means that in a mine 10 m wide and 30 m long about 24 × 10^6 kcal of accumulated cold can be expended. This amount is sufficient to freeze and cool 200 tonnes of meat to -20°C. However, in the next year the admissible annual expenditure of cold decreases by 22 percent, and after another year, by 27 percent. This indicates the limited possibilities of using natural cold.

The situation is quite different in regions with a milder climate, where the number of negative degree-hours decreases to 60,000 to 70,000. In these regions even intense ventilation during the cold part of the year cannot ensure the given temperature in the underground structure. Therefore, in practice, during the entire period of operation, it is necessary to use additional artificial cold. The output capacity of the refrigeration unit can be low, since the heat-flux density from the ground massif stabilized by 5 to 6 yr of maintenance of the structure is about 0.15 kcal/m^2·h at a permafrost temperature of -2°C and about 0.3 kcal/m^2·h at a temperature above -1°C. In the same region the thermal flux density through the enclosure of the ground refrigerator reaches 10 to 12 kcal/m^2·h.

Thus, in practice in any underground structure where there are sources of heat and stable low temperatures required, refrigeration must be provided. The use of natural cold will promote a reduction in the operating time.

Underground structures with a positive temperature inside can be constructed both with conservation of the soil in the frozen state and with thawing of it during operation. However, in contrast to surface structures, in underground structures the first of the mentioned procedures requires the use of complicated, expensive devices. Therefore, in the majority of cases it is necessary to be oriented by adaptation of the structure to thaw-settlement.

In solving the non-steady-state problem of thawing, the method of successive replacement of steady states is used, which permits use of the theory of steady-state heat flow. With this application, exact and approximate solutions were obtained for structures of any configuration with different temperatures of the permafrost at different thermophysical characteristics of the frozen and thawed ground. With an exact solution of the problem, the thickness of the thawed layer under the structure with rectangular transverse cross section was determined; with approximate solution the rectangle was replaced by an ellipse with the same perimeter and a ratio of the axes equal to the ratio of the sides of the rectangle.

The method of conformal mapping was used to obtain the expression to determine the dimensions of the thawed layer. The solutions were integrated numerically, and graphs were compiled for practical calculations. An analysis of the solutions demonstrated complete justification of the replacement of the rectangular transverse cross section by the elliptic one for which the solution of the problem is much simpler, and the results are more convenient for practical use.

The solutions were obtained analytically for buried structures at the surface and for underground structures with great depth when the effect of the surface need not be considered.*

For structures at great depth, the heat discharge into the frozen zone is taken into account by artificial restriction of the zone of thermal effect of the structure. The ratio of the radius of the effect to the depth of thawing depends on the temperature in the structures, the depth of thawing, and the ratio of the sides of the structure. With an increase in depth of thawing, the ratio of the radius of effect to the depth of thawing approaches a constant for specific conditions. On the basis of the solutions obtained, an analysis was performed of the effect of various factors on the depth of thawing of the ground. It was discovered that the thawing process takes place intensely for the entire time of operation of the structure, and in the majority of cases the depth of thawing does not reach values close to steady state. An increase in width of the structure is accompanied by an increase in the thawing layer. Its thickness is affected to a lesser degree by variation in height of the structure. These factors must be considered when selecting three-dimensional solutions during the planning and design process.

The presence of a thermally insulating layer decreases the depth of thawing by the amount equal to the thickness of the equivalent layer. However, the most effective means of regulating the thawing intensity of permafrost is by the temperature inside the structure.

*Shur, Yu. L Calculating the depth of thawing of frozen soils in the bearing ground of underground structures. Voprosy-Inzhenernoy Geokriologii (Problems of Engineering Geocryology), no. 22. Moscow, VESGINGEO Institute, 1969.

STABILITY OF THE SIDES OF OPEN PITS IN PERMAFROST

G. R. GLOZMAN AND I. I. YERMAKOV VNIMI Institute, AND
N. G. BOBOV VSEGINGEO Institute

The strength and stability of permafrost in cuts and also in the sides of open pits as a whole depends essentially on the temperature regime of the ground. The highest temperature of the permafrost is characteristic of the hypsometrically lowest benches. This is promoted by snow drifting, which increases with depth, and flooding of the pits. In addition, when working the mine, deeper and deeper horizons are opened up, the natural temperatures of which increase.[1] The thawing of the rock in the spring-summer period has an effect primarily on the stability of the cuts. The deep seasonal thawing in the benches causes slides of individual blocks of rock with respect to thawed ice veins.

A characteristic feature of the mass of permafrost rock and semirock is that the cracks in it are, as a rule, filled with ice or a fine-grained ice-saturated material. Therefore, the porous individual blocks are cemented into a sort of monolith. Also, the strength of the permafrost in the monolith is greater than the strength of this rock in the thawed state.

The strength tests run on rock and semirock in a monolith show that their shear strength in the frozen state increases by comparison with the shear strength in the thawed state, primarily only as a result of an increase in cohesion. The angle of internal friction of the frozen rock is equal to the angle of internal friction of the thawed rock.

For conversion from the strength of thawed rock and semirock in a monolith to the strength of the rock and semirock in the frozen state on the basis of field and laboratory tests, the following relation is proposed:

$$\sigma_M = \sigma_T(1 + aw), \qquad (1)$$

where σ_M is the compressive strength in the frozen state, kg/cm^2; σ_T is the compressive strength in the thawed state, kg/cm^2; w is the moisture content, in percent; and a is a coefficient that depends on the temperature of the sample.

The long-term compressive strength of the frozen rock and semirock is 0.6-0.8 of the short-term compressive strength.

The determination of the strength characteristics of permafrost with respect to the contacts and in the massif has great significance, since these characteristics are used in calculating the stability of cuts.

The short-term (10 min) strength tests on rock contacts, contacts of rock with ice, and ice-saturated crack fillers also indicate an increase in the shear strength by comparison with the unfrozen contacts, primarily as a result of an increase in cohesion in the frozen state. At the contact of rock with ice and the frozen soil, the cohesive forces are 3 to 6.5 times greater than for similar unfrozen contacts.

The results of the studies of the long-term strength (for $T = -3.5°C$) along the contacts (rock-ice-rock) show that

$$c'_{1t} \simeq c_{10}'; \tan\phi'_{1t} \simeq 0.6 \tan\phi'_{10};$$

where c'_{1t} and ϕ'_{1t} are the strength characteristics (cohesion and friction angle) along the contacts with the interlayers of ice during long-term tests; c'_{10} and ϕ'_{10} are the strength characteristics with respect to the same contacts for 10-min tests.

An increase in strength of the frozen massif by comparison with the thawed one takes place both as a result of freezing of the blocks together along the contacts and as a result of an increase in the strength of the permafrost in the monolith. The strength of frozen marls and strongly fissured limestones is 3 to 4.5 times greater than the strength of the same rock in its thawed state at the same time as the strength of this rock in the frozen monolith is 1.5 to 3 times higher than in the thawed monolith. Thus, the increase in strength of the massif in the case of freezing is 30-40 percent as a result of an increase in strength in the monolith and 60-70 percent as a result of freezing of the blocks together along the contacts.

An analysis of the data from the natural tests permits extension of the known G. L. Fisenko formula for determining the cohesion in the massif[2] for permafrost by introducing the coefficient taking into account adfreezing of the blocks along their contacts, as a result of which there is an increase in the size of the structural module:

$$c_M = \frac{c_k}{1 + a\ln\frac{H}{l} - ab}, \qquad (2)$$

where c_M and c_k are the cohesion in the massifs and in the monolity in the frozen state during short-term testing; a is the coefficient that depends on the strength of the soil in monolithic form and the nature of the jointing in thawed form; H/l is the ratio of the side height to the mean size of the elementary modules in the thawed state; and b is a coefficient taking into account the freezing together of the blocks.

The cohesion of the permafrost in the massif during prolonged application of load is 0.6 to 0.8 of the cohesion of the soil in the massif with short-term application of the load.

Being exposed in the cuts, the permafrost thaws in the spring-summer period within the limits of a layer of defined thickness. During thawing, the cryogenous bonds in the rock are disturbed, which causes a decrease in strength. As a result of weathering, frequently deepened by repeated cycles of thawing and freezing, the rock erodes off the cuts, as a result of which new layers are exposed that have not been subjected to weathering. Thus, the cuts made up of strong rock in the frozen state are severely flattened out.

For cuts in permafrost, as a rule, nonparallel recession of the surface of the cut is characteristic. This is explained by the special thermodynamic conditions at the edges of the cut.

The observations of the smoothed-out cuts as a result of weathering in open mine pits under conditions of permafrost has permitted us to obtain the following relation for calculating the angles of the cuts for defined times of standing:

$$\alpha_t = (\alpha_0 - \phi)\exp(-bt^n) + \phi, \qquad (3)$$

where α_t is the cut angle (above the upper point of the slope) at time t; α_0 is the cut angle at the time of working; b is a coefficient that depends on the weathering rate of the rock (the rate of smoothing of the cut angle); t is the number of years (spring-summer seasons) from the time of working the bench; n is a coefficient that depends on the rock strength in the thawed massifs; ϕ is the angle of the natural slope.

Knowing the slope angle (α_t) at any point in time from the time of the working; it is possible to derive the function permitting determination of the magnitude of the crumbling (Δl_t) of the upper brow of the berm:[3]

$$\Delta l_t = \frac{h(\cot\alpha_t - \cot\alpha_0)}{\sqrt{A}\mu + 1}, \qquad (4)$$

where h is the height of the bench, m; μ is the disintegration factor that is the ratio of the volume of the crumbled rock to its volume when whole;

$$A = \frac{\sin\phi\cdot\sin(\alpha_0 - \alpha_t)}{\sin(\alpha_0 - \phi)\cdot\sin\alpha_t}. \qquad (5)$$

When $\alpha_t = \phi$, that is, in the case of smoothing of the slope angle of the bench to the angle of the natural slope, $A = 1$ and Formula (4) acquires the following form:

$$\Delta l_\phi = \frac{h(\cot\phi - \cot\alpha_0)}{\sqrt{\mu} + 1}. \qquad (6)$$

Considering the comparatively low strength of the thawing layer, the calculations of the stability of the sides of open-pit mines in permafrost must be carried out with respect to the long-term strength characteristics of the rock in the massif in the frozen state with a factor of safety of 1.1 to 1.2 (considering a reduction in strength of the rock of the frozen massif with time).

The calculations of the stability of the sides of the open-pit mines in permafrost (in particular, the sides of the open-pit mines of the Kimberlite pipes at Yakutsk) give high values of the limiting slope angles. In addition, the slope angles of the benches which by the strength characteristics of the rock can assume high values are intensely smoothed with time. Thus, the stability characteristic of the sides of open pits in permafrost is the relation of the angle of inclination of the side stability of the slopes subjected to intense disintegration with time.

Therefore, the stable angles of the slopes of the cuts subjected to intense crumbling must be established beginning with the inclination of the rock toward crumbling and not by the limiting equilibrium. Here, the slope angles of the sides and the limiting position are established according to the angles of the bench slopes and the required works of the berms, and they are checked by the calculations for stability with respect to the schematics taking into account the geological structure of the massif.[4]

In order to decrease the volumes of the rock and semirock stripping when setting up open-pit mines in permafrost, the possibility of giving the sides of the pits high slopes will be used. For this purpose it is necessary to take both preventive measures to prevent the deformation of the berms and the bench slopes and to reinforce the most intensely crumbling and disturbed sections of the benches. In addition, the following recommendations are made:

1. Place the benches at the limiting position with the use of a special procedure (treatment of the initial contour of the objective with preliminary crack-formation).
2. Use of thermal insulation and reinforcing coverings and also the application of other methods of reinforcing the slopes (rods, piles).
3. The construction of drainage ditches and guide berms.

When selecting the method of reinforcing and the materials for doing the reinforcing it is necessary to begin with a valid estimate of the geocryologic conditions and existing methods of controlling the process of thawing and freezing of the rock.

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EFFECT OF CRYOGENOUS PROCESSES ON THE STABILITY OF HIGH TAILINGS BANKS IN POLAR OPEN-PIT MINES

G. V. KALABIN, E. B. KRASNOSEL'SKIY, B. K. OVODENKO, AND V. I. USYNIN
Mining and Metallurgical Institute of the Kola Branch of the USSR Academy of Sciences

The tailings dumped on the slopes in polar regions usually freeze, and therefore the problem of their stability is connected with studying geocryological processes in the frozen tailings. This problem has been encountered at the Central Mine of the Apatite Combine, where the height of the tailings reaches 180-300 m and the rock is dumped to the onset of critical deformations, when the work on the tailings is curtailed and then it is knocked down or slowly creeps to the more gently sloping part of the embankment (Figure 1).

The stability of the tailings is significantly affected by the temperature regime of the tail-

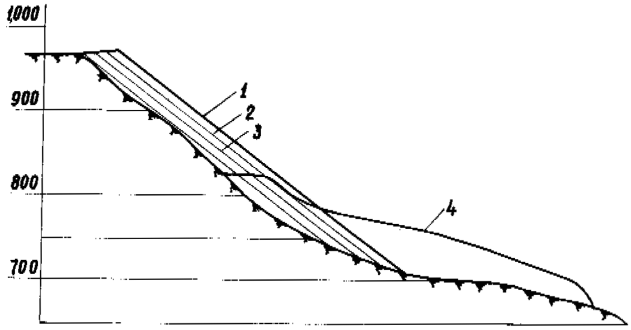

FIGURE 1 Longitudinal section through tailings no. 9 during various periods of dumping to collapse (9 August 1970) and after collapse. 1--9 July 1970; 2--12 February 1970; 3--27 February 1970; 4--12 November 1970.

ings rock, which varies during the year. In order to study the temperature regime of the rock in the tailings, special observations were made for which thermal sensors were installed in boreholes drilled in the tailings and the bedrock. However, among the problems here is to find the depth of penetration of the heat waves, both into the body of the tailings and into the bedrock.

For bedrock (Figure 2a) the amplitude of the extreme temperatures from 10 m and deeper becomes stable and does not exceed 0.9°C, with a mean monthly value of 0.5°C (with a predominant value of 0.2°C). This permits the proposition that for the mining conditions at Khibin, the thickness of the layer with seasonal fluctuations in temperature on the slopes with southern exposure does not exceed 10 m. The observations have established the nature of the rise (Figure 2b) and fall (Figure 2c) of the temperature of the tailings. The depth of penetration of temperature fluctuations during the spring-summer and fall-winter periods is different. During the spring-summer period, this depth exceeds 15 m, and the amplitude of the temperature fluctuations at a depth of 4 m is 0.8° to 1.4°. This is explained by the fact that in the spring-summer period, as a result of thaw water and rainwater getting into the tailings, heating takes place to a depth of more than 15 m. In the winter the entry of water into the tailings stops, and heat transfer is predominantly by convection.

During the process of freezing of the tailings, the cryogenous texture is formed; this is characterized by interlayers and lenses of ice formed both as a result of moisture (up to 15-20 percent) in the rock during the summer from rainwater and the migration of water from the lower horizons to the freezing front and as a result of falling of snow during the process of winter dumping, packing of the snow under the weight of the overburden, and conversion to firn and ice. As a result of freezing during the winter, the strength of the tailings increases. In the summer and fall the tailings are saturated with rainwater, which can percolate to the base as a result of the high porosity of the tailings and under certain conditions can cause thawing of the entire layer of frozen rock dumped during the winter and a sharp decrease in strength of the thawing layer.

An analysis of the dependence of the displacement rate of the tailings down the slope on the air temperatures (Figure 2d) and temperatures in the tailings [Figure 2(3)] demonstrated that, even with an increase in the tailings in

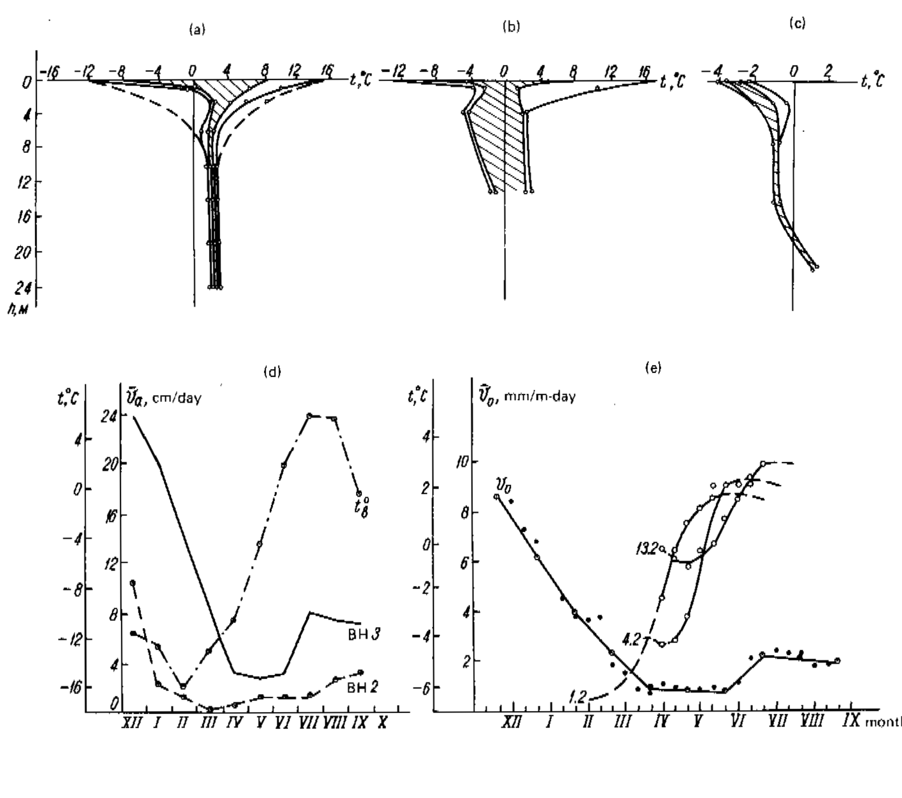

FIGURE 2 Temperature regime and displacement rate of tailings downslope: (a) seasonal temperature fluctuations in bedrock; (b) the same in the tailings (heating); (c) the same in the tailings (cooling); (d) absolute rate of displacement of the tailings along the slope (9*v*) and variation in air temperature (*t*); (e) relative displacement of the tailings downslope and the temperature variation at depths of 1.2 m, 4.2 m, and 13.2 m. The crosshatched section is the range of variation of the mean monthly temperatures; the solid line is the actual data; the dotted line is the proposed temperature curve.

December-March, the deformation rate strongly decreased for air temperature below -9°C. The period of the steady-state deformation rate was observed with negative but rising air temperatures from -9°C to 0°C. An increase in the deformation rate was noted approximately 10 days after transition of the mean monthly air temperatures and temperature in the tailings at a depth of 13 m through zero. In Figure 2d we have the successive heating and upper temperature limit in the tailings.

The observations of the tailing deformations near the adjacent thermal wells located on the tailings site at different distances from the edge demonstrated their successive pulsation during the heating of the rock in the tailings. This is connected with the fact that the layer of greatest loosening of the tailings mass is at the level of the zero isotherm and changes position as this isotherm shifts, leading to a wave nature in the deformations in the body of the tailings. Under the effect of load from the weight of the overburden, the loosened layer is compacted, leaving a secondary zone of loosening. Then the process repeats. This leads to wave development of the deformation. The magnitude of this deformation depends on the amount of snow in the tailings and the temperature of the outside air during the dumping of the tailings.

In order to calculate the stability of the tailings made up of frozen and thawed rock, it is necessary to know their strength characteristics as a function of the temperature regime. The laboratory studies performed at the Institute of Foundations and Underground Structures (N. K. Pekarskaya and G. I. Bondarenko) established that the frozen tailings deform as a plastoviscous body.

During the tests, significant creep was observed with respect to magnitude and also the absence of a sharp transition from a uniform rate of flow to accelerated deformation. Accordingly, it is necessary to consider the development of significant deformations during creep (without destroying the continuity of the rock) when predicting the time of occurrence of critical deformations and collapse of the tailings.

For tailings comprising frozen and snow-covered rock, deformations leading to collapse can occur in the subcritical state as a result of significant deformations of the rock developing with time in the frozen state.

The development of the deformations in the tailings according to the survey data corresponds to deformation of frozen soil. After plastic deformations, the process of progressive flow begins, which ends with complete collapse, destruction of the internal bonds, and collapse of the tailings in a slide. Thus, with respect to the nature of the deformations of the frozen soil obtained in laboratory conditions and the deformations of the tailings obtained using the survey observations, the possibility appears of forecasting the state of the tailings and the period of onset of critical deformations.

THERMAL REGIME OF THE BASE AND CORE OF THE DAM AT THE VILYUY HYDROELECTRIC POWER PLANT*

R. M. KAMENSKIY Geocryology Institute of the Siberian Department of the USSR Academy of Sciences

In regions with severe climate and continuous permafrost, the type of dam in a hydroengineering complex, the structural peculiarities of its individual elements, including the impervious elements, and the joining of the body of the dam to the base and the shoulders are determined to a significant degree by the thermal regime of the structure.

In turn, the formation of the thermal regime of the base and the body of the dam is affected by many factors. These include the shape and size of the structure, the structural design of the lower prism and the impervious element, the type of material, the heat-exchange conditions of the daylight surface of the dam with the atmosphere, and the thermal effect of the reservoir and tailrace. The heat-exchange processes are so complex that a reliable prediction of the thermal regime of the structure is difficult, in the planning and design state, especially since there are insufficient experience and observational data for such structures under the specific conditions. In this connection it must be noted that the dam of the Vilyuy Hydroelectric Power Plant was designed as a thaw-type dam, but there was danger that part of the body of the dam on the lower support prism side could periodically be subjected to freezing and thawing during its operation.

The structural peculiarities of the dam at the Vilyuy Hydroelectric Power Plant and the conditions of its erection have been investigated in quite some detail in a number of papers.[1,2] Let us only note that, to prevent percolation in the base of the dam during thawing of the fissured rock, the connection of the suglinok core to the

*[Refer to p. 594 for a description of the dam.]

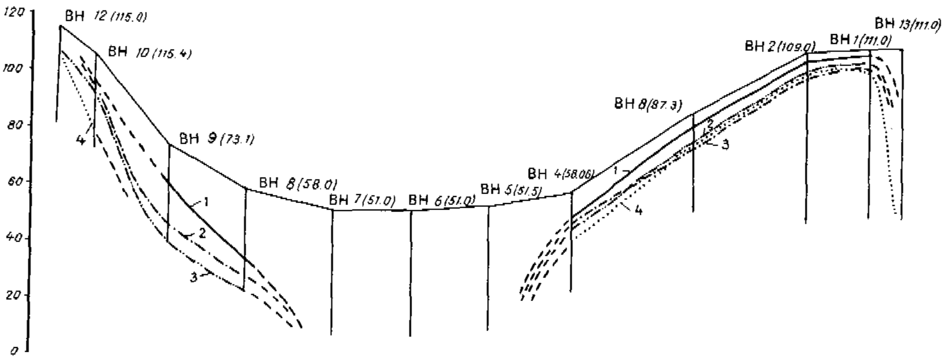

FIGURE 1 Zero isotherms in the base of the dam under the grouting gallery. 1--April 1968; 2--November 1968; 3--November 1969; 4--September 1972.

base was made in the form of a weakly reinforced [concrete] slab with grouting galleries. In accordance with the plans, the grouting of the base during construction of the hydroengineering complex was gradual as the thawed zone increased. Therefore, systematic observations of the varying temperature field in the base of the dam have direct practical significance in addition to purely scientific interest.

The arrangement of the temperature wells, their installation, the type of sensors, and the waterproofing of the strings were developed by the Leningrad Division of the Gidroproyekt Institute. These design solutions were partially implemented in the first steps of the construction (the temperature observations in the supporting prism of the dam and certain other structures); they were partially reworked and somewhat modified after some observations had been made by the Vilyuy Scientific Research Institute of Geocryology of the Siberian Department of the USSR Academy of Sciences. In particular, for observations of the thermal regime of the base of the dam under the grouting galleries and its suglinok core, the wells, the type of sensor, and their waterproofing were changed. For greater reliability, it was decided to install two types of sensors in each borehole: One type was the thermistors used by the Vilyuy Scientific Research Institute of Geocryology, and the other was the ordinary copper resistance thermometers as used at the VNIIG (All Union Vedenyev Scientific Research Institute of Hydraulic Engineering).

Initially 10 temperature wells from 35 to 60 m deep were drilled from the grouting gallery into the base of the dam (Figure 1). Subsequently, another three boreholes were drilled. In order to ensure watertightness, the boreholes were lined with steel casing. The strings of sensors were placed in the casings (10 sensors for holes 35-45 m deep and 12 for 60 m deep). The PSRP-type conductor was used for the strings and the KVRB-type cable was used for remote recording. The precision of the measurements was 0.1°C to 0.2°C. In order to avoid the effect of air convection on the accuracy of the temperature observations and to ensure reliable waterproofing, the wells were poured full of viscous "nigrol"-type* oil; the upper section was filled with bitumen.

Since the drilling and outfitting of the temperature wells in the base of the dam were carried out after erection of a significant part of the dam of the first stage and partial filling of the reservoir, the initial data on the temperature distribution in the rock base reflected a certain thermal condition developing under the effect of heat exchange with the structure and the reservoir. Nevertheless, we took it as the initial state.

In Figure 1 we have the zero isotherms in the base under the grouting gallery during the observation period from April 5, 1968, to September 27, 1972. From the given data, the difference in the thermal regime of the base is clearly shown with respect to the cross section of the dam, which could be predicted in the exploration, planning, and design stages. Deeper cooling of the rock massif on the right bank was recorded at that time (the slope with northern exposure), and increased fracturing of the rock in the base was recorded on the left bank. These initial engineering-geological and frozen conditions determined to a significant degree the nature of development of the thawed zone and the intensity of the heat-exchange processes after completion of the dam.

In the right-bank side of the base after an initial relatively intense thawing of the rock massif, the quasistationary thermal state was established; this is shown not only in the zero isotherms, but also in the temperature fields. Only in the rock massif directly adjacent to the outlet channel (borehole 13) is a sharp increase in the temperature and thawing. Insignificant thawing of the base also continues in the channel section on the side of the talik under the channel.

Another picture is recorded for the left-bank section of the base, where thawing took place very markedly during the first years; this was undoubtedly connected with percolation of the water as a result of increased fracturing of the rock massif. In 1969 the first stage in the

*[A heavy residual of petroleum]

grouting of the base was concluded. This caused immediate retardation of the thawing process and gave a relative stabilization of the temperature field (see Figure 1).

It must be emphasized that the temperature data can serve as an indirect indication of seepage. Thus, according to drillhole 9 (obviously the region of the most fractured rock, if we judge by the nature of the temperature field), the temperature variation at different depths to grouting of the base followed the water temperature variation in the reservoir. After the first stage of grouting, this relation was disturbed. Since the boundary of this zone went beyond the limits of the temperature wells 8 and 9, we have no exact data on the intensity of deeper thawing of the rock massif; however, by the dynamics of the temperature field of the thawed zone and by the nature of the temperature distribution with respect to depth, it is possible with all rocks to talk about the quasistationary thermal state of this part of the base. Of course, this does not pertain to the upper horizons of the massif, where seasonal air-temperature variations are felt in the grouting gallery.

All of the above data pertain only to the middle (relative to the transverse cross section of the dam) part of the base. Since the width of the base of the dam is more than 300 m, the differences in the thermal regime connected with the various heat-exchange conditions and the prevailing effect of certain factors must be clearly traced along the river channel. In the base of the loaded prism, it is possible to assume a thermal state close to that observed in the central part. The base under the low supporting prism experiences the sharp effects of heat exchange in the rock talus. According to the data from the observations performed by the All Union Vedeneyev Scientific Research Institute of Hydraulic Engineering, the most cooled zone of the lower prism is its lower section (the 71.0 level and lower), which was naturally also connected with the formation of the thermal regime of the base in spite of the warming effect of

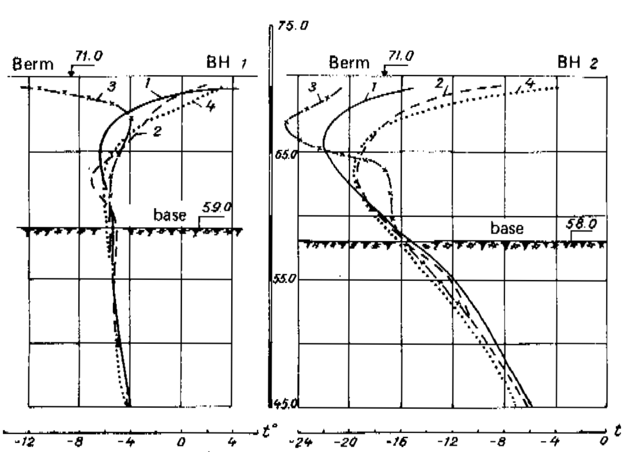

FIGURE 2 Temperature distribution in the lower supporting prism and its base (berm at the 71.0 level). 1--July 1971; 2--September 1971; 3--December 1971; 4--September 1972.

the tailrace and the possible "through" filtration in the rock massif.

In the middle of 1971, at the berm at the 71.0 level, temperature holes were drilled 26 m deep at station 1 + 80 and station 3 + 50, that is, the second well in the vicinity of the talik under the channel. The results of the temperature observations indicate (Figure 2) that freezing and nonuniform cooling of the upper part of the rock base occurred under the lower prism. The greatest cooling is observed in the channel section. Significant nonuniformity of the cooling of the various parts of the lower prism and its base is obviously connected with the difference in conditions of the heat exchange in the rock talus and greater warming effect of the tailrace in the vicinity of 1 + 80.

Thus, the field data indicate significant nonuniformity (in plan) of the thermal regime of the base and its dependence on the initial engineering-geophysical and permafrost conditions and the heat-exchange conditions in the construction process during the first years of operation of the hydroengineering complex. In some parts of the rock massifs of the base, relative stabilization of the thermal state had already occurred, whereas in others the heat-exchange processes still occur quite intensely. The general trend is heating and thawing of the rock in the base under the upper wedge and core of the dam and freezing and significant cooling of the rock (nonuniform with respect to the dam axis) under the lower supporting prism. The grouting of the base as it thaws, excluding the effect of the seepage on the thermal regime of the rock, promotes stabilization of the thermal effect in the zone under the core and cooling of the rocky massif under the lower prism.

For observations of the thermal regime of the suglinok core of the dam, the plan provides for drilling and equipping three slant (at an angle of about 33°) boreholes about 100 m long along the axis of the body at the 2 + 20, 3 + 60, and 5 + 00 stations.

The drilling of the slant holes always presents well-known difficulties, especially in this case, where any deviation from the designed direction could lead to perforation of the impervious core. Therefore, the holes were actually drilled at stations 2 + 20 and 5 + 00, respectively, to a depth (length) of 92 m and 77 m and only at station 3 + 60 to the designed depth. All three holes were equipped with sealed telescopic casing made of steel pipe 56-89 mm in diameter. In order to lower the strings of thermistors into the holes, a length of 1/2 "pipe" was used to which the string was attached as it grew longer. This permitted careful placement of the sensors with respect to length of the hole and also use of the length of pipe for pumping nigrol into the hole. From 16 to 19 sensors were installed in the holes. Matching sensors were installed at certain depths for control. On the whole, the procedure used in equipping the temperature wells was similar to that used to equip the holes in the grouting gallery.

The frequency of the measurements was established during field observations, beginning with

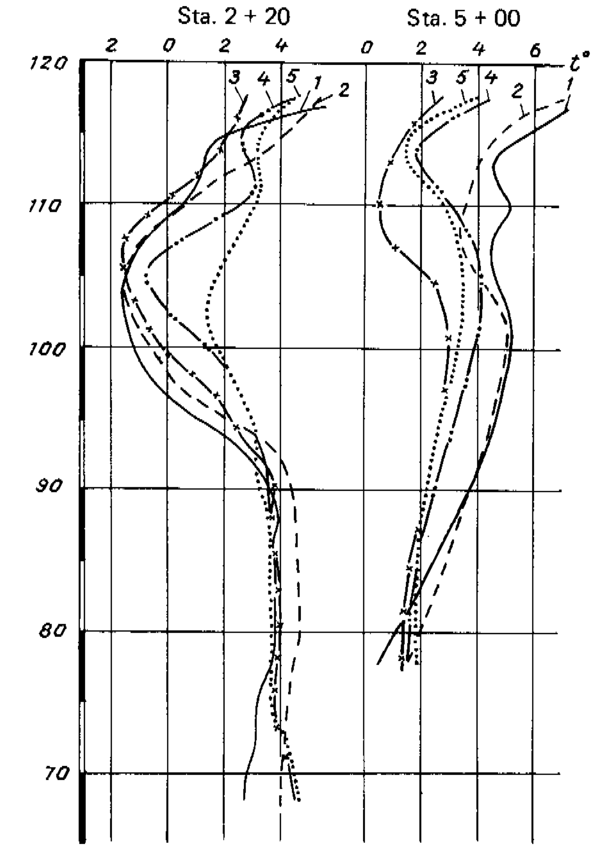

FIGURE 3 Temperature distribution in the core of the dam: 1--September 1968; 2--March 1969; 3--September 1970; 4--December 1971; 5--December 1972.

the intensity of the temperature fluctuations. Initially, after equipping the holes, the measurements were taken daily, then once a week. Subsequently, measurements were taken monthly.

Temperature observations in the core of the dam were started in 1968 and are continuing to the present time. An analysis of the data permits establishment of a general trend in the variation of the thermal regime of the body of the dam and its dependence both on the conditions and technology used in the construction process and on the general heat-exchange conditions of the dam and the environment during the first years of operation.

The suglinok core is affected by the complicated heat-exchange processes taking place in the supporting rock-fill prism, in the fill (which is thermally interacting with the reservoir), and in the base of the dam.

During the first years of operating the hydroengineering complex, the thermal regime of the core of the dam cannot be explained without considering its initial state developed during the construction process under complicated climatic conditions (see the paper by G. F. Biyanov, "Experience in Dam Construction..." in this volume). During erection of the dam, it was known that the layers of suglinok stations (96 to 109) between the 0 + 60 and 3 + 80 stakes were subjected to deep cooling and freezing, which had already been recorded in the first observations (Figure 3). The fact that the construction of the dam significantly preceded the filling of the reservoir also played an important role.

The heat exchange in the lower [downstream] prism of the dam has the sharpest effect on the thermal regime of the core of the dam. Natural air convection causes intense and deep cooling of the coarse fill during the winter in the zone under the body of the dam, so that on the lower boundary of the core negative temperatures can occur. This leads to cooling of the core of the dam until it is partially frozen, in particular up to the filling of the reservoir. This picture was observed in the body of the dam at Vilyuy Hydroelectric Power Plant until the middle of 1970, when the frozen layers of suglinok not only did not thaw, but there was an increase in the freezing zone and general cooling of it also took place; and at stake 3 + 60, freezing of the core also began, although during the initial period stable positive temperatures were recorded. In the 1970 floods, the water level in the reservoir rose sharply to the 116.0 level and then rose from year to year (Figure 4). Work was simultaneously done with respect to placing a "blanket" over the surface of the lower embankment, which reduced the intensity of the convective heat-exchange processes in the rock fill of the supporting prism. Both of these factors changed the heat balance of the core of the dam. The temperature in the core of the dam began to be equalized with respect to depth (see Figures 3 and 4) with general relatively sharp heating of the upper layers, some heating of the lower layers and insignificant cooling of the middle layers of the suglinok.

The temperature observations that were made permit judgment only of the distribution and variation of the temperatures with respect to depth of the core of the dam in its central section. It is quite obvious that in the transverse cross section there is also nonuniform tempera-

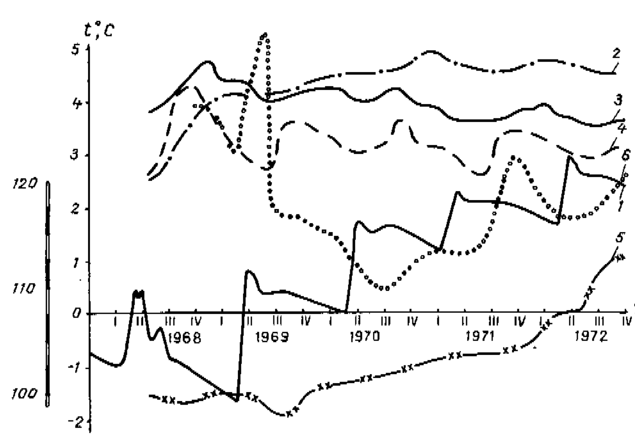

FIGURE 4 Measurement of temperature differences at various elevations on the core of the dam and water levels of the reservoir: 1--water level in reservoir; 2--sta. 2 + 20, 68.3; 3--2 + 20, 75.8; 4--2 + 20, 91; 5--2 + 20, 103.8; 6--5 + 00, 110.4.

ture distribution with a general trend of a reduction in temperature in the direction of the tailrace below water level in the reservoir. It is possible with some certainty to suggest that the layers of the core of the dam adjacent to the rock fill experience seasonal temperature fluctuations that follow the variations in the thermal regime of the reservoir, and the mean annual temperature at the upper boundary of the core of the dam is approximately equal to the mean annual water temperature in the reservoir.[3]

The temperature observation data for 1972 permit us to draw the conclusion of the beginning of stabilization of the thermal regime of the dam at the Vilyuy Hydroelectric Power Plant. The overall thermal condition of the structure corresponds to the design condition.

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PROPERTIES OF THE FORMATION AND QUANTITATIVE ESTIMATE OF THE ICE CONTENT OF GRANITE PERMAFROST AT THE CONSTRUCTION SITE OF THE KOLYMA HYDROELECTRIC POWER PLANT

S. YA. ZHUKOVSKIY, O. S. MAZUROV, AND I. A. PIROGOV Lengidroyekt Institute

The granite massif of the Upper Jurassic age, where the Kolyma Hydroengineering Complex is built, is in the upper course of the Kolyma River within the upper Kolyma highlands. In structural respects it is part of the In'yali-Debinskiy sunclinorium. The area is mountainous with altitudes of 1,000 m to 1,500 m, formed in the Pliocene-Quaternary age as anticlines; the Kolyma valley is narrow, asymmetric, and deeply incised into the bedrock. Its width at the water level of the river is about 120 m; at a height of 100 m, it is up to 700 m. The right side of the valley is terraced and covered with alluvial, fluvioglacial, and colluvial deposits of 3 m to 5 m up to 30 m thick. The left bank of the valley with a slope of 23°-35° is covered with a thin (1-3 m) veneer of colluvial deposits (Figure 1).

For the engineering-geological basis for the design of a dam 124 m high, a spillway channel, a flood gate, an underground machine room, and connecting tunnels, a study was made of the geocryological conditions, jointing, and ice content of the granite serving as the base and the country rock of these structures. The large-scale structural and cryologic surveys, exploratory work, including deep wells and adits, and also exploratory holes 15-20 m deep in the rock, geophysical, thermal (35 wells), and experimental and laboratory research were also performed.

The engineering-geological exploration demonstrated that the granite massif is dissected by original jointing and later tectonic cracks and zones. The original joints are grouped primarily in three systems: two steeply dipping and [one] gently dipping. The steeply dipping joints extend primarily to the northwest and east, that is, along and transverse to the river. Their length is from 2 m to 10 m and more, and the frequency is every 0.5 m to 2.5 m. Their opening is small, and as a rule there is no infilling. The gently sloping (subhorizontal) joints with dip up to 20° have an extent from several meters to tens of meters, the frequency is every 0.4-1.3 m, and they have somewhat greater opening. The tectonic cracks and zones are rare; the thickness of the tectonic zones is seen to be from 0.5 m to 5 m, and 0.5-1.2 m are most frequently encountered. The tectonic joints usually follow the direction of the original joints. In addition to the above-indicated types of joints along the outline of the valley, exogenic joints developed (weathering and relief) that have significant opening and loose infilling.

In accordance with the permafrost zonation of the northeastern part of the USSR (A. I.

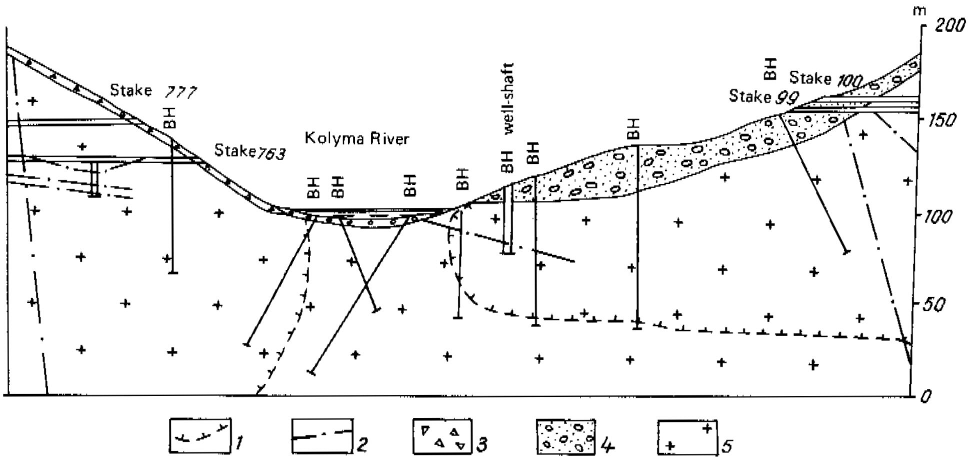

FIGURE 1 Geological section of the Kolyma valley: 1--outline of the permafrost; 2--tectonic zones; 3--colluvial deposits; 4--alluvial, fluvioglacial, and colluvial deposits.

Kalabin, V. A. Kudryavtsev, P. I. Mel'nikov, and P. F. Shvetsov), the investigated area belongs to the region of continuous permafrost broken up by taliks under large streams and at the points of release of the water under the permafrost.

In the area where the structures are located, the shores of the Kolyma River differ from each other with respect to permafrost thickness an depth of occurrence of the water under the permafrost. The least thickness of the permafrost--20 m to 50 m--was recorded on the right bank of the Kolyma in the vicinity of the terraces; its thickness increases to 160 m deeper into the shore. On the left bank, the thickness of the permafrost reaches 300 m. Between the frozen massifs of the slopes of the banks under the Kolyma River channel there is a through-talik.

The formation of the temperature regime of the investigated permafrost was caused by a number of factors among which the slope exposure has prevailing significance: on the left bank of the southern exposure, the mean annual temperatures of the rock are 1°-2° higher, and they are -4° to 7°; on the right bank with a northern exposure they are -6° to -8°C. The thickness of the layer of the annual temperature fluctuations is 20-30 m.

As a result of a detailed study of jointing of the granite and careful documentation of the ice-containing cracks in the adits and exploratory holes and analysis of this material, a number of laws were discovered in the nature of the ice content of the granite and its relation to the jointing, morphology, and history of the formation of the Kolyma valley.

The formation of the ice in the granite massif during its freezing took place along the cracks primarily as a result of filling of them and in the surface part also as a result of expansion of the cracks during the process of multiple freezing and thawing of the rock. With respect to the peculiarities of its formation in the granite, the ice belongs to the vein and injection-cementation types, although sublimation is noted and it is represented in the fissures both by pure ice and ice in combination with various types of minerals and loose fillers. The fissures of different genesis are characterized by different ice content and different ratio of it to the filler. The joints usually contain pure ice of a turbid white color in the form of thin interlayers several millimeters thick. In the thin cracks, the crusts, films, and various layers of sublimation ice are observed. In the tectonic cracks, the ice is in the form of streaks between the surfaces of the fissures, frequently covered with calcite or tourmaline, or it is in contact with a layer of mylonite. The form of the streaks is usually complex, and their thickness varies from 1-5 mm to the first centimeters [*sic*]; then cementation of the mylonite with ice is noted. The cataclastic broken granite is impregnated with ice or is dissected in the complex part by a network of ice veins, and it is the iciest rock (Figure 2).

In the distribution of the ice containing exogenic cracks, two stages are isolated in the vertical section of the granite massif: The first belongs to the zone of intense surface weathering and surface relief, and the second encompasses the massif below this zone. The thickness of the zone of intense weathering of the granite is different in the transverse profile of the valley, and it varies from 2-5 m to 10 m. Its characteristic features are the

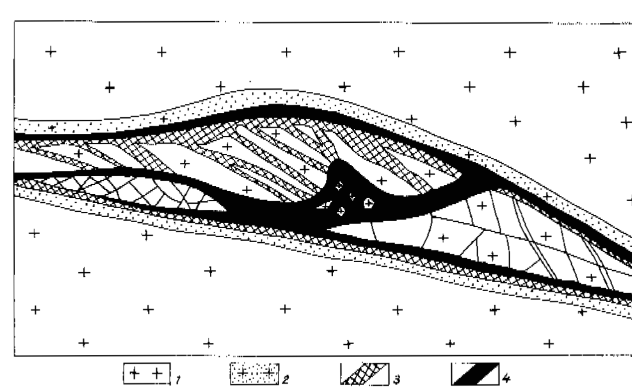

FIGURE 2 Nature of the ice content in the gently dipping tectonic zone: 1--weakly jointed granite; 2--granite impregnated with iron oxides; 3--mylonite; 4--ice.

TABLE 1

Location	Adit Number	Total Number of Fissures/ Percentage Content	Fissures with Mineral Filler	Hollow Fissures	Ice-containing Fissures		
					Pure Ice	Ice with Mineral	Ice with Scree
Right bank	100	267/100	50/18	193/73	18/7	--	6/2
	99	253/100	65/26	153/60	--	11/4	24/10
Left bank	763	88/100	6/7	28/32	46/52	5/6	3/3
	777	36/100	--	--	33/92	--	3/8

NOTE: The number of cracks is indicated in the numerator, and the percentage content in the denominator.

ice in the predominant part of the cracks, the presence of injected loose material, and maximum thicknesses of the ice-containing cracks reaching 10-30 cm.

The ice content of the exogenic cracks below the zone of intense weathering on the sides and the bottom of the valley differs. The exogenic cracks in the sides of the valley contain ice with an admixture of scree, which is the product of disintegration of the granite. The ice cement in such cracks often prevails over the scree filler. The visible inclusions of the ice in the form and lenses and pockets have different dimensions.

The cracks in the rocks at the bottom of the valley are filled with scree mixed with river sand; the ice in them is in the form of ice cement and also forms individual lenses and pockets in the filler. The thickness of the exogenic cracks below the zone of intense weathering reaches several centimeters.

Below the zone of development of exogenic fissures in the retained granites, the following characteristic features are noted in the distribution, frequency, and thickness of the ice-containing cracks:

1. Ice is observed in the cracks of all the genetic types independently of the elements of their occurrence.
2. The greatest frequency and thickness of the ice-containing cracks is associated with the sections of tectonic jointing, and especially the tectonic zones, where the thickness of the ice inclusions sometimes reaches 5 cm.
3. Among the ice-containing cracks, usually cracks 1-5 mm in size predominate.

Depending on the association with one morphological element or another of the valley, the ice content in the cracks also is not identical. At the bottom of the valley, practically speaking there are hollow cracks--all the tops are very ice-containing. The opposite slopes of the valley of the Kolyma River differ sharply from each other with respect to distribution of the ice-containing fissures. On the right slopes, the hollow cracks have predominant development in the granite, and the ice is primarily associated with the tectonic zones. The number of ice-containing cracks here is on the average of 10 percent. On the left bank, the number of ice-containing fissures rarely increases, and it is 60-100 percent of the total number (Table 1).

The indicated differences in the ice content of the granite massif, just as the different thickness of the permafrost on different banks of the Kolyma River, can be explained:

1. By the warming of the massif during the last thermal optimum by water flows from the thawing mountain-valley glacier, the attributes of which in the form of cirques, moraines, fluvioglacial deposits, and others appear on right bank of the Kolyma River.
2. The warming effect of the subchannel talik existing on the right bank and gradually shifting to the left, which is indicated by the suprafloodplain terraces.

Inasmuch as the ice content of the rock massif is predetermined by its fissuring, the quantitative evaluation of the ice content was performed as applied to the coefficient of fissure void as comprising part of this index (Table 2).

TABLE 2

Geomorphological Element of the Valley	Granite in the Exogenic Fissure Zone		Granite of the Lower Exogenic Fissure Zone	
	Coefficient of Fissure Void, %	Ice Content, %	Coefficient of Fissure void, %	Ice Content, %
Right bank slope	1-4.5.0	0.5-2.8	0.4	0.04
Bottom of the valley	4-6 to 11	3-4 to 9	0.2-1.2	0.1-0.6
Left bank slope	1.5-3.5	1.01-2.5	0.02-0.3	0.01-0.2

Establishment of the characteristic features of the formation of the ice content in the rock of the massif and its quantitative evaluation permitted a well-founded approach to the characteristic of the deformation and percolation properties of the base of the structures of the Kolyma Hydroelectric Power Plant.

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CRYOGENOUS PHENOMENA IN EARTHEN HYDROENGINEERING STRUCTURES

YA. A. KRONIK MISI Kuybyshev Institute

This report briefly generalizes the first experiences of complex studies of the cryogenous processes and phenomena in hydroengineering and hydroameliorative geocryology by the MISI Kuybyshev Institute under the direction of Corresponding Member of the USSR Academy of Sciences, N. A. Tsytovich.

On the basis of a generalization of the published data and many years of experience in complex geocryological studies performed during the construction and operation of the first rock-dirt dams in the Far North and certain hydroameliorative structures,[1-5] a provisional classification of cryogenous processes phenomena developing in earthen hydroengineering structures and of definite, frequent danger to them is proposed:

1. Heaving of the surface layers of the ground during freezing with subsequent subsident during thawing; heave of the structural elements and foundations in the case of multiple freezing and thawing of the base and the earthen structures.

2. Specific temperature deformations and crack formation caused by temperature stresses in the frozen and thawing ground of the earthen structures on variation of their temperature.

3. Thermokarst during thawing of the ice-saturated ground of earthen structures and their bases, including partial thermokarst caused by artificial thawing (for example, electric thawing, thermohydraulic thawing, and so on).

4. Solifluction of the embankments of the earthen structures and canals, landslides, and cave-ins of the banks of the canals in the frozen and thawing rock.

5. Ice body formation during freezing and percolation outcrops during thawing of the ground in the fills and bases of earthen structures and canals.

6. Ice formation in the rock material of the earthen structures and the supporting prisms of the rock-earth dams.

7. Frost weathering of the coarse soils and rock fill in the earthen structures and rock-earth dams.

8. Thermal erosion of the ice-saturated em-

bankments of the water-conducting systems and canals, the shores of the reservoirs, and the shore elements of the dams.

Each of these types of cryogenous phenomenon must, in turn, be subdivided into variations and types, depending on the type of structure that it affects and the degree of danger of it for certain earthen structures. All of this requires a more detailed study of the enumerated cryogenous phenomena, the characteristic features of their effect on the structures, and evaluation of their danger both during construction and during operation considering the local climatic and permafrost conditions and the structural peculiarities of the earthen structures themselves.

As a result of analysis of the published Soviet and foreign data on ground heaving (more than 40 types), a prolonged experimental study of the danger of frost to 24 types of ground in Siberia, the Far East, and the Far North, and the complex laboratory and field studies of the cryogenous processes in the rock-earth dams and hydroameliorative earthen structures, it was possible to establish that basically almost all the cryogenous phenomena characteristic of fine-grained soils are in direct and close connection with each other and develop most intensely in the most frost-susceptible, sharply heaving soils. Therefore, the criteria and the provisional limits of heaving soils proposed by the author[5] can be extended in the first approximation to evaluate the degree of danger to the soil in the earthen structures, as to the possible intensity of the development of frost processes in them or with respect to the *general frost danger* to the soils in the more general understanding of this term than the commonly used synonym of frost heave.

The comparison made between these traditional boundaries and the analogous data of other researchers demonstrated that, in spite of some differences, the zones of the weakly, and sharply, heaving soils compared quite closely for the majority of authors studying the heave of soils of different type and genesis in the various countries. This indicates a definite law of division of the soils with respect to degree of heaving, and it permits quite well-founded recommendation of the established criteria and the limits of frost danger to the ground in geocryological practice and, in particular, in northern dam construction. The proposed provisional boundaries of frost danger to soils permits the selection of less frost-endangered or, in general, nonfrost endangered ground to erect earth structures or, in the case of the impossibility of replacing the local soils, taking more directly during the planning and design work of the required antifrost measures or measures for technical improvement of the soil.[1,2,5]

The experience in building and operating the first hydroameliorative structures and dams in the Far North and the results of research performed on them[1-5] demonstrated that the cryogenous heave and the frost-crack formation in the ground of the earthen elements of the dams and the bodies of earthen structures and also solifluction of their slopes, thermokarsting of the bases, and thermal erosion of the shores of canals and the lateral wings of the dams erected in the regions of ice-saturated frozen rock present the greatest danger to these structures.

Let us consider the cryogenous phenomena observed in rock-soil dams during the construction period in the example of the right-bank of the rock-fill dam of the Khantay Hydroengineering Complex.[4]

The basic complexities resulting from frost phenomena during construction of dams on permafrost arise primarily when exposing the excavations and preparing the bases of the dams and also during winter erection of their impervious elements from cohesive soils. On erection of the core of the right-bank of the dam of the northernmost in the world [for the] Khantay-Hydroelectric Power Plant, cryogenous phenomena were observed: (a) rated formation as a result of seepage--springs of the severely flooded base (in the northern lowlands near the previous channel) and seepage from the slopes of the open pits and the foundation of the core (in the northern and southern lowlands, also in the former channel and swamps); (b) thermokarst during development of bases for dams and artificial thermokarst caused by electric thawing of the frozen ground of the saturated layer covering the thawed flooded base of the northern lowlands of the dam; (c) heaving of the cohesive soils by freezing with layer-by-layer winter laying in the core; and (d) frost-crack formation during freezing of the cohesive soils after compaction in the core.

The formation and growth of naleds up to 80-100 cm thick and up to 2,500-3,000 m^2 in area with a mean temperature of about -9°C occurred in the same way as found for the development of natural naleds. The appearance of naleds in the vicinity of the core greatly complicated the performance of the operations in building the dam, and it required additional expenditures for the removal of the naleds.

The artificial thermokarstic sinkholes to 1 m deep also complicated the construction of the dam and made necessary the complete removal of the thawed liquid soils from the core in the vicinity of the greatest development of thermokarst and replacement of it by high-quality thawed soils.

In order to study the heave and crack formation during thawing of the layer of ground laid in the core and also to evaluate the danger of these cryogenous phenomena and their effects on the reduction in quality of the soil of the core, special field studies were made in an experimental section of the dam. The goals of the experiment also included measurement of the specific cryogenous deformations (heave and crack formation) along the three coordinate axes with use of differential heave meters and the crack meters, the heaving of the ground surface layer with respect to stations, and measurements of the air and soil temperature in depth of the layer on the control plot of the experimental section and on the plot developed by the method of complex antiheave salt treatment.[1,2,5] The results of field studies are discussed in detail in Tsytovich *et al.*[4]

The joint analysis of heave and crack formation developed in the freezing ground simultaneously, and, in interrelation to each other, demonstrated that, together with the uplift (heave) of the experimental layer of the core of the dam, folding of the deformed frozen surface takes place. Some expansion of the layer of soil from the direction of the headrace and reduction from the direction of the tailrace are observed.

In the transverse cross section, the compression zone, on the contrary, is located close to the headrace, and the tension zone, closer to the tailrace. Thus, with the possibility of deep cooling (freezing of the core), the transverse frost cracks can probably occur primarily in the zone of contact between the core and the upper filter, at the contact of the core with the sides of the valley in the lower section and in the upper zone of the core itself. In turn, longitudinal cracks (parallel to the axis of the core) can occur primarily at the contact between the core and the upper filter and also in the lower part of the core itself. In the central part of the experimental plot, a tension zone was set up that could cause the formation of a longitudinal crack in the center of the core. The most dangerous case for the dam is the occurrence of the transverse cracks in the core from the direction of the headrace and the longitudinal cracks at the contact between the core and the headrace (or the transition zone).

All of this agrees quite well with the data from observations on the deformations of the Kelsey dikes (Canada), where deep longitudinal frost fissures, requiring annual expenditures on their repair, were formed in the middle of the core, and on the small dikes longitudinal rents occurred along the crest of the embankment.

The studies made also demonstrated that the temperature deformations of the soils in a dam develop according to the known laws of thermorheology,[6,7] and for very coarse soils (classes I and II according to the author's classification) with fine-grained fill, the thermorheologic properties and the amount of fill are predetermined. The coefficient of thermal expansion of the Khantay gravelly supes were found to be 25 to $50 \cdot 10^{-6}$ per deg. This is close to the data of I. N. Votyakov[6] for supes. The limiting values of the temperature deformations, which when exceeded lead to the rupture of the studied soil with the occurrence of microcracks and wide cracks, are approximately equal to 300 to $350 \cdot 10^{-6}$ or 0.3-0.35 mm/m on the surface of the frozen layer.

From the discussion above, it is possible to draw the conclusion that when building earthen hydroengineering structures in the Far North it is not recommended that the most frost-susceptible severely heaving soils be used for the essential elements of the dams and high-quality fills. It is proposed during erection of dams from local materials in the permafrost zone that physicochemical and complex measures be used broadly for technical amelioration of the ground[1,2,4,5] and that the thawed ground be worked into optimal ground mixtures with the introduction of coarse admixtures up to 40-55 percent (of the weight of the mixture).

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DEFORMATIONS OF UNDERGROUND WAVE GUIDES DURING SEASONAL FREEZING OF THE GROUND

A. V. SADOVSKIY AND L. P. SEMENOV Scientific Research Institute of Foundations and Underground Structures
V. M. DMITRACHENKO All Union Scientific Research Institute of Communications

A wave guide is one of the elements of a new multichannel type of communication. It is a metal tube 60-100 mm in diameter. In order to insure normal functioning of this type of communication, increased requirements are imposed on the stability of the spatial position of the wave guide.

Of all the types of deformation, bending in the vertical and horizontal directions has the most significant effect on the normal operation of the wave guide. According to the technical specifications, it is necessary to maintain the rectilinearity of the wave guides in comparatively large sections or a definite magnitude of the radii of curvature in the vertical and horizontal planes. The outstanding feature of wave guides is the low stiffness of the tube in comparison with that of other kinds of pipe. Therefore, even insignificant disturbances in straightness of the tube during the construction process, for example, as a result of nonuniformity of the ground, deviations of the levels of the ground from the design levels, back-filling of the ditches, loads from trucks, and so on are reflected in the quality of their operation.

Such requirements with respect to the admissible deformations were not imposed earlier on any of the known types of pipe; therefore, the nature of the effect of the freeze-thaw processes on the pipes of comparatively low stiffness turned out to be almost entirely uninvestigated.

The winter freezing of the soil is one of the basic natural processes predetermining the selection of depth of laying of the wave guides in connection with the necessity for insuring stability in the presence of nonuniform deformations of the ground. On the other hand, deep laying of the wave guides, especially in the presence of a high water table, appears to be unreasonable.

In order to select the optimal depth, it is necessary to discover how the processes accompanying the freezing and thawing of the ground affect the deformation of the wave guides.

When freezing and thawing wet fine-grained soil, volumetric changes take place--heave and settlement; therefore, any wave guide laid in seasonally frozen ground will be subjected to the strong effect of the surrounding ground. The most unfavorable and worst effects so far as the operation of the wave guide is concerned are the nonuniform spatial displacements of the tube in the ground with residual deformations with respect to the initial position, which leads to distortion of the longitudinal axis of the tube.

Of the factors affecting the magnitude of the nonuniform deformation of the wave guides in the freezing process, it is possible to isolate the following: (a) changes in the engineering-geological conditions along the route; (b) sharp variation of the depth of freezing in individual sections (sections from which snow has been removed, under roads, and so on); and (c) temperature regime and ground freezing rate.

The study of the deformation of underground wave guides was made by the Scientific Research Institute of Foundations and Underground Structures jointly with the Central Scientific Research Institute of Communications in the experimental section several kilometers long under the natural conditions of the Moscow area. The ground along the route consists of heavy suglinoks.

Of the variety of engineering-geological and thermophysical conditions along the route, the following unfavorable combinations were selected for reproduction in the experimental section of the wave guide: (a) sharp changes of the depth of freezing of the ground with shallow (0.6 m) burial of the wave guide; (b) the same with increased depth of burial (to 1.4 m); (c) sharp changes in the ground conditions along the route--combination of heaving and nonheaving ground (suglinoks and sand with shallow burial of the wave guide (0.8 m); and (d) the same with increased depth of burial of the wave guide (to 1.2 m).

The observations of the wave guide deformations were made during three winter seasons after the wave guide was laid and the ditches were filled.

The vertical displacement of the tube was measured by clock-type gauges with accuracy to 0.01 mm; the stresses in the walls of the tubes were measured by strain gauges, and the temperature regime of the ground, by thermocouples.

Observations demonstrated that during the first winter season after completion of construction, small displacements of the wave guide tube upward (2-3 mm) which were caused by the freezing ground, were encountered. In the spring the wave guide settled, in some sections as much as 20 mm to 30 mm. Observations of subsequent years demonstrated that the wave guide experienced seasonal vertical displacements--uplift and settlement of the tubes. The amplitude of these displacements did not exceed ±3 mm to 5 mm.

Unfavorable combinations of engineering-geological conditions artificially created in the experimental section had no significant effect on the distortion of the wave-guide axis during the process of seasonal freezing of the soils. The effect of sharp variation of the depth of freezing of the ground during shallow (0.6-0.7 m) burial of the wave guide turned out to be the most significant. In this section, over a length of 5 m, the change in depth of freezing was 1.1-1.2 m. In the spring one part of the tube turned out to be fastened to the frozen ground, and the other settled the thawed ground. The consequence of this was distortion of the longitudinal axis of the wave guide; the radius of curvature was about 500 m.

The studies made demonstrated that in the general case the deformations of the underground pipe of comparatively low stiffness (diameter less than 80 mm) laid in seasonally thawing ground are made up of the postconstruction deformations caused by the freeze-thaw processes. The postconstruction settlement has, as a rule, greater significance than the subsequent seasonal displacements of the pipe. After damping of the postconstruction settlement, the pipe remains in a comparatively stable position, with residual distortion of the longitudinal axis formed during the first year after construction. The seasonal deformations of the pipe take place in three basic stages: the first is settlement in the nonfrozen ground from the pressure of the overlying freezing ground, the second is uplift as a result of heaving of the frozen soil, and the third stage is settlement during the spring thaw.

The postconstruction settlement of the pipe can be decreased significantly as a result of higher density of the ground and the construction of a sand bedding 10 cm thick. Observation of these facts makes it possible in a number of cases to lay the wave guides within the layer of seasonally frozen ground.

The studies made served as the basis for developing procedural instructions with respect to laying underground wave guides.

TEMPERATURE REGIME OF THE FROZEN SOILS AT THE SITES OF SOME TRANSPORT STRUCTURES

YU. S. PAL'KIN, A. A. TSERNANT, V. A. STEPANOV, AND Z. M. PAL'KINA
Novosibirsk Branch of the TsNIIS Institute

In 1970-1971, at one of the new transport structural sites in the permafrost zone, observations were organized of the temperature regime of the ground for three structures: the fill around the railroad, the bridge, and the culvert designed with the assumption of thawing of the frozen ground.

The observation section is located in the vicinity of insular permafrost, the thickness of which does not exceed 20 m, and the mean annual temperature is 0°C to -2°C. On approaching the floodplain of the river, the ground temperature rises, reaching +2°C. To a depth of 1.5-4.0 m, the frozen ground is made up of subsiding and strongly subsiding deposits during thaw: peat, suglinok, and clay of coarsely layered, rarely thinly layered and massive cryogenous texture; lower down it is supes, sand, and residual [soil from] argillite with a reticular cryogenous texture.

Each experimental site is equipped with several drillholes 12-20 cm in diameter and 8-14 m deep, in which thermal sensors made of copper resistance thermometers are installed. The 9 to 12 temperature sensors are installed with a spacing of 0.5 m to 1.0 m to a depth of 4 m to 5 m and 1.5 m to 2.5 m to the depth of zero annual amplitude. The measurements are taken monthly, and their precision is estimated to be ±0.1°C.

A description is presented below of the experimental sites, and the basic results are given from the observations that are performed until the formation of the quasistationary temperature regimes in the investigated atmosphere-ground-structure systems.

RAILROAD FILL

The observations were organized for two types of fill: the fill for the temporary spur made of clay and argillite poured on the frozen or partially thawed ground, and the fill for the main track made of a sand-gravel mixture washed by the hydraulic method on to the thawing ground. The placing of the fill, 3-8 m high, took place in the summer and the winter of 1968-1969 in layers 0.8-1.2 m thick with compaction by a dump truck and pneumatic rollers. The hydraulic fill 12-16 m high was started in the summer of 1970, and it is being carried out at the present time using floating dredges. The structural design of both types of fill for one of the three cross sections of the experimental section appears in Figure 1.

Observations of the fill for the temporary spur were started 2 yr after the beginning of operation; on the main track they were started before the beginning of its construction, in

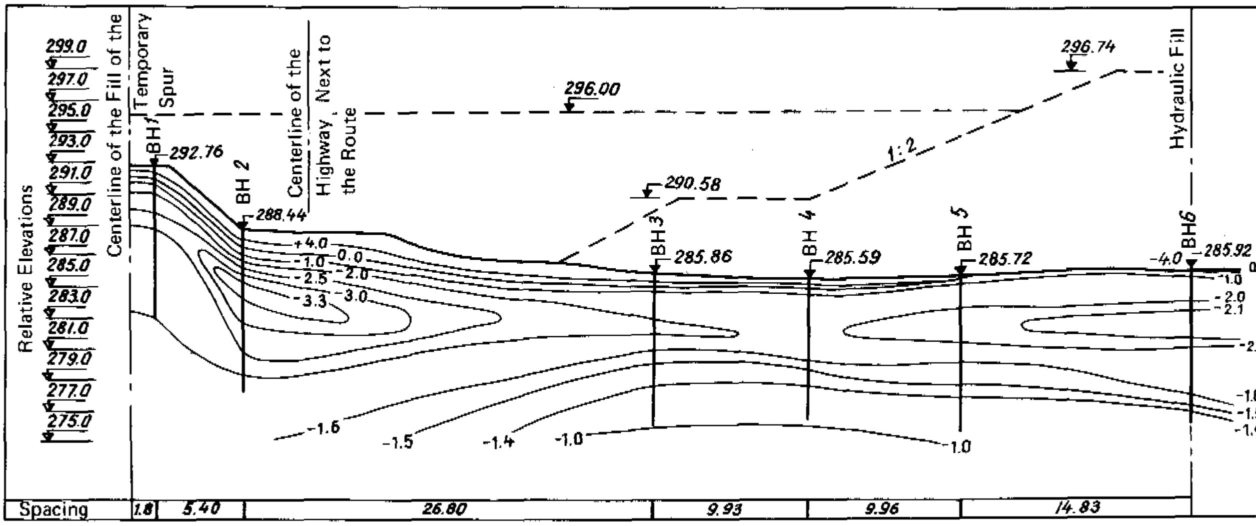

FIGURE 1 Temperature fields in the subgrade of rolled and hydraulic fills on 10 June 1971 (before the beginning of sluicing).

order to measure the temperatures in the subgrade and the body of the fill during the sluicing process. At the experimental site 18 boreholes were installed, from 9 to 12.5 m deep. The temperature measurements were duplicated by inertial mercury thermometers. During the sluicing of the fill, the stationary thermal assemblies were built up.

In addition to the ground temperature measurements, regular observations were made of the air, and water, temperatures in the dredge sites and at the hydraulic fill sites and also the snow accumulations. An analysis of the results obtained (an example of the processing of the results is shown in Figure 1) permitted establishment of the following factors:

1. The directional nature of the thermal processes in the subgrade of the fill (cooling or warming of the permafrost) is determined primarily by the height of the fill, the ground temperature at the depth of the zero annual amplitude, and the mean annual surface temperature. The technological process for constructing the fills, the season of the construction of the fill, the type and characteristics of the ground, and also the moisture conditions (in the absence of the infiltration) are factors actively affecting the stabilization rate of the temperature regime of the subgrade, but having almost no effect on the trend of this stabilization.

2. The hydraulic fill interrupts the natural thermal conditions of the subgrade to a much greater degree than the rolled fill, which is promoted by the convective heat exchange during percolation of the water from the fill areas. This is also affected by the fact that the water temperature at the exit from the areas is 0.5° to 3.0° at the same time as the temperature of the water-soil mixture fed to the fill area was 12°-18°C.

3. With a fill height up to 4 m, the subgrade temperature was 1.0°C to 1.5°C lower than in the seasonal thawing layer at the corresponding depths (see Figure 1) and 2°C to 3°C lower than in the base of the fill more than 6 m high. Considering analogous observation results of other researchers, it is possible to propose the existence of a completely defined critical height of the fill under certain conditions, for which there is a reduction in temperature in the subgrade or retention of it at the natural level and, above this height, an increase in the temperatures and thawing of the permafrost. Under the conditions of the observations made, the critical height of the fill was 3-4 m.

BRIDGE

The abutment and four piers were selected for observations of the thermal state of the ground of the shore and intermediate supports of a multispan bridge. The foundations of the two supports were constructed in December 1970 to January 1971, and the other two in July-August 1971. The foundations are of mass concrete with a depth of 9-11 m and dimensions in plan of 11 × 11 m.

The observations were organized during the period of construction of the foundations and construction of the piers, which permitted establishment of special lines for temperature measurements directly in the body of the foundations with their subsequent growth as the piers were erected. In all, copper-resistance thermometers were installed in 21 wells 8 m to 14 m deep, grouped primarily in five sections perpendicular to the centerline of the bridge.

From an analysis of the results of the temperature measurements, as is shown in Figure 2a, it is obvious that:

1. The thermal effect of the massive concrete foundations of the bridge supports on the surrounding groups is detectable at a distance up

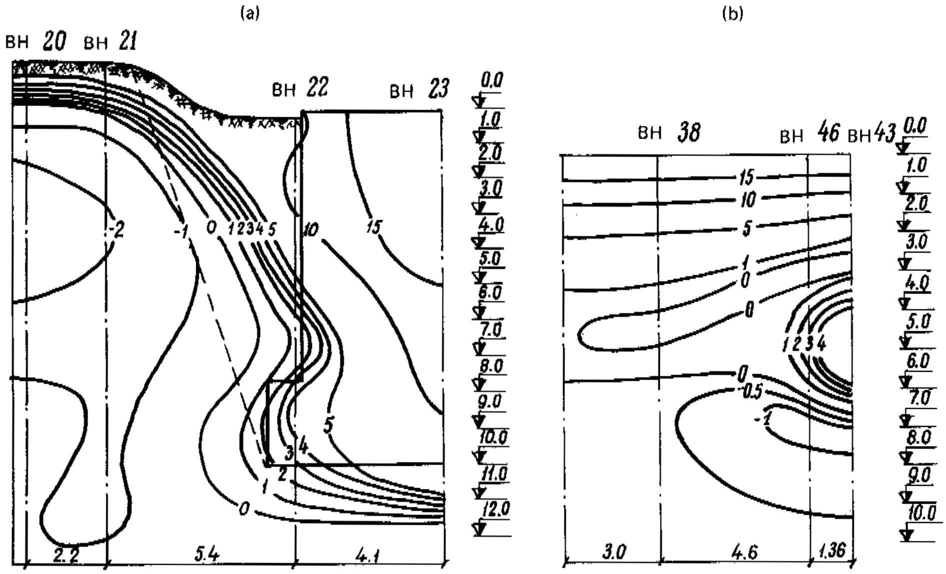

FIGURE 2 Temperature fields in the ground at the foundation of the intermediate support of the bridge as they were on 16 June 1971 (*a*) and for the groundwater level according to the observations on 17 July 1971 (*b*).

to 5 m to 7 m from their faces. However, this effect is felt most significantly at 2 m to 3 m. In this zone the mean temperature of the ground (with negative mean annual surface temperature) is 1.0°C to 1.5°C lower than the ground temperature under natural conditions. As a result of the thermal conductivity of the concrete, in the summer the foundation raises the mean temperature of the surrounding ground by 2°-3° (Figure 2a) and in the winter it lowers it by 3°-4°C.

2. Hardening of the concrete heats the body of the foundation to a temperature of 22°-24°C, which gradually drops at a rate of 2°C to 3°C per month. This leads to a degradation of the permafrost table under the foundation by 1.5 m to 2.0 m (Figure 2a) and general warming of the ground, which, at a frozen ground temperature of above -1°C can cause irreversible thawing.

3. The degree of disturbance of the temperature regime of the ground and the rate of its subsequent stabilization essentially depend on the construction season. When doing the work in the winter, the ground is cooled through the walls of the pit to such a degree that the cold reserve obtained completely compensates for the heat release during hardening of the concrete, and the ground temperature at a distance of 4 m to 6 m from the foundations 7 to 8 months after winter concreting remains 1.0°-1.5° lower than in the case of summer concreting.

CULVERTS

Observations were organized on two culverts constructed in the summer of 1968 on periodically active streams. The height of the fill above the first line was 4 m and above the second, 8 m. Their length was 42 m. The inlets were 1.35 m in diameter. The foundations of the elements were of concrete with a depth from 1.0 m under the midsection of the pipe to 2.45 m under the protruding head.

The experimental section was equipped in April-May 1971 with 13 holes from 8 m to 12.5 m deep with stationary ETS thermal instrumentation. The wells are located along four sections parallel to the longitudinal axes of the pipe.

The results of the observations, which are presented by the temperature fields for each date of the measurements in Figure 2b, provide a basis for the following conclusions:

1. The thermal effect of the culvert on the ground in the body of the fill is felt at a distance of up to 7 m from it, and in the subgrade not deeper than 4 m. The magnitude of the thermal effect zone decreases on going away from the center line of the fill toward the ends of the pipe. During the summer, a thawing aureole is formed around the pipe, the limits of which are approximately an identical distance from the top and bottom of the pipe and somewhat farther to the sides of it.

2. With a fill height above 6 m, the zero isotherm at the base of the pipe rises on going away from the head and then in the midsection it drops again. Here, a zone of plastically frozen ground with a mean annual temperature of 0°C to 0.3°C is formed above the pipe, the boundary of which moves toward the base of the fill on going away from the pipe and emerges with the upper boundary of the permafrost.

3. The greatest warming effect of the pipe on the supporting ground is noted at the emerging head, where, as a result of the convective heat transfer during percolation of the water, a local thaw bowl can be formed. The observations during the maximum discharge (May-June) established the presence of a temporary water-bearing layer in the base of the emerging head at a depth of 8.0 m to 8.5 m fed by the surface water and having a discharge 40 m from the head.

The observation results and the data accumulated on continuing the experimental work are a basis for the methods developed at the SibTsNIIS for predicting the temperature regime of the frozen ground in the foundations of constructed transport structures.

CALCULATING THE SURFACE TEMPERATURE OF PAVEMENTS, ROADS, AND RUNWAYS ON PERMAFROST

V. N. IVANOV *Lenaeroproyekt Institute*

The accuracy of calculating the temperature regime of runway or road pavements, independently of what procedure is used in making the calculations (analytically or by computer), depends primarily on the accuracy of determining the upper boundary condition--the surface temperature.

At the present time the surface temperature of the ground or a pavement is found from the heat balance equation[1,3] from which it follows that

$$t_n = t_B + \frac{R - (LE + B)}{\alpha_k} \qquad (1)$$

where t_n is the surface temperature of the pavement, °C; t_B is the air temperature, °C; R is the radiation balance, kcal/m^2·h; LE is the evaporation component, kcal/m^2·h; B is the heat flux in the ground, kcal/m^2·h; α_k is the coefficient of surface heat transfer, kcal/m^2·h·deg.

The greatest difficulties consist in establishing the calculated values entering into Formula (1) when determining the maximum depth of thawing of the base and the subgrade and also variation of their temperature regime with time.

The utilization of the mean monthly values of the air temperatures over many years and the radiation balance presented in the reference work[2] in the calculations does not guarantee reliability of the results obtained, since in reality significant deviations in these values toward the higher or lower side are possible. The indicated deviations can be determined with given reliability by the methods of mathematical statistics and probability theory. However, great difficulties arise in determining the optimal reliability. The higher the reliability, the greater the one-time expenditures on building the airport or the road, but the less the operating expenditures, and, on the contrary, a lower reliability corresponds to lower one-time expenditures and larger operating expenditures. If we take the equation expressing the total cost of construction and maintenance of the structures as a function of reliability, take the derivative with respect to it, and equate the result to zero, then we obtain the following formula for determining the conditions of optimal confidence level:

$$P_{co} = \frac{5AN}{2MH_{0.683}}, \qquad (2)$$

where A is the cost of the maintenance per square meter of area of the structure, rubles/m^2/yr; N is the service life of the structure before capital repairs, years; M is the cost of 1 m^3 of soil from which the fill is made, rubles/m^2; $H_{0.683}$ [M] is the height of the fill with a confidence level of 0.683.

The transition from P_{co} to the actual magnitude of the optimal confidence level P is made according to Table 1.

As studies have demonstrated, the cost of the operation and maintenance at the airports and on the roads for the majority of places in the Arctic and the Far North is $A \geq 2$ rubles/m^2·yr. For such values of A, the optimal confidence level is 0.95 according to (2).

For calculation of the maximum thawing depth and, consequently, the height of the fill when building the airports and roads by the principle of maintaining the frozen state of the subgrade, it is recommended that the initial data in Formula (1) be designated with a confidence level of 0.95.

As a result of processing the data on the mean monthly air temperatures by the methods of mathematical statistics for 40 yr of observations of several points in the Far North, the magnitudes of the errors were obtained, which, with a confidence level of 0.95, are equal to twice the standard deviation, that is, $\pm 2\sigma$. Analysis of the absolute errors 2σ for several points demonstrated that, within 1 month, they differed very little from each other. Therefore, when determining the calculated mean monthly air temperatures with some reserve, it is possible to take constant values of the errors for each month equal to the maxima from all the stations for

TABLE 1 Values [P] at the Optimal Confidence Level Corresponding to P_{co}

P_{co}	0.4	1.4	2.4	3.4	4.4	5.4
P	0.683	0.75	0.80	0.85	0.90	0.95

TABLE 2 Values of the Maximum Errors $(2\sigma_B)^{max}$ in Determining the Calculated Values of the Mean Monthly Air Temperatures

Month	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
$(2\sigma_B)^{max}$	9.9	10.8	10.1	8.0	4.6	3.1	5.2	3.4	3.6	6.0	8.6	10.0

which the calculations were performed, that is, $(2\sigma_B)^{max}$. The values of the maximum errors are presented in Table 2.

Thus, the calculated values of the mean monthly air temperatures when determining the maximum depth of thawing of the airport or highway fill are defined as the sum of the mean monthly values[2] and the values of $(2\sigma_B)^{max}$ according to Table 2, that is,

$$t_B^{calc} = t_B + (2\sigma_B)^{max}. \qquad (3)$$

An expression was similarly obtained for determining the calculated value of the mean monthly total radiation

$$Q_c^{calc} = Q_c + (2\sigma_p)^{max}, \qquad (4)$$

where Q_c is the mean monthly total radiation determined by the literature,[2] kcal/m^2·h; $(2\sigma_p)^{max}$ is the magnitude of the error corresponding to the confidence of 0.95 defined by Table 3, kcal/m^2·h.

In cases where there are no data[2] on the mean monthly values of the total radiation, it is proposed that they be defined by the graphs (Figure 1) as a function of the northern latitude of the station.

In order to determine the radiation balance, correlations were established $R = f(Q_c)$ for various values of the surface albedo. Considering that in airport practice for roads two values of the albedo or values close to them are most frequently encountered--these are $a = 0.1$ (asphaltic concrete, black gravel) and $a = 0.2$ (concrete, soil cement, gravel, and so on are encountered); the correlations were established with a correlation coefficient of 0.96 between R and Q_c for $a = 0.1$(0.07 to 0.16) and $a = 0.2$ (0.17 to 0.34), which can be represented by the following expressions respectively:

$$R = 0.71\ Q_c - 18\ (\text{kcal/m}^2\cdot\text{h}), \qquad (5)$$

$$R = 0.70\ Q_c - 38\ (\text{kcal/m}^2\cdot\text{h}). \qquad (6)$$

The field observations and experiments performed at airports and roads of the Far North have permitted the establishment of the correlation between the parameter $(LE + B)$ and the total radiation Q_c. This function, which has the correlation factor 0.96, can be expressed by the equation:

$$LE + B = 0.57\ Q_c - 52 \qquad (7)$$

Processing the mean monthly values of the wind velocity by the methods of mathematical statistics demonstrated that the maximum possible deviation of their values from the mean during the summer is with a confidence level of 0.95 within the limits of 2 m/s. Considering that the wind velocity enters into Formula (1) indirectly in terms of the heat exchange coefficient, it is entirely admissible to take the mean monthly values of the wind velocity[2] as the values for calculation.

It has been proved experimentally that the coefficient of convective heat transfer with a pavement surface length >4 m in the direction of the air flow does not depend on the linear dimensions. On the basis of dimensionality theory, a formula has been derived to determine the convective heat-transfer coefficient, which is a function of the wind speed, the surface roughness, the coefficients of thermal conductivity, and the kinematic viscosity:

$$\alpha_k = \frac{0.00058 \cdot V^{1.15} \cdot h^{0.15} \cdot \lambda_B}{\nu_B^{1.15}} \qquad (8)$$

where V is the wind speed at the wind vane level, m/s; h is the roughness of the pavement, m; λ_B and ν_B are the coefficients of thermal conductivity and kinetic viscosity of the air respectively, kcal/m^2·h and m^2/s.

TABLE 3 Values of the Maximum Errors $(2\sigma_p)^{max}$ in Determining the Calculated Values of the Mean Monthly Total Radiation

Months	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
$(2\sigma_p)^{max}$	2	5	15	40	36	45	41	33	21	10	10	1

After substitution in (8) the averaged values λ_B, h, ν_B, we obtain

$$\alpha_k = 2.1\ V^{1.15}. \qquad (9)$$

In order to facilitate the calculations, Equation (9) is represented in the form of a graph (Figure 2). The points denoted by rectangles, which were obtained experimentally by E. D. Bondareva by the balance observations at Leningrad airport, were plotted on the same graph. As is obvious from the graph, the point distribution repeats the nature of the curve of α_k as a function of V, and it confirms the correctness of the derived formula for determining the coefficient of convective heat transfer.

Thus, the determination of the calculated surface temperature is made in the following sequence:

1. The initial data--the air temperature (t_B), the wind speed (V), and the total radiation (Q_C)--are determined from climatic data. In the case where there is no information in the references on the least squares values of the total radiation Q_C, it must be determined by the graphs (Figure 1).
2. The radiation balance (R) is determined by Formulas (5) or (6) as a function of the surface albedo.
3. The coefficient of convective heat exchange (α_k) is determined by graph of Figure 2 as a function of the wind speed (V).
4. By Formula (7), depending on the magnitude of the total radiation Q_C, the parameter ($LE + B$) is determined as a function of the magnitude of the total radiation.
5. The surface temperature is calculated by Formula (1).

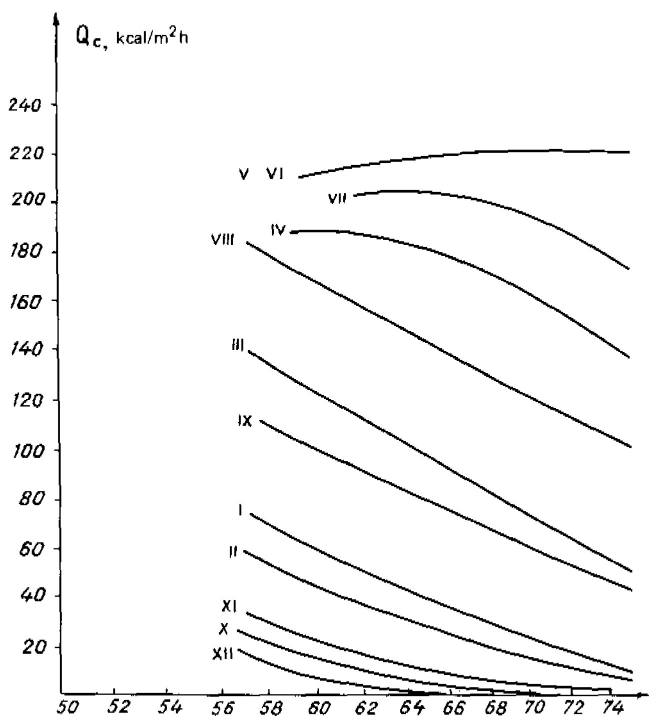

FIGURE 1 Mean monthly values of the total radiation (Q_C) as a function of the northern latitude (N), I, II... XII--months.

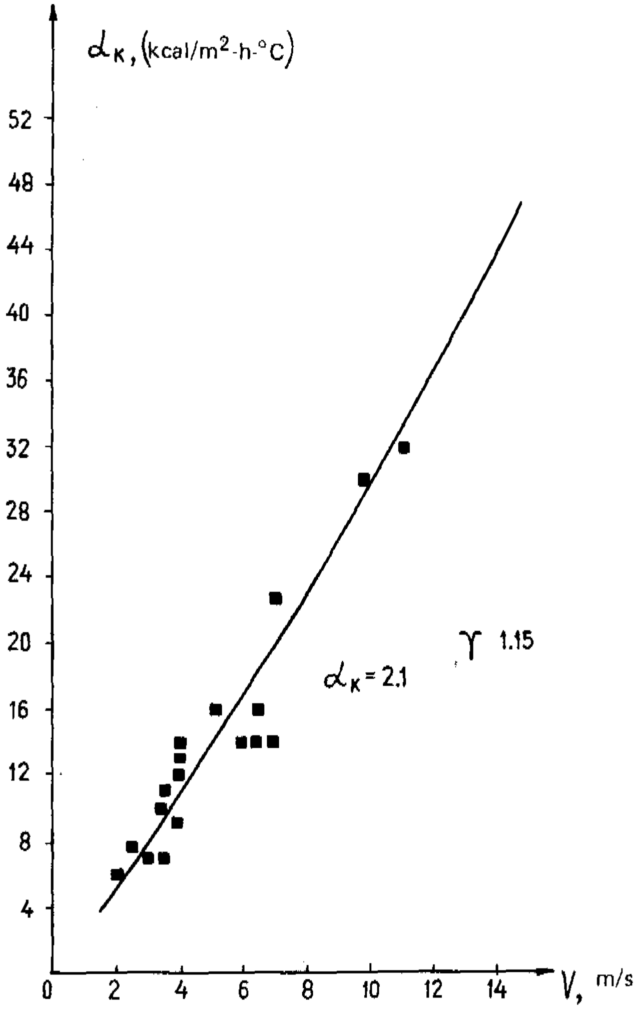

FIGURE 2 Convective heat transfer as a function of wind speed.

The developed procedure can be used to determine the surface temperature at any point in time and also its average values (diurnal, 10-day, monthly, yearly, and so on).

The divergence of the calculated values of the surface temperature of the pavement from those observed under natural conditions is within the limits of 0.3°C to 2.5°C.

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NUMERICAL SOLUTION OF PROBLEMS OF THE STEFAN TYPE FOR ATMOSPHERE-GROUND-STRUCTURE SYSTEMS

YU. S. PAL'KIN SibTsNIIS

With the development of digital computers and the development of effective finite-difference systems of approximating nonlinear boundary problems in mathematical physics, numerical methods of solving the equation of thermal conductivity in the presence of phase interfaces (the Stefan problem) have become extremely widespread.

At the SibTsNIIS Institute these methods are used in connection with calculations of the nonsteady-state temperature regimes in the two-dimensional multiphase atmosphere-ground-structure systems. The solutions obtained[1,2] are attained on a computer in the form of the Led-3 and Led-4 algorithms using the explicit, and implicit, finite-difference approximation methods, respectively.

In this report the indicated schemes are compared as applied to calculating the temperature regime of the frozen, freezing, and thawing ground under buildings and structures.

The finite-difference method provides for replacement of the investigated region by a discrete model, the nodes of which have the coordinates i, j, τ, and the set $t_{i,j,\tau}$ characterizes the desired solution of the equations of the type:

$$t_{i,j}^{k+1} = t^{k}_{i,j} + \frac{\Delta\tau}{C_{i,j}\Delta x_{o.k}\Delta y_{i,j}} \left(\frac{t^{k}_{i-1,j} - t^{k}_{i,j}}{R^{b}_{i-1,j} + R^{b}_{i,j}} - \frac{t^{k}_{i,j} - t^{k}_{i+1,j}}{R^{b}_{i,j} + R^{b}_{i+1,j}} + \frac{t^{k}_{i,j-1} - t^{k}_{i,j}}{R^{r}_{i,j-1} + R^{r}_{i,j}} - \frac{t^{k}_{i,j} - t^{k}_{i,j+1}}{R^{r}_{i,j} + R^{r}_{i,j+1}} \right) \quad (1)$$

(the explicit scheme), or

$$t^{k+\frac{1}{2}}_{i,j} = t^{k}_{i,j} + \frac{\Delta\tau}{2C_{i,j}\Delta x_{i,j}\Delta y_{i,j}} \left(\frac{t^{k+\frac{1}{2}}_{i-1,j} - t^{k+\frac{1}{2}}_{i,j}}{R^{b}_{i-1,j} + R^{b}_{i,j}} - \frac{t^{k+\frac{1}{2}}_{i,j} - t^{k+\frac{1}{2}}_{i-1,j}}{R^{b}_{i,j} + R^{b}_{i+1,j}} + \frac{t^{k}_{i,i-1} - t^{k}_{i,j}}{R^{r}_{i,j-1} + R^{r}_{i,j}} - \frac{t^{k}_{i,j} - t^{k}_{i,j+1}}{R^{r}_{i,j} + R^{r}_{i,j+1}} \right),$$

$$t^{k+1}_{i,j} = t^{k+\frac{1}{2}}_{i,j} + \frac{\Delta\tau}{2C_{i,j}\Delta x_{i,j}\Delta y_{i,j}} \left(\frac{t^{k+\frac{1}{2}}_{i-1,j} - t^{k+\frac{1}{2}}_{i,j}}{R^{b}_{i-1,j} + R^{b}_{i,j}} - \frac{t^{k+\frac{1}{2}}_{i,j} - t^{k+\frac{1}{2}}_{i+1,j}}{R^{b}_{i,j} + R^{b}_{i+1,j}} + \frac{t^{k+1}_{i,j-1} - t^{k+1}_{i,j}}{R^{r}_{i,j-1} + R^{r}_{i,j}} - \frac{t^{k+1}_{i,j} - t^{k+1}_{i,j+1}}{R^{r}_{i,j} + R^{r}_{i,j+1}} \right) \quad (2)$$

(the implicit scheme), where $\Delta x_{i,j}$ and $\Delta y_{i,j}$ are the space steps; $\Delta\tau$ is the time step; $R^{B}_{i,j}$ and $R^{r}_{i,j}$ are the thermal resistances between the junction of the model and the boundary with the adjacent block respectively in the directions s and y, $R^{B}_{i,j} = \Delta x_{i,j}/2\lambda_{i,j}\Delta y_{i,j}$, $R^{r}_{i,j} = \Delta y_{i}/2\lambda_{i,j}\Delta x_{i,j}$; $\lambda_{i,j}$ and $C_{i,j}$ are the coefficients of thermal conductivity and the volumetric specific heat capacity of the block in the model, respectively.

In the first case (the Led-3 algorithm) the solution is stable if

$$\Delta\tau \leq \frac{c_{i,j}\Delta x_{i,j}\Delta y_{i,j}}{4\lambda_{i,j}}, \quad (3)$$

and the calculation is made in the direction $i = 1, j = 1, k = 1..., i = 1, j = \max, k = 1, ..., i = \max, j = \max, k = 1, ..., i = \max, j = \max, k = \max$.

In the second case (the Led-4 algorithm), the stable scheme for the longitudinal-transverse run of the fractional step method is used.[3] According to this method, in the first step k, the approximation of t with respect to x of the second derivative of the equation of thermal conductivity is implicit, and with respect to y, explicit; in the second step $k + \frac{1}{2}$, the opposite is true, that is, the primary calculation is made on making the transition from the step k to $k + 1$ with the auxiliary step $k + \frac{1}{2}$, and an increase in the approximation error in the direction y is compensated for by a decrease in the error in the direction x.

When solving problems of the Stefan type, the presence of the water-ice-water phase transitions at the freezing point t_3 accumulates definite restrictions per time step inasmuch as an increase in $\Delta\tau$ the probability of the "catching" of t_3 at the nodes of the grid decreases. The given fact significantly reduces the advantage of the implicit system over the explicit one. In addition, the approximation of t with respect to the implicit system requires a larger volume of calculations than with respect to the explicit system, and it becomes economically disadvantageous (according to the machine time expenditures) with a value of $\Delta\tau$ less than the fivefold value of the time step of the explicit system.[3]

It was demonstrated[4] that if the system is stable and approximates the differential equation with known order of approximation n, then the order of accuracy coincides with the order of the approximation. The absolute error ε in the solution t (Δx, Δy, $\Delta\tau$) can be estimated[5] in comparison with the solution $t(2\Delta x, 2\Delta y, 4\Delta\tau)$ from the expression

$$\varepsilon \simeq \frac{t(\Delta x,\Delta y,\Delta\tau) - t(2\Delta x,2\Delta y,4\Delta\tau)}{2^n - 1}. \quad (4)$$

The errors in the solutions of a number of problems found by this principle turned out on the average to be 0.01°C (for $\Delta x = \Delta y = 0.25$ to 0.5 m and $\Delta\tau = 6$ to 24 h).

One of the programs implementing the Led-3 algorithm was prefaced by a procedure for searching for the optimal step size $\Delta\tau$ from the condition of the required precision of the solution. It turned out that when $\varepsilon = 0.1$°C this step size is a value that is a multiple of 24 h and satisfies the stability condition from the top (3). Thus, when $\Delta x = 0.4$ m, $\Delta y = 0.6$ m, $c = 400$ kcal/m^3·deg, and $\lambda = 2.5$ kcal/m·deg·h, the optimal value of $\Delta\tau = 8$ h. A decrease in this value (6, 4, 2, 1 h) is felt in the results of the solution only in the second or third decimal place.

Considering everything, it is necessary to draw the conclusion that the explicit system is preferable for solving the majority of Stefan problems.

A comparison of the numerical solutions with observations in the experimental sections and the result of simulation on the hydrointegrator of V. S. Luk'yanov[2,6] demonstrated good convergence of the compared parameters: the ground temperatures, the depth of freezing-thawing, and the rate of advancement of the phase transition boundary (see Figure 1). The fitness of the algorithm was checked on another class of

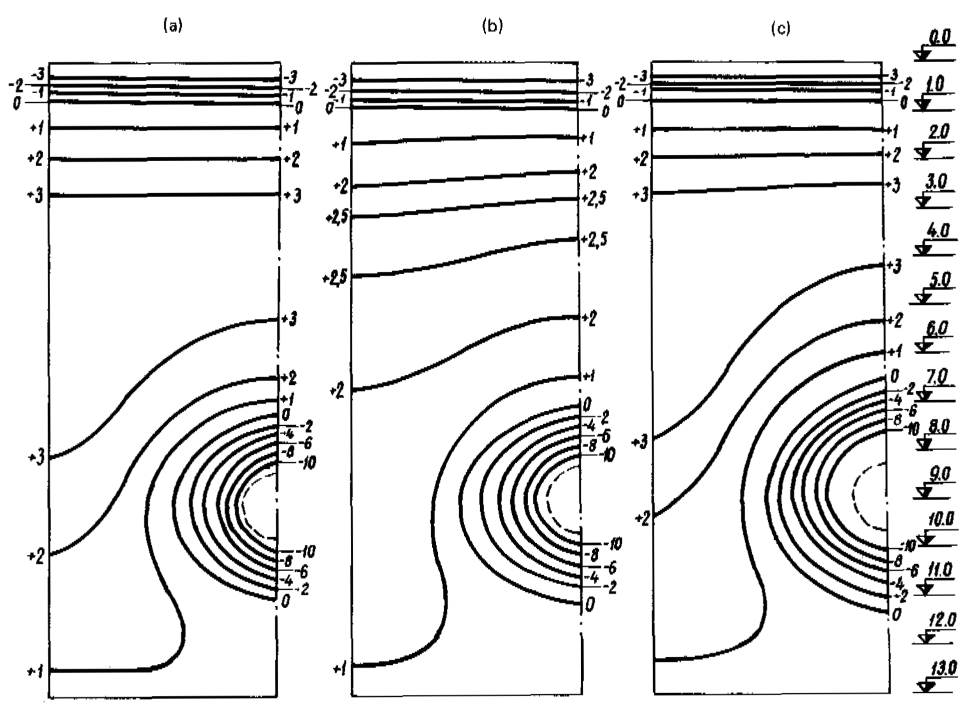

FIGURE 1 Temperature field in an embankment for a water pipe according to the situation on 25 December 1971 obtained (a) by calculation on a digital computer, (b) observations, and (c) simulation on a hydrointegrator [hydraulic analog computer].

thermophysical problems--calculation of transient regimes in solids. The relative error in selecting the numerical solutions of problems of Pekhovich and Zhidkihk[7] turned out not to be more than 1.5 percent.

The algorithm was written in the ALGOL-60 language and has now been tested in many kinds of calculations of the thermal conditions of the bases of embankments, dams, buildings, bridge supports and electric power transmission lines, water lines, buried pipe lines, gas lines, and so on. The annual cycle of the temperature field formation in the investigated region of 120-140 nodal points is calculated in 15-18 min on the M-220 computer (depending on the discreteness of the three-dimensional space of the parameters); the calculation of a 7- or 8-yr cycle lasts 40-50 min.

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*[Available in translation by the Consultants' Bureau "Soil Mechanics and Foundations."]

PROCESSES OF HEAT AND MOISTURE EXCHANGE IN FLOODING OF MEADOWS IN CENTRAL YAKUTIA

P. P. GAVRIL'YEV, I. S. UGAROV, AND ZH. I. RYSAKOV Institute of Geocryology of the Siberian Department of the USSR Academy of Sciences, A. A. MANDAROV, Igarskaya Scientific Research Cryologic Station

This report is a generalization of the results of the 6-yr thermophysical heat balance, water balance, and other studies during the flooding of meadows in central Yakutia.

In the case of basin irrigation, both the heat and moisture exchange of the soil with the ground layer of the air and the processes of heat and moisture transfer in the seasonally thawing layer of ground vary significantly. For example, depending on the degree of dryness of the vegetative period, the surface albedo of the meadow land decreases by 5 to 10 percent, and the effective radiation by 10 to 20 percent. The radiation balance increases by 5 to 25 percent. During drought years, the radiation effect of the irrigation is observed during the entire vegetative period; during normal years, it is observed for 15 to 20 days, and during the wet years, only 10 to 15 days. The maximum possible radiation effect during irrigation in central Yakutia is estimated at 25 to 30 percent. For the irrigated meadows, the expenditures of heat on evaporation from the soil increased by 1.5 to 2 times and constitute about 75 to 85 percent of the radiation balance. The turbulent heat exchange of the soil with the atmosphere decreases by 1.5 to 2 times, and its proportion is about 5 to 20 percent of the radiation balance.

The thawing of the soil under the water (during the period of flooding of the basins) depends on the initial time, the duration and depth of flooding, the initial moisture (ice) content of the frozen ground, and so on. For early flooding periods of the meadows, especially in the first half of May, the layer of irrigation water serves as a thermal insulator and protects the thawed layer of soil from its secondary freezing during frosts. This significantly accelerates the melting of the soil. With a shallow depth of flooding (0.5 m to 0.7 m), the thawing under the water by comparison with the unirrigated meadow increases by 10 to 20 percent; with a large depth of flooding (1 m to 3 m), it lags by 10 to 13 percent with the same values of the initial soil moisture content. The field studies and the results of simulation of a hydrointegrator demonstrated that if the initial soil moisture content is less than 10 to 13 percent, then during irrigation acceleration of the thawing by 10 to 20 percent takes place. For an initial moisture content of more than 13 to 15 percent, no significant change in depth of thawing takes place during irrigation.

As a result of the basin irrigation, the thawed layer of soil accumulates water to its complete moisture capacity. Under the effect of irrigation, the specific heat capacity, the thermal conductivity, and the thermal receptivity of the soil increased by 1.5 to 2 times, reaching 600 to 900 kcal/deg·m^3, 0.70-1.15 kcal/m·h·deg, 25 to 28 kcal/m^2·h 1/2 deg [sic].

The heat flux into the soil is on the average 9 to 11 percent of the radiation balance. The greater part of this total heat flux (60 to 70 percent) into the soil going through the surface is expended on the phase conversion of the pore ice during thawing of the soil, and only 15 percent goes for heating the thawing layer.

Significant variations of the magnitudes of the heat and moisture exchange of the soil with the atmosphere, the moisture, the thermophysical properties of the soil, and the thermal flux into the soil occurring with basin irrigation determine the thermal regime of the seasonally-thawing layer of the soil. The heating of both the root layer and the layer of seasonal thawing of the soil takes place slowly. On the average by the time of sprouting of the grass (at the beginning of vegetation), the depth of thaw reached 25 cm to 35 cm, and the end of May, 55 cm to 65 cm. At the beginning of July (the beginning of the haying season), it reaches 120 cm to 130 cm, and by the end of summer, 160 cm to 180 cm. The active temperature (above 10°) penetrates to a depth of 20 cm only in the middle of July, and to a depth of 40 cm at the beginning of July, that is, at the beginning of the haying season. The maximum depth of penetration of the temperature of 5° is 100 cm to 110 cm. Under the effect of the basin irrigation, the surface temperature of the soil drops, on the average, by 2°-5°, and in clear weather during the daylight hours, by 10°-15°. In the upper layer of soil from 0 cm to 30 cm, the temperature drops by 2°-3°, and in the thawed layer of the soil, below 40-50 cm, the temperature rises by 1°-3°, that is, the irrigation prevents extraordinary overheating of the soil surface and attenuates the negative effect of the frozen undersoil for growth and development of the grass (Figure 1a). Thus, basin irrigation improves the water-thermal regime of the seasonally thawing layer of soil. In some cases, for example, with high initial moisture of the frozen soil and with inefficient irrigation regimes (deep, continuous flooding during the early part of the year), the thermal regime of the soil becomes worse: There is a significant decrease in intensity of thawing, cooling, and supersaturation of the entire root layer (Figure 1b).

During irrigation, the thickness of the layer of soil participating in the hydrologic cycle with the atmosphere decreases by 10-20 percent, amounting on the average to, e.g., 10 to 20 percent, 60-70 cm. Here, about 80 percent of the water is consumed from the upper 40-50-cm layer, which is explained by the excess of easily-accessible moisture in the upper layers of the soil.

It has been established that the accumulation of moisture in the soil by the beginning of vegetation is primarily affected by the fall precipitation. The increase in moisture reserves of the upper layer of the soil also corresponds to migration of the moisture during the fall-winter freezing. The growth of the moisture content of the upper meter layer in this case is more than 20 mm, reaching 100 mm in the presence of groundwater.

On the basis of analysis of the heat and moisture exchange in the soil and the ground layer of

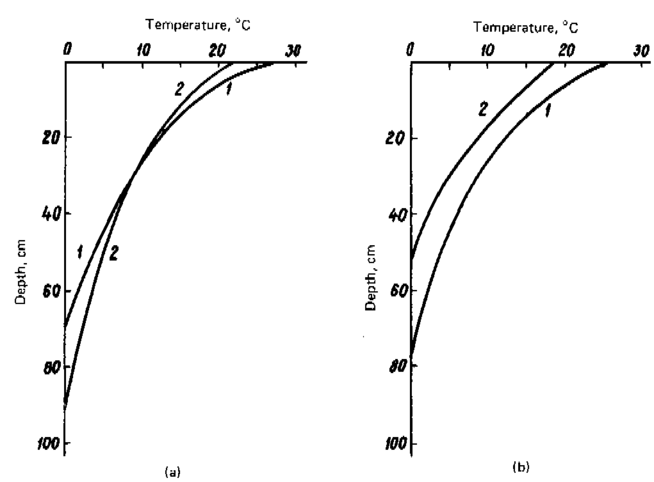

FIGURE 1 Temperature distribution in the thawed layer of soil. (a) Yelechey, 5 June 1967; (b) Sytygan, 1 June 1971. 1--unirrigated meadow, 2--irrigated meadow.

the air, a calculation scheme is proposed for determining the regime of basin irrigation of the meadows in central Yakutia. In the permafrost zones, the optimal irrigation norm V_{op} corresponds to a strictly defined depth of thawing of the soil H_{op}, which must be achieved at the time of removal of the floodwaters from the meadows (the time of drainage of the basin). Considering this, a regional calculation formula was obtained that relates the required depth of thawing to the irrigation norm:

$$V_{op} = [3 + 0.175(\Sigma T_{>5^\circ} + \Delta T)] - Oc + W_{CRM}$$

$$- W_{init} = 10(0.8 - \gamma_{init})AH_{op},$$

where V_{op} is the optimal irrigation norm, mm; $\Sigma T > 5^\circ$ is the total air temperature above 5° for the calculated period in degree-days considering the air temperature correction for the seasonal behavior of ΔT; Oc is the precipitation, mm; W_{CRM} is the moisture, mm in the broken capillaries; W_{init} is the initial soil moisture, mm; γ_{init} is the initial moisture stored in the soil in proportions of a unit of the porosity; A is the porosity of the soil as a fraction; and H_{op} is the required depth for thawing of the soil, cm.

This formula takes into account the factors determining the irrigation regime such as the demand for heat and moisture by the plants, meteorological conditions, and the hydrophysical properties of the local soil.

As the calculations demonstrated, the depth of thawing of the soil during the cold part of the year decreases by comparison with the average year by 20-25 cm--at the beginning of thawing--and to 7-8 cm at the end of June. During a warm year, on the contrary, it decreases correspondingly, within the limits of 5-20 cm and 8-10 cm. The time of reaching a defined thawing depth can vary within the limits of 4-5 days, depending on the meteorological conditions. Beginning with this, it is necessary to take into account the expected meteorological conditions of the future vegetative period, not only when determining the irrigation norm, but also when determining the time of emptying of the basin.

PART VIII

Problems in the Science of Permafrost

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V. M. PIGUZEVA, *Editors*

OPENING ADDRESS

P. I. MEL'NIKOV Chairman of the Organizing Committee and Director of the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences

Ten years have passed since the First International Conference on Permafrost was held in Illinois in the United States. At that conference a summation was made of the results of permafrost research spanning almost a quarter of a century. Everyone attentively studied the proceedings of the conference and used them in research and in practice.

During the past decade there has been further intensive development of science in all fields of knowledge, including geocryology. Everywhere, people have been delighted by the progress in space research achieved by the scientists and engineers of the Soviet Union, the United States, and other countries. The studies of the moon using Soviet equipment and lunar probes, the multiple landings on the moon by American astronauts, and the space studies of Mars and Venus have made it possible to obtain a series of valuable data on the nature of the distant planets. For centuries these planets have attracted the attention of scientists seeking to untangle the mysteries of the cosmos.

The time has come for us, the geocryologists, to participate actively in the space studies of our nearest satellite, the moon, where undoubtedly permafrost is present. The study of the Earth and of near space from Earth satellites for the investigation of natural resources and geocryological and geographic research is of great interest.

We are living in an age of rapidly accelerating scientific and technical progress. The rapid expansion in the production of material goods is giving rise to increasingly intensive exploitation of natural resources, with the result that the effect on the environment is ever more pronounced.

Among the multiplicity of problems facing mankind in its attempts to make maximum use of natural resources, the problems of the North stand alone. The northern environment is exceptionally sensitive to human activity and is easily disturbed. Its protective functions are limited; everything is in a delicate state of balance here.

Unfortunately, the problem of making efficient use of natural resources and also the problem of the interrelation and interaction of man and nature are not always being tackled correctly. It is no longer possible to develop the Far North by adhering to antiquated and fallacious notions that the existence of vast tracks of land and innumerable rivers there makes it unnecessary to be concerned with their purity and proper use. There is an urgent need to achieve a marked expansion of the frontiers of knowledge in the field of environmental conservation and to make efficient use of natural resources. This would make it possible to obtain a much better understanding of the overall situation with respect to the preservation and development of life on this planet.

Geocryologists are especially interested in observations in recent years that point to an increase in the carbon dioxide content in the atmosphere. Calculations made by many researchers indicate that an increase in the carbon dioxide content may cause a global rise in the temperature with all of the consequences arising therefrom, right down to permafrost degradation and melting of the ice. However, despite the increase in the carbon dioxide content in the atmosphere, following the slight rise in the temperature observed during the period 1900-1945, it has now begun to drop.

The dust content of the atmosphere has a more complex, but also a more global, effect on the climate. Calculations have shown that the amount of dust rising annually into the air approaches many millions of tons. The settling of dust on the ice intensifies its melting. The accumulation of dust in the atmosphere alters the intensity of the solar radiation, which could lead to the development of glaciation conditions in the presence of a constant and intensive increase in the atmospheric dust content.

This would give rise to a multiplicity of new problems requiring profound study. Geocryologists must consider these changes and determine the extent to which they influence the climate and the permafrost regime.

The problem of the study of nature and how to protect the environment is becoming increasingly worldwide in its scope: All countries are attaching special importance to it. There is no doubt that through the common efforts of a number of countires this problem will be resolved. The view of nature as an exceptionally intricately balanced system must receive proper recognition in our research.

In preparing for the Second International Conference on Permafrost, we have endeavored to arrange the work so that during the 5 days of the plenary sessions all of the basic achievements in the field of geocryology will be heard and discussed with respect to main areas of research. Accordingly, the Soviet Union was requested to prepare eight sectional papers, and the United States and Canada an equal number of papers. In addition, a paper entitled "Problems and Research Areas in the Field of Environmental Preservation in the North" was prepared and has been included in the conference agenda in view of the special urgency of this problem.

Altogether, 1 general paper and 17 sectional papers will be heard at the conference, the authors of which have compiled outline summaries with respect to the various branches of our science and have analyzed its status. They have also discussed the basic achievements and mapped out the future course of development of the research.

A total of 204 papers have been received from the Soviet scientists and a number of foreign participants at the conference. They have been grouped into seven sections. The delegates from the United States and Canada have prepared 80 papers. Thus, 302 papers have been submitted for presentation.

After completing the plenary sessions there will be scientific excursions the itineraries of which pertain to three different topics. Guidebooks in Russian and English have been prepared for these excursions.

The conference delegates will not only hear the plenary reports, but will also be briefed on the published papers and participate actively in discussing them. Arising from the need to direct the work along certain lines and hold discussions on the most urgent problems of contemporary geocryology, the authors of the papers have prepared these questions, which have been reproduced and distributed to the delegates.

In May of this year F. E. Are and I took the opportunity to visit the United States in order to familiarize ourselves with the research situations there and with the achievements of our American colleagues in the field of permafrost studies. During our 32-day stay in the United States we visited both the East and West coasts and also the central regions of the country and became particularly well acquainted with Alaska. We were warmly received and became thoroughly familiar with the laboratories, the laboratory investigations of permafrost problems, and the results of applying the scientific achievements. We were briefed on the progress made in preparing for the building of an oil pipeline that will cross the permafrost region of Alaska from the coast of the Arctic Ocean to Valdez on the Pacific. It is very pleasing to note that the remarkable improvement in the relations between the Soviet Union and the United States was perceptible everywhere and that favorable possibilities have arisen for the establishment of closer scientific cooperation between the geocryologists of the Soviet Union and the United States.

Last year an agreement on scientific cooperation was concluded between the Soviet Union and Canada in which provision was made for the joint development of several areas of permafrost studies. We must find ways of combining our efforts so as to ensure further integrated progress in solving the most complicated problems in geocryology. This will serve to expedite scientific and technical progress in the development of the northern territories of our countries.

Again, it gives me great satisfaction to welcome you, respected delegates and guests, to this impressive international forum of geocryologists and to wish you success in your efforts and a pleasant stay in the Soviet Union. In the name of the Organizing Committee permit me to open the Second International Conference on Permafrost.

EDITOR'S NOTE: Also presented were welcoming addresses from the Presidium of the Supreme Soviet and the Council of Ministers of the Yakut ASSR; M. P. Gabyshev, vice chairman of the Council of Ministers of the Yakut ASSR; the Presidium of the USSR Academy of Sciences; and M. A. Lavrent'yev, vice president of the Academy of Sciences and chairman of the Siberian Division of the USSR Academy of Sciences.

ADDRESS

T. L. PÉWÉ Leader of the United States Delegation

In the name of the American participants, I express our appreciation to the Soviet organizing committee and in particular, to Doctor Mel'nikov. We value the warm reception given us, but you have arranged weather that is too hot for a meeting of geocryologists.

Before we proceed with hearing the papers let us have a minute of silence in memory of the leading geocryologists of the world who have died since the First International Conference on Permafrost was held in the United States in 1963: Saltykov, Koloskov, Kachurin, and Shimanovskiy (USSR); Jan Dylik (Poland); Robert Berg, K. Terzaghi, and John MacAnerny (USA). The name of

Simon V. Muller, which I mention last, has great significance for geocryologists and, in particular, for me personally.

Professor Muller was born in Russia in 1900 and came to the United States at an early age. He is especially well-known in North American as the man who coined the word "permafrost" in 1943. He supported the study of permafrost, wrote the first book on the subject in English, and was an honored guest at the First International Conference on Permafrost. He has also played a special role in my life: He was my teacher and advisor at Stanford University.

P. I. Mel'nikov, who opened the conference, is an internationally recognized eminent scientist in the field of permafrost. He has devoted many years to the study of permafrost in the USSR and has been invited for preliminary studies of permafrost both in the United States and in Canada. He has received numerous honors for his well-known scientific and organizational work. He is a Corresponding Member of the USSR Academy of Sciences, an honored scientist and engineer of the RSFSR, Director of the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences and Chairman of the Organizing Committee of the Second International Conference on Permafrost, and it is with great satisfaction that I turn the meeting over to him.

BASIC RESULTS OF RESEARCH IN THE FIELD OF PERMAFROST FOR THE PERIOD FROM 1963 TO 1973 AND THE PROSPECTS FOR ITS DEVELOPMENT

P. I. MEL'NIKOV Permafrost Institute of the Siberian Division of the USSR Academy of Sciences

During the last decade geocryological research has continued to progress and has gathered momentum in the scientific institutions and universities of the Soviet Union, the United States, Canada, the Mongolian People's Republic, and a number of other countries. These years have been characterized by intensive development of the natural wealth of the permafrost region, which occupies about half of the territory of the USSR and about 25 percent of the territory of all of the world's dry land, and, together with the region of seasonally frozen ground, about half of it.

Mineral deposits, including diamonds, gold, tin, nickel, copper and rare elements, are concentrated in the permafrost region. Enormous oil and gas deposits have been discovered and are being exploited in western Siberia, the reserves of which are already accounting for a significant proportion of the total volume of output in our country. Important gas reserves have been discovered in Yakutia and in the northern part of the Krasnoyarsk Kray.

The northern territory also has rich deposits of iron ore and coal. In the near future the exploitation of high-quality iron ore and coking coal will begin in southern Yakutia. Recent years have marked the discovery of major oil and gas fields in northern Alaska and Canada.

Thus, the permafrost regions are preeminent with respect to the concentration of valuable and unique mineral deposits. The industrial development of this natural wealth is one of the most important national economic problems facing a number of countries.

In the northern part of the Soviet Union a program for the intensive development of the highly valuable natural resources required for the growth of the economy and for the country's scientific and technical progress as a whole is currently under way. The proportion of these resources in the total output of minerals is constantly increasing. The extraction and refining industry is undergoing intensive development. Hence, the volume of construction and assembly operations must increase by several times. A great deal of attention is being devoted to the development of agriculture in the northern regions and to increasing the productivity of the fields and meadows in order to provide food for the populations of these regions from local resources.

All of this is giving rise to heightened interest in geocryological research, which is making a significant contribution to the economic development of the permafrost region.

During the last decade (1963-1973), geocryological research has been continuously expanding. The scope of this research is indicated by the fact that during this period, in the USSR alone, more than 100 monographs and collections, procedural instructions, and reference aids on the various problems of geocryology have been published. Three texts on geocryology as a whole or individual branches of it have been published for the institutions of higher learning. In 1966 the Eighth Interdepartmental Meeting on Geocryology (Permafrost) was held in Yakutsk, the reports of which were published in eight issues. In 1970 an All-Union Conference on Permafrost was held in Moscow. In 1969 an International

Symposium on Paleogeography and Periglacial Phenomena of the Pleistocene was held in Yakutsk and in Moscow.

National conferences on permafrost have been held in Canada and the United States, where the results of research carried out were summarized, the achievements were noted, and the course of future research was planned.

The progress achieved indicates that geocryology (permafrost studies) has risen to a new, higher level of development and, along with other natural sciences, is conducting basic research without which it would be impossible to solve major economic problems.

In the published reports and papers that were included in the program of the plenary sessions of the conference, the developments in geocryological science during the last 10 yr are summed up, the contemporary status of the science is discussed, and the course of future research is planned.

During the forthcoming research period it will be necessary to concentrate attention on the need for achieving a marked improvement in the efficiency of scientific creativity by upgrading scientific planning and prediction and also for improving the methods and facilities used in scientific research through cooperation and specialization among scientific collectives, by increasing the availability of the materials and experimental facilities needed by science, by further developing and strengthening the relations between scientists, and by arranging for the practical application of scientific discoveries at the earliest possible time.

Let us now review the achievements in geocryology during the last decade from the standpoint of its main subject areas.

GENERAL GEOCRYOLOGY (PERMAFROST STUDIES)

General geocryological research embraces all aspects and modes of development of the permafrost region. Recent years have been characterized by an intensification of regional geocryological research, primarily in those areas that are of greatest interest from the standpoint of their integrated study and mapping, and that are destined for economic development.

In our country, the integrated approach to research and working contacts with geological and production organizations have given rise to the possibility of using deep-well drilling data to obtain necessary information on the composition and structure of permafrost. One-time and annual temperature observations in deep wells have permitted determination of the thickness of the permafrost, the temperature field, and thermal fluxes to within a sufficient degree of accuracy. It has been established that saltwater and brine, which have come to be called cryopegs [bodies of liquid saline water below 0° C], participate in the formation of permafrost. The presence of a 1½-Km layer at negative temperatures, established from two wells in the northern part of Yakutia, is associated with the cryopegs. This remains the thickest known layer of frozen earth materials with negative temperatures anywhere in the world.

The present day concepts of the permafrost region of the Laptev Sea shelf have been substantiated and refined. This makes it possible to visualize the permafrost region of the Arctic Sea shelves as a whole.

The problem of the formation of permafrost as a function of the geological structural, tectonic, and zonal factors and also in connection with altitude belts has been further elaborated. Generalizing monographic summaries have appeared for western and central Siberia, Chukotka, Zabaykal'ye [trans-Baikal], Pribaykal'ye [the near-Baikal region], central Yakutia, and other regions.

A great deal of attention has been given to the study of cryogenic physical and geological phenomena. An enormous amount of factual data has been gathered which shed light on the peculiarities and the extent of these phenomena. Quantitative studies of heaving, fissure formation, solifluction, and thermokarst have been extensively conducted. These studies cannot but refine and extend our concepts of the mechanism involved in the development of cryogenic phenomena, in accordance with the contemporary level of development of the exact sciences. The basic laws governing the shaping of the relief and the reworking of the shores of natural bodies of water and reservoirs consisting of permafrost have been elucidated. Several methods of estimating the intensity and predicting the reworking of these shores have been developed.

A study has been made of the mechanism that leads to the formation of [naleds] groundwater icing. Also, a procedure for calculating the melting rate of the naled that takes into account the surface albedo of the ice and the openness of the horizon has been developed for plains with a continental climate.

The large amount of data on seasonal freezing-and-thawing has also been correlated. A procedure has been developed for determining the quantitative characteristics and calculating the depths of seasonal freezing and thawing of earth materials as a consequence of the radiative heat balance of the surface and the geological and geographic factors. Also, studies have been made of problems pertaining to the dynamics and history of development of permafrost in western Siberia and in other areas.

A great deal of attention has been devoted to the compilation of geocryological maps. In the last 3 yr a 1:5,000,000 scale geocryological map of the Soviet Union and a number of regional maps have been produced. The maps compiled by the Soviet specialists are outline maps. They give a geocryological description of the entire territory of the USSR. The outline map has been recognized as satisfying to the maximum possible degree the requirements of specialists of various types. It must depict the set of characteristics that give the fullest possible representation of the most important features of the mapped area while remaining clear and readable.

The existing geocryological outline maps have been compiled on various principles, and they are therefore not comparable with respect to content. At the present time the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences is working on permafrost maps based on formations. On small- and medium-scale maps formations are being isolated and depicted that, while having a known uniformity within them, are differentiated by the ratio of the genetic types and facies of the deposits and by the composition and cryogenic structure of the latter. The formational principle of geocryological map compilation has made it possible to increase the amount of information provided and at the same time to make it more meaningful.

A comparatively new research field, cryolithology, has been successfully developed. It was established by Soviet scientists in the 1950's and has been recognized by many specialists. The basic achievements in cryolithology can be summarized as follows: The main features of cryolithogenesis--i.e., rock formation in the permafrost region, have been defined. These features are closely connected with the interaction between the geological and cryogenic processes, the nature and character of which are determined by the environment and by the facies conditions of sediment formation.

The geological criterion for syngenetic freezing of deposits has been found to be the degree of diagenesis of the deposits at the time of their transition to the permafrost state. The specific attributes of these deposits have also been demonstrated. It has been proved that the formation of the basic cryogenic structures, ice wedges, and soil wedges takes place under specific facies conditions and is subject to geological laws. The zonal peculiarities of the development of cryogenic phenomena in various deposits have been revealed.

Cryolithological research has made it possible to substantiate a conclusion that is of importance to Earth sciences, namely the continuous existence of a permafrost region in the Earth's crust throughout the greater part of the Quaternary.

One of the basic methods of cryolithology--permafrost facies analysis--has been improved. The guiding cryogenic structures have been isolated with respect to syngenetically frozen subaqueous and subaerial deposits. The permafrost facies technique has been successfully applied in cryolithological research in various parts of Yakutia.

During the last decade the extent of geocryological studies of North America has increased significantly. In Canada and the United States, several major reviews of the status of geocryological research have been published, besides which, problems in geocryology have been discussed in numerous reports, articles, and monographs. Studies of permafrost and the phenomena associated with it have not been confined to the northern part of Canada and Alaska. They have also been made for the Alpine areas of Canada and the United States. The boundaries of the continuous and discontinuous permafrost regions have been defined and correlations have been established between the permafrost and the conditions of its formation. It has been found that the earth materials in the shallow water between the arctic islands of Canada and along the north coast of Canada and Alaska are in a frozen state. Underwater frost mounds have been detected on the shelf of the Beaufort Sea. A great deal of attention has been devoted to the study of ground ice, especially horizontal layers of thick massive, polygonal-wedge, and injected ice. Many new data have been collected that make it possible to reconstruct the geological conditions of the Pleistocene.

During geocryological mapping of the United States and in Canada, in addition to the usual geographic, geological, and geophysical methods of determining permafrost, recently infrared color and radar surveys from the air have been used with varying degrees of success. A start has been made on studying the applicability for geocryological purposes of data obtained by artificial Earth satellites.

In 1965 a 1:2,500,000 scale outline geocryological map of Alaska was published, and in 1967, a 1:7,500,000 scale map of Canada.

THE GROUNDWATER OF THE PERMAFROST ZONE AND ITS INTERACTION WITH THE PERMAFROST

The last decade has also been marked by significant progress in the study of groundwater in the permafrost zone: its composition, conditions of formation, and interaction with the permafrost. In the Soviet Union the extensive development of various types of talik water has been discovered. These include fresh, potable water, and they have been found both in the folded mountainous regions and under platform conditions.

A study has been made of cryopegs discovered within the permafrost region. The composition and high mineralization of the cryopegs is a direct consequence of the processes of cryogenic metamorphism of groundwater. Great importance is attached to these processes in the development of the overall hydrochemical zonation of the vast area comprising the permafrost region, within which not only have the traditional hydrodynamic zones of free, difficult, and very difficult water exchange been isolated, but also a special zone of complex water exchange and, corresponding to it, a zone of water that is variegated in its composition and mineralization because of the cryogenetic processes.

The development of the theory of cryogenic metamorphization and exchange reactions in the water-rock system has made it possible to explain the widespread occurrence of sodium-containing water in subpermafrost aquifers, including those consisting of carbonaceous rock, which are characterized by calcium containing water outside the frozen zone.

An analysis of the groundwater resources of the frozen zone has revealed well-defined zonal and regional patterns of distribution and also an association with climatic factors and the thickness and type of discontinuity of the permafrost. The subsurface runoff reaches the least magnitude, i.e., tenths and hundredths of a liter per second per square kilometer, in the central regions of the Yakut artesian basin where, while thickness of the permafrost is considerable, through-taliks and atmospheric precipitation are minimal.

The decisive factors in the accumulation of the natural resources of groundwater in mountainous areas are the thick series of boulder-pebble and sandy deposits filling the tectonic depressions and the overdeepened sections of the river valleys or the foothill trains and debris cones of which the latter are composed. These are the deposits that form intermediate reservoirs and receive the surface water from the taliks below the riverbeds and the subpermafrost water ascending along the fracture zones. The considerable quantities of water entering here and the speed of movement of this water prevent these reservoirs from freezing. Groundwater travels through them and at the points where the transverse section diminishes; relaxation takes place; and high-discharge springs appear, sometimes forming very large polyn'yas and, in the foothill trains, naled lines. The extent of the naled lines surrounding the Momo-Selenyakhskaya depression on the south is more than 250 km, and the discharge rate of the springs associated with it is measured in hundreds and thousands of liters per second.

The studies connected with working out problems relating to the hydrogeological zoning of the permafrost region are of scientific and practical interest. The integrated approach to solving this problem has made it possible to reanalyze the permafrost-hydrogeological data and to depict them on special maps. It is evident that there is a possibility of significantly increasing the information content of the cartographic data.

A regional estimate has been made of the groundwater resources that for the alpine permafrost regions has become possible as a result of the development of special methods of separating the icing rivers and estimating the icing source reserves. The statistical processing of the data on the distribution of icings in the mountainous regions of the Asian part of the USSR has made it possible to arrive at a more precise definition of the nature of the permafrost-hydrogeological zonation in relation to altitude and to clarify the role and the significance of various natural factors in its formation.

The results of the studies of the groundwater in the permafrost zone of the USSR have been discussed most completely in the general papers, including the regional hydrogeological monographs encompassing the entire permafrost region. Not only do these papers discuss the conditions of propagation, formation, and use of groundwater of all types having regard to the latest factual data. They also contain essays pertaining to the engineering and geological characteristics of the regions. A large group of researchers working in various institutions in the Soviet Union participated in the task of compiling the Siberian monographs. This cooperation of specialists in a study executed from a single plan, and involving a single procedure that gives scope for broad creative initiative, has yielded positive results in terms of the complexity, variety, and novelty of the approaches to the solution of the various hydrogeological and geological engineering problems.

The published monographic summaries satisfy the stringent requirements imposed by science and practical work on general regional studies, and they form a major contribution to the scientific and technical literature.

During the last decade the studies of permafrost groundwater in North America have been expanded significantly. Between 1960 and 1965 groundwater and naleds began to be studied in combination with hydrological surveys in the various river systems of Alaska. A further impetus to hydrogeological research was given by the oil and gas exploration work, the construction of pipelines, and other forms of engineering activity in the northern part of Canada and Alaska. It was discovered that the most favorable conditions for groundwater recharge exist on the arctic coast of America, where the hydrologic cycle is complicated by thick layers of permafrost. An abundance of high-quality groundwater distinguishes the mountain regions, where the spring flow rates along the tectonic fracture lines or near karst caverns reach 120 to 1,140 liters/s. In connection with the problem of supplying water to arctic populations, studies are being made of the groundwater in loose, primarily alluvial, deposits. Problems pertaining to the interaction of seawater and groundwater in the coastal zone and the effect of the permafrost on these processes are also being worked out.

PHYSICAL CHEMISTRY AND MECHANICS OF PERMAFROST

A major part of the plans of the scientific institutions is taken up by theoretical and experimental work on the physics and mechanics of frozen earth materials and ice. Attention is being devoted mainly to the general laws of the physics and mechanics of permafrost and to detailed investigation of the mechanism of the cryogenic processes on the higher molecular-structural level. Much that is new has been introduced into the investigation of the tensile properties of permafrost. The long-term failure of the soil is regarded as a thermally active process, that is, failure of permafrost is the process of the generation and development of microcracks and other defects, while the quantitative characteristic of these processes is the "defect density" and the degree of orientation of the particles in the shearing direction. The change in structure with displacement of the particles depends on the activation energy imparted by the particles. The equation for the creep rate, including the rate corresponding to the beginning of the steady-state flow and also the theoretical formula for the long-term strength of the ground, were obtained from the theory of thermal activation. Interesting results were obtained when establishing the basic laws of thermorheology and thermofracturing of the ground. The temperature variations of the permafrost to -30° can be described by a nonlinear dependence of the hereditary type. A method of determining the parameters of this function has been developed. The anomalously large coefficients of linear expansion of permafrost have been established, which gives rise to the significant value of the thermal deformation and the thermal residual effect phenomenon. Thermal deformations are most significantly affected by secondary short-term temperature fluctuations.

The investigations made are of vital importance when analyzing the stability of fills, earth dams, and underground reservoirs and when underground communications cables, pipelines, and other engineering structures are being installed. Important studies have been performed for the purpose of determining the dependence of the mechanical properties of ice on the conditions of its formation. Three stages of recrystallization of ice during the creep process have been isolated, and each stage determines the manner in which the ice is capable of being deformed and has a significant effect on the mechanism of deformation. These studies are of practical significance, for ground ice must of necessity be used in the foundation base for the structures.

Physical-chemical research has made a significant contribution to the study of permafrost. A vast amount of data has been accumulated that makes for a better understanding of the nature of the processes and the phenomena. For example, the results of X-ray diffraction and spectral and nuclear magnetic resonance research indicate that in frozen soil materials, between the mineral particles and the ice, a multilayered system arises. It comprises adsorbed water where, in turn, the tightly bound layer and the diffuse film are distinguished. It also comprises the mobile surface and contact layers of the ice adjacent to the remaining mass of the ice body. The diffuse film and the mobile surface layer of the ice are the most dynamic in this system. Depending on the characteristic features of the surface of the mineral particles and the thermodynamic conditions of freezing, the relations between these layers and the strength of the permafrost strength vary. It is therefore necessary at the present time for the development of the rheology of permafrost to take into consideration the formation of this special type of structural bond. Several peculiarties have been discovered in the structure of liquid boundary layers. It has been established that it is not only the nature of the solid state, but also the electric field of the binary layer formed during exchange between the solid and liquid phases, that has a significant effect on the physical and chemical properties of water. Experimental data have been obtained that confirm the conclusion that the tightly and loosely bound water correspond spatially to the dense and diffuse parts of the binary electric layer. This conclusion permits a new approach to be made to the problem of clarifying the nature and properties of the bound water in the surface layer, the role of which in the course of the processes related to permafrost can be decisive.

It can be assumed that there is a relationship between the unfrozen pore moisture, its mobility, and the surface charge. A detailed study of this phenomenon will not only be of theoretical, but also of great practical, significance. The results of the physical-chemical analyses will not only make it possible to estimate the degree of ground heave, but also to select a method of decreasing it.

American researchers have developed hypothetical models of the phase interfaces of silicate-unfrozen water-ice and silicate-water-silicate. These models are used to predict and explain the freezing and thawing phenomena. Experimental data have been obtained on the phase composition of a number of standard soil types. It has been established that the phase composition can be determined by the phase potential of the groundwater and the thermodynamic relations determining the magnitude of the lowering of the freezing point of the water. The water-ice phase can be calculated from an empirical equation in which only the temperature and the specific surface are variables.

Studies have also been made of the migration of the water, the ions, and dissolved materials during freezing of the soil, and while it is in the frozen state; the phenomena of ice segregation; heaving; and the electrical conductivity of the permafrost and its strength. Latterly, American researchers have begun to display an interest in the behavior of the thawing soil under load. In the Soviet Union a considerable amount of experience has been accumulated on this question.

THE THERMOPHYSICS OF PERMAFROST

The thermophysical studies that have made notable advances in geocryology include the study of the processes of heat, and material, transfer in frozen, freezing, and thawing earth materials and their energy interaction with the environment. The primary purpose of these studies was to discover the thermophysical laws of the formation and development of permafrost, seasonally thawing and seasonally freezing layers as a function of the geological and geographic conditions of the environment, and also to establish the theoretical principles of controlling these processes in order to ensure stability and longevity of structures, improve the efficiency of excavation and mining operations, and increase the productivity of farmlands.

Initially, the geothermal area predominated in thermophysical research. Basically, this consisted of analyzing the temperature fields of the upper layers of the permafrost solely with regard to conductive heat transfer. The heat-transfer processes were not tied to the mass-transfer processes in the ground itself. The calculation of the interaction between the external and internal processes was effected initially in relation to landform. This method of research continues to be of great practical importance. With minimum expenditures, it permits determination of the thermal state and even the temperature of the ground, and in a number of cases it affords a basis for predicting possible changes in its thermal state as a consequence of economic activity. More recently, attention has begun to be devoted to the external components of heat and mass exchange. This has made it possible to arrive at a quantitative evaluation of the mode of development and existence of permafrost and seasonally frozen ground and also to estimate the effectiveness of various thermal improvement measures with respect to a number of areas. In addition, there has been vigorous progress in what is solely an important applied field--the study of the thermal interaction of various structures with frozen, freezing, and thawing ground.

On the basis of broad theoretical and experimental studies, a great contribution has been made to the study of the thermophysical and mass-exchange characteristics of permafrost. A number of original methods and devices have been developed, and important experimental data have been accumulated.

The efforts of many research organizations have led to the development of new progressive methods of investigating the thermodynamic processes occurring in the permafrost of the Earth's crust, more especially, measurements of the temperature, phase composition, bound water, heat and mass fluxes, the depths of freezing and thawing, and others.

The contemporary level of development of thermophysical research in geocryology is characterized primarily by an effort to discover the general laws and derive functions expressing not only the qualitative but also quantitative relations between the various parameters determining the dynamics of freezing and thawing. The studies to establish the interrelations between the processes of heat and mass transfer and their effect on the temperature fields of the ground are important.

The drilling of a large number of deep wells in the northern regions has provided a basis for stepping up and expanding geothermal research. An important achievement in recent years is the detailed study of the permafrost with a transient thermal regime in many parts of Siberia. A single trend has been noted toward a decrease in the thickness of this permafrost. The transient temperature fields of the ground cannot be characterized by a single temperature. To get a complete picture of this process, it is necessary to study the heat-transfer coefficients in the ground to considerable depths.

In Alaska and Canada, in the course of geothermal research, four basic problems were solved: (1) the effect of the steady-state intrasoil heat flux on the position of the lower surface of the permafrost; (2) the seasonal temperature variations and the depth of the permafrost table; (3) the long-period surface temperature variations, the development of permafrost, and the effect of climatic variations on it; and (4) the effect of the following surface conditions on the temperature of the permafrost: bodies of water, vegetation, topography, and buildings. In particular, it has been established that along the arctic coast of Alaska the thickness of the permafrost varies from 282 to 640 m, primarily as a result of variations in the thermal conductivity of the rock. The problems of determining the temperatures of the permafrost as a function of any hypothetical climatic variations and the inverse problems of determining the climatic variations from data derived from geothermal measurements are being solved. Simplified mathematical solutions are being developed for various geothermal problems that make it possible to estimate the effect of the surface conditions on the permafrost temperature. Progress has been made in computer simulation of the thermal regime of the ground.

GEOPHYSICAL METHODS OF RESEARCH

The effectiveness of geocryological research depends to a considerable degree on the extensive introduction of geophysical methods and on the definiteness of the information derived.

The geophysical methods are being used in hydrogeological studies; in geological mapping of unconsolidated and indigenous deposits and mineral exploration and prospecting; in surveys for the construction of pipelines and industrial and civic structures; and in research undertaken to discover the laws of permafrost distribution.

During the last decade, both in the Soviet Union and in America considerable experience has been accumulated in the study of permafrost by geophysical methods, foremost among which is direct-current electrical exploration. These years marked the completion of a major stage in the study of electrical conductivity and the geoelectrical profiling of permafrost. The overall picture of the distribution of various types of permafrost, as indicated by the electrical conductivity and the nature of its dependence on the primary factors determining the magnitude, have been largely clarified.

It has been established that a decrease in the electrical conductivity of earth materials as a result of freezing is caused by an increase in the sinuousness of its pore channels due to the formation of ice inclusions. Thus, the structure and composition of the permafrost are together primary factors determining its electrical conductivity. The laws of formation of the cryogenic structure control the variation in the electrical conductivity of the freezing ground and explain the initially paradoxical conclusion that the electrical conductivity of the permafrost is not dependent on the temperature within the limits of accuracy of the measurements. This conclusion is valid for most earth materials with low mineralization of the pore solution and is important in that it permits retention of the classical approach to the problems of interpretation.

It is especially necessary to single out the conclusion regarding the conservation of relative differentiation of the ground with respect to electrical conductivity and freezing, emphasizing that frozen unconsolidated deposits of one type, but having a different cryogenic structure and composition, differ significantly in electrical conductivity with other conditions being equal.

The basic types of geoelectrical profiles of permafrost and their seasonal variations have been established and investigated. Computer-based methods of recognizing the types of geoelectrical profiles have been developed.

It has proved possible to ascertain the factors that lower the resolving capacity of electrical prospecting by direct, and low-frequency, current methods, consideration of which is mandatory when solving specific problems. Thus, in electric prospecting of permafrost, results of a general nature were obtained that serve as a basis for expanding both the scope and scale of its application. The methodology of electrical prospecting using direct and low-frequency current techniques for the investigation of permafrost has been developed, as have the variable electromagnetic field techniques.

The ascertainment of the factor of surface polarization, not of homogeneous phases but of heterogeneous materials, is dictating a new approach to the ion-conducting geological formations (in particular, permafrost). This enables them to be considered as surface-polarizable bodies. The observed macroscopic effect and the secondary electrical field associated with it are opening the way to new developments in the study of permafrost.

The use of electrical methods of investigating permafrost has taken on a systematic character and for the mapping of taliks, and, especially for determining the thickness of frozen unconsolidated deposits, it has also become very widespread. In recent years, regional work has been done to determine the thickness of permafrost by vertical electrical logging.

During the last 10 yr seismic exploration has begun to be used on a relatively broad scale, primarily to determine the thickness of frozen unconsolidated deposits. In a number of studies its effectiveness for the mapping of permafrost and ground ice, for determining the position of the permafrost table and also the elastic properties of permafrost in its natural state, and for solving other problems in geocryology has been demonstrated.

ENGINEERING PROBLEMS IN DEVELOPING THE PERMAFROST REGION

Due to the intensive industrial development of the northern regions of the USSR, in recent years scientific workers, builders, researchers, and designers have displayed heightened interest in problems of construction on permafrost. The broad studies of permafrost have played a vital role in providing for the development of methods of stable construction of surface and underground installations in complex geocryological conditions.

The principle of building with directional control over the engineering-construction properties of the permafrost to create a foundation providing for long-term stability of the structure is being more and more widely used. This involves artificial reduction of the temperature of the "plastic-frozen" ground constituting the foundations, maintenance of the "solidly-frozen" state of this ground, preconstruction thawing and consolidation of permafrost when construction entails thawing of the foundation, physical-mechanical and thermophysical methods of increasing the strength and bearing capacity of the ground, and so on. Pile foundations installed in the permafrost by various methods have become widely used. They make it possible to regulate the bearing capacity. Methods of designing piles have been developed that take into consideration the rheological properties of the permafrost.

The first-priority goals are to reduce the labor requirements of the zero-cycle construction operations and use areas previously considered unsuitable as construction sites. It is necessary to make more complete use of the bearing capacity of the permafrost on the basis of a comprehensive appraisal of the nature of its strength and of the laws of deformation and transformation of the properties of the soil in the required direction by thermophysical and physical-chemical means.

The results of the research and the experience acquired in building on permafrost have been drawn together in government standards documents--in the *Construction Regulations and Standards* [SNiPS] and also in the *Manual* relating to these standards. A number of other standardizing documents have also been published. A broad exchange of construction experience has taken place at All-Union meetings and seminars held in Yakutsk, Magadan, Noril'sk, Vorkuta, Tyumen', and Krasnoyarsk.

In the United States and Canada, the principles of building on permafrost have undergone continuous refinement since the 1940's. For proper selection and study of construction sites in permafrost regions, a technique involving remote measurements, geophysical exploration, and laboratory and field analyses has been developed. A procedure for electromagnetic sounding is being developed at the present time.

Studies are being made of various methods of controlling heat exchange at the surface of the permafrost, including methods aimed at changing the surface albedo. Various methods of designing foundations and footings under complex geocryological conditions are being found.

A great deal of attention has been given to the construction of airports and highways. Airports built in diverse permafrost areas, such as Thule in Greenland and Inuvik and Normal Wells in the Northwest Territories of Canada, operate normally. This attests to the reliability of the construction methods used. In Alaska and in Canada the construction of new roads on permafrost is continuing. They are operating successfully. Of undoubted interest is an 82-km stretch of highway between Fairbanks and the Yukon river, which was built 3 yr ago over a time span of only 9 mo. The builders boldly decided to cross several depressions containing thick wedge ice, after originally providing for natural or artificial thermal insulation of the embankments to ensure controlled melting of the ground ice.

The methods of drilling wells, and of building and operating industrial projects and pipelines when the natural geocryological conditions are complex, must be efficient and ensure that the environment is fully protected against possible irreparable damage. Many studies have been made in this area. Suffice it to say that information on the environmental effects of laying pipelines in Alaska has been prepared by the American government and published in six volumes (3,200 pages). The problems connected with drilling, casing, and cementing of operating wells in permafrost are complicated and have not yet been adequately studied.

The thawing of permafrost around the wells through which warm oil flows necessitates the use of a special installation and reinforcement of the casing and wellhead equipment in order to ensure emergency-free operation. In recent years American and Canadian specialists have carried out extensive research to ensure stability of oil pipelines laid in the direction of Canada and through Alaska from north to south, from Prudhoe Bay to Valdez Arm.

In the near future construction work will begin on this oil pipeline, which will have a total extent of 1,270 km with about 50 percent laid in permafrost. Oil with a temperature of 60°-90° will be transported through a pipe 1,220 mm in diameter laid in the ground, along the ground, and above the ground. On individual sections where the permafrost conditions are unfavorable, the pipe laid in the ground will be artificially cooled. This interesting experiment in construction and operation must be carefully investigated in order that later on a more efficient solution will be found with respect to the construction, operation, and maintenance of warm pipelines in the permafrost region.

In the northern part of the Soviet Union, gas pipelines of great length have been built under various, often complicated, natural conditions. Although these lines have been in operation for many years, some complex problems have yet to be investigated and more efficient solutions must be found. It would probably be advisable to combine the efforts of the Soviet and American scientists and engineers in order to solve these complex problems, which are of great interest to the Soviet Union, the United States, and Canada.

PROSPECTS FOR THE FUTURE DEVELOPMENT OF GEOCRYOLOGY (PERMAFROST STUDIES)

In geocryology, as before, the primary goal remains the development of theoretical and experimental research aimed at revealing the general and the special laws of the formation of seasonally frozen ground permafrost and the phenomena accompanying them. Regional studies form the basis for this.

In the USSR, despite the large volume of work that has been undertaken, the Taymyr and the central part of the central Siberian plateau, the Yana-Indigirka and the Kolyma lowlands, the Yakagir and Alazeya plateau, and the Koryaki highlands have not been adequately studied. It is necessary to step up geocryological research in these areas. Intensive development of the permafrost region calls for further development of regional research, even in such comparatively well studied areas as western Siberia, central Yakutia, Pribaykal'ye [the near-Baikal region] and Zabaykal'ye [Trans-Baikal], and so on. The accelerated rates of development of the northern parts of Canada and Alaska will also require much new information about the geocryological conditions.

The geocryological conditions of the alpine regions have been poorly investigated. By rough estimation the total area occupied by alpine permafrost in the USSR is 2 million km^2, while the overall figure for the alpine regions of the world (not including Antarctica and Greenland) is 4 million km^2. The main goal of geocryological study in alpine regions is the accumulation of factual information for use in establishing the basic laws of spatial development of permafrost by geographic zones under marine and continental climatic conditions. Steady-state investigations of cryogenic processes are required, and a start must be made on a permafrost survey.

In view of the great importance attached to the mapping of seasonally frozen ground and permafrost, it is necessary to become thoroughly familiar with the samples of geocryological maps presented at the conference and to exchange opinions about the methods of compiling them. A single mapping procedure must be adopted, in order that the published maps will have the required weighting and be of equal value.

The increasing volume of geocryological research points up the need for developing highly efficient methods of undertaking it. These methods must satisfy the following basic requirements:

1. The research must be carried out at survey level, that is, it must feature the stringent mine working standards of which are mandatory for the scale in question and the same procedure measuring the basic exponents in order that they are comparable to the data of different researchers.
2. The methods employed must be the least costly and entail the minimum expenditure of manpower; that is, provision must be made for a high level of mechanization and automation of this research.
3. A uniform, mandatory procedure must be used throughout.

In the Soviet Union many procedural instructions, directives, and recommendations have been published with respect to geocryological research, interdisciplinary permafrost-hydrogeological and and engineering-geological surveying, and the study of ground ice and the cryogenic texture of permafrost. The time has come to develop a single-standards document pertaining to studies of the permafrost region that takes into consideration the contemporary level of scientific and technical instrumentation.

Further progress in understanding the geological and genetic aspects of geocryology must of necessity involve a more profound and detailed elaboration of the principles of cryolithogenesis. There is a need for a comprehensive study of the peculiarities of rock formation under permafrost conditions and for revealing the specific forms of the interrelation that exists between geological and cryogenic processes. Much more attention must be devoted to mineralogical-petrographic studies and to the study of authigenous mineral formation under various permafrost-geological and facies conditions.

A study must be made of the quantitative laws of development of cryogenic phenomena as a function of the controlling natural factors, and, on the basis of this, procedures must be developed for making quantitative predictions of their occurrence under natural conditions and for artificially controlling them when they do occur.

The history of the development of permafrost is also of great interest. It is necessary to perfect criteria for use in evaluating the geocryological conditions of the past, besides which, procedures providing for absolute dating of the permafrost must be widely applied in these studies.

The most important areas of future hydrogeological investigation of the permafrost region are the following:

1. The study of the laws of groundwater recharge and discharge, that is, the entire set of complex and multifaceted problems of the interrelation between atmospheric, surface, suprapermafrost, intrapermafrost, and subpermafrost water and the conditions of the interaction of this water with the permafrost.
2. The broad study of the groundwater regime and the overall water budget.
3. More profound studies of the chemical composition of groundwater and its peculiarities arising from cryogenic metamorphism.
4. The development of economically efficient and productive methods of improving the quality of groundwater and perfecting the techniques of using it for various purposes.
5. The development of measures designed to ensure efficient use and prevention of pollution of the groundwater of the permafrost region.

The further development of the physical-chemical research revealing the nature of formation and the structure, composition, and properties of permafrost as a multicomponent, multiphase system at below-freezing temperatures is required.

It is necessary to expand the study of heat fluxes in frozen and thawing ground, which was started comparatively recently but has already made a significant solution of the problems of the dynamics of the permafrost region and the history of deep freezing of the Earth's crust.

It is necessary to study the thermal state not only of frozen but also of the underlying unfrozen ground, as well as the thermophysical laws of their interaction. These studies are still in a relatively backward state. The studies of the thermal interaction between the frozen ground and the groundwater are in the same situation. It is necessary to continue the study of the effect of the deep freezing of the ground on the hydrogeological and hydrochemical conditions.

Based on the theory of irreversible processes, it is necessary to consider permafrost as an active formation with enormous reserves of latent energy, which, on being released when changing its thermophysical state, is capable of significantly transforming the surrounding materials and activating powerful processes of mass transfer in the ground.

The thermal processes in the ground are irreversible and transient. They are accompanied by changes of all the parameters of the thermodynamic state, a number of collateral processes, and transport phenomena. The achievements of transient thermodynamics are being inadequately used in geocryology, even though this would bring together advances in the fields of the thermophysics, mechanics, and physical chemistry of frozen, freezing, and thawing ground and open up new paths to the broadening of our knowledge. It is well known that thermodynamics is based on very general laws of nature to which all of the terrestial phenomena are subject, and in order to get a general picture of the geocryological processes in the ground it will be necessary to develop research in this direction as well.

There are grounds for assuming that by using geophysical methods it would be possible to investigate the state of the permafrost much more extensively than at the present time. The future development of geophysical research in geocryology presupposes an increase in the unambiguousness of the results obtained, the development of new methods of investigating permafrost and the improvement of existing ones, an increase in the productivity of the geophysical studies by using aerial surveys in regional investigations, and the mapping of ore-bearing minerals occurring in the permafrost.

The activities of geophysicists in making a detailed study of the relationship between the cryogenic texture of the ground, the nature of its formation and its physical-mechanical properties on the one hand and the acoustic parameters, the specific resistance, the dielectric constant, and electrochemical activity on the other, are of increasing interest.

A number of problems in engineering geocryology arise as a consequence of the construction of large hydroelectric power plants and the reservoirs connected with them. The formation of reservoirs introduces significant changes in the heat and mass exchange and causes thawing of the permafrost. It is necessary to predict the rate and depth of this thawing and the settlement of the ground, percolation of the water through the thawing ground, the reworking of the shores made up of permafrost, and the changes in bottom relief of the reservoirs as the permafrost thaws and sediments accumulate. In the presence of ground ice and icy soils, the solving of these problems is complicated.

One of the most important contemporary problems is the construction of oil, and gas, pipelines. Along routes followed by pipelines, the engineering-geocryological conditions change and are often complicated and difficult for construction. An increase in the diameter of the pipes and, consequently, in their weight, and increases in the power and lifting capacity of the machines used for transporting and laying the pipes, serve to complicate construction still further. The disturbance of the natural vegetative cover by machines leads to an increase in depth of thawing with resulting roadlessness. In areas where ground ice is present, this can initiate the development of thermokarst.

The existing methods of laying gas pipelines--underground, above ground, and surface--lack validity at the present time. Construction experience indicates that the underground procedure is the most efficient, but under certain conditions the surface method is preferable. To fully validate efficient methods of building gas pipelines, it is necessary to expand the research, construct experimental sections where the natural conditions are most complex, and perform comprehensive observations on the gas pipelines constructed.

For the construction of oil pipelines, special methods must be developed. The difficulty encountered in the construction of oil pipelines arises from the fact that the oil being transported [is at, or] must be heated to, high temperatures. Consequently, methods must be developed that will exclude the possibility of heat escaping into the ground and ensure that the ground will be kept frozen. Specialists in the United States and Canada are performing interesting experiments in the vicinity of Inuvik, as a result of which valuable data will be obtained that will help to solve the problem of how to build oil pipelines efficiently in the permafrost region.

The improvement of the methods of constructing gas pipelines and the development of efficient methods of constructing oil pipelines are main problems on which the attention of the researchers must be concentrated.

In connection with the increase in the rates of development of mineral deposits in northern regions and also the incorporation of increasing volumes of rock in the material to be refined, which is due to a progressive lowering of the grading requirements for minerals and the combined extraction of new rare and dispersed components, thus entailing a continuous and abrupt stepping up of the refinement rate, an increase in the efficiency of mineral extraction is required. This can be achieved:

1. By increasing the power of the equipment and producing new structural designs for machines or components capable of working frozen deposits and disposing of them with a high rate of productivity.
2. By improving the existing methods and developing radically new methods of thawing or reducing the strength of permafrost. On the basis of a comprehensive and thorough study of the physical-mechanical and thermo-physical properties of the permafrost, the optimal conditions of preparing large masses of ground for excavation must be defined.

The development of effective methods of maintaining large volumes of earth materials in the thawed state during the winter is an equally important problem. The maintenance of the ground in the thawed state during the cold part of the year permits an increase in the extraction of the minerals and makes it possible to extract minerals year 'round.

The conservation of the environment and efficient use of natural resources is one of the most important problems of modern times. Great concern is being aroused among the peoples of all countries by the pollution of the air and water basins, by soil erosion, and by the damage inflicted on the animal and plant kingdoms. In connection with the increase in emissions into the atmosphere, the carbon dioxide content is increasing constantly. In the opinion of many scientists, as a result of the so-called greenhouse effect, this can lead to an increase in the temperature of our planet and, as a result, to the melting of the glaciers and permafrost and to other undesirable consequences.

Among the numerous problems confronting man in his effort to make maximum use of natural resources, ensuring at the same time the maintenance of their potentiality for reproduction or for regeneration in the required direction, the problems of the North stand alone.

The permafrost region is one of the dynamic factors quickly reacting to changes in the thermal regime of the landscape. The northern environment is exceptionally sensitive to human activity and is easily damaged. Its protective functions are limited; everything is in a delicate state of equilibrium. The restoration of natural landscapes in the permafrost region takes place very slowly, often requiring decades, and the process is not always reversible. As a rule, alterations in the soil and plant cover and in the soil moisture conditons lead to the development of irreversible processes, foremost among which are solifluction and thermokarst processes. These processes usually lead to the destruction of the previously formed landscape, sometimes with total or prolonged loss of its biological productivity.

Thermokarst and thermal erosion processes also occur under natural conditions. They are manifested in the formation of lake depressions, in the destruction of the riverbanks and seashores made up of icy earth materials, and in other transformations of the relief. These natural processes gradually die down as thermodynamic equilibrium is reached. Under human influence these phenomena can reach catastrophic proportions and encompass vast areas.

The character of the permafrost region, as determined in the course of research on particular subjects, undergoes significant changes during development of the region, both by virtue of the natural dynamics of the landscape mantle and under human influence. The alterations of the natural environment caused by human activity are so profound and widely ranging that a need has arisen for the working out of geocryological predictions, that is, for making qualitative and quantitative estimates of such alterations in the geocryological situation, which will occur in the course of developing a region and which must therefore be foreseen in the plans. In addition to the environmental protection measures, these plans must also provide for recultivation of the land. One of the main tasks in this area is the development of measures for protecting the northern environment and the establishment of a single geocryological prediction procedure.

In geocryology, as in other sciences, the volume of information is constantly growing. The processing of the flow of incoming data requires further mathematical formalization of the geocryological concepts, for only on this basis is the broad introduction of mathematics and comprehensive use of computers in geocryology possible.

The formidable problems facing geocryology call for a unification of the efforts of scientists and specialists of various countries. During the past 3 yr, we have engaged in combined research with Mongolia and Czechoslovakia. There is an agreement between the USSR and Canada calling for combined studies in a number of fields, which will run until 1980. We have a mutual interest in arranging for combined research with the United States and other countries.

The Organizing Committee hopes that this International Conference will serve to deepen mutual understanding and expand and reinforce the creative scientific relations among the geocryologists of all countries, provide each of its participants with information about the latest theoretical and methodological achievements in geocryology throughout the world, and direct attention to the most important problems of future research. The scientific papers presented at the conference have been published, and the discussions will undoubtedly afford opportunities for comprehensively investigating and identifying new problems in geocryology, arising as a result of modern scientific and technical progress.

THE THERMOPHYSICAL PRINCIPLES OF THE FORMATION OF PERMAFROST

G. V. PORKHAYEV Research Institute of Foundations and Underground Structures, and V. T. BALOBAYEV Permafrost Institute of the Siberian Division of the USSR Academy of Sciences

Soil is a multicomponent dispersed material of complex composition and structure, the pores of which are partially or completely filled with water in the liquid or gaseous phase. When soil freezes, part of the water is converted into ice, as a result of which the soil acquires significantly different properties. Ice under normal conditions can exist only at negative celsius temperatures. Therefore, the temperature appears as the most important characteristic of permafrost.[34]

The temperature is one of the principal thermodynamic parameters of the state of any material and serves as an object of investigation in thermodynamics. But since the thermal regime of the ground varies in space and time, it is constantly in a nonequilibrium thermal state. These states of the material are considered as the thermodynamics of irreversible processes.[11] The principles of the thermodynamics of irreversible processes have been worked out only in the last two decades, and the methodology in applying them to complex dispersed systems has yet to be developed. This fact has dictated the elaboration of preeminently the thermophysical principles of the formation of permafrost.

From the viewpoint of thermophysics, the temperature field of the ground is determined by the conditions at its boundaries and the properties of the earth materials. In the case of the upper layer of the Earth's crust, the upper boundary is the "daylight" surface, and the lower boundary is some provisional surface at a fixed depth.

The Earth has its own internal sources of heat, which cause the existance of a heat flux in the Earth's crust directed toward the surface. It is manifested as the boundary condition on the lower surface. The density of the ground heat flux varies within narrow limits from 0.02 to 0.07 kcal/m^2-h, depending on the age of the geotectonic structures.[27,41] The younger they are, the greater the magnitude of the heat flux. For all practical purposes, this heat flux is independent of the thermal state of the Earth's surface and the composition and properties of the ground. It is stable over large areas, and within the Quaternary at least there has been no variation with time.

The main source of thermal energy at the Earth's surface is solar radiation, and that part of it which is accumulated by the surface is called the radiation balance. On the average the magnitude of the radiation balance for the year increases from 10 kcal/m^2-h near the Pole to 45-50 kcal/m^2-h at a latitude of 40°.[1] About two-thirds of this heat is absorbed during the process of heat exchange between the ground surface and the atmosphere, and for all practical purposes it is irretrievably lost. The subsequent condensation of water vapor in the upper layers of the troposphere leads merely to a slight decrease in the adiabatic temperature gradient in the atmosphere, which has almost no effect on the temperature of the Earth's surface. The remainder of the energy of the radiation balance is expended on heating the atmosphere. In the mean annual heat balance there is no heat flux to the ground. On the contrary, the Earth loses heat in an amount equal to the ground heat flux.[9]

The mean annual surface temperature of the Earth is established as a result of the heat and moisture exchange with the atmosphere, constituting the thermodynamic level of these processes.[34] Here the ground heat flux, the magnitude of which is approximately one thousandth of the radiation balance, cannot have and has no effect on its formation. The surface temperature is determined by the magnitude of the external heat balance components. The larger they are, the higher the surface temperature of the Earth. The ratio of the expenditures of heat on evaporation and on the heat flux to the atmosphere affects its magnitude, but it is not the determining factor. Other conditions being equal, the decrease in the expenditure of heat on evaporation and increase in the heat flux to the atmosphere is accompanied by an increase in the surface temperature.

The mean annual value of the radiation balance is less than 30-35 kcal/m^2-h in the permafrost region. This means that the processes of external heat and moisture exchange for smaller values of the radiation balance take place at a below-freezing mean annual surface temperature of the Earth. The decrease in radiation balance from south to north is accompanied by a decrease in the mean annual surface temperature.

A negative surface temperature is a necessary condition for the development and existence of permafrost. In the steady state-thermal regime the thickness of the permafrost is greater with lower mean surface temperature, lower ground heat flux, and more thermally conductive rock. This in its most general form is the qualitative picture for the conditions of permafrost formation.

The variety of natural conditions and types of surfaces in the permafrost region can introduce changes into the above-presented picture. Most of this region is characterized by continental conditions of external heat exchange for which the ordinary heat-balance equation is valid. However, in low regions that are adjacent to the arctic coast and in the European part of the USSR, a powerful advective heat transfer prevails, which is commensurate with the elements of the heat-balance equation. In this case the latter can contain an advective component with the corresponding sign.

The characteristic features of the surface temperature situation develop in the mountainous regions. They do not extend beyond the scope of the already-mentioned laws, but the heat-balance components, especially the radiation balance, are determined to a significant degree by the nature of the relief, slope orientation, and altitude. The variation in radiation balance resulting from differences in angle of incidence of solar radiation on slopes of different orientation and altitude corresponds to the relation of surface temperature distribution to relief. This creates an extremely patchy distribution of the permafrost in mountainous conditions.

Unfortunately, the opinion is quite widespread that the mean annual surface temperature is determined by the proportion of solar energy that reaches the ground in the form of heat. It derives from the concept of the heat-exchange process at the surface during various seasons of the year, although it is not entirely valid. Under equilibrium mean annual conditions the heat flux can be zero, but depending on the heat-balance components we can have an entire range of existing below-freezing surface temperatures. The direction of variation of the heat flux in the ground as an index of variable heat-exchange conditions on the Earth's surface is of great importance.

With the current accuracy of measuring the components of the external heat exchange and moisture exchange, the construction of any theoretical model for calculating the mean annual surface temperature of the Earth on the basis of them would probably be premature. The basic difficulty here is in determining the transfer coefficients, which vary within very wide limits. Furthermore, in the vast area constituting the permafrost region there are as yet few stations equipped for radiation and heat-balance measurements radiation and even fewer special observations stations in the various terrain conditions. The problem of investigating components of the external heat exchange and hydrologic cycle and their relations to surface temperature in various natural combinations remains urgent in the near future.

The measurements of ground heat flux, which, along with the surface temperature, determines the morphology and thermal regime of the permafrost have been started relatively recently and should be expanded considerably. The initial uncoordinated data indicate that their value is unusually high and that they significantly expand the concepts of the state and dynamics of the thermal processes in the permafrost region and their interaction with the underlying thawed material.

A negative mean annual ground surface temperature is a necessary condition for the development and existence of permafrost, but it is not sufficient in itself nor is it unique. By the surface of the Earth we mean the interface between the free atmosphere and the upper organic-mineral layer of the lithosphere or hydrosphere, in which the basic processes of reciprocal transformation of solar energy are realized. This is the ground cover (moss, peat, turf, snow, and others) or, in their absence, the surface of the soil itself. The ground cover, as a rule, is poorly heat conductive; it inhibits deep freezing of the ground in the presence of a negative surface temperature and in some cases completely protects them from freezing. In this respect the role of the snow, moss, and peat is well known.[21,40]

Without discussing its characteristics, it can be stated, for example, that the vast regions of Trans-Baikal and Mongolia owe the existence of permafrost to the thin snow cover, whereas, in the severe conditions of western Siberia, permafrost is nonexistent in many areas because of the heavy snowfall.

In studying the effect of the natural ground cover on the underlying temperature, the physical-geographic approach has prevailed. Although the general thermophysical methods of calculating this effect have presumably been developed, the heat-transfer mechanism and the thermophysical properties of the ground cover have as yet hardly been studied. The snow cover has been the subject of the largest number of papers,[22,35] but their results pertain to areas where permafrost is absent and which are frequently unsuitable for adaptation to the specific conditions of snow accumulation in the severe northern regions. The situation is even worse with respect to the study of the thermophysical properties of moss and peat in the frozen state. All of this is retarding the development and application of methods of making thermophysical calculations and forecasting ground temperature under natural conditions.

As a result of the inclination of the Earth's axis of rotation to the ecliptic, the arrival of the solar radiation at the ground surface and the heat- and moisture-exchange processes on the surface undergo periodic seasonal variations. In the permafrost region they are immensely important for natural-biological activity and human existence. Because of the seasonal variations in surface temperature during the summer, the upper layer of the frozen ground thaws and is heated to temperatures that permit development of relatively rich fauna and flora. Seasonal temperature fluctuations, seasonal thawing, and freezing have an impact on all spheres of human economic activity in these regions and actually determine all of the natural phenomena and processes, including many aspects of the permafrost formation. This role of the permafrost has aroused steady and prolonged scientific and practical interest in it. As a result, thermophysical principles of the processes of seasonal thawing and freezing of the frozen ground have been established and the theory of calculating them by the so-called Stefan solution has been developed. Whereas the exact solution still meets with considerable difficulties, although machine algorithms have been obtained for a number of versions,[8,28] a large number of approximate solutions fully satisfy problems in engineering practice.[2,24,34,36,42,43] The approximate methods take into account all of the most important defining parameters and are brought to a sufficient degree of accuracy and in correspondence with natural conditions. Forming the basis for the majority of them are the methods proposed by L. S. Leybenzon[23] and M. Ye. Shvets[44] or V. A. Kidryavtsev's method of the ground heat storage balance.[14]

As a result of calculating the heat exchange between the ground and the atmosphere, it has become possible to provide theoretical predictions of the depth of seasonal thawing under variable climatic conditions.[2,4]

The phase transformations of water into ice and ice into water are extremely energy-consuming. Therefore, during the summer the basic amount of heat coming from the surface is absorbed in the process of phase transitions, and in the winter it constitutes a significant part of the heat loss from the ground. The annual ground heat storage of the seasonally thawing layer proves much larger than that outside of the permafrost region. However, the seasonal temperatures in this layer and at its surface do not correspond to the ground heat storage. This is explained by the fact that the ground heat flux in the presence of seasonal temperature fluctuations at the surface is largely determined by the depth and rate of movement of the phase boundary, and not by the external heat-exchange characteristics.

The formation of the permafrost temperature regime during seasonal thawing and freezing takes place in a highly complex manner and depends greatly on the phase processes.[37] The surface temperature fluctuations are nearly harmonic, but below the seasonal thawing layer this dependence on time is so distorted that the application of the harmonic solution to the thermal conductivity equation for the layer of seasonal temperature fluctuations becomes impossible. The thermal regime of the permafrost in this layer is largely determined by the heat fluxes that come through the phase boundary. As there is no reliable procedure for determining them, the problem of finding the temperature field in the layer of seasonal fluctuations has not yet been solved, and the need for a solution remains urgent. Without this, it is impossible to realize the transition from the temperature conditions at the surface to the temperature of the permafrost at the bottom of the layer of seasonal fluctuations--the mean temperature of the frozen ground. The problem is complicated by the fact that in the course of the year, with the phase transitions the thermophysical properties of the ground and surface cover vary along with the properties of the surface itself. These characteristics sometimes give rise to significant differences in the mean annual surface temperature and the mean temperature of the permafrost.[14]

It is evident that the mean permafrost temperature is influenced by a wide variety of natural factors, each with a different role. Chief among these are the amount of solar energy accumulated at the surface and the conditions of surface heat and moisture exchange. They determine the surface temperature of the Earth. Next in importance are the physical-geographic conditions, including the geobotanical, soil, geomorphological, hydrologic, and other characteristics of the Earth's surface combined into landscape groupings. They give rise to the surface conditions of the thermal transfer of energy to the ground.

Next are the geological conditions characterized by the compositon, structure, and properties of the ground in the layer of seasonal temperature variations that determine the rate of seasonal thawing and freezing and of heat transfer in the permafrost.

The effect of these conditions cannot be considered alone. They are closely interrelated and to some degree are mutually conditioned by the indivisible natural energy balance.

As soon as the mean ground temperature falls below freezing, permafrost begins to form. The depth of winter freezing comes to exceed that of summer thawing, and a layer of frozen ground forms that increases as the temperature decreases. The rate of the deep freezing of the ground is determined by the rate of its mean temperature variation, the ground heat flux, the thermophysical properties of the ground, and the amount of pore and fissure moisture capable of turning into ice at the given temperature.

At present the initial stage of deep freezing can be observed only near the southern limit of the permafrost region and in isolated areas emerging from the water. Permafrost first made its appearance over a large area several hundreds of thousands of years ago. Since that time there have been repeated variations in heat-exchange conditions at the surface and the thermal regime of the permafrost. It would appear that these were not accompanied by complete thawing of the permafrost. These variations are very slow and are detected only by indirect signs. The heat-transfer processes in the ground, however, and the freezing and thawing are characterized by a low rate. If the temperature variation in the surface layer takes place more slowly than the freezing or thawing of the ground, the temperature field in the permafrost region follows the changing conditions at the boundary, maintaining the thermal regime close to equilibrium. This regime characterizes the permafrost in dense, slightly jointed deposits that contain little water and are primarily of post-Jurassic age.

The thickness of the equilibrium permafrost corresponds to the contemporary conditions of surface heat exchange and ground heat flux with the existing accuracy of these measurements. It is greater with lower negative temperature of the frozen ground, less ground heat flux, and higher thermal conductivity of the earth materials. With the existence of positive or negative sources of heat both within the permafrost and below it, the thickness of the permafrost can be determined to a significant degree by its density.

Their widespread distribution and the simplicity that characterized the thermophysical interpretation of the stationary temperature fields were the main reasons for the thorough studies that have been made of this permafrost. The fact that specialized integrated geothermal studies of permafrost began only recently, prior to which time the permafrost was investigated from the general geocryological standpoint, the primary object being merely to ascertain the thickness with no attempt made to analyze the temperature fields and heat fluxes, has played a not unimportant part in this. Even today, most of the concepts of the thermal regime of the permafrost are based on the unconditonal assumption of its remaining stationary in the simplest unidimensional case.

The expansion of the geography of the research, the application of new quantitative methods, and the development of the thermophysical principles of the processes of deep freezing of the Earth's crust have demonstrated the narrrowness of these comcepts. On the basis of the variety and territorial variability of the physical-geographic and geological conditions, the temperature field of the upper layers of the Earth's crust is three-dimensional. On the plains, the three-dimensionality of the temperature field is mainly due to the nonuniformity of the mean permafrost temperature in adjacent areas.[38] As a result, the permafrost base acquires a complex configuration. The effect of the temperature nonuniformities decreases exponentially with depth; it is therefore most noticeable for small thicknesses of permafrost. This clearly points up the inadequacy of data obtained at one point only. The reliability of correlating data over large areas is more valid in the case of thick permafrost and falls to zero in the case of thin permafrost. A distortion of the unidimensionality of the normal temperature field of frozen earth materials is also caused by lithological-petrographic and tectonic factors, but it is encountered more rarely and is manifested more weakly.

In alpine areas these factors are manifested to an even greater degree. The very strong influence of the geometry of the relief is imposed on them here. When all other conditions are uniform, the effect of the relief is a redistribution of the heat flux beneath its various elements. It becomes concentrated under the valleys, as a result of which the ground acquires a higher temperature. Conversely, beneath the peaks the heat flux diminishes, which leads to a reduction in the temperature there. Accordingly, less thick permafrost forms in the valleys than under the peaks. As the temperature of the permafrost decreases with altitude, the contrast in the distribution of its permafrost thickness becomes even more pronounced. At the same time, in a number of mountain systems of the permafrost region the law that results in a reduction of the temperature with altitude is disturbed by radiation and orographic inversions that promote pronounced cooling and deeper freezing of the earth materials in the valleys.

The complexity of the surface temperature distribution in the mountains, and also the effect of the relief and the hydrogeological setting, have placed major difficulties in the path of investigating the morphology and the thermal regime of the permafrost. Actually each element of the relief at each altitude is matched by its own thermal regime and permafrost thickness. However, the problem cannot be simplified. The redistribution of the heat flux under the effect of the relief by no means indicates absorption or generation of it. In valley-slope-peak geomorphological systems the total mean heat flux remains invariant, being equal to the ground heat flux. If the mean surface temperature of this system is known, it is possible simply to determine the mean thickness of the permafrost for the given relief. It is a more generalized index of the geocryological conditions of the mountainous regions.

The construction of the overall thermophysical picture of the formation of permafrost under alpine conditions has made for a heightening of interest on the part of geocryologists in these regions and elaboration of the regional mechanisms of its development in areas that until recently were blank spots on the map.[33] Serving as the basis for this were numerous geothermal data which, in many alpine areas, made possible the creation of a spatial concept of the temperature fields of the ground, analytical solutions to the problem of the effect of the relief, and methods of mathematically simulating two-dimensional and three-dimensional temperature fields.[5,6,12,13,37,38]

As already stated, during the period in which permafrost has existed its temperature has varied. The nature of these slow, long-term variations remains unexplained. They can hardly have been strictly periodic, for the surface temperature of the Earth forms under the influence of too many factors, and the main source of heat--the absorbed solar radiation--is determined not only by cosmic but also by planetary and local ground conditions. Most probably the extrema of the surface thermal regime corresponded to the superpositioning of several single-phase effects in definite time intervals.[32,39] The long periods of relatively stable thermal state are not excluded. All of this took place against a background of general cooling of the climate. In view of this, permafrost must be regarded as a dynamic formation with its own history of existence and development. It is natural that we are interested in the thermophysical aspects of its development.

The temperature field of the ground within the confines of the approximately 2- to 3-km upper layer of the Earth's crust in which the processes of deep freezing developed and are developing cannot retain any traces of the thermal effects of bygone eras. The ceaseless process of temperature equalization, however slowly it proceeded in the ground, erases them to the point where they are imperceptibly small. According to some estimates, however, information about the thermal conditions of the last several tens of thousands of years must have been retained. It is precisely the evolution of permafrost that has led to this. In poorly-cemented, highly jointed, and unconsolidated rocks the freezing and thawing rates are at least an order lower than the heat-transfer rate in them. They are a highly inert system in which the thermophysical phase processes develop with a long delay.

The thicker the permafrost, the more inert it is and the more remote the epoch to which the conditions preserved in it correspond. Here we can only talk about nonstationary temperature fields and the disequilibrium permafrost corresponding to them.

Studies in recent years have demonstrated that disequilibrium permafrost is extensively developed where there are thick deposits of unconsolidated and weakly-cemented rocks of Quaternary, Neogene, and Cretaceous age. It was initially discovered in western Siberia in the form of the deep-lying, thick permafrost disconnected from the surface.[7 15] Later, it was found in Yakutia and in the northern part of eastern Siberia in the form of anomalously thick permafrost not corresponding to the contemporary conditions of heat exchange at the surface. As a result of detailed thermophysical studies of it in Yakutia it proved possible to establish characteristics and criteria for isolating it.[3]

Its insignificant rate makes it impossible to ascertain directly the variation in the temperature field of the earth material with time. Neither are we any closer to the goal by analyzing the temperature distribution with depth, which is typical of a nonstationary regime. This is because the three-dimensionality of the temperature field gives rise to the possibility of obtaining a temperature curve of any shape, even under stationary conditions. This is frequently overlooked, when deliberately interpreting stationary temperature distributions with depth that have maxima, minima, or gradient-free sections, such as a degradational or aggradational material. Unfortunately, errors of this type are frequently encountered. From the equation of thermal conductivity it follows that the nonstationary thermal regime is characterized not only by rock temperatures that vary with time but also by the variation of the heat flux by coordinates as a result of the heat capacity properties of the rock. The observation of the latter has no connection with the necessity for measurements during various time intervals. The basic cause of the nonstationarity of the thermal regime of the rock is the variation with time of the boundary conditions at the surface.

In this case the total heat flux must increase or decrease with distance from the surface. In the permafrost the variation of the heat flux in space is even more pronounced, as a result of the high energy capacity of the phase transitions of the water both within and at the boundary with the thawed ground. The existence of the clearly defined lower phase boundary is confirmed by a large number of factual data. At this boundary the heat flux under the conditions of a nonstationary regime, varies unevenly. An increase in it, at the transition into the melted zone, is indicative of thawing and a decrease of freezing of the rock.[14] Thus, in slowly occurring nonstationary thermal processes the heat flux is the principal and effective criterion for detecting them.

The measurements and analysis of the heat fluxes in frozen and the subjacent thawed ground were begun recently, but they have clearly demonstrated what extensive possibilities they offer for obtaining valuable additional information. It is to be regretted that their scope is still limited. The basic parameters of the thermal state of the frozen ground continue to be the temperature and its gradient.[14,38] Although the latter is directly linked to the heat flux, the strong dependence on lithologic-petrographic factors and the extraordinary variability caused by this within the confines of even a small space significantly restrict the limitations of its use.

The investigation of the heat fluxes have to all intents and purposes demonstrated that our concepts of permafrost as being resistant, stable natural formations, in no way correspond to reality. Many researchers have made statements about this, but they have not had at their disposal the necessary data.

It turned out that over the entire area in which permafrost is present in the USSR, a decrease in its thickness is observed predominantly as a result of thawing from below under the influence of the ground heat flux.[3] There are no data indicating that the opposite is the case. Apparently, this process is global in scale. The known regions containing anomalously thick permafrost, the thermal regime of which does not correspond to the contemporary conditions at the surface, occupy a large area and there is a definite trend for it to increase as further regional studies are undertaken.

Disequilibrium permafrost is azonal with respect to the physical-geographic conditions at the surface, for its development is linked with the specific type, composition, and properties of the geological deposits and with the paleoclimatic conditions. Within central Yakutia, where they have been studied best of all, their thickness is two or three times greater than it should be in an equilibrium regime, and the thawing rate reaches 2-3 cm/yr. The temperature field within the confines of this permafrost is characterized by an almost equilibrium regime. Accordingly the study of it should be accompanied by studies of the thermal regime of the underlying thawed ground.

The theory of the thawing of disequilibrium permafrost is at the stage in which a thermophysical method of reconstructing paleoclimatic conditions is being worked out on the basis of the parameters of their contemporary state and a detailed investigation of their composition and properties. The basic advantage of this method is the possibility of obtaining quantitative paleoclimatic data.[3] The preliminary results of research in this area point to the existence of lower temperatures of the earth materials and the air during the preceding cold period than follows from the conclusions of traditional paleogeography. The continuation and expansion of this research will undoubtedly lead to a broadening of our knowledge of the conditions and processes of deep freezing of the lithosphere.

Not all of the changes in the surface temperature have a cardinal effect on the state of the permafrost. With great thickness of the permafrost even relatively long-period fluctuations cannot reach its base or affect its character of distribution and development.[14] In this case mention must be made of the nonstationarity of the temperature field within the permafrost which, on the whole, is in a state of equilibrium. The simplest example of this is afforded by annual temperature fluctuations.

The key factor in these variations in the thermal regime and permafrost thickness is to be found in the natural dynamics of the environment. However, we are living in an age when the availability of power to meet the demands of society is increasing so rapidly that the day is not far off when it will become commensurate with the energy released by certain natural processes. Already, giant projects for the transformation of nature are being proposed and implemented. Primarily, they encompass the surface conditions and the climate, and, consequently, the temperature of the ground. The study of the thermophysical history of the natural development of permafrost affords a basis for making a scientific forecast of the effect of the transforming activity of man.

The potentialities of thermophysics for the investigation of permafrost have not yet been exhausted. By relying on the well-developed theory of thermal conductivity, the rapidly developing methods of thermal modeling and similarity, and also modern experimental equipment, and by using the powerful tools that we now have in the form of electronic computers, thermophysics is finding an increasingly wide field of application in geocryology, to the extent that it has become firmly looked into such traditionally geological-geographic subject areas as regional investigations, the study of cryogenic phenomena, cryogenic texture formation, and structure formation. In spite of this, we have to recognize the limitations of thermophysical methods when describing the processes in dispersed earth materials undergoing freezing and thawing. This is a consequence of the one-sideness of the approach to the study of these earth materials. The thermophysical aspect of these processes is being investigated in isolation from others that are inseparably linked with it. At the same time, it is well known, for example, that heat transfer in the ground is accompanied by transfer of moisture and dissolved material. The heat transfer takes place both by thermal conductivity and by moisture migration and diffusion and, conversely, the moisture flux and the flow of dissolved material are determined not only by the moisture and concentration gradients, but also by thermal diffusion and thermal moisture conductivity.[18].

These and the processes of molecular transport similar to them, which are interrelated and mutually caused, are investigated by the thermodynamics of nonequilibrium states. In contrast to classical nonequilibrium thermodynamics, the parameters of state are not regarded as being under equilibrium conditions but as continuous functions of coordinates and time.[11,25] Heat-transfer theory is a special case of nonequilibrium thermodynamics. The latter is predicated on the following basic laws of nature: The mass conservation law and the energy conservation and conversion law and also the principle of the origination of entropy. The rate of entropy origination multiplied by the absolute temperature proves to be equal to the sum of the products of the fluxes for the corresponding thermodynamic motive forces. The fluxes characterize the irreversible processes under the conditions of nonequilibrium state, and the forces are caused by nonuniformity of the system (the presence of potential gradients) or by deviation of the internation parameter of state from the equilibrium value. The fluxes (heat, material, charge, momentum, and so on) are determined not by the action of the sole force corresponding to them, but of all of the thermodynamic forces, and are linear functions of them.

The known natural processes of transfer are subordinate to this linear function within a wide range of conditions. Thus, in order to describe interrelated processes, the system of linear

Onzager equations has been derived. The coefficients of proportionality of the flux to the thermodynamic forces are called kinetic coefficients. They are linked by a reciprocal relation indicating the equivalence of the kinetic coefficients corresponding to cross effects, that is, to effects that are mutually caused. Examples of such effects are diffusion in the presence of a temperature gradient and heat transfer when a concentration gradient exists.

The conservation laws, the energy balance equation, the equation of state, and the linear Onzager equations completely describe the nonequilibrium thermodynamic state of any system and the irreversible transfer processes in it. With their help we can obtain a matched system of differential equations in partial derivatives for the varying states of a material medium, which can be solved under the necessary initial and boundary conditions.

In accordance with the Curie principle not all of the processes of molecular transfer can interact with each other and be described by an interrelated system of differential equations, but only those which are determined by the actions of the thermodynamic forces that have an identical tensor nature. Thermodynamic forces can be scalars (the zero rank tensor), as, for example, the forces of chemical and phase transformations, vectors (a first-rank tensor), that are the forces of heat and mass transfer, and tensors (a second-rank tensor), to which the forces of momentum transfer belong. The forces having identical tensor rank or an even difference in ranks can interact. For example, heat and mass transfer cannot be combined with the processes of chemical and phase transformations and be accompanied by cross phenomena. The application of the Curie principle will help us to gain a better understanding of the interaction of thermodynamic phenomena in a complex system. Dispersed earth materials constitute one such system.

The calculation of the thermodynamic forces gives rise to the greatest procedural difficulty, as this requires a knowledge of the entropy source based on the Gibbs equation, which takes into consideration all of the necessary parameters (the second law of thermodynamics).

For the most general thermodynamic investigation of freezing and thawing earth materials, it is necessary to find the equation of their thermodynamic state that relates all of the most important parameters. This is a formidable but necessary task. At the present time the study of the following freezing and thawing phenomena of earth materials is being conducted in disconnected subject areas: thermophysical, physical-chemical, electrophysical, physical-mechanical. Nonequilibrium thermodynamics forms the basis for consolidating the data derived by these studies and for creating a single picture of interrelated processes. It seems to us that the solution of such an essentially important problem as heaving during ground freezing can only be obtained in this way.[45]

One of the initial achievements using the methods of nonequilibrium thermodynamics was the derivation of the differential equations of mass and energy transfer and, as a partial case of them, a system of interrelated equations of heat and mass transfer in dispersed media.[25] They found a use in geocryology for the investigation of moisture transfer during ground freezing and for explaining their specific cryogenic structure.[18] There are a number of solutions to these equations.[29,30] However, they do not entirely correctly describe the processes of heat and moisture transfer in the case when a moving phase boundary is present. It is known that with the existence of this boundary the moisture transfer in the thawed zone takes place more intensely than in the absence of it. An attempt is being made to take this fact into account, on the assumption that the values of the coefficients of moisture conductivity and the thermal gradient are larger than when the ground is merely thawed. This is a purely artificial and formal procedure that fails to explain the essence of the phenomenon. The point is that the expressions for the mass thermodynamic force when deriving the transfer equations must be different for completely thawed earth materials, for thawed earth materials during the freezing and thawing process, and for freezing earth materials.

The phase transition heat of the water has an enormous effect on the processes of heat transfer in thawing and freezing ground and, of course, cannot have any effect on the interrelated moisture-transfer process. The sign of the heat source of the phase transitions also plays a role. Consequently, in the case of thawing and freezing ground and its frozen and thawed components there must be different transfer equations.

Practical requirements connnected with the economy have led to attention being concentrated on the study of transfer processes in the layer of seasonal temperature fluctuations. These processes develop very intensely here. They are accompanied by significant energy cycles and are characterized by the absence of an external pressure field and relative free variation in volume. Analogous transfer processes occur during thawing of permafrost from below. It is true that they take place under other conditions when there is a powerful hydrostatic pressure field with complete water saturation of the earth materials and, because of skeletal rigidity, no possibilities for significant volumetric changes. Their intensity and specific energy expenditures are two or three orders lower. However, they last 10^4 to 10^5 longer and therefore lead to highly significant changes in the thermodynamic state of the ground.

In regions where permafrost is currently undergoing thawing from below, a compression-vacuum mechanism of moisture transfer to the phase boundary is observed.[3] As a result of the strong hydrogeological sealing of the subpermafrost beds, the inflow of water from the surface through the talik zones cannot completely compensate for this flux, which leads to the appearance of a hydrostatic pressure deficit.[19,20] Observational data indicate that it reaches 10-30 atmospheres and encompasses the deep beds. Under such conditions even the gravitational moisture turns out to be partially bound as a result of which the water-saturated beds of sand and sandstones are characterized by insignificant water losses, and, generally speaking, those which lie directly under the frozen ground do not indicate the presence of water in them. These processes are dynamic. With a reduction in pressure, part of the gases dissolved in the water are released, and the ground contains a signficant amount of gas, including methane. The opening of the subpermafrost horizons by wells often leads to intensive but short-lived gas emission.

The rigid skeleton of the ground when the conditions are such that there are significant fluctuations of the hydrostatic pressure acquires a capability of transforming the crack-pore space. This means that the processes of deep freezing and thawing of the ground are accompanied by irreversible changes in their structure and properties. All available information points to the repeated nature of these processees. Therefore the end result of such changes can be significant.

The water transfer determines the hydrochemical composition. The influx of a large quantity of fresh surface water during the thawing process leads to a decrease in the salinity of the subpermafrost water, which is confirmed by the field data.

As is obvious, the processes of heat and moisture transfer during thawing of the permafrost are not of only theoretical interest. Nevertheless, insufficient attention has as yet been devoted to them. The theory of these processes is still awaiting its discovery. In the light of the foregoing factors it is necessary to change the concept of the role of permafrost in the evolution of natural processes. Their interaction is considered to be in large measure one-sided when the dependence of the formation of permafrost on the geological, hydrogeological, physical-geographic, and other conditions is emphasized and nothing is said about the natural transforming role of the developing permafrost itself. The narrowness of the strictly thermophysical analysis of permafrost, in which these conditions take on a singularly external, preset appearance, determining the formation of the temperature field but not dependent on it, is manifested in this. Here we are again convinced of the advantages of the thermodynamic approach to the broad investigation of the formation and evolution of permafrost. Even when on the basis of our limited knowledge the thermodynamic analysis cannot be carried through to completion, the partial functions derived are of undoubted interest. This is clearly manifested in the application of the Le Chatelier thermodynamic principle to a series of natural phenomena of interest to us. According to this principle, any process caused in an equilibrium system by an external effect or other primary process is always directed towards a decrease in the results of these effects. Its generalization to the case of nonequilibrium processes is substantiated by the stability of the stationary states corresponding to the minimum production of entropy with respect to disturbances of the individual parameters of state. In the case of irreversible processes, there is no equilibrium state; they approach the stationary regime.

The process resulting in the freezing of thawed ground is a process involving its cooling and transition to the frozen state. According to the Le Chatelier principle the secondary process of moisture transfer can be directed only toward a decrease in this effect. This means that the moisture flux must be directed towards the freezing front, for only in this case will it promote additional heat transfer to the cooled zone, and the incoming water will decrease the freezing rate. Hence, it follows that the accumulation of water in the freezing and frozen zone is a thermodynamically caused phenomenon. On the other hand, during thawing of the permafrost the resulting thawed ground is heated. Accordingly, in this ground there can be no moisture flux towards the thawing front, as this would mean more heating of it and faster thawing, which contradicts the Le Chatelier principle. What has been said does not pertain to the filtration fluxes of the water, which are caused by other forces not connected with the thermal regime.

If at the initial point in time the ground was characterized by a stationary thermal regime, then subsequent heating or cooling of it could have been accompanied by physical-chemical processes in a specific direction only: With an increase in temperature, there would have been reactions entailing heat absorption, and with a reduction in temperature-reactions entailing the liberation of heat.

The Le Chatelier principle makes it possible to estimate which processes can occur and in which direction when there is a variation in the thermodynamic state of the system.

It seems to us that surface geographic complexes are not only determined by the existing conditions, including the thermal conditions, but also by the contemporary direction of their change. Under identical conditions, but when the directions of their variation are different, the surface geographic complexes must be different, for they do not fulfill identical functions. For example, during warming periods, soil coverings and vegetation will develop at the surface that would have prevented heating of the ground. The universality of the laws of thermodynamics enables us to perform similar analyses for any of the natural phenomena associated with variations of the parameters of state.

In the application of the methods of nonequilibrium thermodynamics, an important role is played by knowledge of the transfer coefficients. Significant progress in the study of the thermal coefficients resulted in the extensive use of thermophysical methods.[16,17,31] Here, however, not everything has yet been done. The problem of establishing the analytical relation between them and the determining parameters on the basis of experimental data remains urgent. Little attention has as yet been given to the thermophysical properties of dense, weakly-cemented, and unconsolidated but deeply occurring earth materials in the thawed and frozen state, or to the properties of earth materials with a significant content of organic matter. We are not satisfied with the slow progress made in investigating the thermophysical coefficients of the soil covers, as this inhibits the development of the methods of calculating heat transfer in the active layer in the natural situation.

Of prime interest in geocryology are the transport processes that are interconnected with the heat transfer, that is, the processes defined by thermodynamic motive forces or vectors. Usually these are the gradients of various types of potentials: electric, chemical, moisture, pressure, and so on.

Their study must be accompanied by the investigation of the corresponding transfer coefficients: electrical conductivity, moisture conductivity, filtration, diffusion, thermogradient, thermoelectric, thermodiffusion, and so on.[18]

Direct transfer coefficients are investigated by the appropriate scientific disciplines to which they are basic. Here there are well-developed methods for determining them and sufficient practical material has been gathered that, unfortunately, pertains mainly to thawed ground. For the frozen, freezing, and thawing ground the indepenedent development, collection, and systematization of data are necessary.

In the field of investigation of cross coefficients subordinate to the Onzager reciprocity principle, very little has as yet been done. The thermogradient coefficient has been better studied, and the rest are waiting for their researchers.[26]

The deep freezing of the Earth's crust is an objective natural process of variation in the state of the Earth's thermodynamic shells.[10] Its governing principle is the thermophysical conditions that become established during the given stage of development of the planet in the lithosphere, atmosphere, and hydrosphere. Appearing as one of the most powerful energy-consuming processes, it stimulates the activity of a large number of natural physical phenomena. Advances in the study of the thermodynamic and thermophysical principles involved in the formation of permafrost are one of the prerequisites for the successful development of all of the subject areas of geocryology.

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REGIONAL LAWS OF THE DEVELOPMENT OF PERMAFROST IN THE USSR

N. A. GRAVE AND V. V. BAULIN Production and Research Institute on Engineering Studies for Construction

The fact that the permafrost region is confined to the cold, circumpolar, and alpine zones of the Earth and the evolution of this region as a function of historical climatic changes reflect the basic general laws of the phenomenon known as "permafrost."

Permafrost, being one of the components of the geological-geographic environment, a consequence of the complexity and variety of the latter throughout the vast territory of the USSR, is highly heterogeneous with respect to composition and properties that, in the various parts of the region of its occurrence, very often cannot be explained solely by the zonal laws of its development.

The peculiarities of permafrost, namely, the fluctuations in areal distribution, the thickness, the temperature, the iciness, and other properties of permafrost, also depend on the regional conditions. Therefore, the study of the regional laws of development of permafrost is a particular expression of the general zonal geophysical conditions of the "permafrost" phenomenon in a specific geological-geographic environment.

The regional laws of permafrost are investigated in natural regions isolated on the basis of a multiple analysis of data on the morphotectonic, geomorphological, and lithological structure of large areas, considered from the standpoint of the geocryological peculiarities of the historical development of these areas in the Quaternary and comtemporary periods.

The complex regional characteristics of the permafrost region, together with the laws of the development of cryogenic phenomena that were discovered when studying the regions, makes possible a reliable prediction of the geocryological setting for the area in question in the event of various types of structures being erected and used there and enables us to provide for the most efficient principles and methods in the design, construction, and use of the structures.

The general zonal laws of the development of permafrost in the USSR are most competely reflected in the 1:5,000,000 scale geocryological map of the USSR compiled by I. Ya. Baranov in 1970 (Baranov, 1972). It is produced in accordance with the principles discussed earlier (Baranov, 1965). By comparison with his previous map compiled by him in 1956, Baranov in the new one devoted more attention to the composition and the genetic types of frozen deposits, which reflect the regional differences in the permafrost region. On the basis of the new data, many boundaries have been more precisely defined. The overall layout of the geocryological zonation remained as before.

Three zones have been isolated on the continent: the northern zone (with central and southern subzones) and two zones in the shelf region--Aroto-oceanic and Aroto-continental-oceanic.

In the USSR, another line of research in the identification of the geocryological regions is being developed, which, in turn, contains two different approaches that complement each other. The first is reflected in the papers of A. I. Popov, who compiled a schematic map of the permafrost-geological regions of the USSR in 1958. The distinguishing of "permafrost regions" was based on the morphotectonic structural attributes of the permafrost caused by the freezing process itself. The second was established in 1958 by P. F. Shvetsov and then used by N. A. Grave, P. I. Mel'nikov, *et al.* when investigating specific regions. Assuming that the leading attributes of permafrost are the lithologic composition, the structure, and the position in the macrorelief and mesorelief, Shvetsov proposed the concept of geocryological formation.

A series of geocryological formations was isolated by N. A. Grave in the Chukotsk-Koryak area and in Kamchatka and by P. I. Mel'nikov in Yakutia, basically from the geomorphological attributes. The zonal peculiarities of the permafrost of each formation were manifested in the isolation of the regions of development of the particular formation.

In assessing the existing attempts to isolate the geocryological regions in the USSR, it can be stated that the difference in the abovementioned approaches are indicative of the authors' attempts to provide a complete representation of the basic peculiarities and properties of the permafrost on the maps, as seen from the viewpoint of each individual author with respect to the leading attribute.

Apparently, the most efficient approach in the present phase of available concepts will be to isolate geocryological regions on the basis of morphotectonics, which will entail identification of smaller regions within these regions characterized by the zonal and formational peculiarities of the permafrost.

It should be noted that on the basis of regional investigatons of individual areas during the last 10 yr, a number of new theoretical problems have been solved. Among these, mention must be made of cryolithology--an area of geocryology that investigates the formation of permafrost as part of the overall geological process. All types of ground ice have been investigated: segregated ice, wedge ice, injected ice, and so on. These concepts were discussed most completely by A. I. Popov, Ye. A. and B. I. Vtyurinymi, Sh. Sh. Gasanov, Ye. M. Katasonov, and others.

The study of the relation between cryogenic processes and tectonics is an entirely new subject area that has received the support of the majority of the researchers. In spite of the fact that the idea that tectonics have an effect on engineering-geological processes in the general form was advanced long ago; only recently was it specifically corroborated for the northern parts of the Soviet Union. In the plains areas, this field of enquiry originated in western Siberia and received impetus from the oil and gas exploration work there.

Previously, the effect of the tectonic conditions on the formation of permafrost was investigated chiefly with respect to mountainous regions in connection with the study of groundwater. Today, these problems are also being solved for the platform regions. The available data indicate that the tectonics have a varied effect on the permafrost. It is movements of the Earth's crust that determine the prevalence of processes of accumulation or denudation of the sediments and, consequently, the formation of syngenetic (in the harsh climatic zone) or epigenetic permafrost, wedge ice, and so on. It is known that the thickness of the permafrost often decreases above the arches of anticlinal uplifts. The uplifted regions are frequently associated with frost mounds, ridged relief forms, etc.

The effect of the tectonics on the permafrost is so obvious that it is no longer possible to analyze the regional characteristics of permafrost without analyizng the tectonic conditions. One such analysis was performed earlier for the folded mountainous systems.

In connection with the rapid economic development of the North and the novel direction this has taken, problems of forecasting variations in the permafrost following a disturbance of the natural conditions in individual regions have come to be considered more extensively. Not only the variety of frozen ground, but also the various forms of economic activity, are being taken into account.

Finally, a new field of enquiry in geocryology--paleocryology--was formulated. The progress made in recent years in elucidating the laws of freezing and thawing ground in the Pleistocene are to a great extent linked with the paleocryological research, which is proceeding in two directions: (1) the analysis of the permafrost itself and the discovery of attributes characterizing the conditions of freezing in the past; (2) the study of traces of cryogenic processes in the various stratigraphic beds of the thawed Quaternary deposits and, based on this, restoration of the history of the perennial freezing of the deposits.

The papers and reports at this conference will significantly supplement and refine our concepts of paleogeography, predicated solely on geological data (the reports by A. A. Velichko; V. V. Berdnikov; M. S. Ivanov and Ye. M. Katasonov; and V. V. Baulin, G. I. Dubikov and Yu. T. Uvarkin).

In recent years the study of the interrelation of permafrost with the landscape as a whole (a report by G. S. Konstantinova) and, in particular, with the vegetation (a report by A. P. Tyrtikov) has continued. It is demonstrated that the regular alternation of plant associations can lead to perennial freezing or thawing of the ground, even in the case of unchanging climatic conditions. This fact is of the greatest importance in the southern parts of the permafrost region, where the slightest natural change in the natural conditions will lead to degradation or to neogenesis of permafrost. However, the frozen permafrost itself is one of the factors influencing the spread of the forest (a report by V. V. Kryuchkov).

Some major studies have been made that pertain to the reworking of coastlines and lacustrine shores composed of marine soils (the papers by F. E. Are and Ye. N. Molochushkin). The conclusions of these authors, which are based on regional information for the Laptev Sea littoral, are typical for the entire arctic coast. Three interrelated but independent processes have been isolated which govern the rate of reworking of the shore: thermal abrasion, thermal denudation, and thermokarst. The specific manner in which each of these shows up leads to destruction of the shoreline by up to 4-5 m per year. Some interesting new data have been obtained on the temperature regime of a disintegrating shore. *In situ* observations and simulation of the thermal processes have made it possible to establish the rates of variation of the ground temperature and the take at which the ground thaws beneath the bottoms of reservoirs.

The geocryological laws ascertained for the various regions are reflected in geocryological maps (both composite and specialized), which show the variations of the most important indices characterizing the permafrost. The procedure for compiling composite permafrost maps, using as an example the Tyumen' Oblast, was the subject of a paper by A. V. Vostokova.

Although the geocryological maps of western and central Siberia and the Mongolian People's Republic presented at the conference were not compiled by a single procedure, they nevertheless reflect sufficiently the regional laws of development of permafrost in these regions, and the above-mentioned procedure for compiling composite maps is one way of efficiently developing unified regional geocryological maps.

During the last 10 to 15 yr, regional geocryological studies have been performed by a whole series of Soviet research organizations, and the choice of the regions was determined primarily by their practical importance for the development of the country's productive resources.

The most recent data presented in the papers at the conference extend and supplement our knowledge of the regional peculiarities of permafrost in the USSR, and they are discussed hereunder with reference to individual regions.

THE EAST EUROPEAN PLAIN

The formation of the permafrost in this area is subject to strong influence from the Atlantic, which has resulted in a sublatitudinal trend of the geocryological zones.

The fluctuations of the climate, both in the Pleistocene and in historic times (including the variations recorded directly by meteorological observations), have caused repeated shifts of the permafrost zones.

At the present time the southern permafrost boundary extends along latitude 67° north on the meridian of the Kanin Peninsula and latitude 65° in the Urals part of the plain. The ground temperature falls to -5.5°, and the thickness is 400-500 m (the report by N. G. Oberman).

In recent years, geocryological research (under USSR Gosstroy [State Construction Committee] and the Ministry of Geology) has been accelerated as a result of the exploration of new mineral deposits. The latest collective summary, entitled "The Geocryological Conditions of the Pechora Coal Basin," was published in 1965. The new data are partially summarized in a volume of a monograph on the hydrogeology of the Komi ASSR and the Nenetskiy National Okrug ("Hydrogeology of the USSR," Vol. 52, 1970) and also in collections and individual articles. The dependence of the distribution and thickness on the tectonics and the hydrogeological conditions is noted. The most detailed studies of permafrost were performed in the vicinity of Vorkuta from a study of the paleocryological conditions of the Pleistocene over the central and eastern Europe as a whole and certain of its regions (reports by A. A. Velichko and V. V. Berdnikov). Based on an analysis of traces of the permafrost processes, an attempt is made to create a schematic outline of the climatic variations of the Pleistocene and isolate the basic stages of the cryolithomorphism.

THE WEST SIBERIAN PLAIN

In the last 10 yr the western Siberian plain has become a unique experimental reference source for permafrost studies. It is here that the most complete study has been made of the zonal and regional peculiarities of the processes involved in the freezing and thawing of the ground. Major studies are being conducted that relate to the plotting and compilation of permafrost sketch maps, doing construction-oriented surveys and surveys connected with the exploration of gas deposits. The results of the research have been summarized in monographs, collections, and individual articles.

A paper by V. V. Baulin, G. I. Dubikov, and Yu. T. Uvarkin and a report by V. V. Baulin and A. L. Chekhovskiy summarize a great deal of information derived from investigations of this unique region. Based on an analysis of the laws governing the distribution and texturing of the permafrost and their connection with the sum total of natural conditions, a study was made of the latitudinal-zonal variations of the permafrost and the laws governing its formation.

The basic zonal features of the texture and distribution of the thick contemporary permafrost of western Siberia evolved throughout the second half of the upper Pleistocene and Holocene, that is, after the age of the boreal transgressions. The extremely cold conditions of the Zyryanka period caused intense freezing of the ground far beyond its current distribution. Evidence of this is to be seen in the traces of permafrost in the Zyryanka sediments. The very considerable warming of the climate in the Holocene led to partial thawing of the ground to latitude 68°-69° N. Later, the ground again froze, but south of the 66th-67th parallel the two frozen layers did not merge, and south of the 61st-62nd parallel permafrost did not in general form at the surface. As a result of these climatic variations, on the west Siberian plain a successive alternation is observed from continuous to two-layered permafrost. The thickness of the upper layer, formed after the climatic optimum, decreases from 40-80 m to 10-20 m. The lower (relict) layer of permafrost was preserved until the pre-Holocene cold aged. Its table gradually dips southward from 50-100 m to 100-150 m. The thickness of the relict frozen layer varies from 50-100 m to 200-300 m southwards of latitude 62°-61° to the region between the 60th and the 55th parallels. The table of the relict frozen layer lies at depths of up to 200 m; the base is at a depth of 300-400 m. The relict permafrost investigated in recent years is unique in the permafrost zone in that it has been discovered only in western Siberia.

The climatic optimum had a decisive influence on the cryogenic texture of the permafrost and the development of frozen land forms. Thus, syngenetically frozen Pleistocene sediments have been preserved only to the north of the 68th-69th parallel. This parallel is the distal limit of many of the landforms connected with the freezing and thawing of the ground (thermokarst formations, frost mounds, and so on).

The upper horizons of the permafrost are closely linked with the dynamics of the contemporary climate and vegetation (the report by Ye. B. Belopukhova and A. P. Tyrtikov). A specific characteristic of the temperature regime of the permafrost of western Siberia is the existence of a broad (300-400 km) zone in which the temperature ranges from 0° to -1° to -2°C. The emergence of this zone is connected with a rise in the mean annual air temperature from the end of the last century and the "zero curtain" phenomenon. During the last 10-15 yr, however, a cooling of the climate has been observed that causes a lowering of the ground temperature, as observed by direct measurements and a southward extension of the permafrost region.

The distinctive features of the development of frost mounds, thermokarst, ground ice, and so on are also linked with the history of the region (the reports by Yu. T. Uvarkin and by I. I. Shamanova and V. P. Yevseyev). The rate of development of the thermokarstic processes increases in a southerly direction to 68°-69° north latitude.

The research in western Siberia has established the dependence (for the plains areas) of all of the permafrost processes and formations on the tectonics. This is most clearly exemplified by the thickness of the permafrost.

A decrease in the thickness of the permafrost is observed above third-order structures. The difference in the depth of occurrence of the permafrost base on the limbs and in the free part of the uplift reaches 150-200 m. In the case, however, where in the arch part of the fold there is a gas bed which is a thermal screen, the depth of occurrence of the permafrost base may be even greater. The process that leads to cooling of the ground above a gas bed may become more pronounced as a result of adiabatic expansion of the gas.

The relation between the depth of occurrence of the permafrost base and the geothermal gradient which, under the conditons existing in the western Siberian plain, is proportional to the heat flux, is manifested very clearly. The magnitude of the geothermal gradient decreases from 5-6° to 1.2-2° in the west-east direction and the thickness of the permafrost increases from 200-300 m to 400-500 m in the same direction.

THE CENTRAL SIBERIAN PLATEAU

Geocryologically speaking, despite the striking manifestation therein of a latitudinal zonation, this vast region differs significantly from western Siberia in the composition of the permafrost. This is due to the geological conditions of the plateau and the pronounced effect of the variously composed groundwater on the texture of the permafrost (the paper by S. M. Fotiyev, N. S. Danilova, and N. S. Sheveleva and the reports by N. S. Danilova, M. A. Piotrovskiy, G. E. Rozenbaum, N. G. Oberman, and N. B. Kakunov *et al.*). The geocryological conditions of the northwestern part of the plateau are discussed in a monograph (N. S. Sheveleva and L. S. Khomichevskaya, 1967); those of the eastern half of the plateau are discussed in a monograph on the basements of Yakutia (K. F. Voytkovskiy *et al.*, 1963) and the region's hydrogeological peculiarities ("Gidrogeologiya SSSR" [Hydrogeology of the USSR], Vol. XX, 1970).

A specific characteristic of the geological conditions of the central Siberian plateau is the existence of a stage of cooled rock and a stage of rocks containing saltwater with negative temperature (cryopegs). Usually these stages occur directly beneath the permafrost. A cryogenic series consisting only of permafrost is present in the eastern and southern parts of the region. The permafrost, passing into cryopegs downward along the section, is most typical of the plateau and widespread in its central and northern parts. The stable and extremely severe climatic conditions, combined with the shallow (20-200 m) occurrence of saline water, have resulted in the formation of a thick (more than 1,400 m) stage of flooded strata with negative temperatures. In terms of the below-freezing temperatures of the groundwater (to -5°), the Olenek Basin has no analogs anywhere in the world.

The two-stage cyrogenic series, consisting of permafrost more than 100 m thick and a stage of temporarily frozen ground more than 100 m thick, was formed within the Anabar Massif of crystalline rock.

The paleoclimatic conditions have had a decisive effect on the formation of the contemporary geocryological situation. An analysis of the state of the cryogenic series and the paleogeographic conditions of the Quaternary has made it possible to distinguish two geocryological zones: a northern and a southern. In the northern zone the severe climatic conditions persisted throughout the entire cooling epoch, causing continuous freezing of the strata. The individual warmings of the climate did not essentially change the perennial freezing process. Consequently, the cryogenic series in the northern zone dates from more ancient lower Pleistocene times, has a continuous distribution, is thicker (500-1800 m and more), and has a lower temperature (from -2° to -16°). In the southern zone, throughout the Quaternary period the processes of perennial freezing and thawing have replaced each other repeatedly.

At the present time, predominating in the southern zone is a late-Holocene cryogenic series, which is of moderate thickness (to 100-200 m) and is characterized by a continuous, distribution and by high (from 0° to -2°) negative temperatures. In the northern zone the processes of frost fracturing are not only developed in the river valleys but also on the interfluves in ground of varying composition. The thick ancient syngenetic wedge ice is characteristic of the northern Siberian and Yakut provinces. In these provinces the thermokarst process is most widely developed. In the southern zone, along with the contemporary wedge ice, thick pseudomorphs replacing wedge ice and also relict forms of thermokarst are extensively present. Solifluction formations are one of the predominant forms of the tundra zone.

The geocryological conditions of the Trans-Baikal and the Near-Baikal regions have been studied in recent years by the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences; the Institute of the Earth's Crust, also of the Siberian Division; USSR Gosstroy; the Ministry of Geology; and Moscow State University. The results of the research are summarized in the monographs, articles, and collected works on the hydrogeology of the USSR.

A distinguishing feature of the paleogeographic setting of the region is the stability of the markedly continental climatic conditions throughout the Holocene and possibly the Pleistocene, which have had a decisive influence on the permafrost.

A peculiarity of this vast region is the contrast in the geocryological conditions, determined both by the latitudinal zoning and the altitudinal belts, in conjunction with the complex geological-tectonic situation (the reports by T. N. Kaplina and O. P. Pavlova, and R. Ya. Koldysheva).

Among the regional factors, the most important are the composition of the ground and the tectonics, the effect of which shows up in the absolute elevations of the relief, in the nature of its ruggedness, and, finally, in the geocryological conditions. High-altitude (altitudes above 1,500 m), medium-altitude (1,000-1,500 m), and low-altitude (less than 1,000 m) regions have been distinguished, within which the specific laws governing the formation of permafrost are noted.

The high-altitude zones are characterized by distinct evidence of altitudinal belts and maximum freezing of the strata. On the ridges the thickness of the permafrost reaches 500-900 m, and the temperature drops to -6° and lower. In the intermontane basins made up of unconsolidated flooded sediments, in spite of the effect of the temperature inversion, the thickness of the permafrost does not exceed 300 m and the temperature drops to -5°.

In the medium-altitude zones, the altitude belts and the temperature inversion compensate for each other. There are therefore no sharp differences in the geocryological conditons of the basins and the watersheds. The maximum thickness of the permafrost increases northwards from 100 to 300 m, and the temperature of the strata is -1° to -3°.

The lowlands of Trans-Baikal are classical regions. It is this temperature inversion that predetermines the severe geocryological conditions in the valleys. In the valleys the thickness of the permafrost increases from 10-50 m in the South to 150 m in the North. The temperature of the strata in this area increases from -0.5° to -2°.

On the watersheds in the southern part of the area of low-mountain relief permafrost is absent, while in the North the permafrost is discontinuous and the thickness is as much as 100-150 m. The mean annual temperature is between -1° and -2°.

In Trans-Baikal and the Near-Baikal regions, an abundance of tectonic fractures and associated large icings (a volume of ice of up to 15 million/m^3) are observed. In central Yakutia, consisting of loose Meso-Cenozoic deposits, the formation of permafrost occurred under the conditions of a markedly continental climate.

Our concepts of the geocryological conditions of central Yakutia and the laws of their development are based on fundamental research carried out by the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences and also by Moscow State University, the USSR Ministry of Geology, and other organizations. As a result of these studies, at the Permafrost Institute a geocryological map of the Yakut Republic, drawn to a scale of 1:5,000,000, was compiled and published for the first time ("Gidrogeologiya SSSR" [Hydrogeology of the USSR], Vol. 20). Some monographs and collections have also been published.

By studying the key sections of the Quaternary deposits in central Yakutia (Mount Mamontova) and in the coastal lowland, which entailed the use of permafrost-facies analysis, it has been established that the permafrost in these areas is of lower Pleistocene age. This is also indicated by the results of geothermal observations performed in recent years in Yakutia.

In central Yakutia the climatic fluctuations that are known to have occurred in other regions throughout the Pleistocene and the Holocene did not cause significant transformations of the cryogenic texture of the permafrost. This explains the widespread occurrence of ancient and recent syngenetic icy deposits. Depending on the latitude, the mean annual temperature of the ground ranges from minus 5°-7° in the northern regions to minus 1°-2° in the southern regions.

The differences in the composition of the ground, the tectonic and the geomorphological structure, and the hydrogeological conditions have resulted in major variations in the depth of freezing of the upper part of the loose sedimentary mantle. This ranges from 100-200 m to 400-650 m. The morphology of the present-day surface of this territory is largely determined by such processes as frost fracturing, thermokarst, and heaving. The processes of ice melting are accompanied by the formation of alasses, which often create a distinctive landscape.

NORTHEASTERN USSR

The geocryological conditions of this region are formed under the influence of the Asian anticyclone, the cold Arctic Basin, and the Sea of Okhotsk. As a result of the temperature inversion, an unusual altitudinal geocryological zoning is clearly manifested.

The most extensive geocryological studies in the Northeast were performed by the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences, the Northeastern Integrated Research Institute, Moscow State University, the Ministry of Nonferrous Metallurgy, and certain other organizations.

The integrated geocryological studies were performed in the mountainous regions of the Northeast (by the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences). It was there that the regional peculiarities of the heat balance of the Earth's surface, the relationship between deep freezing of the Earth's crust and recent and ancient glaciation, the peculiarities of the deep freezing of the tectonic depressions in the Northern Polar Hemisphere, and the role of the altitudinal zoning in the development of permafrost were first revealed and substantiation was obtained for the hypothesis regarding deep freezing of the alpine zones of the Northeast (more than 1,000 m) and significantly less freezing in the cold intermontane basins.

Perennial freezing could have originated by at least the beginning of the Pleistocene period. In contrast to the more westerly regions, in the Northeast the Pleistocene and Holocene climatic variations were more weakly manifested.

The peculiarities of the region's geocryological conditions were caused by the active manifestation of tectonic processes and the effect of groundwater. Associated with this are significant variations in the thickness and temperature of the permafrost, the development of taliks (the paper by K. A. Kondrat'yeva, S. F. Krutskiy, and A. B. Chizhov). In the lowlands of the Northeast, consisting of series of unconsolidated deposits, the effect of the latitudinal zoning is appreciable. The mean annual temperature of the continuous permafrost rises towards the south from -12° to -7°. The thickness of the permafrost decreases in this direction from 500-600 m to 300 m. On the Pacific Coast the temperature of the ground increases to between -6° and -1°, and the thickness decreases to 100-300 m. The iciness of the deposits making up the lowland reaches 40-50 percent and more (basically segregated and wedge ice).

In the mountainous regions of the Northeast, depending on the composition of the freezing earth materials and the nature of the altitudinal zoning, the mean annual temperature of the ground varies from between -5° to -8° in the basins to -10° and below within the alpine zones. The temperature inversion is exhibited most clearly as a function of the region's geographic position within the 700 to 2,000 m altitude range. There are significant variations in the thickness of the permafrost. These range from 100 to 700 m and more.

In the areas consisting of unconsolidated sediments, the extensive occurrence of polygonal relief forms, heaving, and thermokarst processes is typical (the paper by S. V. Tomirdiaro, V. K. Ryabchun, and Ye. N. Molochushkin). The specific nature of the mountainous regions is manifested in the intensive occurrence of solifluction and other cryogenic processes. The complex tectonics of the region have resulted in the formation of naleds of unique dimensions (to 50 km^2 and more) in the groundwater springs. Under the conditions obtaining in the Northeast there is an interrelation between recent glaciation, naleds, and permafrost.

The results of the geocryological studies in the Mongolian People's Republic derived by the joint Soviet-Mongolian geological research expedition (1967-1971) are of interest. In essence we have the first major geocryological survey for the area comprising the Mongolian People's Republic (the paper by A. F. Gravis, S. I. Zabolotnik, A. M. Lisun, and V. L. Sukhodrovskiy). A basic regional peculiarity of this area is the close connection between the deep freezing of the ground and the laws of altitude zoning (the "Alpine" type of permafrost, after L. A. Yachevskiy).

In accordance with the tectonic structure of the region and the peculiarities of its relief, three altitude belts in which permafrost develops are distinguished. These are: island, discontinuous, and continuous distribution. The law of latitude zoning is manifested here solely in a lowering of the absolute elevations of the boundaries of the altitude zones in the northerly direction.

According to data derived from extrapolation of geothermal measurements in shallow boreholes, the greatest permafrost thickness, amounting to as much as 1,000 m, must be expected in the high-altitude zones of Khangay and the Mongolian Altay.

The data adduced have not confirmed the hypothesis concerning the southward shift of the contemporary permafrost boundary in Asia. On the contrary, it should be moved northward to some extent.

In the paper there was no mention of certain mountainous regions (the Tien-Shan', the Altay, the Pamirs, and so on), the geocryological conditions of which are of undoubted interest in permafrost studies, but are not fundamentally important from the standpoint of clarifying the history of the perennial freezing of Eurasian deposits.

The analysis of the laws governing the distribution, cryogenous texture, and history of development of the permafrost in Eurasia shows that the general law of development of permafrost throughout the Quaternary is a latitude zonal and altitude zonal variation of the basic properties of the permafrost. However, the specific natural conditions of each region correct for the manifestation of these two types of zoning, which often leads to contravention of the basic law.

The regional studies have made it possible to establish the basic tenets of the law of development of cryogenous processes and phenomena in the Earth's crust. The qualitative interrelations between the individual elements of the set of natural conditions and the perennial freezing of the ground were clarified. A geocryological map of the USSR to the scale of 1: 5,000,000 has been compiled, which unfortunately is not yet ready for publication. There are, however, geocryological maps for certain regions.

Definite progress has been made in resolving problems pertaining to the history of development of permafrost. It is on the basis of regional geocryological studies and data on the geology, paleogeography, and paleocryology, primarily in the Quaternary period, that the laws of perennial freezing have been ascertained and the contemporary anomalous characteristics of permafrost have been explained. The ever-increasing effect of tectonic processes on the zonal geocryological conditions is also being explained, and these studies are already taking shape in a special area of geocryology--the history of permafrost.

Unfortunately, there are as yet no collated studies concerning the history of the formation of permafrost throughout the Northern Hemisphere. In our view, one of the most important tasks of this conference is to find a solution to the problem of compiling a geocryological map of the Northern Hemisphere, as well as a monograph on the development of permafrost in the Quaternary period in the northern regions of Eurasia and America.

GENESIS, COMPOSITION AND STRUCTURE OF PERMAFROST AND GROUND ICE

A. I. POPOV Moscow State University, and YE. M. KATASONOV Permafrost Institute of the Siberian Division of the USSR Academy of Sciences

The many-sided problem of the genesis, composition, and structure of permafrost and ground ice includes questions that for the most part are resolved independently of one another. It can be discussed in differing degrees of detail, either with emphasis placed on one or other of the numerous aspects of the problem or comparatively uniformly for each part of it. In the light of the tasks confronting the international conference on permafrost, it is evident that a more complete clarification of the current status of this problem is desirable. However, so varied are the questions and fields of inquiry that constitute it that it would hardly be correct to investigate them in strict isolation from one another. In focussing attention on certain of the most prominent areas, it is more expeditious to dwell on the overall problem of the genesis and development of the permafrost and its subordinate ground ice and regard them as geological formations whose structure and composition facilitate the task of their genetic validation. It is inevitable that questions pertaining to the age of these formations must be touched on here, although in this case they are not independent.

THE OVERALL STATUS OF THE PROBLEM OF THE GENESIS OF PERMAFROST

During the last decade, about a hundred papers directly or indirectly pertaining to this problem have been published in the USSR. The majority of them contain a description of ground ice and the processes occurring in the freezing and frozen earth materials. Many of these questons are also being investigated by other sections of the conference, and we shall therefore dwell upon the geological aspects of the problem under discussion, which is: the genesis of permafrost and seasonally freezing earth materials and the laws of formation of the ground ice contained in them.

The main point in the study of ground ice as geological formations is to ascertain the character and nature of the relationship between the ground ice and the country rock. Both scientifically and practically speaking, the genetic relationships are of fundamental importance--genetic in the sense that the ice is formed in deposits that have not ceased to be influenced by the environment in which they have been accumulated and continue to form as an independent series under certain cryological and geological conditions. In such cases a three-way dependence arises: (a) between the genesis (flood plain, lacustrine, residual, etc.), composition, and structure of the earth materials; (b) between their composition and structure (unconsolidated or consolidated, insufficiently or excessively wet loams, silty sand, sandy loams, containing a high concentration of peat, etc.,) and the ice forming in them.

The three-way genetic relationship is not applicable to series of bedrock. As a result of deep alterations in the earth materials, the middle link--the primary structure of the deposits--drops out of the scheme, and therefore the genesis of the deposits loses its significance as a factor directly influencing the course of the cryogenous processes. The mode of accumulation and diagenesis, for example, of Paleozoic limestones, marls, or "argillite clays" has no effect on their freezing regime or on the formation of ground ice. The latter fills fissures and voids, the formation of which in the pre-Quaternary earth materials was associated with various phenomena that do not always have a direct bearing on the origin of the earth materials themselves. The cryogenous factor in bedrock, as we shall see later, has proved to be solely a destructive factor.

The series of unconsolidated, primarily Quaternary deposits, have a different cryogenous structure. Standing out especially clearly are deposits that were formed under permafrost conditions--in the presence of a permafrost substrate. These deposits did not undergo significant changes; they retained the basic features of the primary structure. The accumulation and transition into the permafrost state of each of their genetic and facies varieties took place under the same or very similar permafrost and geological conditions, which predetermined the following: (1) the composition and moisture content of the sediments, (2) the length of time these sediments were in the thawed state and the degree of their postsedimentation changes, (3) the freezing regime--in the seasonally thawing layer or taliks, (4) the nature of the cryogenous processes and phenomena, and (5) the position of the freezing front and the shape and orientation of the ice inclusions.

The commonness of the conditions of accumulation and freezing is the most important feature of the rock-forming process in the permafrost region. In all of its stages the hypergene, sedimentation, postsedimentation, and cryogenous process interact with each other. According to Ye. M. Katasonov, specific cryolithogenic deposits develop in which the formation of the ice is subject to definite geological laws.

However, the understanding of the genesis of perenially frozen and seasonally freezing rocks is not restricted to the case that has been presented, but is more extensive and will be discussed in greater detail later.

THE MAIN TRENDS IN THE STUDY OF THE GENESIS OF PERMAFROST

Ordinarily when explaining genesis, that is, the processes and the conditions under which sediments accumulated, and when demonstrating the relation between these conditions and the composition and structure of the earth materials, geologists display little interest in cryogenous phenomena. Some of them do not even touch on ice formations. Others regard them as epigenetic and, consequently, not involved in the genesis of the earth materials.

Conversely, when making a detailed study of ice interlayers, lenses, and veins, many geocryologists note their relationship with the country rock, even though they see no reflection of the specific nature of the rock-formation process in the permafrost region in the cryogenic structure of the series. Little attention is paid to genesis, that is, the accumulation and lithification of the sediments under specific conditions.

In recent years there has been a heightening of interest in the study of the petrographic, mineralogical, and geochemical peculiarities of the composition of permafrost, in clarifying the role of the freezing processes in the disintegration and aggregation of the mineral particles in the formation of loess-type soils. Permafrost-facies investigations, which make it possible to discuss the composition, structure, and ice of Quaternary deposits in connection with their facies classification, have also undergone further development. Several subject areas have been singled out in the study of the structure of permafrost. Foremost among these are the thermophysical and the permafrost-geological areas.

The thermophysical approach to the interpretation of the cryogenous structure of permafrost is the most traditional one. It is based, on the one hand, on the experimentally established laws of the formation of ice as a function of the grain-size composition of the soils, the rate of their freezing, and the temperature gradients,[42,43,51,52,53] and, on the other hand, on mathematical calculations explaining, for example, the development of frost fissures and their associated ice and earth wedges[11] and also of the cryogenous structures.

The primary objects of thermophysical studies are cryogenous (permafrost) phenomena and processes. To be numbered among these studies, for example, are the establishment of the dependence of the thickness of the ice inclusions on the mean annual temperature of the country rock,[25] the conclusion regarding the alternation of layers with increased iciness and those with little ice in permafrost that has frozen from the top, as a result of an abrupt change in the temperature conditons at the surface;[13] the examination of the effect of the thermal regime of the earth materials on the conditions of frost fracture and on the development of fissure-polygonal phenomena;[39] the distinguishing of three types of syngenetic freezing--northern, temperate and southern--based on the thermophysical character.

The cryogenic-geological subject area considers the structure of the permafrost primarily from the point of view of its freezing characteristics in a specific geological situation.

The concept of epigenetic and syngenetic types of freezing, the "ice lattice which becomes thinner with depth" as an index of the cryoepigenesis of the series, was introduced.[30-32] It was demonstrated that a lattice of this type, indicating deep migration of moisture, is present in epigenetically frozen finely dispersed deposits, whereas when the earth materials have frozen syngenetically, primarily in floodplain alluvium, the ice inclusions are uniform within the entire series. Numerous examples have demonstrated the syngenetic growth of the wedges, and the theory of their frontal growth in an upward direction has been proposed.

Later, the concept of the processes involved in epigenetically freezing series was expanded and it was proved that large flat sheets of injected ice are capable of forming at great depths.[4,12] N. G. Bobov[3] and V. V. Baulin[2] extended the theory of "deep" migration of water, advancing the hypothesis concerning the formation of the same sheet ice by the segregation method at depth.

Ye. M. Katasonov,[18,22] G. F. Gravis,[8] and M. S. Ivanov,[16] are developing the geological approach, which is based on the use of the permafrost-facies analysis technique. The primary object of permafrost-facies studies are not the processes nor the cryogenous phenomena themselves, but the genetic varieties, the facies of whose permafrost Quaternary deposits are characterized by specific processes and phenomena. The ice interlayers and lenses, the beds and veins, the various soil wedges, and the involutions of these deposits, equally with those of their attributes such as composition, sedimentary layering, and so on, are considered in connection with genesis and the facies association of the country rock.

The permafrost-facies area in the study of the structure of permafrost is closely intertwined with the thermophysical, general geocryological area inasmuch as it pertains to cryogenic phenomena and ground ice. However, on the theoretical and especially the procedural level, it is closer to lithology--the study of sedimentary rock. This area has been developed in the papers of a number of researchers.

Katasonov[17,18] developed the permafrost-facies analysis technique and demonstrated that the development of cryogenous phenomena is determined primarily by the genesis and facies association of the deposits. He proposed the distinguishing of subareal and subaqueous cryosyngenesis. A diagram is presented that illustrates the formation of cryogenous structures in sediments that freeze from the direction of the frozen substrate under surface conditions and beneath bodies of water;[22] the possibility of the burial and transition of individual ice floes or ice fields in the coastal sediments is demonstrated; it is established that ice and soil wedges form simultaneously, side by side, under identical climatic but different facies conditions.[19] The permafrost-facies analysis of the Quaternary deposits making up the key sections of the Yana, the Omoloy, the Aldan, and the Yenisey made it possible to substantiate the conclusion concerning the continuous formation of permafrost beginning at least with the middle Pleistocene.[20,21]

N. N. Romanovskiy[36-38] used the permafrost-facies analysis technique when studying the Quaternary series of Bolshoi Lyakhovskiy Island. A detailed description was given of the composition, ground ice, and genesis of the deposits of erosion-thermokarstic basins. He was the first to give a permafrost-geological description of the coastal deposits and to detect in them specific soil wedges or "frozen envelopment structures," the formation of which is attributed by the author to the thawing of ice wedges as a result of an increase in the depth of seasonal thawing and filling of the fissures with sediments.

Using as an example the Indigirka lowland, Yu. A. Lavoushin[26] discussed the genesis, composition, and structure of the permafrost alluvium of the subarctic zone. In his view, the formation of wedge ice is primarily connected with the deposits of the channel sandbars and "laidas" (low-lying coastal areas bordering the northern seas in the USSR, usually swampy with hummocky relief, permafrost often close to the surface, flooded during high tides)--this is how this author interprets the sediments of a polygonal floodplain. An especially detailed description was presented on the sedimentary characteristics of the permafrost. Also, the facies conditions accompanying the origination of "envelopment structures," for which the term subaqual pseudomorphoses was proposed, were more precisely defined.

Using the permafrost-facies analysis technique, Gravis[8,9] investigated the slope deposits of Yakutia, in respect of which he worked out the most complete geological-genetic classification of the cryogenic structures; distinguished and characterized the geographic versions and facies of contemporary permafrost colluvium, etc.; and demonstrated that the solifluction and landslide deposits are characterized by deformed cryogenous structures, distorted, and asymmetric ice and soil wedges. The study of the sections showed that in spite of the prevailing concepts, slope deposits are widespread and consist of series of up to 20-30 m thick. The permafrost-facies analysis method was used to analyze these series, distinguish alternating beds with differing soliflunctional and colluvial accumulations, and draw a well-substantiated conclusion that periodic variations in climatic humidity occurred in the Pleistocene. These climatic fluctuations did not lead to degradation of the permafrost, but radically altered the slope sediment formation conditions.

V. A. Usov[45-47] made a significant contribution to the knowledge of the cryolithogenesis of marine series. He investigated the conditions of accumulation and freezing and also of the cryogenous structure of the deposits forming within the confines of an offing (foredelta), of shallow lagoons and tidal marshes, that is, coastal plains that are flooded at high tide. These facies complexes are distinguished in sections of the upper Quaternary series of the Yenisey North and it was shown[47] that the soil wedges in the Kazantsevskiy stratigraphic layer similar to those described by V. V. Baulin and L. M. Shmelev[1] are genetically associated with tidal marshes and the sheet ice with lagoon deposit.

M. S. Ivanov[15,16] presented data on the soil wedges forming in the sediments of the marginal, periodically flooded parts of the delta--in the tidal marsh deposits. He described in detail the contemporary syngenetically freezing deposits that form in an offing with depths of not more than 1.5 m. Their characteristic features are noted to be as follows: (1) salinity of the soil; (2) high content of vegetable detritus and driftwood; (3) comparatively small amounts of ice inclusions, consisting mainly of horizontally or obliquely oriented broken lenses; and (4) the presence of silt-like interstices together with the ice inclusions.

In the basins of the Yana and the Omoloy, G. Ye. Rozenbaum[35] investigated the processes involved in the formation of cryogenous structures and polygonal wedge ice, considered in relation to the distinctive features of development of the facies of the alluvium and delta deposits.

CRYOLITHOGENESIS--THE LEADING LITHOGENIC PROCESS IN THE ZONES OF STABLE COOLING OF THE EARTH

During the last decade more and more attention has been devoted to cryolithogenesis, that is, to the geological processes associated with the origination and the dynamics in the Earth's crust of ice of both rock and mineral in the composition of polymictic permafrost and seasonally freezing earth materials. In other words, cryolithogenesis is the set of processes in the cryolithosphere, that is, in the zones of stable cooling of the Earth, causing the development of cryogenous earth materials and cryogenous relief. Cryolithogenesis is thus the engendering of characteristic physical-geographic zonal conditions.

To a considerable extent cryolithogenesis itself determines the appearance of the environment as a whole--the polar zone in the Northern Hemisphere--and all of Antarctica in the southern hemisphere. It also leaves a definite imprint on the course of exogenic processes when the conditions are those of a humid zone, chiefly in the Northern Hemisphere and everywhere in alpine regions.

The study of cryolithogenesis, like the integral science of lithogenesis in the zones of stable cooling of the Earth,[33] which is known as cryolithology, is understood differently by various specialists. The concepts of "cryolithogenesis" and "cryolithology" coexist in the literature with certain other concepts and terms that more or less resemble them or differ from them.

The term "cryolithogenesis" was introduced by Katasonov[17] as the synonym for the concept of the "lithology of frozen Quaternary deposits." It was he who defined the problems of the new scientific discipline as being the study of the structure and genesis of deposits forming under permafrost conditions. These deposits are the result of the interaction of cryogenesis and lithogenesis during all stages of the soil-forming processes. The order of superpositioning and the very nature of the cryogenetic and lithogenetic processes are determined by the precipitation forming medium and by the permafrost-geological conditions of accumulation and freezing, as a result of which deposits (facies) are formed that differ in their composition and cryogenic structure. The permafrost facies-analysis method facilitates the distinguishing of these deposits in the sections and the ascertaining of their genesis.

Cryolithology, which originated as a result of the development by Katasonov of the permafrost-facies analysis technique, did little or nothing to elucidate the mineralogical-petrographic and geochemical aspects of this scientific discipline. The publication of the well-known monograph by N. M. Strakhov[41] became the turning point in this respect.

V. A. Zubakov[14] proposed that, in addition to the humid, arid, and icy climatic types of lithogenesis, a fourth type--cryogenous lithogenesis--should be distinguished. The latter is divided by him into the following subtypes: cryohumid (glacial) and cryoarid (permafrost). Corresponding to the first subtype are seven cryogenous formations, and to the second type, three. A brief description of these formations is presented.

A. I. Popov[33,34] discusses cryolithology as a division of geocryology that studies the permafrost phenomena in the Earth's crust from the geological standpoint. Cryolithogenesis is the permafrost geological process. It includes the following: (1) cryogenous diagenesis, the geological essence of which consists in the formation of ice and the resulting transformation of the mineral substrate itself, i.e., the sediments and previously lithified loose soils (compaction, dehydration, and so on); and (2) frost weathering--the disintegration and staged transformation of any rock to the ultimate fine-grained product (in silt-cryopelite), as a result of seasonal freezing and thawing. It is due to the operation of these geological permafrost processes--cryodiagenesis and cryogenous weathering--that cryogenous earth materials are formed: cryolites that are monomineral glacial soils and cryolithites that are polymineral icy solids, both as a result of cryodiagenesis, and cryoeluviites, which are the products of frost weathering. Epigenetic and syngenetic freezing of earth materials are discussed as independent types of cryolithogenesis.

N. A. Shilo[50] distinguishes the type of periglacial lithogenesis. The latter takes place in the regions that are characterized by below-freezing mean annual temperatures, a deficiency of precipitation and the occurrence of water in the solid and liquid phases. With reference to his previous papers[49] devoted to the distinctive features that characterize the formation of placers in the permafrost region, N. A. Shilo notes the role of periglacial (cryogenous) processes in the chemical and physical-chemical changes of the residium constituting the seasonally thawing layer in water catchment basins. The transformation of the aluminosilicates is as follows: (1) hydromicas--beidellite-montmorillonite, which in general is caused by an acid or neutral medium; (2) muskovite--hydromicas-kaolinite (rarely); ferromagnesium silicates--hydromicas. Attention is being given to the development of ice wedges and lacustrine thermokarst in the lowlands of the sub-Arctic.

Like N. A. Shilo, Sh. Sh. Gasanov[6] proposes the distinguishing of a fourth type of lithogenesis--cryolithogenesis--and touches upon aspects of the physical-chemical weathering of the residium in watersheds. He suggests that the leading factors of "slope denudation in the cryosphere" are frost sorting and solifluction. The ultimate runoff bodies of water in the cryosphere are described: The predominance of terrestrial sediments and the hydromicaceous-montmorillonitic composition of the argillaceous minerals, etc., are noted.

N. N. Lapina *et al.*[27] and I. D. Danilov[10] used the term "polar lithogenesis" mainly in order to describe deposits of arctic bodies of water from the point of view of their granulometric and mineral-petrographic composition, the content of organic matter and salts, the formation of concretions, etc.

G. P. Mazurov[28] understands cryolithogenesis to mean the set of physical, physical-chemical, and chemical processes involved in the alteration and transformation of any earth materials and minerals during freezing and thawing, as well as those in the frozen state. He limits cryolithogenesis to the layer of seasonal freezing and thawing.

Thus, the concepts of "cryolithology" and "cryolithogenesis," like that of "genesis," are defined differently, which is natural, for the researchers representing the different areas in geocryology are discussing different facets of the study of the permafrost of the Earth's crust. The differences of opinion can be overcome if we do not become confused and try to delimit the phenomena of a different order and to give a clear definition of them.

All researchers consider that cryolithogenesis is the lithogenesis (pariglacial, polar) in the zones of stable cooling of the Earth at below-freezing temperatures, that is, in the permafrost region and the region of deep seasonal freezing.

Beginning with the generally accepted definitions of the terms lithogenesis and lithology (*Geologicheskiy Slovar* [Geological Dictionary], 1960), Ye. M. Katasonov considers that it would be advantageous to expand the proposed formulation and present it in a more detailed form. According to his definition, cryolithogenesis is the set of geological (hypergene, sedimentation, postsedimentation) and cryogenous processes occurring in all of the stages of soil formation and leading to the evolution of permafrost deposits. Cryolithology is the science that treats of soil formation in zones with below-freezing temperatures, the science that deals with cryolithogenesis. It studies the permafrost, i.e., the cryolithogenic deposits forming in this zone.

In Katasonov's opinion, a zone of negative temperatures differs radically from a zone of seasonal freezing. The deposits forming within these zones' boundaries are similar in composition and in the appearance of certain earth materials that are subject to freezing and thawing. With respect to all other characteristics, however--the conditions of formation, structure, and state--they differ markedly. Therefore, the set of hypergene, sedimentation, postsedimentation, and cryogenous processes leading to the accumulation of seasonally freezing deposits is logically known as seasonal cryolithogenesis, emphasizing not so much its similarity to as its difference from cryolithogenesis proper. Seasonally freezing deposits, by virtue of their intermediate position in the overall soil-formation scheme, will obviously become a subject for investigation in one of the divisions of cryolithology. This is how Katasonov sees it.

A. I. Popov agrees with Katasonov that the restriction of cryolithogenesis solely to the participation of freezing in the sediment accumulation and cryolithogenic permafrost-formation process narrows down this concept unjustifiably. According to Popov, cryolithogenesis is any effect of the cryolithogenic factors on a sediment in the process of being formed, on unconsolidated rock and dense crystalline rock, etc., when this sediment or rock is undergoing changes--in the form of primary or secondary lithification or is being subjected to frost weathering; that is, when the freezing itself is interpreted as a geological process. He considers it inadvisable to distinguish the periglacial type of lithogenesis as though occupying an intermediate position between the icy and humid types of lithogenesis (the latter after Strakhov).[41]

The presence of water in the liquid phase during the warm season of the year, either in glaciers or in any event near them, and also the amount of precipitation that, in glacial regions as in the humid zone, exceeds evaporation, all of this makes it impossible to regard as rigid the distinguishing of the icy type of lithogenesis as an independent type of rock formation.

The geological activity of glaciers (and the snow patches or nivation remaining throughout the summer), especially as its resultant, are so closely dependent on cryogenous factors that unquestionably the so-called icy type of lithogenesis must be considered together with cryolithogenesis, and not treated independently of it. Thus, once it is admitted that it is meaningless to distinguish the glacial type of lithogenesis, it follows that there is no point in distinguishing the periglacial type. Only the cryogenic type of lithogenesis or cryolithogenesis, the zone of which is located northwards of the humid-type zone, deserves to be distinguished independently.

Aspects of the geographic zonation of cryolithogenesis were investigated by A. I. Popov, using northern Eurasia as an example, in view of the extensiveness with which this zonation is most fully manifested. Within this vast expanse of land and water the contemporary zonation and regional picture of cryolithogenesis and the distribution of its morphogenetic types are determined by the following basic geological-geographic and historical factors:

1. The factor of latitudinal zonation, i.e., a geographic zonation of higher rank, is predominantly responsible for determining the spatial structure of cryolithogenesis, although it is not the only operative factor.
2. A subordinate factor that essentially is also a zonation feature, but of lower rank and oriented at an angle to the latitudinal zonation. As applied to Eurasia, this factor is the land-to-sea ratio, that is, the factor of the degree of continentality of its various parts.
3. The neotectonic factor, determining the basic trend in the development of exogenic processes, under prevailing denudation or accumulation of sediments or the tapering off of both processes and the establishment of relative stabilization of sedimentation and ablation. Here it is also necessary to consider the vertical zonation factor.
4. The substratum factor or lithologic factor.
5. The historic factor, that is, the varying displacement of the natural zones in different places of the vast continent of Eurasia during the Pleistocene and Holocene. This factor is also operative in other respects.

Whereas the latitudinal geographic zonation factor determines the basic features of the spatial texture of cryolithogenesis, all of the rest (except the historical factor), operating together, by means of their combination and superpositioning on one another, give rise to major provincial differences in the types of cryolithogenesis. The division of the cryolithogenetic regions into districts is based on consideration being given to the organic connection between all of these factors.

Cryolithogenesis, as a geological phenomenon, is taken to be a residual process in some cases and a diagenetic process in others. Corrresponding to each of them is a definite set of cryogenic formations--earth materials and relief forms.[33]

The peculiar structural characteristics of the cryogenic earth materials forming regularly patterned permafrost permit the distinguishing of the aforementioned two principal cryogenic complexes (or types)--epicryogenic and syncryogenic.

The structural peculiarities are caused both by the method of freezing--epicryogenic (after the formation of the earth materials) and syncryogenic (during the sediment accumulation process)--and by the temperature conditions under which freezing occurred, and, consequently, by the varying degree of activity of the freezing process.

The temperature conditions under which freezing takes place and the associated differing degree of activity of the freezing process, both of which predetermine the emergence of specific ground patterns and typical structural features, make it possible, on the basis of the latter, to distinguish the genetic layers along the vertical in epicryogenic and syncryogenic series of earth materials.

The geographic zonation features of cryolithogenisis of the second and higher orders are due primarily to the relation between the genetic layers typical of the epicryogenic and syncryogenic series in the various subzones.

In the zonal sense, however, epicryogenic series have the greatest significance, and therefore the distinguishing of the subzones is most strikingly demonstrated in cryogenous formations of this type. It does not follow from this that the syncrogenic type drops away from the field of view of the investigator attempting a zoning of the processes of cryolithogenesis and the formations originating as a result of them.

As has already been stated, the subdivision into genetic layers is carried out in relation to the temperature regime of the series, this regime predetermining the variable course of the freezing process at different levels along the vertical and giving rise to a varying morphological effect at these levels. In this sense the layers distinguished in basement rock massifs display a definite variation from one subzone to the other, but they are still more significant in the sense of zoning the subtypes of an epicryogenic series corresponding to the series of frozen loose sediments and earth materials.

Within series such as these, the following layers are distinguished from top to bottom: (1) active cryohypergenesis (with a phase change), (2) active, and (3) passive cryodiagenesis.

If we exclude the narrow strip of the shelf of the polar seas where cryolithogenesis develops rather specifically and azonally, the entire dry land of the northern Eurasian continent, which is subject to the effect of cryolithogenesis, affords an example of its manifestation to a sufficiently zonal degree, and this zonation is primarily expressed in a north to south shift of the relations between the aforementioned genetic layers of the epicryogenic series.

The overall trend in the north-south zonal variation of the nature of cryolithogenesis is associated with the variation of both the mean annual temperature and, to a large extent, the temperature gradients in the fall and winter season. Here, we have in mind both the contemporary stage and the period when the permafrost was formed.

On the basis of the corresponding cryostructural characteristics, within northern Eurasia four subzones of sublatitudinal orientation--polar, subpolar, boreal, and subboreal--are schematically isolated. Each of them is characterized by its own thickness of the cited genetic layers, by winter temperature gradients within their confines with a definite thickness of snow cover, by distinctive morphological attributes in the form of cryogenous weathering and cryogenous structures, and by large ice formations such as cryoliths, etc.

It is to be noted that within the polar and, in part, the subpolar subzones, frost fissure formation, which stands out mainly as a diagenetic factor and leads here to the formation of polygonal wedge ice and the associated low-center polygonal relief, stands out as a frost-weathering factor within the boreal and especially the subboreal subzones, promoting the formation of polygonal-ground wedge systems accompanying loess-type soils and the formation of polygonal hummocky-sinkhole relief. This regular pattern, along with the others, illustrates exceptionally clearly the zonal nature of the processes of cryolithogenesis.

The general trend in the zonal variation of cryolithogenesis from north to south in northern Eurasia, caused both by the mean annual temperature and by the temperature gradients in the winter, is expressed in the succession of geographic subzones, with a predominance in the north of products of cryodiagenesis, and in the south, of products of frost weathering.

The laws tracing the zonal structure of cryolithogenesis, as exemplified by northern Eurasia, are now being schematically revealed and, as already noted, the numerous aforementioned factors markedly complicate the overall picture of the spatial structure of this already complex natural phenomenon.

COMPOSITION AND STRUCTURE OF CRYOGENIC EARTH MATERIALS

In recent years, significant progress in the study of the composition and structure of cryogenic earth materials has been noted. This is to a considerable extent associated with the in-depth study of argillaceous minerals and their transformation under the influence of freezing and the organic materials in the freezing processes, which is of great importance in gaining knowledge of the essential principles of the processes of frost weathering.

Another very important area is the in-depth study of the microstructure of frozen ground and ice, which permits us to draw significant conclusions for use in the interpretation of their genesis.

The progress achieved in the aforenoted areas does not remove the problem of studying the macrostructure of frozen and thawed cryogenous earth materials, for the formation of these peculiarities proceeds in its own way, not always and not wholly connected with the origination and development of the fine structure. At the same time, a knowledge of the major structural features is of great importance in solving the problems of cryolithology and geocryology as a whole, both in scientific and practical terms.

For example, it is known that the structure of the frozen and relict, no longer frozen, Quaternary deposits is widely used for different types of paleogeographic and, in particular, paleoclimatic syntheses. Here, as a result of the inadequacy of the morphogenetic criteria applied to cryogenous formations of differing genesis, significant miscalculations and errors are assumed. It is sufficient to indicate the isomorphism of certain cryogenous and noncryogenous formations (for example, the wedge-type pseudomorphs replacing polygonal wedge ice and wedge-like formations caused by the convective instability of the liquid sediments) as the cause of numerous errors on the part of geologists and the erroneous conclusions arising from them.

It must be acknowledged that the most important problem in cryolithology is the study of the laws governing the transformation of the mineral constituent of the earth materials under the conditions of the cryosphere. The solution to this problem is connected primarily with the study of the composition and properties of the sedimentary formations emerging in the presence of stable and long-term freezing of the Earth's crust.

During the last decade V. N. Konishchev has obtained many data on the distinctive features of the composition of the deposits that have formed in the cryogenesis zone--the loesslike formations, the moraines, the glacial-marine deposits, and the synthetic permafrost of the Yana-Indigirka lowland. These data indicate that the most important compositional features of the sedimentary formations characteristic of the cryosphere originate during the stage when matter is being mobilized in the weathering crust, that is, in the course of the formation of cryogenous residuum. In turn, as indicated by the results of studying typical sections of the latter, its main qualitative characteristics are primarily connected with the effect of multiple alternating processes of freezing and thawing on the various types of parent rock.

The analysis of the present-day concepts of the interaction of water with various groups of minerals in relation to temperature and ice formation in the ground, and also experimental data, have enabled V. N. Konishchev to arrive at the conclusion that the cryogenous disintegration of the rock is associated with two groups of processes--the cleaving effect of the ice, leading to coarse comminution of the primary minerals, and hydration disintegration (basically physical-chemical), which leads to significant transformations of the structures of the inherited argillaceous minerals. Along with the difficulty of synthesizing argillaceous minerals from the products of weathering of massive-crystalline rocks,[44] this leads to enrichment of the cryogenous residuum and the products of its next redeposition with colloids--hydrated organic mineral compounds consisting of degraded forms of argillaceous minerals, amorphous groups of SiO_2, Al_2O_3, Fe_2O_3, and organic matter.

As a result of investigating the microstructure of natural frozen ground and an experimental study of artificial mixtures subjected to the freezing and thawing cycles, V. V. Rogov obtained data permitting a more exact evaluation to be made of the thin structure of these rocks during freezing and some of the new structural peculiarities of ice-cement and segregated ice. Moreover, Rogov developed an entirely original method of studying the permafrost microsections under room conditions. The value of this microstructural research for gaining an understanding of the genesis of frozen and previously frozen rock is difficult to overestimate.

During the last decade other important studies have been made of the structure and composition of the frozen contemporary and relict cryogenous formations. The problem of the formation of cryogenous structures was originally examined in connection with the peculiarities of the migration of water in the finely dispersed soils and in connection with the heaving phenomenon in them. Sh. Sh. Gasanov[6] classified cryogenous structures according to the granulometric composition of the soils.

For the first time a published study was made of the mechanism involved in syngenetic sediment accumulation and segregated ice formation.[33] It was established that the increase in the syncryogenic permafrost (as a result of nonthawing of the base of the active layer during sediment accumulation) takes place, not by an annual increment of the frozen layer equal in thickness to that of the sediments accumulated during the year, but by the transition to the permafrost state of entire "packets" of layers in which reticular and layered cryogenic structures are well defined. A similar syncryogenesis procedure arises from the relationship between the rates of sediment accumulation and cyclical variation in the depth of thawing of the active layer over a number of years--and the cyclical nature of the climate. This problem is dealt with later by L. N. Maksimova.[29]

Attention has been specifically devoted by researchers to problems pertaining to the cryogenous structure of the seasonal freezing and thawing layer, that is, the active layer. As a result of detailed studies, Ye. A. Vtyurina has isolated the classes, subclasses, and types of cryogenous composition of the rock in the layer of seasonal thawing, which enables it to be mapped according to the peculiarities of the cryogenous structure.

It is also possible to include within this category the elucidation of the laws governing the formation of the cryohypergenesis layer (this is, the active layer) on the basis of cryostructural characteristics (the report by N. V. Tumel' and Yu. V. Mudrov and the report by Ye. M. Naumov on the effect of cryogenous factors on the soil).

In recent years a number of papers have been published by G. M. Fel'dman and V. G. Melamed on problems relating to research into the thermophysical principles of the formation of cryogenic structures in which methods of calculating the thickness of ice streaks over the entire depth of the active layer are proposed.

Problems pertaining to the composition of frozen and freezing rock have been treated in many papers published in recent years. These problems are discussed specifically in the reports by F. N. Leshchikov and T. G. Ryashchenko and also by M. N. Uskov, as applied to argillaceous soils and minerals, which, as is well known, are of very great interest in the study of cryogenic phenomena. We have the least data of all about the ice and the dense (crystalline, metamorphic and so on) rocks (the report by S. Ya. Zhukovskiy *et al.* in this volume).

It has already been noted that indications of past freezing and the formation of cryogenous rock, as expressed in large wedgelike structures or in involution and cryoturbation structures, are widely used for paleogeographic reconstructions and climatic stratigraphic synthesis. However, the procedures for using them in this way have not yet been perfected, and their reliability needs to be improved.

Nevertheless, for paleocryological synthesis, it is possible to use both contemporary "live permafrost" and "fossil" residual traces in the deposits and the weathering crust.

Fossil "frozen ground" is especially popular among specialists in Quaternary geology. Actually, the traces of the cryogenous processes in the active layer and the underlying permafrost layer are most clearly recorded. These are chiefly polygonal ground wedges, which, under the conditions of deep seasonal freezing and thawing, are an important index of the degree of severity and continentality of the climate in the past.

Degradation of polygonal wedge ice can lead to the emergency of ground pseudomorphs substituting for ice wedges. Unfortunately, the fact of isomorphism with wedgelike shapes caused by convective instability (see the papers by A. G. Kostyayev) complicate the diagnosis of the pseudomorphs. A serious attempt to surmount these difficulties when studying pseudomorphs substituting for ice wedges has been made in recent years by F. A. Kaplyanskaya and also by A. A. Arkhangelov in the Kolyma Basin. A convincing analysis of pseudomorphs indicating the presence of permafrost in the lower Pleistocene deposits in the northeastern USSR is contained in a paper by A. A. Arkhangelov and A. V. Sher (in this volume).

Later, in the section on ground ice we cite new data pertaining to the presence, not only of pseudomorphs, but also of the same "live" syngenetic polygonal wedge ice in the middle and, perhaps, lower Pleistocene deposits of northeastern USSR, the main paleoclimatic significance of which is the evidence it provides of the continuous existence of permafrost, probably since the beginning of the Pleistocene. Some of the studies made in the last decade of the composition and structure of permafrost have been noted earlier.

Questions pertaining to the genesis, composition, and structure of frozen ground and earth materials are discussed in reports and papers devoted to broader topics. On the genetic level the paper by N. N. Romanovskiy on the effect of the latest tectonic movements on the formation of permafrost is of interest. Also meriting attention is a paper by N. F. Grigor'yev, which reveals the role of cryogenous factors in the formation of coastal-marine placers. A report by B. I. Prokopchuk on the formation of diamond placers under permafrost conditions is close to the latter in terms of its subject matter.

GROUND ICE AS AN INDEPENDENT PROBLEM

Ice occurring in earth materials is called ground ice regardless of whether it forms large ice bodies or is scattered in the ground in the form of fine crystals. The problems of the general classification of ground ice can be regarded as already solved, and, therefore, the last decade has not yielded any fundamentally new principles in this respect. Only certain refinements have been introduced into the existing classifications, of which the most complete is the classification by P. A. Shumskiy.

Polygonal wedge, vein, injected, and segregated ice and ice-cement deserve the greatest attention from the standpoint of their widespread nature and the extent of their involvement in the permafrost. B. I. Vtyurin pointed out (in this volume) that sill-forming ice amounts to 10 percent of the total reserves of the apparent ground ice (without allowing for the ice-cement) and that, of this ice, it is only the polygonal wedge ice and the sheet deposits of segregated and injected ice that are the most widespread and of great practical interest. According to his calculations the reserves of apparent ground ice in the USSR amount to about 19,000 km^3, and, on the Earth as a whole, about 35,000 km^3. The reserves of polygonal wedge ice, estimated from the sill-forming ground ice, amount to about 1,000 km^3.

A number of general problems pertaining to ice formation in rock and the formation of large ice bodies have been discussed by Ye. A. and B. I. Vtyurin.[4,5]

Of special interest to geocryologists and cryolithologists studying ground ice are problems relating to the genesis of polygonal wedge ice and thick sheet ice. Polygonal wedge ice, the most important component of thick syncryogenetic series, stands out as the most independent genetic type of ground icing, forming a massive lattice, typically of monomineral ice rock over the broad expanses of northern Eurasia, Alaska, and Canada.

The debate about the genesis of polygonal wedge ice is continuing to this day. Undoubtedly the complex mechanism resulting in the development and syngenetic growth of this type of ice during the course of sediment accumulation has not yet been fully revealed, although efforts to explain it have been made by many researchers in recent years.

Accordingly, it is important to note that the concept of the so-called wedge mechanism of development of this type of ice as the only true mechanism (according to B. N. Dostovalov and P. A. Shumskiy) is not completely acceptable if we take into account the structural peculiarities of the contacts in the ice wedges (the frequent absence of deformations in the country rock), the presence in the ice of horizontal layering elements, and the peculiarities of the ice structure. Nevertheless, some researchers, apparently oblivious of the major difficulties encountered in explaining the genesis of wedge ice, continue to regard its wedge mode of development as the only valid one. Of course, the role of this mechanism is initiative, important, and in many respects determining, but it is not the only one and is certainly complicated by the so-called frontal growth mechanism that is recognized by many researchers.

Among the serious attempts being made to overcome the impasse in explaining the mode of syngenetic growth of wedge ice is the hypothesis of V. N. Konishchev and A. D. Maslov.[23] According to this hypothesis the frontal growth of the ice wedges takes place by squeezing out of the ice as a result of lateral pressure being exerted by the frozen country rock with variation of the temperature regime. However, this hypothesis also is not entirely acceptable, for it does not explain a number of peculiarities in the structure of the entire ice-mineral complex.

According to A. G. Kostyayev,[24] the wedge-type and dome-shaped uplifts of the layers above the ice wedges are caused by the pressure of the wedges themselves or the parts of them protruding upward. He believes that this phenomenon can only be explained in terms of the development of mechanical displacements in the ice-frozen rock system. The ice, being a more fluid material (under long-term stresses) than the country rock, is slowly squeezed upward by the rigid polygonal skeleton.

An attempt at providing a physical substantiation of the squeezing out of the ice wedge by the country rock is made in an article by Ye. V. Artyushkov. Unfortunately, these two hypotheses are characterized by the same weaknesses as the preceding one.

V. I. Solomatin's[40] studies were conducted along other lines. He concluded that segregated ice formation plays an important role in the formation of ice wedges, and that the ice wedges increase in width as a result of water migrating laterally from the country rock. He was the first to propose a classification of the structures of polygonal wedge ice. The structural peculiarities are determined by the introduction of inclusions from the country rock (due to segregated ice formation at the lateral contacts) and the overlapping rocks (as a result of the processes of frost fissuring and the formation of elementary veins) and also by the redistribution of inclusions, primarily gas inclusions, under the influence of oriented lateral stresses.

Sh. Sh. Gasanov,[7] in describing the self-regulating mechanism affecting the increase in width of ice wedges because of the gradual variation in the dimensions of the frost fissure as the ice wedges grow, a consequence of a decrease in the summated coefficient of thermal deformation of the ice and the frozen ground, arrived at the conclusion that as the wedges increase in width the fissure becomes narrower and the depth of its penetration decreases. A gradual accumulation of these quantitative indices takes place until at a certain stage they attain critical values, preventing any further increase in width of the wedges.

A paper by N. N. Romanovskiy[39] is devoted to explaining the effect of the temperature regime of the rocks on frost fissure formation and the development of the polygonal wedge ice. He shows that the mechanism of syngenetic polygonal wedge ice formation varies in accordance with the permafrost-temperature zonation. When there is a low mean annual temperature, besides the formation of ice in the frost fissures, frontal growth also takes place as a result of the accretion of streaks of segregated ice on the upper surface of the wedges. When there is a high mean annual temperature, frontal growth is absent or limited and it does not show up in the structural character of the ice wedges.

Progress made in the study of the Cenozoic stratigraphy (including that of the Pleistocene) of the Yana-Indigirka and Kolyma lowlands has made it possible for V. N. Konishchev[23] to speak of two accumulative frozen syngenetic series with two generations of polygonal wedge ice of different age (except for the terrace ice of course). These two ice complexes are estimated to be of upper and middle Pleistocene age, although a Lower Pleistocene age of the lower stage of the series is not ruled out. The cardinal paleocryological significance of this fact has already been noted.

The initial datings of the absolute age of the ground ice in the northeastern part of the USSR were obtained by the radiocarbon method (V. N. Konishchev, Yu. A. Lavrushin, and I. Ye. Timashev). The oldest ice investigated on the basis of these determinations dates from more than 60,000 yr ago.

The further development of methods of determining the absolute age of ice is necessary, the most promising of which is obviously the thermoluminescence method. Full cognition of the mechanism of development of polygonal wedge ice and especially its syngenetic type is a matter for the future.

The study of the sheet ice characteristic of epicryogenic series has permitted the discovery of several types in which either segregation processes or injection processes or combinations of them have played a part (V. V. Baulin, G. I. Dubikov, N. G. Bobov, and Sh. Sh. Gasanov). An unusual type of sheet ground ice which is a variation of ground icing has been described for the polar Urals by V. A. Usov.

The study of the genesis and distribution of sheet deposits of ground ice in the various parts of the permafrost region through the use of various methods of investigation has made it possible to trace the formation of this type of ice from nucleation to melting.[3] The aggregate of features, in the author's opinion, confirms the segregation method of their formation which, to be more precise, should be termed multiple segregation.

However, the investigation of the cryogenous structure of the Quaternary deposits of the Yamal peninsula, including the sheet ice, led M. M. Koreysh to the conclusion that many aspects of the theory of cryogenous structure, and in particular that of the segregation and injection genesis of the bulk of the sheet ice deposits in marine Quaternary deposits, are without a firm foundation on account of the limited progress made in developing the physical theory of the crystallization of the ice in dispersed rock. He points out that one means of investigating the cryogenous structure and genesis of frozen Quaternary deposits could be the experimental simulation of the freezing process under laboratory and field conditions.

The immense importance of pressure head migration of water, both in the formation of fine inclusions and of major deposits of ground ice, is emphasized by G. F. Gravis.[9] In his opinion, segregation and injection processes frequently take place simultaneously and are closely intertwined with each other. Injection also takes place in argillaceous rocks.

In a number of cases allowance must be made for the presence of sheet ice of diverse origin, i.e., marine, river, or icing, which has been buried by sediments, and in certain mountainous and foothills regions, of glacier ice as well, although there are few reliable data on buried ice. The texture-forming ground ice--segregated, injected, and ice-cement--has been discussed in part in the preceding sections and there is no need to dwell upon them here.

In conclusion, it must be noted that, in the study of the structure of permafrost, until recently the method of visual observation and description has predominated, which continues to be of paramount importance and cannot be replaced by any other method. However, the development of microscopic methods of studying cryogenous structures and textures is already making it possible to discover new laws in the structure of permafrost and to draw important conclusions of a genetic order. In the near future, therefore, serious attention should be devoted to the introduction of precise petrographic methods in the study of the structure of frozen and freezing rocks.

The study of the relations between cryogenous structure and the conditions of freezing will call for a combining of the thermophysical, physical-chemical, and petrographic subject areas in the study of permafrost development.

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GROUNDWATER OF THE PERMAFROST REGION OF THE USSR

N. I. TOLSTIKHIN Leningrad Mining Institute, and O. N. TOLSTIKHIN Moscow Economic Research Laboratories of VNIIVO

The decade that has passed since the previous international conference on permafrost (in the United States) has been marked by an intensification of the development of the natural resources of the North and Siberia, i.e., the whole of the vast area constituting the permafrost region of the USSR. The discovery of new deposits and the acceleration in the rates of development of the known gas and oil deposits in western Siberia; the gas and diamonds of Yakutia; the ore deposits of the Noril'sk area; the ore and placer deposits of gold and nonferrous metals of central Aldan; the basins of the Yana, Kolyma, Indigirka, and Anadyr' rivers; and also the deposits of the northern Urals, Chutkotka, and Transbaikal has been the driving factor in the expansion and deepening of the research aimed at making efficient use of water resources, including groundwater resources. The development of the latter, in turn, has necessitated an increase in the rates of production of hydrogeological surveys and exploration; the development of new, rapid methods of performing these types of operations and of ways of utilizing and maintaining wells, irrespective of the permafrost-hydrogeological and climatic conditions; and also forecasting changes in the amount and composition of the groundwater with various inflows and partial freezing of the water-bearing horizons (aquifers).

In view of the fact that the rates of development of the permafrost zone are outstripping the pace of its investigation, a need has arisen for making a number of generalizations of the available data. These have found expression in the compilation of hydrogeological and hydrochemical maps of the USSR, drawn to various scales; a map of the mineral water and natural reserves of goundwater in the USSR together with a monographic description of them; a map of the usable reserves of groundwater in the USSR; the multivolume publication *Hydrogeology of the USSR*, which incorporates hydrogeological regional maps; and a number of major studies conducted in individual areas of hydrogeological research (thermal and mineral water, icing, the formation of groundwater, including saltwater and brines as sources of chemical raw materials, and so on). These studies have been performed by the territorial geological administrations of the USSR Ministry of Geology under the guidance of the All-Union Scientific Research Institute of Hydrogeology and Engineering Geology (VSEGINGEO), the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences, the territorial administrations of the USSR Hydrometeorological Service under the direction of the State Hydrological Institute, the All-Union Scientific Research Geological Institude (VSEGEI), Moscow State University, the Leningrad Mining Institute, the Water Problems Institute of the USSR Academy of Sciences, the Scientific Research Institute of Geology of the Arctic, the Central Institute of Treatment at Health Resorts and Physiotherapy, and numerous planning, construction, and branch institutes and organizations.

The methodology of permafrost-hydrogeological research incorporates a fundamentally new approach to hydrogeological survey operations, which consists in isolating key sections. This method, which is being widely used in hydrogeological mapping throughout the USSR as a whole, has been further developed as applied to regions characterized by various permafrost conditions. The efficient combination of detailed research in these sections, together with interpretation of data obtained over the remainder of the area and the extensive use of air photo interpretation and geophysical and computer techniques, has made possible a marked increase in the speed and economic efficiency of permafrost-hydrogeological surveys, as well as standardization of the information obtained. Studies aimed at the determination of an efficient set of exploratory operations for water under the conditions obtaining in platform regions containing thick permafrost, where the main object of the exploration is the water of taliks beneath lakes and riverbeds, must be regarded as a continuation of the survey operations. The stage-by-stage conduct of air-photo interpretation, geophysical exploration, and geological-geothermal studies, and later of exploratory or direct prospecting-exploitation drilling, has made it possible to identify these water resources with maximum reliability and at minimum cost.

The studies of the natural groundwater resources were conducted in two main areas: the first is connected with identifying the peculiarities of formation of the river runoff hydrograph and the principles of its analysis and with isolation of the groundwater component; the second, with the study of the processes of naled formation and determination of the magnitude of the naled feed as an index of the natural groundwater resources. The studies in this second area called for methodological validation and conducting a set of routine observations of naled formation under various permafrost-hydrogeological conditions. These studies have made it possible to arrive at a more precise definition of the methods of calculating the natural resources, including the naled feed and the runoff, and of the interrelation between naled phenomena and the overall permafrost-hydrogeological zonation in mountainous regions, thus contributing to the further development of the theory of the formation of both groundwater runoff and naleds.

The new progressive methods have made possible a marked expansion in the scope of regional hydrogeological research and a deepening of the theoretical hydrogeological studies associated with the processes of cryogenesis, as well as a more accurate definition of the effect of the permafrost region on the formation of groundwater.

The concepts of the structure of the negative temperature zone and its hydrogeological significance have been substantially broadened.

Negative temperature zone (cryolithosphere), permafrost region and cryopeg (body of liquid saline water below 0°C associated with permafrost) zone. The negative-temperature zone* is made up of two zones: the upper zone contains water predominantly in the solid phase (the frozen zone), while the lower zone contains negative-temperature saltwater and brines (the cryopeg zone). In continental deposits the section of the negative temperature zone is usually expressed only by the frozen zone, and the cryopeg zone is either absent or negligibly small and cannot be taken into consideration in the plotting of hydrogeological curves. On the other hand, within areas made up of lagoon and marine deposits (especially salt-bearing or gypsiferous deposits), the cryopeg zone may greatly exceed the thickness of the frozen zone. Sometimes a transitional zone in which permafrost lenses alternate with cryopeg lenses is observed between the frozen and cryopeg zones.

The most convincing information concerning the structure of the negative-temperature zone was obtained in the plains regions of the eastern part of the Siberian platform. The observations of the rock temperatures and the state of the water included in the rocks indicated the presence of two types of sections formed in similar climatic situations but under different hydrogeological conditions. The first (in continental Cretaceous deposits) is characterized by matching thicknesses of the negative-temperature zone and the frozen zone, below which at a depth of approximately 400-650 m there is fresh subpermafrost water with a positive temperature. The second type of section is formed in the carbonaceous gypsum-bearing and saline deposits dating from the lower Paleozoic to the upper Proterozoic, where the thickness of the permafrost zone decreases to 180-200 m, while below there is a cryopeg zone of up to 1,250 m thick. Hence, the total thickness of the negative-temperature belt is as much as 1,450 m.

Ground cryopegs, in addition to being widely distributed in the northern part of eastern Siberia, are abundant on the islands of the Arctic Ocean, along its coastline, in the artesian basins of the extreme northeastern part of the USSR, being discovered away from the sea, and in subaqual hydrogeological structures. Cryopeg lenses have also been detected in the permafrost region itself. In the upper part they are often connected with human activities. Once originated, they may develop and migrate downward and laterally as a result of cryogenetic processes and the effect of the gravitational factor in a varying temperature field.

When discussing the problems of permafrost distribution, in addition to its traditional breakdown into continuous, discontinuous, and insular, from the hydrogeological standpoint the nature and distribution of taliks is of special interest, that is, of areas of melted (thawed or unfrozen) earth materials among frozen earth materials, through which an interconnection between subpermafrost, intrapermafrost, and suprapermafrost water is possible. The studies of the last decade have added much that is new to the theory of the formation of taliks and primarily to questions pertaining to their genetic classification and also the boundary conditions that are necessary and sufficient for their formation. In particular, on the basis of their genetic characteristics, five types of taliks have been distinguished, depending on the nature of the key process leading to their formation or

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* By the structure of the negative temperature zone we mean the distribution of the phase state of the water in its cross-section.

preservation. These are thermal radiation, underwater, chemogenic, technogenic, and vulcanogenic taliks.

Thermal radiation (or rainwater-radiation) taliks are formed under the influence of radiation heat and the heat from percolating atmospheric precipitation. They are for the most part typical of an area of discontinuous permafrost. The presence of such open taliks is a criterion for the distinguishing of discontinuous or continuous permafrost. Usually they serve as reliable paths of groundwater feed.

Underwater taliks are the main taliks in a region of continuous permafrost. However, along with thermal-radiation taliks they are also widely prevalent outside this region. Their formation can be associated either with the heating effect of a body of water in the absence of a percolation flow (certain taliks below lakes and estuaries or on coastlines) or with the effect of the percolation flow (the taliks below a riverbed, certain open taliks below lakes, debris cones, thermal springs, and so on). Frequently the permafrost region is affected by both factors in combination. The presence of taliks below a lake when platform structure conditions prevail is not always sufficient for an interconnection to exist between the subpermafrost water and the suprapermafrost and surface water. This linking is only achieved in the event of the open talik being coupled with a tectonic fracture zone or with a favorable lithologic section, in particular, the absence of water-confining strata. Open taliks under riverbeds formed in the presence of a percolation flow in alluvial river deposits and the jointed zone underlying them cut across areas of great length, sometimes transversely to the strike of the structures, and are of crucial importance to the achievement of an interconnection between the subpermafrost and the surface waters.

Chemogenic taliks occur as a result of the exothermal reactions of the oxidation of coal, sulfide ores, or other geochemical processes taking place with the release of heat. However, they can also be formed without the release of heat, for example, with the dissolving of rock salt and the formation of cryopegs, and they can have a negative temperature. These taliks are rarely encountered and do not govern the development of an interconnection between the groundwater and the surface water on a broad scale.

Technogenic or artificial taliks have become increasingly prevalent. They are connected with various types of engineering structures: heating systems, operating wells, buildings with high heat loss, and so on. In addition, the destruction of the turf cover, the flooding of reservoirs, the building of water conduits, and land reclamation measures frequently lead to regeneration of "natural" underwater and other types of taliks.

Vulcanogenic taliks are not widespread, and they have been little studied. The analysis of the effect of open taliks on the interconnection between surface and groundwater has made is possible to separate the permafrost zone into cellular and the block-cellular zones that are typical of plains richly studded with lakes, and block zones that are typical of mountainous countries and plains with few lakes where the block massifs of the permafrost zone (even if they occupy up to 95 percent of the entire territory of a mountainous country) are separated from each other by talik crevices.

For a long time it was assumed that the area of water basin necessary and sufficient for the origination of open taliks under riverbeds in hydrogeological massifs consisting primarily of terrestrial rocks is measured in thousands of square kilometers. This figure (when the conditions are such that coarse clastic material is accumulating in the river valley and on the slopes of the watersheds) was later reduced to 500 km^2. Information derived from direct exploration for water and a statistical analysis of the distribution of large naleds and river basins, as indices of the presence of open taliks, studies of the snow cover distribution, icing regime, and river runoff in the watershed parts of river valleys of mountain structures has made possible a revision of the boundary conditions of the formation of open taliks in the mountain river valleys. In particular, with respect to the Verkhoyanskiy (Upper Yana) massif, the following probability of the appearance of open taliks was discovered as a function of the water collecting area: 29 percent for less than 50 km^2, 30 percent for 50 to 100 km^2, 77 percent for 100-150 km^2 and 88 percent for 150-200 km^2.

However, the change in composition and worsening of the percolation properties of the alluvial deposits as the rivers exit onto the plain in many cases lead to them becoming more deeply frozen and to transition of the open taliks to closed, suprapermafrost taliks. These phenomena are even observed in such large rivers as the Yana and the Indigirka. They are indicative of an extremely close genetic interrelation and interdependence between the hydrological, hydrogeological, and permafrost indexes within unitary hydrogeological strucutres and combinations of them.

The studies of the processes involved in the deep freezing of minerals, the structure and development of the negative-temperature zone in the permafrost region, its discontinuous character, and the formation of taliks have made possible a greatly enlarged concept of hydrogeological structures.

The laws governing the distribution and formation of groundwater in various hydrogeological structures where there is deep cooling of the Earth's crust. The basis for the analysis of the processes occurring in this case is the relationship that exists between the thickness of the sedimentary mantle of the artesian basins or the zone of regional jointing of hydrogeological massifs and the thickness of the negative-temperature zone, having regard to the latter's structure, distribution, and history of development. In this context, let us begin by considering hydrogeological massifs--structures that are predominantly of the water-bearing type--and then artesian basins with a predominantly stratal type of groundwater, remembering at the same time that intermediate types of structures also exist with respect to their water-bearing characteristics.

It has been discovered that as the severity of the permafrost situation in the hydrogeological massifs intensifies, namely, a reduction in the degree of discontinuity and an increase in the thickness of the permafrost region, at first there is localization of the groundwater charge and discharge foci and the formation of a cryogenous head. Their further freezing, combined with the processes resulting in permafrost disintegration of the rocks under the influence of the fluctuation of the permafrost base during the course of its establishment and development, leads to the formation of a single subpermafrost water-bearing zone containing pressure water (the cryogenous basins of pressure fissure water) or, conversely, to differentiation of the unitary water-bearing system of the massif into many localized systems in individual or several adjacent river basins (hydrogeological massifs of superdeep freezing). The former occur under dissected plateau conditions if the depth of dissection correlates with the thickness of the permafrost region. The latter occur under conditions of heavily dissected relief in which the thickness of the permafrost region greatly exceeds the depth of downcutting of the river network. Finally, in the extreme case of deep freezing of massive crystalline rocks, the calculations indicate that hydrogeological structures form in which groundwater in the liquid phase can only exist during the summer. This is solely suprapermafrost water. The possibility of the existence of suprapermafrost closed taliks is very small. These hydrogeological massifs have come to be called cryogeological.

The deep freezing of artesian structures that have a simple hydrogeological section containing fresh water within the entire sedimentary mantle leads in many cases to complete freezing of the basin mantle and regeneration of their structure as artesian (cryological basins). With freezing of artesian basins with a complex hydrogeological section, the thickness of the permafrost region becomes greater than the thickness of the fresh water zone, whereupon a zone of cryopegs emerges at its base. These structures have been defined as cryoartesian.

The described conditions of freezing encompass all or part of the artesian basin, thereby reflecting a latitudinal zonation in the development of the permafrost zone. It is understandable that the conditions of formation of the groundwater of such structures are appreciably different from those of hydrogeological structures located outside the permafrost region. This difference is exhibited for the most part in the variation of the hydrodynamic conditions of formation of the groundwater.

Hydrodynamic systems under the conditions obtaining in deeply freezing structure. The study of hydrodynamic systems on the procedural and theoretical levels has been carried out separately for artesian basins and hydrogeological massifs. When investigating the latter, it was discovered that, in addition to the latitudinal zonation, in alpine regions the altitudinal permafrost-hydrogeological zonation is of decisive importance in the formation of a hydrodynamic system of hydrogeological massifs. Within the continuous permafrost zone the following four altitudinal belts have been distinguished: hydrothermal accumulation, percolation and inflow, transit and accumulation, and discharge.

Within the first watershed belt of hydrothermal accumulation, the amount of heat that has accumulated in the aqueous flows will be sufficient to initiate the formation of suprapermafrost and, subsequently, open taliks in the river valleys. Correspondingly, in the second belt, by virtue of the presence of open taliks and the belt's high position in the relief, the determining hydrogeological process is percolation and the inflow of surface water, subpermafrost water recharge, and the resulting formation of pressure heads of this water. In addition, this is the belt where the most pronounced seasonal variations in groundwater levels occur, causing the water-absorbing regimes of the taliks to alternate with water-discharging regimes. In the third belt, which is confined to the midsection of the slopes, the predominant factor is goundwater transit, although, along with the continuous increase in the flow rate downward and along the extent of the open taliks, the groundwater continues to accumulate. The lower discharge belt is confined to the foot of the mountain systems, and it is characterized either by the direct discharge of groundwater to the surface by way of foothills debris cones or the overflow of this water into the adjacent artesian basins. The predominance of one or other of these processes depends on the permafrost and hydrogeological conditions in the foothills part of the massif and the limb of the artesian basin. The discharge of the groundwater ordinarily occurs when the latter has no highly permeable horizons or aquiferous zones of fracture. It is precisely this zonation that characterizes many of the hydrogeological massifs of northeastern USSR, the alpine structures of the northern Transbaikal, and certain other regions.

A comparison of the above-described permafrost-hydrogeological zones with hydrodynamic zones indicates that the upper belt is an aeration zone; the second belongs to a zone of seasonal variations in the groundwater level and the formation of groundwater heads; the third corresponds to a saturation zone; and the fourth corresponds to a discharge zone.

Under the conditions obtaining in the discontinuous permafrost zone, the watersheds are frequently in the thawed state, whereupon the two upper belts merge into one, encompassing both the aeration zone and the zone of seasonal fluctuations in the groundwater levels. In complex mountain structures that have stepped relief, the altitudinal zonation may reoccur.

The altitudinal permafrost-hydrogeological zonation of the alpine structures in the southern regions of the permafrost zone, in particular the hydrogeological massifs of central Asia, displays a highly distinctive character. In the area of their water divide, depending on the presence, thickness, and state of the glaciers, a coincidence of the hydrothermal accumulation belt with the percolation and inflow belt can be expected. The transit belt extends beyond the limits of the permafrost region. The fact that the hydrogeology of this alpine region has hardly been studied at all means that these suggestions can only be advanced on the basis of indirect signs, in particular, on the presence of a naled belt--the characteristic sign of the transit and accumulation of the water contained in the hydrogeological massifs of the permafrost zone. In the alpine region of central Asia they occupy the uppermost part and are in direct contact with glaciers. In hydrogeological massifs expressed by plateaus and highlands, the altitudinal zonation is replaced by a latitudinal zonation.

The hydrodynamic system of artesian basins of the cryolithosphere, in addition to being determined by general factors, namely the relative positioning of the recharge and transit regions, the geodynamic stresses connected with the lithification of the rocks and certain other processes, is also determined by the distribution and character of the discontinuity and the direction of development of the permafrost region. In this respect, special attention has recently been devoted to the low piezometric levels of the upper subpermafrost aquifers. The debate over the causative factors in the formation of these low levels, which has been conducted in print during the past decade, is still far from being exhausted. The absolute values of the low levels to 333 m, which constitute a unique phenomenon in nature, are proving to be a regular consequence of an aggregate of tectonic, geochemical, and geothermal processes: namely, the variation of the hypsometric position of the bed in relation to the recharge or discharge region, the decrease in the volume of underground fluids occurring as a result of geochemical and lithochemical processes, and the increase in the pore-space volume as a result of degradation of the permafrost region. In a number of cases the latter is the primary cause. In particular, it has been discovered that the regional decrease in the thickness of the permafrost zone during the last 20,000 yr, as indicated by an entire set of observations, including geothermal and hydrochemical observations, is leading to a sharp reduction in the stratal pressures of the upper subpermafrost layers and to the establishment of depression peizometric surfaces extending for many hundreds of km^2. The stratal pressure of the deeper layers, which are separated from the upper layers by dependable water-impervious beds, does not reflect this effect. These conditions have been discovered in the vast area comprising the eastern Siberian and western Siberian artesian regions, in the Kolyma artesian basin, and others.

The effect of the permafrost region on the formation of the hydrodynamic system of artesian structures is not only confined to these. Thus, in the artesian basins of the Penzhino-Anadyr' artesian region the piezometric surface of the upper subpermafrost layers clearly reflects the regions of external recharge and the centers of internal recharge of aquiferous complexes caused both by the tectonic structure of the basins and the distribution of the taliks within the permafrost region.

The effect of the discontinuous or insular permafrost zone is less pronounced if we exclude the local cryogenous head occurring during partial freezing of the upper aquifers.

The study in this context of the hydrodynamic zonation of the artesian basins of the continuous permafrost region has revealed that it is not enough to distinguish three traditional water-exchange zones there: a free, a complex, and a highly complex. The presence within the permafrost region of open, closed, and completely closed taliks has led to the necessity for distinguishing within its boundaries a zone of complex water exchange where the groundwater contained in the taliks can be in a highly varied interrelation with the surface water and the subpermafrost water. It will be demonstrated below that this is also confirmed by hydrochemical data.

The cryogenic metamorphization of groundwater and the conditions giving rise to its chemical composition was the next most important area of hydrogeological research in the recent past. The study of intrapermafrost and subpermafrost water, together with the study of the thermal regime of the permafrost and the direction the permafrost process towards freezing or thawing, have made for a marked expansion of the concepts of cryogenous metamorphization of groundwater and also of the significance and the scale of this phenomenon, which is of regional importance to the subpermafrost layers. By cryogenous metamorphization we mean the change in the mineralization and composition of the groundwater under the influence of freezing or thawing. Accordingly, in the first case the general directivity of the cryogenous metamorphization is oriented toward an increase in the mineralization of the groundwater, and in the second case, toward a decrease of it.

In addition, the cooling of the groundwater and, in particular, its freezing and melting lead to redistribution of the ions as a result of precipitation of part of the compounds into an insoluble precipitate, primarily of calcium [salts], and to relative enrichment of the residual solution by more soluble sodium compounds. In the especially favorable situation combining stable low temperatures and the filling up with freezing groundwater, ultimately a tenfold or hundredfold concentration of the natural solutions takes place.

Thus, the leaching out of the sulfides and the cryogenous concentration observed in one of the gold-sulfide deposits in the Indigirka River basin led to the appearance of groundwater with a mineralization of more than 250 g/liter and a typical "mining" composition. Much more frequently, the mineralization of the water of freezing taliks falls within the range of tens of grams per liter, with its magnitude being a function of the openness of the talik and its emergent subterranean stream or body of water, the dimensions of the catchment area, and the ratio of the processes of evaporation and precipitation occurring within it, the content of water-soluble salts in the perenially frozen and the seasonally thawing deposits, and the water-salt regime of the body of water or stream.

The development of the theory of cryogenic metamorphization and exchange reactions in the water-rock system has led to the explation of the widespread distribution of saline water in subpermafrost aquifers, including those made up of carbonaceous rock, which are characterized outside the permafrost zone by calcic water. It has been discovered that it is precisely these processes that are connected with the origin of the thick zone of fresh or extremely weakly mineralized groundwater, which is characteristic of the continental deposits, and even of a zone of complex water exchange, for example, the Pechora and the Lena-Vilyuy artesian basins. The permafrost concentration processes also explain the abrupt rise in the mineralization at the base of the freshwater zone, which has been noted in a number of wells of the Yakut artesian basin. The hydrochemical nature of the section in the described cases agrees well with the presence in the rocks of secondary calcite and mirabilite inclusions, and in some cases, of halite, which are regarded as products of freezing. These processes also explain the formation of calcium chloride water in the cryopeg zone at depths where such water does not usually occur (the Yakut and the Tungussy artesian basins). Studies of cryopegs are receiving special attention on account of the possibility that they are exerting a cooling effect on the Earth's interior and because of the prospects for the development of gas extraction from gas hydrate deposits, which are very closely connected with the thermal regime of the gas-bearing beds.

Studies of the salt composition of naleds have come to be of great practical importance. In particular, it has been discovered that, as a result of cryogenous metamorphization when the water on a naled or on river ice freezes in a thin layer on the surface of the naled, the mineralization of the ice undergoing formation can vary by more than three orders, although in the bulk of the ice it is lower than in the water making up the ice. For example, the mineralization of the ice of the Ulakhan-Taryn spring varied from 13 mg/liter to 5 g/liter (in small sections at the end of the freezing currents of water), the mineralization of the spring water being of the order of 200 mg/liter. The studies also demonstrated both the possibility and the presence of ions migrating directly in a body of the ice oriented toward the surface of the naled in the winter and towards its base in the spring and summer.

The salt formation associated with the icing process, the transition to insoluble precipitate of part of the salts during freezing of the water and their removal by melt water in the form of solid river runoff, leads to regional freshening of the surface water of naled regions and, in consequence, of the groundwater with which they are fed. The scale of these phenomena is significant both in quantitative and regional terms. Calculations have shown that the naleds of Yakutia can remove calcium carbonate into precipitate at the rate of 2-4 $\times 10^3$ tons/yr. According to other data the naleds of southern Verkhoyan'ye alone precipitate salt in amounts of up to 15 $\times 10^3$ tons/hr. A part of these calcium salts remain on the naled sites, making up the building material for the microorganisms developing there, and a part of them are removed in the form of solid river runoff.

Natural groundwater resources. An analysis of a map of the natural groundwater resources in the area of the USSR encompassed by the negative-temperature belt indicates the distinct zonal and regional patterns on their propagation, the connection with climatic factors, and the thickness and type of discontinuity of the permafrost zone. The smallest values, amounting to tenths and hundredths of a liter per second per square kilometer, are attained by the groundwater runoff in the central regions of the Yakut artesian basin, precisely where the maximum thickness of the permafrost region is noted, as well as its weak discontinuity and minimum quantity of precipitates. This is the territory of the Lena-Vilyuy artesian basin, which is characterized by extremely low piezometric values of the subpermafrost water levels. Similar permafrost-hydrogeographic conditions of the Kolyma artesian basin and the artesian basins of the maritime zone are also exhibited in the low moduli of the groundwater runoff (less than 0.5 liters/s-km^2). At the western edge of the eastern Siberian artesian basin and in the central part of the western Siberian artesian basin the groundwater runoff increases to 3 liters/s-km^2, forming a positive anomaly in the vicinity and along the periphery of the Putoran Plateau, and again it drops to 0.5 liters/s-km^2 and less in the direction of the north coast. To the south and southwest of the region of maximum groundwater runoff of the Lena-Vilyuy basin, within the Aldan hydrogeological massif and the Patomo-Vitim hydrogeological folded region, the groundwater runoff again increases to 3 and 5 liters/s-km^2 respectively, which is connected with the increase there in the amount of precipitation and the discontinuity of the permafrost zone. It increases to a lesser degree in the northeastern part of the USSR, where the isolines of the groundwater runoff moduli clearly emphasize its connection with the relief, increasing along the periphery of the mountain structures to 2 liters/s-km^2. In the Koryak hydrogeological massif, the positioning of the isolines of the runoff modulus is determined by its adjacency to the Bering Sea--the main source of atmospheric precipitation in this area.

A more detailed analysis of the data on the runoff of individual river basins and the results of groundwater surveys in the Northeast and in the Noril'sk region have demonstrated the pronounced lack of uniformity in its distribution between the individual basins and have confirmed that, other conditions being equal, the decisive factor in the accumulation of the natural groundwater reserves in the mountainous regions is the great thicknesses of the boulder-shingle and sand deposits in the overdeepened sections of contemporaneous and ancient river valleys, foothills debris cones, intermontane depressions, and block tectonic zones of subsidence. It is precisely these deposits that form the intermediate reservoirs that receive the surface water and the groundwater of the taliks beneath riverbeds, and also the subpermafrost water rising through the fractured zones. The large quantities of incoming water and high velocities of its movement prevent the freezing of these reservoirs and ensure that the rocks remain in a thawed state throughout the year. It is via these reservoirs that the transit of the groundwater is effected, as well as its redistribution. Moreover, it is due to the variation in the facies composition of the deposits in the direction of a reduction in the filtration properties downstream along the river valleys, or because of the decrease in their thickness and transverse section that the groundwater discharges onto the surface. High-yield springs sometimes appear, forming very large polyn'uas or naled lines. Such conditions are especially characteristic of foothills boulder trains. Thus, the naled lines framing the Momo-Selennyakhskya depression from the south extend for a distance of more than 250 km, and the yield of the springs associated with it is measured in hundreds and thousands of liters per second.

Interesting results of the exploration of the talik water of the loose Quaternary deposits were obtained in the Noril'sk region. Beyond the Arctic Circle, in the extremely severe permafrost and climatic setting, an exceptionally promising basin has been discovered, the groundwater of which is becoming the main source of the water required by the plants of the Combine. The water-bearing taliks here also coincide with tectonic zones of increased jointing, in which the voids exceed 6 percent.

The localization of the thick accumulations of these deposits primarily within the river valleys, and the regular variations in thickness and composition of these deposits as indicated by benchmarks are directly connected with the above-investigated permafrost and hydrogeological zonation. However, they also give rise to extreme nonuniformity in the flooding of the hydrogeological massif if the thickness of its permafrost zone exceeds the thickness of the zone of regional jointing. Actually, under conditions of a highly dissected relief and a thick permafrost zone, the watershed areas prove to be deeply frozen and drained and the groundwater is concentrated directly in the taliks of the river valleys. In sections where the taliks are closed it is concentrated in the jointed zones. It is as though the latter succeed the river valleys in plan view. In turn, the valley taliks join the taliks of the piedmont debris cones or intermontane depressions, forming linear water-bearing systems that are separated by very slightly water-bearing or nonaquiferous blocks. The hydrogeological massifs made up of karstic rock or highly jointed effusions and also vulcanogenic superbasins constitute an exception. In these structures highly flooded zones also form in cracks of the bedrock. Thus, karstic manifestations of groundwater were observed in the Selennyakhskiy massif, but they are most pronounced on the Aimo-Uchurskiy watershed where the karst forms directly at the surface.

The high percolating properties of the karstic rocks and the abundant precipitation give rise to favorable conditions for the formation of watershed rain-radiation taliks, the high degree of flooding of the Aimo-Uchurskiy karstic plateau, and numerous naleds and springs, the most important of which--the spring in the Selinde river valley--has a discharge of about 4 m^3/s during the critical winter period, and is the largest karstic spring in the area comprising the permafrost region.

Tectonic fractures also play an important part in the formation of the groundwater resources. Their role, however, is different. On the one hand, it is the zones of increased fracturing that are associated with faults, and, on the other, thick accumulations of unconsolidated deposits or conversely, a sharp reduction in their thickness. Therefore, an unambiguous solution to the problem of the role of the fracture in each individual case is not always possible without conducting special studies.

In artesian basins, especially of the platform type, the situation with respect to the fresh groundwater resources is greatly complicated in cases where cryopegs occur at the base of the permafrost region. It then becomes necessary to direct one's attention exclusively to the water of the taliks under riverbeds or lakes. However, the alluvial deposits of lowland rivers are appreciably less water-bearing, and the quality of the water included in the taliks under riverbeds and, in particular, under lakes by no means always conforms to the requirements imposed on drinking water. In addition, the variegated composition of the water complicates the choice of the talik and leads to an increase in the amount of exploration work. But, these complications are gradually being overcome, and more reliable criteria are being developed for predicting the occurrence of talik water suitable for water supply.

With regard to the natural groundwater resources of the permafrost region, it is necessary to note the sharp seasonal variations in their values, which are greatly in excess of such variations outside the permafrost region. They are manifested most clearly in the taliks of river valleys of mountainous regions where the level of the groundwater during the winter falls by tens of meters, and its resources are used up. The reasons for this, in addition to the steep slopes of the valleys and the natural runoff of water not being recharged by the atmosphere and the ground during the winter, are the naled control processes that, even under the most adverse conditions, are capable of accumulating in the winter almost all of the water in the taliks under the riverbeds. These unfavorable consequences can be lessened to some extent. In particular, experiments with the passage of the surface (river) water into the subpermafrost taliks in the northeastern part of the USSR have demonstrated that here thawing out of the frozen ground takes place in a downward direction, there being an increase in the volume of the subpermafrost reservoirs and an improvement in the quality of the ground water, along with an increase in reserves.

In recent years there has been an expansion of studies concerned with estimating the mineral water reserves, including the therapeutic, thermal, and industrial reserves. These studies have proved to be especially productive in the extreme northeast of the USSR, in the Transbaikal, and in the southern part of eastern Siberia.

The research in the aforementioned areas has posed a number of problems and questions that must be solved by hydrogeologists. The working out of the designs of water intakes, water mains and various tupes of drains and sewers not subject to freezing, and also of measures to combat icing phenomena and their consequences and to achieve control over icing processes and effect thermal and water improvments and, conversely, to combat mine and shaft water is only a fraction of the practical difficulties that must be overcome if there is to be a successful solution of the basic problem with respect to exploiting or controlling the groundwater. The solution of these problems, however, along with theoretical questions in hydrogeology, has posed new questions and problems, which we shall now proceed to discuss.

Future research. The most important task in future theoretical research will be to clear up the problem of subpermafrost water recharge and discharge, in other words, the whole intricate and diversified set of questions pertaining to the interrelation of atmospheric, surficial, suprapermafrost, subpermafrost, and intrapermafrost water, considered in the context of their interactions with the frozen ground. This main subject area can be divided into several special areas:

1 Defining more precisely the actual nature of the hydrologic cycle involving the upper subpermafrost layers and the suprapermafrost and intrapermafrost water. The elaboration of this area, in turn, must lead to the working out of the boundary conditions for calculating the filling up of the subpermafrost horizons when using their groundwater and regulating the subpermafrost water reserves by storing it. The latter question is deserving of special attention, both by reason of the enormous depletion of subpermafrost water in winter, as already mentioned, and because of the absence of reliable sources of water in the vast areas occupied by the artesian basins in which the cryopeg zone is situated. The initial experiments on storage when there is an abundance of fissure water have yielded promising results but require further elaboration. Given the existence of the low peizometric levels that typify artesian basins, water-injection wells could operate for a long time without the use of pumps. This could create lenses of fresh subpermafrost water floating on saltwater and, during the short summer, form reserves of fresh water to be used in the ensuing winter. Water-supply sustems of this type are used in arid and semiarid regions to obtain groundwater. They could also be set up to obtain artesian water. However, storage of groundwater under such conditions is complicated by the significant 10°-20° temperature difference between the pumped water and the environment, as a result of which there are reasons to expect abundant precipitation of calcium carbonate from the water and the deposition of a layer of silt along the periphery of the water-collecting wells and the necessity in a number of cases for the pumping of fresh water into an environment with negative temperatures and filled, moreover, with saltwater and brines. Thus, besides the usual hydrogeological problem, it is also necessary to solve the hydrochemical and thermophysical problem of the behavior of the saltwater-freshwater interface and to work out temperature and hydrodynamic operating criteria of the system, which will unconditionally ensure that it is maintained in the melted state. The hydrogeologists have only just embarked upon the task of solving this problem. Nevertheless, storage of groundwater constitutes a change-over from the passive use of the "gifts of nature" to the active control of natural pressure head systems and, in this context, merits assiduous attention.

2 Ascertaining the orientation and rates of movement of subpermafrost water in the upper aquiferous layers of the large artesian basins and the conditions under which this water is discharged at low piezometric levels, and even the very nature of the latter.

3 Identifying the regions of discharge of the artesian water. In a number of cases the regions in which the groundwater of the artesian structures in the permafrost region is discharged remain unclear. The enormous areas with relatively low and negative piezometric levels obviously indicate the absence of groundwater discharge over vast expanses and call for precise definition and mapping of the contours of the latter. Recently, new data have been obtained with respect to the internal recharge and discharge regions of the artesian basins of the Northeast. Such studies of the centers of groundwater intake and discharge appear promising, as the usual mechanism involving overflow of groundwater from hydrogeological massifs into adjacent artesian basins when the conditions are such that the latter are frozen to great depths, cannot play the all-encompassing role that is observed in structures outside the permafrost zone. However, it is impossible to resolve the extremely complex problems of groundwater feed and the formation of artesian basin hydrostatic pressure heads in the permafrost region by on site observations alone. These studies must be supplemented by theoretical calculations and simulation with subsequent approximation of the model to the actual structures and hydrodynamic conditions existing in the artesian basins.

It is on these that depend the correctness of choice of the conditions under which the aquiferous layers will be utilized; the designs of the intake systems that will be called upon to store the groundwater; the planning of various types of underground storage facilities for petroleum products and natural gas, such as are widely used abroad; and, finally, the disposal of industrial waste products.

4 The study in hydrogeological massifs, of water-bearing complexes of coarse sandy and gravelly pebbly and pebbly rubbly alluvial, proluvial (foothills debris cones), and fluvioglacial

deposits, the so-called intermediate reservoir rocks accumulating the greater part of the groundwater resources of these structures and of the water-bearing zone of effective jointing underlying such deposits as before, this study is deserving of special attention.

5 The further study of the conditions of recharge of hydrogeological massifs and similar structures with a predominantly jointed type of aquifer (the vulcanogenic superbasins, and so on). Primarily this calls for the working out of hydrothermal problems pertaining to the formation of open taliks in the recharge zone, that is, in the watershed expanses. It is they that must reveal in its entirety the role of condensation and of congelation ice formation in rocky streams, and also the significance of flat runoff and of linear riverbed zones of hydrothermal accumulation. One of the end results of the studies must be clarification of the minimum areas of the runoff bases that are necessary and sufficient for the formation of open taliks ensuring percolation and inflow of surface water, while, in the case of recently glaciated areas (the Arctic Islands and alpine structures), the working out of problems pertaining to the connection between subglacial and subpermafrost water and the potential utilization of these waters is of interest.

6 The study of questions pertaining to the recharge and discharge of groundwater in hydrogeological structures that are subject to deep freezing can only be effected if it is predicated on *the study of groundwater regimes*--one of the least-developed areas in studies of the groundwater of the cryolithosphere. This applies especially to the study of subpermafrost water, the investigation of the regimes of which was until recently limited to the freezing over of test wells. The successful surmounting of the freezing of deep wells drilled in frozen ground has opened wide vistas for carrying out routine observations of subpermafrost water and has presented an opportunity for setting up stations for the systematic study of the deep aquifers in the permafrost zone. The initial results of observations in central Yakutia and also in the Vorkuta region of the Pechora coal basin have revealed a completely unexpected close dependence of the heads of subpermafrost water and their levels on the surface water regime. This significantly alters the prevailing concept of the conditions that govern the interrelation between subpermafrost and surface water in platform artesian structures. However, the results of these studies, conducted on the southern limb of the Yakut artesian basin in a region of recharge, where the thickness of the permafrost ranges from 200 to 300 m, cannot be extended to its central regions, to say nothing of artesian structures such as the western Siberian, Kolyma, or Penzhino-Anadyr' artesian basins. Deserving of special attention are studies of the groundwater regime in the coastal regions, the thermal and mineral springs and the water of the intermediate reservoir rocks.

7 The theoretical basis for studying the groundwater regime calls for a *more precise understanding of the aeration zone* and of its role in shaping the composition and regimen of the groundwater in the upper layers. The present tendency to define the aeration zone as part of the layer of seasonal thawing is not in accordance with the numerous hydrogeological opinions partly discussed above. Actually, within a thin permafrost zone, even if it is continuous in distribution, from time to time the presence of a conduction band of subpermafrost water is manifested, the variations of which are accompanied by, or cause variations in, the permafrost base. In the vast areas constituting the karstic plateaus, the deep-seated occurrence of the free level of subpermafrost water is often observed, the recharge of which is accomplished over the area containing the karstic water by way of numerous local taliks. These taliks are temporarily water-bearing only during periods of abundant precipitation or of snow melting. In the river valley taliks--the main pathways in the filling up with subpermafrost water--with the lowering of the level of the intrapermafrost water of the taliks that occurs during the winter, a dehydrated zone originates that, on the genetic plane, corresponds to the notion of the aeration zone. Within this zone it is evidendly proper to include the thick series of sands comprising the terrace deposits, the vapor content of which corresponds to their air-dry state. These examples alone indicate how complicated is the problem of defining, it would seem, the established notion of an aeration zone as applied to an area of deep freezing of the Earth's interior. It is necessary to study the peculiarities of the aeration zone of the subterranean cryosphere, its structure and pathways, and also the methods of defining and evaulating it in the hydrogeological context.

8 The studies of the groundwater regime of the permafrost region are inseparable *from the study of the overall water budget*. The budgetary investigations are presently confined to very small drainage areas and are characterized by an incomplete series of observations--they do not encompass groundwater studies--of the water beneath riverbeds and that which is even deeper. In addition, these investigations are closely linked with the study of natural resources and the reserves of groundwater that are available for use.

9 The comprehensive exploitation of water resources would be unthinkable without making a *further study of the chemical composition of the groundwater and its peculiarities, which are caused by cryogenesis.* Of interest in this connection are regional and theoretical studies aimed at ascertaining the position and significance of the permafrost region in the shaping of the overall hydrochemical zonation. There have been superficial studies of questions pertaining to the geochemistry of the solid phase of groundwater, namely, ground ice of differing age and origin. The water of closed intrapermafrost lenses and saltwater and brines with negative temperatures (cryopegs) have evoked particularly strong interest. The study of these groundwater lenses, the nature of many of which is at present not open to interpretation, would facilitate the

identification of new patterns of development of the permafrost region with time and determination of the effect of this region on the groundwater. The observed relationship between some intrapermafrost lenses containing highly mineralized water and brines containing gold-sulfide deposits, and also the specific chemical composition of the groundwater of such lenses, indicate the unusual results of the simultaneous effect of the sulfide minerals' oxidation mechanism and the permafrost concentration. In this respect the studies of the composition of the groundwater have importance in hydrochemical prospecting, and this has been proved in practice. However, on the theoretical level the problems of the migration of ore mineralization components when the conditions are such that there is multiple freezing-and-melting have not been resolved at all and there are currently no data for use in the interpretation of hydrochemical anomalies.

The studies of the chemical composition of groundwater are closely intertwined with the study of the composition of the wedge ice, naleds, and salts formed on their surfaces. The studies and preliminary calculations have demonstrated the characteristic distilling role of naleds. More precise definition of the scale of these phenomena and their potential role in the formation of the carbonate content of coastal-marine deposits could lead to entirely unexpected results.

One of the applied but very important questions pertaining to the exploitation of groundwater is the development of *economically efficient and productive methods of improving the quality of groundwater,* which, either because of its overall mineralization (1-3 g/liter) or because of the content of individual components, cannot be used for drinking unless it is specially treated. These studies are inseparably connected with *the problem of the conservation of the natural water in the permafrost region.* The available facts with respect to the interrelation between subpermafrost and suprapermafrost water indicates that the so-called "forzen water-confining stratum" gives no guarantee that the purity of subpermafrost water will be preserved with discharge of industrial wastes into river valleys. Furthermore, it could be disturbed as a result of the warming effect of the sedimentation tanks and the biochemical and other reactions occurring in them. All of these problems relating to the conservation of natural water have as yet been untouched by research.

10 A special area of research is thermal water and the prospects for using it. In recent years a great deal has been done with respect to the study of thermal mineral-water pools. A tangible result of this research is the modern comfortable health resort at the Tal'skiye springs. However, hydrothermal research is now faced with new problems, the solution of which must improve the conditions attending the development of frozen placers. These problems consist in the development of efficient systems whereby geothermal energy can be used to thaw the placers. The exploration of natural hydrothermal systems and the creation of artificial hydrothermal systems are necessary for their solution.

Among the multiplicity of problems relating to the formation of thermal springs, it is possible to distinguish the main ones. The first is the working out of the Earth's hydrothermal zonation and the classification of groundwater according to temperature, ranging from the supercooled water of the permafrost region to the hot water occurring at great depths. It is against the background of the hydrothermal zones of the Earth's crust that exhibit a regular alternation with depth that hydrothermal anomalies are distinguished. The second is the study of the nature of anomalously high and anomalously low heat fluxes and, in particular, formulation of the modes of heat transmission in the cryopeg zone. The third is the water balance of the hydrothermal systems and the relation between vadose water and juvenile water. Recently, many researchers have supported the view that most of the thermal water is of vadose origin, and it is on this premise that schemes have been devised for the establishment of artificial hydrothermal systems. The fourth is the structure of the hydrothermal systems, in particular, the nature and distribution of the catchment basins, the conditions attending the formation of the pressure heads, and the recharge and discharge of thermal springs. However, the study of the drainage basins of hydrothermal systems would be unthinkable without making a study of the problems of recent mineral formation within the confines of these systems, both in the natural state and as a result of the lowering of the pressure that occurs during tapping by water- and steam-emitting wells. In this respect much has been done in the course of exploring the Pauzhetskoye water- and steam-emitting thermal springs in Kamchatka and the Goryachiy beach in the Kurile Islands. The problem consists in the fact that data on mineral neoformations and the secondary alteration of the rocks must be more closely coordinated with studies of the structure of the catchment basins and their properties. These studies will be of great importance in predicting the performance of the artificial hydrothermal systems.

PHYSICS AND MECHANICS OF FROZEN EARTH MATERIALS AND ICE

N. A. TSYTOVICH Moscow Construction Engineering Institute

Recently there has been a marked expansion in the USSR, in research in the cryology of the Earth, and especially in geocryology--the study of permafrost. Research projects in geocryology are included in the plan of 48 Soviet scientific institutions, in which 124 individual projects are being investigated, of which about half are in the physics and mechanics of frozen earth materials.

The leading research institutions are the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences (Yakutsk), the All-Union Research Institute of Hydrogeology and Engineering Geology (VSEGINGEO, Moscow), the Foundations and Underground Structures Research Institute (NIIOSP, Moscow), the Industrial Research Institute for Engineering Surveys in Construction (PNIIS, Moscow), the Northeastern Integrated Research Institute (SVKNII, Magadan) and others. Among the higher educational institutions are Moscow State University, the Moscow Construction Engineering Institute (MISI), and others.

Studies connected with construction of the following types of installations under permafrost conditions have been conducted: road, industrial, and civil construction; heat and gas supply systems (including the laying of main gas pipelines); power engineering and hydraulic engineering construction (thermoelectric and hydroelectric power stations); and others. It would be impossible to present a detailed discussion of all of the studies, even with respect to this particular problem (the physics and mechanics of frozen earth materials). In addition, not long ago several interesting regional conferences on geocryology, the findings of which were published at the end of 1972, were held in advance of the Second International Conference on Permafrost. Also, research in the physics and mechanics of earth materials and ice performed in the United States and Canada will be discussed elsewhere in these volumes.

Not being in possession of all of the necessary data on research in the physics and mechanics of frozen earth materials and ice conducted during the last few years, the author has had to limit himself to the elucidation of only a few of the questions, although in his opinion they are leading ones that must be further elaborated during future research.

THE MAIN TRENDS IN THE ELABORATION OF THE PHYSICS AND MECHANICS OF FROZEN EARTH MATERIALS AND ICE AND SOME RESULTS

The main trends in the most recent research in this field are as follows:

1. Elaboration of the "basic principles" of the physics and mechanics of frozen ground, which were previously formulated by the author as general laws ascertained by a purely experimental method predicated on an analysis of the results of direct observations and measurements and determining for the most part the physical and mechanical properties of frozen earth materials and ice, as well as the physical and mechanical cryogenous processes that take place in them (with any variation in the thermodynamic conditions, e.g., temperature and pressure).[4]
2. The following detailed studies of the mechanism of cryogenous processes, on a new and higher (molecular-structural) level:

   (a) On the basis of the physical chemistry of surface phenomena occurring during interaction of the individual phases in the formation of frozen earth materials (solid, fluid, gaseous), utilizing the laws of absorption and ion exchange, the dynamics of irreversible processes and the like.

   (b) On the basis of crystal and structural analysis of frozen earth materials, considering their structure and rearrangement as an indicator of the stress-strain state of frozen earth materials and ice, and taking the activation energy of the molecules to be the driving force when making the transition from phenomenological functions to analytical formulations of the physical relations controlling the mechanical processes in the frozen earth materials and ice.

With respect to the first area, these principles, that is, the general laws determining the peculiarities of cryogenic properties and processes in earth materials and ice, can be formulated as the following fundamental tenets:

1. The amount, concentration, and properties of the nonfrozen water and ice that are always present in frozen earth materials (especially in dispersed materials) vary with a variation of the external influences, being in dynamic equilibrium with the latter (N. A. Tsytovich's well-known principle).
2. The migration of the water in freezing earth materials is a molecular process of moisture transfer that originates when there is any disturbance in phase equilibrium as a result of a variation in the external influences.
3. The fluidity (relaxation and creep) of the frozen earth materials and ice, as a process of rearrangement of their structure is a constant factor determining the behavior of frozen earth materials and ice under the protracted influence of a load, which must always be taken into consideration when studying the origin, formation,

and course of development of the mechanical processes in them.

4. The instability of the physical and mechanical properties of frozen earth materials and ice under the influence of varying thermodynamic factors and time shows up essentially in the behavior of the frozen earth materials and ice, both because of natural external influences and when various types of structures are being erected on the permafrost.

5. The compactibility of high-temperature frozen earth materials, as a result of their migration-viscous consolidation when laden, causes (even when they are maintained at a negative temperature) appreciable settlement of the foundations of the structures.

6. The structural instability observed during thawing of permafrost causes a cumulative variation in its porosity, and, in the case of ice-rich permafrost, appreciable settlement and subsidence, both in terms of magnitude and lack of uniformity.

7. The significant nonlinearity of the variations in the porosity and water permeability of thawing ground, which is caused by its degree of fracturing during freezing, is the factor determining the appreciable settlement of weak, thawing earth materials under load and the rate at which this process takes place with time.

These seven basic principles of the physics and mechanics of frozen earth materials and ice can provide a foundation for the study of cryogenous variations in the physical and mechanical properties of earth materials and ice and for the study of the cryogenous mechanical process originating in these substances, thus furnishing a basis for building up a knowledge of the physics and mechanics of frozen ground.[5]

At the present time a number of the basic principles of the physics and mechanics of frozen earth materials (unconsolidated sediments) have been analytically formulated and are being used successfully, not only in the scientific sense, but also for practical purposes connected with the planning and erection of structures on permafrost.

With respect to the second area let us note only the promising proposals contained in papers presented by Soviet scientists at the Second International Conference on Permafrost. Interesting proposals were advanced in papers by S. S. Vyalov (Research Institute of Foundations, USSR Gosstroy), S. Ye. Grechishchev (VSEGINGEO), K. F. Voytkovskiy and V. N. Golubev (Problems Laboratory of Moscow State University).

In his paper entitled "Long-Term Destruction of Soil as a Thermally Activated Process," Vyalov discusses the fracture of frozen earth materials as a process of nucleation and development of microcracks and other defects. The quantitative characteristic of these processes is the "defect density" and the degree of orientation of the particles in the direction of shear, the variations in structure during displacement of the particles depending on the activation energy imparted to the particles.[2] On the basis of the theory of thermal activation contained in the paper by Zaretskiy and Vyalov,[2] an equation was derived for the creep rate, including the rate corresponding to the initiation of the steady-state flow, and also a theoretical formula for the long-term strength of earth materials. It is important to note that the comparisons of the analytical calculations with the experimental data adduced by the authors give good results both for frozen and nonfrozen earth materials.

S. S. Vyalov[1] proceeded from the equation proposed earlier by him in determining the strength of frozen soils:

$$\tau = \frac{\beta}{\ln \frac{t + t^*}{B}},$$

where β, B, t^* are the parameters of the long-term strength equation and t is the time to rupture.

On the basis of the theory of thermal activation, the theoretical values of the parameters β and B are presented as functions of the Boltzmann constant, the Planck constant, the characteristic curves of the particle concentration and the defect density, and the value of the constant t^* is taken to be equal to unity.

Vyalov discusses the question of calculating the temperature variation from a periodic law corresponding to the actual variation in the permafrost temperature and makes a comparison between the calculated value of the stress-rupture strength of the frozen soils at a variable temperature and values of strength corresponding to variations in the depth and also in the mean annual temperature of the permafrost.

The aforementioned results, discussed by Vyalov, can be of great practical importance from the standpoint of making a correct choice of the design resistances of frozen ground when the foundations of installations are laid in it. The constants required for the calculations necessitate that these be determined for a number of individual types of soil.

New results of the studies are presented in an interesting papaer by S. Ye Grechishchev, entitled "Basic Laws of Thermorheology and Thermal Cracking of Ground." It is shown that on the basis of the experimental data the temperature variations of frozen ground (to -30°) can be described by a nonlinear dependence of the hereditary type, and the method of determining the parameters of this dependence is outlined. The experiments (S. Ye. Grechishchev and I. N. Votyakov--described in papers presented at this conference) also revealed anomalously large coefficients of linear expansion of frozen earth materials; it is these that are responsible for the significant magnitude of the thermal deformations and the thermal aftereffect phenomenon. Grechishchev showed that the thermal deformations are most significantly affected by secondary short-term temperature fluctuations. We would point out that the results of the studies conducted can be of great significance in analyzing the stability of embankments, earth dams, underground reservoirs, airstrips and roads, and also in the laying of underground communications cables and other underground structures.

It is important to note that currently the majority of the studies on the mechanics of frozen soils and ice conducted when analyzing the deformations of the frozen ground with time begin with the theory of hereditary creep (the papers by S. S. Vyalov, S. Ye. Grechishchev, and also V. A. Kudryavtsev *et al.*) and that on the basis of the crystal and structural analysis and defect theory the process of the flow of frozen soils and ice is regarded as a thermally activated process, the indicator of which is the variation in the structure of the frozen earth materials and ice. Thus, in a paper by K. F. Voytkovskiy and V. N. Golubev entitled "Dependence of the Mechanical Properties of the Ice on the Conditions Attending its Formation," three stages of recrystallization of the ice are distinguished in the creep process: growth of the crystals, relative stabilization of the structure, and crushing of the crystals. Here, each stage determines the nature of deformability of the ice and has a significant effect on the deformation mechanism.

Some of the papers and reports pertain to the investigation of certain regional types of frozen soils: saline (N. A. Tsytovich, Ya. A. Kronik, K. F. Markin *et al.*) and coarsely-clastic (V. N. Taybashev, N. A. Tsytovich, and Ya. A. Kronik), which is of practical significance in evaluating the suitability of permafrost as foundations and as the medium in which various types of structures are erected under the conditions being considered.

In conclusion, I should like to note that the discussion at this conference of the problems of the physics and mechanics of frozen soils and ice will be conducive both to the general advancement of the science and also to the practical application of its achievements in the economic development of regions where permafrost is present.

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PHYSICAL AND CHEMICAL PRINCIPLES GOVERNING THE STRUCTURE, COMPOSITION AND PROPERTIES OF FROZEN GROUND

B. A. SAVEL'YEV Moscow State University

Frozen ground is a multiple-component, multiple-phase system existing at negative temperatures, all of the constituents of which are held together by structural cohesion. A peculiarity of frozen ground is that it contains water in three states: gaseous, liquid, and solid. The ratio between the various states of the water is determined by the complex interaction of the molecules making up the water with each other, with the molecules of the mineral skeleton, and also with the ions, molecules, and particles of materials in the liquid phase. It therefore depends on the magnitude of the negative temperature, the form and size of the active surface of the mineral skeleton, and the types and concentrations of the materials dissolved in the liquid water. The physical and chemical processes occurring in rocks, such as cation exchange and migration, lead to a change in this ratio. The properties of frozen ground (electrical, thermal, mechanical, and so on) are to a large extent determined by the relation between the water in the various states and, above all, by the amount of nonfreezing water consituting a liquidlike anisotropic interlayer between the solid components. The thickness of this interlayer may vary as a function of the above-mentioned conditions: It can range from a monomolecular water-layer held firmly to the surface of a foreign body by the enormous forces of cohesion to a layer that does not differ in its properties from free water and does not strengthen the rock, but weakens it.

Consequently, the properties of frozen ground are to a large extent determined by the physical and chemical processes occurring in it.

During the last decade, geocryology has been notable for a pronounced intensification of the physical-chemical approach to the study of permafrost.[3,7,13,15,16] Many of the theoretical, experimental and generalizing articles in the physics and chemistry of freezing, frozen, and thawing rocks were contained in 30 collections entitled *Merzlotnyye Issledovaniya* (*Permafrost Research*), which were published by the Permafrost Department of Moscow University, and also in the transactions of the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences. The two collections of articles, published by Moscow University in 1970 and 1972, are of undoubted interest.

The fact that the physics and chemistry of frozen ground have in recent years attracted the attention of geocryologists is apparent from the *Trudy VIII Vsesoyuznogo Soveshchaniya po Merzlotovedeniyu (Transactions of the Eighth All-Union Conference on Permafrost)* (Yakutsk, 1966) and *Tezisy Dokladov Vsesoyuznogo Soveshchaniya po Merzlotovedeniyu (Papers Presented at the All-Union Conference on Permafrost)* (1970).

Progress in the science of rheology, including the rheology of frozen ground, is currently being achieved by taking into consideration the thermodynamic principles of a solid. Simultaneously, a study is being made of aspects of the kinetics of disintegration during creep. The thermodynamic and kinetic factors in the disintegration or strengthening of frozen ground must be estimated by taking into account physical and chemical compositions, structure and processes of energy, and mass exchange. The heterogeneous processes occurring at the discontinuity surfaces between the phases and components in the frozen ground and the three-dimensional processes reflected in the structure, composition, and all of the properties of frozen earth materials were revealed by modern physical methods.

SOME OF THE PHYSICAL METHODS USED IN GEOCRYOLOGY

The intricate structural complex of frozen ground is being investigated through the use of crystallographic, X-ray, and spectrum methods of studying minute local peculiarities of structure and the ultrasonic method of measuring cyrstalline structural parameters. In permafrost studies wide use has been made of the electron microscope, by means of which the mineralogical composition of a rock is made manifest.

Along with the customary calorimetric method, the physical composition of H_2O component in its three phases is being investigated by means of nuclear-magnetic resonance and ultrasonic X-raying, and by the dissolving of ice in alcohol or other solvents. The chemical composition of rocks with a sufficiently high concentration of dissolved substances has approved chemical-analytical methods of analysis.

In the case of rocks with a low content of dissolved materials, which are undoubtedly reflected in the shaping of the structures of rocks and the textures of the ice inclusions, latterly atomic-absorption analysis (spectrum analysis) has been used. As indicated by studies of the texture of natural ice bodies conducted under the author's supervision when investigating the structure of ice by the ultrasonic method, a possibility arises of determining the summated values of the structural parameters of the crystals, the phase composition of the water, and the chemical composition of the dissolved ions and their phase transformations at eutectic temperatures. The use of the isotopic-analysis method makes it possible to draw conclusions as to the genesis of the rocks and to study the migration of material.

A very important characteristic of a frozen rock is its specific and active surfaces, for the evaluation of which adsorption methods are used (physical adsorption, desorption, and chemisorption).

For describing the processes of compaction and porosity, the gamma-logging and gamma-neutron-logging methods of observing natural objects are used. The use of the method whereby measurements are obtained of the electric potentials between the frozen and thawed part of the rock makes it possible to reconstruct the migration and heave processes.[9]

The studies by I. B. Savel'yev[11] demonstrated that the use of nucelar magnetic resonance for determining the content of nonfrozen water in frozen rock makes it possible to obtain more reliable and exact results than is the case with measurements by the calorimetric method. Even in sandy specimens this method makes it possible to detect slight variations in the nonfreezing water as the temperature drops. The nuclear magnetic resonance [NMR] method permits measurement of the phase composition of the water in any of the negative temperature ranges of a frozen rock, whereas the calorimetric method is applicable to temperatures not lower than -40°C. In addition to determining the phase composition, the NMR method affords an opportunity of estimating the energetics of the various layers of nonfreezing water, which correlates well with the spin-spin relaxation time of the protons.

Ye. V. Volkova[18] used the NMR method to ascertain the difference in the pattern of redistribution of the moisture in soils, relating the migration of water with differing degrees of adsorption to the mineralogical composition.

In order to study the pattern of distribution of the trace elements in the small layers of ice forming in the rock, I. B. Savel'yev[11] was the first to use the atomic-absorption spectrophotometer built by the Hitachi Company (Japan) in permafrost studies. This spectrophotometer has a high degree of selectivity and sensitivity.

Yu. Ye. Slesarenko[19] demonstrated the possibility of using the ultrasonic method for determining variations in phase composition in saline ice, correlating this process with the velocity of propagation of longitudinal and Rayleigh elastic waves, with the periods of the signals corresponding to these waves and ultimately with the modulus of elasticity. Slesarenko found that the irregular variations in the above-mentioned parameters coinciding with the various negative temperatures corresponding to the eutectic precipitation from the solution of individual chemical components make it possible to use the ultrasonic method to study the phase chemical transformations in saline ice with an undisturbed texture at any selected instant in time.

The study of the electrical properties of frozen rocks in alternating electromagnetic fields conducted by A. D. Frolov and his student B. V. Gusev is of undoubted interest. He succeeded in establishing the basic characteristics of the frequency and temperature dependence of the electrical properties of a frozen rock on its phase composition and in determining more precisely the conditions of propagation and absorption of electromagnetic waves in rock.

At the present time efforts are being made to use the induced polarization method to study the structural peculiarities of frozen rocks in order to ascertain the spatial arrangement of the ice bodies in permafrost. M. P. Sidorov and D. A. Fridrikhsberg[20] succeeded in formulating a theory that makes it possible in the first approximation to explain the nature of induced polarization as being due to differences in electrochemical activity in a capillary-porous ion-conducting system. With passage of a current in a heterogeneous frozen system, concentration variations take place in the connecting microsections, leading to the emergence of diffusion potentials, inasmuch as the adsorbed layers of bound water, including the electrolytes of the pore water, are held within significant negative temperature limits. A. M. Snegirev and L. L. Lyakhov,[21] in checking the experimental laboratory and field studies of the polarizability of dispersed frozen rocks, established its relationship with the physical and chemical properties. They demonstrated that the ice and its interface with the nonfrozen electrolyte exert a major influence on the initiation of the induced polarization processes.

The method used by I. G. Yarkin[22] is that of measuring the difference between the natural electric potentials originating at the interface between the frozen and thawed rocks, which makes it possible to discover the electrical nature of the migration of moisture toward the freezing front.

According to Gibbs, an isolated system is absolutely stable if with any possible and infinitely small variation in its state, when there is constancy of energy conservation, the inequality $dQ/T < 0$ holds true, where dQ/T is the increment of the entropy. In the case of some finite variations, when the increment of the entropy is greater than 0, the system can be considered relatively stable--metastable. The probability of the nucleation of crystals (of a new phase of the H_2O component) in metastable solutions rapidly diminishes with a decrease in the concentration of the component particles, although, in individual local sections of the system, owing to various fluctuations, the probability can exceed the limiting probability. The formation of the solid phase from highly supersaturated solutions will depend on the relation between the rate of accumulation of the molecules and their ordering rate. The metastable state does not correspond to stable thermodynamic equilibrium under these conditions, although a definite time period persists, being determined by the relaxation phenomenon, which in turn is caused by kinetic processes. There are two opinions with respect to the nature of origin of the crystal. According to one of them freezing begins in a single-phase medium at points of random collisions between molecules with the formation of centers of a new phase. In order that the new phase can grow, it is necessary that its vapor pressure be lower than the vapor pressure of the medium from which the new phase originated (the homogeneous theory). The proponents of the heterogeneous theory maintain that, although the process of homogeneous nucleation of the crystal is theoretically probable, in reality, owing to the presence in the liquid of foreign particles, from which in practice it is impossible to be completely free, it is these particles that cause the crystallization.

We will consider the kinetics of the phase transitions of water, using as an example the melting of ice (the transition of ice to water). The distinguishing feature of the crystal lattice of ice is the fact that it pertains to materials with a molecular structure; it is characterized by a regular arrangement of the molecules with periodic recurrence in space (remote ordering), but with a greater mobility of the hydrogen atoms than of the oxygen atoms. Consequently, in order to covert the ice to water, it is necessary to break the bond and remove the water molecule from the node (the potential well) by a distance beyond which elastic return is impossible (overcome the potential barrier) and to form a new aggregate system--a liquid--from such molecules. The activation energy of the molecule in the node of the ice lattice at negative temperature and normal pressure is less than the energy that must be expended upon overcoming the potential barrier, and if it were not for the elastic-vibrational and rotational movements of the molecules in the lattice, there would be no irreversible jumps. The irreversible jump (the translational movement of the particles) originating as the self-diffusion phenomenon, arises as a result of the fluctuations in the elastic vibrations of the molecules, promoting in some cases attenuation of the potential barrier and intensification of the activation energy. A rise in the lattice temperature of the ice will be accompanied by an increase in the range of the elastic vibrations of the molecules and, consequently, an increase in the irreversible jumps and a decrease in the bonds in a unit mass of ice. It must be noted that part of the molecules can be introduced reversibly into the vacant nodes, which leads to healing of the structural defect. For any negative temperature there must be a constant ratio between the number of free molecules n and the number of attached molecules m per unit volume (n/m). The coefficient n/m increases with a rise in the ice temperature and reaches a maximum at 0°C. At this temperature the energy coming into the ice will not be expended on increasing the heating, but rather on rupturing the bonds and increasing the number of unattached molecules per unit volume. When 80 cal/g are admitted to the ice and the number of broken hydrogen bonds reaches about 11-13 percent of their total number, the material enters into a new state--the solid is transformed into a liquid.

With the material in the liquid state a new singularity arises: on rupture of 11 percent of the bonds, the remote ordering disappears inasmuch as the settled lifetime of the molecules in the ice lattice decreases sharply to an extremely small value $\tau = 1.7 \times 10^{-9}$ s (at 25°C). Simultaneously, an abrupt increase in the drifting of unattached molecules is observed. Along with the aforementioned forms of movement of the liquid, group displacement appears. A possibility arises of breaking the bonds in integral units, and, under the effect of the collisions of the surrounding molecules, their displacement takes place. Thus, the self-diffusion process in the liquid D is made up of individual and group displacements:

$$D = A_1 l^{-\frac{E_1}{KT}} + A_2 l^{-\frac{E_2}{KT}},$$

where the index 1 pertains to the displacements of individual particles and the index 2 to the displacement of cells (groups), E is the activation energy of the discontinuity. Inasmuch as the molecular bonds and the spatial arrangement of the molecules and their settled lifetime inside the displaced cells remain approximately the same as in the principal molecular lattice, these groups of molecules cannot be called associations. Associated molecules, in contrast to the original material, have a different structure, and the nature and strength of the bonds are different.

With the disappearance of remote ordering in the structure, isothermal melting ceases and the influx of energy will be spent on changing the ratio between the number of unbound molecules and the number of molecules attached at the particular instant n/m. The energy previously expended on breaking the remote-ordering bonds (decreasing the settled lifetime of the molecules) will subsequently be spent on increasing the temperature of the material. In spite of the fact that water does not have remote ordering of the molecules, its state is such that it belongs among loose bodies with a mean statistical coordination number approximating to 4, which is close to the coordination number of the ice lattice. The distance between the nearest molecule in the ice lattice is 2.76 Å, while in water the distance between the attached molecules (at the particular instant) at 1.5°C is 2.90 Å; at 83°C it is 3.05 Å. Accordingly, it would appear that the density of water must be less than the density of ice; in reality, however, it is 10 percent higher. This is explained by the fact that the water density is not only a resultant of the attached molecules, but also of the molecules drifting at the instant in question from one node to another.

In a single-component water system, after cooling to below the region of stability of the high-temperature liquid phase, local sections of a new stable solid phase occur. Under the influence of the thermodynamic factors, these sections, once their magnitude exceeds some critical size, we will call them centers of the new phase, will begin to grow and will fill up the entire volume.

The sections of homogeneous medium that are smaller than the critical dimensions, that is, the nucleating centers, should decrease in size and disappear. However, as a result of fluctuations of the various system parameters away from their equilibrium towards the opposite side of what is thermodynamically acceptable, a favorable situation arises for the nucleating centers to continue to grow until their dimensions correspond to their stable growth. Thus, the growth of the nucleating centers is a result of the constant influence of the thermodynamic forces and of fluctuational variations in their sizes. The nucleation rate of the centers of the new phase J, in other words, the number of nucleations converted per unit time in a unit volume of the metastable system into centers of the new phase, is determined by the equation of the type:

$$J = \frac{L}{v_0}\left(\frac{KT}{h}\right)\exp\left(-\frac{EN}{RT}\right)\exp\left[-\frac{16}{3}\;\frac{\pi\sigma^3 N}{RT(\Delta F_0)^2}\right], \tag{1}$$

where σ is the surface tension at the phase interface; ΔF_0 is the variation of the free energy on formation of a unit volume of the new phase; h is the Planck constant; L is determined by the structural peculiarities of the new phase and it varies between 1 and 10; v_0 is the specific volume of the new phase; E is the activation energy of the transition of the atoms through the phase interface; r_{cr} is the radius of the critical nucleating center.

The size of the critical nucleating center r_{cr} is established by the expression:

$$2\sigma/\Delta F_0 = r_{cr}. \tag{2}$$

THE PHYSICAL-CHEMICAL NATURE OF ADSORBED WATER

The process whereby water is adsorbed by the surface of a solid, including the surface of soil particles, is due to the formation of nonautonomous liquid coatings. Their nonautonomous nature is manifested in the fact that the vapor pressure of such coatings is lower than the elasticity of the free liquid, and, with separation of the adsorbed liquid from the adsorbent, it becomes unstable. Possessing as it does differing degrees of binding as it moves further away from the surface of the foreign body, the adsorbed water is not converted to ice at any one temperature, as with a three-dimensional phase, but gradually becomes a solid phase, with the highest crystallization temperature always being below 0°C. The composition and structure of the adsorbed liquid coatings are caused by the field of the adsorbent, the metabolism at the liquid-solid state interface, and the external thermodynamic conditions. In the absence of exchange between the adsorbed water and the adsorbent, when the situation is such that both of these phases are in a state of energetic and mechanical equilibrium, and their compositon and structure remain practically unchanged, it becomes possible to express this heterogeneous system by an equation of the type:

$$F d\sigma_n = -\left\{ S^{(\sigma)} - m_n^{(\sigma)} S_M^{(\alpha)} - m^{(f)} \left[S_M^{(\beta)} \sum_{i=1}^{n-1} \left(x_i^{(f)} - x_i^{(\beta)} \right) \left(\frac{\alpha S_M}{\alpha x_i} \right)^{\beta} \right] \right\} dT + \left\{ V^{(\sigma)} - m_n^{(\sigma)} v^{(\alpha)} - m^{(f)} \left[v^{(\beta)} + \sum_{i=1}^{n-1} \left(x_i^{(f)} - x_i^{(\beta)} \right) \left(\frac{\alpha v}{\alpha x_i} \right)^{\beta} \right] \right\} dP - m^{(f)} \sum_{i,k=1}^{n-1} \left(x_i^{(f)} - x_i^{(\beta)} \right) g_{i,k}^{(\beta)} \cdot dx_K^{(\beta)} \ldots, \quad (3)$$

where F is the area of surface of the discontinuity; σ is the surface tension, S is the entropy, S_M is the molar entropy, and the indexes are as follows: (n) refers to the adsorbent material, (α) to the three-dimensional phase of the adsorbent, (β) to the liquid or other state (ice or a gaseous medium interfacing with the adsorbed layer on the opposite sides of the adsorbent, (f) to the adsorbed liquid layer, (σ) to the entire surface layer (including the liquid coating and the surface layer of the adsorbent; V is the volume, v is the molar volume: $[V^{(\sigma)} - m_n^{(\sigma)} v^{(\alpha)}]$ is the volume of the adsorbed water, $x^{(f)}$ is the mol fraction of the i-th component in the adsorbed liquid coating, $x_i^{(\beta)}$ is the mol fraction of the i-th component in the liquid (or other state) interfacing with the adsorbed liquid, m^f is the total mass of the adsorbed liquid, T is the absolute temperature, P is the pressure

$$g_{i,K} = \frac{\alpha^2 g}{\alpha x_i \cdot \alpha x_K},$$

where $g = G/m$ is the molar thermodynamic Gibbs potential, the index K refers to the critical state; G is the thermodynamic Gibbs potential for the flat surface of the discontinuity.

$$G \equiv U - TS + P_N V - \sigma_n A,$$

where U is the energy. The elucidation of the mechanism of interaction of water with the surface of a solid body is connected with the study of the nature of the adsorption centers. The knowledge of the essence of these processes in such complex heterogeneous systems as soils, due to the superpositioning of a multiplicity of uncontrolled factors on the investigated phenomenon, is attended by enormous difficulties. Accordingly, researchers are trying to discover the basic laws of the phenomenon of the interaction of water with the adsorbent in simple model systems.

The most convenient models, in the opinion of V. F. Kiselev, L. A. Ignat'yeva and V. I. Kvlividze[4] are the oxides, as these are the principal components of the vast majority of rocks and their surfaces are accessible to study by the methods of nuclear magnetic resonance and infrared spectroscopy.

The mechanism whereby water is adsorbed by oxides, according to contemporary thinking, is based on the notion that the oxides have hydrate coatings, which contain the hydroxyl groups, valence-bonded to the surface atoms of the water molecules, these atoms interacting in different ways with the oxide atoms. In turn, the water molecules are linked together by hydrogen bonds.

The idea that the interaction between the OH group of hydrated surface oxides and the adsorbed water molecules is accomplished by the mechanism of the hydrogen bonds has not been confirmed by data resulting from determinations of the heats of adsorption in a region of small surface coverings. It turned out that the heat of adsorption is within the 15-20-kcal/mole range, which cannot correspond to the formation of hydrogen bonds. Kiselev, Ignat/yeva, and Kvlividze[4] ascribe the high values of the initial heats of adsorption in the oxides to the presence of the coordination bond mechanism. For example, the $Si(OH)_4$ tetrahedrons are coordination unsaturated, and in accordance with the conditions of cation shielding can encompass the two water molecules. It must be noted that, when investigating the adsorbed mechanism of the bond, no allowance was made for the chemical processes on the surface; and, in particular, their effect on the electrochemical and ion-exchange properties was not considered. In natural disperse bodies there is cation exchange that leads to complication of the adsorbed processes.

By using the nuclear magnetic resonance method, it proved possible to detect a significant decrease in the mobility of the water molecules in the initial adsorption region.[5] With an increase in the amount of water the mobility of the molecules increases and is closely connected with the structure of the adsorption layer. The effect of the geometric dimensions of the cavities on the structure of the water was traced on the crystalline adsorbent--zeolite. The structure and the geometric cavities of the crystal lattice of zeolite, in which the water is present, were investigated by means of X rays. When heated, the zeolite is dehydrated without rupturing the alumino-silicon-oxygen skeleton. The state of the water in zeolite is closely connected with the geometry of the lattice cavities.

Zeolite contains isolated water molecules rigidly attached, in pores with diameters from 4 to 7 Å in which the molecules have two rotational degrees of freedom and in pores with diameters of up to 13 Å in which the water molecules have all of the rotational and translational degrees of freedom. With small surface coverings a molecule of water cannot form the four bonds,* which is a necessary condition for rigid fixation of the molecules. In this case the water is evidently only connected to the cations of the skeleton by electrostatic forces. Therefore, the mobility of the molecules increases at higher temperatures. The increase in the number of adsorbed molecules leads to an increase in the average number of bonds per molecule. As a result of the emergence of the hydrogen bonds between the molecules, the adsorption energy decreases. The mobility of the adsorbed water molecules decreases in the zeolite series containing K, Na, Ca, Li, and Mg, which agrees with the ideas of O. Ya. Samoylov[12] regarding negatively and positively hydrating ions. The structure of the layer of adsorbed water at the surface is connected with the geometry of distribution of the active centers, but it will differ from the structure of the ice and the stereometry of the free water.

B. N. Dostovalov[2] considers that the layer of adsorbed water consists of three zones. The first is directly adjacent to the adsorbent and characterized by the transition to the solid state; as a result, bonds that are "broken" during melting form anew. This concept of the structure of the layer of water closest to the adsorbent does not take into account the effect of the geometry of the active adsorption centers and their wide diversity, as a result of the differing structure and composition of the adsorbents. In the second zone an increase in mobility and a decrease in activation energy as compared with the free water must occur, which, in his opinion, is caused by a decrease in the dimensions of the association and must lead to significant changes in the properties of the water in this zone. The origination of the more mobile intermediate layer of the water in the event that on one of its interfaces there is a third layer less subject to the effect of the fields of force is improbable. Owing to the self-diffusion processes, the absence of rigid attachment in the water molecules of the third zone must of necessity lead to dynamic equalization of the mobility with the molecules of the second zone, that is, the second zone of more mobile water will be in a more stable state in relation to the third. The origination of the more mobile intermediate layer of adsorbed water under these conditions can only be conceived of in the event that there are present in this layer ions with larger crystal-chemical radii in relation to the ions of the other two zones, that is, when there is special differentiation of the ions in the adsorbed water layer.

The proposal advanced by Dostovalov regarding the presence in the water of associated molecules with a differing degree of stability and duration of existence is not confirmed by the experimental studies in which the spectrum, and X-ray, methods were used. It has been established (quantitatively) that, along with the individual discontinuities of the molecules, group molecules are also encountered, both in the case of homogeneous media and in solutions. However, the duration of the attached state of the water molecule in the entire mass and in the individual broken block (cell) is the same. Thus, there are no grounds for regarding the displaced blocks (cells) as associations with properties other than those of the remaining mass of the water.

Whereas in studying the structure and properties of the adsorbed water when there is only small coverage by it of the adsorbent surface the methods of spectral analysis and nuclear magnetic resonance are being successfully used, in the case of large surface coverings the necessity has arisen for using other methods. In clays, for example, the study of the properties of the bound water is based on the electrokinetic characteristics of the clay. The basis for this method is the concept of the mechanism involved in the formation of the boundary layer, which is due to the presence of an electric charge on the surface of the solid particles and to the origination of a double electric layer composed of hydrated ions and molecules of bound water. The sign and magnitude of the electric charge depend on the size of the particles, the composition of the mineral, and the crystal-chemical structure, which is undoubtedly reflected in the structure and properties of the adsorbed layer of water.

In contrast to the spatial structure of the crystal lattice with the saturated bonds between particles of different signs forming the structural coordination, in the boundary regions, for various reasons noncompensated charges of the particles appear. One of the reasons for the origination of the additional sources of the charges is dissociation of the molecules of the solid surface of the clay particles into ions. Some of these ions diffuse into the disperse medium, and the remaining ions connected with the solid phase serve as an additional source of particle surface charge. In some cases completion of the crystal lattice of the minerals occurs as a result of adsorption of the ions from the solutions. This is due to the origination of stronger chemical bonds, which serves to prevent their inverse solution. The adsorption of the ions from the solution increases as a result of the polarization of the

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*The water molecule can form four tetrahedral hydrogen bonds.

surface atoms by the adsorbed ions (Van der Waal's or electrostatic forces of the bonds). The formation of the active centers of the surface also depends on the presence of the OH group in the crystal lattice of the majority of clay minerals. With an increase in the pH of the medium, the OH groups are capable of exchanging their hydrogen for metallic ions, in other words, they must contribute to cation exchange.

Thus, at the interface between the loose rock and the solution, adsorption centers form that are qualitatively different as a result of the dissimilar crystal-chemical structure of the mineral surface, the different compositions and properties of the solutions in contact with them, and the variable dynamics of the physical and chemical processes in the boundary layers.

As a result of the excess of ions of the same sign or of a like surface charge, electrostatic attraction arises along with a corresponding orientation of the water dipoles, ions, and other polar particles of opposite charge. Thus, a double electric layer originates that is similar to a capacitor consisting of two plates of opposite sign. The sign and magnitude of the charge of the mineral particles (the inner plate) are formed as a result of the large number of qualitatively different bond centers. The formation of the properties and the structure of the "outer plate" of opposite sign depends on the magnitude of the charge of the "inner plate." The layer of water molecules and ions held near the "inner plate" at a distance of several angstroms is called the adsorbed layer. Apparently, the thickness of the adsorbed liquid layer does not exceed 20 Å although some researchers accept a greater thickness--100-150 Å. The mobility of the excess ions of the double electric layer results in the origination of additional conductivity.

In the opinion of Ye. M. Sergeyey, R. I. Zlochevskaya, R. S. Ziangirov, and A. N. Rybachuk,[13] the presence of a significant concentration of exchange cations in the strongly bound part of the adsorbed water determines its nonsolvent action, owing to shielding of the water molecules through the action of the surface particles and cation fields.

Based on a calculation of the distribution of the surface potential and the electric capacitance of the electric double layer of clays by its diffuse and tightly bound ("wall") parts, these authors concluded that the tightly-bound water does not participate in the electrokinetic processes. There is also the view that the adsorbed water has an enhanced capacity for solution in comparison with unbound water as a result of an increase in the dipole moment, and, consequently, of an increase in the dissociation of the water molecule under the expitaxial effect of the substrate.[1] The other part of the adsorbed water, the most mobile part, is called diffuse. Its capacitance is about 30 percent of the adsorption water. The specific capacitance of the tightly bound water varies little for the various finely dispersed soils; in the diffuse layer the greatest magnitude is detected in askanite gels and the smallest, in kaolin.

In the argillaceous minerals there is a definite relationship between the type of crystal lattice, the crystal-chemical peculiarities of the surface structure and the positioning of the various active adsorption centers. Two basic groups of argillaceous minerals are distinguished. One of them (the montmorillonites and hydromicas) includes minerals with electrically nonequilibrium lattices, the neutralization of the charges of such lattices is accomplished by exchange cations (Na^+, K^+, Mg^{2+}, Ca^{2+}, and so on) or nonexchange cations (most frequently K^+).

The basal surfaces of these minerals are of the same type and consist of the oxygen atoms of tetrahedral grids. The second group of argillaceous minerals (kaolinite and halloysite) is characterized by electrically equilibrium crystal lattices, and therefore having no isomorphic substitutions. With montmorillonite clays, for example, up to 80 percent of the capacity is accounted for by isomorphic substitutions in the lattice, while the exchange capacity of the kaolinite minerals is due to rupture of the bonds along the surfaces of the particles.[8]

There is no single opinion regarding the structure of the adsorbed water. Macey (1942) considered that bound water has the structure of ice; P. A. Shumskiy[17] supported the view that bound water is a solidlike overlay of modified ice, the single layer being likened to UP ice. Hendrichs and Jefferson (1938) considered that the molecules of bound water are arranged in layers. In each such layer the molecules are linked together in a hexagonal grid with tetrahedral charge distribution around the water molecules. Each side of the tetrahedron coincides with the direction of the hydrogen bond. The water lattice is held directly against the surface of the argillaceous mineral by oxygen atoms, the latter forming hydrogen bonds with the hydrogens of the water located on the outer edge of the lattice.

The results of X-ray diffraction, spectrum and nuclear magnetic resonance studies of the adsorbed water, obtained in recent years, force us to reject the hypothesis regarding the solidlike states of bound water. The density of bound water depends primarily on the concentration and nature of the active centers and is not equal to the density of ice. The anisotropy in the structure of the bound water does not correspond to the three-dimensional equivalence in the positioning of the unit cells of the ice lattice in space. Arguing against an ice structure is the reduced freezing point of the bound water, which remains unfrozen to very low temperatures. The difference between the structures of bound water and of ice was confirmed by the X-ray diffraction analysis. The width of the nuclear magnetic resonance lines of the protons of bound water with low moisture is much less than for ice. No discontinuous broadening of the nuclear magnetic resonance lines is observed when the temperature drops below the freezing point of water.

According to G. B. Bokiy (1961), the surface of the single crystals of argillaceous minerals is formed by the bases of the silicon oxide tetrahedrons. The water molecules are positioned in the cavities of the hexagonal mineral lattice. At the level of the lower oxygen atoms at the bottom of the well there are OH groups of the octahedral layer binding two Al and Mg atoms. The proton of the OH group is turned outward and can form hydrogen bonds with the water molecules getting into the well. Inasmuch as the distance between the centers of the hexagonal wells is 5.5 Å and the distance between the molecules in ice and water is about 3 Å, the adsorbed water molecules cannot form hydrogen bonds with each other and create a film of water of normal structure. In addition, the water molecules bound to the OH group by the hydrogen bond accomplish vibrational movements lengthwise along the axis of the well and jump from one well to another, being reminiscent of the translational movements of the molecules in a volumetric liquid. Thus, in terms of its state, bound water is closer to a liquid than to a solid. The high value of the heat capacity and entropy of the bound water at low humidities can be attributed to the higher intensity of the vibrations and jumps of the molecules in the hexagonal cavities as compared with the vibrations and jumps in liquid water. With an increase in moisture the bonds between the molecules grow and this is accompanied by a decrease in the heat capacity and the entropy of the bound water.

Low and Anderson (1958), who studied the mobility of exchange cations (Na^+, K^+, and Li^+) in bentonites, found that there is a relation between the structure of bound water and the type of cation.

In the first batches of the water the adsorption of the water by the kaolinite surface was evidently caused by the coordinating bonds of the Al or Si atoms. Subsequently, adsorption can be originated in the OH groups by the hydrogen bonds.[14]

EFFECT OF ADSORBED WATER ON THE STRUCTURE OF ICE INCLUSIONS

In soils, ice crystals can form from water that is adsorbed and not bound, and from the solution and vapor contained in the freezing body. Recrystallization is encountered which is caused by internal thermodynamic, thermochemical and migration processes. The conditions giving rise to the nucleation and growth of the crystals from the liquid medium are its supercooling or a change in the concentration of the substances dissolved in it. The supercooling of the liquid medium or the supersaturation of the vapor-line phase are caused by various factors. Foremost among these are changes in the temperature, pressure, volume, concentration, and so on. The formation of the ice crystals is caused by the internal molecular processes occurring in the liquid medium. The origin and growth of the crystals depends on the rate of accumulation of the molecules near the crystallization center and their ordering in the ice lattice. The accumulation rate is determined by the selective translational motion of the water molecules in the liquid, and the rate of their ordering at the nodes of the crystal lattice depends on the energy of the water molecules themselves and on the force field arising near the mineral substrate.

The crystallization gives rise to a process of differentiation of the rock and to the formation of a diversified texture (massive, layered, reticular). Inasmuch as loose mineral rock differs in its composition and structure, its effect on the adsorption of water, migration, and crystallization will also differ. Consequently, during freezing the origination of an ice structure corresponding to the surface composition of the mineral skeleton must be expected to occur. Unfortunately, for a long time it was believed that the surface of a solid heteromorphic base is passive with respect to the shaping of the strucure of the forming ice (Shumskiy, 1953). It follows from this that the mineral and chemical compositions, the degree of dispersion, the presence of uncompensated charges, the dislocation of the mineral crystal lattice, and so on, have no effect on the originating ice structure.

In order to ascertain the effect of the substrate surface on the shaping of the ice structure, I. B. Savel'yev[10] simulated the freezing of water in certain frozen fine soils in a cooling chamber and made an ensuing study of the structure of the ice formations in the contact layers and in the more remote horizons of the ice cover. He established the presence of contact layers of ice forming with a specific structure, which differs from that of the remainder of the ice situated more remotely from the mineral surface. The thickness of the contact layer of ice depends on the seepage of moisture into the ice forming at the boundary with the adsorbed liquid layer. Evidently the thickness of this layer can range from fractions of a millimeter to 5 mm. Owing to the orthotropic growth of the crystals, they are much larger at a distance of 3 cm from the substrate surface than in the contact layers. In the structure of the ice cover the surface of the base has a pronounced effect on the dimensions and shapes of the crystals and almost no effect on the character of the prevailing orientation of the optical axes. The crystal dimensions decrease in the following sequence as the degree of activity of the base increases: glass--quartz plates--frozen sand--frozen kaolin--frozen askanite gel. With lowering of the freezing point, the sizes of the crystals decrease and the effect of the base surface on the ice structure becomes weaker. The adduced results of experimental studies of ice coatings by Savel'yev indicate the active effects of the frozen base on the structure of the ice. Evidently, at relatively high negative temperatures, when the intermediate liquiform layer between the ice surface of the frozen rock and the ice is sufficiently thick, the bond strengths between the water molecules farthest removed from the mineral substrate will decrease significantly and will approximate to that of the bond between the free water molecules. Under these conditions a contact layer of ice may not even originate at all, in which case the ice coating will have a uniform structure throughout the entire mass inasmuch as the force exerted on the ice by the horizon of the molecules of the adsorbed water layer situated closest to it will prove to be inadequate.

It was on the basis of X-ray diffraction analysis, measurement of the magnitude of the ice polarizability, and spectroscopic studies that the position of the oxygen atoms frozen into the crystal lattice of the ice was established. As for the spatial distribution of the hydrogen atoms, this problem has yet to be resolved. Because of the weak scattering of X rays by hydrogen atoms, it has not proved possible to use the X-ray analysis technique to determine their location in the ice lattice. P. G. Owston and K. Lonsdale (1948) postulate that the hydrogen atoms in the crystal lattice of ice are continuously migrating in the space between each pair of oxygen atoms. This type of hypothesis was predicated on the Laue diffraction patterns of the complex diffuse pattern of the ice originating during prolonged exposure.

That picture of the ice structure is more complex than it is for other solid crystalline bodies gave the above-mentioned authors grounds for believing that it had originated either as a result of grouped movement of the oxygen and hydrogen atoms or of grouped movement of disoriented molecules. With a reduction in temperature the diffuse pattern disappears and below -78°C it is not seen at all. It is evidently the exceptional mobility of the hydrogen atoms leading to rupture of the hydrogen bonds in the molecular structure of the ice lattice that explains the exceptional fluidity of the ice and the fact that it lacks a stress-rupture strength. At temperatures below -78°C, a weakening of the mobility of the hydrogen atom in the ice occurs: Its hardness increases discontinuously, and it goes from the liquidlike state to the solid state (according to Academician R. A. Rebinder's Classification). Under these conditions the ice must have a stress-rupture strength.

The study of nuclear-magnetic resonance spectra of polycrystalline ice carried out by V. F. Kiselev, V. I. Kvlividze and A. B. Kurzayev[6] pointed to the presence on the surface of the ice body of a mobile layer with a temperature below the melting point (T_{melt}). It was demonstrated that the impurities and defects of the crystal lattice of the ice are not the reason for the appearance of the mobile phase near the surface. A mobile layer is detected experimentally on the ice-gas surface and the ice-foreign hydrophobic solid-state surface. The mobility of the lattice elements on the disordered surface layer proved to be appreciably higher than in the body of the crystal. Lowering of the temperature below -5°C is accompanied by a noticeable decrease in mobility in the surface layer. Presumably the high mobility of the lattice elements on the surface of the ice is due to the exceptional mobility of the hydrogen atoms detected by Owston and Lonsdale. Despite the fact that the mobility of the molecules on the surface of the ice is closer to the mobility of the water molecules than to the mobility values within the ice body, it is too early to assume that the mobile layer on the ice surface is quasiliquid and lacks the elements of an ice lattice structure. In contrast to the adsorbed water on a foreign body in which there is a clearly defined interface between the phases of the two different components, there is no clear interface between the moving surface layer and the remaining mass of the ice. On the contrary, a gradual lessening of mobility from the surface into the interior of the ice body is traced here, and at a specific distance the mobility stabilizes upon reaching a minimum.

When studying the distribution of trace elements in fresh ice coatings on frozen foundations by means of an atomic-absorption spectrophotometer, I. B. Savel'yev[11] discovered that the cation content in the ice (Ca^{++}, Mg^{++}, Na^{+}, K^{+}) increases with distance from the contact with the foreign body. In this way, he established that the mineral particle not only affects the structure of the contact ice but also the redistribution of the trace elements in the ice.

Based on the foregoing it is our contention that the originality of structure in the boundary region of the frozen soil consists in the multilayering that occurs between the mineral part of the skeleton and the ice. The multilayered system includes the adsorbed ice, which in turn consists of the tightly bound layer and the diffuse shell of the mobile surface ice and contact layer of ice situated adjacently to the remaining mass of the ice body. The diffuse shell and the mobile surface layer of the ice are most dynamic in the frozen system. These can undergo significant transformations as a result of variations in the thermodynamic conditions and through the influence of physical-chemical and chemical processes.

THE SPECIFIC NATURE OF THE MIGRATION OF MATTER IN FROZEN SOIL

During the freezing and thawing of soil and when it is in the frozen state, redistribution processes involving the migration of matter occur as a consequence of various external and internal factors. It was found by observation that water in the vapor and liquid phases migrates primarily toward the freezing front. In a number of cases, migration of ice inclusions from a zone of larger pressure to regions of smaller pressure is noted. The redistribution of the different phases of the water is accompanied by a relative displacement of the mineral parts in the soil. In addition, ionic displacement of the individual components of the soluble salts has been detected.

Vapor displacement of water occurs in soil with an appropriate degree of porosity as a result of the origination of a vapor pressure gradient caused in turn by the temperature or pressure gradients or by the physical-chemical processes.

The migration of water in the liquid state in soil when a nonuniform force field is present is governed by the different number of jumps of the molecules of water and ions and of group jumps in the forward and reverse directions along the gradient force line. In this case the resultant of the translational movement of the particles in different directions for the given volume of liquid is not equal to zero. The liquid will be displaced in the direction in which the particle jumps are more numerous. The migration rate is caused by the mobility of the molecules, their "settled lifetime" (the period during which the molecules are at the node of the instantaneously attached matrix), and the self-diffusion magnitude (the translational motion). Thus, in the liquid state there is a single internal mechanism of movement in the soil; whereas the causes of migration are numerous. Chief among these are the gradients of the temperature, moisture, pressure and density, the concentrations of the solutes, and surface tension caused by the differing mineralogical composition of the soil skeleton, meniscus forces, gravitational forces, and so on.

The main modes of migration of matter is coarsely skeletal, highly porous, and fissured rocks are by pressure head and pressureless percolation of water and movement of vapor-air flows.

The transport of matter in fine-grained soils having strongly developed interfaces is in large measure due to electrostatic and physical-chemical interactions between the molecules and ions of the adsorbent and the molecules and ions of the surface and volumetric phases of the porous solution. The rate of migration of the water in noncohesive soils depends on the content of soluble ions and the degree of moistening. It is known that, depending on the dimensions of the crystal-chemical radii of the ions in weak solutions, the activation energy and mobility of the water molecules will range within wide limits. As the crystal chemical radii of the ions increase, the mobility of the water molecules becomes greater.

When the moisture does not exceed the maximum-hygroscopic capacity, the migration of moisture takes place mainly in the vapor state. Here, migration both at positive and negative temperatures is characterized by closely similar quantitative indices.

If the moistening zone lies between the maximum hygroscopic and the maximum molecular moisture capacity, then the movement of moisture is accomplished in both the liquid and the vapor state.

A moistening zone exceeding the maximum molecular moisture content is characterized by an abruptly varying rate of migration in thawed and frozen soils. In thawed soils, as the moisture content rises migration increases and reaches significant values with a water content close to saturation. During freezing, the moisture moves up from the thawed layers towards the crystallization front. In frozen soils, migration occurring as a result of freezing of the basic, most mobile part of the moisture decreases significantly on account of a decrease in the porosity, the surface activity, and the lessening of mobility of the molecules in the adsorbed water.

Depending on the rate of freezing and of migration of moisture towards the freezing front, the frozen soil displays a highly distinctive structure (massive, layered, or reticular). The phenomenon of moisture migration and simultaneously freezing is accompanied by an increase in the volume of the soil system, which is known as heaving.

Only minimal attention has been devoted to the migration of matter in the solid state (we are speaking primarily of ice) under the influence of a stress field in the rock. This is partly due to the fact that this process takes place slowly owing to the high viscosity of the ice. It is possible to propose a twofold mechanism of movement of ice under pressure: (1) the migration of the ice occurs primarily as a result of displacement of the matter in the surface mobile layer in the direction of lower pressures; (2) in the zone of higher pressure a part of the ice is transformed into bound water, which in the presence of the surface tension gradient shifts towards the lower pressure zone and freezes. As a result, in the lower pressure region the ice will increase as a result of a decrease in it in the region with higher pressure.

THE PECULIARITIES OF FORMATION OF THE STRENGTH OF PERMAFROST

The freezing of noncohesive soils is accompanied by a sharp rise in their resistance to mechanical effects in comparison with soils in the thawed state. The frozen soil has a new form of bond, which is due to the formation of ice. The strength of the permafrost is determined by the strength of the mineral particles and ice inclusions and the size of the contact bond between the ice and the mineral skeleton. The mineral particles have the greatest resistance to deformation and rupture due to the applied force. As for the strength of the ice and its contact bond with the mineral constituent, depending on the composition and structure of the permafrost elements and the thermodynamic conditions, the adhesive bonds can prove to be either stronger or weaker than the cohesive strength of the ice.

The strength of the permafrost is not simply the sum total of the strength of its parts. This property arose not as a result of the adhesiveness, but owing to the special form of the structural bond. It was noted that under the special thermodynamic conditions and moisture, attending the formation of permafrost a contact layer of ice originates, whereas if the circumstances were different it could not do so. In addition, at a relatively high negative temperature the ice surface is characterized by a higher mobility of the molecules than is the case when they are inside the ice body. Between the ice and the mineral part there is a layer of unfrozen adsorbed water, the thickness of which varies within wide limits as a function of the variation in the negative temperature.

The adhesive bonds will be stronger than the cohesive strength of the ice if the formation of the permafrost occurred at low negative temperature. Under these conditions, the contact layer of the ice is formed, and the adsorbed intermediate liquidlike layer will be less thick with bond strengths of several thousand kilograms per square centimeter; that is, a tightly bound layer of water orginates. On contact with the tightly bound water on the ice surface the mobility of the molecules becomes even less than the mobility inside it. In this case the adsorbed water is a factor strengthening the permafrost. As Savel'yev's[11] experiments showed, in the presence of contact ice the separation of the ice from the frozen mineral in all cases displayed a cohesive character, that is, separation of the ice took place along the ice at the interface between the contact layer of ice and the basic mass of the ice. As a result of the separation, a layer of ice 1-3 mm thick was left on the surface of the substrate. This layer of ice had the structure of the contact layer. The strength of the bond between the contact layer and the remaining mass of ice becomes greater with an increase in the number of crystals per unit volume and with a reduction in temperature.

The cohesive strength of the ice becomes greater than the adhesive bond in the presence of a substantial layer of adsorbed water, which can originate either with a rise in the negative temperature or because of an increase in the concentration of the dissolved salts in the diffuse layer of the liquidlike interlayer.

The increase in thickness of the liquidlike layer will be accompanied by a weakening of the bond of the molecules of this layer at the boundary with the ice and an increase in mobility of the molecules of the surface layer of ice. As a result there is a reduction in the strength of the ice bonds with the mineral interlayer through the intermediate liquidlike layer, and the latter changes from being a factor strengthening the adhesive bonds to a factor weakening them.

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FUNDAMENTALS OF GEOCRYOLOGICAL SURVEYING AND FORECASTING

V. A. KUDRYAVTSEV Moscow State University, and I. A. NEKRASOV
Permafrost Institute, Siberian Division of the USSR Academy of Sciences

INTRODUCTION

One of the most important natural factors that must be taken into consideration is the search for solutions to problems relating to the efficient placement of productive forces is permafrost, occupying as it does more than 80 percent of the area comprising Siberia and the Far East. It is for this reason that the development of this area is being preceded by the study of geocryological conditions. For purposes of long-term planning, it is necessary that such studies should encompass at the earliest possible date all of the eastern regions of the country. Moreover, in order to solve specific problems pertaining to the construction and placement of industrial complexes, more detailed studies of smaller areas are required.

In view of the fact that geocryological studies are being conducted on a continuously increasing scale and by various departments, in recent years the question of the need for the standardization of permafrost studies, that is to say the working out of unified regulations governing their conduct, has come to be especially urgent. These regulations must call for a clear definition of the studies to be undertaken, their sequence, and the methodology of determining the individual parameters. This is necessary in order that the results derived in the course of the research can be standardized and easily compared, and that the entire sweep of the problems can be elucidated, both from the standpoint of general geocryological research and for the solution of applied problems confronting planners, builders, and users. In addition, these studies must be carried out as quickly as possible and by the cheapest possible means.

The geocryological survey is a strictly regulated set of office and field operations enabling us to obtain an overall picture of the geocryological conditions in a specific area or region. The procedure followed in the fieldwork is always the same in that it entails unswerving compliance with the norms established for mineral workings that are mandatory for the scale in question. The survey is conducted both as a means of making a general study of the natural conditions and for specific utilitarian purposes, in relation to which its content and methodology can be varied to some extent.

Thus, the geocryological survey furnishes all of the source information needed for a general geocryological description of the area being studied, as well as the data required for the design and construction of various installations and for solving all of the remaining problems connected with the industrial development of the area.

The foundations of geocryological surveying were laid in Russia when in the late nineteenth century Russian engineers concluded that it was necessary to draw up special instructions regulating the study of "eternally frozen ground." The special commission for the study of the permafrost region, which was set up at the request of the Siberian Highway Construction Administration, compiled the first highly detailed manual entitled "Instructions for Studying the Frozen Ground in Siberia". This was published in 1895 under the editorial supervision of I. V. Mushketov.[63] Seventeen years later a second edition entitled "Instructions for the Study of Frozen Ground" was published. The editors were L. A. Yachevskiy and P. I. Vannari.[64]

The next stage in the creation of a procedural handbook for geocryological research was the publication of a "Collection of Instructions and Program Directives for the Study of Permafrost and Frozen Grounds" prepared by the co-workers of the Permafrost Commission of the USSR Academy of Sciences jointly with the Main Administration of the Northern Sea Route, the Fundamenstroy Trust, and the Main Geological Administration of the People's Commissariat of the Full Industry.[160] It was edited by V. A. Obruchev and M. I. Sumgin. The appendix of the collection contains a "Design for the Study of Permafrost for Construction Purposes," in which the following problems facing the "permafrost" researchers are formulated:

a) Determining the depth of occurrence of the permafrost table;
b) Determining the type of permafrost in terms of its occurrence at depth (continuous, layered, and so on) and its geographic distribution;
c) Determining the nature of the relief of the permafrost table;
d) Studying the thermal regime of the active layer and the permafrost series;
e) Studying the regime of the suprapermafrost water and other types of water;
f) Studying the ice inclusions in the permafrost and the dynamic processes in the active layer (heave mounds, naleds, and so on);
g) comprehensive investigation of the physical and technical properties of the soil in the active layer and the permafrost (p. 255).

The concept of the geocryological survey dates from the 1940's and was formulated by V. F. Tumel'.[178] In his opinion, "...the permafrost survey...includes the aggregate of field operations aimed at collecting the information required for the generalized evaluation and cartographic representation of the permafrost habit of the areas being investigated" (p. 135). The concept was borrowed by the geocryologists from related sciences in which the concepts of the geological, hydrogeological, geomorphological soil, and other types of surveys had been formulated and strictly regulated for a long time. In all of the surveys a mandatory requirement which also distinguishes the survey from other research--is clarity and definiteness of purpose and also the systematization of research work when the region to be surveyed is covered by a more or less uniform network of transportation routes or mineral workings, all being strictly subordinate to the scale of the survey.

Subsequently, questions pertaining to geocryological surveying were investigated by many organizations and researchers. Until its dissolution in 1962, the V. A. Obruchev Permafrost Institute was at the forefront of all of the geocryological research in the USSR. It was here in particular, in papers by I. Ya. Baranov,[5-8] V. K. Yanovskiy,[217] P. F. Shvetsov,[196,197] *et al.*, that many of the problems in geocryological surveying were formulated and set out in detail. Here also, a new procedural instruction ("Geocryological Field Research," 1961) and "An Instruction relating to the Study of Ground Ice" (Shumskiy, Shvetsov, and Dostovalov, 1965) were compiled.

In the early 1950's Departments of the Geocryology and Geography of the Polar Countries formed (now the Departments of Cryolithology and Glaciology) were formed at Moscow State University. Problems in geocryological surveying have ranked high among the investigations by the scientists of these department.[55,77,85,94,100,102,104,105,117,136,139,140,142,207-209,215]

The protracted studies by the scientists of the Geocryology Department culminated in the issuance of an "Instructions and Procedural Information for Carrying Out Integrated Permafrost Hydrogeological and Engineering Geological Surveys," which are now in general use, not only among geocryologists, but also in the geological subdivisions of the USSR Ministry of Geology.

Also concerned with problems of geocryological surveying are the All-Union Research Institute of Hydrology and Engineering Geology (VSEGINGEO),[25,122] the Institute of Permafrost Studies of the Siberian Division of the USSR Academy of Sciences, which was formed in 1962 at Yakutsk; the exploration trusts and institutes of the USSR Gosstroy;[184] and, in particular, the Production and Scientific Research Institute for Engineering Surveys in Construction, which is publishing recommendations relating to individual aspects of geocryological surveying.[26,58,182]

As a result, the importance of geocryological surveying has become universally recognized as a means of gaining an understanding of the permafrost region of the USSR and it has been adopted as a tool by all of the organizations engaged in differing degrees with permafrost research in the arctic and subarctic zones of the country.

It remains only necessary to emphasize that, even though it has recently become generally recognized that during the geocryological survey a study must be made of the general and partial laws governing the formation and development of the permafrost zone, its relation to the environment and its place in the overall history of the Earth's geological development during the Quaternary, such an understanding of the basic goal and content of the geocryological survey is by no means always encountered.

Some researchers, particularly specialists in engineering and allied fields who do not have geocryological training, consider that the basic task of the geocryological survey is to record the morphometric characteristics of seasonally and perennially frozen series and describe the phenomena accompanying them, thereby demonstrating that it is entirely sufficient to present a picture of the distribution; mode of occurrence; temperature regime; composition, texture, and cryogenous structures of the permafrost and seasonally frozen rock; and to describe all of the geocryological phenomena encountered. In spite of the fact that these characteristics are bound up with environmental conditions, as a rule they are presented in their own type of photograph, that is, the total conformity of the geocryological conditions to the natural situation is recognized without any explanation being given of the various peculiarities recorded at the time of the study.

It is entirely obvious that this approach cannot lead to an understanding of the phenomenon being investigated, as knowledge of a phenomenon implies to understanding the laws governing its development. In the above-cited case, only the morphometric characteristics of the permafrost zone are studied, or, in other words, only what can be measured and described. Obviously, in the final analysis this procedure will lead to negation of the possibility of gaining an understanding of the phenomenon.

At this stage in its development geocryology has overstepped the limits of this primitive concept of the geocryological survey and has swung over to the study of the global and particular laws of development of the permafrost region and the entire range of geocryological conditions, seeing them as the result of the overall course of development of the Earth during the Quaternary.

In the course of the geocryological survey, the laws governing the formation of the permafrost and its interrelation to the environment, the peculiarities of the radiation-heat balance of the area being investigated, the geological structure of the region, and the composition of the earth materials are determined.

Also ascertained during the survey are the causes that in the area in question gave rise to precisely these rather than other geocryological conditions, as well as the role of each of the geological-geographic factors in the genesis of these conditions. Here, the evaluation of the factors is made not only from the qualitative but also from the quantitative point of view.

In the course of the geocryological survey, it is not only the effect of individual factors or particular laws that must be discovered and investigated, but also the nature of their interrelation and the effect of the entire range of natural conditions on the formation of the permafrost and of the cryogenic phenomena and processes, in other words, the general laws governing its formation and development.

It is based on the knowledge of these laws, considered in relation to the peculiarities of the geological-geographic situation, that the overall picture of the geocryological conditions for the region as a whole and for each of its constituent areas is compiled.

Central to the methodology in geocryological surveying is the affirmation that a close interrelation exists between the permafrost conditions and the geological-geographic environment or types of landscape. It is this that determines the content and sequence of the geocryological survey. The landscape technique itself came to geocryology from allied sciences (geography, geobotany, soil science, geomorphology, and so on).

Landscape, in accordance with the concepts of physical geography, must be taken to mean a homogeneous section of the land surface, encompassed by natural boundaries within which the natural components (geological structure, hydrogeological peculiarities, relief, climate, surface water, soil, vegetation, and the area's pattern of habitation) form an interconnected and interconditioned unity characterized by definite regional and zonal peculiarities, in conformity with which region constituting the lithosphere originates and develops in all of its manifestations.

As permafrost is also a component part of the landscape, its origin is likewise influenced by regional and zonal factors.

The regional factors are the geological structure of the region, its tectonic and geomorphological peculiarities, the nature of the hydrogeological structures, the neotectonic movements, and the peculiarities of development of the region during the Quaternary.

The zonal factors are in turn dictated by the latitude, the relation between the continents and oceans, on the one hand, and the peculiarities of the surface radiation-heat balance, on the other, the combined effect of which determines the thermal condition of our entire planet.

It is obvious that in different areas where a uniform type of landscape prevails and like regional and zonal conditions are noted, the same types of permafrost will originate and these will have identical or closely similar characteristics.

From this it follows that the basic procedure in a geocryological survey consists of identifying the types of landscape encountered in the area being investigated, determining their geographic range and limitations and obtaining descriptions of both the mechanisms of permafrost formation and of the permafrost conditions as a whole within each of the landscapes.

Accordingly, both before departing for the field, using as a basis the interpretation of air photo survey data, and later under field conditions, a general geological-geographic study of the region will be made and the landscapes or terrain types that are characterized by specific set of natural conditions will be distinguished. Then within each landscape, key or reference areas will be distinguished, which will provide for a detailed study of particular laws, that is, the effect of individual factors on the origin of the permafrost. Also, the cryogenous processes and phenomena will be identified and quantified.

The key areas are analogs of the reference sections and outcrops used in a geological survey. They must characterize the planning conditions for each type of landscape.

In these areas, as has already been pointed out, a thorough study has been made of the permafrost conditions and the laws governing their formation, both as a function of each factor of the geological-geographic environment and of their combined manifestation in the form of general laws. In particular, studies have been made of the laws giving rise to the heat exchange at the Earth's surface and to the temperature regime of seasonally frozen and seasonally thawed layers, the composition and texture of the latter, cryogenic structures, the moisture regime, and also the cryogenic processes and phenomena associated with them. A qualitative and quantitative evaluation is being made of the effect of the various elements of the set of natural conditions on the formation of these characteristics. In the key areas a study is being made of the laws governing the formation of permafrost with primary consideration being given to its peculiarities in relation to (a) the geological structures, the stratigraphy, lithological-facies classification, and geological-genetic typing of the frozen deposits; and (b) the tectonic characteristics, nature of the hydrogeological structures, and geomorphology and character of the Quaternary deposits. Proceeding from this, in each key area a determination is made of the genetic types of the permafrost (syngenetic, polygenetic, epigenetic), its cryogenous structures, the overall iciness and the properties of the soils and rocks in the frozen and thawed state, and also of the variation of their properties during freezing and thawing. Special research is under way that relates to a study of the temperature regime and thicknesses of the permafrost, and of the time-related dynamics of these characteristics considered in connection with the general course of the region's geological history. The cryogenetic age of the permafrost is being estimated.

To this end, in the key areas exploratory boreholes are being drilled and fully sampled and a series of geophysical studies are being performed, as are electrical profiling and sounding, well-logging, and geothermal research. Detailed studies are being made of the composition, structure, and texture of the seasonally and perennially frozen ground and of the hydrogeological characteristics of the region.

The specific nature of the geomorphological and landscape studies associated with geocryological surveying, and also the requirements imposed on the choice of the key areas, are reflected in the existing instructions and procedures and in a large number of articles and reports describing these studies.[2,22,68,69,76,176, 101-105,112,189]

One of the principal survey methods must be considered to be the formulation of stationary and random geothermal observations, using all of the accessible mineral workings, inasmuch as stationary geothermal studies make it possible to obtain additional information characterizing the permafrost conditions of the region. In geothermal studies in deep wells, it is necessary to take a core from typical horizons and make in profile subsequent determinations of the thermal conductivity of the rocks. Ultimately, this makes it possible to estimate the magnitude of the heat flux from the Earth's interior.

In geocryological surveying increasingly extensive use is being made of modern research techniques. Foremost among these is aerial photographic surveying,[154] which enables us to isolate many of the peculiarities of the geocryological structure of the areas, for in many cases even ground ice can be mapped, having only air photo survey data as the source material. In recent decades, both spectrozonal and color air photo surveys have been more and more widely used in geocryology, their applicability in principle having been domonstrated sometime ago.[154] Action has been taken to initiate the use of centimeter-wave-band radar for geocryological surveys.[16] The time is not far off when we shall be using data obtained from Earth satellites; such data are already being widely used in allied sciences.[21,84,123]

When measuring ground temperatures, heat fluxes and the thermal constants of earth materials in geocryology increasingly wide use is being made of semiconducting sensors and both automatic and semiautomatic recorders.[39,40,51, 52,135,159]

In geocryological surveying, extensive use is being made of various geophysical methods that permit determination of certain parameters of the permafrost region. A separate paper describing these methods will be presented by B. N. Dostovalov, A. T. Akimov, and V. S. Yakupov. On the whole, in evaluating these methods it must be acknowledged that they are still in the nature of auxiliary techniques and frequently require confirmation by drilling and driving operations.

Actinometric and gradient observations have been widely applied in geocryological surveying. Observational data on the radiation heat balance for each type of landscape or key area are making it possible to correct the relations between the various factors and clarify the most general laws of development of the permafrost in a given region.[33,34,43,132-134,195,207,208]

In recent years, at the Department of Permafrost Studies at Moscow State University, in connection with geocryological surveying, increasing attention has been directed to the study of the interaction between the permafrost zone and groundwater (both suprapermafrost and sub- and intrapermafrost water). It is largely the hydrogeological situation that determines the temperature regime, the distribution and mode of occurrence of the permafrost and also its cryogenous structures, iciness, and other properties. As a rule, the genesis, distributuion, occurrence, and types of taliks that are present are also determined by the thermal interaction of the groundwater and the permafrost. The nature of this interaction is both qualitatively and quantitatively investigated during the survey.

In contrast to normal conditions, permafrost gives rise to both lithological and cryogenic water-impervious beds that, in combination with water-bearing complexes, form specific hydrogeological structures. The latter are characterized by a high mobility of the boundaries of the water-bearing complexes and the water-impervious beds, which is due to the dynamics of the permafrost and its thermal interaction with the groundwater. It is this that determines the specifics of the positioning of the groundwater feed and discharge areas; their percolation paths, regime, and dynamics; and also their chemistry, reserves, and resources.

Obviously, where permafrost is present problems of hydrogeology cannot be solved without making a geocryological study of the region and also a geocryological survey. Accordingly, in some cases the latter type of survey is performed as an interdisciplinary permafrost-hydrogeological survey.

The main subject of research during the geocryological survey is the matter itself, i.e., the permafrost. A study is being made of the processes leading to its formation; its genesis, cryogenous structure, and texture; and the relation of the latter to the overall course of the region's geological history during the Quaternary. Recently a new branch of geocryology known as cryolithology has originated in the USSR. It discusses the peculiarities of lithogenesis under the severe northern conditions obtaining in the permafrost region. The theoretical principles worked out by this school, headed by A. I. Popov,[48,146] are basic when performing the geocryological survey. In the light of this new field of study, petrographic methods of investigation are assuming increasing importance.

In the case of epigenetic permafrost, the standard cryogenous structures have been defined. A description has been given of the country rock in relation to the stratigraphy, genesis, and lithological composition of the permafrost for the various permafrost-temperature zones. The general laws governing the positioning of lenses of segregated ice in relation to the nature of the freezing, both for periodic harmonic fluctuations of the heat exchange and for the constant temperature at the Earth's surface have been ascertained. A physical-mathematical theory of the manner in which ice lenses form during the freezing of thick layers of unconsolidated sediments associated with perennial temperature fluctuations at the Earth's surface has been worked out. A study has been made of the petrographic criteria that distinguish epigenetic ice from syngenetic and other genetic varieties. A procedure has been developed for making petrographic studies of permafrost and, in particular, a special frost chamber has been built to house a polarization microscope that enables petrographic studies to be conducted at room temperature and also in summer in the field.

Much has been done with regard to the study of syngenetic permafrost. Based on B. N. Dostovalov's theory of the laws governing the formation of frost fissures in rocks undergoing freezing, the theoretical principles of the formation of syngenetic and epigenetic wedge ice have been worked out. The petrographic-structural peculiarities of wedge ice have been defined for both lateral and frontal growth of the wedges. The relation between the peculiarities of formation of the thinly laminated ice structures of syngenetic permafrost in the massifs between ice wedges has been clarified, as has the relation of these structures to the freezing of the layers of seasonal thawing from below and to the annual heat storage zones at the foot of the seasonally thawed layer.

Based on the regional study and the permafrost survey in the various permafrost-temperature zones, the latitudinal zonation with respect to both epigenetic and syngenetic cryogenic structures has been determined.

In the permafrost region the engineering-geological conditions are to a large extent dictated by the geocryological. In the engineering geology context the nature of the geological structure of the region is materially supplemented by the distribution and occurrence of the permafrost. A description of the geological-genetic types of rocks does not give a complete picture of the region's geological engineering conditions unless an adequate study has been made of the enumerated types of permafrost, the conditions leading to its formation, and also the cryogenic structures.

The character of the iciness of the rocks and the laws governing their distribution and occurrence to a large extent determine the physical, physical-mechanical, and rheological properties of the rocks. The study of the variation of these properties during the freezing of thawed and the thawing of frozen rocks is coming to be of vital importance. The cryogenic processes and phenomena associated with the processses of freezing and thawing of rocks are of great significance in geological engineering. Heaving, frost fracturing, icings, thermokarst, solifluction, landslides, slumps, and cave-ins, the nivation processes of thermal erosion and thermal abrasion, etc., to a large extent determine the geological engineering conditions of the region being investigated. Here it is important to note that the permafrost conditions are wholly determined by the latitudinal zonation and the altitude belts constituting this region. Thus, for example, each of the permafrost-temperature zones has its own specific character as regards the distribution and occurrence of the permafrost, the composition of the cryogenic structures and properties of the permafrost, and also the cryogenic processes and phenomena.

It is therefore obvious that in the permafrost region the geological engineering conditions are intimately connected with the geocryological conditions. It is for this reason that a geocryological survey often takes the form of an interdisciplinary permafrost engineering-geological survey (an engineering-geocryological survey).

A major achievement in the field of geocryological surveying is that in recent years the initial proposals have been submitted with respect to standardizing the quality of geocryological surveys, even though this is not yet mandatory, as geocryological surveying is outside the scope of the State survey series. Research by the Department of Permafrost studies at Moscow State University has shown that "maps can be regarded as meeting the standard if the patterns reflected on them correspond to the contemporary level of knowledge with respect to this question and are adequately corroborated by factual information and if their graphical representation corresponds to the scale of the map." Then again, "the question of the standard to be met by the survey and the amount of factual information it requires must be solved in each specific case" (pp. 25, 26).[102]

Thus, the geocryological survey is the necessary basis for solving any of the problems pertaining to the industrial development of an area in the permafrost region.

An equally important aspect of the task of effecting a geocryological survey is elaborating the general theoretical principles of geocryology (permafrost studies). Only under field conditions is it possible to obtain a knowledge of the general and special laws governing the formation and development of the seasonally and perennially frozen earth materials, and only under natural conditions is it possible to verify in practice the theoretical synthesis worked out during the course of the survey. The interdisciplinary engineering geocryological and permafrost-hydrogeological survey of key areas has become indispensable to the regional study of geocryological, engineering geological, and hydrogeological conditions in the case of large areas. Only on the basis of such an interdisciplinary complex survey is it possible to talk about compiling composite maps of the seasonal freezing and thawing of rocks, general geocryological maps, and also geological engineering and hydrogeological maps for individual regions, the USSR as a whole, other countries, and the entire world.

GEOCRYOLOGICAL MAPPING AND GEOCRYOLOGICAL ZONING

The need for geocryological maps is now generally recognized. Since small- and medium-scale geocryological maps are a synthesis of our knowledge of the development of the permafrost zone within the confines of large regions or even over the Earth's surface as a whole, it follows that they make it possible to depict very clearly the morphology of the permafrost zone and of the permafrost, frozen, and cooled rocks constituting it; to reveal the patterns and also the various types of anomalies in the structure of the permafrost zone; to link the morphology of the permafrost zone with the physical-geographic and geological conditions of its development in a particular area; and to give a clearer representation of the nature of the heat-exchange processes leading to deep cooling and freezing of the upper horizons of the lithosphere. Thus, small- and medium-scale geocryological maps are of great theoretical importance and at the same time are needed, above all, by the geocryologists themselves. They are also needed by researchers studying the physics of the Earth, the propagation of radio waves and earth currents, gravitational and magnetic field anomalies, as well as by those engaged in microseismic zoning, etc. In this case the geocryological map makes it possible to gain an in-depth understanding of the laws governing the devleopment of the Earth's cryosphere.

Medium- and large-scale maps are needed by practical workers, since without such maps the efficient solution of economic problems pertaining to the placement of productive forces, industrial complexes and centers, and also of areas for agricultural development over the greater part of Siberia and the Far East would be unthinkable.

Geocryological mapping is inseparable from the entire history of development of geocryology in our country, so much that the first geocryological maps and plans were issued at the beginning of this century, chiefly with respect to regions in southern and central Trans-Baikal, where in those years major survey operating were unfolding in connection with the construction of the Trans-Siberian Railroad. However, geocryological mapping and geocryological zoning have been most extensively applied in the commission and committees on permafrost and more recently at the V. A. Obruchev Permafrost Institute.

By the late 1930's, Baranov[5] and Tumel'[177] had dealt quite fully with the question of the content and methodology of compiling geocryological maps. Pushed to the forefront was "the pattern of distribution of the frozen earth materials which are to be indicated by conventional symbols. When precise information is available, individual islands and lenses of permafrost or the taliks within continuous permafrost are to be outlined on the map" (Baranov, page 109).[5] Geocryological mapping progressed in those years at a fairly rapid rate, so much so that by 1950 some 239 geocryological maps and plans had been compiled throughout the country.[57]

In recent years the content and methodology of compiling geocryological maps have been discussed and refined on numerous occasions.[3,6,7,24,23,29-32,47,62,74,75,80,81,86,87,89,99,101,121,163,165,166,174] The maps themselves are now being compiled by many organizations of various departments.

Composite maps of the deductive and hypothetical types are being compiled by the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences (Yakutsk); the Department of Permafrost Studies of the Northeastern Integrated Research Institute of the Far Eastern Scientific Center, USSR Academy of Sciences (Magadan); an analogous department of the Production and Scientific Research Institute for Engineering Surveys in Construction, USSR Gosstroy (Moscow); and the Department of Permafrost Studies and the Department of Cryolithology and Glaciology, Moscow State University, which are sponsoring major study programs in various parts of the Arctic and sub-Arctic. Individual projects in the field of geocryological mapping are being undertaken by the Institute of the Earth's Crust, Siberian Division of the USSR Academy of Sciences (Irkutsk), and the All-Union Scientific Research Institute of Engineering Geology and Hydrogeology (VSEGINGEO), USSR Ministry of Geology (Moscow).

The vast majority of the maps being compiled are published in the form of text maps and inserts in books issued mainly by "Nauka," "Nedra," and Moscow University presses, and it is only the units of the geocryological maps that are included in regional atlases that go through the map plant of the Main Administration for Geodesy and Cartography.

It is significant that there are now several areas in which practical use can be made of geocryological maps--in dredge mining of placers, in the solution of water supply problems, in major construction or the construction of highways and railroads, and in the construction of various linear installations such as gas pipelines, overhead electric power lines, and so on.

On the whole the geocryological mapping of the area constituting the USSR is progressing very satisfactorily. In recent years several small-scale maps of the geocryological zoning of the USSR have been published that form an important supplement to the geocryological map of the USSR (1959) compiled by Baranov,[6] and also to the large series of maps for individual regions and areas of our country.

Thus, A. I. Popov,[143] proceeding from the tenet that "ground ice is the basic genetic feature of permafrost" (p. 59), constructed a "Map of the Underground Glaciation (permafrost) for the European Part of the USSR and Siberia." Both black and white[144] and color versions[143] were produced. On this map (one of the varieties of the geocryological zoning maps) the genetic types of the ground ice in the upper layers of the Earth's crust are indicated and the boundaries of the temperature zones are drawn (after Tumel').

A map of the geocryological zoning of the USSR to a scale of 1:50,000,000 was produced at Moscow State University by V. A. Kudryatsev and K. A. Kondrat'yeva.[56] A zone of individual islands of permafrost measuring up to 100 m thick and several zones of permafrost from 100 to more than 500 m thick are portrayed on it.

Another map of the geocryological zoning of the USSR, drawn to a scale of 1:20,000,000, was produced at the Institute of Permafrost Studies, Siberian Division, USSR Academy of Sciences, by I. A. Nekrasov.[126] On the map three types of permafrost regions are distinguished: subaereal, subaqueous, and interstitial. Within the subaereal type the following subtypes are indicated: the continuous, discontinuous and insular permafrost zones. The basic parameters of these types and subtypes, the permafrost zones, and their areal extent in the USSR are presented in the legend.

Based on roughly the same principles numerous maps have been produced and published in recent years, together with accompanying monographs and generalizing studies on the geocryological conditions of the various regions of our country. Such maps, for example, have been compiled for the Pechora Coal Basin,[37] for the West Siberian lowland,[14,17,106,211,212] for the Yenisey North,[202] for central Siberia,[50] for Yakutiya as a whole and for its individual regions,[46,78,79,118,119,129,157,185,190] for Cis-Baikal and Trans-Baikal,[26,73,109,110,130,131,167-169] for the Far East,[20,72,215] for the northeastern USSR,[67,197,216] and for Tyan'-Shan'.[41,42] (Gorbunov, 1967, 1970).

Mention must be made of the map of the geocryological zoning of the Verkhoyansk-Kolyma mountainous area,[197] as it incorporates the initial attempt to isolate the geocryological formations previously proposed[196] as the basis for geocryological zoning. On the whole, the principles of geocryological zoning have been reflected in many papers.[38,53,110,116]

The geocryological map for the continent of Asia (to a scale of 1:50,000,000) was compiled by N. A. Marinov,[114] and the map of the entire northern hemisphere (to a scale of 1:30,000,000) by I. A. Nekrasov.[127]

At the present time work is being completed on the compilation of two geocryological maps covering the whole of the USSR to a scale of 1:5,000,000. This is being done at the Industrial and Scientific Research Institute for Engineering Surveys in Construction[10] and at the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences.[166] The geocryologists of the Faculty of Geology at Moscow State University are preparing for the compilation of a geocryological map of the USSR to a scale of 1:2,500,000.[31]

Many special geocryological maps have been published for various areas. For example, G. S. Gravis and I. V. Klimovskiy[44] compiled a permafrost-lithologic map for one of the regions of Trans-Baikal; P. A. Solov'yev[164] represented on a map the thickness zonation of the seasonally thawing layer for western and central Yakutia; I. A. Nekrasov[128] proposed a map of the taliks of the Anadyr' river basin; A. I. Popov[147] compiled a map of the cryogenic rocks of western Siberia; V. V. Rudavin[158] represented on his map the various cryogenic processes and formations; B. A. Vtyurina[28] compiled a map of the seasonally thawing layer of central Siberia; the types of cryogenic series and the cryogenic structure of the permafrost zone in central Siberia are represented on maps compiled by S. M. Fotiyev[186] and N. S. Danilova;[49] a map of the zonation of the mean annual rock temperatures for the Yana-Indigirka interfluve was compiled by K. A. Kondrat'yeva *et al.*;[190] the zoning of the thermokarst land forms was the subject of a map by Yu. T. Uvarkin *et al.*[183]

Let us remember that it was V. A. Kidryavtsev[92] who initiated the production of specialized geocryological maps in that he compiled a series of maps of the temperature zoning of the permafrost region in the USSR.

It would appear that in the immediate short-term geocryological mapping will be undertaken, as now, by many departments and institutions. However, the necessity for an integrated approach, even with specialized surveys, is promoting an awareness of the need for treating as a first-priority problem the planned, methodical, and programmed coordination of the geocryological mapping of the USSR.

This problem has been repeatedly discussed at the All-Union Conferences on Geocryology and at more specialized conferences and seminars, and a start has already been made on its practical realization.

GEOCRYOLOGICAL FORECASTING

When making a geocryological and especially a geocryological-engineering evaluation of an area for construction purposes, the deciding factor is forecasting variations in the geocryological conditions. The planning, construction, and use of installations in the permafrost region are only possible on the basis of data from geocryological forecasting, that is, by taking into consideration the geocryological conditions that will originate as a result of the industrial development of the area, rather than those which exist at the time when the geocryological survey is conducted. The geocryological forecast is predicted on a study of the specific and general patterns of formation of the permafrost and its properties and also of the cryogenic processes and phenomena. In the forecast, mention must be made of the expected variations in the geocryological conditions. There must be not only a qualitative, but also a quantitative, estimate of the variation, for example, of the temperature regime, the depth of seasonal freezing and thawing, the formation of taliks and their specific dimensions, the variation in the areal extent of the permafrost, and its degradation and aggradation.

The geocryological forecast also includes forecasting the variations in the cryogenic structures and the settling of the ground during thawing, as well as the variations in its physical-mechanical properties as a function of the temperature variation, and so on. The study of the qualitative and, more particularly, the quantitative relations not only affords an opportunity to make known the predicted engineering geocryological assessments with respect to individual areas and the region as a whole, but also to determine the principles and methods to be adopted with a view to controlling the geocryological processes in the interests of optimal construction.

As with the geocryological survey, so also is the geocryological forecast the brainchild of Soviet geocryology. The first attempts at devising a geocryological forecast are to be found in the "degradation theory" of M. I. Sumgin.[171-173] They were subsequently developed by V. A. Kudryavtsev,[90,91,95,97] Kudryavtsev *et al.*,[103] G. V. Porkhayev and V. K. Shchelokov,[149-152] Shvetsov,[198] and many other researchers. At the present time the various aspects of geocryological forecasting are being successfully worked out at Moscow State University, at the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences, at a number of the institutes of the USSR Gosstroy (Industrial and Scientific Research Institute for Engineering Surveys in Construction, Research Institute for Foundations and Underground Structures, etc.), in the survey trusts of the RSFSR Gosstroy (TsTISIZ,* etc.), as well as at the All-Union Research Institute of Hydrology and Engineering Geology. . . . of the USSR Ministry of Geology.

The problems of geocryological forecasting have not yet been completely resolved, and much remains to be done in this area.

Geocryological forecasting can be global or partial. *Global* means forecasting changes in the geocryological situation over the Earth as a whole, which are associated with fluctuations of the heat-balance components under the influence of long-term (measured in thousands of years) fluctuations of the heat exchange in the lithosphere-atmosphere-cosmos system. *Partial* means forecasting changes in the geocryological conditions within individual regions and areas. These changes are associated with variations in the natural situation arising from natural variations of the heat balance of the landscape mantle (short-term variations of the heat exchange) and from disturbances of this equilibrium due to human activity. The latter field of engineering is currently attracting increasing attention on the part of researchers.

In the field of global forecasting, the contemporary level of research is such that only very general conclusions and suppositions are possible. This forecasting would be impossible if it were not for paleogeocryological reconstructions, in the course of which, first and foremost, the history of the permafrost zone is reconstructed and, on the basis of this, trends are noted in the short-term development of the permafrost zone. M. I. Sumgin[171-173] was the first to do such forecasting in geocryology. By adhering to the concept that "permafrost is the product of the glacial period," he demonstrated that the present era is a time of widespread permafrost degradation.

Similar syntheses were embarked upon by other researchers[20,45,60,71,92,141,145,170,179,201,203,212] who are united by the procedure of doing paleogeocryological reconstructions based primarily on logical synthesis, from time to time corroborated by palynological or cryolithological observations.

With a view to using them in paleogeocryology and global forecasting, a second group of researchers have tried to apply computer techniques and modeling, taking as their starting point the rhythms exhibited in the development of the natural setting. Such syntheses were made for western Siberia by V. V. Baulin[11-13] and A. A. Sharbatyan,[191-193] who use the modeling technique to explain the origin of the second, buried layer of permafrost. For the mountains of the Stanoyoye highlands, similar calculations were performed by Yu. G. Shastkevich[194] and later by I. A. Nekrasov *et al.*[128]

The principles of the partial geocryological forecast have been worked out in greater detail and discussed in the literature. This type of forecast[4,18,35,93,97,103,149-152,187] is being developed in three basic areas:

(a) The variations in the geocryological situation, which are caused by the construction and use of separate structures and artificial coverings occupying small areas are calculated;

(b) The variations in the geocryological situation, which are dictated by the inherent dynamics of natural factors are calculated;

(c) The variations in the geocryological situation originating under the combined effect of a group (complex) of installations, on the one hand, and of the factors that lead to environmental changes caused by the economic development of an area on the other (drainage, alteration of the natural surface, alteration of the plant associations, pollution of the snow cover and air space, etc.) are determined.

At the present time, attempts are being made in the forecasts to take into account short-term variations of the air and ground temperatures. Thus, for example, analysis of meteorological data indicates that throughout Siberia as a whole the 2- and 3-yr half-periods of warming, when the amplitude of the mean annual air temperature variations reaches 3°, are clearly marked. Similar temperature fluctuations are also noted for the 11 and 33 year periods of variation in solar activity.

The research performed by P. F. Shvetsov and I. V. Zaporozhtseva[199] and A. A. Konovalov[83] demonstrated that the 3-yr rise in temperature of the surface layer of the ground is being felt, for example, in the European North to a depth of 13-15 m. At a depth of 5 m the ground temperature varies to the extent of $\pm 0.5°$, while seasonal thawing during the warm periods is almost twice what it is during the cold periods.

The principles of the forecast associated with the economic development of an area have been worked out in much greater detail, as deep disturbances of the thermal equilibrium of the natural grouping are noted here. The majority of the changes occurring here in the heat storage cycles and the moisture exchange of the covering layer of the ground display an irreversible character. Numerous examples indicate[54,115,155,162,188] that the economic development of an area can lead both to degradation and to development of the permafrost zone (a decrease in temperature, an increase in thickness, and so on) at one and the same place.

The basis for the geocryological forecast in this case is the analysis of the variations of the components of the radiation heat balance of

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*Expansion unknown. Possibly Central Trust for Engineering Surveys in Construction.

the underlying surface occurring both as a result of the natural short-term changes in the natural processes and as a result of the development of the area. In the latter case the following variations are normally analyzed: (a) the total reflected radiation occurring as a result of distrubance of the natural surface albedo; (b) the amount of heat used for evaporation, both on account of drainage or flooding of the terrain and as a result of a decrease or increase in the total incoming radiation; (c) the amount of heat used for direct heating of the ground; (d) the magnitude of the effective radiation.

The heat-balance observations indicate that in the Trans-Baikal and in the mountains and depressions of the Stanovoye highlands more than a third of the heat reaching the Earth's surface in the summer is spent on evaporation, the forest canopy being penetrated by 2.5 times less than the open spaces.[34,195] Proceeding from this, it can be assumed that in the southern regions of Siberia the cutting of the forest and the drainage of the ground that accompany development will inevitably lead to a rise in the ground temperature. For example, in the Chara depression, in the areas of well-drained ground all that was necessary for the permafrost to completely thaw within a short span of time was to eliminate forest planting,[34] and in the vicinity of Chita all that was necessary for islands of permafrost to begin to form in one case[54] or for the permafrost to begin to thaw in another was removal of the snow from the site.[115] Destruction of the soil cover and of peaty soil also leads to similar results. As A. P. Tyrtikov[181] notes, in this case, in the sparse forests of Siberia, as a rule the soil temperature at a depth of 20 m rises during the growing period by 6°-8° as compared with natural conditions. The depth of the summer ground thaw in this case is 2.0-2.5 times greater.

Pollution of the snow in the vicinity of construction sites also leads to an increase in the temperature of the ground and hence to earlier disappearance of the snow. Experience shows that removal of the snow cover in May and June when negative air temperatures prevail increases thawing even at latitudes of about 80°.[199] Today, many areas of the permafrost region are covered by observations entailing the deduction of quantitative relations. This makes it possible to forecast an alteration in the geocryological situation with fairly high accuracy. In this case to begin with the forecast amounts mainly to a forecast of general variations of the ground temperature regime, both over the whole of the area to be developed and at the base of individual buildings and installations. Moreover, it is precisely the ground temperatures that determine its state (frozen, thawed, and so on), and hence bearing capacity. Today, many of the problems of this type of forecasting have already been resolved. The practical extensions are being used in the national economy for the planning and design of buildings and settlements,[148] in the building of railroads,[59,60] and in mineral prospecting.[200] Special maps forecasting variations in the geocryological conditions are being compiled.[15,18,61] The procedures for simulating these problems on electronic and analog computers are being successfully worked out.[108,115,124,187,188]

Over a 20-yr period the calculation methods used in geocryological forecasting have been developed at the Department of Permafrost Studies of the Faculty of Geology at the Moscow State University. Based on these solutions of the Stefan problem, mathematical systems have been worked out both for a general geocryological forecast connected with the normal course of development of natural conditions, for a specific forecast connected with human industrial activity, and in particular with major hydraulic engineering construction projects (the Bratsk, Ust'-Ilim, Nizhne-ob', and Vilyuyskaya Hydroelectric Power Plants), railroad construction, and the like. A series of approximation formulas have been worked out that make it possible for a geocryological forecast to be compiled directly during a permafrost survey in the field and also for the forecasting data to be corrected in the course of the planning and design work. Finally, the geocryological forecast is refined when the office work is being done. At the Department of Permafrost Studies, algorithms for computer calculations have been developed for various problems of geocryological forecasting (see the collection "Merzlotnyye Issledovaniya" [Permafrost Studies], Moscow State University, 1961-1972).[120]

REPORTS AND PAPERS PRESENTED AT THE SECOND INTERNATIONAL CONFERENCE ON PERMAFROST

Eight papers and 11 reports published in the Volume 6 of the proceedings of the Conference were selected for investigation in the session entitled "Geocryological Surveying and Forecasting."

Paradoxically, not a single paper was devoted to the survey as such--the standards it must meet, its stable requirements, systematic character, and so on. Only in the paper by G. A. Golodkovskiy *et al.* and the report by A. B. Chizhov are certain aspects of geocryological research discussed.

The first paper, entitled "The Specific Character of Permafrost-engineering-geological Studies Associated with Prospecting for Mineral Deposits" discusses experience gained at the Department of Engineering Geology of Moscow State University, which for a number of years has been engaged in engineering-geological studies of the deposits of the Noril'sk district. The authors emphasized the need for making a geological-structural analysis when isolating the engineering-geological rock massifs and studying distribution patterns of frozen and thawed rocks and their temperature regimes. At the same time, they point out that the problem of doing engineering-geological research on deposits consists of studying all of the natural factors determining the interaction of the installations with the geological environment.

A list of problems is presented which must be incorporated in a geological engineering research program. The role of permafrost as the most important factor determining the geological engineering and hydrogeological conditions of the deposits is emphasized. It is recommended that the permafrost-engineering-geological studies be carried out simultaneously and that they be integrated with the exploration of the deposits, which will lead to a substantial lowering of costs and to an improvement in the quality of the data obtained during the field investigations. It is interesting that the authors point out that the information derived from the integrated permafrost-engineering-geological studies will in turn permit the mode of occurrence of the ore bodies to be predicted and will provide supplementary information which will aid in solving problems pertaining to the genesis of the deposits.

Some aspects of the formation of water-reserve taliks and of the temperature regime in ascending flows of groundwater are discussed in a report by A. B. Chizhov entitled "Problems of Investigating Permafrost and Groundwater as a Complex Self-Regulating System" is presented for discussion. It is suggested that natural systems, including the solid phase of water, should be called cryosystems. The high level of organization of cryosystems makes it necessary to study them in three interrelated aspects: structural, functional, and historical. In order to resolve this problem, in the author's opinion it is necessary to combine traditional geological methods with mathematical simulation. As an example, the author considers a number of questions pertaining to the formation of the pressure-filtration taliks under various zonal conditions and their dynamics.

The majority of the papers and reports (those of Yu. A. Avetikyan and A. T. Akimov; A. N. Bogolyubov, V. I. Dzhurik, F. N. Leshchikov, V. V. Bogorodskiy *et al.*; O. K. Voronkov and G. V. Mikhaylovskiy; Yu. D. Zykin and Yu. I. Baulin; V. P. Mel'nikov) are devoted to examining new methods of investigating various parameters of the permafrost region, chiefly by geophysical methods.

It will be recalled that on the whole the problems of using geophysical methods in geocryological surveying are dealt with in a paper by A. T. Akimov, B. N. Dostovalov and V. S. Yakoupov.

In some of the reports presented at the session, geophysical techniques are recommended as methods for the mapping and zoning of the permafrost region (the reports by Yu. K. Agafonov *et al.*, L. G. Tsibulin, A. G. Krasovskiy, and N. A. Irbe). The principles of compiling small-scale geocryological maps are discussed in greater detail in a paper by A. V. Gavrilov and K. A. Kondrat'yeva and also in a report by B. I. Vtyurin.

The report by Gavrilov and Kondrat'yeva discusses the principles of compiling small-scale maps. The production of the small-scale geocryological maps of the USSR and its individual areas is one of the urgent tasks of modern geocryology. Currently a number of scientific research organizations and individual researchers are working in this field. Therefore, the discussion of this problem at an international conference is undoubtedly of value.

Obviously, the authors are right in insisting that the fundamental principle of mapping and explaining the laws governing the formation and development of permafrost must be through the representation of its basic properties, these properties being the resultant of heat-exchange processes occurring under specific geological-geographic conditions.

The suggestion about compiling two interrelated maps representing permafrost and seasonally freezing and thawing ground is deserving of attention. Represented on the permafrost map are the types of permafrost that were tentatively distinguished on the basis of similarity of structure and basic characteristics and commonness of historical development, and on the maps of the seasonal freezing and thawing, the types of seasonally thawing and seasonally freezing rocks as defined by V. A. Kudryavtsev. In this connection a great deal of attention is being devoted to mapping the natural conditions of permafrost formation.

It should be noted that this mapping technique has been used for a number of years at the permafrost department for compiling geocryological maps of various parts of Yakutia and is being used successfully at the present time for compiling a permafrost map of the Soviet Union on a scale of 1:2,500,000.

The report by B. I. Vtyurin is devoted to the principles of mapping permafrost from the standpoint of its cryogenic structure. The author points out the advantages to be gained from mapping permafrost of varying composition, in that this reflects the "order of bedding of horizons with differing structure." The basis for the mapping is the classification developed by the author by which all of the permafrost series are divided according to the nature of their composition into monogenetic and polygenetic types, and, according to their structural characteristics, into epigenetic and syngenetic types. In the Vtyurin's opinion, bed-forming ice only complicates the cryogenic structure, which is determined by the nature of the ground ice inclusions.

A new approach to engineering-geocryological mapping was proposed in a report by V. P. Solonenko, who suggested the principle of a mandatory study being required of the behavioral characteristics of structures erected on a frozen foundation in seismic zones, this study to be made in the course of the survey. He proposed a new area of research, to be known as engineering seismocryology, and outlined the problems facing the geocryologist.

Three papers (V. A. Kudryavtsev *et al.*, P. F. Shvetsov and N. G. Bobov, and N. Ye. Zarubin and O. V. Pavlov) discuss certain aspects of geocryological forecasting in areas undergoing development where there is permafrost.

The paper by V. A. Kudryavtsev *et al.* discusses a wide range of problems associated with the procedure for compiling a permafrost forecast. The basis of any permafrost forecast is the permafrost survey, in the course of which a study is made of the mechanisms by which geological-geographic factors influence the formation and dynamics of the seasonally freezing ground and permafrost. The two-sided interrelations that have been found to obtain betwen the components of the natural environment and the permafrost characteristics are used directly in design diagrams.

It is proposed that special attention should be given to the structure of the radiation balance and its variation with the economic development of an area. This is because of the fact that an alteration in the permafrost conditions is directly associated with an alteration in the radiation-heat balance.

The paper goes on to discuss the basic problems involved in permafrost forecasting and methods of solving them. In this connection, extensive use is made of the experience accumulated in the Department of Permafrost Studies. An analysis is made of the peculiarities of the permafrost forecast in the various stages of the investigation in connection with the scale of the permafrost survey and the permafrost-engineering-geological survey.

In a paper by P. F. Shvetsov and N. G. Bobov entitled "Scientific and Procedural Principles of Forecasting the Engineering-Geocryological Phenomena in Areas Undergoing Development" it is pointed out that the key factor in this forecast is the geocryological survey. The authors present a classification system for the main types of influence exerted by industrial activity on the geocryological conditions. Here, the geothermal type of effect is taken as the basis for the classification. It is suggested that a distinction should be made between the direct and indirect effects of engineering works, as expressed by their density.

The report includes a map-diagram of the breakdown of the permafrost region and indicates the possible variations in the permafrost conditions as a result of economic development (the indirect effect). In the view of Shvetsov and Bobov, the complexity of geocryological forecasts is dictated by its random characters.

Besides the above-listed reports, some of the reports and papers discussing problems in geocryological surveying and forecasting were sent to other sessions of the Conference. Among them was a paper by M. A. Minkin devoted to the peculiarities of geocryological surveying along main gas pipeline routes (Vol. 7) and also papers by A. N. Khrustalev and V. M. Gorbachev on the variation of the external heat-exchange components in the lithosphere-atmosphere system during building under the northern conditions; a paper by A. Ya. Kurennaya, again devoted to changes in the radiation-heat balance of various landscapes accompanying the economic development of an area (as exemplified by the Polousnyy ridge and Ukandinskaya depression); and a report by A. A. Sharbatyan, "Formulas for Estimating the Minimum Periods of Aggradation and Degradation of a Permafrost Zone," which formulas are based on a thermophysical analysis of the laws of freezing and thawing the Earth's crust in the simplest sample situations. Previously a somewhat different procedure had been proposed for such calculations.[96]

In conclusion, it is apparent that in spite of the planned development of geocryological research in the USSR, which only began in the mid-1930's, much progress has been made in the field of arranging and implementing this research. Important advances are also evident in the field of geocryological forecasting.

At the same time, it must be noted that many of the problems in surveying and forecasting have been inadequately resolved and that the principles elaborated with respect to geocryological surveying are only slowly taking root in the daily activity of geocryologists.

The following must be designated immediate tasks in the further development of geocryological surveying and forecasting procedures:

1. The history of development of permafrost, considered in relation to the overall geological history of development of the areas being investigated.
2. The cryolithogenesis of perennially frozen sedimentary strata as one of the basic characteristics of geocryological conditions.
3. The use of petrographic research techniques when studying the cryolithogenesis of permafrost.
4. The use of geophysical research techniques such as microseismics and acoustic methods.
5. The use of the entire range of aerial photographic techniques, including infrared, and radar surveying.
6. The field methods of studying the surface radiation heat balance considered in relation to the formation of the geocryological conditions. Providing a more precise definition of the effect of the various components of the natural conditions on the formation of the individual components of the surface radiation heat balance.
7. The improvement and further development of mathematical techniques for determining the extent to which individual factors in the natural environment affect the formation of the thermal regime, the depths of seasonal freezing and thawing of earth materials, the heat cycles of the ground, and other geocryological characteristics.
8. The study of the quantitative characteristics of formation of geocryological processes and phenomena.
9. The working out of unified requirements for the compilation of geocryological maps of various scales; giving a more precise definition of their content, load, and conventional symbols; and the development of a single, generally accepted legend for geocryological maps.
10. Further development of the procedure to be followed when compiling a geocryological forecast as the logical conclusion of the geocryological survey; the compiling of geocryological forecasting manuals and procedures.

The need for scientific coordination of research in the field of geocryological surveying and forecasting on a Union-wide scale is being acutely felt. It is most desirable that this coordination should be achieved under the auspices of a Scientific Council for the Cryology of the Earth, of the USSR Academy of Sciences.

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GEOPHYSICAL METHODS OF STUDYING PERMAFROST

A. T. AKIMOV Production and Scientific Research Institute for Engineering Research in Construction, B. N. DOSTOVALOV Moscow State University, and V. S. YAKUPOV Permafrost Institute of the Siberian Division of the USSR Academy of Sciences

As a promising tool for the study of permafrost, geophysics has attracted attention from the very beginning of the history of geocryology.[44,45] By now a great deal of experience has been accumulated in the study of permafrost by geophysical methods, more particularly electrical and seismic exploration (the study of geothermics will not be considered in this survey). For the solution of some problems, it is also possible to use differences in the density, magnetism and other properties of the rocks, that are caused or intensified by the differing variations in ice content.

We will consider successively the results of studies of the electric and elastic properties of permafrost, certain general problems relating to the methodology of electric and seismic exploration when permafrost conditions prevail, and, finally, the main problems in geocryology that are being solved by geophysical methods.

ELECTRICAL PROSPECTING

The majority of the permafrost studies involve electrical prospecting techniques, foremost among which are direct current methods. Electrical prospecting is now a routine procedure, and, for the solution of a number of problems, it is being widely applied. Among these are mapping the boundaries of the thawed and frozen rocks, determining the thicknesses of frozen soils, and large-scale mapping within the confines of mineral deposits and areas designated for development. The successful application of electrical prospecting has been materially aided by the results of *in situ* vertical electrical sounding studies of the electrical conductivity of permafrost and of the geoelectric section of the permafrost. The initial results led to the notion that permafrost is a single layer of high resistivity, which is primarily a function of the temperature and is almost independent of the lithologic structure of the series.[4,27] This explains why the problem of investigating the structure of permafrost in the vertical section has been discussed at such length[28] and why a number of researchers are continuing to regard it (see, for example, Borovinskiy,[19]) as primarily a problem of formulating a theory of electrical prospecting of layered media with parameters that are a continuous function of the depth of occurrence and of developing the requisite apparatus for interpreting the observational data.

Simultaneously, the search has continued for simplified models of the geoelectric section, on the basis of which it would have been possible by vertical electric sounding curves to obtain a correct estimate of the thickness of the permafrost[5,9] or its individual layers.[52] An effort to construct a model of the section based on the following assumption turned out to be fruitful: A variation in the electrical conductivity of coarse-grained soils during freezing occurs only in the range of significant phase transformations, that is, in the 0°C to 2°C range where the electrical conductivity is a linear function of the temperature, which maximizes its effect.[52] The calculations showed that the effect of the layer with a temperature of between 0°C and -2°C in the upper part of the section can be ignored if its thickness is 10 or more times less than the thickness of the frozen soils with a lower temperature. In the northeastern part of the USSR and eastern Siberia, where the thickness of the soils is more than 10 m, this condition is almost everywhere satisfied throughout the length of the field season. It was this model of the geoelectric section that eventually made it possible to work out the procedure for determining the thickness of frozen soils by the vertical electric sounding method.[53] As a result, studies to this end are being undertaken on a large scale in northeastern USSR, and they have yielded a vast amount of information on the *in situ* electrical conductivity of frozen ground and on the geoelectric section of permafrost. By arranging for vertical electric sounding at outcrops of basement rock in summer and in frozen soils of known thickness in winter, after the seasonally thawing layer had become completely frozen, numerous data were obtained that indicate that with low mineralization of the pore water, in the negative temperature range the electrical conductivity of permafrost of highly variable composition remains constant within the limits of accuracy of the observations. Lithologically uniform permafrost with a temperature that increases with depth is depicted on the vertical electric sounding curves by asymptotic sections ρ_k = const, that is, by electrically-uniform layers.[56,57] The resistivity of the freezing rock is obviously a function of the temperature with increased mineralization of the pore solution even when as a result of a relatively small number of sufficiently large pores or of a low moisture content the rock contains practically no free water. The overall increase in resistivity in these cases is small or nonexistent. The primary cause of the increase in the electrical resistivity of freezing rocks consists in an increase in the sinuousness of the pore channels of the rock: a lengthening and shortening of a number of the current conducting paths, as a result of the formation of ice inclusions. The increase in resistivity when there is a sufficiently high content of free water in the rocks takes place discontinuously at the temperature of formation of the cryogenous structure.[11,18] The magnitude of the discontinuity in the case of rocks is limited from above the resistivity of the rocks during their transition from the thawed to the frozen state usually increases by no more than 10 times, inasmuch as the volume and shape of the ice inclusions forming in them are rigidly limited by the volume and structure of the pore space. The resistivity of soils, depending on their grain-size distribution, the type and degree of development of the originating cryogenous structure, and also the cryogenous composition, increases by tens and hundreds of times. Consequently, such important permafrost characteristics as the type of cryogenous structure and the cryogenous composition are among the principal factors determinng its electrical conductivity. The mechanisms leading to the formation of the cryogenous structure govern the variation in the electrical conductivity of the freezing rock and, in particular, explain what at first glance is the paradoxical conclusion that the electrical conductivity of the permafrost is independent of the temperature within the limits of accuracy of the observations.

A comparative study of the electrical conductivity of a wide variety of earth materials in the thawed and frozen states has made it possible to establish another important principle: that of conservation or, in the case of sedimentary rocks, even an increase in relative differentiation in terms of their electrical conductivity after freezing. This principle emphasizes that, other things being equal, frozen soils of the same composition but of varying cryogenous structure and texture differ appreciably in their electrical conductivity.[57]

Thus, in the general case, permafrost can be represented as the aggregate of layers that are uniform in their electrical conductivity and perhaps also anisotropic, the boundaries of which coincide with the cryolithological boundaries and the boundaries of the permafrost itself. The type and the parameters of the geoelectric section of the permafrost are to a large extent determined by the frozen soils and undergo seasonal variations due to the occurrence, development, and freezing of the seasonally thawing layer. The most widely prevalent types of geoelectric section of the permafrost are the KQ, AKQ, and KQQ in summer and the Q, KQ, QQ in winter--the layer of thawed subpermafrost basement rock being taken as the reference layer.[57]

The normally large variation in electrical conductivity as between thawed and frozen rocks, the high degree of differentiation in the electrical conductivity of permafrost that differs in its lithologic composition or cryogenous structure and texture, and the possibility of retaining the classical approach to problems of interpretation, by virtue of the electrical conductivity of lithologically uniform permafrost (with uniformly distributed iciness) being almost independent of the temperature, make electrical prospecting a powerful tool for investigating the morphology, cryolithological composition, and cryogenous structure of permafrost, and, as a result, a means of evaluating its various physical properties.

At the same time, there are a number of factors that predetermine some degree of decrease in the resolving capacity of direct current and low-frequency current electrical prospecting and in the accuracy with which its data can be quantified. These must be taken into account when estimating the feasibility of postulating and solving specific problems by electrical prospecting methods involving direct and low-frequency currents. Of these factors the following are particularly noteworthy:

(a) The shielding by the thawed (suprapermafrost) layer of the horizontal interfaces in frozen soils;
(b) The substantial shielding effect of the frozen soils. In the majority of cases this makes for difficulties in interpreting the structure of the section, the usual or commonly encountered elements of which are the K and Q elements;
(c) The rapid variation in area of the section parameters and frequently its structure, a consequence of the variability of the properties of frozen soils;
(d) The existence of seasonal variations in the type and parameters of the geoelectric section of the permafrost: here, however, it is important to note that after the seasonally thawed layer has become completely frozen the solution of the problem is often simplified;
(e) The wide range of action of the equivalence principle;
(f) The possibility of layers forming in which the resistance varies with depth as a function of the temperature variation (as the concentration of the pore solution increases) or the iciness.

A decrease in the bounding effect of some of these factors to the known limit is possible by improving the existing direct and low-frequency alternating current techniques and the procedures for interpreting the derived results (the discrete values, technique, the use of recognition programs for determining the type of geoelectric section, and so on). The radically new ways of increasing the resolving capacity of electrical prospecting with respect to permafrost consist of using methods in which the high resistivity layers do not act as a shield. These are primarily methods based on the use of electromagnetic fields of fairly high frequencies.[1,2,10] Unfortunately, the dielectric constant of permafrost remains poorly studied. It is obvious that in the frequency range above 100 kHz the dielectric constant of permafrost decreases after freezing. Correspondingly, the degree of differentiation of rocks with respect to their dielectric constant after freezing must also decrease, approximating to that of dry rock if the concentration of the pore solution is low. A lower frequencies, according to the findings of A. D. Frolov and B. V. Gusev, who studied the electrical properties of sand, kaolin, and mixtures of them in the frozen state under laboratory conditions in the 1-100 kHZ range, frequency and temperature dispersions of the dielectric constant and a loss factor are observed, the dielectric constant of frozen sand at a temperature of -2.2°C, and a reduction in frequency to 1.5 kHz increasing to 110.[48] (Frolov, 1973).

The polarizability of permafrost, which has only recently become an object of investigation, also remains a poorly studied parameter. The initial results afford a basis for conjecture that permafrost of varying lithologic composition has a significantly different polarizability.[33,34,39,40,49]

On the whole, in the area of electrical prospecting of permafrost, a number of general results have been obtained that provide a basis for an expansion of both the scope and the scale of its application and make it possible to distinguish ways of further improving its resolving capacity.

SEISMIC EXPLORATION

Seismic exploration has become relatively widely used during the last 10 yr for solving one of the most important and widely encountered problems--determining the thickness of frozen soils.[12,36] A whole series of laboratory studies have been made of the velocities of elastic vibrations in permafrost. The most interesting of these are *in situ* studies of seismic wave velocities in permafrost by means of seismic logging of wells and vertical shafts, etc., which have yielded the following results. In the 0°-6°C range the stratal velocity of longitudinal waves in frozen loose deposits remains constant within the limits of accuracy of the observation[37,46] and varies discontinuously during transformation from the frozen to the thawed state. In dense rocks (shales, sandstorms, hornblende, granitoids, tuffs, etc.,) the stratal velocity of longitudinal waves is identical at positive and negative temperatures. Jointed rocks constitute an exception.[22] The stratal velocity of longitudinal waves in frozen soils, according to data obtained in the northeastern part of the USSR,[37] varies from 2,900 m/s (clay) to 4,600 m/s (coarsely clastic glacial deposits), its range of variation being greater than for the same deposits in the thawed state. The degree of contrast between the cryolithological varieties of the frozen soils in the seismic sense is frequently intensified by variations in their density. The longitudinal wave velocities in dense rocks range from 4,700-4,800 m/s (cross bedded sandy shale series) to 6,000-6,500 m/s (igneous rocks). The range of variation of longitudinal wave velocities in the frozen soils of the Bol'shaya Zemlya tundra and western Siberia is 600-4,000 m/s.[8,14,25]

More recently, seismoacoustic methods have begun to be used in determining the physical-mechanical properties of permafrost in drillholes.[6-8] Using these techniques it proved possible to determine the dynamic Young's Modulus and the Poisson's ratio of permafrost; their variation with depth, in plan, and in time; to estimate the consistency indexes of permafrost; to study the velocity distribution in the upper layers of the permafrost; and, on the basis of correlations, the strength and deformity characteristics of frozen soils.[30]

According to laboratory studies of samples, both the electrical resistivity of permafrost and the velocity of the elastic vibrations in them vary markedly with temperature.[4,27,47] It is true that some exceptions have been reported.[11,26] Let us also note the satisfactory convergence of the data from electric and seismoacoustic logging with the data from field methods and the wide discrepancy between the results of laboratory and field determinations of the physical properties of permafrost. The discrepancy between the results of studying these physical properties in samples and in large massifs of rock undoubtedly calls for special study. It appears that the reasons for this discrepancy lie in the fact that the vast majority of laboratory studies of the electrical conductivity of permafrost and the velocities of elastic vibrations in them as functions of the temperature are conducted at a high rate of cooling, without taking into account either the stressed state of the earth materials in the massif or the supercooling of the water in the material and the time required for its transition of ice and, finally, without observing the principle of similarity, which prevents the results obtained from being extended to large volumes of rock or soil *in situ*.

This question is partially touched upon in the report by A. N. Bogolyubov,[16] who concluded that even in permafrost in the natural state the electrical conductivity is a function of temperature. On the one hand, however, the author presents only the final results of the interpretation of his field data, thereby making it impossible for the reader to judge the correctness of the conclusions drawn; on the other hand, he gives no explanation of the numerous vertical electric sounding curves on which lithologically uniform permafrost with a temperature that increases with depth is depicted by asymptotic sections ρ_k = const., i.e., by uniform electric layers--neither does this conclusion require any additional assumptions. These curves were obtained by A. N. Bogolyubov himself.[15]

The general problem of the study of the morphology, cryolithological composition, cryogenous structure, and properties of permafrost breaks down into a number of special problems. These are discussed below:

1. *Mapping thawed and frozen ground in plan and determining the position of the upper boundary of the permafrost.* The solution of this problem is the most common example of the use of geophysics for the investigation of permafrost. Studies to this end have become very common in connection with surveys before the construction

of civil and industrial pipelines, determining the conditions for the working of mineral deposits, searching for groundwater, the need for checking the state of the bases of dams and other structures, checking the natural and forced thawing of placer areas, etc. The mapping of thawed and frozen rocks in plan--the contouring of taliks and permafrost islands--is being successfully realized by means of electrical profiling on direct and low-frequency alternating current and radio frequencies for all types of rocks, the resistivity of which in the thawed and frozen states is different. Somewhat greater difficulties are associated with the search for and contouring of closed taliks (in the permafrost or under the layer of winter freezing). Conditions favoring the detection of such taliks depend on their size, the difference between their electrical conductivity and that of the surrounding permafrost, the geoelectric section, and the position of the talik in the section. Also holding promise for this purpose are methods based on the use of relatively high-frequency electromagnetic fields. The most suitable periods for setting up the work structure of the geoelectric section are prior to the emergence of the thawed layer and after the layer of winter freezing has completely thawed out.

The mapping of thawed and frozen loose deposits is possible by seismic exploration techniques, using for this purpose the sharp attenuation of direct or refracted waves and the variation in their apparent velocities.

In some cases methods can be used that are not based on the differences in the physical properties of thawed and frozen rocks. An example is magnetometry for the tracing of taliks confined to sites of tectonic disturbances. Instances are known of the successful application of radioactive techniques. Thus, S. G. Solopov,[42] in studying the natural gamma radiation, demonstrated the complete correlatability of the boundaries of the radioactive fields of surface deposits with the permafrost boundaries: Above the permafrost islands minimum gamma radiation values, the composition of the constituent rocks remaining unchanged. The increase in the radioactive field above the taliks is due to the favorable conditions of groundwater circulation, linked with which is the removal of the radioactive material to the surface. In this respect the permafrost is a shield, with the result that the radioactive field above it is minimal. Since the talik zones are not only regions of active exchange of deep and surface water but also regions of intensive convectional air exchange, it is natural to assume that for permafrost mapping in areas of sporadic permafrost, the emanation method is equally suitable.[24]

It is more economical to trace the variations in the depth of occurrence of the upper boundary of the permafrost by electrical profiling, with checking of the results by the vertical electric sounding method. The thickness of the taliks in frozen loose deposits is determined by the vertical electric sounding technique, and most precisely by the method of correlating refracted waves (CMRW), by means of which it is also possible to determine the shape of the contact surface between the thawed and frozen rocks in open taliks. The thickness of the taliks in the frozen rocks can be determined by the vertical electric sounding method with a favorable geoelectric section when the difference between the thawed and frozen rocks is sufficiently large. The distinguishing of open taliks is done by the vertical electric sounding method. The linear dimensions of the open part of the talik must be greater than twice the thickness of the permafrost. If the thickness of the permafrost is less than the thickness of the loose deposits, then the open nature of the talik can also be established by the correlation refraction method.

Duplicate observations enable us to trace the seasonal and perennial variations in the position of the boundary between the thawed and frozen rocks and its displacement as a direct result of human influence on the environment. Inasmuch as seepage flows cause and sometimes accompany rapid thawing of the permafrost, one of the methods of checking the state of the foundations of large structures is to observe the natural electric field. This makes it possible to distinguish areas with different thawing rates and thus to identify the most dangerous ones. An efficient means of conducting such checks is, along with measurement of the temperatures, investigation of the space surrounding the well by electric and acoustic logging.

2. *Determining the thickness of the permafrost--one of the principal tasks in geophysics.* The urgency of this problem arises from the fact over vast areas frequently there is not a single sufficiently-deep well and in the near future there will be no change in this situation. This problem is only now being satisfactorily solved by the vertical electrical sounding method.[57] The first promising results were obtained by other electrical prospecting methods, in particular, by electromagnetic frequency sounding. The relative cheapness of arranging for geophysical observations enables us to study the regional patterns of distribution of the thickness of the permafrost, the altitude zonation, and so on. Serving as an example of studies of this type are the results obtained by the vertical electric sounding method at the northern tip of the Verkhoyansk Range.[58]

Determination of the thickness of the permafrost by the vertical electric sounding method is possible under the following conditions: corresponding to the lower boundary of the permafrost is an electrical interface with a resistivity ratio of 1/2 and less; as it is situated in a sufficiently thick series, which is uniform in lithologic composition, there is no doubt as to the nature of the boundary being defined. The shielding effect of the frozen soils or their upper layer, if the entire frozen series is made up of soils, is small enough not to obscure the lower boundary of the permafrost. Subject to the permafrost thickness being sufficiently great ($H > 100$ m), these conditions are most easily satisfied and checked when it consists of rock, especially in folded regions. The interpretation of the vertical electric sounding curves in this case is simple technically: the position of the lower boundary of the permafrost is determined by means of a two-layer set of master curve measuring grids $\rho_1 > \rho_2$. The known difficulties are associated with establishing and taking into consideration the anisotropy in the case of horizontally layered media.

In regions where the thickness of the permafrost is not great, conditions for determining the position of its lower boundary are favorable if it is situated in soils. In effect, the problem amounts to determining the thickness of the frozen soils. The interpretating of the vertical electric sounding curves is usually attended by considerable technical difficulties, but it can be done without the enlistment of additional data in the case of K type ($\rho_1 < \rho_2 > \rho_3$) or AK type ($\rho_1 < \rho_2 < \rho_3 > \rho_4$) sections when there is a sufficiently high resistivity of the second or third layer respectively, by means of the discrete values method.[53,57]

Determining the thickness of the permafrost by means of the radar and geosonar techniques is still a matter for the future. The estimate of the potentialities of the radar method which was made by V. V. Bogorodskiy *et al.*[17] and the initial experimental studies of A. T. Akimov[7] with respect to the investigation of taliks and permafrost by the geosonar method after the hope that under favorable conditions these techniques will be successfully used for the rapid and detailed study of the behavior of the upper boundary of the permafrost, the lithologic analysis of the permafrost, and the determination of its thickness.

The possibilities of determining the thickness of the permafrost by the reflected wave method are evidently limited by the lack of any difference between the velocities of elastic vibrations in thawed and frozen dense rocks. In areas consisting of unconsolidated or of weakly cemented deposits, the thickness of the permafrost, if it is great enough, can be determined by the reflected wave method. No examples of the successful application of the reflected wave method for this purpose are known to have been reported in Soviet literature. In a paper by Yu. K. Agafanov *et al.*,[3] the possibility of making a quantitative estimate of the variation in the thickness of the permafrost by reference a test well is demonstrated. In the special case where the thickness of the unconsolidated deposits barely exceeds the thickness of the permafrost, in estimating the position of its lower boundary the CMRW can be used.[21]

Determining the boundaries of the permafrost by means of well-logging operations is not only a necessary expedient when there is significant mineralization of the pore solution, giving rise to a change in the position of these boundaries in comparison with the temperature boundaries, but also in the general case, inasmuch as the restoration of the natural temperature regime of the rocks through which the well has been made requires a long period of time, sometimes as much as several years. The boundaries of the permafrost can be determined by a number of well-logging techniques: various versions of electric, dielectric, cavernometric, acoustic, seismic, radiometric, and other loggings. In favorable cases the problem can be solved by using one of these methods (resistance logging, seismic logging, acoustic logging), and in more complex situations, by joint interpretation of data obtained by various techniques. Instances are known of the successful use of acoustic logging of cased wells for this purpose.[20]

3. *Mapping permafrost of different lithologic composition, cryogenous texture, and cryogenous structure.* The constancy, and among sedimentary rocks even the increase in the relative differentiation of the frozen ground in terms of its electrical conductivity, makes possible successful resolution of this problem by electric prospecting techniques. Electrical profiling has become one of the indispensable components of the sum total of geological prospecting activities connected with the surveying and exploration of ore deposits. Not only are the lithologic varieties of the rocks, the veined bodies and dikes mapped, but also the tectonic dislocations. In some cases, freezing increases the difference in electrical conductivity of the rocks (unchanged extrusive rocks and sandstones, shales, and sandstones), and their mapping is greatly facilitated. In other cases, the reverse is true: the difference between the rocks in terms of their electrical conductivity is less in the frozen state than in the thawed state (sandstones and granitoids, sandstones and modified extrusives). By varying the spacing of the power supply units or the field frequency it is possible to map either over a layer of frozen rocks or over a layer of thawed rocks (the subpermafrost layer). The fact that this choice is available improves the changes of successfully resolving the problem. It is also of interest in that the search for relatively deep-lying ore bodies with high resistivity--a not common target in the northeastern USSR--is a difficult task.

Various versions of electric profiling are being used: for the mapping and exploration of poorly-conducting ore bodies, for the most part direct and low-frequency alternating currents are used. In the search for bodies that are good conductors, electromagnetic fields of relatively high frequency and the natural field method are used. In searching for and exploring ore deposits of the vein type and also coal deposits, the method of the arbitrary component of the current density or the inductance reception method is effective.[35,54] This method is suitable when searching for targets that have both a higher and a lower electrical conductivity relative to the country rock, and neither case necessitates grounding of the receiving line. It can therefore be used in any type of cover: in rock waste, in snow, and so on.[43]

It is the unconsolidated deposits that undergo the most profound changes as a result of freezing, and it is these that constitute one of the primary objects of research in geocryology. To this must be added the fact that the structure, the electrical conductivity, and the thickness of the individual layers of the frozen deposits determine in large measure the resolving capability of electrical prospecting. By electric profiling it is possible to map frozen soils that differ with respect to the lithological composition, the type, and degree of development of the cryogenous structure and the cryogenous texture, which also suffices for genetic differentiation. Usually this is achieved by tracing the boundaries of a layer with a definite electrical conductivity. Sometimes the mapping of areas with a definite type of geoelectric section is possible. The results of the electric profiling must be carefully checked by vertical electric sounding method. Because of the complexity of the geoelectric structure of frozen loose sediments, this latter method is frequently the basic technique for their mapping.

The longitudinal wave velocity can be used for mapping frozen soils of differing lithologic composition, and within the confines of areas that are uniform in composition, their iciness can be used.

Wedge, injection, and other large sills of ground ice are an important and fairly widespread component of the cryogenous structure of frozen soils, thus determining their geotechnical properties and the scale and rate of progress of such frequently irreversible processes as thermokarst. Recently, areas containing large sills of ground ice have frequently been used for the building of industrial and civil structures and as sources of potable and industrial water. It is therefore necessary to know both their areal extent and the positions of the various ice sills. The individual bodies of wedge ice are traced by detailed electrical profiling studies. The correlation of the anomalies presents certain difficulties in view of their large number, the presence of false anomalies as a result of the shielding effect, etc. Therefore, ascertaining the positioning of the wedge ice system and determining the thickness of the individual ice bodies, if this is great enough, necessitates undertaking a large volume of highly detailed work (a dense network of profiles of differing orientations with a spacing of 1-2 m between observations is sufficient). The mapping of thick wedge ice and the estimating of its thickness can be done from the longitudinal wave velocity when tracing the upper boundary of the loose deposits. A specific example is described in a paper by O. K. Voronkov and G. V. Midhaylovskiy,[23] who simultaneously attempted to estimate the vertical extent of the wedge ice. An attempt to map wedge ice by means of a micromagnetic survey that made it possible to establish a detailed picture of its positioning also proved successful.[13] In this instance the diamagnetism of the ice, its wedge shape, and the shallow depth of occurrence of its upper edge, as well as the confinement of the wedge ice to fine-grained soils, which usually have a somewhat higher intensity of magnetization, favored the use of a magnetic survey. It is noteworthy that the phenomena associated with seasonal thawing and freezing had no effect on the techniques employed in the micromagnetic survey or on the results derived from it.

When determining the thickness and depth of occurrence of injection and other sheet ice of fairly large dimensions in plan, good results are provided by seismic prospecting.[38] In the sheet ice multiple reflected waves are formed, the period of which is directly proportional to the thickness of the ice. The depth of occurrence of the ice is determined by the system of counter refracted wave hodographs. It is known that some successes have been achieved in the mapping of injection ice by means of electric profiling and vertical electric sounding.

The possibility of studying the variation in the lithologic composition of permafrost in the vertical section by geophysical ground techniques depends on the type and the parameters of the geoelectric section when we are talking about electrical prospecting and on those of the seismogeological section in the case of seismic processing; it diminishes as the thickness of the permafrost becomes less.

For investigating the lithologic section of wells sunk in rocks and soils, the usual array of logging techniques is used. For logging by the resistivity method, in the case in question some additional possibilities emerge: comparison of the data obtained by small and large probes and accordingly pertaining to the thawed and frozen regions of the earth materials, respectively, makes it possible to estimate the ratio between the electrical conductivities of the materials in the thawed and frozen states and thereby furnishes an additional parameter for diagnosing the material and estimating the structure of its pore space.

When studying the structure of frozen unconsolidated deposits in the vertical section by geophysical ground techniques, the following results can be obtained. The vertical electric sounding technique is used to determine the type of geoelectric section. In particular, in winter after freezing of the seasonally thawed layer, the vertical electric sounding technique permits the establishment of an even thinner structure of all or part of the section.[56] This is a difficult problem, however, and is not always solved solely by the data from the vertical electric sounding. For a sharp improvement in the reliability of its solution, it proved possible to use computerized recognition programs.[57] It is advisable to make a comparative study of the section and of the ratio between the electrical conductivities of loose deposits in both the frozen and the thawed states, that is, outside and inside the taliks.

In unconsolidated deposits the correlation method of refracted waves is used to distinguish the horizontal interfaces at which an increase in the velocity of the longitudinal waves takes place. The combined use of the vertical electric sounding technique and the correlation method of refracted waves makes it possible to establish practically all of the basic lithologic and permafrost boundaries in loose deposits. The problem of determining the thickness of frozen loose deposits by geophysical methods will be discussed below.

The lithologic section of frozen loose deposits in wells is studied by a range of techniques. For this purpose, dry wells drilled with blown air or wells with drilling mud consisting of a mixture of oil and clay are preferable. The total moisture content and density of frozen unconsolidated deposits are determined by radioactive logging techniques; when there is a nonuniform moisture distribution in the section, electric logging techniques--resistance logging, well potential and J logging, electromagnetic frequency logging, and acoustic logging--are applicable. Acoustic logging makes it possible to determine the velocity of the longitudinal and transverse waves by which the dynamic Young's modulus and Poisson's ratio are calculated. The aggregate of data on the moisture distribution, iciness, density, apparent resistances, velocities of the longitudinal, transverse, and Rayleigh waves, temperature, and other parameters with respect to the well log make possible a detailed cryolithological analysis of the frozen soils, the distinguishing of talik zones, and the provision of a sufficiently complete evaluation of the frozen unconsolidated deposits for them to be put to practical use in geotechnics and geophysics.

4. *Determining the thickness of frozen unconsolidated deposits* is of great practical importance when surveying areas designated for the building of industrial and civil structures and pipelines, and when conducting search and exploration activities for placer gold, tin, and diamonds. It is a widely encountered problem.

The thickness of frozen soils is determined by the vertical electric sounding method and the correlation method of refracted waves. In the case of the most widespread type of geoelectric logging--type K in relation to a layer of frozen earth material used as the reference (in summer)--the vertical electric sounding method is used to find a technical solution to the problem but only after the seasonally thawing layer has become completely frozen, that is only at the most unfavorable time of the year. In summer it is necessary to know the type of geoelectric section and the resistivity of the frozen unconsolidated deposits. The major technical difficulties arising in this connection were overcome by the development of the two most widely used types of vertical electric sounding curves, K and AK with relatively high values of ρ_2 and ρ_3 for interpreting vertical electric sounding curves, and the method of discrete resistivity values based on the use of the general principle of equivalence $T_{1,2}$[55] and permitting the simultaneous determination of the thickness and the electrical conductivity of the frozen unconsolidated deposits.[53,57] The mean error as determined by the results of a subsequent check in areas where the thickness of the deposits is more than 10 m is ~12 percent.[31]

For making highly delicate and complex estimates in the course of the interpretation, indicator systems obtained by means of the computerized Kora-3 recognition program are highly effective.[57]

Determining the thickness of frozen unconsolidated deposits by the correlation method of refracted waves rests on multiple tracing of the principal refracting boundary using extended systems of operations.[12] Under favorable conditions the accuracy of determining the thickness of the frozen soils using seismic exploration is ±5 percent ±5 m; the minimum absolute accuracy is ±5 m.[36]

Equally unfavorable for use is the section involving the vertical electric sounding method and the correlation method of refracted waves in which frozen unconsolidated deposits with a high resistivity and a high velocity of the longitudinal waves (for example, a layer containing wedge ice) occur in frozen deposits with a lower resistivity and a lower velocity or in thawed deposits. The combined use of the vertical electric sounding method and the correlation method of refracted waves does, however, make it possible to distinguish such areas and determine the overall thickness of the soils in their individual layers. When the thickness of the frozen soils is greater it is advisable to include gravimetry in the aggregate of geophysical operations. The radar sounding method is highly promising: the immediate purpose of its use is to determine the thickness of the frozen unconsolidated sediments.

5. *Investigating the physical properties of frozen ground*, both by direct determination and through correlation, is an exceptionally important, independent area. An understanding of the physical properties of frozen ground is necessary not only as a basis for the use of existing geophysical methods and the development of new ones for the investigating permafrost, but also when solving a wide range of geotechnical problems, extending all the way from problems of construction to seismic microzoning and including such problems as managing the properties of frozen ground, its disintegration and working, heating, and melting, the design and equipping of communications lines and facilities and of other electrical systems, etc. The problem of seismic microzoning of the permafrost region was the subject of a paper by V. P. Solonenko.[41] It analyzes the distinctive features of the seismicity of the permafrost region and proposes the establishment of a new area of research be known as seismogeocryology. Solonenko distinguishes five types of permafrost, which he uses as the basis for seismogeocryological zoning. The specific features of the engineering-seismocryological conditions of each type of permafrost are outlined. The effect of earthquakes on the course of the geocryological processes--solifluction, thermokarst, and icing formation--is demonstrated.

The study of time variations and of variations in the physical properties of permafrost is important both for predicting changes in the geocryological conditions and for selecting optimal solutions of specific applied problems.[10] Included under this heading is a paper by N. Ye. Zarubin and O. V. Pavlov,[29] in which a study is made of the problems of predicting the behavior of the seismic properties of permafrost both as a result of its seasonal variations and under the influence of building development.

Advances in the methodology conducting geophysical studies of permafrost are also exceptionally important for the solution of many geological prospecting problems by geophysical techniques. When it is a question of investigating the frozen zone of a rock series, problems of a geocryological and geological nature coincide. In other cases, allowance must be made for the presence of permafrost when estimating the possibility of solving a specific problem, selecting the geophysical methods, and deciding on the procedure, but also when interpreting the observational data. In electrical prospecting this is usually due to the shielding effect of the frozen unconsolidated deposits and the formation of an additional inter-ace--the lower boundary of the permafrost, which shields to a certain depth all of the other subjacent boundaries, and in the case of layer by layer interpretation calls for a knowledge of the parameters of the external layer.

Similar difficulties--for the most part when the permafrost consists of unconsolidated sediments--are encountered in seismic prospecting. These are: complication of the wave picture as a result of the formation of an additional interface and the necessity of determining its position in order to arrive at a true estimate of the mean velocity; the possible emergence of a horizontal velocity gradient as a result of a variation in the thickness of the permafrost and of additional vertical interfaces (taliks); the origination of diffracted waves and interference waves as a result of the complex nature of the geocryological structure of the region being investigated.

In recent years, the so-called remote-sensing methods of studying natural resources--using airborne or space platforms--have undergone rapid development. Essentially, there are the infrared, radiothermal, and radar survey techniques as well as radar sounding.

The infrared survey is conducted in the 1 to 14 micrometer range. In modern infrared scanning systems the spatial resolution is 1.5 milliradians, and the temperature resolution is 0.3-0.5°C at the 20°C level. The effectiveness of the method increases significantly with multiple repetition of the survey inasmuch as this makes it possible to use the time and the random variations of the thermal contrasts of the surface. The infrared survey makes possible separation of the surface deposits in terms of their lithologic composition and moisture content.

The radiothermal survey is conducted in the centimeter frequency range; the depth of occurrence of the radiating layer can be as much as several meters. The results of experimental studies enabled G. Ya. Chernyak and O. M. Myaskovskiy[50] to conclude that the radiothermal survey is suitable for mapping surface deposits with respect to lithology, moisture content, and degree of salinity and for separation of frozen and thawed soils.

The radar survey makes possible separation of surface deposits with respect to lithological composition inasmuch as the image contrasts depend on the capacity of the reflecting surfaces for diffuse scattering and also on the electrical properties of the soils and rocks.[51] Most promising of all is radar sounding which, by overcoming technical difficulties, may make it possible to solve many problems: determining the thickness of frozen unconsolidated deposits and analyzing them with respect to composition and vertical section, mapping, and determining the thickness of the permafrost, and so on.

It seems to us that the hopes expressed when geocryology was in its infancy as regards the enormous potentialities of geophysics have been largely justified. Today, the vast majority of the standard and special problems of geocryology are being solved by geophysical methods. Nevertheless the potentialities of geophysics are far from being fully utilized and the extent to which geophysical techniques are being included within the sum total of geocryological research is, in our opinion, inadequate.

The further development of geophysical methods of investigating permafrost must be channelled in the following basic areas:

(a) The development and introduction of new methods that have a high resolving power and effecting improvements in the resolving power of existing geophysical techniques;

(b) Sharply increasing the speed of execution and the efficiency of geophysical studies by converting to airborne (infrared surveys, radiothermal radiation methods, radar, aerial electrical prospecting) and abovewater surveys (geosonar studies in the shelf zone and in lakes and rivers);

(c) Automating the methods of processing the observational data and ensuing the more complete extraction of the information through the extensive use of mathematics and modern computers and its representation in the most accessible form for use.

The further development of geophysical procedures for the study of permafrost in the areas enumerated above corresponds in very large measure to the requirements of the industrial development of the North, inasmuch as it allows for the exceedingly extraordinary complex nature of the cryolithological composition and cryogenous structure of the permafrost and the requirement for detailed studies to be carried out as quickly as possible, often over vast expanses--when surveying routes for gas and oil pipelines, electric power transmission lines, roads, and so on--under the harsh conditions of the North.

The outlook for the future development of geophysical methods for investigating permafrost is linked first and foremost with electrical prospecting, techniques, including the infrared region of the variable electromagnetic field spectrum, and with seismic prospecting and seismo-acoustics. The potentialities of seismic prospecting and seismoacoustics are only being utilized to a moderate degree, and much progress can be expected in this area. Electrical prospecting by different variable field techniques, primarily those in the radiowave band, is already being applied in practice. The extent to which theoretical problems and those concerned with apparatus have been worked out, and the level and rate of expansion of our knowledge of the electrical properties of permafrost, as well as of the structure and parameters of the geoelectric section, also afford a basis for anticipating that important advances will be made in the near future. The development of the infrared, radiowave, and radar survey techniques offers special promise.

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PRINCIPLES OF CONTROLLING THE GEOCRYOLOGICAL CONDITIONS DURING CONSTRUCTION IN THE PERMAFROST REGION (SURVEY OF RESEARCH IN THE USSR)

S. S. VYALOV Scientific Research Institute of Foundations and Underground Structures

INTRODUCTION

The origin and growth of the science of geocryology are primarily the outcome of engineering requirements that have resulted from the economic development of the permafrost region. Actually, although the existence of permafrost has been known since the early 1800's (the studies of Middendorf *et al.*), the systematic study of it only began when the engineers encountered the inherent difficulties that characterize construction on this type of ground, during the building of the Trans-Baikal and Amur railroads (the 1900's). The establishment of geocryology as an independent science dates from the 1930's and 1940's, which marked the beginning of intensive economic development of the Soviet North and Northeast. It was during these years, with the publication of monographs by Tsytovich and Sumgin, Liverovskiy, Bykov and Kapterev *et al.*, that the principles of the engineering branch of the science of frozen ground--engineering geocryology--were formulated. I think I am correct in saying that in the United States and Canada the establishment of geocryology and especially its engineering branch was similarly connected with engineering projects in the northern regions. In particular, it played an important role in the building of the Alaska-Canada Highway and in the construction of industrial enterprises and various installations in Alaska and the northern territories of Canada.

The problems of engineering geocryology consist primarily of studying the conditions and laws that govern the interaction of frozen ground with various types of engineering structures and, secondly, of devising methods of exerting directional control over the properties of frozen, thawing, and freezing ground and the processes occurring in this ground.

This report presents a review of the present status of engineering geocryology in the USSR. The review has been compiled on the basis of papers and reports presented at the Second International Conference on Permafrost (Geocryology) and takes into account the experience accumulated by research, design, and construction organizations.

THE PRINCIPLES OF BUILDING ON PERMAFROST

In engineering practice two basic principles governing the use of perennially frozen ground have been clearly manifested. In the USSR they have been codified by the State normative documents ([SNIPs] "Footings and Foundations" ... 1967):[32] The ground must be maintained in the frozen state or it must be allowed to thaw. The choice of one principle or the other is determined by the permafrost-geological conditions at the site, the technological flow chart and design diagram for the structure, and the conditions under which it will be used.

The first principle is a standard example of controlling the properties of the ground when we use the harshness of the northern environment itself and in supplying a specific amount of cold to the ground, provided the requisite thermal regime in which the absence of unacceptable deformations to the structure is assured. The idea of controlling the properties of the ground in the desired direction is manifested on an even larger scale in the artificial cooling of the frozen ground constituting the foundation or in the freezing of thawed ground, creating a permafrost curtain, etc., when we alter these properties abruptly and increase the strength and bearing capacity of the ground by freezing it, and thereby insuring its water imperviousness, etc.

As for the second principle, the problem can be defined as follows: In preceding years construction entailing the thawing of frozen ground was accomplished in such a way that this thawing was allowed to take place during use of the structure. In order to prevent structural deformations due to settlement of the thawing ground, a number of design measures were taken. It was as if the structural elements had been adapted to the behavior of the thawing ground, that is, a passive method of construction was used.

However, construction practice demonstrated that this method is not efficient, inasmuch as in most cases the settlement of the thawing ground is so pronounced that the structures, in spite of their designs having been adapted, exhibited inadmissible deformations.[18] Therefore, construction entailing thawing during use of the installation is now permitted only in such frozen ground as results in moderate settlement during the thawing process and which does not exceed the threshold values established by the norms. In the case of slumping (ice saturated) ground we go to the active method of construction by providing for artificial, prethawing of the frozen ground constituting the foundation and its consolidation. As a result, we obtain a foundation in which the deformation of the installation does not exceed the permissible values.

Finally, among the methods of controlling the properties of frozen ground, it is necessary to include those involving a physicochemical effect, which are highly promising. In this case, by adding various reagents to the thawing or freezing ground, we can alter its strength, permeability, and other properties.

MAINTAINING THE GROUND IN THE FROZEN STATE

The principle of maintaining the frozen state of the ground has been used in the USSR since the 1930's--in Yakutsk at the present time an electric power plant is functioning that was built in 1935 according to this principle[47] (Figure 1).

The principal measure insuring preservation of the frozen state of the ground is the provision of a cold ventilated air space.[8] These

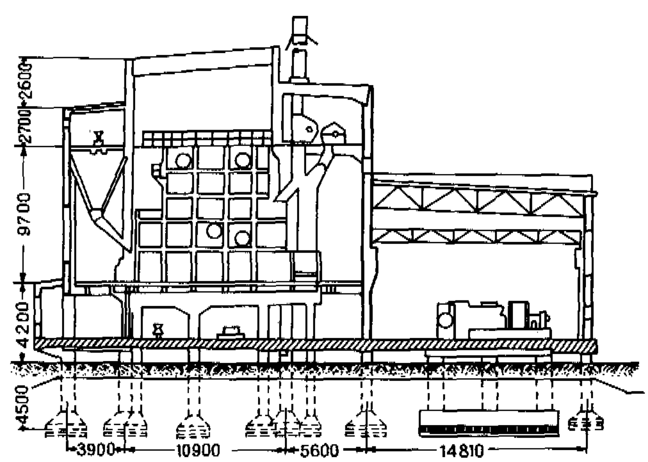


FIGURE 1 Electric power plant built at Yakutsk in 1935, with maintenance of the foundation in the frozen state.

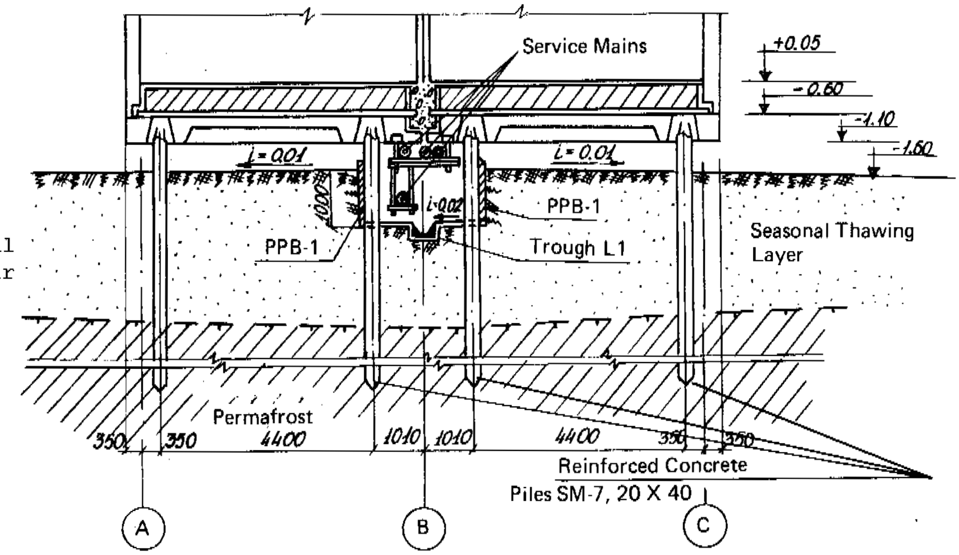


FIGURE 2 Five-story residential building with low ventilated air space (development of the foundation design).

air spaces are made with heights ranging from 0.5 m (low) to 1.0-2.0 m (high). Design solutions of air spaces are presented in Figures 2 and 3. In the case of the low air spaces the main piping is laid in special channels (Figure 2); in the high air spaces the piping is suspended from above. In recent years there has been a trend towards the building of high air spaces and the construction of special floors for the housing of engineering plants. In some cases an unheated lower floor is provided, which is used as auxiliary space (Figure 4).

For wide-span structures, in which the construction of air spaces is structurally difficult, the maintenance of the foundation in the frozen state is insured by cooling it with outside air during the winter through pipes or ducts laid under the floor. This procedure was first used (with the participation of the author) in 1939 for restoring the frozen state of the foundation under the Communications Center in Magadan, the faulty design solution for which had led to thawing of the permafrost. Examples of cooling by means of tubes and ducts are presented in Figures 5 and 6. Proposals providing for local cooling of the ground by admitting cold air into hollow piles have been worked out. The analyses of the operation of the cooling devices and their structural solutions are presented in the paper by Shishkanov *et al.*[49]

A variation of the examples of controlling the ground temperature regime is to furnish beneath the building naturally-ventilated air spaces made of coarsely skeletal material, the pores of which allow cold outside air to circulate.[42]

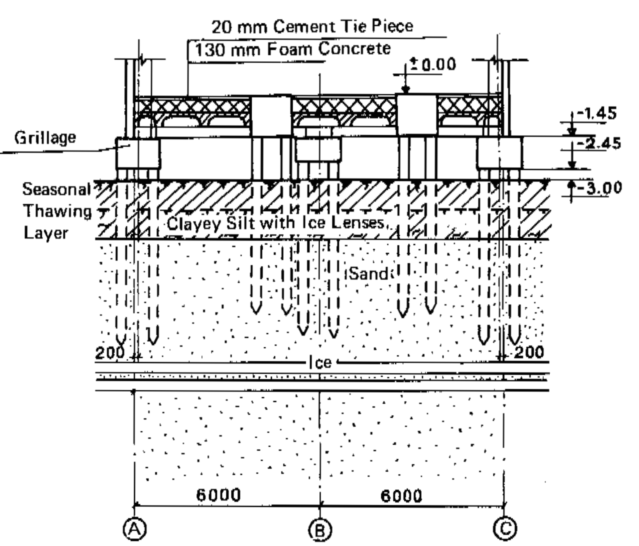


FIGURE 3 Industrial building with high air space (development of the foundation design).

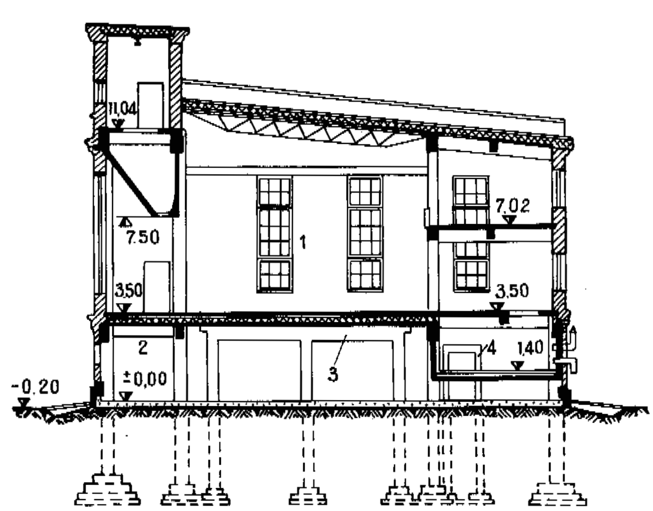


FIGURE 4 Boiler house with cold first story used as zonal space.

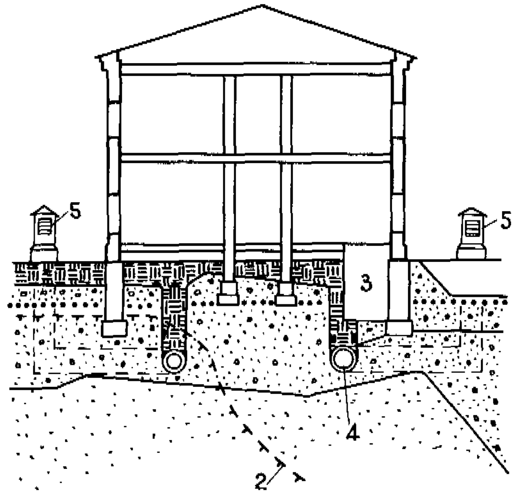


FIGURE 5 Restoration of the frozen state of the ground beneath the Communications Center in Magadan. 1--permafrost boundary prior to beginning construction (1935); 2--boundary of the thaw basin (November 1937); 3--foundation for heating boiler; 4--cooling ducts; 5--intake and exhause shafts.

COOLING PLASTICALLY FROZEN GROUND

A further development of the method of maintaining the perennially frozen state is that of cooling the frozen ground and lowering its temperature to subnormal values. This method is used in cases where the sites consist of plastically frozen ground.[8]

This kind of ground, having a temperature close to 0°, is highly unstable and can thaw even when there are only slight thermal effects. Furthermore, this ground has a low bearing capacity. Therefore, the "Norms" ("Footings and Foundations" ...1967)[32] require that when building in accordance with the first principle the temperature of the plastically frozen ground be lowered so that the ground becomes solidly frozen. In this way, a firm safeguard is provided against unforeseen thermal influences and the bearing capacity of the foundation is increased markedly. The effectiveness of this procedure is clear if only

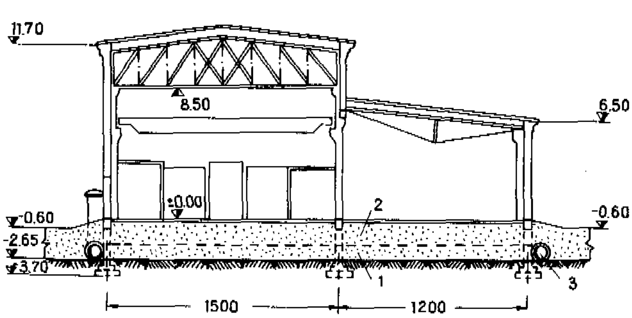


FIGURE 6 Cooling the ground forming the foundation of the mechanical assembly shop: 1--cooling by pipes; 2--fill; 3--air intake.

from the fact that on lowering the temperature, for example, from -0.5° to -2° the pile-ground adfreezing strength increases from 0.5 to 1.5 kg/cm^2, that is, it triples.

The cooling of the ground is achieved by both natural and artificial means. The possibility of natural cooling arises from the fact that, due to the influence of the vegetative and snow covers, as a rule the mean annual ground temperature is higher than the mean annual temperature of the outside air. For example, at Igarka the air temperature is -7.7°, whereas the mean annual temperature of the frozen ground ranges from -0.2° to -1°. At Noril'sk, with a mean annual air temperature of -10.1°, the ground temperature at the zero amplitude depth ranges from -3° to -7°; in some areas it even rises to -0.2°.

Observations have established that in the majority of cases the presence of a high ventilated air space not only insures maintaining the foundation in the frozen state, but it can also appreciably lower its temperature. This phenomenon is observed at Yakutsk, where in a correctly built and correctly used high air space a raising of the permafrost table by 0.3 m and more is to be seen. Also at Yakutsk, the cosntruction of a ventilated air space provided for elimination of a 4-m deep thaw basin beneath the old library building over a 9-yr period. At Noril'sk the construction of high (1.0-1.5 m) crawl spaces for buildings erected in areas containing plastically frozen ground is resulting in a lowering of the temperature from -0.2° to -3°. A similar phenomenon is noted at Igarka.

However, cooling by building an air space requires a relatively long period of time. Thus, at Noril'sk the temperature under one of the buildings with a high air space dropped at a depth of 5 m from -3.7° to -5° after 2.5 yr and to -6° after 5 yr.

Estimates have shown that for the calculated ground strength it was necessary to take the strength corresponding to the state of the ground soon after transfer of a load to it. Therefore, prolonged lowering of the foundation temperature leads to an increase in the safety factor but not to a lightening of the foundation design. Consequently, along with the lowering of the ground temperature during use of the building, it is advisable to cool it before or during construction, having lowered and maintained the calculated ground temperature until the steady state as determined by the effect of the air space is established.

The simplest procedure is preliminary cooling of the ground by systematic removal of snow from the site. Tests at Igarka demonstrated that in this event the mean annual ground temperature at a depth of 5 m can be lowered from -0.8° to -3°. An additional effect can be provided by shading and thermally insulating the ground surface during the summer.

But it must be kept in mind that lowering of the ground temperature by the aforementioned examples of very simple types of thermal modification can only be achieved after 1 to 3 yr. Therefore, this procedure is applicable when there is sufficient time to implement prebuilding measures, for example, when developing large areas it is possible to install main piping in individual sections and thereby secure an operations lead time of 1 to 2 yr on the erection of the buildings. Usually it is necessary to accomplish the cooling process within a shorter period. In these cases it is necessary to accomplish the cooling process within a shorter period. In these cases it is possible to effect freezing through the medium of natural cold emanating from the bottom of the foundation trenches. This procedure was implemented, in particular, at Vorkuta when erecting the service building and the feed bin. The supply of cold air was force fed.

Highly efficient cooling can be achieved by using hollow piles as the foundations and feeding cold air through them, as mentioned earlier.[49]

Cooling the ground with cold air fed by a fan into holes drilled for the subsequent installation of piles (Figure 7) provides a good effect.[28]

At an outside air temperature of -20°, the required cooling time is 22 days, and at a temperature of -50°, 5 days. The method was successfully used at Noril'sk and Igarka. The method of cooling the boreholes with dry ice was also used there.

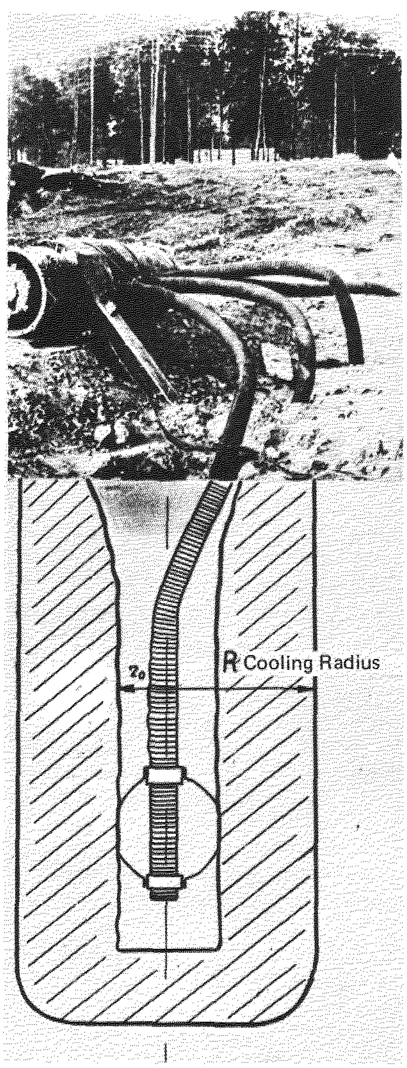


FIGURE 7 Cooling the ground by admitting cold air to boreholes drilled for subsequent installation of piles.

Under favorable permafrost conditions and for essential installations it is more advisable to use artificial freezing by means of brines. This method is applicable where discontinuous permafrost is present, when a need arises for the freezing of taliks, when erecting earth dams, when it is necessary to build a frozen core of a dam, etc. Finally, the method is widely used when driving underground workings in ordinary unfrozen rocks and when weak quicksands are encountered, in which case they are artificially frozen in order that tunnels can be driven in them.

SELF-REGULATING FREEZING DEVICES

The high cost of cooling the ground by brines or air ventilation has impelled us to seek more efficient methods. One such example is the self-regulating cooling device, a characteristic feature of which is the automatic achievement of heat exchange between the cold outside air and the ground. The cooling agent in such devices can be a two-phase (gas-liquid) or single phase (liquid) medium. In the former case the gas (propane) circulating through the tubes of the device as a result of the temperature difference between the outside winter air and the ground, completes an evaporation-condensation cycle, thereby transmitting cold to the surrounding ground. In the latter case the liquid circulating as a result of the temperature difference supplies cold to the ground by convection. The vapor-liquid unit was proposed in the United States by Long, and at the present time the USSR is conducting experimental studies of similar devices where the use of Freon as the cooling agent is being tested. The liquid unit in which the heat exchange agent is kerosene was proposed by S. I. Gapeyev.[10,11] He developed three structural designs for the unit--single-tube, double-tube, and multiple tube. The rate of circulation of the kerosene in the tubes reaches 40 cm/s when there is a temperature difference of 40° between the outside air and the ground and 1 cm/s when the difference is 10°. The units are being successfully used to restore the frozen state of thawed ground. They were used in particular to eliminate the consequences of unforeseen thawing of the ground beneath a part of the foundation of the Yakutsk Electric Power Plant. The units are also used to maintain the frozen state in certain unfavorable sites, for example, under bridge supports (Figure 8).

The Yakutsk Institute for the Planning and Deisgn of Diamond Mining Installations and the Permafrost Institute of the Siberian Division of the USSR Academy of Sciences developed modifications of coaxial liquid cooling units during the period 1964-1970.[2] These devices are used both independently and when mounted on piles. The latter case is highly efficient, inasmuch as, on the one hand, it does away with the need

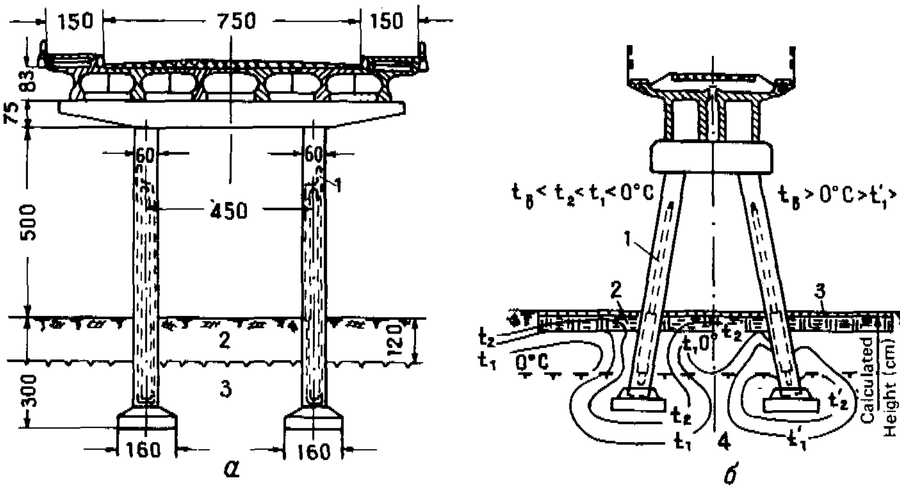


FIGURE 8 Bridge supports incorporating S. M. Gapeyevs double-tube (a) and single-tube (b) cooling units.

for additional work in order to bury the unit in the ground and, on the other hand, more favorable operating conditions are set up for the pile, which itself generates a cooled zone of ground around it by means of which the bearing capacity of the pile foundation is substantially increased. The experimental data demonstrated that, after installing the "self-cooling" piles, the temperature of the plastically frozen ground, which under natural conditions is -0.2°, dropped to -2° during the spring and summer and to -20° in the winter.

Today, "self-cooling" piles are being quite extensively used for construction in areas containing plastically frozen ground near the town of Mirnyy. The savings resulting from the improvement in the bearing capacity of the pile amount to 6-8 percent of its cost.

THAWING OF THE GROUND

As already noted, according to the Construction Norms and Regulations, construction on permafrost that is subject to thawing during use of the installation is permitted in those cases where the settlement of the thawing substructures does not exceed the limiting values. Otherwise, a prethawing (preconstruction procedure) is used (if it is impossible to maintain the frozen state of the ground).[8,32] This method was proposed quite long ago, but only recently has it been applied in practice.

As is well known, the method is based on the fact that, as a result of prethawing, the most perceptible part of the settlement is eliminated due to melting of the ice and consolidation of the ground under its own weight; the magnitude and rate of subsequent deformation of the foundation are accordingly greatly diminished.[8,22] If the consolidation of the prethawed ground under its own weight proves insufficient, then it is artificially consolidated, either mechanically or by electroosmosis, etc. Where necessary, hardening of the thawed ground is accomplished by introducing chemical additives.

One of the first examples of preconstruction thawing was the thawing of the ground in 1961 at sites with discontinuous "permafrost" at Vorkuta. On these sites two buildings had to be erected on substructures consisting in part of ice-saturated permafrost suglinoks and in part of thawed ground. In order to avoid appreciable uneven settling, it was decided to prethaw the permafrost by means of an electrical thawing technique, and then to consolidate the thawed ground by electroosmosis. The thawing experiment was completely successful. At a thaw depth of 10 m the settlement of the thawed ground was from 40 to 100 cm. After erecting the building, the subsequent settlement was from 5 to 30 cm, and this gradually became less. The building has functioned normally ever since.

Later, the prethawing technique became widely used at Vorkuta. A number of residential and administrative buildings and industrial installations were erected in this manner. Moreover, the thawing technique itself has been improved in that electric water heaters, electrochemical heating, etc., are used. It turned out that during the thawing process, owing to the increased capacity of the frozen silty ground to be displaced, consolidation of the thawed ground under its own weight is quite good.

A highly labor-intensive stage of the work on the thawing of the ground is the drilling of boreholes for the installation of electrodes. This problem is solved by using vibrating pile drivers and vibrating hammers, which substantially accelerate the drilling of the hole--the driving rate of tubes measuring 73-108 mm in diameter is 0.3-1.0 m/min when they are driven by a vibrating hammer.

In the Magadan Oblast prethawing is used when building structures on ice-saturated coarsely clastic soil. The thawing is effected hydraulically, using river water of natural temperature or industrial waste water. The time required for thawing is one-and-a-half to two-and-a-half months.

The problem of prethawing, including the describing of the various procedures, and also the design techniques are discussed in an article by Z. F. Zhukov and V. D. Ponomarev.[22] We would point out that in this paper a new formula is proposed according to which there are three types of thaw-induced settlement: that due to disintegration of the cryogenous structure during thawing, another due to consolidation under the influence of its own weight, and the third due to consolidation resulting from the load imposed by the structure. The first two types of settlement occur during prethawing; the third, after construction.

PHYSICOCHEMICAL METHODS OF CONTROLLING THE PROPERTIES OF FROZEN GROUND

Among the most promising avenues of controlling the mechanical properties of frozen and thawing ground are physicochemical methods.[38,39]

As is well known, depending on the magnitude and the sign of the charge, the adsorbed ions on the surface of mineral particles can have both a coagulating and a dispersing effect on the ground structure, altering the ratio between the ice and the tightly bound, unfrozen water. They can also affect the strength of the bonds between the ground components. The strength of the frozen ground will vary in conformity with these changes. Thus, by influencing the adsorbed capacity of the frozen ground in one way or another or alternatively the chemical composition of its liquid phase, it is possible to alter the strength of this ground in either direction, i.e., to increase it or decrease it. The practical application of these methods for varying the adfreezing bond strength between the ground and the foundation surface can be summarized as follows.

It is known that adfreezing bond strengths play both a favorable and an unfavorable role in engineering practice. The favorable role is exhibited during laying of a foundation in permafrost when the adfreezing bond strength takes a significant part of the external load transmitted to the ground through the foundation.

Adfreezing bond strengths play a particularly important role during the building of pile foundations. In this case the bearing capacity of the foundation is basically determined by these strengths. It is possible to increase the adfreezing bond strength in two ways: by physicochemical treatment of the pile surface or, in view of the fact that the pile is lowered into a liquid soil solution that is subsequently frozen, by physicochemical treatment of this solution.

In the first case, activation of the foundation surface can be achieved,[38,39] for example by treating this surface with hydrochloric or sulfuric acid and by removing the upper layer of concrete by sandblasting (the adfreezing bond strength increases by 25-50 percent). In the second case it is necessary to increase the degree of dispersion of the ground by treating it with various reagents so that the ice film at the contact between the ground and the pile surface is diminished and the adfreezing bond strength is thereby increased. Thus, when treating the ground with cement, organosilicon liquids, or iron oxide, the adfreezing bond strength is increased by 1.5 times.

The unfavorable role of the adfreezing bond strength is manifested in the effect of frost heave on the foundation. It is known that the magnitude of the tangential forces of frost heave is determined by the adfreezing bond strength between the heaving ground and the foundation surface and, consequently, a decrease in the adfreezing bond strength leads to a decrease in the heaving forces. This can be achieved[38,39] by decreasing the wetting of the foundation surface with the water adsorbed by the ground particles, for which purpose this surface must be coated with a high molecular compound that has hydrophobic (water repelling) properties. Experiments have shown that coating the foundation surface with a film of epoxy resin decreases the heaving forces by 2-3 times.

Another method of decreasing the heaving forces is to alter the heaving properties of the freezing ground by making it hydrophobic, replacing the multivalent cations in the exchange complex by univalent cations, decreasing the magnitude of the specific surface of the mineral particles by using active texture-forming agents, and, finally, by introducing salts and texture-forming agents into the ground. The treatment of the ground with reagents containing the fluorine ion and carbamide resins has yielded good results.

Physicochemical methods can also be used to advantage for strengthening thawed ground. It is known that cementation or the strengthening of argillaceous soils with resins is only moderately efficient in view of the difficulty of injecting solutions into these soils. Owing to the increased capacity for dispersion, in the case of perennially frozen, silty argillaceous soils this problem is more successfully solved during their thawing.

THE INTERACTION BETWEEN FREEZING SOILS AND FOUNDATIONS

Basic research into the problem of foundation heave has consisted of studying the tangential forces of frost heave of the ground acting on the foundation and of working out measures promoting a decrease in these forces.[16,24,25,33,34,40] Experimental studies have established that heaving occurs even in non-water-saturated argillaceous soil, the magnitude being directly proportional to the moisture content of the ground and reaching a maximum when the soil is completely saturated. At the present time, a schematic representation,[8] according to which a layer of freezing ground with increasing volume exerts a pressure on the overlying (frozen) and underlying (unfrozen) layers of the ground, thus raising the former, is taken as the design principle for determining the tangential forces of heaving. When a foundation is present, the freezing layer of the ground, being bonded to the foundation surface by the adfreezing forces, at first bends and then, as the depth of freezing increases and the heaving forces develop, it breaks away from the foundation surface (when the heaving forces exceed the adfreezing forces) and begins to slide along it. Thus, whereas at a certain distance from the foundation the heaving forces are equal to the weight of the overlying ground, near the foundation their magnitude increases and becomes equal to the limiting long-term shear strength along the surface of the foundation. Usually the tangential heaving forces reach this value on freezing to a depth of two-thirds of the total depth of the seasonally frozen layer.

The calculation for the resistance to the heaving forces is predicated on the condition that these forces do not exceed the load on the foundation plus the ground-foundation surface adfreezing forces prevailing in the region where it is sunk below the freezing layer. If the foundation is made in the form of an upright with an anchorage shoe, then when determining the anchoring forces, along with the adfreezing forces and the weight of the overlying ground, it is also necessary to take into account the pressure transmitted to the anchor by the heaving forces.

Several modifications of the foregoing system have also been proposed. In one case[35] the freezing layer of the ground is considered to be a rigid body sliding along the surface of the foundation, and the tangential forces of heaving are taken to be equal to the contact forces of friction at the foundation-ground interface developing during sliding. In an other system[16] the freezing ground is considered to be an elasto-viscous body that bends during heaving and tries to "extract" the foundation. This system can be likened to a semiinfinite beam (of variable height equal to the thickness of the frozen layer) on an elastic Winkler base. At the free end of the beam, forces are applied that vary according to the height and are equal to the tangential heaving forces.

The basic problem in all cases is determining the design values of the heaving forces. Relevant tests were conducted in the laboratory and under field conditions. In some of them (the field tests) the heaving forces were determined by equalizing them with a load transmitted to the test foundation and increased gradually as the depth of freezing increased (the method was proposed by N. I. Bykov in 1940). Another method (V. F. Zhukov and S. S, Vyalov), which was first used by the author at Igarka in 1950, consists of transmitting the heaving forces of the foundation to a beam secured to two supports anchored in the permafrost in such a way that it is possible to determine the heaving forces by the bending of this beam or from the readings of a dynamometer. This test procedure has now been adopted as the basic one.[41] It can be used not only for determining the peak (design) values of the heaving forces, but also for following their development during freezing of the ground.

In recent years (1958-1971), a large number of field investigations have been performed by the Central Scientific Research Institute of Transportation Construction of the Ministry of the same name.[35] Field tests were also conducted (1964-1967) in Irkutsk.[16] Based on a summary of all of the available test data,[25] it is proposed that the grounds should be subdivided according to the degree of heaving and that the following differential values of the tangential heaving forces should be adopted: for strongly heaving ground, 1.0 kg/cm^2; for moderately heaving ground, 0.8 kg/cm^2; for slightly heaving ground, 0.6 kg/cm^2. However, in another paper[35] significantly higher test values of the heaving forces are adduced: for silty sulginoks (under the conditions pervailing at Skovorodino) they average 1.7 kg/cm^2. In a paper by Tsytovich,[47] (1973), based on a generalization of the experimental data, it is proposed that design values of the heave forces ranging from 0.8 to 3.0 kg/cm^2 be adopted, depending on the foundation material (wood or iron--concrete), the confluent pattern of the seasonally frozen layer, and the depth of freezing.

At the present time the question as to which values of the tangential heave forces should be included in a new normative document is being vigorously debated.

DESIGNING FOOTINGS AND FOUNDATIONS

The basic principles of design for footings and foundations are standardized in the Construction Norms and Regulations,[32] and they are discussed in detail in the manual on these norms.[8]

The calculations consist of estimating the thermal regime of the ground interacting with the installations (the thermophysical calculation) and estimating the resistance of the permafrost foundation to the loads exerted by the installation, having regard to the dynamics of the temperature regime and the development of rheological processes in the ground with time (the static calculation).

During construction entailing maintaining the ground in the frozen state (principle I) the tasks involved in the thermophysical calculations are as follows:

1. ascertaining the measures required to insure maintenance of the foundation in the frozen state and performing the thermophysical calculations for the devices needed for this purpose;
2. determining the thermal regime of the permafrost foundation and its variation with time;
3. selecting the calculated values of the permafrost foundation temperatures in order that they can be used in the static calculations;
4. determining (by calculation or experimentally) the depth of seasonal ground thawing in order that a subsequent calculation for foundation heaving can be made.

G. V. Porkhayev[36,37] developed a method of designing ventilated air spaces and cooling ducts for buildings with the floor on the ground. A modulus of ventilation equal to the ratio of the area of the ventilation holes to the building area is taken as the value determining the air space regime. The values of this modulus required to ensure that the foundation remains frozen are determined as a function of the temperature inside the building, the outside air temperature, the air temperature in the air space, the thermal resistance of the plinth course, and the mean annual wind velocity. The heat liberated by the sanitary-engineering pipelines is also taken into account. The design of the cooling tubes and ducts amounts to determining the depth to which the tubes are embedded, the thickness of the fills in which they are laid (under the floor of the building) and the spacing between them, the maximum depth of thaw of the fill, the volume of ventilating air, and the permafrost temperature that results from the cooling. V. S. Luk'yanov's hydraulic integrator can be used for the thermophysical calculations.

The static calculations for the foundations pertain to two limiting states--the bearing capacity (strength) of the permafrost foundation and the deformations. The essentials of these calculations (as determined by Vyalov) are as follows. The calculation for the bearing capacity of foundations used in accordance with the first, taking into consideration the reduction of strength that occurs during the creep process, amounts to determining the load which, under the given temperature regime, will not cause progressive flow of the soil during a preset period (the useful life of the structure) leading to loss of stability and disintegration. The values of the normative resistances of the frozen ground for which these conditions are satisfied are specified in the Norms[32] and also in the manual for use with them. These values are derived from the results of field and laboratory experiments performed at Igarka by the author of the paper. The calculation techniques are discussed.[8,19,20]

For solidly frozen ground the calculation using normative resistances ensures the absence of inadmissible settlement, and therefore the calculation for deformations is not made (except in special cases). Plastically frozen ground has a perceptible compressibility and is capable of significant settlement when loaded.[23] With this type of ground a calculation for deformation is therefore needed. In view of the creep phenomenon, this calculation amounts to determining the foundation dimensions in respect of which the settling will not exceed the limiting admissible value for the structure in question.

The thermophysical calculation of foundations used in the thawed (or prethawed) state, i.e., in accordance with the second principle, consists of determining the thaw zone beneath the structures and the development of this zone with time.

This problem was first investigated in an approximate form by N. A. Tsytovich. Present day calculation techniques are based on the following assumptions: With a definite approximation it is permissible (in solving practical problems) to assume that the transfer of heat to the ground is accomplished solely by convection. In this case the problem amounts to the three-dimensional Stefan problem, given the existence of preset boundary conditions on the free surface of the ground and within the confines of the structure. However, even in this simplified statement the problem is fairly complex. Therefore, certain approximations are used when solving it. The 1967 Norms[32] recommend the use of G. V. Porkhayev's[37] solution or V. S. Luk'yanov's hydraulic analog method. The first of these methods makes it possible to determine the approximate outline of the thaw zone under the structures at any given point in time, including the final steady-state position. A numerical solution of this problem on a computer is presented in a paper by Pal'kin.[34] The analog method enables consideration to be given to any boundary conditions (including those variable in time), the nonuniformity of the ground, and so on, and also makes it possible to study the dynamics of the process.

The static design of the foundations, made in accordance with the second principle, is made from the deformations: the highest value of the settlement and its differentials must be such that they do not cause deformations in the structure exceeding the limiting ones for the structure in question. The calculation for settlement is based on the two-term formula proposed by N. A. Tsytovich[47] which separately considers settlement due to thawing and settlement due to consolidation.

Whereas columnar or pile foundations are used for structures built in such a way that the ground forming the base is maintained in the frozen state, in the case of structures erected on thawing ground it is recommended that foundations in the form of strips, cross strips, or slabs be used. In designing these latter the combined action of the base and structure must be taken into account. This calculation technique is illustrated in the paper by Mel'nikov *et al.*,[30] in which methods of determining the reaction pressures of the thawed ground on the base of the foundation are presented. The thawed ground is considered to be a Winkler base, which corresponds to its actual behavior and which was checked experimentally. The curvilinear outline of the thaw zone along the length of the foundation is considered, the depth of thawing varying with time. Accordingly the bed coefficients of the Winkler foundation are assumed to be variable. The design diagram of the foundation amounts to that of an elastic beam to which a load is applied from above, while from below it is influenced by forces that are equal to the reaction pressures of the ground. The calculation on such a beam is by the methods of structural mechanics.

PILE FOUNDATIONS

The combining of a ventilated air space and pile foundations proved to be an exceptionally successful union, ensuring an increased bearing capacity of the foundations and an enhancement of their operating reliability when building with preservation of the permafrost. These circumstances, and also their production under factory conditions, have led to the extensive use of piles, which have now become the commonest type of foundation when building according to the first principle (maintaining the frozen state of the foundation). Pile foundations began to be used in the USSR in the 1940's and 1950's

(Igarka and Yakutsk). Wooden piles had formerly been used. The late 1950's and early 1960's marked the conversion (at Noril'sk and Yakutsk) to factory-built reinforced concrete piles with prefabricated or monolithic high grillages. This type of pile is currently the most widely used. Metal piles are used only in special cases.

The piles are sunk after predrilling of holes filled with mud or, alternatively, after prethawing of the ground by a steam "needle." The latter method is now outmoded, although it is still being used in Yakutsk, where the ground has no coarsely clastic inclusions and the low temperature of the permafrost insures adfreezing of the steamed slush to the pile. All of the well-known shortcomings of the method (its non-industrial character, the resulting pollution of the site, the difficulty of centering the pile, and the protracted nature of the freezing process) raise the question of whether the method should not be completely discarded.

In the presence of plastically frozen ground, the piles are driven into a small-diameter leader hole or directly into the permafrost. This method, which was first used at Igarka in the 1950's and was later refined by Targulyan[44] and Yeroshenko,[21] derives from the premise that the kinetic energy of driving is converted into thermal energy, with the result that the frozen ground thaws and the driving is facilitated. The pile-driving techniques are discussed in a paper by Markin and Targulyan.[29] It is noteworthy that the drilling of the holes in the permafrost is the most labor-intensive aspect of the work on the installation of pile foundations. One way of improving drilling efficiency is replacement of the impact-cable method currently used by the impact-rotational and leader method. Thermal and thermomechanical drilling methods are also being tested.

The methods of installing pile foundations in permafrost were discussed in papers by Goncharov *et al.*,[15] Dokuchayev and Markin,[20] Yeroshenko,[21] and Targulyan.[44] The design of the piles is initially predicated on the ultimate shearing strength of the ground along the lateral surface of the pile (the ultimate adfreezing strength) and the ultimate resistance of the ground to the normal pressure under the end of the pile. The calculation technique and the calculated values of the ultimate strengths themselves were proposed by the author and included in the Norms.[32] The method is discussed in detail in papers by Vyalov *et al.*,[8] Dokuchayev,[19] and Dokuchayev and Markin.[20] This method is the calculation for the bearing capacity, but with solidly frozen ground at the strength values stipulated in the Norms[32] it also satisfies the conditions of the calculation for deformations inasmuch as with a load not exceeding the indicated ground resistance values, there will be no deformations due to sustained creep and the overall settlement of the piles is required. It is important to note that by making the calculation for creep settlement of the piles, based on the condition that during a preset period (the useful life of the structure) this soil settlement will not exceed the limiting value, the bearing capacity of the piles can be increased in comparison with that calculated by the bearing capacity. This calculation can be predicated on the following relation proposed by the author:

$$S = U\left[\frac{A}{1+|\theta|^{\lambda}} \cdot \frac{N}{Uh^{t}}\right]^{-\alpha \frac{1}{m}},$$

where S is the settlement of the pile; N is the load; θ is the mean annual ground temperature [C] without a minus sign; t is the time; U is the perimeter of the pile; h is the depth of its embedment in the permafrost; A, λ, m, and α are the parameters determined from pile tests.

Some data on tests of piles for settlement are adduced in the paper by Markin and Targulyan[29] in which a study is also made of the important problem of the performance of piles under horizontal load.

Piles have proved to be an exceptionally efficient type of foundation for residential and administrative buildings, including 12-story residential buildings with loads of up to 100 tons on each pile (8-12 m long).[26]

DEEP SUPPORTS

The expansion of industrial construction in the North and the necessity of erecting large, wide-span industrial structures designed to carry heavy loads has called for the use of foundations with an increased bearing capacity, i.e., of up to 600-700 tons and more per support. It is suggested that deep supports in the form of piles, tubular piles, and pedestal piles should be used for such foundations. A schematic representation of a pedestal pile used at Noril'sk is given in Figure 9.[48]

The use of deep piles under a heavily loaded structure is especially efficient in cases where a site overlain by basement rock is within range, it being considered expedient to rest the piles on the rock to depths right down to 50 m. In such cases, the supports used at Vorkuta are in the form of circular-cylindrical pits. At Noril'sk they are in the form of tubular piles measuring 600 mm in diameter, prefabricated end-bearing piles measuring 40 × 40 cm, and piles cast in place. When using end-bearing piles, thawing of the ground was permitted. The problems of installing deep supports are discussed in papers by Kolyada *et al.*[26] and Zhukov and Ponomarev.[22]

THE SPECIAL CONDITIONS AFFECTING CONSTRUCTION IN PERMAFROST REGIONS

Ground Ice as Foundations for Buildings

The extensive occurrence of ground ice in the North makes the question of its use as footings for buildings one of extreme urgency. Until

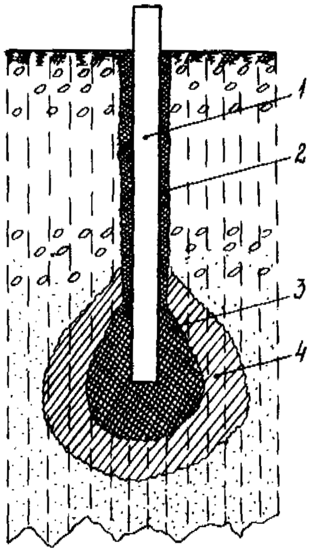


FIGURE 9 Pedestal pile: 1--reinforced concrete pile installed in a drilled hole; 2--injected concrete casing; 3--pedestal; 4--ground hardened by injection.

recently, sites with inclusions of ground ice were not used for construction. However, the layout requirements of areas undergoing development, the technological conditions governing the operation of the structure, and other such circumstances are posing the problem of the necessity for developing such sites and, accordingly, of working out methods of installing foundations on ground ice. The studies undertaken[9] have established that ice can serve as a sufficiently reliable footing, provided that specific measures are foreseen and that the requirements of calculating for creep are satisfied.

As is well known, in contrast to frozen ground, ice does not have a stress-rupture strength and is capable of continuous flow under load. Therefore, the calculation for deformations of foundations based on ice is a decisive one.

The calculation is based on the existence of two critical stress values in the ice. Until the attainment of the first of them the deformations develop so slowly that they are almost negligible. The values of the resistance of the ice to shear (its shear strength) along the lateral surface of the piles R_{shear} and to the normal pressure beneath the end of the pile or beneath the foot of the foundation R, which correspond to the first of the aforementioned critical stresses and on the basis of which it is possible to determine the permissible load on the pile or column foundation, have been established experimentally. These values are as follows at ice temperatures of -1.0°, -1.5°, -3.0°, and -8°C.

R_{shear} (kg/cm^2) 0.25, 0.35, 0.4, 0.55

R (kg/cm^2) 0.4, 0.9, 1.3, 2.3.

The other critical stress corresponds to the delimitation of the deformations (of a viscous flow of ice) to deformations developing at an approximately constant rate and to those developing at a continuously increasing rate. As the latter type of deformation cannot be permitted for the footings of a structure, the second critical stress is taken as the upper stress-rupture strength delimiting the magnitude of the load on the foundation. The calculation for deformations is made for a load, the magnitude of which is between these two limits.

In all cases, direct transfer of the load to the ice is not permitted (Figure 10). The pile foundations are installed in large-diameter drillholes, with the gaps filled with a sand-lime mud. In this case the calculation is made for the lesser of the two shearing strengths, namely, the shear of the pile with respect to frozen mud and the shear of the frozen mud with respect to ice. The columnar foundations are installed in a fill of frozen sand and in the calculation the footing is considered two-layered. Structures resting

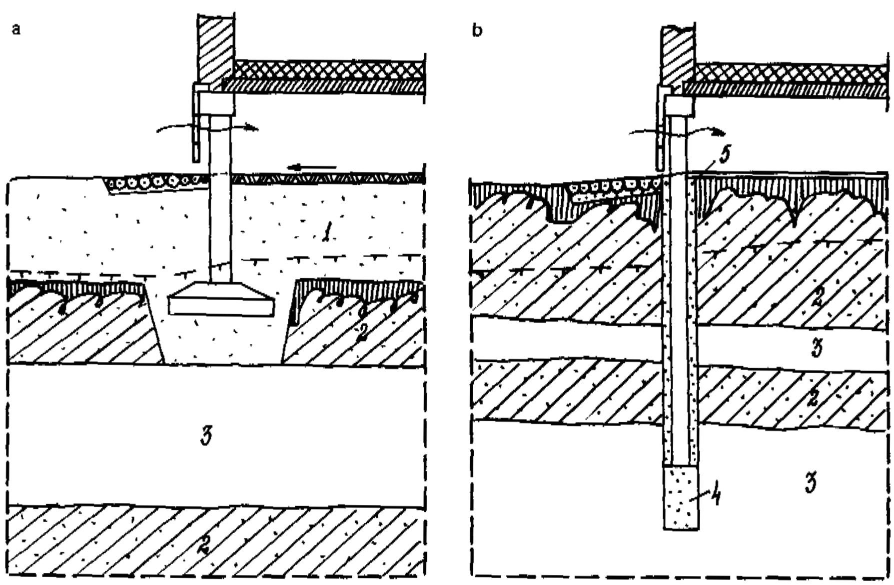


FIGURE 10 Layout of (a) columnar and (b) pile foundations on ground ice.

on ground ice have already been erected and are in normal operation at Tiksi, Amderma, Cherskiy, Pokrovskiy in the Yakut ASSR, and elsewhere (Figure 11). Based on the study by Vyalov *et al.*,[1] at the present time instructions are being published relative to the planning and design of the footings and foundations of buildings and structures on ice-rich permafrost, including ground ice.[45]

Saline Permafrost

On the Arctic Coast, in Yakutia and in many other places in the Far North, saline permafrost is widespread. The salinity of this ground can reach 1.5 percent and more. Such ground has a number of specific properties that must be taken into consideration during construction. Foremost among these is the lowered (right down to -5°) freezing point and the correspondingly higher content of unfrozen water. Hence, the strength of saline frozen ground is appreciably lower and the deformative properties are higher than in nonsaline ground (at the same temperature). For example, a salinity of 0.5, 1.0, and 1.5 percent lowers the resistance of frozen suglinok (at -2°) from 8 kg/cm^2 (nonsaline ground) to 4.3, 2.4, and 1.7 kg/cm^2, respectively.

Studies by Velli and Karpunina[6] and Velli[4] have established the values of the basic strength characteristics of saline ground, which are necessary for designing foundations (the resistance to normal pressure and the shear resistance along the lateral surface of the pile). These values, obtained for different types of ground and at differing degrees of ground salinity and temperature, formed the basis for the "Instructions."[45] In order to increase the bearing capacity of the piles, the "Instructions" recommend the drilling of holes that have larger diameters than the size of the pile and their filling with nonsaline sand-lime mud in a similar manner to that indicated in Figure 11 for piles in ice. In this case the calculation is performed for the shearing strength of the pile with respect to the frozen nonsaline mud (having increased strength) and the mud with respect to the surrounding saline ground (having regard to the increased adfreezing area of the mud).

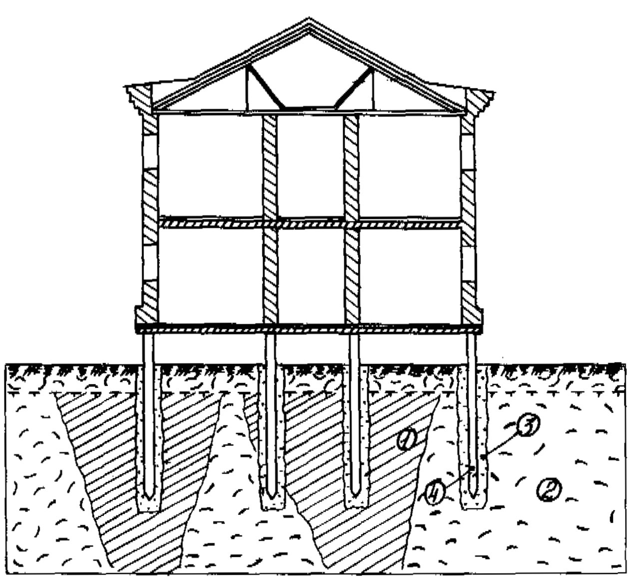


FIGURE 11 School at Pokrovskiy (Yakut ASSR) constructed on ground ice; 1--ice; 2--suglinok; 3--sand mud surrounding piles; 4--piles.

REGIONAL PECULIARITIES OF CONSTRUCTION

The general principles of using permafrost as footings for structures are valid for the entire permafrost region. In this connection, however, consideration must be given to the regional peculiarities of the area in question. First and foremost, it is necessary to distinguish areas with continuous thick low-temperature permafrost; areas with thin, high-temperature permafrost; and, finally, areas containing permafrost islands. The first of these regions is the most favorable in the sense of ensuring stable construction. The groups forming this region are solidly frozen and, as a rule, construction on them is accomplished while maintaining the frozen state of the foundation. The provision of an air space in this zone not only ensures maintenance of the natural temperature of the permafrost, but even a lowering of it. An example of such a region is Yakutsk.[7] The bulk of present-day construction (since the 1930's) has come to be based on the principle of maintaining the ground in the perennially frozen state; with rare exceptions, piles were used as the foundations. Initially, the air space was made low, but in view of a number of shortcomings of this design, conversion has been made to air crawl spaces (often excessively high) and at times even to the construction of a technical (engineering) half-story or an unheated story with the administrative services located in it.

Another example of a region with low-temperature ground is Noril'sk,[26,27,48] although the geocryological conditions of this region are more varied. Along with a thick 400-m series of permafrost with temperatures of -2° to -4° and less, areas are encountered with thin high-temperature frozen ground that has a discontinuous distribution both in plan view and vertically. The basic building principle adopted at Noril'sk, as at Yakutsk, is that of maintaining the ground in the permafrost state, which is achieved by constructing high air spaces. During the initial period of construction at Noril'sk, columnar foundations were used, but later the builders converted to using piles, on which all of the residential buildings are now erected, including the multistory buildings. At the same time, the presence of basement rock at attainable depths has made it possible to found many of the Noril'sk structures and, in particular, the majority of the industrial installations on rock, using both ordinary and deep (to 50 m) supports (Figure 12).

A standard example of a transitional zone between low-temperature and high-temperature fro-

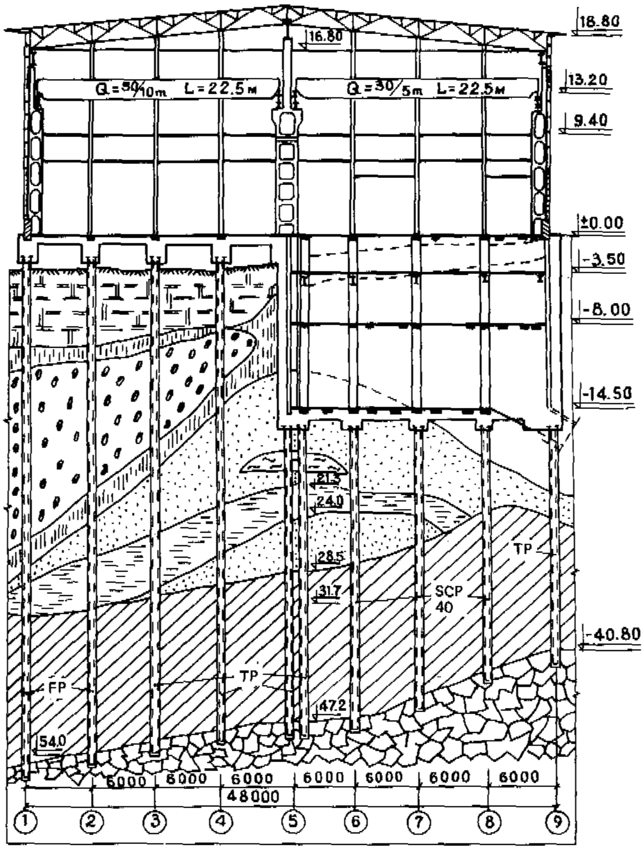


FIGURE 12 Industrial building at Noril'sk founded on deep supports cutting through a series of sedimentary ice-saturated frozen ground and resting on rock (basalt). Types of support: TP--tubular piles 60 cm in diameter; FP--filling (built in place) piles; SCP-40--40 × 40 - cm^2 chamfered piles.

zen ground is the area comprising the town of Mirnyy,[5] where ground exists with a temperature ranging from -3.5° to 1.0° and higher. Nevertheless, even under these conditions the principle of maintaining the frozen state of the foundations is taken as the basic construction principle.[17,31] In this area as well, for sites with low-temperature frozen ground the provision of a high air space (from 0.5 to 2 m) not only ensures maintenance of the ground in the permafrost state, but even a lowering of its temperature in comparison with the value. During the initial years of building at Mirnyy the footings beneath industrial and civil buildings were of the columnar type and embedded to a depth of 3-5 m. Subsequently the Mirnyy builders likewise converted to using piles. Reinforced concrete piles up to 14 m long installed in previously drilled holes are used.

Footings that take horizontal, dynamic, and heavy loads imported by nonequilibrium equipment are made in the form of a bunch of "frozen" piles with a high monolithic grillage.

As for industrial structures entailing a wet technological process and heavy floor loadings, they are usually placed on sites with comparatively competent rocks: on trap dikes or dense limestones.

A much more complex situation obtains with regard to the development of the Mirnyy sites containing high-temperature frozen ground. Although observations have established that even the normal development of such sites by erecting buildings with a high air space, constructing roads, clearing the snow, etc., lead to a lowering of the ground temperature (from -0.5° to -1°), the cooling effect is exhibited only after a prolonged time interval. Therefore, the builders of Mirnyy have elected to use a more efficient procedure, cooling the ground by the "thermopiles" of the Gapeyev-Makarov design.[3] These are the reinforced concrete piles with a built-in closed heat-exchange system in the form of two concentric tubes filled with kerosene. A number of buildings have been erected on such piles, and observations have shown that in one winter they resulted in the ground temperature of the foundations being lowered from -0.2° to -15° during the winter and to -2° in the fall.

A characteristic (and highly favorable) feature of the buildings of Mirnyy and other cities and settlements in western Yakutia is the care that has been taken in the engineering preparation of the building sites. This consists in providing a gravel cover 1 to 1.5 m thick in the zone affected by the construction machinery. On completion of the pile-sinking operations the fill is leveled in accordance with the planned elevations and the concrete floor of the ventilated air space of the future building is laid over it, thereby providing protection against surface water getting into the ground and making possible more rapid cooling of the ground through the air space. In the presence of heaving ground, the latter is replaced by gravelly material over the entire depth of seasonal thawing.

The areas comprising the cities of Vorkuta and Igarka serve as an example of a region with discontinuous permafrost. The permafrost conditions in these areas are highly varied and complicated. The depth of the permafrost table varies from 0.5-2 to 10 m and more, with the result that in some places this surface merges with the layer of seasonal freezing. In others, it is overlain by a layer of thawed ground. Thus, in the city of Vorkuta itself the thawed ground takes up 80 percent of the entire area, and in the vicinity of the city, 50 percent. In this area the permafrost has a temperature ranging from -0.2° to -1° and only in isolated places does it drop to -2.5°. The depth of seasonal freezing and thawing is from 0.5 to 2-3 m. The most typical ground consists of silty argillaceous material and, more rarely, of silt and sand. The ground is heavily ice-saturated, and, on thawing, considerable settlement occurs. Such permafrost conditions greatly complicate construction activity in these areas. For areas with discontinuous permafrost or where this ground occurs in the form of individual lenses and islands, thawing is used. At Vorkuta this method was first used in 1968 when building the two above residential buildings. Subsequently, the prethawing technique was used at Vokuta for various structures, including large industrial complexes.

At sites with continuous and shallow permafrost, the method used is that of maintaining the permafrost state of the foundations and providing a high air space and pile foundations (driven into leader holes or directly into the plastically frozen series). The construction of buildings on fills of crushed rock has become very common.

When the rocks are present at the depth attained, the foundations are laid on these rocks. In such cases, for civil buildings pile foundations are used and for industrial buildings designed to take heavy loads--deep support pits.

The Magadan Oblast is highly complex and varied with respect to its geocryological conditions where all of the main types of permafrost are encountered: permafrost islands with temperatures from 0° to -1.5°; discontinuous permafrost with a ground temperature of up to -3.5°; continuous permafrost with a thickness of 150-500 m and temperatures ranging from -4° to -9°. In the first zone, to which, in particular, the Magadan region belongs, construction is by either the thawing of frozen ice lenses or by the cutting through them and laying the foundations on the underlying (at a depth of 5-15 m) solid ground of the conglomerate, sandstone, or granite type. In the latter cases the most useful type of foundation is the end-bearing pile with a pedestal (the camouflet pile).

The methods of construction in the second and third zones depend on the type of ground and the shape of the structure. In areas where the ground consists of shales and sandstones that have a low compressibility during thawing (0.665-0.0001 cm/kg), it is permissible to build and allow uncontrolled thawing to occur during use of the structures. But in zones of intensive tectonic dislocations these rocks disintegrate and are characterized by a high ice content, with the result that considerable settlement occurs (up to 4 cm/m thickness). The residuum and detritus of schists and sandstones represented by ice-saturated suglinoks and supesses result in even greater settlement during thawing (up to 7 cm/m). In such cases efforts are made to erect the building while maintaining the frozen state of the foundations. However, if this cannot be done for structural or technological reasons, then prethawing is resorted to. It should be noted that in the Magadan Oblast hydraulic thawing is used for the most part. The builders have borrowed this method from the miners.

In the mountain relief zones and primarily in the mountain river valleys, gravel beds (it is these areas in particular that the shingle deposits were first to be developed) are widespread. As a rule, these deposits are highly subject to ice-heave and when thawing they settle to a depth of 4 cm/m. The shingle is characterized by high water permeability and therefore their maintenance in the frozen state is impossible in the event of flooding. For these reasons the building of industrial structures entailing a wet technological process is done on such ground by prethawing it. It should be noted that frost heave is even typical of certain rocky permafrost and that cores of frozen andesite or other rocky material that give the appearance of being strong externally break down into gravel upon thawing. This ground can settle up to 2 cm/m during thawing. Such pseudo-solid rocks are highly typical of permafrost regions (they are encountered, for example, in western Yakutia, Trans-Baikal, and so on). A cautious approach is necessary when using such ground as foundations for structures without maintaining it in the frozen state.

THE VARIOUS FORMS OF CONSTRUCTION

It is known that, in many cases, unforeseen deformations of buildings erected according to the principle of maintaining the ground in the permafrost state are due to the faulty installation or incorrect use of service mains. Therefore, a great deal of attention must be devoted to the procedure followed in laying piping systems, leading them into buildings, etc. Using as examples Noril'sk and Mirnyy, we will consider how sanitary engineering mains are laid when building according to the principle of maintaining the ground in the permafrost state.

Four methods of laying service mains (water lines, sewage and heating systems) are possible:[1,26,50] directly in the ground, in underground open and closed sewage conduits (utilizer ducts), along the surface of the ground, and underground. At Noril'sk all four procedures have been tested, and on the basis of many years of operating experience the people of Noril'sk have arrived at the following conclusions: The laying of heat-releasing service mains (heating and sewage systems) directly in the ground is inadvisable, inasmuch as in this case thawing of the ground surrounding the mains, made deeper by the rapid runoff of rainwater along the length of the thawed zone, promotes degradation of the permafrost in areas under development and on running the lines near buildings (in particular, at connection points) it causes thawing of the ground under the foundations. In addition, with this method of laying, nonuniform settlement of the ground takes place with thawing around the pipeline, which, on the one hand, leads to surface cave-ins, spoiling of the roads, etc., and, on the other hand, to frequent damage to the line itself. Finally, the servicing of deep pipelines is severely hindered; laying the water line and heating systems above the ground on trestles serves as a hindrance to use and freedom of movement in the area under development and interferes with the architecturally esthetic appearance of the city. This method does not solve the problem of laying service mains.

Laying the systems in closed ducts is free of some of the above shortcomings, but it hampers use of the lines. In addition, with this procedure the installing of reliable thermal insulation is complicated, which often leads to thawing of the ground and also to thawing around the duct; the installing of two-stage open manifolds (utilidors), which are widely used in Canada and the United States, is the most efficient method. All forms of engineering systems are laid in them, including electric cables, communication lines, and so on (Figure 13). In the event of damage the water is diverted through a manifold to a point outside of the area being developed. For this purpose a sewer is laid (along the axis of a street) with the appropriate gradient. The connections are realized through insulated diversion pipes (single-stage manifolds). The laying of lines in two-stage manifolds is done at Noril'sk in densely built areas of the city. The high cost of the manifold (made of prefabricated reinforced concrete blocks) is made up for by the convenience of operation and by the assurance that the ground surrounding the pipes is maintained in the frozen state. Inasmuch as observations have shown that in the region where the two-stage manifolds are laid, as a result of winter air circulating through them a 3°-4° lowering of the mean ground temperature took place (at the bottom of the manifold). At the same time the Noril'sk builders contend that in the relatively undeveloped areas of the city, in the suburbs, the settlements, and so on, surface laying of all of the heat liberating lines is justified. In this case the lines with the appropriate thermal insulation are laid on sliding supports that lie on wooden cribbing 0.5 m high and more or on prisms made of gravel, slag, etc. Two piles serve as fixed feet. While we are in agreement with the conclusions of the Noril'sk people regarding the economy and operating convenience of this type of laying, at the same time it should be noted that the architectural appearance of the settlement is by no means enhanced by such lines.

In all of the methods of laying service mains, their internal separation is effected by suspending the piping from the plinth course of a ventilated air space.

According to the findings of G. L. Gomelaue' and I. Ye. Gur'yanov, the service mains at Mirnyy and other settlements in western Yakutia are laid by two methods. In major settlements such as Mirnyy all of the systems are laid in prefabricated reinforced concrete manifolds similarly to the way that this is done at Noril'sk. The manifolds are laid to a depth of 2.5 to 5.0 m and measure from 1.8 to 2.5-3.3 m in cross section. The combined laying of the heating systems is confined to the manifolds 3.3 m in cross section. Inasmuch as the air temperature in the manifold is not controlled (although the manholes are opened in the winter), in several areas not far from Noril'sk thawing not leading to deformations of the manifold was observed.

In the settlements of Aykhal and Udachnaya the architectural layout provides for the joining of individual buildings by covered walkways and laying of the service mains is accordingly confined to these walkways.

HYDRAULIC ENGINEERING CONSTRUCTION

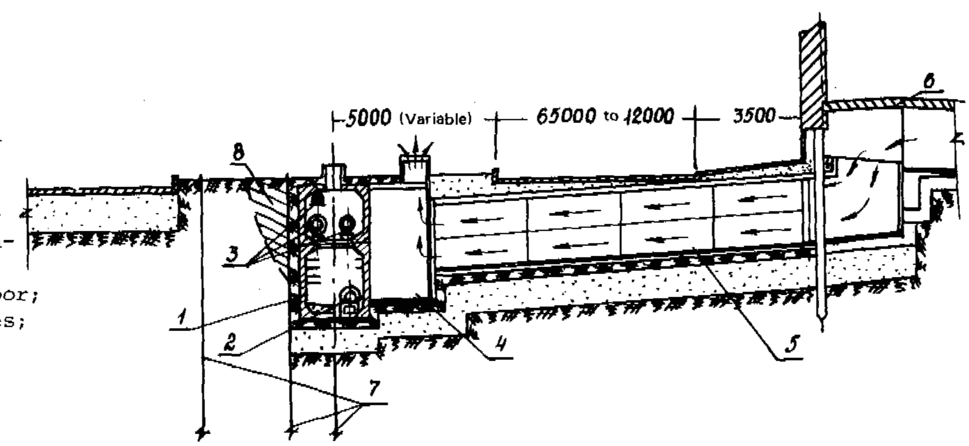
The building of hydraulic engineering structures in permafrost is a highly complicated and responsible problem--both on account of the uniqueness of the structures in terms of volume cost, and service life and because of the pronounced effect of the structures on the environment.

By now, considerable experience has been gained in the Soviet Union in the design and construction of hydraulic-engineering structures in permafrost, and efficient procedures have been worked out for the building of such structures.[2,13,47]

At the time of working the building of two large electric power plants with high-head dams made of local materials is nearing completion--the Vilyuy Hydroelectric Power Plant with a dam 74 m high and a crest length of 600 m, the construction of which is nearly finished, the first stage having been handed over for use in 1967, and the Khantay Hydroelectric Power Plant, with a dam 65 m high and a crest length of 420 m. A rock-fill dam of the Kolyma Hydroelectric Power Plant measuring 124 m high and 790 m long is being designed. The construction of a number of medium- and low-head dams from local materials is under way.

The building of the dams is based on same two cardinal principles governing the use of frozen ground that are observed in civil and industrial construction: namely, maintaining the ground in the frozen state (the "cold" version to use the terminology of the hydraulic engineers)

FIGURE 13 Standard design of main sewer and connection to a building (Noril'sk): 1--two-stage main sewer; 2--sewerage; 3--water supply line and heating system; 4--basement of connection; 5--connection to the building; 6--plinth course floor; 7--monitoring temperature holes; 8--possible thaw zone.



and allowing it to thaw (the "warm" version). However, when building dams on permafrost it is necessary not only to insure stability of the base and of the dam itself, but also to foresee the measures that will be needed to achieve the preset percolation regime.

Low- and medium-head earth-fill dams are usually built by the method of preserving the frozen state of the base and creating a frozen impervious core, although in some cases their building is effected by the "warm" version. When erecting high-head rock-fill dams as a rule the alignments of the hydraulic-engineering structures are selected in such a way that their base is on rock. In this case the necessity for maintaining the ground in the frozen state no longer exists and freezing the core of the dam proves to be no more effective than the construction of an ordinary impervious curtain. Therefore, in the USSR high-head dams for permafrost regions are designed according to the "warm" version.

When maintained in the frozen state, low- and medium-head earth-fill dams can be built on any type of foundation, including those which are subject to settlement.

However, it must be taken into account that in the presence of a reservoir, preservation of the frozen state of the ground is possible only in the central part (including the core) and in the downstream slope of the dam. Therefore, if beneath the upstream slope the base is made up of ice-rich settling soil, and especially if it is intruded by ground ice lenses, then it is recommended that this ground either be replaced by nonsettling ground or that it be prethawed.

The freezing of the impervious core of the dam is effected by one of the following methods:[13]

(a) layered freezing of the body of the earth dam during filling in winter using natural cold;

(b) freezing the downstream face by natural methods after filling it using winter cold, the provision of "cold wells" of the type designed by M. M. Krylov, etc. In this case the freezing is achieved during operation of the dam;

(c) artificial freezing of the body of the dam and its foundation after the building of the latter (but before raising the headrace), using different types of cooling units.

The third of these procedures has become the most widespread. Fans that operate in the winter and feed cold air through ducts into the body of the dam are used as the cooling units. Less frequently, artificial freezing is done with brine used as the cooling agent. The question of using self-regulating coolers of the Long or Gapeyev type, which was mentioned earlier, is being investigated.

Cooling of natural cold gives a good effect for dams with a head of up to 25 m. With a higher dam it has been found to be more difficult to provide for its freezing by ordinary means. In this case it is necessary to use the two-stage cooling unit and to freeze the base of the dam from a cooling tunnel.

In a number of cases it proves sufficient to effect artificial freezing only during the initial years of operation of the structure, after which a stationary temperature regime of the dam is usually established. Forced cooling is required only periodically. Sometimes, however, artificial cooling by cold winter air must be carried out systematically.

Highly vulnerable sites in hydraulic engineering complexes with earth dams are the weirs and water discharge points. A number of examples can be cited in which the unsuccessful design of these components has resulted in deformations and even disintegration of dams.

By following the writings of Tsytovich *et al.*,[46] Gluskin and Ziskovich,[13] and Bianov,[2] we will consider some examples of design solutions for low- and medium-head dams on permafrost.

The first earth dam, the frozen state of which was achieved by the freezing of soil brought in during the winter and sustained by natural cold, was erected as long ago as 1792 in Petrovsko-Zabaykal'sk. This dam, measuring 9.5 m high and 320 m long, has endured for 137 yr in its original form. Another similar dam 7 m high and 90 m long was built in 1941 near the village of Ege-Khaya (in the northern part of the Yakut ASSR). Although this dam was filled in the summer, its subsequent freezing from the base (the permafrost temperature was -7° to -8°) ensured "cold" version operation of the dam. In 1961, for the purpose of supplying water to the city of Mirnyy, a dam measuring 3 m high and 285 m long was built on the Irelayakh River. By way of an experiment, a frozen core of local earth materials was first built in this dam. The fill of the core was accomplished in the winter and involved flooding with water and freezing. To provide for more rapid freezing of the dam, it was furnished with pits, which were encovered in the winter and thermally insulated in the summer. The data from the temperature observations revealed that the dam has remained frozen to this day. A dam in the Arctic (on the Portovyy Creek), which was built in 1942 and rebuilt in 1955, is also in good condition. This dam, measuring 7 m high and 190 m long, was erected by the freezing of compacted filled soil and subsequent cooling from the direction of the downstream slope. The upstream slope and the crest of the dam were prevented from thawing (by thermal insulation, etc.).

The dams at the Poselkovyy Springs[13] are an example of freezing of the body of the dam by drilling M. M. Krylov type "cold wells" (on the downstream slope).

As stated earlier, the most widely used method of building dams according to "cold" version is their artificial freezing. One of the first dams built by this method is the earth dam measuring 10 m high and 130 m long on Dolgyy Lake, which was erected in 1941-1943 for the purpose of supplying water to the city of Noril'sk. The freezing of the body and the base of the dam was initially achieved by the use of brine circulating through two rows of vertical wells furnished with freezing pits. For various reasons, however, the freezing turned out to be inefficient and was replaced by forced ventilation using cold winter air passing through holes running along the crest of the dam and simultaneous equipping with cooling tunnels of the ice storage type designed by M. M. Krylov on the downstream face and their natural ventilation in winter through intake and exhaust shafts. The downstream slope was thermally insulated by a layer of sawdust. In accordance with a similar principle, a dam 7 m high was erected in the 1950's in the vicinity of Noril'sk on the Nalednaya River. The cooling of this dam was achieved by forced ventilation using winter air passing through holes located along the crest of the dam and by cooling of the surface of the downstream slope through natural circulation of winter air beneath a curtain built above the slope.

The history of the construction of the dam on the Myaunzhde River (Magadan Oblast) is not without interest. This dam, measuring 7.2 m high and 860 m long, was built during the period 1952-1959 and involved freezing of the body (including the core and its spurs) and base by forced ventilation using cold winter air passing through cooling holes arranged in a single row. This system, however (which is periodically actuated rather than annually), did not result in freezing of the dam and was subsequently supplemented by an artificial cooling system using brine passing through an additional row of freezing holes and by grouting of the fissured rock forming the base of the spillway.

A more successful example is the erection of two dams on the Irelyakh River and the Oyuur-Yurege River. The first of these, which was designed to supply water to the City of Mirnyy, is 21 m high, 8 m wide at the top, and 115 m long at the bottom. At the base of the dam there are permafrozen marl sands with interlayers of dolomites, limestones, and marls which are strongly (to 60 percent) ice saturated; lenses of ice 15 cm thick and more are encountered. The dam was filled with suglinok in order that the suglinok core would form a permafrost foundation cutoff. A layer of sand was poured over the sandy-silty body and a rock fill was constructed. A layer of peat was laid on the downstream slope and the crest of the dam. Thawed earth materials were used to fill the body of the dam. Freezing was done subsequently in order to build the core and cutoff of the dam and maintain the permafrost state of the base. The freezing apparatus consisted of a system of columns installed in wells ranging from 9.5 to 26.5 m deep and totaling 327 wells in all. The cooling was achieved by feeding cold air in the winter blown by 7 fans, each of which served from 16 to 30 wells. Three years after the cooling units began to function, a frozen curtain measuring 10 m thick was formed in the body of the dam, following which the operating time of the devices during the winter was reduced--they were switched on only at outside air temperatures of -20° to -25° and lower. Today, the dam, being in the frozen state, is functioning normally. A transverse section of the dam, showing the isotherms characterizing the temperature regime 2 yr after commissioning, is shown in Figure 14.

The dam on the Oyuur-Yurege River (western

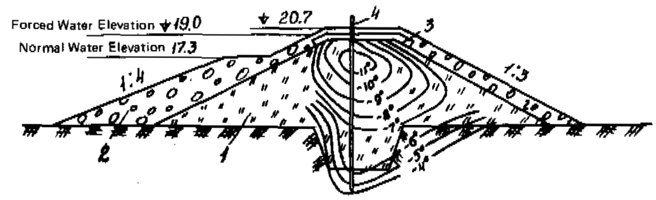


FIGURE 14 Temperature regime of the dam on the Irelyakh River: 1--suglinok; 2--rock fill; 3--peat layer; 4--freezing columns of the frozen curtain.

Yakutia) was built in 1972 by the same procedure as the dam on the Irelyakh River. The core and the cutoff of the dam were filled with local suglinok, the cutoff cutting through the loose ice-saturated deposits and resting on hard basement rock. The freezing system is analogous to that adopted for the dam on the Irelyakh river.

On both dams the spillway is in the form of an open canal in a bypass of the dam.

As already noted high-head rock-fill dams for the hydroelectric power complexes in the Far North are erected on a rock foundation in accordance with the "warm" version. In this case also, however, there are a number of characteristic features associated primarily with the organizing of the construction operations, especially the very large-scale earthwork undertaken under harsh climatic conditions and with the use of local silty-clay soils as fill for the body of the dam. The first experiment of this type was the Vilyuy Hydroelectric Power Plant with an output of 650,000 kW and incorporating a rock-filled dam with a total volume of 5,000,000 km^3.[2] The base of the dam consists of diabases, which on the banks are covered by a layer of loose deposits. An impervious curtain is provided by a coarse gravel and suglinok apron with a two-layered filter of crushed gravel. The apron connects with the rock base, a concrete slab structure, and a gallery from which the cementing of the talik below the bed was carried out (Figure 15).

A feature of the construction was the year-round discharge of suglinok soil, including during winter when outside temperatures dropped to -40°. The suglinok was prepared in a quarry in the early spring and placed in stockpiles where it became heated during the summer, accumulating reserves of heat. To prevent them from surface freezing, the piles were treated with calcium chloride or sodium chloride. Prior to working the stockpiles, during the winter they were thawed by electric heating; the heating units used consisted of aircraft jet engines. By adhering to a clearly defined technological procedure, a record was achieved in the laying of the apron material: up to 3,000 m^3/day and on the average 2,000 m^3/day.

The dam took a head of 50 m in 1967 and has since been functioning normally. It is interesting that although the dam was erected by the "warm" version, observations of the temperature regime indicate freezing of the dam from the direction of the downstream slope, which is due

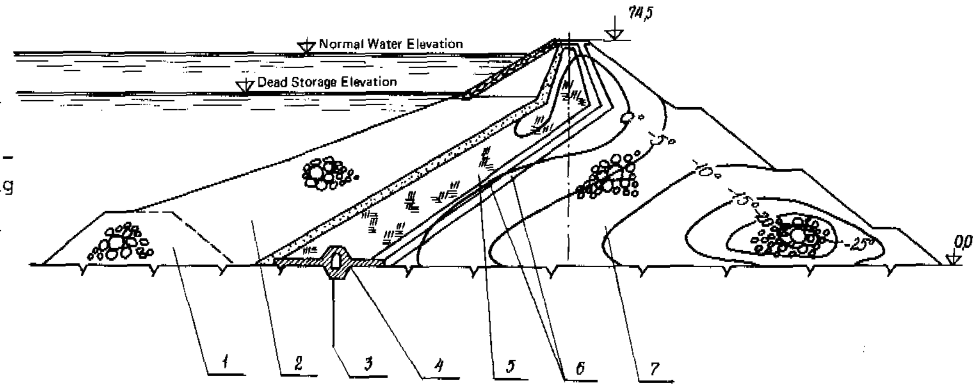


FIGURE 15 Dam of the Vilyuy Hydroelectric Power Station with isotherms drawn in. 1--Cofferdam; 2--rock ballasting of apron; 3--grout curtain; 4--grout cooler; 5--apron of suglinoks; 6--filters; 7--rock fill shell.

to rapid cooling of the rock-fill prism as a result of convection of the outside air through the fill pores. Figure 15 includes the isotherms plotted from the results of the temperature observations in the body of the dam.

The experience gained in the building of the Vilyuy Dam was used successfully in building the dam for the Khantay Hydrolectric Power Plant, which also has a core of gravelly-shingly suglinok soil.

PIPELINE CONSTRUCTION*

The discovery of important gas deposits in the North must have the effect of gradually shifting the main regions of pipeline construction into regions of widespread permafrost.

Already some large diameter trunk gas pipelines passing through permafrost regions have been built in the USSR. These are to be followed by even larger construction projects, including both gas and oil main pipelines laid over permafrost and spanning vast distances.

A main pipeline is a complex construction project consisting of the pipeline proper, the compressor stations, the electric power transmission lines and communication lines, the roadways along the route, residential settlements, and auxiliary structures. In a permafrost region the building and operation of a structure such as a trunk pipeline is a complex problem. This is because the routes for the pipelines pass through regions with diversified permafrost and geophysical conditions, the laying of pipelines is inseparably linked with their effect on the natural environment, and the building work is done in remote, uninhabitated areas.

A gas line was laid in a zone of active heat and mass exchange where the ground is characterized by a highly dynamic state of the freezing conditions. Even in the case where the temperature of the gas in the pipe is close to the ground temperature and its effect on the permafrost is slight, important environmental changes can occur during construction as a result of route-clearing operations, disturbance of the vegetation, the passage of the mechanized columns laying the pipelines, and so on.

* This section was written jointly with L. P. Semenov.

Even more significant changes in the permafrost conditions are caused by the thermal effect of pipelines transporting a product with a positive temperature. As a result, the temperature regime of the ground changes and there may even be an alteration in the sign of the mean annual ground temperature; the moisture content, density and thermophysical properties of the frozen and thawed rocks are also changed, as are the depths of seasonal freezing and thawing. All of this gives rise to variations in the rate of development of engineering-geological processes and the extent to which they manifest themselves.

With the thawing of silty ice-saturated permafrost, [gas] pipelines may "float." As a result of the thermal effect of the pipeline or the route preparation measures, thermokarst phenomena--settling, slides, and gullies--may develop. A simple change in the regime of the natural suprapermafrost runoff can lead to swamping of an area, the formation of thick naleds, and a number of other irreversible changes.

The list of problems arising in connection with the construction and operation of pipelines in permafrost regions could be continued, but the adoption of particular design solutions necessitates taking into account a large number of factors, not only those existing at the time of the survey and building of the gas pipeline, but also situations that could arise in the future as a result of the operation of the pipeline and the engineering development of the area.

The planning, design, and construction of pipelines in northern regions includes the following stages:

1. Choosing the optimal variant of the pipeline;
2. Engineering-geological and the permafrost surveys;
3. Choosing efficient design solutions, laying techniques, and working procedures;
4. Choosing efficient operating conditions and methods of controlling the thermal regimes of the pipeline;
5. Predicting changes in the permafrost processes and working out methods of controlling them.

In this report we shall avoid mentioning a whole series of other problems that are not directly connected with the problems of geocryology but are purely technological, for example, low-temperature welding, insulating the pipe, etc.

The above listed basic problems confronting the builders of gas pipelines in the regions of the Far North are discussed in papers by Biyanov,[2] Spiridonov *et al.*,[17] Dertsakyan *et al.*,[17] and Minkin.[31]

The planning and design of gas pipelines is usually carried out in three stages: in the first stage (the engineering and economic justification stage) the volume of capital investments is determined, the optimal variant of the route is chosen, and the fundamental engineering problems are solved. In the second stage (the contract design stage) engineering surveys are made and design solutions are chosen. In the third stage (the production of working drawings) the individual designs are precisely defined and illustrated in detail.

The choice of the optimal variant of the gas pipeline is based on the minimum capital expenditures on construction, operation, and maintenance, one of the most important indexes being minimum consumption of metal (in the form of pipe).

The concept of the optimal route, in addition to the normal elements such as minimum length, favorable ground, and hydrogeological conditions and the minimum number of crossings of water barriers and artificial structures, also includes special factors associated with the presence of permafrost: the mode of occurrence and temperature regime of the permafrost, the thickness of the layer of seasonal thaw, the moisture content and iciness of the ground, the possible development of permafrost processes.

A comparison of the competitive variants of the route is made on the basis of the landscape zoning method. Initially, a map of this zoning is compiled to a scale of 1:100,000, plotted on which is information on the geological structure, the permafrost conditions, and possible development of frost action, as well as data derived from air photo surveys and exploration, including drilling and geophysical operations and an examination of the landscape.

Inasmuch as the choice of the optimal variant of a gas pipeline is based on the minimum capital investments required for surveying, construction, and operation, it is recommended[17] that when choosing this variant all possible competitive design solutions should be considered. For this purpose the landscape zoning map is broken down into a number of sections that are typical with respect to geological and permafrost conditions, for each of which possible laying procedures are designated. Because of the large number of possible variants, the choice of the optimal one involves the using of elctronic computers. For computer calculation purposes, a digital model of the terrain is compiled, the landscape, geological, and permafrost characteristics and methods of laying being encoded for each section. This processing is done during the first stage of the planning and design.

In the more detailed stages the calculations can be repeated so that data on the permafrost-ground conditions obtained during the surveys can be taken into consideration.

There are also other methods of choosing the optimal variant, for example, a procedure in which the problems of selecting the otpimal route and deciding on the design solutions are resolved separately.[31] In this case, at the outset the optimal direction of the route is chosen, for which purpose a computer calculation is made wherein the permafrost, geological, hydrogeological, and other natural factors are analyzed. Later, depending on the complexity of the permafrost conditions along the selected route, the design solutions for the pipeline are decided upon. When using this method, the calculations call for an area permafrost survey to a scale of 1:100,000 and the use of high-speed computer techniques. The determining of the laying procedures, types of supports and other engineering solutions is done during the next stage of the design work after the completion of detailed engineering-geological and permafrost-ground studies on the selected route.

In the designing of pipelines a great deal of attention is devoted to permafrost engineering-geological surveys. In contrast to other projects, surveys for pipelines are made during all of the design stages. The inaccessibility of the northern regions and the short summer season make it almost impossible to carry out detailed engineering-geological surveys during the initial stages of designing pipelines. Therefore, during the first stage, small-scale permafrost maps are compiled on the basis of air-photo interpretation, the use of geophysical methods, and individual core samples. Although these maps are initially used only in connection with the technical-economic justifications, during the first stage the thermal effect of the pipeline on the permafrost in the zone immediately affected by it and the intensity of development of frost heave, settling, icing, and other cryogenic processes.

In addition, allowance must be made for the possible occurrence of groundwater, centers of groundwater discharge, and swamps.

The forecast of the changes in the permafrost conditions is made in two stages--an approximate and a detailed stage.

The approximate forecast involves ascertaining the direction in which the cryogenous and the engineering-geological processes will develop under the changed conditions. To arrive at a quantitative estimate of the manifestation of these processses when the dynamics of the temperature regime are known, approximation formulas are used. In performing the calculations the patterns of development of the permafrost, determined during the small-scale studies, must be taken into consideration. The requisite design characteristics are selected after a study of library materials and other reference literature. The results of the calculations are taken into account when determining the volume and the types of field studies, the depth of drilling and digging exploratory holes, the disposition of mineral workings and sites of stations, etc.

The temperature regime of the gas in the pipe has an appreciable effect on the choice of the design solutions and methods of laying the line. This is the criterion used to determine whether the sections of a gas pipeline are designated "warm" or "cold." Included in the first category are the sections of pipeline in which the gas temperature is above or equal to -1°C. The remaining sections are "cold."

Depending on their position in relation to the surface of the ground, gas pipelines are divided into underground, surface, and aboveground (Figure 16). At the present time the following pipeline design solutions and methods of preparing the foundations are widely used:

(a) with the underground method--laying within the layer of seasonal thaw or partially within the permafrost either with or without the preparation of a foundation and also semiburied laying in an embankment;

(b) with the surface method--laying the pipes directly along the ground (leveling pads) or in fills (with or without thermal insulation under the pipeline);

(c) with the aboveground method--linear laying of the pipe in slightly cambered sections on various types of supports--on piles, earth pedestals, etc.

Because of the variety of permafrost conditions along the route, a one-only method of laying is difficult to apply and usually the various methods are based on the surveys; a small-scale permafrost map is used as the basis for geological-geographic microzoning, and within each microzone a detailed study is made of the permafrost distribution patterns with respect to area and depth, the shaping of the temperature regime, etc. During the second stage of the surveys, based on the map of the microzoning, key sections are designated in which detailed permafrost studies along the proposed route of the pipeline will be made.

Engineering-geological and permafrost studies cannot be conducted continuously over the entire route on account of their high labor-intensiveness and the shortness of the season and also as a result of the compressed working periods. Therefore, a certain number of standard sections are normally isolated in the course of the survey work.

A distinctive feature of permafrost surveys, as emphasized in the paper by Spiridonov *et al.*,[43] is the interrelation that exists between them and the proposed design solutions of the pipeline. For example, the degree of detail and volume of the permafrost studies must be decided upon after taking into consideration the design solutions, which, in turn, are ultimately decided upon only after detailed permafrost studies have been made, having regard to possible changes in the permafrost conditions along the route during operation of the line.

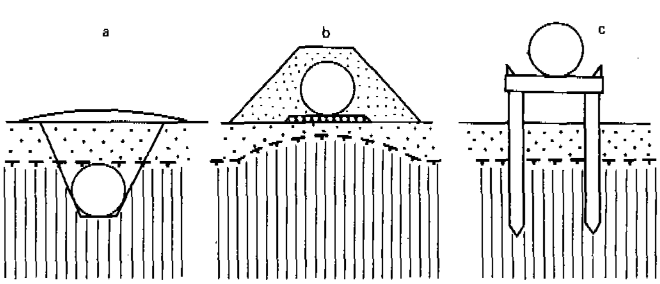


FIGURE 16 Methods of laying trunk pipelines on permafrost: a--underground; b--surface; c--aboveground.

While on the subject of predicting changes in the permafrost conditions along the route when the pipeline is in use, it is also important to emphasize the following possible changes that should be taken into account when designing and building a pipeline: Changes in the natural conditions along the route as a result of mechanized development of the area (removal of vegetation, leveling of the terrain, and changes in snow accumulation). Even if the entire pipeline is laid on the surface, at the approaches to river crossings, the pipe may have to be laid in the ground.

Underground laying is done within the seasonal thawing layer or with partial burial in the permafrost. Semiburied laying is also done; this entails embanking the pipe with earth. With underground laying of "cold" pipelines, the basic problems that must be solved are the effect of the forces of frost heave, the temperature induced deformations of the pipeline, and the origination of additional stresses in it as a result of adfreezing of the ground to the deformed pipe. With "warm" pipelines the basic problem is thawing and differential settlement of the frozen ground and the resulting deformation of the pipeline, settlement of the ground surface, and other disturbances of the natural relief. Until now, underground laying has been confined to "cold" sections and to those instances in which the route has passed through an area with relatively undisturbed relief and has not crossed a large number of watercourses, given a comparatively low level of groundwater, the presence of sandy and supessy soil with a low ice content, and no inclusions of large ice lenses and wedge ice. When building underground gas pipelines it is necessary to take into consideration the intensity of their thermal effect on the frozen ground and, on the other hand, the mechanical influence of the thawing ground on the pipeline, having regard to the possible erosion of the pipeline foundation as a result of a disturbance in the natural regime of the suprapermafrost water.

Surface laying is done in terrain with undisturbed relief and no evidence of thermokarst, frost mounds, landslides, etc. Its use without an embankment is recommended only in the "cold" sections of the route. Its use in the "warm" sections normally leads to thawing and settlement of the ground. It is therefore recommended that straight sections of pipeline be laid on a fill of soil or on a layer of moss and that thermal insulation be placed beneath the pipe. The sections of a surface pipeline used for the compensation of length changes are usually elevated on supports above the surface of the snow cover.

Laying in a fill is applicable in sections consisting of soil with sufficient bearing capacity after thawing or in "cold" sections. The method is frequently used to effect the transition between above ground and underground sections.

Aboveground laying is applicable whatever the surface topography (with the exception of floodplains, which are inundated during the debacle) and whatever the gas temperature. The pipeline is positioned on pile, surface, ground, and other supports. The type of supporting structure--roller, sliding base, etc., depends on the method of compensating for longitudinal deformations of the line caused by the air or gas temperature, and by the internal pressure. Depending on the design solution, pipelines may be with or without compensation for the longitudinal deformations. For example, some types of design solutions provide for partial or periodic adjustments in the summer.

Until recently, when building trunk pipelines on permafrost the aboveground and surface methods of laying have been by far the most common.

However, a trend towards the more extensive use of the underground method of laying is apparent; added to this is the possibility that with an increase in the pipe diameter and, consequently, in its rigidity, the traditional approach to selecting the method of laying may have to be revised. Whereas previously, in the planning and designing of pipelines the method of laying was adapted to the local permafrost conditions and it proved possible to lay the aboveground line so as to fit the surface topography, in future the method of laying will be determined primarily by the rigidity of the pipe and the greater magnitude of the admissible radius of curvature.

In the USSR, studies are currently being conducted with a view to ensuring reliable operation of pipelines in permafrost regions. These studies, along with problems pertaining to the technology and organization of construction, encompass a wide range of problems relating to the interaction of the pipeline with the permafrost. These include:

(a) studies of the thermal and mechanical interaction of pipelines with permafrost;
(b) studies relating to the prediction of variations in the permafrost conditions as a result of construction and operation;
(c) the devising of rapid survey techniques;
(d) studies of optimal operating conditions;
(e) estimating the environmental effects of laying pipelines.

In conclusion it might be well to point out that although engineering geocryology now has at its disposal quite good resources for solving the main problems of today and, in part, those of tomorrow, the further large-scale development of the North is posing new and more complex problems.

These and other engineering problems are predetermining the following areas of study in engineering geocryology: The development of more efficient methods of achieving directional control over the properties of permafrost and the use of these properties for engineering purposes, including various ways of cooling the earth materials constituting the foundation by prethawing, chemical hardening, etc.; the improvement of building methods at sites with distinctive geocryological conditions such as discontinuous permafrost and permafrost islands, permafrost with a temperature close to 0°, saline permafrost, and ice-rich ground, including areas containing buried ice; the study of the behavior of frozen ground subjected to dynamic effects and the development of antiseismic construction methods on permafrost; improving the calculated designed bearing capacity of foundations, primarily pile foundations, through the more complete utilization of the mechanical properties of permafrost, advances in foundation building techniques, and the use of improved calculation techniques; the use of deep supports and other efficient types of foundations for heavily-loaded structures; improvement of the methods used for engineering design of foundations and the establishment of a combined rheological and thermophysical method of calculation, which will make it possible to take into account the variation in temperature regime of the soil with time, etc.; the manufacture of highly efficient machinery desinged for northern use for the working of permafrost; and maximum mechanization of zero cycle operations and the development of the highly efficient methods of working the permafrost.

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PROBLEMS AND AREAS OF RESEARCH IN NORTHERN ENVIRONMENTAL CONSERVATION

P. I. MEL'NIKOV Permafrost Institute, Siberian Division of the USSR Academy of Sciences and O. N. TOLSTIKHIN Moscow Economic Research Laboratory of VNIIVO

The efficient use of natural resources and protecting the human environment against irreversible changes capable of making it unfit for habitation by future generations are becoming a matter of constant concern to an ever-increasing number of individuals, governments, and international organizations. Among the multiplicity of problems besetting civilized man in his efforts to make maximum use of natural resources while retaining the ability to regenerate them or transform them in whatever direction is required, the problems of the North stand alone. They are especially complex in the low-temperature frozen zone, which is characterized by deep cooling of the Earth's interior and the perennially frozen state of the earth materials in the upper part of the lithosphere.

Studies by Soviet scientists are making it possible to formulate these problems and map out basic areas of research, thereby providing a theoretical basis for overcoming the contradictions between man and the natural settings of the North. These natural settings with their unique landscapes and well-known potentialities for regenerating living resources, including timber, wild and domestic animals (reindeer husbandry and horse breeding), as well as their as yet unknown recreational potentialities, are deserving of undivided attention.

The problems of environmental conservation in the North have their origin in many cases, of which the following are the most important:

(a) the historic pattern of development of the landscapes making up the North;

(b) the contemporary permafrost-geological conditions, which in many respects are dictating the special character of the circulation of moisture in the hydrosphere-lithosphere system;

(c) the contemporary physical-geographic conditions, the predominant factor of which is the harsh cold climate;

(d) the abundance and variety of the natural resources, the development of which involves the mobilizing of equipment and energy and quantities of manpower.

Paleogeographic and paleocryological studies indicate that the shaping of the contemporary landscapes of the permafrost region took place in a natural setting that differed from that of today. In any case there is every reason to believe that in the comparatively recent past the climatic conditions were much more severe. This favored the formation of vast polygonal plains containing thick wedge ice and led to regional ice-richness of the mantle formations. The most recent climatic warming gave rise to a stressed thermodynamic state of the ice-rich permafrost. This state of the upper part of the Earth's crust is due to the fact that the heat losses to the atmosphere no longer compensate for the heat arriving from the Earth's interior, a consequence of which is progressive regional heating of the permafrost and melting of its included ice, beginning at the lower boundary limit of the permafrost region.

An analysis of hydrogeological, geothermal, geochemical, and hydrochemical information derived from studies of sections of deep wells in many parts of western and eastern Siberia unambiguously indicates that the reduction in the thickness of the permafrost zone in the last 20,000 yr as a result of the rise of its base ranges from tens of meters in folded mountainous areas to hundreds of meters in areas where platform continental deposits predominate. At the upper boundary of the permafrost region, this stressed thermodynamic state is manifested in that, as a rule, many of the disturbances in the nature of the underlying surface, such as changes in the soil and plant cover and in the moisture regime of the ground, lead to the development of irreversible processes and primarily to solifluction and thermokarst processes. These processes lead to disintegration of the previously formed landscape, sometimes entailing complete or prolonged loss of its biological productivity.

Thermokarst and thermal erosion processes also take place under natural conditions. They are manifested in the formation of troughs, lacustrine basins, and cemetery mounds; in the disintegration of riverbanks and seashores consisting of icy earth materials; and in other more or less perceptible transformation of the relief. Under natural conditions, however, usually they do not attain catastrophic proportions and gradually diminish when dynamic equilibrium is achieved between erosional and accumulative processes, between the processes of thawing and freezing. But the influence of man is such that these phenomena can have a catastrophic character and encompass vast areas; cutting of the mossy cover by tracked vehicles and the felling or burning of trees can serve as an impetus to irreversible changes of this type. Such seemingly negligible influences, which in another natural setting might have had no consequences whatever, lead to complete or partial disintegration of the landscape in some types of frozen landscapes.

Under natural heat-exchange conditions, all of these changes and disturbances can only take place as a result of a variation in the heat and water balance, a disturbance of the established equilibrium between the inflow and outflow components of the particular natural landscape.

Changes in natural landscapes, which are even more intensive and difficult to predict in terms of their consequences, can occur in the event of an artificial increase in the thermal load on the environment occurring as a result of the harnessing and use of large masses of energy or implementation of some of the existing plans for a global amelioration of the arctic climate.

Hence, the first problem is formulated as follows: *the working out of scientific criteria for use in maintaining the regional thermodynamic equilibrium in the system comprising the permafrost zone and the atmosphere, and the principles of controlling this system for conservation and improvement of natural landscapes and for increasing the regeneration of their living resources.*

The means of solving this problem are seen to lie in the development of efficient methods of artificially controlling the heat and mass exchange between the Earth's surface and the atmosphere within areas undergoing development, as a rule in the direction of stabilizing the temperature regime of the perennially frozen earth materials. Methods of building civil and industrial structures whereby the ground constituting the foundation is maintained can be regarded as a first step in implementing the results of scientific research in this area in the frozen state, widely used in construction practice. The successful experience of building large modern cities and industrial complexes will provide adequate information for the devising of efficient methods of controlling thermal processes when developing northern regions. Above all, however, it must be used for the planning and construction of transportation arteries: trunk highways and railroads, pipelines, and power transmission lines. Long-term developments in this area must take into consideration, on the one hand, a continuous growth of the power being made available to industrial enterprises in the North and, on the other, the opportunities provided by the chemical industry for using new types of thermal insulating material.

It would seem that when determining the criteria to be used in maintaining the regional thermodynamic equilibrium in the permafrost atmosphere system and the principles of controlling this system, it is advisable to consider two possibilities. One of them is the development of efficient procedures for overcoming the unfavorable consequences stemming from disturbances of the underlying surface from a growing heat load. The second is the development of procedures aimed at reducing the number and degrees of the disturbances of the underlying surface by decreasing the intensity of the thermal load on the environment, this to be achieved through more extensive utilization of hydroclimatic resources. In particular, this is possible in the mining of placer deposits, which is one of the main consumers of energy. Hydrothermal management techniques as a means of thawing clastic frozen deposits are highly propitious, both from the standpoint of protecting the environment from disturbance and thermal pollution caused by the construction of power plants and the laying of transportation routes and power transmission lines to connect with them and of implementing the procedures required when other methods of thawing and mining placers are being used.

The regional studies conducted in the permafrost zone have revealed that everywhere the latter is discontinuous. This even includes the so-called continuous permafrost zone, where taliks are also universally present, the origin and existence of which is closely linked with bodies of water and watercourses. Although the areas covered by these taliks do not usually amount to 5 percent of the area of the continuous permafrost zone, it is precisely these taliks which in many respects determine the character and the biological productivity of the frozen landscapes. The preservation of these taliks, given the conditions of the thick, low-temperature permafrost zone and the harsh climatic conditions of the North, is wholly determined by the capacity of the natural water to accumulate during the summer and transport during the winter the substantial heat fluxes that are both necessary and sufficient for maintaining a state of equilibrium between the thawed and the frozen earth materials. Correspondingly, changes in the magnitude of the thermal load on the talik zones or in their natural water-thermal regime also lead to irreversible changes in another thermodynamic system: that of the frozen and thawed earth materials. The thawing and especially the freezing processes occurring here are capable of changing the character of the natural landscapes in the permafrost zone and of disturbing the biocenoses associated with them.

The most striking example of the transformation of a landscape associated with the taliks of river valleys into a permafrost landscape are the changes originating when the mining placer deposits involves the use of powerful excavating equipment and water-monitors for sluicing the deposits. As a result, within the confines of the placer and frequently over a much larger area the united surfaces of the floodplain and suprafloodplain terraces are broken up and reworked into technogenic relief consisting of low hillocks. In the process, aleuritic material (melkozem) is washed out of the gravelly-shingly deposits and carried downstream by the current. By the material getting into the surface water absorption zones, it is conducive to their silting up and to a worsening of the percolation characteristics of the talik aquifer. In the valley entirely new hydrogeological conditions originate due to the presence of the well-drained low hills consisting of gravelly shingly and bouldery material, washed down by percolation, and also of depressions in which the percolation characteristics of the reworked alluvium are reduced to a greater or lesser extent by the melkozen accumulation processes and by the plugging of their pore space. Correspondingly, the fragmentation of the formerly united subchannel flow, which had provided for the existence of the talik itself, is rapidly manifested in freezing of the now technogenic deposits, comprising for the most part well-drained hills and hence the depressions between them. This freezing can extend downstream if the worsening of the percolation properties of the alluvium reaches a threshold below which the speed and, through it, the thermal capacity also of the subchannel flow becomes so low that it is no longer able to maintain its natural geothermal regime. If we consider that in mining the channel and floodplain deposits of the river valleys within the talik zones, the most valuable of the relic poplar forests in the North are destroyed, the very existence of which is due to the water-bearing taliks, it becomes obvious that the conditions stemming from mining procedures are not conducive to their natural regeneration. Moreover, after stabilization of the new temperature regime and freezing of the massifs of technogenic deposits in the river valleys, the artificial recultivation of the technogenic relief becomes an almost unrealizable procedure. In place of flourishing valleys covered with deciduous forests providing a significant part of the food required by hoofed animals, rabbits, and birds, lunar landscapes arise that break up the hitherto intact taiga into lifeless rocky strips. It is still unknown how they will affect the surface and groundwater reserves and other natural factors determining the northern landscapes.

Of course the special case discussed does not encompass all aspects of the second problem touched upon, but merely indicates very clearly the consequences that can arise as a result of an artificial disturbance of the hydrothermal state of the rocks in the talik zones, and thus makes it possible to formulate the essence of this problem as follows: *the working out of scientific criteria for use in maintaining a regional thermodynamic equilibrium of a system comprising frozen rocks and thawed rocks, and the principles of controlling this system for the conservation and improvement of natural landscapes and for increasing the regeneration of their living resources.*

The means of solving this problem are seen to be primarily in the development of efficient methods of optimizing the hydrothermal regime of the river valley taliks. This can be done by an increase in the thermal load as a result of harnessing the thermal effluent from power plants, controlling the river runoff, and implementing other measures aimed at intensifying the use of energy resources. Control of the frozen rocks-thawed rocks system in such cases is not seen as an end in itself, but as a means of overcoming the unfavorable effects of the technogenic load on the environment and providing for a simultaneous improvement of the natural landscapes and an increase in their biological productivity.

The successful solution of problems relating to optimizing the hydrothermal regime of taliks is closely linked with another research area--groundwater resources and their control. In particular, experience gained in the prolonged operation of talik water intakes in certain valleys indicates a shifting of the boundary between the thawed and frozen rocks in the water intake zone in the direction of the frozen rocks and an increase in the thickness of the talik slit. The operating reserves of the talik water also increase and their quality improves. All of these changes have taken place merely as a result of an increase in the speed of the subchannel flow in the vicinity of the water intake and without any special measures aimed at enlarging the talik and improving its parameters. A project for the future, however, is the creation of artificial pools of fresh groundwater in the frozen ground or in the cryopeg zone--containing bodies of liquid saline water below 0° and frequently forming the lower part of the permafrost section. The sparse data currently available and also preliminary calculations show that the possibility of artificially controlling the groundwater resources is entirely practicable and, as was noted in the generalizing paper on the fifth section, in a number of cases it is in this area that the only workable way is envisaged for supplying water to the populated places and enterprises in certain parts of the permafrost region.

All that has been said with regard to solving the second problem pertains to artificially changing or "improving" the natural equilibrium of the frozen rocks-thawed rocks system, the practicability of which is predetermined by the increase in the power supply per unit of production in the northern regions. It is also necessary, however, to consider the studies aimed at the preservation of this equilibrium under the conditions associated with intensive development of the natural resources and its restoration in the case of a forced temporary disturbance. Here again we run up against studies aimed at determining both the potentialities and efficient methods of achieving hydrothermal improvement and control of the hydrothermal balance of the surface together with maximum use of the hydroclimatic resources, but from a somewhat different point of view.

The successful solution of these two problems is organically linked with a third: *the development of efficient methods of recultivating the land.*

This problem, which has come to a head in all of the mining industrial regions without exception, looms especially large in the permafrost region. Actually, whereas under normal conditions the problems of recultivation can be solved during any stage of exploitation of the deposits and even years after they have been worked out, as was demonstrated in the example of mining a placer, the recultivation of the land in the North must normally be done immediately after mining it, or in any case during same summer. Otherwise the mass of earth brought to the surface will freeze. Inasmuch as the thickness of the winter freezing layer is usually greater than the thickness of the summer thawing layer, this freezing progresses until the short-term permafrost completely merges with the basic mass of permafrost. After this, recultivation of the land becomes almost impracticable, or in any case requires such expenditures of energy and so much manpower and technical facilities that the consequences of this procedure may prove to have just as unfavorable an effect on the environment as the mining of the deposit.

What has been stated does not exhaust the exclusiveness of the recultivation problem in northern areas. Under normal conditions, the solution to the problem ultimately amounts to the creation of efficient forms of artificial landscapes and optimal geochemical conditions for restoration of the soil and plant cover. Moreover, in the presence of a cold, harsh climate, the artificial restoration of the hydrothermal balance that existed prior to the mining of the deposits and provided for the development of the talik ecosystems of the river valley is of immense importance for successful recultivation.

Of undoubted interest is the solution of other practical and procedural problems of recultivation in the permafrost region where certain methods can be used that would be unacceptable under ordinary conditions. These include filling the depleted workings with water, the subsequent freezing of which eliminates the stressed state of the roof and enables all types of recultivation work to be done at the ground surface, right down to hydrothermal improvements; the creation of a frozen substrate in the leveled massifs of depleted rock, thereby insuring optimal conditions of flooding of the restored soil cover; and other methods.

Success in solving this problem is largely determined by the economic efficiency of the recultivation measures. It is no accident that these problems began to be solved in the southern regions where restoration of the land for agricultural use is an observable and comparatively easily evaluated economic problem. The situation is different in the mountainous taiga and tundra regions of the North, where the economic efficiency of developing biocenoses within the territories affected by the working of open-pit mines and placer deposits, or by underground mines has not been investigated and can hardly be evaluated by existing methods. Here it can only be discussed in the general context of making efficient use of the natural resources of the North.

The presence only a short distance below the surface of a regional, frozen, water-impervious bed, combined with the negative mean annual air temperatures and very low winter air temperatures, give rise to an absolutely unique hydrochemical situation in the North and to the extensive occurrence of processes resulting in cryogenous metamorphization of the groundwater, including the water of the active layer, the suprapermafrost, and the intrapermafrost taliks. The overall trend of these processes is towards an increase in groundwater mineralization and changes in the relation between the basic ions, which ultimately leads to an increase in the sodium chloride and magnesium sulfate content of the water and to salinization of the soil with hydrocarbonates and sulfates of calcium and magnesium. Under natural conditions these processes are in equilibrium with the lixivation processes and do not usually cause irreversible changes in the groundwater composition or salinity of the soil profile. Such processes originate only in the closed drainage depressions of areas with a sharply continental climate where mineral lakes and primarily soda lakes form. But under the effect of human economic activities, these changes are taking place everywhere. In particular, the increase in the mineralization of the water of the active layer caused by domestic or industrial effluents is leading to a situation in which this water does not completely freeze at zero temperature. Its cooling below zero degrees causes differentiation of the solution with squeezing out of the salts at the freezing front and, accordingly, concentration of the residual solution. The degree of concentration is determined by the salt composition and the cooling temperatures. The multiple repetition of the freezing and thawing cycle leads to a situation in which a layer of groundwater with a uniform chemical composition and mineralization is broken down into sections of weakly mineralized water and lenses of high mineralization that have a negative temperature in the winter. These are the so-called "seasonal cryopegs." These phenomena are common in settlements not provided with sewers, on livestock farms, and in areas where untreated waste is discharged. For example, in the city of Yakutsk and its environs, during its 350 yr of existence numerous lenses of briny and brackish water consisting primarily of magnesium sulfate and sodium chloride have formed, the mineralization of which ranges from grams per liter to 80-100 g/liter, the most concentrated brines being confined to the older part of the city, whereas weak brines and brackish water are encountered around the perimeter and in the recently developed areas.

During the freezing period, the polluted and concentrated water of the active layer may be squeezed out into the suprapermafrost and intrapermafrost taliks and lower the quality of the talik water, which is used for the water supply. It cannot be ruled out that this is precisely the nature of the high concentrations of biogenic components of the talik water mineralization in the alasses. These have long been used by the people of Yakutia for pasturing livestock and for the positioning of winter corrals and pens. In addition, the salinization of the water and soil rapidly leads to loss of the plant cover, primarily the trees and brush, and under these conditions regeneration of the vegetation is laborious and inefficient. This circumstance hampers the implementation of landscaping measures in the towns and settlements and leads to an expansion of steppe and semiarid landscapes characterized by low biological productivity.

In addition, the presence of the highly mineralized groundwater has an adverse effect on the foundations of buildings in that it leads to destruction of the concrete in the seasonally freezing and thawing layer, which in turn causes deformation of the buildings.

The occurrence of the permafrost close to the surface makes for an extremely thin aeration zone and thus greatly reduces its capabilities for natural purification of the water percolating from the surface. Thus, the upper aquifers of the talik water remain unprotected against surface pollution and the squeezing of the polluted water of the active layer into them. This imposes special requirements on the application of fertilizers and herbicides, the percolation of which below the soil profile becomes almost impossible, and in the absence of a sufficiently rapid surface runoff it is inevitable that there will be a gradual increase in their concentration to the point where this leads to a reduction in soil fertility and perhaps also to the poisoning of birds, small rodents, and other animals linked to them by the ecologic chain.

Especially unfavorable with respect to pollution is the situation of the water in the subchannel taliks in mountainous regions which, together with the surface and subpermafrost taliks, form a single water pressure system reacting sensitively to the surface water pollution. Hence, a fourth problem is formulated that is closely interrelated to the three preceding ones: *the working out of scientific criteria for use in maintaining a stable natural hydrochemical equilibrium in the system comprising the soil and the soil, ground, and surface water, and the principles of controlling this system in the direction of overcoming the unfavorable consequences of the development of northern areas and generating optimal conditions for the conservation and improvement of landscapes and for the regeneration of its living resources.*

The ways of solving this problem are seen to be primarily in expanding and deepening the studies of cryogenous metamorphization, the geochemical conditions influencing the genesis of the soil water and groundwater, and the scattering haloes of the pollutants. The application of various types of fertilizers and the use of domestic wastewater for irrigation under the conditions obtaining in the North call for independent studies, the content of which is to a large extent determined by the specific nature of the permafrost-hydrogeological situation--the absence of an aeration zone, the climate, the cryogenous processes, and the delayed course of the oxidizing and exchange reactions--that is, the sum total of the natural factors distinguishing northern areas from the central and southern belts.

The working out of all of these cardinal problems with the object of providing a theoretical basis for the natural conservation measures must be done in such a way that it is inseparably linked with the economic activity associated with developing the natural resources of the North, proceeding from the fact that ever-increasing quantities of people, equipment, and energy are being mobilized for this purpose. Furthermore, it must be predicated on the solution of practical problems aimed at reducing the adverse effect of the technogenic load on the environment. Essentially, in terms of conservation research, the traditional engineering problems--the effect of local conditions on engineering structures--must be solved in reverse order--the effect of the engineering structures on the environment.

It has been proved theoretically and in practice that in most cases surface laying of various types of pipelines is possible when a specific kind of protective casing is placed around the pipe. It is the basis of these developments that pipelines have been built which in part run directly along the Earth's surface. The initial operating period of these pipelines has demonstrated that neither the high amplitude of the temperature fluctuations, reaching 100° and more, nor other factors have any appreciable effect during this period on the operational reliability of these structures. However, continuous surface movements of the pipelines, which are due to the effect of these temperature amplitudes in causing changes in their lengths, have led to the destruction of the soil cover over long distances and, correspondingly, to a change in the surface hydrothermal regime. The latter, in turn, has led to the development of thermokarst processes. These processes have caused destruction of the natural landscapes in the zone occupied by the pipeline and approach routes. They have even posed a threat to the structure itself, which here and there has been left hanging above areas of thawed ground that have settled. Another widely known fact is that in certain northern areas pipelines have led to the death of reindeer as a result of fragmentation of the summer and winter grazing areas of the reindeer, for which the lines proved to be an insurmountable obstacle.

One of the common methods of neutralizing wastewater that has been partially purified or not purified at all is its dilution with river or lake water. Here, the completion of the water purification process is achieved both as a result of dilution as such and due to the self-purification processes taking place on the basis of the oxidation reactions and the biological activity of the microorganisms in the natural water. This is the situation under ordinary conditions. In the North the rivers carry much more organic matter in the water and in the winter, as a result of the length of the freeze-up process and the significant reduction in the water content, oxygen starvation takes place, which in some years causes wholesale dying of the fish even under natural conditions. Of course, this water cannot neutralize the wastewater, which inevitably leads to its poisoning. But this is not all. Experiments performed with a view to determine the breakdown rate of many pollutants as a function of water temperature have demonstrated that when this is decreased this breakdown time increases significantly. Consequently, both the composition and the temperature of the river water are unfavorable for the normal course of the wastewater dilution and self-purification processes in the bodies of water into which this runoff takes place.

A third unfavorable factor in this respect is the low water content of the rivers during winter. It is sufficient to cite the mean value of the winter river runoff in the permafrost zone relative to the annual runoff: It does not exceed 5 percent. Many of the rivers, even comparatively major ones, the Yana included, and in some years and at some stream gauging sections, the Indigirka as well, freeze up completely in the winter. Under these conditions, when all of the fish stocks of the river are concentrated in certain unfrozen "holes," a discharge of wastewater means the poisoning of the entire river and all of the living organisms using its water.

River water, however, under the conditions obtaining in the permafrost zone, is intimately linked with the intrapermafrost and subpermafrost water. This is because water-absorbing taliks are concentrated in the river channels and under the river terraces. It is in winter, during the period of maximum depletion of the natural groundwater reserves, that the absorption of the river water is at a peak. It is known that this is precisely the time when many water-removing taliks become water-absorbing. Therefore, the discharge of wastewater in winter, in addition to poisoning the river water, in many cases can also lead to poisoning of the groundwater, which during the summer low-water period will put this poison water back into the river system. In time, control over the discharge of polluted water will be achieved to the extent of its being more evenly distributed throughout the year, causing systematic poisoning of the river water by effluent, which perhaps will be dumped during a brief time period.

The waters of northern rivers and lakes, like those of Lake Baikal, frequently have a very low level of mineralization, to which their contemporary biocenoses have adapted. Therefore, nominal purification of the water, even to the extent of upgrading its mineralization to the potable standard, may prove insufficient for maintaining the productivity of the existing ecosystems in which, due to the discharge of even comparatively pure water, unforeseen and irreversible changes may occur. All of this indicates that for the northern rivers their own wastewater purification norms must be worked out, and in some cases this must either be completely neutralized and recycled or the pumping of untreated industrial waste into deep water-bearing layers will be necessary.

There is also the attractive prospect of providing for wastewater storage directly in the permafrost by generating artificial taliks in it and replacing the water from melted fresh ice by wastewater or the use of existing closed taliks for this purpose. Although use of untreated or highly toxic water in this way is entirely realistic, a very careful evaluation must be made, both of the permafrost-hydrogeological conditions of the talik and the reliability of its insulation from the subpermafrost, suprapermafrost, and intrapermafrost aquifers. Allowance must be made for the fact that with the passage of time and a lowering of the temperature of the wastewater, it will become partially frozen in the talik reservoir as a result of which the remainder of the solution will be concentrated to such an extent that it ceases to freeze even at low negative air temperatures. A lens of highly concentrated brine will form that will be capable of dissolving the ice cement of the rocks and of migrating under the influence of the gravitational factor in a variable temperature field in the direction of the most icy and potentially permeable rock. Such migrating lenses of concentrated domestic discharge are known to exist, in particular, in the vicinity of old Yakutsk. If the wastewater has a higher temperature, rapid thawing of the permafrost may begin, and it will be important to be certain that the wastewater storage talik will not merge with taliks or aquifers, the water of which is used for drinking or economic purposes.

One of the normal methods of neutralizing a flue gas discharge is its dispersion in the air, the intensity and degree of which is determined not only by climatic factors (wind speed and direction), but also by the distance of the site of the discharge from the surface, that is, by the height of the smokestack. In the case of a sustained wind direction and at normal air temperatures in many cases the dispersion of the smoke leads to the desired results. Moreover, the fallout of solid particles to the Earth's surface does not cause the additional consequences that are associated with permafrost and the climatic peculiarities of the North.

It is known that in winter, over the vast expanses of the northern part of eastern Siberia, a stable anticyclonic state of the atmosphere is established, which is characterized by calm cold weather and temperature inversions. Evidence of this is to be seen in the fact that cold air masses from the slopes of the promontories and the river valleys stream toward the intermontane depressions and valleys, which is just where the cities and settlements are usually located. On the one hand these processes hinder the dispersion of the flue gas discharge, and on the other they promote its settling directly in the inhabited area. The solid particles of smoke become crystallization nuclei for water vapor, resulting in the development of ice fogs or smogs. These differ from ordinary ones in that they have a lower temperature and are much more stable and toxic. These ice fogs, which last for many weeks, lower the visibility on the roads, which under the conditions of winter glaze ice leads to high accident rates in transportation, lowers the regularity and safety of aircraft flights, and complicates the life of city dwellers. In covering the snow surface the solid particles of flue gas discharges reduce its reflecting capacity and lead to early removal of the snow cover, which at low nighttime air temperatures has an adverse effect on plant growth.

What has been said also applies in large measure to discharges of vapor from the cooling towers at thermoelectric power plants. Whereas under normal conditions these vapors are scattered over comparatively small distances, on cold still days the trails of frozen fog extend for many kilometers, causing virtually the same consequences as result from flue gas discharges.

It is this that gives rise to the purely practical problems of developing systems of fuel combustion, raw material processing, and cooling that will completely rule out discharges of solid and gaseous particles into the atmosphere. It is in the North that the need is more pressing for the innovation of existing advanced techniques.

The conservation of the natural environments in the North is inseparable from the development of their resources in the broadest sense of the word. Accordingly, the study of the natural resources must be oriented in the direction of optimizing their use for the benefit of man. What has been done along these lines, for the most part concerning building on permafrost, can be regarded merely as first steps in this direction. Here are some of the areas of research that will be of interest in the immediate short term. Some of these studies are already being conducted, and the implementation of their results will not long be delayed. Others must be discussed in the order in which the problem is stated.

1. The use of the natural frozen substratum of the upper part of the Earth's crust and the processes of naled formation for the building of underground storage reservoirs with natural or forced winter ventilation insuring the requisite drop in temperatures to -10° to -15° where the permafrost temperatures do not reach these values. The theoretical justification and the exploratory and heat-measurement studies have demonstrated the high profitability of building such reservoirs, which would require almost no additional power input, thus eliminating all of the ensuing thermal and other technical loads imposed on the environment.

2. The use of percolating water intakes for intraseasonal regulation of the operating water reserves, raising water levels, lowering the mineralization, and improving the composition of the intrapermafrost and subpermafrost aquifers, as well as providing for artificial underground talik water storage reservoirs based on the use of hydroclimatic resources. The development of these procedures will provide for an enhanced capability of using groundwater and river floodwater as sources of water supply with minimal removal of silt beneath the reservoir, streams, and channels, thereby avoiding the development of thermokarst and water erosion processes, which are inevitable when installing surface reservoirs, canals, and surface pipelines. The available sparse data on percolation water intakes and an analysis of the information on the operation of ordinary water intakes tapping more than two aquifers and on certain reservoirs indicates that the use of methods providing for the artificial replenishment of groundwater resources at the expense of surface resources is entirely practical, and in terms of environmental conservation they are highly promising.

3. The use of naled control in order to intensify the river runoff during the summer low-water period. For all of the promise that it apparently holds this method of auxiliary control of the river runoff carries without it the possibility of overdrainage and irreversible freezing of the pebble bed, which, in turn, causes destruction of the floodplain and terrace forests in river valleys. In this context studies must be made not only of the consequences of naled regulation, but also of the rate of water takeoff from the subpermafrost and intrapermafrost aquifers, together with the consequences of using other methods of controlling or intensifying river runoff (eliminating the evaporating surfaces of the interfluves, etc.).

4. The use of the desalting action of the icing-forming process to improve the quality and reduce the mineralization of weakly brackish and fresh groundwater that does not correspond to the All-Union State Standard and also purification of wastewater not subjected to other means of purification. Obtaining table salt by freezing has long been practiced by the local population of eastern Siberia. The natural processes of naled formation and the differentiation of the natural solutions into the fresh ice and saline components associated with them suggest the possibility of the industrial development of this process, which is especially promising for the purification of wastewater. The purification plants must include storage basins in which the summer discharge of the enterprise accumulates and natural purification of the water occurs as a result of gravitational differentiation, and also freezing basins from which the salts crystallizing on the surface of the ice are removed mechanically and those migrating to the base of the ice in the form of solutions are pumped off for secondary processing.

5. The use of freezing and melting processes for directional control of the groundwater regime for purposes of land improvement or for controlling unfavorable cryogenous processes (heave, icing, frost fracturing of the ground, etc).

Although these processes are finding practical applications, for example, permafrost belts to control naleds, deep freezing for the sinking of holes in flooded strata, increasing the bearing capacity of piles, the building of impervious barriers to prevent percolation from reservoirs, and other sources, the opportunities for directionally controlling this system are far from being exhausted.

In evaluating the natural resources of the south coast of the Crimea, it is customary first and foremost to speak not of its mineral wealth and, perhaps, not even of its vineyards, but of the health resorts, therapeutic, and tourism potentialities: the sea and the climate--these are the principal natural resources of Crimea, the use of which is growing rapidly with the rise in the standard of living. In discussing the natural resources of the North, it is usual to estimate them in millions of cubic meters of gas and tons of oil and gold, carats of diamonds, cubic meters of wood, and hundreds of thousands of priceless pelts from fur-bearing animals. At the same time, with the increase in the population, especially in the industrially developed regions of Siberia and the Far East, the rise in the overall standard of living of the Soviet people, and the expansion of tourist relations abroad, a definite trend is to be seen in the shifting of large scale tourism away from the southern and central regions towards the eastern and northern regions. This trend is a natural outcome of the desire of the city dwellers to remove themselves during vacations from the ordinary pattern of life, to obtain as many new experiences as possible, and also physical activity that will have a favorable effect on their health and insure fitness during the next period between holidays. The North and the East offer maximum opportunities in this respect. The tourist excursions on the motor vessels through the Arctic Ocean and on the Lena, the Ob', and the Yenisey undertaken in recent years, the enormous number of organized tours, and the even greater number of unorganized groups of tourists--all are a consequence of this trend, which we are seeing in own time. There is every reason to assume that in the future this trend will become even more evident.

From this an entirely distinct problem arises--evaluating the recreational potentialities of the northern and eastern parts of the USSR and the recreational significance of the various landscapes in the North and the East, and also envisaging ways of expanding tourism in the foreseeable future and the measures that will be needed to ensure optimal opportunities for the tourists and maximal economic and extraeconomic (health, esthetic, and so on) benefits of their stay on location. It appears that, in addition to the ordinary traditional journeys on board motor vessels, rafting and boating on the rivers, and hiking, there could be a marked expansion of journeys on horseback, which have been successfully initiated in the Altay Kray, and also springtime journeys on reindeer-drawn sleds (March-April). They would permit extension of the total tourist season to 5-6 months, which would increase the profitability of the tourist bases and the employment of guides, instructors, and drovers.

Organized tourism imposes an additional load on the environment and calls for the creation of wildlife sanctuaries and national parks. But they are not only needed for recreational purposes, but also as areas in which to preserve the natural conditions of heat and water exchange, the plant communities, and the animal world. The study of all these would enable us to see the changes taking place in the external environment under the influence of man and to orient these changes in the necessary direction. The inability of the northern landscapes to withstand external influences and the widespread development of intensive irreversible processes that are changing the face of the taiga and tundra make it mandatory to regard the setting up of wildlife sanctuaries and national parks as a most important measure in which all Soviet people and especially the developing people of the North are interested.

STUDIES AND CALCULATIONS OF THE TEMPERATURE REGIME OF THE GROUND AND STRUCTURES BY THE HYDRAULIC ANALOG METHOD

V. S. LUK'YANOV All-Union Scientific Research Institute of Transport Construction, Ministry of Transport Construction

The use of the analog method in hydraulic integrators in studies and calculations of the temperature regime of the ground and structures was proposed by the author in 1934. In 1936 the possibility of using this method to obtain a sufficiently exact reproduction of the temperature regime of the ground as observed at the Skovorodino Permafrost Station was demonstrated.

The method is based on the principle of obtaining approximate numerical solutions of systems of equations describing the heat balances of interacting elementary volumes or assemblies by bringing into play an easily observed hydraulic process described by the same systems of equations. It was found that the hydraulic analog method performed on hydraulic integrators makes possible highly flexible changes in the design diagrams and boundary conditions.

The method and the equipment (the hydraulic integrators) have been continuously refined. At the present time the method is being used in many fields of science and technology. The UGL hydraulic integrators have been plant-manufactured for more than 20 yr. In geocryology the method is especially important for studying the movement of heat and matter in porous media.

With refinement of the equipment, development of the theory and accumulation of experience, the efficacy of the method in geocryology is constantly expanding.

The basic difficulties in determining the temperature regime of the ground are as follows:

(a) Selecting the correct design diagrams for the highly complex thermal processes in the ground;
(b) Designating the design parameters and coefficients;
(c) The practical application of analog computer mathematics facilities.

Having regard to the complex initial and boundary conditions, the heterogeneity of the medium and the latent heat of a change in the state of aggregation at some constant temperature of freezing and melting of water substance, design diagrams, which are based solely on the theory of heat transfer by thermal conductivity, sometimes prove unsatisfactory. On occasion it is necessary to allow for the freezing and melting of water in the ground occurring in the negative temperature range in conformity with a given law. In many cases along with the thermal conductivity it is necessary to allow for the heat transfer by the percolating water. These very important factors complicate the calculations. It is only relatively recently that they have begun to be taken into account, using the method of hydraulic analogies.

For correctness of the forecasts many design parameters, for example, the coefficients of thermal conductivity of the soil, the snow cover, and the thermal insulation, and especially the heat transfer coefficients at the surface, must be defined on the basis of *in situ* measurements and the solution of inverse problems on integrators.

When there are complex design diagrams the problems can only be solved through the use of computers: analog (hydraulic integrators and electrical integrators) and electonic digital computers.

For studies and fast calculations based on the new design diagrams, hydraulic integrators are most convenient and for large numbers of calculations by developed schemes--digital computers. When using analog computers, digital computers, and similarity theory, it is advisable to find generalized solutions.

In addition to the direct solution of problems in research and production, the use of the hydraulic analog method contributes to a better understanding of the thermal processes, the development of engineering intuition, and a creative approach to solving practical problems in various fields of human activity under harsh climatic conditions. Therefore, the hydraulic integrators are widely used in higher institutions of learning and in research and production organizations.

The use of the hydraulic analog method in determining the temperature regime of various engineering designs and structures is in some respects simpler. However, the diversity of these problems is also very great. For example, concrete work requires predicting the temperature regime of the structures, not only from the point of view of estimating the conditions of the concrete maturing as a material, having regard to the exothermy of the hardening process, but also of estimating the thermally stressed states originating in them both in the course of their manufacture and when they are in operation.

The hydraulic analog method must be more extensively used in developing new structural designs of residential buildings and premises intended for various uses under harsh climatic conditions by estimating the thermal-engineering properties of the buildings as a whole.

It is obvious that even such processes as the percolation of water and the diffusion of substances described by the same differential equations as the thermal processes are also calculated by the hydraulic analog method, for example, the processes of soil settling resulting from squeezing the water out of the skeleton, which are discussed in the interesting paper by Mr. Morgenstern.

ON THE PERMAFROST AT THE SUMMIT OF MOUNT FUJI

K. HIGUCHI AND Y. FUJII Water Studies Institute, Nagoya University, Japan

Permafrost observations at the summit of Mount Fuji have been conducted since 1970.[2,3] The results of the observations make it possible to draw the following conclusions.

The presence of permafrost at the summit of Mount Fuji (7,365.3 m above sea level, 35°21' north latitude, 138°44' east longitude) is confirmed not only in a small area near the meteorological station of Fuji-san, where the existence of permafrost has been observed in the summer and also previously, but along the entire perimeter of the principal crater. The lower limit of permafrost passes along the north slope at an altitude of about 2,800 m and along the south slope at about 2,900 m. The mean annual air temperatures at these elevations are -1.4°C and -1.8°C, respectively, which agrees with the mean annual temperature (-1°C and -2°C) at the southern limit of the discontinuous permafrost zone in the high latitude region.[1]

The thickness of the active layer at the end of July is 50-130 cm; smaller values are noted in the area where in the winter the thickness of the snow cover was greater. The thickness of the active layer, l, is approximately proportional to the square root of the magnitude of the thawing index Ω_T (°C-days) after disappearance of the snow cover as indicated in the equation:

$$l = 6.44\sqrt{\Omega_{T^\circ}}$$

At the permafrost table in a lava rock area, transparent ice was observed. From its shape it can be called "wedge ice."[4,5]

The thickness of the permafrost at the summit was hardly measured at all, but from an approximate estimate it is 40-60 m. In a repeated calculation of the freezing and thawing of the rock, the thermal regime of the ground, the water content, and the amount of snow fallen were assumed to be the same as they are today at the summit of Mount Fuji. On the basis of the calculation, it can be concluded that the effect of the snow cover on the thickness of the permafrost is small inasmuch as the snow that fell during the warm season was deeper than in the cold season.

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RELATION BETWEEN THE TEMPERATURE OF ROCKS AND THE HEAT-EXCHANGE CONDITIONS AT THEIR SURFACE

G. Z. PERL'SHTEYN VNII-1 (All-Union Research Institute No. 1)

One of the basic areas of thermophysical studies of the permafrost region is ascertaining the quantitative relations between the temperature regime of the rocks and the heat-exchange components at their surface. Without a knowledge of these relations, it is impossible to determine or predict and regulate the temperature field of the permafrost region.

At the same time, the heat-exchange processes occurring at the surface of the rocks are exceedingly complex, and the number of stations performing a complete set of heat-balance observations is small both in Canada and Alaska and in the northern and eastern parts of the Soviet Union. It was apparently these circumstances that led Dr. L. V. Gold (Canada) to give a pessimistic evaluation of his report on the prospects of the quantitative approach to the solution of this problem.

I do not share this pessimism. A procedure has been developed in the Soviet Union that makes it possible to relate the basic factors of the "external" heat exchange to the temperature of the underlying surface. The essence of this procedure is as follows:

All of the values entering into the classical equation of the radiation-heat balance of the Earth's surface have been divided into two groups. The first group includes the regional heat exchange characterized by constancy in large areas. For their determination the few actinometric stations that already exist are sufficient. Also in the first group we include the air temperature at an altitude of 2 m, which varies regularly at all of the meteorological stations.

The second group is made up of the microclimatic characteristics that depend on the temperature of the underlying surface T_0 (the turbulent heat flux to the atmosphere, the long wave radiation of the surface, the heat flux to the rock, and evaporation). By the use of well known physical laws (exchange coefficients) these variables are expressed as explicit functions of T_0. Unquestionably, many of the relations being used (for example, Newton's law) have a highly approximate nature. However, as progress is made in adjacent sciences, the form of the relations will be more precisely defined.

Thus, we shall consider the heat-balance equation as an expression containing one unknown T_0 with respect to which the given equation is resolvable.

The use of this method for analyzing the laws governing the course of thawing of the permafrost at the experimental sites in the Undy River Valley (Trans-Baikal) and in the vicinity of Cape Schmidt has yielded highly promising results.

I should also like to mention the need to allow for convective heat transfer by the percolating atmospheric precipitation or groundwater. In areas such as the northeastern part of the USSR, where impermeable coarsely clastic soil is widespread, convective heat exchange frequently plays a determining role. In these cases we can assume an inaccuracy when stating that the presence of a negative mean annual temperature at the surface is sufficient for permafrost to form under it. Under the influence of percolation flows, complete thawing of the seasonally frozen layer often takes place during the short northern summer. In this case, the mean annual surface temperature can be below freezing.

RELICT FORMS OF POLYGONS ASSOCIATED WITH WEDGE ICE IN SOUTHERN SCANDINAVIA (SWEDEN AND DENMARK)

H. SVENSSON Faculty of Physical Geography, Lund University, Sweden

INTRODUCTION

In southern Scandinavia the existence of pseudomorphs replacing ice wedges (relict ice wedges) is known from the mineral workings that have been driven through fluvioglacial deposits of late glacial origins (Wurms, Vistula, Valday). The shape and the structural peculiarities of these ice wedges have been studied by Danish and Swedish scientists since the end of the 1930's.[1-7]

Inasmuch as the polygons with wedge ice are a typical phenomenon characteristic of large areas, for mapping the distribution of the ancient polygons, data of purely local significance which can be obtained from isolated holes sunk in gravel are insufficient. In order to establish the presence of relict polygons with wedge ice and, therefore, create a basis for a more precise delineation of the boundaries of the periglacial regions in the past, a survey of large areas is required.

AIR PHOTO SURVEYS AS A MEANS OF DETECTING AND MAPPING POLYGONAL SYSTEMS

The air photo survey has been used for vast exploration work of this kind. I should like to present and discuss some results of the investigations of polygonal systems with wedge ice because the description of the shape and the explanation of the phenomenon itself must be based on morphological and structural data obtained in areas of contemporary permafrost.

In particular, let us consider the Laholm Plain on the West Coast of southern Sweden--an area that was subject to late glaciation (Wurms and Vistula), and part of western Jutland in Denmark--an area located outside the glacial boundary of the late glaciation.

The Laholm Plain

Polygonal systems with relict wedge ice were first observed on the Laholm Plain by Svensson in 1962 and 1964.[8,10] Here the soil is sandy and consists mainly of fluvioglacial alluvial deposits. At the beginning of the concluding period of glaciation, the region was covered by the sea, and in the course of uplift of the dry land during the late glacial period the sea gradually reworked the upper part of the soil. The region is almost completely cultivated.

The cultivation of the ground helps in defining the boundaries of the polygons, which stand out as well-defined lines or are in the form of tracts characterized by inhomogeneous field vegetation. Initially, there was disagreement with respect to the origin of the lines. Some have suggested that they are traces of an old agricultural system, boundaries of ancient fields, traces of drainage systems, etc. Numerous test holes cutting across the lines of the polygons detected on the aerial photographs have made it possible to discern that these lines are underlain by pseudomorphs of wedge ice (Figure 1).

The polygonal structure was first detected by a vertical aerial photographic survey from an altitude of 4,000 m during the photogrammetric mapping of Sweden. Later I took additional photographs from a light aircraft. The system of polygons is only distinguished after periods of drought. This summer (1973) there was no precipitation in the area for an especially long time. I have several slides made from the air which I hope will give a clear picture of the phenomenon (Figure 2 and 3).

The polygons are outlines by zones of denser

FIGURE 1 Horizontal and vertical section of boundary of relict polygon on the Laholm Plain.

and taller vegetation than the remaining part of the field. Inasmuch as the contrast is a function of the viability or degree of maturity of the crops, it is most convenient to use the part of the infrared spectrum (the lower infrared region), which is especially sensitive to differences in the reflectivity of the leaves for photography from the air and the poorly-defined polygonal system.

The polygons develop every dry summer at the same point and serve as proof of the fact that the lines are confined to the zones with a higher moisture content.[12]

As already stated, the soil consists mainly of sand. The vertical sections made along the lines of the polygons, nevertheless, show that the wedges contain finer-grained fractions than the soil as a whole. Some of the wedges contain a large amount of sand piled up by the wind and also silt. The structural composition of the earth materials making up the wedges and also the country rock shows that, as a result of the high degree of capillary action, the earth material in the wedges has a high moisture capacity. This makes it possible to create favorable conditions for plant growth during dry spells.

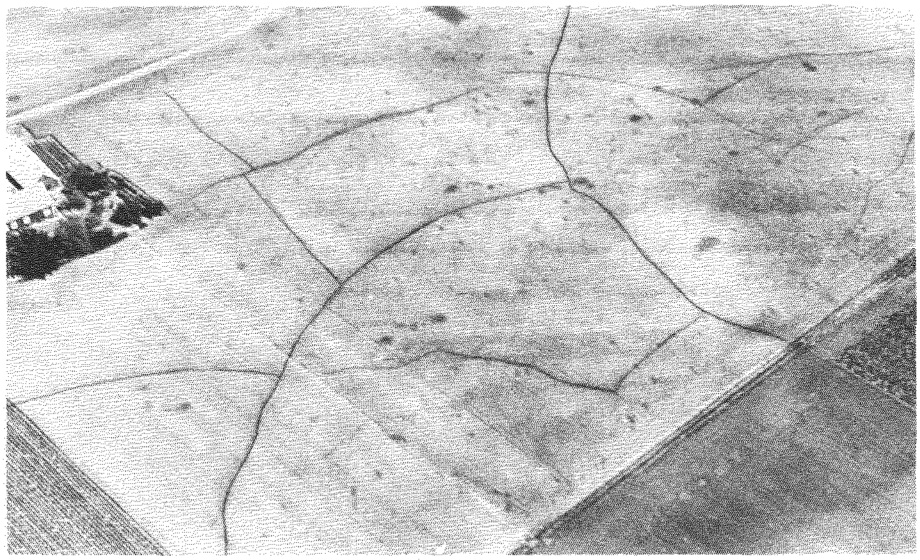

FIGURE 2 Oblique aerial photograph of relict polygons with wedge ice in a field of spring-sown grain on the Laholm Plain.

Denmark

The area occupied by Skoberg Bakke, of which a special study was made, was not subjected to late glaciation.[9] It is a leveled area of Riss moraine (Saale, Dnieper period). The composition of the soil is highly complex, but it is chiefly fine-grained and argillaceous. The Danish polygons are smaller than those of the Laholm Plain. There is also a difference in outline: the white lines contrast with the black lines on the black and white photographs,[14] and this difference can be explained by comparing the composition of the soil in the two regions.

Significance of the Soil Differences

In the Laholm region the soil making up the wedge is richer in fine-grained fractions than the surrounding soil. With the Jutland polygons the situation is reversed. On the Laholm Plain during droughts the vegetation at the edges of the polygons is denser and higher, which gives the outlines a dark color. In Jutland, on the contrary, the distinguishing of the polygons is due to the poor growth of vegetation along their edges (coarse-grained soil fractions) forming an indistinct system. The Jutland structure is distinguished most clearly when there is low vegetation and during the season it is easily blanked out when the vegetation becomes denser. Within a week a clearly defined network can completely disappear.[14]

During the protracted period of the past glaciation (Wurms, Wisconsin), the Saale moraines in Jutland were affected by permafrost processes, and this caused fragmentation of the polygons into smaller parts. On the Laholm Plain in the vicinity of the present coast of western Sweden, permafrost existed only at the end of the glaciation. Therefore, the polygonal structures are not as well-established here.

FIGURE 3 Polygonal structure on cultivated fields of the West Coast of Sweden.

USING THE RESULTS OF MAPPING RELICT POLYGONS WITH WEDGE ICE IN PALEOGEOGRAPHY

The study of polygonal systems by means of aerial photographs depends on the accuracy of interpreting the resulting information. However, the distribution of the polygons associated with wedge ice, the rate of their recurrence, and the geometric characteristics observed on the photographs give additional and more precise information about the periglacial regions of the latter than isolated holes can provide.

In the case of Denmark the polygons can be used when delineating the boundaries of the last ice shield in areas where geomorphological proof of the moraine lines is lacking.

In the case of Sweden the detection of the lowest (most westerly) polygons is very important from the point of view of chronology. This would make it possible to use dated coastal elevations for calculating the exact time-scale needed for the formation of the polygon. In solving this part of the problem paleogeographic and paleoclimatic factors can be utilized.

With respect to this summer (1973), with its long dry spell, we shall make a special effort to find the lowest, that is, the youngest and most indistinct polygonal boundaries of the West Coast of Sweden. As already stated, the analysis and explanation of the relict characteristics of permafrost must be based on data obtained in regions where permafrost at the present time exists. For estimating the age of polygons with relict wedge ice in the coastal part of Sweden, it is very important to know how close together or at what altitude above sea level the polygons with wedge ice are forming in regions of contemporaneous permafrost. Currently, as we have no data on the formation of wedge ice (ice wedges) in the northern part of Scandinavia, we should be grateful if scientists here at the conference who are familiar with the processes of formation of contemporary wedge ice in coastal regions would provide us with some information on this problem.

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PALEOCLIMATIC SIGNIFICANCE OF MEASURING THE TEMPERATURE OF PERMAFROST

V. CERMAK Geophysics Institute, Czechoslovakian Academy of Sciences, Prague

Valuable information can be obtained about the climatic conditions of the past from various fields of science, such as geology, astronomy, paleomagnetism, oceanography, paleobotany, history, and archaeology.[8] In addition, it is possible to reproduce the past temperature conditions on the surface by the temperature conditions under the surface at the present time. The temperature field of the Earth can be considered as the characteristic field of a solid conducting halfspace that depends on the boundary conditions. Any significant change in temperature on the surface extends downward so that the temperature existing at the present time at depth bears some "imprint" of the climate of the past. When actively studying the recorded temperature by comparison with the depth, these traces of the temperature of the past can be found and reproduced using the corresponding models of the possible climatic sequence.

The effect of the climate of the past on the geothermal gradient has been known for a long time. The general mathematical formulation is discussed in detail by Carslaw and Jaeger.[2] The theory of the climatic correction of the heat flux was advanced by Birch[1] and since that time it has been used by Jessop;[7] Cermak and Jessop[4] in terms of the step function; Ciaranfi *et al.*[5] beginning with a finite Fourier series; and Gold and Lachenbruch[6] beginning with an exponential series.

The thermal conductivity of the ground is low; therefore its sensitivity to the surface phenomenon is delayed. At depths of up to several hundred meters traces of the Pleistocene are still noticeable.

In Figure 1 we have a theoretical model of the sudden change in temperature on the earth's surface and the corresponding temperature distribution in the ground and the gradient.

There are many other factors that can affect the temperature of the series: the movement of water, radiogenic heat, mineralization, intrusion, uplift and settlement, erosion, uneven topography, or local variation in the thermal conductivity. Under favorable conditions, when the effect of all of these factors can be ruled out or substantiated, the recording of the deep temperature can be of some value in paleoclimatology.

Any reproduced climatic model based on analyzing residual temperatures gives in the best case only a possibility of obtaining the ground temperature, which differs significantly from the mean annual temperature. The effect of the precipitation, snow cover, drainage, vegetation, relief, and so on has yet not been determined. These microclimatological conditions should be considered from the point of view of the energy budget; however, such a study does not enter into the scope of this paper. Nevertheless, for a proper study of permafrost the existing ground temperature is a decisive factor for the development and evolution of permafrost.

Undoubtedly there is a relation between the data obtained and the geographic position of the investigated areas, and it is very difficult to generalize such results.

The temperature data used for this analysis were obtained in two boreholes in the northeastern part of Ontario: Kapuskasing (49°25' north latitude; 82°23' east longitude) and Hearst (49°41' north latitude; 83°32' east longitude) (Figure 2). These sections are located south of the boundary of the discontinuous permafrost in an area where the mean annual temperature is a little higher than 0°C. Both boreholes pass through granite and gneiss rock of the Canadian Shield.[3]

In order to calculate the climate of the past it is necessary to solve the inverse geophysical problem, that is, from the known corollaries to determine the initial cause. On the assumption that all of the conditions of (2) are satisfied, the general equation of thermal conductivity can be simplified:

$$\frac{\partial T}{\partial t}\ (z,t) = x\frac{\partial^2 T}{\partial z^2}\ (z,t), \qquad (1)$$

where $T(z,t)$ is the temperature; z is the depth; t is time; x is the coefficient of thermal diffusivity. In order to solve (1) the boundary conditions must be exactly determined.

$$T(z,0) = f(z), \qquad (2)$$

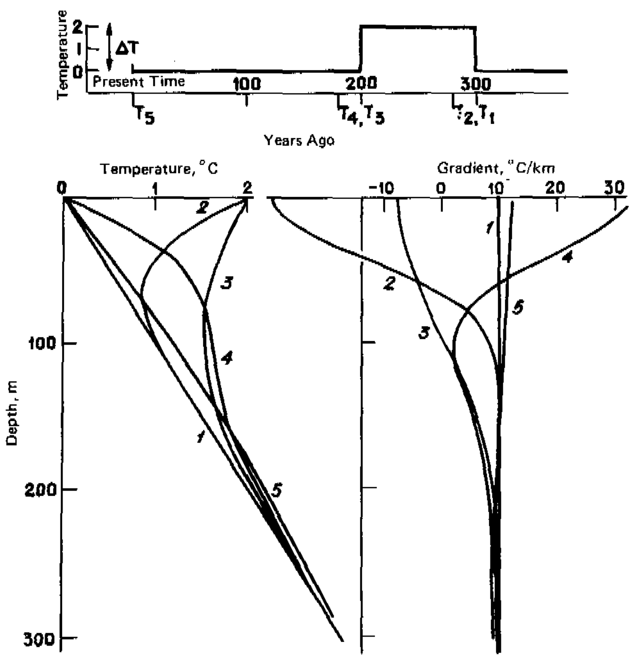

FIGURE 1 Simple model of the climatic variation during the period between 200 and 300 yr ago and the corresponding temperature at depth and the gradient at various times.

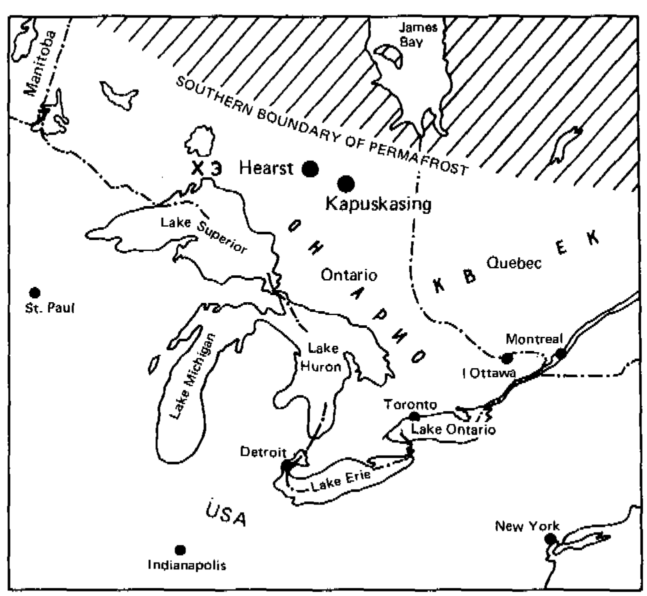

FIGURE 2 Location of drilling sites. Zone of discontinuous permafrost crosshatched.

$$T(0,t) = g(t). \quad (3)$$

The function $f(z)$ can be adopted so as to depict the initial temperature distribution

$$f(z) = T_0 + G_z, \quad (4)$$

where G is the steady-state stationary geothermal gradient; $T_0 = T(0,0)$ is the initial temperature at the Earth's surface. The function $g(t)$ indicates the temperature fluctuations at the surface caused by climatic variations; this function is unknown. In order to determine it, the general form of $g(t)$ expressed by the step function is assumed

$$g(t) = \sum_{i=1}^{n} T_i \left\{ v(t - a_i) - v(t - b_i) \right\}, \quad (5)$$

where T_i is the mean temperature at the Earth's surface in the i-th interval; a_i is the beginning; b_i is the end; $U(x)$ is the unit Heaviside function. If we use (4), (5), then (1) can be solved by a Laplace expansion in order to give the expression for the temperature deviation (the residual temperature $T = T(z,t) - T(z,0)$, caused by the climate of the past)

$$\Delta T(z,t) = \sum_{i=1}^{n} T_i \left\{ \operatorname{erf}\left(z/\left[4x(t - b_i)\right]^{\frac{1}{2}}\right) - \operatorname{erf}\left(z/\left[4x(t - a_i)\right]^{\frac{1}{2}}\right) \right\}. \quad (6)$$

If (6) is differentiated having regard to the depth, then the deviation in the geothermal gradient is given by the expression:

$$\Delta G(z,t) = \sum_{i=1}^{\pi} T_i \left\{ \frac{\exp[z^2/4x(t - b_i)]}{[\pi\chi(t - b_i)]^{\frac{1}{2}}} - \frac{\exp[-z^2/4x(t - a_i)]}{[\pi\chi(t - a_i)]^{\frac{1}{2}}} \right\}. \quad (7)$$

The measured profiles of the temperature distribution are shown in Figure 3; the calculated residual temperature gradient for the Kapuskasing well is given in Figure 4. The study of the gradient curve shows that the effect of the climate of the past becomes more noticeable; the individual peaks on the curve correspond to the climatic extrema.

The detailed descriptions of the processing of the temperature data, calculation of the gradient, and selection of the models are adduced by Cermak.[3]

The residual temperatures of the calculated gradient are obtained for m individual points (z_j); these values are equated to the right-hand side of Eq. (7). If n appropriate intervals (a_i and b_i) of different temperature of the Earth's surface T_i are selected taking into consideration the climatological data, we obtain a system of m equations in n unknowns for T_i

$$\Delta G_j(z,t) = \sum_{ij} R_{ij} t_i, \quad i = 1,2 \ldots, n \cdot j = 1,2 \ldots, m, \quad (8)$$

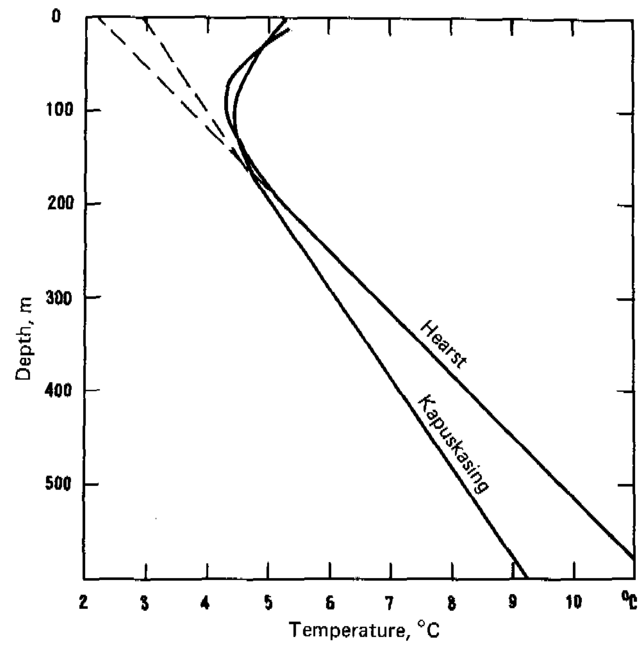

FIGURE 3 Temperature profiles in boreholes.

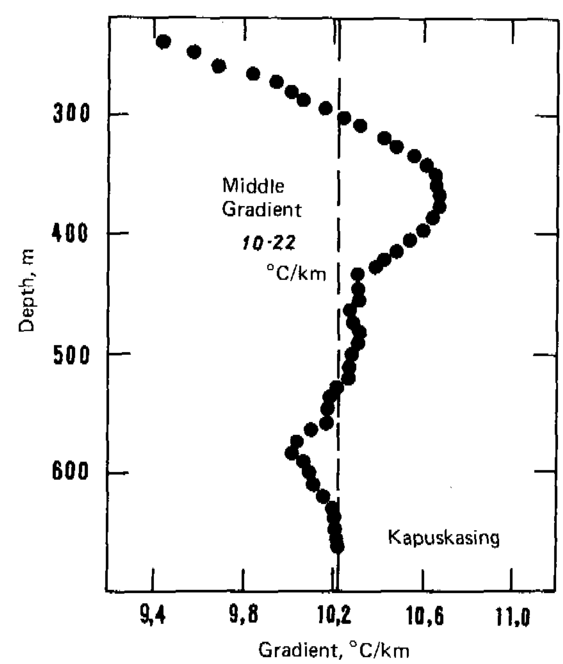

FIGURE 4 Temperature gradient in the Kapuskasing borehole. The gradient was calculated by the least squares method in the overlapping profiles of 15 items of information on the successive temperatures (at intervals of about 100 m).

where R_{ij} is the term of the proportion shown in parentheses in (7).

This system is complicated, and its solution changes readily. In order to overcome this, the approximate method is used.

The measured gradient is established for the climatic variations that existed to the year 500 B.C. The times of the climatic events during the last thousand years were approximately determined, and these data are used to form the set of possible values for a_i and b_i for each interval of interest. At each individual stage only two intervals are selected to form the incomplete climatic model; this is equivalent to accepting $n = 2$ in (8). All possible pairs of equations are solved and the statistical criteria and optimal values for a_i and b_i and T_i for both intervals are selected for this combination, which gives the minimum standard deviation. Then the geothermal gradient is established for the climatic effect just calculated and the next series of two remote intervals is processed in the same way. After the entire processing is ended, the values obtained for a_i and b_i for each interval can be substituted in the general system (8) and m equations in n unknowns solved for T_i. The results obtained for both wells by both of these operations are given in Figure 5.

The calculation indicates two different climatic extrema of opposite sign during the last thousand years and a noticeable rise in temperature on the Earth since approximately 1850 A.D. According to our data from paleoclimatology, these climatic events coincide with the so-called small climatic optimum (670-1,070 yr ago) and the small glacial age (175-475 yr ago), that is, in the interval of A.D. 900-1,300 or A.D. 1,500-1,800.

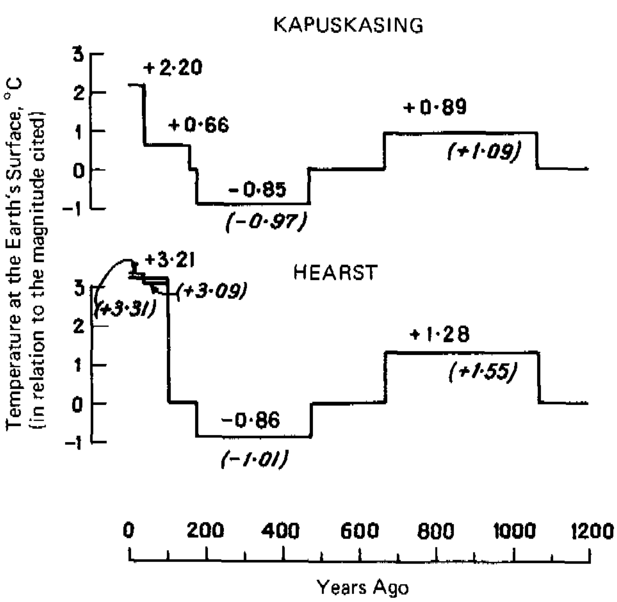

FIGURE 5 Calculated models of the ground temperature during the last thousand years. The numbers in parentheses correspond to the approximate method described in the text: the numbers without parentheses are the solution of system (8).

The working hypothesis of a uniform temperature of a rectangular pulse (the step function) in the past can serve only as the most simplified approach to the solution. In reality, every two definite ages are separated by a short period when the mean temperature at the beginning of the warming increases, and at the end it decreases. Thus, the largest temperature anomaly is probably to be expected in the middle of the two extrema.

The excess temperatures of +1.5°C for the small climatic optimum and/or -1.0° for the small glacial age were possibly greater in the culmination stages of both of these periods with respect to one or two factors. Favorable conditions for this obviously must exist in the regions of continuous permafrost where the groundwater is held in large quantities and where the theory of thermoconductivity applies without restriction. However, in this case the investigation must be limited with respect to surface variations, the effect of which was not reflected at the permafrost base; the total latent heat liberated or consumed during freezing and thawing could disturb the sensitive residual temperatures of the past.

In conclusion, the author expresses his appreciation to Mr. M. W. Chandler, who read the original version of this paper[3] and made useful critical comments.

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PERMAFROST ZONATION OF THE EUROPEAN NORTHEAST OF THE USSR

N. G. OBERMAN Ukhta TGU

Some peculiarities are noted in the distribution of the permafrost in this region, both horizontally and vertically.

Along the entire coast, on the islands of Vaygach and Kolguyev, there is a 10-50-km-wide strip of dry land within which the permafrost with a normal thickness of 10-100 m is underlain by a zone of cryopegs, with a thickness ranging from several tens to many hundreds of meters.

In the interior of the continent, towards the south, the permafrost increases in thickness; the depth of occurrence of its base gradually descends to 300-500 m. This trend is most clearly exhibited in the regions with a thick (more than 250 m) mantle of loose (Jurassic-Quaternary) deposits, which, as is known, are characterized by the greatest thermal inertia.

In areas where the permafrost is fairly thick, it was found to be multilayered and the presence of relict massifs was recorded. The latter was traced in several dozen wells from geological documentation of cores, counterboring of ice samples in the mine shafts, the presence of negative geothermal gradients in the positive temperature zone at depths of 100-150 m and also from geophysical research data (the vertical electric sounding method). The roof of the relict permafrost occurs in the 50-350 m depth range; the thickness of the thawed layers separating the frozen ones is 100-200 m.

In areas where the permafrost thickness exceeds 300-350 m, from time to time shortages of stratal pressures of the fresh permafrost water are noted. Their levels in each case occur at absolute elevations of -10 to -70 m (the absolute elevations of the well heads are 50-350 m). The available data indicate that the pressure deficit arose for the same reasons as in Siberia, which was reported on by O. N. Tolstikhin.

South of the present limit of contemporary permafrost relict permafrost may be present at the surface (judging from some indirect data). This is most likely in areas with a thick loose mantle.

The peculiarities of structure and thickness of the permafrost in the Soviet European Northeast are a consequence of the nonuniform duration and different climatic characteristics of the post-transgression period of permafrost formation in the various parts of the region. This hypothesis agrees with the paleogeographic data but needs to be confirmed by thermophysical calculations.

Based on these concepts, jointly with V. S. Zarkhidze we constructed a map of the permafrost zonation in the Soviet European Northeast in accordance with the principle of the simultaneity of the beginning of the post-trangressional period of its development. Five permafrost zones were distinguished, each of which is characterized by specific features with respect to the structure and thickness of the permafrost. An advantage of this principle of zoning is that it rules out the merging in one zone of variously aged permafrost massifs distinguished by structure and thickness. In addition, using these zones as an example it is possible to visualize the life history of formation of the permafrost in the region during the successive stages of its development.

MORE ABOUT THE GENESIS OF THE YEDOMA* DEPOSIT

V. K. RYABCHUN Northeastern Integrated Research Institute of the Far Eastern Scientific Center, USSR Academy of Sciences

In Professor A. I. Popov's paper it was stated that the accumulation of thick syngenetic wedge ice took place in floodplain alluvium. This hypothesis has been in existence for 20 yr.

However, a group of Magadan geocryologists led by Academician N. A. Shilo and the head of the Department of Permafrost Studies of the Northeastern Integrated Research Institute,

*Translator's note: type of terrain. In Yakutia: eroded terraces or ridges and low mountains. In Kamchatka: low ridges or hills up to 200 m in height, with Quaternary deposits. In Arkhangel'sk region: dense forest without reindeer pastures.

S. V. Tomirdiaro, has proved that this is not alluvium.

The following is the evidence against the alluvial hypothesis:

1. The coevality of deposits occurring at geomorphological levels of differing ages.
2. The absence of riverbed facies, which contradicts the generally known structure of alluvial suites of increased thickness.
3. The absence of peat, tree trunks, branches, chips, and even remains of wood bark which are characteristic of alluvial deposits.
4. The absence of the remains of fishes and mollusks.
5. The abundance of bones and even tissue remains of tundra and steppe xerophilous terrestrial fauna, in particular horses and antelopes.
6. The enormous total iciness of the deposits, reaching 90 percent by volume. This contradicts the thermokarst theory of Professor V. A. Kudryavtsev. According to his theory this ice must have been eliminated long before the attainment of these dimensions. In the structure of the yedomas discontinuities in sediment accumulation have been recorded. They indicate the generation of thermokarst, which was quickly suppressed under the dry and cold climatic conditions. So much so that it failed to melt the ice whatever the conditions of the flooded landscape.
7. The higher salinity of the yedoma deposits. On the average, it is 10 times the salinity of the contemporary floodplain alluvium of the rivers of northeastern Siberia.
8. The development of yedoma in various tectonic structures which could not have formed identically during the upper Pleistocene, which is a requirement for substantiation of the alluvial hypothesis.

As Tomirdiaro shows, the ancient landscapes of the plains at the time of formation of the yedoma were reminiscent of dry and relatively snow-free tundra steppe. Here the displacement of the steppe and tundra forms, for example, antelopes and oxen, bison and reindeer was on a scale that today is nowhere impossible in principle.

In view of the conditions of freeze-drying of the climate discovered as a result of the glacioeustatic retreat of the shelf sea and the emergence of deserts, which in the case of the lower course of the Lena River has been demonstrated to be the case by geologists of the All-Union Aerogeological Trust, the yedoma deposits could only have been formed by cryogenous-eolian processes. According to N. A. Shilo the cryogenous factor in the sediment accumulation consists in an increase in the thickness of the series as a result of the accumulation of ice and its self-filling of the hole in the ground squeezed out by the growing ice wedges.

Thus, we do not have alluvium, but a cryogenous-eolian formation.

CRYOGENOUS-EOLIAN GENESIS OF YEDOMA DEPOSITS

S. V. TOMIRDIARO Northeastern Integrated Research Institute of the Far Eastern Scientific Center, USSR Academy of Sciences

It was Academician N. A. Shilo who in 1964 was the first to speak out against the predominant hypothesis in our science of the alluvial origin of the vast upper Pleistocene loess-ice sheet complex of the eastern Siberian plains. He proved that this complex, which for the sake of brevity we shall call a yedoma, not only does not belong to the floodplain facies but in general it is not alluvium. He also showd that the squeezing out of the country rock by the ice played an enormous role in the formation of the yedoma. According to Shilo it is as if the ice wedges themselves filled up the holes left by the soil they had squeezed out and have continued to grow in the soil squeezed upward. This has happened without there being any link up with the river floodplains and alluvial sedimentation. The studies that our institute has made in the northern part of Yakutia have demonstrated that this squeezing-out process was accompanied by eolian transport of the powered rock. It turned out that the yedoma differs in a number of fundamental respects from alluvial series and suites. These differences were mentioned in the speech by V. K. Ryabchun. It was also discovered that the yedoma has many attributes in common with the ordinary loess coverings of western Eurasia:

1. Yedoma, like loess, accumulated in the dry cold crust of the upper Pleistocene, especially during the Zyryanka and Sartanskoye periods, in a setting of abruptly stepped landscapes near the regions of eolian deflation.

The existence of now-buried wind-sculptured deserts in the northern and central Yakutiya was established by the geologists of the All-Union Aerogeological Trust. The information was published by V. V. Kolpakov in the transactions of Moscow State University.

2. Yedoma, like loess, is present in the form of thick deposits not only on the high terraces but also on the interfluves, its thickness frequently increasing with the height of occurrence. The fact that the yedoma forms just such thick masses is noted on the permafrost map of Yakutia displayed in the foyer of this conference hall.

3. Yedoma, like loess, is characterized by exceptional uniformity and persistence of the grain-size and mineraliogical composition of the earth materials throughout all of the sections investigated.

4. Yedoma, like loess, does not display zonal differences, in spite of the enormous extent of its regional distribution: from the Aldan to the Arctic Ocean.

5. Yedoma contains buried soil. The Yakut scientist B. S. Rusanov even discovered a beaver dam in yedoma loess buried by the eolian process.

6. The soil composition of the yedoma, like that of loess, is carbonaceous and powdery, contains 60-70 percent silt, is porous when it thaws, and contains vertical prismatic jointing from 2 to 3 m high in thawed cemetery mounds.

7. Like loess, yedoma contains a great number of bones of the steppe animals of the mammoth complex and no remains whatever of fish and mollusks. Thus, according to all the criteria of the Kruger definition of loess, the yedoma soil deposits must be considered to be loess. Yedoma differs from ordinary loess in its high ice content both in the veins and in the partings forming the cryogenous layering. In addition, instead of the tubular pores in loess, which are cavities left by rotted roots of herbaceous vegetation, the yedoma contains these roots themselves, frequently penetrating it from top to bottom. However, these are not fundamental differences. Actually, as has already been established, the cold loess of western Eurasia contains pseudomorphs replacing melted wedge ice. According to A. A. Velichko, these pseudomorphs display signs of syngenesis. Indeed, there is no logical basis for assuming that the wedge ice grew into loess, even in the interglacial epochs. But it was during the glacial epochs that the loess accumulation itself occurred, that is, the veins and the loess must only have formed syngenetically. This means that prior to its melting in the Holocene the ordinary eolian loess of the West was a copy of the yedoma--it contained both syngenetic wedge and ground ice and also the tiny roots of herbaceous vegetation decomposing after the Holocene melting. Thus, we arrive at the conclusion that not only is the yedoma not an alluvial formation, but in general neither is it a regional, narrowly eastern Siberian phenomenon. It is a remnant of the immense primary upper Pleistocene loess-ice sheets that were present throughout the entire periglacial zone of Eurasia and America but have remained intact to this day only in Yakutia, Chukotka, and Alaska, as in the only permafrost region that did not thaw in the Holocene. Thus, the phenomenon of subaerial formation of yedoma without the involvement of alluvial processes, as established by Academician N. A. Shilo, existed in the upper Pleistocene throughout the vast areas of the then periglacial regions. The involvement of eolian processes in this phenomenon and the explicit capacity for conversion of yedoma into ordinary loess during its subaerial thawing forces us to regard yedoma as primary loess. On the 1:5,000,000 scale map of the quaternary deposits on the yedoma in Alaska is depicted as an eolian cover and that in eastern Siberia as lacustrine-alluvial deposits. Indeed, this is the one and the same formation, which should be considered a cryogenetic-eolian relict cover.

COMMENTS ON THE SPEECHES BY V. K. RYABCHUN AND S. V. TOMIRDIARO

According to many researchers the ice (yedoma) complexes of central and northern Yakutia are deposits from swampy polygonal plains. Often their sections include earth materials containing remains of moss, which as is known does not withstand salinization. These complexes are underlain by lacustrine sediments or riverbed alluvium. For some reason the speakers have not cited a single key section; they are not using the data available in the handbook entitled *Kriolitologiya Tsentral'noy Yakutia (Cryolithology of Central Yakutia)*. The editorial board is not certain that it is proper to give space to an argument of this type.

JOINT SIMULTANEOUS RESEARCH ON VARIATIONS IN THE PHYSICAL AND PHYSICO-MECHANICAL PARAMETERS AT CONJUGATE POINTS IN THE UNITED STATES, THE USSR, AND CANADA

A. T. AKIMOV Production and Scientific Research Institute for Engineering Surveying in Construction

In recent years plans have been drawn up for joint geocryological research between the USSR and Canada and between the USSR and the United States. Accordingly, it is desirable to include in the program of these studies an item pertaining to a simultaneous study of variations in the physical parameters of permafrost at conjugate points in the permafrost regions of the USSR, the United States, and Canada. By conjugate points we mean points that are identical with respect to climatic, external landscape, temperature and lithologic conditions, and are similar with respect to cryogenous textures and composition.

As the parameters of the study the following are planned: the electrical resistivity (σ) according to data from measurements in direct current and in a wide range of frequencies; the dielectric constant (ε); the polarizability; the propagation speeds and damping decrement of longitudinal, transverse, and surface waves; Young's modulus; Poisson's ratio; the density; and the moisture content and temperature of the permafrost. Among the methods of studying these parameters, the following types of logging in dry wells are proposed: electric, dielectric, seismoacoustic, and thermoelectric, as well as measurements using a gamma-gamma density meter and a neutron measurement of water content.

For conversion to true values of the parameters obtained and for their extension over an area, it is also necessary to use surface electric and seismic prospecting techniques.

The highly fruitful studies of the natural magnetic field at the ends of one magnetic line in Sogr (Arkhangel'skaya Oblast) and on Kerguelen Island (Indian Ocean) are well known. These studies were conducted in accordance with the plan for joint Soviet-French research. It is on the basis of the positive experience gained from these and other studies (for example, research connected with the Third International Geophysical Year) that this proposal is presented for consideration.

The physical and physico-mechanical characteristics of permafrost are not constant, and this must be taken into account in the many practical areas of human endeavor.

The physical parameters of permafrost vary in time and with depth. Their fluctuations are annual, long-term during the period of sunspot activity (8- to 12-yr variations) and secular (100- to 1,000-yr variations and more). The amplitude of these variations is greatest during the annual cycle and primarily in the layer of seasonal freezing and thawing. Thus the electrical resistivity during the freezing of rocks can increase by almost an order, and during the freezing of loose deposits by 2-3 orders. On the average, the speed of propagation of longitudinal waves during the freezing of rocks increases by 1.5-2.0 and more times, and with loose deposits it increases by 5-6 times and more. The Young's modulus and the Poisson's ratio of unconsolidated materials accordingly increase by a factor of 1.5-2.0, and there is a severalfold increase in the dielectric constant and the temperature. Even the density under certain permafrost conditions (for example, in the Krasnoyarsk loess) can vary by more than 1.5 times. In the layer of annual temperature variations, the variations of these parameters are of course smaller, but here also, especially in the regions where the annual temperature fluctuations penetrate deeply into the ground (due to the harshness of the sub-Arctic and continental climate), they are quite large. Thus, according to the data from routing geophysical observations at the two points in the Far North, the resistivity and temperature of the upper layers of unconsolidated deposits in the permafrost vary by a factor of 10 and more times throughout the year. The other parameters vary correspondingly. In the regions with a less severe climate the unconsolidated deposits of the upper layers of the permafrost undergo all of the transitions from an elastic body to an elasto-plastic and an elasto-visco-plastic body during the annual cycle.

The high percentage of deformations of buildings in the cities of the Far North is well known. It is also fairly high in the regions with deep (to 4 m) seasonal freezing. Much damage is caused to underground structures and various types of line installations. An analysis of their causes shows without any question that they are not always a consequence of nonuniform settlement and heaving occurring as a result of appreciable and undesirable disturbances of the permafrost conditions that had not been foreseen during the planning stage. The literature contains descriptions of deformations of buildings occurring when the foundations of the structure were maintained in the frozen state. They arise from the significant amplitudes of the variations in the parameters and the fact that they are not uniform in the various sections of the building site. Naturally, these circumstances also give rise to changes in the bearing capacity of the foundation soil, which differ in terms of time. The integral unindirectional emergency of the deforming forces originating here, acting on the accumulations principle, also leads to the above deformations when the foundation soil is in an apparently good condition. This is especially the case in areas in which the region of significant parameter variations extends to areas of deep freezing.* Hence, we are forced to conclude that the foundations of very important structures must be laid (buried) below the boundary of the region of significant parameter variations. The depth of this region is simply determined by using data from geophysical observations. The study of the variations in the physical parameters is especially important when determining the scale of intensity of areas in seismically active regions. Having regard to the above indicated variations, as is well known, the scale of intensity in individual areas of seismic microdistricts can vary by 1-2 units.

The results of determining the physical and the physical-mechanical characteristics of the range and depths of the variations in the permafrost (σ, ε, E_d) are used as the basis for planning and designing new electromechanical devices for breaking up and working the permafrost, for stipulating efficient operating regimes for these devices, together with the technical procedures to be followed in working the ground, and also for establishing optimal operating conditions of a number of electric and radioengineering structures. They are further needed to obtain a more precise understanding of the propagation of electromagnetic fields in permafrost regions, for improving the accuracy of interpreting data resulting from geophysical observations and the plane table calculations, and for insuring optimal conditions of laying various underground utilities. One of the ultimate goals of studying the parameters of permafrost is the development of reliable methods of monitoring the prediction of the variations in the geocryological conditions occurring as a result of human economic activity.

Of particular interest is the study of the direction of dynamic trends of permafrost. Are they synchronous in Canada, the United States, and the USSR or are they asynchronous with a phase shift, and what is the precise explanation for this? Also, are they cophasal in the Southern Hemisphere (Antarctica) and Northern Hemisphere (the Arctic)? None of this is known at the present time.

The study of the range of variations in the parameters and the depths at which they exhibit significant variations at reference (key) points is also important in many other studies of a scientific and practical nature.

It is desirable to organize the routine stationary observations, initially at two points, one of which should be in a region with cyclonic and other in a region with anticyclonic climatic features. At each observation point areas are selected which are two extreme representatives of permafrost, differing in terms of its temperature regime and ice content, but which have similar landscape complexes, lithologic characteristics, geological structure, temperature regime, and so on. In order to insure comparability of the research data, the observations must be conducted in accordance with a single coordinated program and with standardized equipment in holes somewhat deeper than the layer of zero annual variations in the temperature amplitudes. Furthermore, at each reference (key) point, there must be one well the depth of which exceeds the thickness of the permafrost and which is suitably equipped for long-term observations. The studies in it must be conducted by means of standard logging techniques (resistivity logging, spontaneous polarization technique, seismoacoustic, gamma, gamma-gamma, neutron, etc.).

The proposed studies will also be aimed at achieving mutual advances in the building of remote-controlled equipment with automatic recording of the readings of the sensors and the development of standard portable dry-hole logging-equipment. The latter must be characterized by a high resolving capacity, thus permitting investigation of all changes in the conditions, structure, and properties of the permafrost, and the portable equipment developed on the basis of it must correspond to all the requirements of the geocryological studies in the field.

*By the region of significant variations of the physical and physico-mechanical parameters we mean the part of the permafrost cross-section in which the amplitudes of the variations in its parameters reach consistently maximum values. In permafrost it is located in the upper part of the layer of annual temperature variations. With respect to thickness, it is always less than the depth of this layer. In the regions with deep freezing this region extends from the surface to a depth of 2.5-3.5 m.

PROBLEM OF THE PROCESS OF INTRASOIL CONDENSATION IN THE PERMAFROST REGION

V. V. KLIMOCHKIN Institute of Permafrost Studies, Siberian Division of the USSR Academy of Science

As a rule, hydrogeological research entailing determination of the groundwater budget components of the region under study also reveals the water resources forming in the soil as a result of condensation of water vapor from the air.

Up to the present time this water has not been included in the budget, although, according to our data, it makes up from 10 to 30 percent of the volume of the percolation water and, of course, must be taken into consideration. Experimental studies in very different climatic zones have indicated the important role played by condensation water in the overall groundwater reserves. In the permafrost region these studies have only just begun, but from the preliminary results of the observations it is clear that the rate of condensation in ground underlain by permafrost is appreciably higher than it is in similar deposits in the southern regions, where there is no permafrost.

For purposes of comparison we present below the data on condensation rates for various regions.

Central Buryatia (Tarbagatay)--the condensation rate is 45 cm^3 per day per m^3 of rock, mean unit runoff of condensation groundwater in suglinok-supes deposits is 0.65 liters/s from 1 km^2; Kola Peninsula (Apatity)--22 cm^3 per day per m^3; mean unit runoff in analogous formations--0.4 liters/s from 1 km^2. Leningrad Oblast (Novosel'ye)--25 cm^3 per day per m^3, mean unit runoff--0.46 liter/s from 1 km^2; Central Yakutia (Yakutsk)--60 cm^3 per day per m^3, mean unit runoff--0.85 liters/s from 1 km^2.

The data obtained by other researchers correspond to ours also, which have shown that without allowing for condensation processes the study of the groundwater budget, natural resources, and so on gives incorrect results.

A working model of a condenser has been constructed at the Institute of Permafrost Studies, and the conference participants can familiarize themselves with its design.

DECREASE IN INTENSITY OF PERMAFROST PROCESSES IN 10,000 YEARS OF POSTGLACIAL TIME AND THE OCCURRENCE OF PEAT IN THE ALTITUDE RANGE OF 1-100 METERS ON THE NORTHWEST COAST OF GREENLAND (IN THE VICINITY OF THULE AND INGLEFIELD LAND)

J. MALORY Director of the Arctic Research Center, Paris

The region occupied by Thule and Inglefield Land is located on the northwest coast of Greenland between 76° and 79° north latitude.

This region, which is made up of pre-Cambrian rock, was freed of glaciers 8,000 to 10,000 yr ago, that is, some 4,000 to 6,000 yr earlier than the more southerly region of northern Quebec. This peculiarity is explained by the specific geographic and oceanographic conditions--the existence of an unfrozen area of sea, the so-called "Norsveter," to the west of Pond Fjord. The height of the aggradation marine terraces reaches 200 m. The contemporary climatic conditions are severe--the air temperature rises above 0°C only for 8 weeks a year. In February the temperature drops to -40°C. The climate is dry--the annual precipitation is less than 100 mm. Part of the snow is swept from the plateau by anticyclonic winds emanating from the ice caps. In addition, part of the remaining snow evaporates at the very beginning of spring, even before it melts.

Thus, the surface layer of the permafrost is very dry. In addition, strong air temperature inversions are sometimes noted on the plateau. In certain parts of Inglefield Land, the vegetation is better developed in the interior of the plateau than along its periphery.

Finally, the air temperature fluctuations, in part caused by the frequent cyclonic southwesterly winds in the summer, affect only the surface layer of the permafrost.

The detailed geomorphological studies in Inglefield Land have shown that the permafrost processes within the Proterozoic peneplain are of little consequence.

This conclusion could come as a surprise if we consider the duration of the disintegration of the rock during the Proterozoic, the Paleozoic, Mesozoic, and Tertiary times and also the intensive denudation that was able to occur throughout the Quaternary, at least in the interglacial periods. However, the impression is created that an equilibrium profile developed in the rocks making up the peneplain, and that the layers outcropping at the surface offer the maximum resistance to the weathering and denudation processes. The lower the peneplain, the more clearly is this mechanism manifested.

Furthermore, the fact that during the 8,000 yr that have elapsed since glaciation, cryogenous processes have not had a significant effect on the peneplain, we are inclined to attribute this to the water-protective role of the glacial smoothing. In order to fill the pores of the rock, little water is required; therefore, on the surfaces smoothed by the glacier, the frost weathering has been sharply attenuated.

Cryogenous processes (weathering) are only active in certain rocks (clays, crystalline schists), under certain exposures swing rise to asymmetric development of the relief-forming processes (southern exposure--active; northeastern exposure--less active), and at sites of tectonic shattering of the rock and on scarps.

The impression is created that on lake shores or coastlines the permafrost processes are more active than in the interior of the plateau.

In the vicinity of Thule, a 400-m coastal spur made up of almost horizontally occurring pre-Cambrian rocks has retreated by 70 m since the degradation of the glacial cap 8,000 to 10,000 yr ago.

At the Center for the Study of Freezing and Thawing Processes, the National Center for Scientific Research in Paris, we have conducted experiments on rocks under conditions similar to those that prevail at Thule. With the exception of limestones, microporous and macroporous rock, in homogeneous carbonaceous rocks and crystalline schists, were not disintegrated by frost weathering after frequent deep or moderate cooling and moderate or strong heating in the presence of dry and moist air conditions and with complete or partial submersion in the water. From this it is evident that the specific microparameters or weather factors making rocks accessible to frost weathering continue to be unknown.

I do not wish to deny the presence of a large number of products of the frost weathering on the surface of the plateau. It is important simply to note the extraordinary complexity of the processes taking place. Under natural conditions the frost weathering of rocks takes place with difficulty. The disintegration of the rocks during their cooling is hindered by various regulating processes. Frost weathering exists, but it proceeds very slowly.

Considering all that has been stated, we are inclined to the view that frost weathering of rocks under natural conditions is the result of the prolonged interaction of many factors. Some of them are probably still unknown. Thus, for example, the strength of the rock can be reduced by repeated temperature fluctuations on their surface, the effect of atmospheric acids (nitric acid), biological corrosion (bacteria, fungi, and lichens); variations in the crystal lattice as a result of multiple repeated hydration and dehydration phases, etc. The origination of microfractures, which can hardly be linked with frost weathering, may be due to the absorption of water during melting of electrically-charged snow patches remaining throughout the summer and acting as an aggressive electrolyte. It must be remembered that there is a magnetic trap near the Van Allen radiation belt in the arctic regions. What effect the cosmic particles have on the rocks has not yet been established. All of these factors may not prove to be so important, but we have singled them out in order to emphasize that studies in this area should be encouraged.

The internal equilibrium state of matter is being continuously disturbed. The variations in the state of the rocks results in a redistribution of their constituent parts. "Inasmuch as the melting of ice is caused by a decrease in system volume, this process if accompanied by the formation of pores and cracks, upon which a redistribution of moisture takes place in the rock, that is, it leads to additional migration.* The kinetics of these processes are not understood.

Increasing the surface strength of the rocks, these factors (the variable effect of which is still unknown) will undoubtedly contribute to the water penetrating the rocks by way of the cracks.

Thus, in many cases, freezing and melting merely completes the fragmentation of the rocks, which, under the influence of an entire set of factors, has already been in progress for a long time.

At the time of our latest expeditions to Thule, in 1967-1969 and 1972, we took samples of peat from 14 beds in the environs of Thule. Radiocarbon dating and palynological studies were made. Bearing in mind the subject of this brief report, we would note only that all of the occurrences of peat were located on the coast in the 1-100-m altitude range and that the most ancient of them dates from 4,700 yr. On the upper plateau at an altitude from 200 to 400 m, according to our data there is no peat.

Our preliminary conclusions are the following. The peat bogs located below 180-200 m could only have formed after the transgression. They began forming at least 5,000 yr ago (Point No. 35 in the vicinity of Thule-Dundas, Wolstenholm Fjord, 6 km west of the present edge of the ice cap, 3 km from the sea, lower level, absolute alti-

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*B. A. Savel'yev. Peculiarities of ice melting processes in an ice sheet and in permafrost. Problemy Severa, no. 1. Moscow Izd-vo AN SSSR, 1958.

tude 95 m, depth at which sample taken 2 m, absolute age 4,700 yr).

So far as I know, in the 100-400-m altitude range there are no peat bogs. More careful studies, which I should like to make, will possibly reveal shallow peat bogs in the 100-200-m altitude range. However, I still do not have the facts; this is merely a suggestion.

The reasons for the absence of peat bogs in the 200-400-m altitude range could be either the late freezing of these regions from ice or the excessively steep slope that had formed as a result of initial exceptionally rapid glacio-isostatic uplift of the land and had prevented at first stagnation of the water and peat formation.

During the next few years, I should like to perform systematic studies of the relief of regions in the 200-400-m altitude range that are situated above the most ancient boundary of the transgression and were freed of ice very early, about 6,000 to 7,000 yr ago. It is there that the problem arises of estimating the role of steep glacio-isostatic slopes in the development or absence of peat bogs 6,500-8,000 yr ago.

In the laboratory of the National Center for Scientific Research, the dating of shells gathered in the vicinity of Thule 60 m from the mouth of the three watercourses is in progress. These data will characterize the rates of isostatic uplift of the land.

ORIGIN OF THE ICY SILTSTONES OF NORTHERN YAKUTIA

V. N. KONISHCHEV *Moscow State University*

The genesis of a siltstone series, including thick ice veins, is one of the most interesting and central problems in cryolithology. In recent years various researchers have formed diametrically opposite opinions on the origin and conditions under which these deposits evolved. It is necessary to make a number of remarks with respect to the eolian hypothesis and also the arguments advanced by its proponents. It is believed that the icy siltstones cover the interfluves, the variously aged geomorphological levels of the Primor'ye lowlands, and the mountain slopes in the form of unbroken mantle. The basis for this view is the exceptionally widespread development of the thermokarst relief forms consisting of alasses and lakes that break up the initial surface into individual yedoma outliers. In other words, it gives a picture of the very simple geological-geomorphological structure of the Primor'ye lowlands of the Yakutia.

Detailed geological survey work conducted in recent years has made it possible to establish that in the lowlands and foothills the original self-contained surfaces of the lower, middle, and upper Pleistocene and Holocene accumulations that consisted of variously aged series of alluvial and lagoonal marine (in the North) sediments of the sandy- and clayey-silty composition extended over large areas.

Although the icy siltstones are distributed within the lower and middle Pleistocene surfaces, they do not form a continuous cover but occur as lenses of different size embedded (incised) in lower and middle Pleistocene deposits, respectively.

In addition to the series of icy silstones that are highly uniform in composition, in the Yano-Indigirka lowland deposits are widely represented that are ice-rich but predominantly of sandy composition, including thick polygonal-wedge ice and sometimes lenses of injected ice. A series of icy sands 25-30 m thick makes up an independent upper Pleistocene level, which is traced along the valleys of an entire series of rivers. The geomorphological position and lithologic peculiarities of the series of icy sands leave no doubt as to their alluvial origin. The spore-pollen spectra and faunistic remains of these two types of icy series are identical, which is indicative of their coevality. This indicates that at the time of formation of the exceedingly icy Pleistocene deposits the large and medium-sized rivers continued to function, a series of icy alluvium of predominantly sandy composition having accumulated in their valleys. This whole intricately built, lithologically dissimilar and variously aged set of deposits has been intensively reworked since the beginning of the Holocene right up to the present time and is being fragmented by thermokarst, erosion, and solifluction processes. In places the reworking processes have attained such grandiose dimensions that it is difficult to establish the actual ratios between series of different age. In more than half of the cases the yedoma outlies are made up of sand or stratified sandy siltstones of middle Pleistocene or lower Pleistocene age, the ice content of which is appreciably lower than that of the ice-saturated upper Pleistocene series.

Nevertheless, the disintegration of these outliners is also proceeding quite rapidly, inasmuch as, along with the thermokarst processes, an important role in the expansion of the alasses is played by solifluction and nivation processes.

We associate the formation of the icy siltstone series with the development of ancient alas relief forms. Their processes facies analog is the subaerial syngenetic icy deposits of aleurolitic composition forming on ground with low center polygons constituting the alas floodplain of the contemporary alasses. The development of the alas forms is taking place asymmetrically, coupled with both erosion-thermokarst and denudation-aggradation processes that, depending on the external conditions (climate and tectonics), can lead either to a deepening and expansion of it or to its filling. In the course of development of the alas, a regularly patterned constructed series arises--at the base the subaqueous lacustrian deposits, which are overlapped by the deposits and facies of the polygonal alas floodplain. In contemporary alasses the latter have a thickness to 5-7 m; the potential thickness is determined by the depth of down cutting of the alasses and is 20-30 m. These ratios could also have obtained in the past. The icy siltstones are the upper horizon of the ancient subaqueous-subaerial complexes of the alas deposits and are characterized by basically the same features of vertical facies-cryogenous structure as are the deposits of the contemporary. Thus, both types of upper Pleistocene icy deposits have contemporary facies analogs.

The peculiarities of composition of the icy siltstones are attributed to the uniqueness of the cryogenous weathering of the original series, the products of the most recent redeposition of which were the material from which they were formed.

The cryogenous reworking of supes-suglinok soils leads, as is well known, to the origination of a secondary product consisting almost entirely of coarse aleuritic fractions.

The mineralogical analysis of the sandy fraction (0.1-0.05 mm) of the siltstones revealed that they have just as high a content of heavy minerals, including ore minerals, as do typically aqueous and slope deposits. This contradicts their eolean origin.

As an argument in favor of the eolian origin of the siltstones, some researchers (S. V. Tomirdiaro, V. K. Ryabchun) cite the higher mineralization of these deposits along the Oyyagosskiy Yar section. It must be remembered that the siltstones here are embedded in the lagoonal-marine deposits, the mineralization of which is twice that of the siltstones. Naturally, with their disintegration and ensuing redeposition on the surface of the ancient alas floodplains, the secondary series being formed will also be very highly mineralized.

In addition, in the coastal regions marine salts can get into the continental deposits during the pulverization process and the increased salinity could be due to the existence of an exposed sea surface rather than to drained shelves. As numerous analyses have demonstrated the mineralization of the siltstones in other areas is very low and does not exceed that of various types of deposits (slope, alas). In the well-known Mus-Hai ratio and in a whole series of sections on the Yano-Omoloy interfluve, the siltstone mineralization does not exceed 3 to 4 mg/eq per 100g, although in individual cases it increases to 10 mg/eq. Thus, the mineralization of siltstones depends on the mineralization of the deposits from which they were formed during the course of their reworking.

Also contradicting the eolian genesis is the exceptionally high content of organic material in the form of plant detritus and humus compounds.

The species composition of the faunistic remains peculiar to icy siltstones and sands does not contradict their alluvial and alas genesis. It reflects the variety of landscape conditions at the time of their formation--ranging from the relatively dry more ancient alluvial surfaces, on which animals such as the antelope could exist, to the hydromorphous landscapes of a low center polygon floodplain consisting of ancient alasses and the valleys of major rivers inhabited by reindeer, bison, etc.

ORIGIN OF THE DEPOSITS OF THE YEDOMA SUITE ON THE PRIMOR'YE FLOODPLAIN OF NORTHERN YAKUTIA

A. I. POPOV   Moscow State University

S. V. Tomirdiaro refutes the previously postulated lacustrian-alluvial genesis of the icy aleurities of middle to upper Pleistocene age widespread on the plains and in the foothills of northern Yakutia and frequently containing a lattice of thick ice veins. As is known, in the literature, this icy aleurite complex is called "yedoma." Tomirdiaro ascribes to it an eolian genesis and states that it was formed under dry cold climatic and dry tundra-steppe conditions. He calls them loess deposits and compares them with the loess of the Russian plain.

Tomirdiaro's position on this question, which is presented as a new well-founded concept worthy of replacing the old one, has been discussed in the speech by V. K. Ryabchun at this conference.

Without repeating all of the arguments advanced by the authors of the eolian and dry tundra steppe hypothesis (they have been published), I consider it necessary to present the facts and the logical conclusions derivable therefrom, which sharply contradict the aforementioned hypothesis.

First and foremost, it is necessary decisively to refute the suggestion that the yedoma aleurities occur as a blanket on the different elements of the relief. They comprise terraces of various levels and their benches are recumbent or composite in structure, as has been well demonstrated by V. N. Konishchev.

Ordinarily ice richness of the aleurities (saturation with streaks of segregation ice) and also their significant peat content and enrichment with minute plant remains are incompatible with a dry climate and the xerophylous landscape. The fine plant detritus is unquestionably a product of floodplain sediment accumulations.

The lattice of polygonal wedge ice forming a massive ice frame in the aleurites could not have originated under dry surface conditions. It is well known that polygonal wedge ice is currently forming only when the moisture content of the substrate is fairly high, that is, in floodplain, swamp, or alas deposits and also certain slope deposits. In this case it is hardly possible to ignore the principle of actuality.

The lenses of layered allochtonic and only partially autochtonic peat embedded in the yedoma aleurites are clearly of lacrustrine-alluvial-swampy origin, being the facies version of the valley complex. Large plant remains--branches, brushwood, and at times entire driftwood logs are encountered in the aleurites.

The statement by a number of authors regarding the absolute homogeneity of the aleurities is not well founded: layering is usually pronounced, although sometimes it is of a latent nature. Most significant is the fact that the aleurites form interrelated facies varieties differing in their attributes (layering, texture, and so on) and corresponding on the whole to a lacustrine-alluvial complex. The nondetection of deposits of a riverbed facies is also apparent.

The yedoma aleurites are made up of fine dustlike particles, but this is not loess in the usual sense of the term. They have neither the porosity nor the vertical columnar separation and other loess attributes. A comparison of the yedoma aleurites with the loess of the Russian plain is groundless in terms of many of the attributes.

The salinity of the aleurites, which in Tomirdiaro's opinion is such an important index of their eolian origin, is local and, as V. N. Konishchev points out, it is borrowed from the underlying marine deposits. In the known Mus-Khay outcrop of the Yana River and in a number of other places the aleurites are not saline.

The preservation of animal remains in the aleurites (primarily mammoths and rhinoceroses, rarely horses and other animals) and frequently their soft tissues would be impossible under the conditions of a dry active layer. This effect could be achieved only under conditions of high moisture in the active layer.

P. I. Shestakov (1914), who studies the fatty tissue of the Berezov mammoth embedded in similar deposits, concluded that the carcasses of the mammoth had been soaked for a long time in water at low temperature (before it is encased in the permafrost). This fact hardly correlates with the dry eolian sediments.

It is known that the yedoma aleurites contain numerous bony remains of animals of the so-called mammoth complex. The cold-living, frequently tundra forms, such as the mammoth, the reindeer, the ox, bison, horse, woolly rhineroceros, lemming, and so on, are markedly predominant in it. The xerophilous antelope is now added to this list. On the basis of the latter, it is concluded that the landscape was of the dry steppe type when it was inhabited by the mammoth complex.

One can only be amazed at the rashness of such a far-reaching conclusion. It is as if the Ussuriyskiy tiger had become extinct long before our time and on the basis of finding it in the deposits of the Far East we had drawn a landscape of Bengal jungles for the snow-covered Sihote-Alin in winter. In all probability we shall have to regard as possible the adaptation of the antelope to tundra conditions during the middle and upper Pleistocene in the northern part of Yakutia.

The icy yedoma aleurites are drawn towards the large valleys of the Yana, Indigirka, and Kolyma rivers, although (according to V. N. Konishchev and M. A. Velikotskiy) they are limited in their distribution to the interfluves. It is this fact, in our view, that affords the best evidence of the alluvial (floodplain, lacustrine) genesis of the icy yedoma aleurites, taking into consideration, of course, all of their aforementioned characteristics.

All that has been said above does not mean that the question of the origin of the yedoma deposits has been finally decided and that there are no unresolved problems. Of course, a great deal must still be done in order to fully understand the conditions under which they formed and, in particular, the peculiarities of formation and the mechanism of syncryogenic growth of the polygonal-wedge ice during the course of yedoma sedimentation. The age of the yedoma complex is also not entirely resolved.

MECHANISM RESULTING IN CONVEX SURFACE FORMS OF ARCTIC PLAINS

B. V. UTKIN All-Union Research Institute of Roads and Highways

In considering any natural phenomenon, in most cases the researcher is first of all interested in its external characteristics--the shape. Visual perception gives the observer almost 90 percent of the information from external sources. This proves the correctness of taking the geometric factor into consideration not only in the structure of matter, which J. D. Bernal pointed out in his time (1962), but also when studying large-scale natural phenomena.

It is on the basis of this geometric characteristic that we have lumped together the following ice and ground-ice formations, which more or less clearly outlines convex or planar-convex surfaces and have them called lens ice: seasonal ice mounds, ice mats of ice dams, perennial ground-ice mounds, or "bulgunnyakhi," pingos, quicksand and peat mounds, frost salses [mud volvanoes], and ice lenses in the ground at the base of installations.

In the course of the freeze-up, in the backwater section, as in the places with a smooth water surface, the elements of the solid phase freeze together and form a primary convex ice cover. In the initial stage the process involved in the shaping of the ice lenses is not connected with thermodynamics: the frost merely consolidates the convex shape of the mobile equilibrium and the backwater curve in the form of a stable ice shell. If the solid phase turns out to be frazil, an ice mat originates over the ice dam. If the free crystals of the frazil ice freeze together, the primary layer of the shell of an ice mound originates. This shell is able to grow in the event of the channel becoming frozen to the bottom, in accordance with the hypothesis of S. Taber, under the influence of crystallization forces.

The mechanism whereby frazil-ice mounds form is simpler, inasmuch as there are no crystallization forces--a continuous shell is absent or it does not participate at all, for the movement of frazil ice on the rivers is not always accompanied by severe frosts.

During the movement of microflows of free water or quicksand in the ground, according to M. I. Sumgin the backwater sections and the subsequent processes resulting in the formation of ice lens-micromounds can originate as a result of the extinguishing of energy at the points of narrowing of the subterranean ground channel by rock fragments, collision of the subterranean flow with other flows, etc.

Under laboratory conditions, the ice mound was obtained, which had originated as a result of extinguishing the kinetic energy of the radially directed (toward the center) water flows. The source of supply was snow saturated with water occurring in the form of an annular ridge in a pan heated from below and set in the open at an air temperature of -14°C.

In contrast to the generally accepted method of investigation in geocryology, in which attention is mainly directed towards explaining the origin of a particular permafrost phenomenon, the goal has been set of revealing the mechanism involved in the shaping of a convex surface that is characteristic of all such mound formations. The icing mound was taken as the basic object of investigation, in that it is the most typical representative of an ice lens and yields the largest amount of information.

It has been established that the convex surface, which is characteristic of an ice lens, is predetermined by the convexity of the free surface of the liquid flow (the water, the quicksand, or the peat mass, the brine, etc.) at the backwater points, that is, at the points where its kinetic energy are extinguished.

It is known from hydroaerodynamics that in such cases the extinguished (lost) kinetic energy translated into work directed towards overcoming an obstacle on the path of the gravity flow--the hydraulic local drag, which had a braking or blocking effect on it. The equivalent of such effects on the gravitational flow is the inelastic oblique impact of two systems (spheres).

The amount of extinguished (lost) kinetic energy depends on the ratio of the velocities of the systems, the ratio of their masses, and the magnitude of the collision angle.

The effect of extinguishing the kinetic energy is estimated by the formulas of impact theory, which are presented in the paper. When changing from similated to natural conditions, the flow rates of interacting flows are considered equal to the masses of the corresponding spheres.

Under natural conditions in the presence of groundwater flows directed toward the center, such formations can originate in depressions and basins formed during melting of the ice in the surface layers of the permafrost. The annually recurring process of the formation of a state of ice lenses displacing the frozen layers of melkozem upward, together with the ice lenses previously originated therein, can lead to the occurrence of perennial mounds, and in particular to "bulgunnyakhi." A model of the process of bulgunnyakhi growth appears analogous to the American method of constructing buildings by assembling each floor on the ground, beginning with the top ("upside down") with subsequent raising of them to the planned elevations by hydraulic jacks.

The origination of the convex surface forms of the Arctic plains are in turn conducive to sorting of the rock material as a result of its creeping along the layer of thawing melkozem towards the edges of the mound, which leads to the development of solifluction processes and, in particular, to formation of "sorted circles" and other tundra relief forms (according to G. Khegbom, O. Nordenshel'd, E. Shenk *et al.*).

The sharp transitions from frosts to thaws and again to frosts, which are typical of many areas with unstable winters, especially the central belt, promote the origination of ice lenses in the ground, and in particular in railway and roadbeds.

In such cases, ice lenses or micromounds (according to Sumgin) form in the flows of freezing meltwater, which being impeded by the temporary seasonally freezing layer--the "bottom"--moves along its surface in the form of microflows. This movement of meltwater is promoted by the longitudinal gradient of the road, the accumulation of wet snow on the shoulders, and also the moving motor vehicles or trains themselves. The water feeding the ice lenses can in this case go under the ice shell sideways, along the surface of the bottom (according to P. Saipl) or be sucked through the capillaries (according to R. Ryukli) from the as yet unfrozen layers below (according to N. A. Puzakov).

Thus, the ice lens forms in gravity flows (water, quicksand, etc) at the backwater locations caused by various obstacles--giving rise to local hydraulic drag.

The convex shape of the surface of the ice lenses is predetermined by the convexity of the freezing surface of the backwater curve flow. The growth of the ice shell is caused by the crystallization forces in the open system; the process is possible in a wide range of negative temperatures and flow rates within the limits of applicability of the D. Bernoulli equation.

EXPERIMENTAL STUDIES OF THE EFFECT OF PRESSURE OF THE MECHANISM OF THE FROST HEAVE AND THE GROWTH-MELT RELATIONSHIP OF ICE IN THE GROUND AND IN GLACIERS*

F. J. RADD AND D. H. ORTLE   Continental Oil Company, United States

The various phenomena of frost heaving are well known to the builders of roads and airfields, when installing foundations in the cold regions of the Earth. Agronomists and geologists also know about the effect of this phenomenon, as a result of which the upper layers of the soil structure are disturbed and erosion occurs. This phenomenon appears to be quite simple. Some of the wet, fine-grained soil increases in volume when freezing, giving rise to deformations--known as frost heave, the force of which is so great that it is capable of disintegrating the roadbed and the foundations of buildings.

Many papers have been published on the frost-heave phenomenon. The United States, Canada, Sweden, the USSR, and the countries that are studying the properties of soil in the temperate climatic regions and in the permafrost regions have undertaken numerous special studies aimed at discovering the nature of this phenomenon. There is a multiplicity of theories, hypotheses, and experiments that have a bearing on the problem of explaining the effect of frost heave. In spite of a fairly large volume of data having been accumulated, no single opinion exists on the question of the mechanism of frost heave. It is for this very reason that a need has arisen to conduct some experimental studies relating to the following basic problems:

1. If we consider pressure as a factor influencing the freezing of the ground, then what is the relation between the pressure and the formation of ice lenses? Is there proof of the existence of a special property that is not determined by the ice content and that is not predictable thermodynamically?
2. Is there something specific with respect to soil containing fine particles such as clay, clay-silt, or silt? How do other systems of solid particles behave?
3. Is there anything specific with respect to water at the time of formation of ice lenses in the clay and the clay-silt? Do other (2-1) liquid-solid systems manifest such properties?
4. In what way and why is the process of the formation of lenses only a special case of crystal growth?
5. What is the interrelation in this case between pressure-volume-temperature and time?

DEFINITION OF THE PROBLEM

The clays, silts, and the fine-particle soil resembling them can increase noticeably in volume during freezing. It is characteristic that the ultimate water content in frozen ground is 4-5

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*[Refer to a paper on p. 377 in the North-American Contribution Volume by the same authors.]

times greater than the original content. The expansion of the volume of the water-ice system is approximately 10 percent and takes place when there is an intrinsic moisture content in the freezing ground and when water is seeping in from outside. The sheet capillaries continuously forming and persisting at the ice-clay interface are characteristic of systems by which an additional amount of water is continously reaching the freezing interface. The main problem is to ascertain the reason for the formation of layered capillaries at the water-ice interface. In a fine-grained system in which ice lenses form, the process of ice growth itself is obviously regulated by direct and axial heat fluxes. On the one hand, in a coarse-grained system the regulation takes place as a result of local, higher thermal conductivity of the solid particles in comparison with the specific thermal conductivity of the ice being formed. These differences are illustrated in Figure 1. Thus, the regulation of the freezing front is evidently determined by the size, shape, packing, and properties of the enlarged solid particles and not by the requirements of the growing ice front. On the other hand, in a fine-grained system the particles are equal or smaller in size than the growth step at the ice lens surface and, consequently, they cannot cause the appearance of projections with a higher specific conductivity beyond the ice front. In this case the entire growing ice surface can be supplied with liquid without disturbance of the heat flux and just as the ice front requires it. The planar ice front is constantly maintained in the layered capillary. Without this important transporting capacity of the fine-grained system (lowering of the freezing point of the boundary layer and the tensile stress of the film), we would not be able to have a layered capillary or the lens-type crystal growth. The tension of the water film does not present a problem--it is approximately 1,500 to 6,000 atmospheres. It appears to us that lowering the freezing point of the boundary layer liquid creates a vacuum and makes the pore space the primary factor in the behavior of the fine-grained systems (as a result of increased specific thermal conductivity of the particle). In the coarse-grained system, the particle size becomes decisive and not the magnitude of the vacuum. Thus, we decided that in our experiments relating to the effect of the pressure we would study the melt-growth relationship at varying pressure.

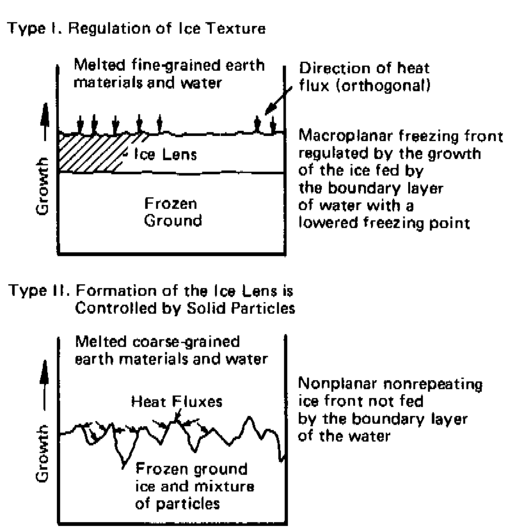

FIGURE 1 The two main types of freezing of the ground.

Experiments were first performed to study the formation of ice lenses in silica gel, in ordinary sand, and in diatomaceous earth: In the first and last of these experiments lenses formed. Studies were also made of the formation of ice lenses in three anhydrous liquids with Linde high purity alumina: Without water, without clay, and without alumina we were still able to obtain "ice" lenses in nitrobenzene and in o-chloroaniline and formamide. Genuine ice lenses were formed in fine carbon-black, MgO and the [SiC] carborundum screened through a number 600 mesh. From these experiments we concluded that the formation of the ice lenses is not limited to the clay-water system, but it is a common property of a system of finely dispersed rocks with constant inflow of cold and liquid. The decisive factors are the regulation of the interface by the growing ice itself, the characteristics of the macroplanar heat flux, and the liquid film feed at the interface rather than the specific properties of the solid particles and the liquid.

EXPERIMENTAL UNIT (INSTRUMENTATION)

The design of the unit must meet the following requirements: The upward force that develops during growth of the ice lenses must be measured in spite of the fact that the ice lenses and the frozen ground will freeze to the wall of the chamber; the water must enter the soil freely and leave it for the feed source when high soil pressures are developed in the chamber; the soil must be protected from escaping; some devices will be needed for determining the growth of the ice lenses and the resulting compaction of the soil as the pressure rises. The cooling source must provide close temperature control. Figure 2 illustrates the unit in which most of our experiments were performed. This is a water-permeable chamber with constant volume. The cooling source was a large copper cylinder using Freon-12 refrigerant in the coils and insulated on the outside. An integral stainless steel piston 28.575 mm in diameter and 25.4 mm long was used as the bottom of the cell and conducted heat for the cooling force. A 38.10-millimeter hole was drilled in the lucite board 152.40 mm^2 in area and 72.60 mm thick so that it could serve as the transparent container. The chamber walls about 63.5 mm thick provided sufficient strength when testing for the pressure foreseen by the experiment. In addition, they were well enough insulated to insure that the chamber should not sweat on the outside during the investigation.

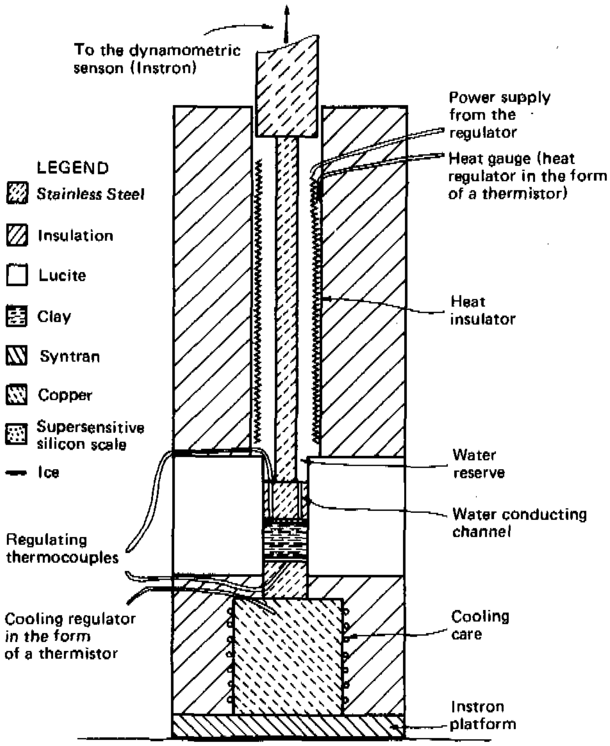

FIGURE 2 Schematic representation of the chamber used to investigate an ice lens.

The upper part of the chamber comprised a layer of filter paper as protection against the permeability and a cermet steel wafer 28.575 mm in diameter and 3.175 mm thick. This wafer was supported by a stainless steel disc 28.575 mm in diameter and 25.4 mm thick with six 3.175 mm holes drilled through it. The water tank was the space between the inside walls of the lucite board and the upper and lower plugs. The openings in the upper plug and the space between the cermet particles of the permeable wafer were channels for the groundwater supply in the chamber while the filter paper and the cermet stainless wafer held the soil. In the chamber constructed in this way, the ice lenses and the soil below it could freeze to the wall of the chamber without affecting the developed pressure. This occurred because the clay above the ice lenses could move freely and transmit the pressure developed from compaction of the soil at the time of growth of the ice lenses. Inasmuch as the chamber walls were made of a lucite block and had polished faces, we were able continuously to observe the growth of the lenses and the other physical phenomena occurring inside.

For measuring the temperature on the inside surface of the upper part and the bottom of the chamber, copper-constantan thermocouples were installed. The internal temperatures of the clay were not measured. The temperature of the space above the chamber and around the rod transmitting the pressure to the dynamometric Instron pickup was constant in order to prevent fluctuations of the temperature at the front of the ice lens caused by variation of the temperature in the housing. The experimental chamber and the equipment connnected with it were assembled on the Instron Company's TT-1 physical test model installed on planking at the point normally used for the dynamometric compression sensor. This sensor was placed above the bottom and attached to a cross piece secured in such a way that it could move above the experimental chamber. The Instron model was used to record the force developed by the growth of the ice lenses. It was also used to maintain the previously selected load. In spite of the fact that the clay was compacted for many hours, the model indicated an increase in force above the established point of growth of the ice lenses for several days more. Most of our experiments were recorded on color film (Bolex 16 mm) at specific time intervals. The film turned out to be very valuable, for some of the physical processes and factors that could not be observed visually during the 30-day experiment were recorded on the film and could be closely examined in 3 1/2 minutes.

RESULTS OF THE EXPERIMENTS

For one experiment the sample of saturated Transco cavern clay [silt] was loaded into the chamber by a spatula; it was evened off at the bottom of the chamber and was compacted by the upper stainless steel plugs of the chamber until the layer became 20.32 mm thick; then the lower stainless steel plug was inserted. All of the remaining samples were loaded in the same way. This experiment lasted 400 h and a pressure of 84.372 kg/cm^2 was achieved. By using a lucite chamber we were able to observe the growth of the ice lenses. During this experiment the necessity for linear regulation of the cooling became clear because the temperature variation rate had decreased sharply by the end of the experiment. Then another experiment was performed that lasted 200 h and resulted in an increase in pressure to 68.552 kg/cm^2. It became obvious that the rate of development of the freezing front was closely connected with the developed pressure. Our next experiment with the same equipment and type of soil was programmed for a longer period. It lasted 700 h, and the pressure was increased to 94.215 kg/cm^2. For our next experiment, a 16-mm movie camera with a telephoto lens was installed so that pictures could be taken of the Transco clay-silt through the lucite wall of the chamber. The automatic timer ensured that the frames were taken every 12 min. This experiment lasted 1,000 h and a pressure of 105,465 kg/cm^2 was achieved. In Figure 3 we have the relation between the temperature, pressure, and time, and in Figure 4, the relation between the volume of the ice lenses, pressure, and time. The data with respect to volume were obtained from the measurements made by the Bolex photographs. As can be seen from the volume curve, the ice lenses still continued to grow for approximately 800 h to a pressure of 91.403 kg/cm^2.

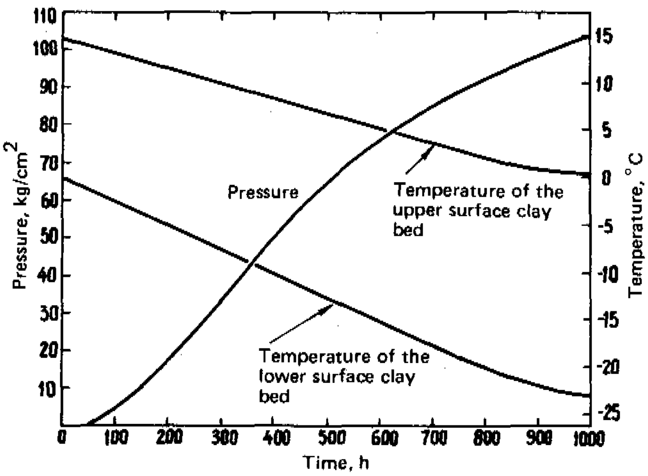

FIGURE 3 Growth of the ice lens in the Transco clay-water system. Pressure-time-temperature curves of the Instron instrument.

At the beginning of the experiment it was noted that in the ice lenses there are vertical columns of air through which small pieces of clay move upward. Here the space under a piece of clay was filled with pure ice. When the ice lenses became clear, the movement of the clay particles through the now solid ice lenses was no longer observed.

The next question to which we needed an answer was the following: Will ice lenses form and grow in a clay system already under high pressure? In order to find the answer, we performed the same experiment as the last except for the fact that prior to the beginning of freezing the clay was subjected to a pressure of 70.31 kg/cm^2 by the Instron machine. The Instron load cycle regulators were installed so that the pressure drop, as a result of compacting the clay, could force the instrument to apply a load that would again raise the pressure to 70.31 kg/cm^2, although a rise above this value would not cause any corrections on the part of the Instron instrument (Figure 5). As is obvious from these curves, the pressure actually increased to 126.558 kg/cm^2, but with a small delay at the beginning of the pressure rise. The initial point of 70.31 kg/cm^2 is considered to be the equilibrium point of the melting on the P-T curves. After 275 h of the experiment the pressure began to rise, thereby indicating the beginning of lens formation. At the bottom of the clay when the ice lenses had formed, the temperature was -7.78°C. This experiment was also recorded on movie film at specific time intervals.

As a result of the experiments, two questions arose: Was the delay at the beginning of the growth of the ice lenses a result of the absence of nucleation points of ice crystals that could have continued to grow or was the delay caused by a reduction in the freezing point by means of the pressure applied to the clay? In order to answer these questions, we decided to grow a lens of corresponding size before beginning to apply steady pressure and to trace whether it continued to grow. At the beginning of this experiment of the lower surface of the clay under a pressure of 2.812 kg/cm^2, an ice lens 2.54 mm thick was grown and then a pressure of 105.465 kg/cm^2 was applied. The clay began to compact and the temperature at the bottom of the chamber began to drop, and 15 min later it began to drop in the upper part of the chamber. Then after about an hour, at the same pressure, the ice lens began to melt, and after 5 h it had completely melted (Figure 6). Figure 7 shows a continuation of the experiment in growing the ice lens as a result of which, after about 200 h there was a rise in pressure above 105.465 kg/cm^2. This helped to convince us that in our system the freezing point dropped under pressure.

We had to establish the curve for the freezing point depression under pressure in the Transco clay. The temperature corresponded to

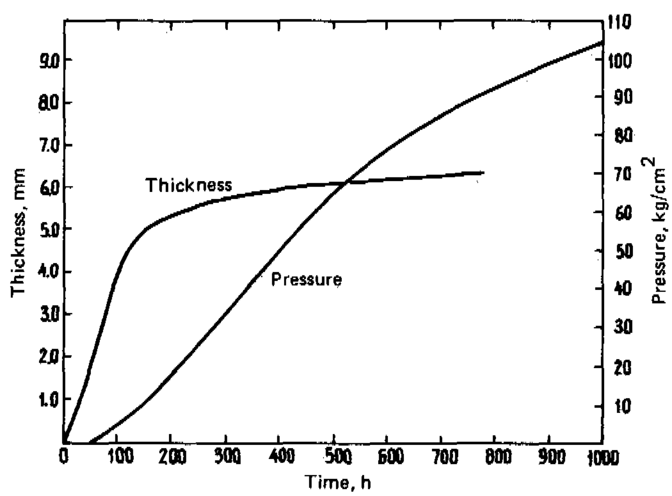

FIGURE 4 Pressure-time-thickness curves with growth of the ice lens in the Transco clay-water system.

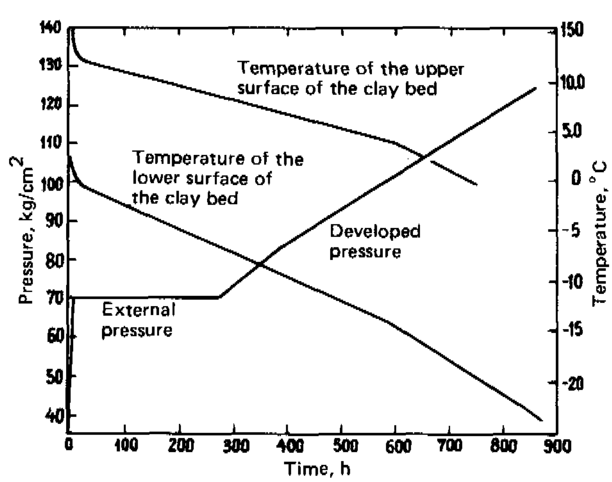

FIGURE 5 Growth of the ice lens in the Transco clay-water system at a pressure of 70.31 kg/cm^2.

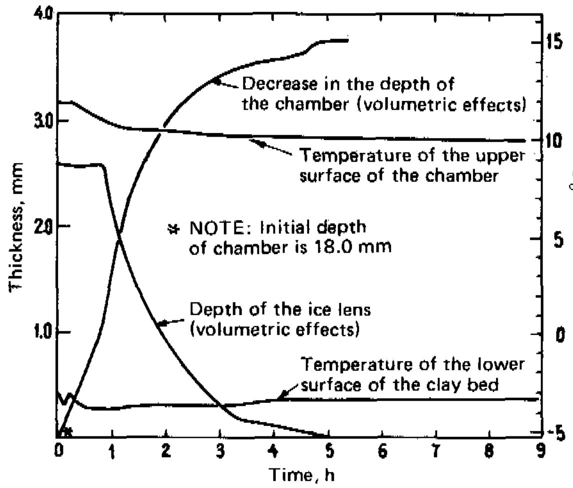

FIGURE 6 Melting of the ice lens and temperature effects at a pressure of 105.4 kg/cm^2 during growth of the ice (approximately 4.2 kg/cm^2).

the beginning of growth of the lens at the bottom of the chamber for pressures of 0, 70.31, and 105.465 kg/cm^2. Our next experiment was designed to obtain the points 35.155, 105.465, 140.62, and 175.775 kg/cm^2. Although the points for 70.31 and 140.62 kg/cm^2 were unfortunately missed, to the already existing points with respect to temperature we added 35.155, 70.31, and 175.775 kg/cm^2 (Figure 8).

During the last experiment the pressure was increased from 175.775 to 182.81 kg/cm^2, but at the low temperatures required to increase the pressure from the 210.93 kg/cm^2 point, the cooling unit operated poorly.

We do not consider that 182.81 kg/cm^2 is the maximum pressure that can be developed by the ice lens in the Transco clay, but we are certain that this is the limit for the experimental unit that we used.

DISCUSSION OF THE RESULTS

It is noticeable that there is a systematic $P(\frac{dv}{dt}) = k$ curve, for the pressure-time response curve in six analogous experiments. In Figure 9 we see that the slower the freezing rate, the higher the final pressure in our experiments. The development of the ice lens pressures at reduced temperatures (and high pressures) confirms the effects of lowering the freezing point of the boundary layer and the important values of the empty spaces in the fine-grained particle system.

The pressures developed in the frozen clay and frozen sand turned out to be entirely different in magnitude. The final pressure ratios were 35:1 or more, giving a higher value on control of the ice front with fine particles than on control with coarse particles. No large differences were noted between the freezing of the large particles of coarse-grained and fine-grained sand having particles of regular acicular shape.

Figure 6 shows the effects of the melting caused by a pressure of 105.465 kg/cm^2 to which the ice lens was subjected which was grown at a pressure of 4.219 kg/cm^2. Hence it is evident that sudden melting took place as a result of applying a high pressure. After checking all other possible experimental explanations of this phenomenon, it was discovered that the only correct one confirming the results is the view that the melting is caused by the free liquid pressure.

Although point 3 (See Figure 8) at a pressure of 175.775 kg/cm^2 indicates the actual rise in pressure, contrary to the other cases there was no clearly visible ice lens. Nevertheless, the clearly recorded pressure rise shown in Figure 10 indicates that it is completely substantiated experimentally. The intake of water and its effects were not limited, and the ultimate tensile strength (~6,000 atm) was not even remotely achieved in the course of the investigation (the maximum was 200 atm).

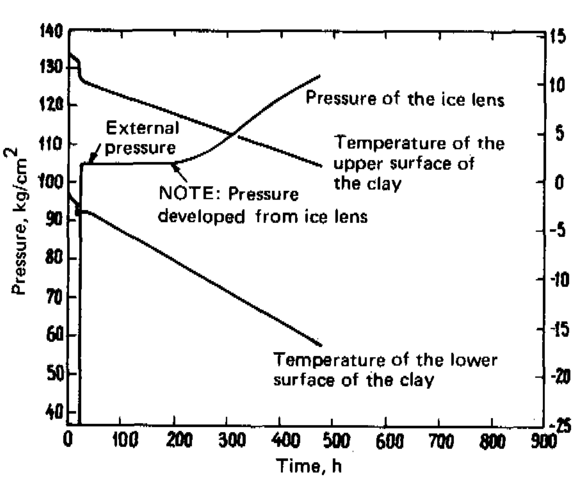

FIGURE 7 Growth of the lens at a pressure of 105.5 kg/cm^2.

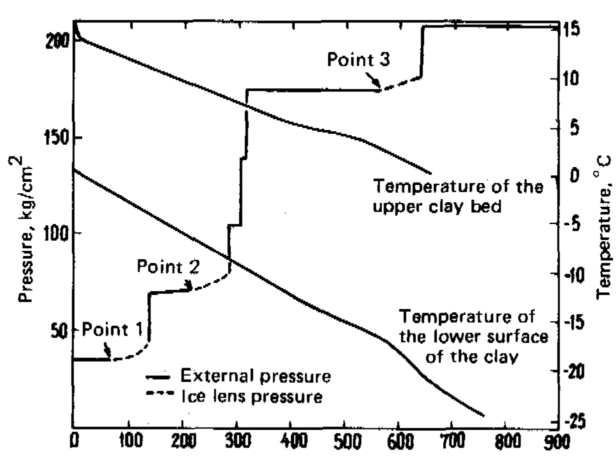

FIGURE 8 Multiple point experiment of the lowering of the freezing point as a function of pressure on formation of an ice lens.

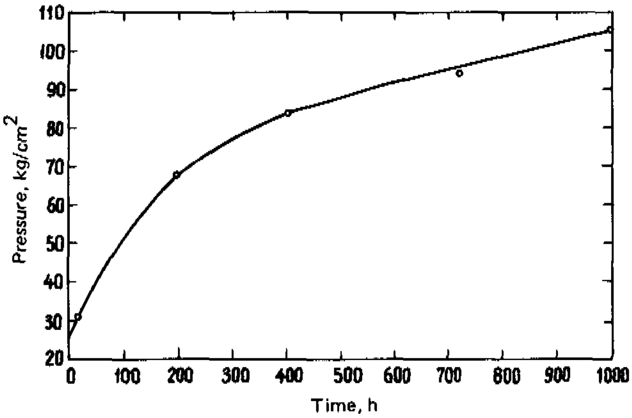

FIGURE 9 Dependence of the maximum pressure reached in related experiments as a function of freezing using identical freezing conditions.

The curve of freezing point depressions (see Figure 10) can be taken as experimental confirmation of Poynting's views, based on his calculations of the vapor pressure for water and ice. Technically (in the special sense) it is possible to see that for the formation of an ice lens at a definite pressure it is necessary to have a definite specific temperature. In other words, the melting point of the ice in equilibrium contact with the free liquid is a linear function of the pressure.

It can easily be seen that this linear curve is the locus of the equilibrium points of the free water and the solid ice under various pressures and that an experimental curve is appropriate thermodynamically to calculate the experimental curve for the single-component system at its melting point. In spite of the fact that we are using solid ice and water (liquid) to obtain these conclusions, the principles used here are applicable to any comparable single-component system and the relations of its pressure and temperature under parallel conditions. This occurs in the case where the solid state is under pressure and the liquid is not.

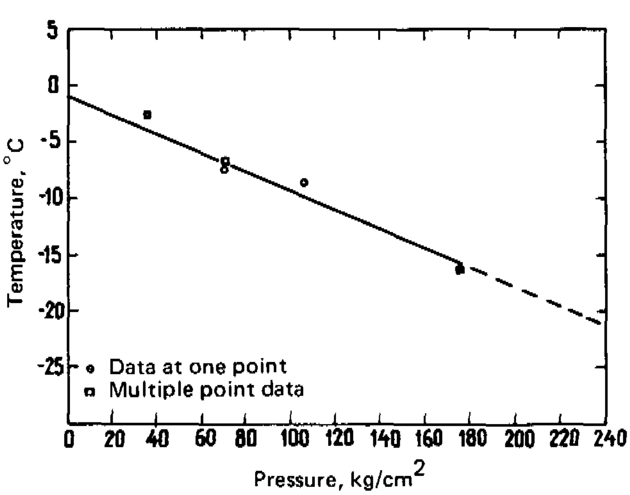

FIGURE 10 Summary (generalizing) curve of the freezing point depressions as a function of pressure on formation of the ice lens.

Beginning with an analysis of the thermodynamic free energy it is possible to draw the conclusion that

$$V_I dP - S_I dT = -S_W dT,$$

$$\frac{dP}{dT} = \frac{S_I - S_W}{V_I} = -\frac{\Delta S_f}{V_I},$$

where f is fusion.

$$\frac{dT}{dP} = \frac{-T(V_I)}{\Delta H_f} \quad \text{or} \quad \frac{-T\ V_S}{\Delta H_f}$$

(more generally, where V is the volume of the solid state). For ice at 0°C, the specific volume is 1.0907 cm^3/g

$$\frac{dT}{dP} = -\frac{(273.2)(1.0907)}{(79.8)(41.293)} = -\frac{297.98}{3295.2} \approx -\frac{1}{11}\ °K/atm,$$

where 79.8 is the heat of fusion, cal/g; 41.293 is the conversion factor, cal to cm^3/atm, $dT/dP = -V_s/S_f = K$ is the curve denoting the solid state under pressure, and the liquid not under pressure has many non-soil applications. One of them is used for describing the large glacier mass, the foot of which melts at various temperatures in accordance with the applied pressures. Another is used inside the glacier where melting can fuse particles together. A third is used for the compaction of powdered metals, ceramics, and salts not quite reaching their melting points. It is interesting that this curve says nothing about the nature of the insoluble solid particles in contact with the ice or of about the method of applying pressure to the solid ice, although the liquid has a capability of coming and going.

The correspondence between theory and experiment is as follows: The theoretical value is 0.0904° K/atm: the experimental value is 0.0938° K/atm. Possibly this method of describing the free liquid and solid state under pressure could be useful for experimental measurement of the entropy of fusion for various substances. Of the greatest importance is that the correspondence also gives double confirmation of the theoretical proof: first, that this method of freezing the chamber (from bottom to top) is justified, since the recorded error in the force will not depend linearly on the pressure; second, that the pressure-time data indicate the absence of any perceptible specific effect of the solid state comprising fine-grained particles--the results are represented only by the thermodynamic properties of the water-ice system.

In the experiment at a pressure of 105.465 kg/cm^2 (see Figures 3 and 4) when using slow speeds, a quite surprising phenomenon was observed: The clay particles under the ice lens moved vertically through it possibly with cavities or gas bubbles if it is assumed that the flow is moving upward along a moving cavity. Since diffusion of the dissolved material is not caused, this phenomenon is different from the well-known case of the Kirkendall metallurgical diffusion in the solid phase. In our experiments there were always pressure-temperature gradients which, in our view, were the cause of the proposed dislocation-vacancy flow. For several days the solid particles were diffused upward; the new ice lens was then clear and no other flows in the solid phase (solid state) were observed. The presence of stresses in the ice lens was detected by X-ray diffraction studies.

The photographs of the inverse representation of the laue X-rays were obtained with liquid nitrogen cooling; the ice samples were shielded by a Milar film located at a distance of 3 cm from it. The splitting of the laue spots is indicative of crystal breakup under pressure, whereas their smearing is the result of the retained residual stresses. The grain sizes of the ice lenses, as the studies of polarized light demonstrated, were quite large so that the absence of clear laue spots indicates strong deformation of the ice lattices. Such lattice deformation results were not surprising, because that by growing the ice lenses under pressure, in particular a varying pressure, one could not expect the formation of the solid state free from deformation.

In all of the experiments involving the application of high pressure attention was directed to another phenomenon--noted by photography--which is that these experiments clearly proved the presence of water below the freezing front. This migrating water may contribute to the formation of an ice lens and move (be displaced). While the reserve of this boundary layer is limited, probably in the small cavities between the particles the question of its presence does not arise. This water, which lowers the freezing point of the boundary layer, may be one of the causes of the limitation in growth of the ice lenses at increased pressures, for example, at 175.775-182.81 kg/cm^2. Although the effects of conservation of the water and of diffusion of the clay particles were recorded on film, when showing the slides the phenomena were not noticeable. The best data were obtained by slow-speed photography. (They were seen best of all when showing the frames at specific time intervals.)

CONCLUSION

Based on the results of several experimental studies of the pressure-time and growth-melting effects on the formation of the ice lens, the following conclusions can be drawn:

1. The maximum pressures obtained when using free water-ice under pressure were limited solely by the capacity of the unit to create low temperatures at precisely controlled rates. This is correct at least to a pressure of 210.93 kg/cm^2 and possibly to higher values.
2. The curve of freezing point depression as a function of temperature corresponds precisely to the thermodynamic propositions. These data are more than an order higher than the data predicted by the Clapeyron equation (~12 times) = (dT/dP). The calculation could be used for many equivalent states and single-component systems. The experimental method of deriving this curve also shows for this individual special system the decisive role in the formation of an ice lens that is played by the freezing point depression of the boundary layer.
3. The formation of an ice lens is a special method of growing a crystal that is not limited by transition of water to the solid-liquid state or to sand particles in fine-grained soil. Evidently this is a law of nature and the result of the ice growth by means of regulating the deposition in fine-grained systems and by the effect of forming an ice front is greater than by regulating the deposition of the solid particles in coarse-grained soil systems.
4. The diffusion of the solid particles in the presence of a steeper temperature gradient can be encountered in ice that was recrystallized under the effect of changing stresses and temperatures. This flow of solid particles moving in the cavities at steeper temperature gradients should be completely repeated in the other solid systems, and it is an absolutely feasible method of purification.

MOISTURE TRANSFER AND ICE FORMATION IN FROZEN AND FREEZING GROUND

V. A. KUDRYAVTSEV AND E. D. YERSHOV   Moscow State University

Experimental and theoretical research conducted at the Department of Permafrost Studies of Moscow State University under the direction of Professor V. A. Kidryavtsev have demonstrated that the migration of moisture in freezing and frozen ground is a necesary, but insufficient, condition for lens ice formation. The sufficient conditions are determined by the magnitude and character of development of the shrinkage and swelling volumetric-gradient stresses and deformations in the ground. With this approach it appears possible definitely to indicate the site of formation of an interlayer of ice in soil and also to discover the conditions of nucleation, further development, and cessation of growth of an ice lens. Experimental studies of the moisture-transfer and ice-formation processes in frozen and freezing samples of the ground serve as confirmation of this.

The study of the moisture-transfer process in frozen ground by means of a temperature gradient (grad t = 0.5-3.0 deg/cm) was accomplished by generating a unidimensional stationary thermal flux in it. The experiments were performed in the -0.7°C to -9°C temperature range at the ends of a frozen sample. The intricate structure of the total moisture flux in frozen fine-grained soil of massive cryogenous texture, including diffusion and thermal diffusion flows of non-freezing water and vapor diffusion, has been established. A quantitative estimate has been made of these flows and of the moisture-transfer coefficients in frozen ground or different grain size distribution and mineralogical composition, indicating a growth in the flow of unfrozen moisture with a rise in the negative temperature and a steepening increase in the values of the temperature gradients.

Moisture transfer in ice-saturated, fine-grained soil is accomplished only in liquid form as a result of the unfrozen water gradient (grad w_u). A decrease in the w_u value in the higher temperature zone disturbs the thermodynamic equilibrium between the ice and the water, leading to melting of the ice and replenishment of the reserves of unfrozen moisture. With a decrease in the degree of ice saturation of the soil ($g < 1$), the total moisture-flux density in it increases substantially, which, however, cannot be explained solely by an increase in the proportion of the vapor transfer. It is evidently connected with the process of free ice formation taking place in the ice-saturated samples, and when the soil pores become saturated with moisture the migration flux proves to be proportional in the first approximation to grad w_u. With complete filling of the pores of the soil with moisture ($g = 1$) the value of grad w_n, or more precisely, the potential gradient of the unfrozen water, determines not only the migration flux density but also insures that, as a result of the disrupting pressure of the thin film (P_{dis}), structural adhesion of the frozen ground (P_{adh}) is overcome, so as to provide for the necessary additional space occupied by the migrating moisture. Here, one should take into account that the shrinkage deformation (h_{sh}) and stresses (P_{sh}) developing in the dehydrated zone of the frozen ground probably contribute in large measure to the origination of weakened zones and the development of the process of excess ($g = 1$) ice accumulation in them. For 10-20 days the shrinkage of the dehydrated frozen ground leads to the origination of a network of macrocracks even in the presence of a sufficient amount of pore ice. This is due to the inertness of the phase transformations of ice to unfrozen water in the dehydrated zone of frozen ground, which causes dehydration and aggregation of the clayey particles.

The fact that in both the ice-saturated and ice-unsaturated clay samples a narrow zone of intense (excess) ice generation forms in the -1.5°C to -3.5°C temperature range and subsequently (with 10-20 days) a continuous interlayer of segregated ice of up to 0.5-1.0 mm thick (h_L), which is perpendicular to the heat and moisture flux is of interest. The heave value (h_H) was usually fractions of a millimeter here, which enabled it to be expressed in the form

$$h_H = h_L - h_{sh}. \quad (1)$$

The sufficient condition for the lens ice generation and also the location (x) of the nucleating interlayer of ice is determined from the following expression:

$$P_{dis}(x) + P_{sh}(x) > P_{adh}(x) + P_0(x). \quad (2)$$

The value of P_0 is the external (usual) pressure at the ice generation boundary.

The ice-formation mechanism discovered in frozen ground is also wholly confirmed in the case of freezing ground of differing composition, structure, and properties. Here, experiments have simulated the unidirectional freezing process of a semi-infinite soil core with a constant negative temperature at its surface. The laboratory studies demonstrated that the formation of ice lenses occurs when shrinkage stresses and deformations develop in the dehydrated frozen ground. The shrinkage deformation of thawed ground at the freezing front can reach more than 2-3 cm, and the increase in the dry unit weight can reach 0.2-0.4 g/cm^3 and more. In accordance with expression (1), the heave value of ground, the freezing of which is of the layered cryotexture type, is in large measure determined by the total magnitude of shrinkage of the dehydrated zone and can be greater than, equal to, or even less than the value of h_{sh}. Hence it is understandable why, other conditions being equal, ground that has small shrinkage deformations (for example, compacted or rapidly freezing ground) exhibits greater heave values. The mutual positioning and thickness of the ice lenses are in this case substantially influenced, not only by the conditions of heat and mass exchange in the freezing ground and the magnitude of the shrinkage deformations and thickness of the dehydration zone (h_{deh}), but also by the nature of development of the shrinkage stresses (2), which is largely determined by the freezing rate. Thus, with a higher freezing rate of the ground, causing more intensive migration of moisture to the freezing zone, significant uncompensated shrinkage stresses develop, which give rise to a more dense generation of weakened zones of the soil with small shrinkage deformations. In turn, this predetermines the formation of the more closely spaced but thinner ice lenses, that is, the origination of a dense, thin-lensed layered structure.

At a certain critical value of grad w_{max} the less intensive development of the shrinkage deformations with a higher ground dehydration (freezing) rate can lead to the occurrence of maximum permissible uncompensated shrinkage stresses causing cracks in the dehydrated zone. In the final analysis this causes the formation of a reticular cryogenous structure during freezing. For qualitative estimation of crack formation in dehydrated soil, it is possible to use the expression for determining the limiting tangential stress of the crack formation (P_{crack}) or the limiting value of the Kirpichev moisture-exchange criterion $K_i{}^*$, which is directly proportional to the value of P_{crack} ($K_i{}^* = AP_{crack}$);

$$P_{tp} = \frac{\beta E_z}{1 + \beta w_{init}} \text{ grad } w_{max} , \qquad (3)$$

or

$$K_i{}^* = \frac{z}{w_{init}} \text{ grad } w_{max} = \frac{J_w h_{deh}}{K_w w_{init}} , \qquad (4)$$

where E is the model of the shear deformation; β is the coefficient of linear shrinkage; w_{init} is the initial moisture content of the dehydrating zone of the ground; z is the distance between the shrinkage cracks; J_w is the density of the migration flux of the moisture to the ice generation front; K_w is the moisture-diffusion coefficient.

Expressions (3) and (4) permit us to discover a number of laws of the crack-formation process. For example, with an increase in the freezing rate insuring large values of J_w and grad w_{max} in the dehydrating zone, crack formation will take place with smaller values of z and h_{deh}, that is, a finer thin-lens reticular cryogenous structure must be formed here.

It is evident that a massive cryogenous structure can form either in the absence of moisture migration in the freezing ground or when it is present, but without satisfying conditons (2) or (3), which are sufficient for ice lens formation.

Thus, consideration of the necessary and sufficient conditions for cryogenous texture formation, that is, the close tying of the heat and mass exchange and the mechanical aspects of the freezing process, not only permit discovery of the laws of the formation of the cryogenous textures, but also enable us to work out a classification (genetic) scheme for the formation of various types of cryogenous textures depending on the composition, structure, and structural-mechanical properties of a fine-grained soil and its freezing regime.

SOME REMARKS ON THE STATE OF RESEARCH IN FRANCE WITH RESPECT TO VARIATIONS IN SOIL STRUCTURE DURING FREEZING

J. AGUIRRE-PUENTE National Scientific Research Center, France

The freezing of a dispersed medium is investigated in connection with the effect of frost on roads. The study has been performed in three areas:

1. Theoretical developments and experimental studies on small models at the Aerothermal Laboratory of the National Scientific Research Center.
2. Measurements directly in the field in the course of highway utilization.
3. Simulation of winter conditions and a study of the state of the road surfacing and the structures making up the soil at the Laboratory of Geomorphology of the National Scientific Research Center.

All of this research was agreed on, organized and begun subsequent to 1963. The articles on these studies were published only in the last 4 yr, but, unfortunately, for the most part only in French.

The results obtained make it possible for engineers to take measures against undesirable effects of frost on the soil in roads and to formulate criteria that must be satisfied when building new roads.

The results of the theoretical research and the experimental work on small models are of the greatest interest for discussion in this session.

We have tried to explain the mechanics of the ground-freezing process and, after generalizing our experience, arrived at the following conclusions.

For understanding the mechanics of the ground-freezing process, it is necessary to consider three elementary phenomena and their combined interaction: namely, heat transfer, mass transfer, and interfacies phenomena near the freezing point.

The heat and mass transfer can be studied by means of the Fourier and Darcy equations. However, investigation of the interaction between these transfers and the interfacies phenomena is still not possible. Indeed, the suction potential excited near the water-ice interface is not yet fully understood and cannot be conclusively explained.

The study of the interfacies phenomena must take into consideration three phases: in saturated ground, in a solid substrate, and in ice. It is convenient to begin the study on models, which facilitate tracing of the freezing phenomenon on the pore scale.

When the model is established, it will be necessary to study the adsorption of water on the walls. Then it will become possible to study the interaction of these three elementary phenomena.

This preliminary study is the topic of one of the papers at the four meetings. We have the opportunity to obtain quantitative data that will enable us to understand the mechanics of ice segregation and the heaving process. An experimental test was performed later using an artificial membrane and the freezing of aqueous suspensions containing small-diameter particles.

STATIONARITY OF THE TEMPERATURE FIELD OF THE PERMAFROST REGION AND DETERMINATION OF ITS THICKNESS BY GEOTHERMAL DATA

A. A. SHARBATYAN Institute of Water Problems, USSR Academy of Sciences

In a survey of North American literature on the temperature conditions of permafrost, Dr. A. H. Lachenbruch cited the well-known unidimensional stationary problem of the Fourier theory of thermal conductivity as one of the most important problems in the geothermy of polar countries. He noted that in spite of the exceptional simplicity of the solution to this problem, containing as it does only arithmetical operations, it is in the interpretation of its results, which in practice forms the basis of any discussion of the temperature regime of the Earth's crust, that misunderstandings most frequently occur.

From this solution it is easy to determine the depth of crustal freezing by the known values of the mean perennial surface temperature, the heat flux from the Earth's interior, and the thermal conductivity of rocks in the frozen state. It is clear that this determination holds true only when there is stationarity of the geothermal field, that is, when the field is independent of time. In the case of the steady-state linear heat flux, its value also remains constant with respect to depth. Dr. Lachenbruch cited some very interesting geotemperature curves plotted from results of measurements at four points on the coast of northern Alaska.

The upper portions of these curves reveal clear traces of sub-Arctic warming dating from the beginning of this century and, therefore, for reconstruction of the mean perennial temperatures of the region being investigated, the author extrapolates the straight segments of the profiles to their intersection with the Earth's surface.

In directing attention to the poor correlation between the mean perennial temperature and the depth of freezing, calculated by the rectilinear sections of the temperature profiles, the observed divergences are attributed by the Author to the variability of the iciness of the permafrost along the section, which has a strong effect on its thermal conductivity and, consequently, the position of the zero isotherm in relation to the Earth's surface. The geothermal data are considered by the Author to indicate practical constancy of the heat flux from the interior at all of the measurement points with a possible error of 10 percent in their results. However, there is no information in the paper on the direct comparison of the heat flux with respect to both sides of the freezing boundary.

The results of the mathematical simulation of the evolution of the permafrost region, which began in the Department of Permafrost Studies at Moscow State University in 1958 under the direction of Professor V. A. Kudryavtsev and continuing today at the Institute of Water Problems of the USSR Academy of Sciences, permit of a different explanation of this problem. Above all, the rectilinearity of the temperature profiles in the frozen and thawed parts of the massif does not mean that they are stationary. As a result of the high energy consumption associated with the processes of phase transformations, the temperature distribution on both sides of the freezing boundary is indistinguishable from a stationary distribution in many essentially nonstationary processes. These processes are widely known in thermophysics and hydromechanics and have come to be called quasistationary, as numerous examples of simulating the most diverse trends in the secular development of permafrost have shown that the quasistationary type of variation of temperature profiles is the most characteristic feature of the evolution of the permafrost region. The remarkable deviations of the temperature distribution from linear are characteristic of only very short segments in the history of formation of permafrost.

If we exclude the effect of the relatively short-term temperature fluctuations affecting the upper layer of rocks to a depth of 100-200 m, then the clearly defined nonstationary temperature distribution can be associated with episodes in the thermal history in which the thermal and water regime of the Earth's crust varies appreciably and unidirectionally. The periods of "adjustment" of the temperature field usually last several hundreds of years, and the development of the permafrost itself takes place over many thousands of years. Considering the relatively large error in measuring the geothermal gradients, it can be assumed that over more than 95 percent of the period of formation of the permafrost region the temperature distribution in the frozen and thawed rocks cannot be distinguished from a quasi-stationary one. This quasi-stationary mechanism of development of the permafrost region forms the basis of deriving approximate formulas for estimating its aggregation and degradation periods in the most typical limiting cases, which is the topic of my report published in the first issue of the papers.

It is important to emphasize that the strict equality of the heat fluxes at the freezing boundary is the exception rather than the rule. In the case of degradation of frozen relicts this difference is especially great, for after separating the permafrost region from the active layer the temperature distribution within it quite quickly becomes gradientless, and the heat flux density becomes zero. Below the freezing isotherms the heat flux density at all times differs appreciably from zero, although as the lower phase interface rises the temperature gradient in the thawed rocks decreases somewhat: the more this is so the less icy are the thawing layers. In the intermediate situations, which we most frequently encounter during geothermal studies of the permafrost region, the differences in the heat fluxes on both sides of its lower boundary are not so pronounced. Nevertheless, it sometimes can be discovered without any ambiguity in its meaning.

Of the four summary geothermograms cited by Dr. Lachenbruch, only two extend to the region of thawed rocks, both of which, in my opinion, clearly illustrates the indicated lack of agreement between the heat flux densities at the freezing boundary and indicate its displacement upward in the Prudhoe Bay area and downward in the Point Barrow and Cape Simpson areas. The cause of this lack of uniformity in the development of the permafrost region of northern Alaska is not so much the secular variation of the climate as the displacement of the shoreline, which freed the last two areas of land from seawater much later than the first area. The intensive reworking of the icy coasts of the Arctic seas apparently well masks the traces of their smooth displacement and the view that the shoreline in this region is stable may prove to be just as illusory as the "stationarity" of the freezing boundaries considered above.

SOME ASPECTS OF USING MODERN STRUCTURAL PHYSICAL METHODS OF INVESTIGATING MINERALS AND ROCKS IN CRYOLITHOLOGY

M. N. USKOV Institute of Permafrost Studies, Siberian Division of the USSR Academy of Sciences

In our opinion, without knowning the mineralogical composition of the permafrost and the physico-chemical factors that resulted in this composition, and without having discovered the crysto-chemical peculiarities of the minerals forming and coexisting in the permafrost region, it is impossible to determine with confidence and interpret correctly the properties of frozen and the thawed soils, to say nothing of trying to modify and control them in the direction required for economic activity.

However, in spite of the recent heightening of interest on the part of geocryologists in cryolithology and in investigating the processes of cryolithogensis, in the cryolithological literature and in practice there are still no data on the authigenic minerals (that is, the minerals formed in this sediment and the conditions corresponding to it), which are almost the only witnesses to the physico-chemical parameters and the course of their variation during the different stages of lithogenesis to which the sediment in question has been subjected.

One of the reasons for this, as was noted in the general paper by P. I. Mel'nikov discussing the prospects of research in geocryology, is that insufficient use is still being made of modern, exact physical methods of structural analysis. These include: *diffraction* studies, involving the use of X rays (for the most part) and the electron-microscope with electrophotography, *spectroscopic* studies, namely infrared, nuclear magnetic (from resonance spectroscopy); and also such specific *mineralogical* types of analysis as the differential thermal technique (especially in the low temperature range), which can be regarded as a model of the phase transformations of various types of bound water during freezing and melting.

The "visual method of investigation" continues to predominate in geocryology, and, as was noted in the paper by A. I. Popov, one should not and must not talk about the diagenesis of the sediment that it underwent before freezing and especially the cryodiagenesis--the diagenesis of the sediment during the freezing period.

First and foremost, it is necessary to ascertain the presence of diagenesis itself through the appearance of certain authigenic mineral indicators, to outline its physical and chemical parameters from the crysto-chemical characteristics of these minerals and their structures, and on this basis to work out the argument to be presented in favor of particular geological or geocryological synthesis.

Listed below, for example, are the conclusions that can be based on a study of the Pleistocene permafrost of central Yakutia (the upper reaches of the Taaty River) involving the following techniques: X ray (to discover the authigenic mineralization in thin sections measuring less than 1 micron); grain-size distribution with subsequent probability-statistical processing of the results by means of dispersion, variation, and correlation analyses (to estimate the dynamics of the sedimentation medium, in other words, the dynamic activity of the transporting medium and the sediment accumulation); and also a study of certain geochemical indicators such as C_{org}:

1. A characteristic of the diagenetic reworking of Pleistocene lacustrine, swampy, and alluvial sediments before their conversion to the permafrost state is the widespread prevalence of the montmorillonitization process. This process is two-stage--at first the hydromicas are invaded and later the chlorite and even the kaolinite.

2. One of the peculiarities of the cryogenous effect on the soil is the consolidation of the particles--aggregation of the pelitic component of the deposits to aleuritic.

3. Based on the closeness (in terms of percentage content) and similarity (in terms of specific character) of the mineralogical composition of the soils and also on the dynamics of their sedimentation and subsequent evolution, it appears incorrect to us to isolate two lithogenetic complexes in this and similar permafrost series in central Yakutia (above the alluvial deposits): subaqueous (without ice wedges) and subaerial or sub-ice (with ice wedges), as was done in the paper by M. S. Ivanov and Ye. M. Katasonov. According to the results of our study of these rocks, they belong to the same lithogenetic type and the facies conditions at the time of their formation were closely similar.*

*The aqueous silts with the shells of mollusks and subaerial deposits with the remains of the surface vegetation *in situ* cannot but differ in the dynamics of the accumulation. All of the researchers class them with various lithogenetic types of sediments (Editor's note).

FORECASTING VARIATIONS IN GEOCRYOLOGICAL-ENGINEERING CONDITIONS WHEN BUILDING ON PERMAFROST

R. M. SARKISYAN Production and Scientific Research Institute for Engineering Surveying and Construction

Prior to embarking upon large-scale building developments in an area containing permafrost, a geocryological-engineering forecast is worked out with the object of ascertaining the characteristics of the condition and the phsyical, thermal, mechanical, and rheological properties of the frozen ground and also the composition, rate of development, and scale of manifestation of the cryogenous processes and phenomena resulting from the combined effect of the structure and the engineering measures.

In the foregoing statement the geocryological-engineering forecast comprises two problems, which are solved consequently. The first reveals the variations in the geological-engineering conditions caused by the effect of the structure itself; the second, the composition and types of engineering measures restricting and optimizing the direction and rate of development and also the scale on which undesirable changes in the geocryological-engineering conditions are manifested.

The first approach solves the problem of the geocryological-engineering forecast; the second, the problem of controlling and optimizing the geocryological-engineering conditions so as to insure structural stability and environmental conservation.

The procedure for controlling the variations in the geocryological-engineering conditions and their optimization must be presumed to be in the solution of the first problem. The possibility of this derives from incorporating in these problems various types of "resistances," simulating the role of the engineering measures, for example, thermal insulation at the contact between the heat source and the permafrost, radiant energy reflectors, heat emitters (a cryohydrate mixture, the vegetation with increased heat of transpiration, shading facilities, etc.) located on the surface or within the frozen ground, fields of force (electric, electromagnetic, etc.). The effect of these is to change the direction and rate of development of the cryogenic processes, as well as the characteristics of the properties of the frozen ground being controlled, and also to influence the course of the cryogenous processes (improvement measures), etc.

Because of the special character of the geocryological-engineering forecast, it utilizes, on the one hand, a geocryological approach ensuring a comprehensive analysis, a largely high-quality assessment of the construction conditions, and consideration of all of the basic factors in the formation and development of the frozen ground; and, on the other hand, a technical approach permitting numerical quantification of the change in the principal characteristics of the condition and properties of the frozen ground constituting the foundations.

For the geocryological-engineering forecast, it is usual to consider and solve the problems pertaining specifically to an area and a structure with known technological operating regimes, a definite engineering measure and also a definite mode and time of performing the construction operations.

The geocryological-engineering forecast is based on the results of the geocryological survey and zonation of the area, including the free-hand sketching of the positions of the construction sites on the terrain, together with the data on their technological peculiarities, and on the proposed plan for executing the construction work, having regard to the existing fleet of construction machinery, and so on.

The information about the natural setting, as derived from engineering surveys, must include a description of the region's climate and its geomorphological and hydrogeological structure; the peculiarities of the plant cover; and the laws governing the formation and development of the cryogenous processes and phenomena, including those that are of special importance for predicting the laws governing their formation and development under geocryological conditions that have been altered by construction.

The positioning of the construction sites and the data on their technological characteristics are of importance for designating the boundary conditions, for technical problems solved subsequently, and for the forecast. The analysis of the construction operations performance plan is subject to the same goal.

The basic problems of the geocryological-engineering forecast amount to establishing, for example, the following characteristics:

Ω--the mean annual sum of degree-hours on the surface of the area after its development. This value can be approximately determined by the formula:

(a) for the summer

$$\Omega_S = 720\sum_i \left(t_B + \frac{-0.04LE - 10}{\alpha} \right) i \ \text{deg}\cdot\text{h}$$

where 720 h is the duration of the month; t_B°C is the mean monthly air temperature for the i-th month; $B = S(0.76\text{-}0.81a)$ kcal/m^2 · h is the mean intensity of the radiation balance for the i-th month with a mean diurnal intensity of the total radiation of S kcal/m^2 · h; a is the albedo of the underlying surface; L kcal is the heat vaporization of 1 kg of water; $LE = 0.02(1 + 0.22)KD$, kcal/m^2 • day is the mean heat of evaporation for the i-th month; 10 kcal/m^2-h is the mean monthly heat flux to the frozen ground during the summer; $\alpha = (4 + 1.9V_i)$ kcal/m^2 • h • °C is the mean heat-exchange coefficient for the i-th month; V_i, m/s is the mean wind speed;

$$K = \frac{w - w_P}{w_L - w_P}$$

is a coefficient taking into account the effect on the evaporation of the moisture content w of a clay soil with a plastic limit w_P and a liquid limit w_L; for a clay surface with $w > w_L$ it is assumed that $K = 1$; for a surface of sandy or coarse, clastic soil it is assumed that $K = 0$.

D_{mm} is the deficit of the moisture content of the air

(b) For the winter,

$$\Omega_w = 720\Sigma_i t_{bi} \text{ deg·hr.}$$

The mean annual permafrost temperature is also a calculated value. For the area being built upon it can be determined by:

$$\frac{\Omega_w + \frac{\lambda u}{\lambda f}\Omega_s + \rho w R_c H}{1 + (0.08\kappa H + 0.35\kappa^2 H^2)(\sqrt{1 + 2\sqrt{N} + 2N} - 1}, \frac{1}{T}$$

where

$$H = \frac{-3 + \sqrt{9 - \frac{\lambda_u \Omega_s A}{\lambda_f t_0}}}{1.9A\sqrt{\frac{C_u}{\lambda_u}}}, m$$

H is the depth in m of seasonal thawing in the area being developed;

$$N = \frac{\pi R^2 C\lambda_f C_f}{T},$$

a dimensionless factor taking into account the effect of snow cover on the permafrost temperature.

$$A = 0.31 - 11.2\frac{w_I}{C_f t_0},$$

a dimensionless factor taking into account the latent heat of fusion in the depth of seasonal thawing of the ground.

$$\kappa = \sqrt{\frac{\pi \cdot C_f}{\tau \cdot \lambda_f}}, \frac{1}{m}$$

the damping decrement of the temperature fluctuations in the permafrost.

$\tau = 8.760$h, the duration of 1 yr.

Ω_c deg·h is the sum of deg·h on a snow surface during the winter; in the absence of a snow cover $\Omega_c = \Omega_w$;

$$w_I = (w - w_u)\gamma_s,$$

is the ice content of the frozen ground

$\rho = 80$ kcal/kg is the latent heat of fusion of ice.

The first approximation of the value of H may be taken as:

$$H' = \sqrt{\frac{2\lambda_u \Omega_s}{\rho w_I}}, m.$$

The forecast also necessitates estimating the magnitudes of ground heave during freezing, the settling of the ground during thawing, and the amount of soil confronted by artificial obstacles on the solifluction slopes. These and other characteristics are estimated by the available formulas, which also include the resistances.

The altered geocryological-engineering conditions obtained must satisfy the requirements for stability and operating reliability of the structure. Otherwise, the set of expedient measures that is considered advisable will be implemented for improvement and optimization of the geocryological-engineering conditions.

However perfect in theory may be the procedure followed in an engineering geocryological forecast, it cannot be recommended for use until it has been confirmed experimentally.

THE STRAIN-HARDENING OF FROZEN GROUND DURING THE CREEP PROCESS

N. K. PEKARSKAYA Scientific Research Institute of Foundations and Underground Structures

At the present time the problems of the rheology of frozen ground have been worked out in considerable detail. It has been demonstrated that in the course of sustained creep,* weakening of the bonds takes place, with the result that the strength of the ground decreases. In addition, simultaneously with the decrease in strength, an inverse phenomenon occurs, namely strain-hardening of the frozen ground. (Vyalov, 1959).† Obviously, depending on the deformation conditions one or another of these phenomena may predominate.

In order to discover the effect of the strain-hardening of frozen ground, special experimental studies were performed in which we began with the assumption that strengthening of the bonds, that is, strain-hardening should be exhibited to the greatest degree with deformation under the conditions of damped creep,†† that is with $\sigma < \sigma_\infty$.

The experiments took the following form. A series of identical samples of frozen soil (clay and fine sand) was subjected to deformation under conditions of damped creep, after which relaxation took place and subsequent determination of the "instantaneous" and the long-term resistances of the samples to uniaxial compression and shear were made. The results obtained were compared with the strength of a sample not subjected to preliminary deformation.

Illustrated in Figure 1 is the envelope of the largest stress circles of the frozen ground ($\theta = -2°$) obtained under rapid-loading. Curve 1 was obtained for the preliminarily deformed sample; curve 2--for identical samples that had not been subjected to preliminary deformation.

From the data adduced it is apparent that, as a result of deformation under the conditions of damping creep, the instantaneous strength increased by 30 percent. The long-term strength of the samples subject to preliminary deformation under the conditions of damping creep will also increase (by about 20-24 percent), which is evident from the long-term strength curves (Figure 2).

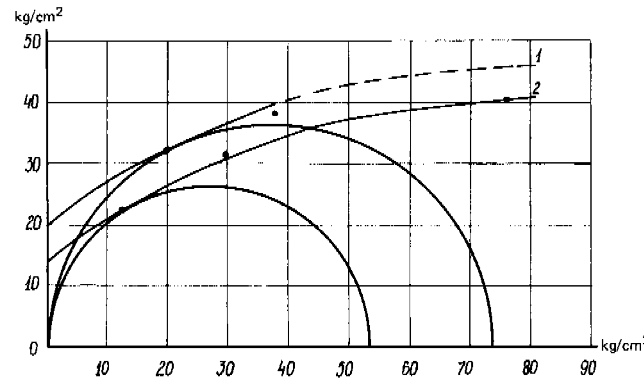

FIGURE 1 Envelopes of the largest stress circles of (1) strain-hardened and (2) unstrain-hardened frozen ground. Silty sand.

The dissimilar strength of the frozen ground obtained for different loading history corresponds to the various deformative properties of this ground. Thus, at one and the same stress values the rate of steady-state flow of the unstrain hardened samples was $V = 0.086$, and of strain hardened samples, 0.001 mm/min, that is, almost 2 orders less. The magnitude of the deformations at each given point in time was greater for the unstrain hardened samples, respectively.

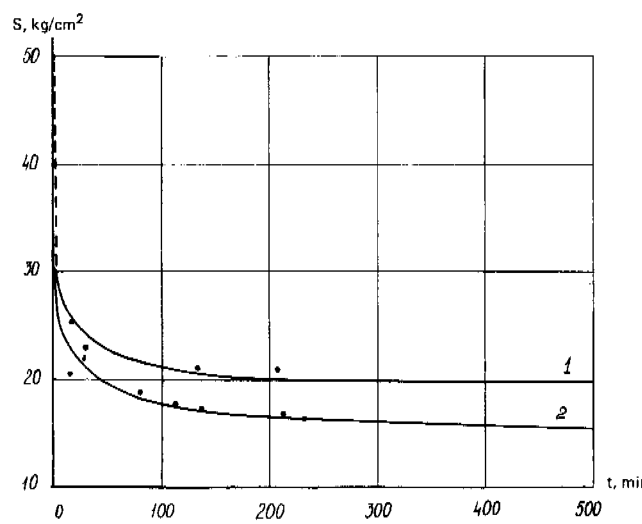

FIGURE 2 Curves of the long-term strength of (1) strain-hardened and (2) unstrain-hardened frozen ground.

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*Sustained creep is deformation of the ground with the stresses (σ) exceeding the limiting rupture strength of the given soil (σ_∞).

†S. S. Vyalov. Reologicheskiye svoystva i nesushchaya sposobnost' merzhlykh gruntov (Rheological properties and bearing capacity of frozen ground). Moscow, Izd-vo AN SSSR, 1959.

††Damped creep is deformation with the stresses (σ) less than the long-term ultimate strength ($\sigma < \sigma_\infty$).

All of this indicates that with deformation under damped creep, there is a change in the properties of the frozen ground, which becomes a stronger material than it was in the initial state (see Figures 1 and 2). In other words, the strength characteristics and deformability of the frozen ground depend on the history of the latter's deformation and loading.

Therefore, it does not matter which method of testing is used to determine the strength characteristics, including the most important of them--the ultimate long-term strength of the frozen ground. It is evident that the method of determination must begin by taking into consideration the loading regime, which corresponds to the conditions of subsequent utilization of the frozen ground.

The study of the laws governing the variation in the mechanical properties of frozen ground as a function of the conditions of its loading and deformation affords an opportunity of increasing the bearing capacity of the ground by directional control of these conditions, that is, of making fuller use of their strength characteristics.

PROTECTION OF PERMAFROST DURING CONSTRUCTION AND OPERATION OF PIPELINES

H. O. JAHNS, T. W. MILLER, L. D. POWER, V. P. RICKI, T. P. TAYLOR, AND J. A. WHEELER   Esso Company, United States

I should like to tell you about some of the field tests of thermal-piles for laying pipelines aboveground. These calculations will be useful for the construction of such pipelines for both oil and gas in the discontinuous permafrost zone in order to combat premature destruction of the permafrost. This could occur as a result of a disturbance of the ground surface or prolonged climatic variations.

Our experiments were performed in order to justify the planning of an oil line through Alaska that will cut across fairly large areas of discontinuous permafrost where the permafrost temperature is about 0°C. In the report that you will find on page 673 of the North American Contribution and on page 227 of the volume containing abstracts of reports, theoretical propositions and also laboratory data on the use of thermal piles as supports for pipelines are presented.

Experiments involving the use of thermal piles of two different designs have been performed: an open system using air convection and a closed system using warm pipes, which in principle are similar to the Long [Thermo] pile but are much smaller in diameter.

The first slide shows a schematic representation of the air atmospheric pile. This device, which is used for laboratory testing, has the following structure: The open tube is buried in the pile so that the cold winter air can enter and circulate in it and in this way heat will be removed from the surrounding soil. In summer this column of air is thermodynamically stable, but at this time of year the pile is kept closed in order to prevent forced circulation of the air in it, under the influence of the atmospheric pressure and the wind. The next slide shows a field installation of such a pile near Fairbanks. In the background we see the piles of the second design, using convection. These piles measuring 46 cm in diameter are sunk to a depth of 5 m in silty soil, which was initially thawed to a depth of 4 m and then frozen. The permafrost is warm and highly unstable. Last winter we measured the temperature around these piles.

The next slide shows the temperature at a depth of 5.5 m at the ground-pile interface and a distance of 2-4 ft or 60-120 cm from the pile. The time scale is in months and days. In the upper part the air temperature is indicated. Note that after a month of operation of this pile, at a distance of 2 ft (60 cm) from it the temperature had dropped below 0°C. At the end of winter the freezing front had almost reached 120 cm. We may assume that this freezing zone remains all summer and increases in the subsequent years. The next slide shows the experimental thermal piles with radiators in the upper part and a thin tube 2.5 cm in diameter filled with a cooling agent, e.g., ammonia, and also the upper part of the cylinder of frozen mud measuring about 50 cm in diameter, forming around the thermal pile during the laboratory tests, which lasted about a week.

In the next slide we see the experimental pile with two heating tubes attached to the sides. Other types of radiators were tested that are perhaps more suitable in the building of pipelines. This slide shows a long vertical radiator. The last slide shows the measured temperatures near the pile in ice-rich soil. In this pile two tubes about 8 m long are inserted. The data were taken at a depth of 4.5 m where the ground was initially frozen, but at a temperature close to 0°C; significant cooling was observed at a distance of 120 cm from the pile. This device operates more efficiently than the atmospheric pile shown earlier.

The field studies basically confirm the suggestion that thermal piles of this type can be used in the construction of a pipeline aboveground and under any conditions.

LIQUID-VAPOR HEAT TRANSFER DEVICES OF THE "LONG" THERMO-PILE TYPE FOR COOLING AND FREEZING THE GROUND WHEN BUILDING IN REGIONS WITH HARSH CLIMATES

N. A. BUCHKO, V. V. ONOSOVSKIY, AND V. S. SOKOLOV
Leningrad Technological Institute of the Refrigeration Industry

For the artificial cooling of the ground forming the foundations when building in permafrost regions, it is advisable to use natural winter cold. This kind of cooling can be achieved through the use of outside air with natural or forced ventilation or by seasonally active heat-transfer devices involving natural circulation of the heat-transfer agent. The second method is characterized by the high degree of efficiency of the heat-exchange processes, the absence of energy expenditures, automatic performance, and operational reliability.

Two types of heat-transfer units with natural circulation of the working medium are known: liquid and vapor-liquid units. In the best liquid units (the double-tube Gapeyev kerosene unit, the coaxial Makarov liquid units, and the liquid units built by the Thermo-Dynamics Company) the heat is transferred from the ground to the air in the course of natural convection of the liquid. In the vapor-liquid units (the direct freezing columns of Blier and Zinman, Badyl'kes, and the thermal piles of Long) the heat is transferred on completion of the evaporation-condensation cycle by the low-boiling-point liquid.

These heat-transfer units are beginning to be used for various heat engineering purposes; in heat engineering they are called single-phase and double-phase heat siphons, respectively, and the latter are also known as heat pipes.

When using these devices in construction, although a definite competition is possible between the two versions referred to, no one has any reason to doubt that as a result of the phase transformation the internal hydrodynamic and thermal processes in the devices filled with low-boiling-point liquid are much more efficient than in the devices filled with trickling liquid.

At the Leningrad Technological Institute of the Refrigeration Industry comprehensive studies were made for a number of years of the methods of making practical use of devices filled with low-boiling-point liquid, which will hereinafter be referred to as thermopiles. Two interrelated problems have been solved: an internal problem in which the processes of heat exchange and hydrodynamics occurring in the thermopiles were studied, and an external problem in which studies were made of the processes involved in the shaping of the temperature field in ice-soil massifs around individual thermopiles and thermopile systems.

The testing of the efficiency of the thermopiles and the working out of the designs of the individual units were carried out by the Institute jointly with other organizations in the course of field testing in a number of northern regions.

The internal processes were investigated under laboratory conditions on models under stationary conditions with observance of geometric, thermal, and hydrodynamic similarity. The experiments performed made it possible to establish the relations describing the processes of evaporation and condensation of ammonia, propane, Freon-12, and Freon-22 under the specific conditions of the thermopile. The heat-flux density at the wall of the thermopile was used as the characteristic of the intensity of the heat-exchange processes. The optimal amount of working medium charged in the thermopile and the maximum permissible angles of inclination of the evaporator tubes ensuring cooling efficiency with depth were also established.

Besides the above-enumerated factors, the high efficiency of the thermopile is also determined by the virtual isothermality of its surface in contact with the ground. The constancy of the temperature is due to the constancy of the pressure with the height of the pile, which follows from theoretical calculations of the hydrodynamic resistance to the flow of vapor and is confirmed by laboratory and field experiments. This fact makes it possible to simplify the external problem with respect to determining the temperature fields around thermopiles of greater length (the effect of the temperature variation in the upper layer of the ground can be neglected).

A method of conjugate solution of the problems is proposed: the heat transmitted from the ground to the thermopile, the heat transfer inside the thermopile, and the heat transmitted to the outside air. Here, the equation of thermal conductivity is approximated by an explicit finite-difference scheme. The phase transformations of the free water are considered to be in the corresponding range of negative temperatures. The calculations were performed on a digital computer, the programs were mainly compiled in the Algol-60 algorithmic language and enable the temperature fields to be calculated for individual thermopiles (a unidimensional temperature field) and also for linear and cylindrical single-row systems of thermopiles (two-dimensional temperature fields). In the calculations it is possible to vary the following parameters: the diameter of the thermopiles, the spacing between them, the thermophysical properties of the ground and its initial temperature, the mean annual temperature and the amplitude of the seasonal variations of the air temperature, the heat-transfer coefficient to the air, the length of the evaporator, and the ratio of the condenser and evaporator surface areas.

As a result of the studies, a procedure was developed for the heat-engineering calculation of thermopile systems for cooling and freezing the ground. Recommendations were worked out with respect to the efficient design of thermopiles for a number of intended uses and sufficient experience was accumulated in performing the practical operations. Thus, there are grounds for believing that sufficient preconditions have been created for scientifically justified design and use of these cooling units in the building on permafrost of such major industrial installations as dams made of local materials, oil and gas pile lines, electric power transmission lines, radio relay communications towers, mine shafts, bridges, and other structures.

CRYSTALLIZATION MECHANISM OF UNFROZEN WATER IN VERY FINE-GRAINED FROZEN SOILS

A. A. ANANYAN   Moscow State University

It is known that the crystallization of water in a tank is a resultant of the speed of formation of the first ice crystals that appear after supercooling has begun.

In contrast to water in a tank, the crystallization of water in fine-grained soils takes place in conformity with the structural peculiarities of water in dispersed media, is dependent on the difference in the energy level of its component molecules, and is determined by the temperature conditions. A number of studies have demonstrated that under the influence of the activating nuclei on the surface of the soil particles, the water molecules contained in them are directionally oriented, the energy level of the molecules changes, and the structure of the water is distorted. Since the orientation effect gradually attenuates with distance from the activating nuclei, the degree of distortion of the structure decreases correspondingly.

Our nuclear magnetic resonance spin echo studies of the relaxation times of protons of water contained in soil have confirmed that the energy nonuniformity of the molecules is a function of the distance to the activating nuclei of the surface of the soil particles. It is shown that for one and the same type of soil the relaxation times of the protons of water depend on the moisture content. The greater the moisture content the longer the relaxation time and the higher the energy level of the molecules. Conversely, with a decrease in the moisture context and in the thickness of the water film on the surface of the mineral particles, the relaxation time of the protons is shortened and there is a decrease in the energy level of the molecules.[1] For example, in bentonite with a specific moisture content of 10.2 to 100 percent, the average number of molecular layers on the surface of the rock particles ranged from 0.7 to 6.5; the spin-spin relaxation time (T_2) was accordingly between 0.27 and 2.00 milliseconds. In kaolin, with a moisture content of 5.0 to 52.6 percent and the number of molecular layers of the water ranging from 6.6 to 66.5, T_2 was accordingly in the 1.7 to 6.8 milliseconds range.

As is well known, for a liquid (melt) to crystallize it is necessary for the arrangement of the molecules in the liquid (melt) to correspond to their arrangement in the crystal. The distortion of the liquid structure is a hindrance to crystallization. For crystallization of unfrozen water in fine-grained soils it is necessary for some force to restore the structure of the water and for the force to remove the water molecules from the influence of the activating nuclei on the surface of the mineral particles. In the case of electro-osmosis in the frozen soil, it has been shown[2,3] that the crystallization in the cathode zone of the previously unfrozen water is caused by the following: Under the influence of the applied electric field the molecules of the unfrozen water of the anode zone are removed from the sphere of the influence of the surface forces and migrate into the cathode space where the structure of the water is restored and the molecules are grouped into an ice structure. Let us consider the forces acting in the unfrozen water. As already noted,[4,5] the effect of the activating nuclei on the surface of the mineral particles on the translational motion of the nearest water molecules leads to a variation in the energy state of the molecules and to distortion of the water structure. These forces are designated long-range. In solutions contained in rocks, ions of dissolved salts may be present. The interaction between the ions and the water molecules is also classed with the long-range forces, the energy of which is estimated in tens of kcal/mole. On the other hand, these forces are counteracted by the short-range forces between the water molecules. These are the hydrogen bonds striving to arrange the nearest water molecules in a definite order and form tetrahedral structures. The short-range forces (about 4.5-5 kcal/mole are appreciably weaker than the long-range forces. Thus, at positive temperatures of the water contained in finely grained soils, the long-range forces distorting the structure of the water and hindering its crystallization when the temperature drops predominate.

Upon a variation in the temperature, as is well known, the effect of the directive forces is determined according to Keesom[8] by the expression

$$\phi_{\text{oriented}} = -\frac{2}{3}\,\frac{P_e^2}{r^6}\cdot\frac{1}{kT},$$

where P_e is the electrical moment of the interacting particles; k is the Boltzmann constant; r is the distance between the interacting particles (in this case between the activating nuclei on the surface of the mineral particles and the water molecules or between the water molecules).

The expression cited indicates that with a reduction in temperature the effect of the orienting forces increases. However, as has been shown,[7] with a reduction in temperature the effect of the short-range forces striving to orient the molecules into tetrahedral structures, relative to each other, that is restoring the structure of the water, increases more rapidly than the effect of the long-range forces distorting the structure of the water and hindering crystallization. As a result, the ratio between the effect of the long-range and short-range forces changes in favor of the latter. For a certain number of oriented molecules, when the temperature is below 0°C the forces of interaction between the water molecules predominate over the forces of interaction between the water molecules and the activating nuclei on the surface of the mineral particles; here the structure of the water is restored. At the temperature of the medium, the molecules of such water are grouped into the ice structure and the soil passes into the frozen state.

A further reduction in temperature leads to the restoration of the structure and transformation into ice of still newer quantities of unfrozen water. It is important to note that the lower the temperature, the more difficult it is to make the transition from water to ice, since at a lower temperature part of the unfrozen water with the more oriented molecules remains. In addition, with a reduction in temperature there is an appreciable increase in the viscosity of the water, which also serves as a hindrance to the possibility of regrouping of the molecules and restoration of the water structure. Therefore, with a drop in temperature, some portion of the molecules will be so oriented that the water molecules are no longer capable of being grouped into an ice structure, and the water remains in an amorphous and probably vitreous state.[6]

To sum up, the following conclusions can be drawn:

1. The orienting effect of the activating nuclei on the surface of the mineral particles on the translational motion of the water molecules in the soil distorts the structure of the water, which hinders crystallization and lowers the freezing point of the soil.
2. With a drop in the temperature of the frozen soil the crystallization of the unfrozen water is caused by a change in the ratio of the long-range and short-range forces in it. The more rapid increase in the effect of the short-range forces in comparison with that of the long-range forces when there is a rise in the temperature leads to restoration of the structure of the water and accordingly to its crystallization.
3. Some portion of the most oriented water molecules, being incapable of grouping themselves into the structure of ice at low temperatures, remains in an amorphous and probably vitreous state.

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MOVEMENT OF WATER IN FROZEN GROUND

R. D. MILLER Agronomy Department of Cornell University, United States

I should like to make a few remarks with respect to the movement of water in frozen ground. At the preceding meetings it was noted that this movement is determined by the film flow. Of course, in unfrozen films surrounding mineral particles there is a flow, but I am sure that the ice phase is also conducive to the transport of water in frozen ground. Serving as possible confirmation of this is the simple experiment that I performed in 1966 at the Geotechnical Institute in Norway.[1]

The device which I called the "ice sandwich" was a lens of ice placed between two sheets of filter paper with very small pores. On both sides of the "ice sandwich" there was water in the chambers. The device was submerged in a bath with a temperature somewhat lower than 0°C so that the liquid phase was somewhat supercooled.

When the pressure on the water in one chamber rose, the water disappeared, while in the other chamber the amount of it increased.

The water was displaced through the space filled by the ice, which can be represented as a large ground pore filled with ice. In my opinion, this is the result of melting of the ice in one direction and growth in the other. The ice may appear to be motionless because its boundaries are stationary, but in reality it is moving. The transport of continuous films of water through a space filled with ice does not occur. When a weak aqueous solution was poured into the input chamber, the following happened. If the solution pressure was less than its osmotic pressure (π), transport took place in the direction of the solution (simulation of osmosis). If the pressure of the solution was greater than the osmotic pressure, then the movement took place from the solution in the direction of the pure water (simulation of return osmosis). If the pressure was identical with the osmotic pressure, transfer did not take place.

I have derived some simple equations for calculating the heat and water transfer through an "ice sandwich," and we are now beginning the study of such transfers for the cases where the ice is replaced by frozen ground.

Those who are interested in studying electrical potentials formed during the freezing process can probably make successful use of the "ice sandwich."

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A STUDY OF THE RHEOLOGICAL PROPERTIES OF ICE USING A PRESSUREMETER

YU. K. ZARETSKIY AND A. M. FISH Scientific Research Institute of Foundations and Underground Structures

Experiments were performed using a D-76 pressuremeter designed by NIIOSP [Scientific Research Institute of Foundations and Underground Structures], which makes it possible to control the accuracy of measuring the radial deformations of a borehole both before and during the test process. The instrument comprises a pressure system, a measuring system, and a probe. The pressure is monitored by two manometers measuring low pressures of up to 1.6 kg/cm and high pressures of up to 40 kg/cm^2. The measuring system comprises two measuring cylinders with d = 40 and 12 mm. The probe is a steel cylinder surrounded by a rubber sheath. The outside diameter of the probe is 73.5 mm and h = 320 mm. The transfer of pressure is effected by compressed air. The measurement of the radial deformation is based on the fluid volume pumped into the probe of the pressuremeter, the flow rate of which is recorded by one of the measuring cylinders. The error in measuring the radial deformations was ±0.0008 mm.

The experiments were performed in September 1971 on the Bol'shoy Azau glacier (El'bruz) at an altitude of 3,700 m above sea level in ice at a depth of 2.0-2.5 m with respect to two load patterns: for creep--at constant pressures--and with stepped loading. The pressure step was 2 kg, and the duration--2, 10, and 30 min. The figure shows the family of creep curves of the ice, each of which was obtained from tests in individual holes at a specific constant pressure.

By means of the pressuremeter tests of ice, the parameters were determined, which enter into the rheological equation of state. The latter was based on the aging theory and it was demonstrated that this theory best corresponds to the experimental data for the loading regimes σ = const and $\sigma = V_\sigma t (V_\sigma = \text{const})$. According to Yu. K. Zaretskiy's hypothesis, the intensity of the shear deformations e_i depends on the intensity of the shear stresses σ_i and the time t in the following way:

$$e_1 \ (t) = \frac{B\sigma_i^n t^\lambda}{T - (\sigma_i - \sigma_{t(\infty)})^n t^\lambda \ \eta(\sigma_i - \sigma_{i(\infty)})}. \quad (1)$$

Here n, λ and B, T are the parameters; $\sigma_i(\infty)$ is the limiting shear stress which in accordance with the Mises-Botkin strength hypotehsis is: $\sigma_{i(\infty)} = c_\infty + \sigma \tan\phi$; $\eta(\sigma_i - \sigma_{i(\infty)})$ is a function equal to zero when $\sigma_i \leq \sigma_{i(\infty)}$ and to unity with $\sigma_i > \sigma_{i(\infty)}$. In the theoretical solution of the problem of expansion of a cylindrical hole in ice of radius $r = a$, it is assumed for the purpose of simplicity that the internal friction angle ϕ is zero; the dilational deformation is also close to zero, and in the region of plastic deformations forming around the hole $a \leq r \leq r_L$, the creep deformations take place so rapidly that their development with time can be ignored, assuming in accordance with (1)

$$T = (\sigma_i - c_\infty)^n t^\lambda \text{ or } \sigma \equiv c_s(t) = c_\infty + \frac{T^{1/n}}{t^{\lambda/n}}. \quad (2)$$

In the region $r > r_L$ in accordance with (1) we shall have:

$$l_i(t) = \frac{B}{T} \sigma_i^n t^\lambda. \quad (3)$$

Thus, from the pressuremeter experiments five parameters are subject to determination: n, λ, B, T, C_∞.

The procedure for processing the data from pressuremeter studies, as developed by A. M. Fish, is based on Zaretskiy's theoretical solution in which the angle of internal friction ϕ is taken to be zero. At pressures on the outline of the borehole p not resulting in the emergence of a plastic region, the relative displacement of the hole wall $\varepsilon_r = U/a$ is expressed by the formula

$$U/a = \frac{B}{2T} t^\lambda \left(\frac{p - \gamma H}{n}\right)^n, \quad (4)$$

from which it is evident that the experimental data in the load range of $p \leq H + nc_s$ must represent a family of parallel lines in logarithmic coordinates, the tangent of the slope angle of which is equal to the nonlinearity index n. From this first step in the processing the values of $n = 1.4$; $\lambda = 0.47$ and $B/T = 0.3 \times 10^{-5}$ (cm^2 min$^{-\lambda}$/kg^n) were obtained. The second step in the processing of the experimental data at pressures greater than $(\gamma H + nc_s)$ is based on the formula

$$U/a = \frac{B}{2T} t^\lambda c_s^n \exp\left(\frac{p - \gamma H}{c_s} - n\right) \quad (5)$$

and consists in "linearizing" the semilogarithmic coordinates. The family of straight lines obtained in this way is characterized by a variable slope angle, the cotangent of which is equal to the cohesion. The cohesion values in the coordinates << $c_s t^{\lambda/n} - t^{\lambda/n}$ are depicted in the form of a straight line that corresponds to formula (2). The values of the parameters of this line give $c_\infty = 2.6$ kg/cm^2 and $T = 32$(kg^n min$^\lambda$/cm^n).

The above-described procedure for pressuremeter testing was also extended to the case of step loading. The derived values of the rheological parameters and the ice strength closely agree with the values determined under laboratory conditions with a uniaxial hole. The existing general solution to the problem of the expansion of a cylindrical cavity in a medium having internal friction, cohesion, and a nonlinear compression diagram (1) is used to determine the rheological constants of dense clay soil (including frozen ground).

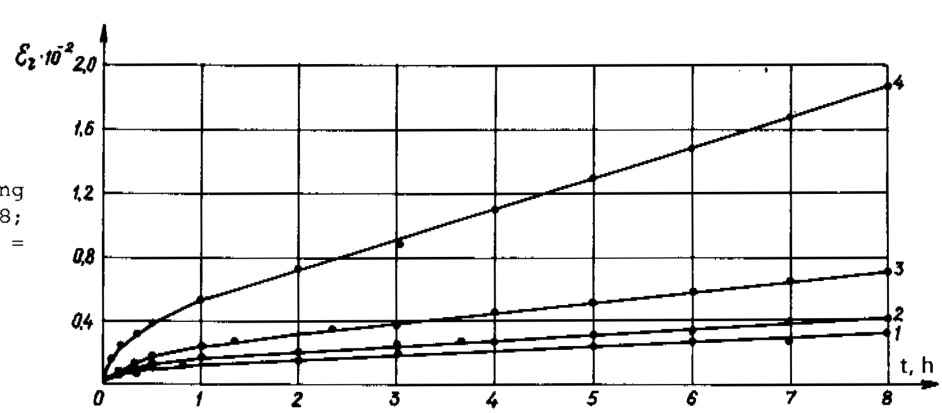

FIGURE 1 Relative deformation of the borehole at the following pressures (kg/cm^2): 1--p = 2.8; 2--p = 11.8; 3--p = 14.8; 4--p = 17.8.

STUDIES OF VARIOUS MANIFESTATIONS OF FREEZING PROCESSES IN TRANSPORTATION

V. P. TITOV, P. I. DYDYSHKO, N. A. TSUKANOV, AND M. V. AVEROCHKINA
All-Union Scientific Research Institute of Railroad Transportation

Many of the physical processes in an earth roadbed influencing its service life and stability are caused by seasonal and annual temperature fluctuations in the soil. In order to predict the development of these phenomena and to develop measures to eliminate or decrease their harmful effects on the operating conditions of the roads, it is necessary to predict the temperature regime in the earth roadbed under the various conditions that influence its thermal interaction with the environment. An effective means of performing these calculations is Professor V. S. Luk'yanov's method of the hydraulic analog, using a hydraulic integrator.

The most important prerequisites for obtaining reliable forecasts of the temperature regime of the ground are the use of design computation diagrams and design computation parameters, which most clearly reflect the active conditions of heat exchange. Accordingly, in our studies a great deal of attention has been devoted to the development of methods of calculating the temperature regime of an earth roadbed, which take into consideration the effects of solar radiation, surface water, and groundwater.

In order to calculate values of the thermophysical parameters of the ground that take into consideration the specific conditions of place and time, extensive use is made of hydraulic integrator techniques for solving inverse problems. These techniques are highly promising in terms of estimating the effect of various surface covers on the temperature regime of the ground.

One example of such calculations are studies of the use of foamed plastic to combat heave. By means of calculations on the integrator, the basic parameters of the paving required to ensure its effective operation during possible deviations of the cliamtic characteristics from the mean annual values are ascertained. It was found that the arrangement of the thermal insulation has an appreciable effect on the temperature regime of the ground to depths of 4-5 m from the surface, leading to a decrease in the temperature gradients in thermally insulated ground by a factor of 1.5 to 2.0. This must have an effect on the reduction in the amount of moisture migrating to the freezing front.

On the strength of *in situ* data, the hydraulic analog method was used to investigate the thermal field in seasonally freezing ground around the pipelines, and determinations were made of the characteristics of the zone of influence of this heat source on the condition of the ground massif, considered in relation to the size of the source, the depth of its positioning, the protection afforded by thermal insulation, the intensity of frost heave, and other factors. The ratio between the magnitude of the mean winter heat flux with an undisturbed soil temperature regime and the additional heat flux brought to the interface under the influence of a linear heat source was determined. Within this zone the frost heave factors depend strongly on the temperature gradients determining the quantitative indexes of the intrasoil moisture transfer (see Table 1). The results of the studies make it possible to designate the engineering measures needed to prevent uneven frost deformations of structures located in the zone of thermal influence of a pipeline.

In situ observations and theoretical studies of the temperature fields of the embankments of earthworks have made possible a more profound understanding of the processes occurring in the ground during freezing and thawing. Thus, the studies of Sigafous and Hopkins showed that during freezing and heaving of the ground a displacement of the surface of the massif takes place in a direction close to the perpendicular to the surface delimiting the frozen and thawed ground. During subsequent thawing the ground shows a tendency to settle under the influence of the force of gravity. As applied to the soil in the embankments of earthworks, this phenomenon leads to the formation of fissures. They occur near the intersection of the plane delimiting the embankment and the horizontal surface adjoining it. In these cases where an embankment forms when building the structure in natural bedding, the frost fissures occur, as a rule in the embankment and rarely on the horizontal area beyond the embankment. This is due to the difference in the intensity of ground heave in the active layer subjected to repeated freezing and in the ground located beneath the active layer and which had begun to freeze after the structure was built. The cracks are oriented along the embankment and are situated one above the other.

In fills, these frost fissures occur both on the embankment and on the horizontal area beyond the embankment.

It is also necessary to consider the following phenomenon, which is characteristic of embankments. As a result of structural variations in freezing and thawing ground, its strength proves to be markedly reduced, especially during the first hours and the days after melting of the ice inclusions. It is known that the ice inclusions formed during freezing and heave of the ground are not limited by plane surfaces. This fact leads to the origination in the thawing ground of large pores that are easily filled with water. The dimensions of these pores and

TABLE 1 Dependence of the Moisture Accumulation Indexes in Freezing Ground on the Temperature Gradients

	Silty Suglinok, W_L = 19.9%		Silty Suglinok, W_L = 14.9%	
Temperature Gradient, °C/m	Intensity of Frost Heave	Increase in Moisture, %	Intensity of Frost Heave	Increase in Moisture, %
0	0.03	1.8	0.01	3.2
3.0	0.05	2.7	0.03	5.0
4.0	0.07	3.4	0.05	4.9
5.0	0.10	4.4	0.07	5.8
6.0	0.12	5.3	0.10	6.7

the degree of permeability of the soil are predetermined by the outlines of the ice inclusions and the above-described directions of displacement during the freezing and thawing of the ground massifs.

The degree of stress relief of the thawing ground depends on a number of factors: the magnitude of the heave, the thawing rate, the possibility of efflux of the freed water, and so on. A quantitative estimate of the stress relief of the ground can be obtained from the coefficient of frost sensitivity, which is the ratio between the ground strength before freezing and its strength after thawing. The magnitude of this coefficient increases in the case where there is a dynamic (vibrational) load acting on the thawing ground. With other unchanged indexes, the degree of stress relief of the ground is proportional to the magnitude of the relative heave and the vibration intensity. In time the strength of the ground increases as a consequence of drying and strengthening of the cohesional structural bonds.

As a result of these ground-softening phenomena and the peculiarities of its displacement along the embankment during thawing, a number of deformations occur, which can be divided into four varieties. Three of them are equally typical of permafrost regions and regions with no permafrost, and one is typical only of regions with no permafrost: water-eroded surface layers of the soil manifested in sheet and stream erosion; runoff of the thawed layers of the ground through the underlying frozen layers; runoff of completely thawed ground during wetting by surface water; and runoff of frozen ground (sometimes together with the snow cover) as a result of thawing of the ground from below, from the freezing boundary.

It is especially important to dwell on the deformations of the subgrade of a railroad. With high train loadings and an insufficiently thick layer of ballast, during the initial period of operation, depressions form in the subgrade. In transverse section their shape is analogous to the outline of the normal stress diagram from the train loading. The different dimensions of these closed depressions and their slopes lead to nonuniform distribution of the percolating moisture.

Since within the confines of the ballast depression the prewinter soil moisture content throughout the entire active layer is from 2 to 6 percent higher than in the adjacent sections, a frost boil occurs during winter. It has been found that for total elimination of frost boils occurring in this way, an effective measure is grading the ground below the level of the bottom of the ballast depressions or the provision of a vapor barrier above the unevennesses.

In an experimental section containing ballast depressions with closely located groundwater (less than 1.3 m deep), grading of the foundation site was done. As a result, the magnitude of the frost heave was reduced from 110-140 mm to 20-30 mm. Tests were made on vapor barriers of "Brizol," polyvinylchloride, and other materials placed over the depressions within the confines of the ballast prism at a depth of 0.45 m from the surface.

SELECTION OF MEANS AND METHODS OF BREAKING UP FROZEN GROUND ON THE BASIS OF ITS ACOUSTIC CHARACTERISTICS

I. P. BALBACHAN TsNIIOMTP Institute, Central Scientific Research Institute for Foundations and Materials

The execution of earthworks under conditions of seasonal freezing and permafrost calls for the optimal selection of the means and methods of working the frozen ground and the singling out of their logical field of application, depending on the ground conditions.

As is well known, the degree of difficulty encountered in breaking up frozen ground is determined by its resistance to short-term forces and by the plastic deformations preceding disaggregation.

The resistance of the frozen ground to the applied force depends not only on the strength, but also on its viscosity, which is due to the cohesive forces. Of great importance here is the type of soil.

Thus, whereas in the case of frozen sandy soil its resistance to disaggregation increases as the strength increases, with clay soil its resistance to disaggregation depends in large measure on the viscosity or cohesion. Using as an example the shattering of frozen ground by explosions, it is known that with high-strength frozen sandy soil when obtaining identical quality of shattering the specific consumption of explosive is 1.5 times higher than it is for the shattering of seasonally freezing clay soil.

Existing geological-engineering survey methods are used to determine the characteristics of the frozen ground and permafrost from the point of view of providing foundations for buildings and structures and not the working of it. Therefore, as a rule, the upper layer of the frozen ground to be worked is not investigated. Furthermore, these methods make it possible to obtain data on the condition of the ground in individual areas only and give no idea of the overall condition of the soil in the massif with respect to the total depth of freezing and the areal extent of the frozen layer.

During investigations of frozen soil samples under laboratory conditions, their natural structure and iciness are disturbed and the properties vary, which is due to the removal of pressure and the limitations in the sizes of the samples.

Studies conducted in artificially produced samples do not reveal the overall pattern of variation of the properties of the frozen ground and permafrost according to the depth of the layer being investigated, where there is an interrelation between soil structure, cryogenous texture, moisture content, and temperature, having regard to the influence of the surrounding massif.

The method of estimating the density of the frozen ground in the massif using a striker or Research Institute of Road Construction (DORNII) densitometer to determine resistance of the ground to disaggregation does not meet modern research standards, as the readings of the number of impacts can depend on the presence of rock inclusions and ice lenses and their location in the ground. Also, it does not permit us to obtain data on the nature of the frozen ground in the massif and the variations in its properties according to the depth of the layer being worked. Therefore, in our opinion it is inadvisable to estimate the difficulty of working frozen ground on the basis of the readings of a DORNII striker.

As is well known, data from studies of frozen ground are used to determine the soil groupings, which are included in the Construction Norms and Regulations (SNiP), the United Standards and Costs (YeNIR), and various other standard documents. In our opinion, for determining these groupings the most reliable indexes are those of their resistance to drilling. This method, however, also gives approximate results inasmuch as the "monitoring gauge," the drilling time for which is recorded by chronometry is located below the region being drilled, that is, below the upper layer of the frozen ground.

Accordingly, the following physical methods of investigating the soil in a massif appear more promising: acoustic, electric, radiometric, etc. Since the most important physico-mechanical characteristics of frozen ground are the strength characteristics (the tensile, compressive, and shearing strength and the cohesion) and the deformation (the elastic modulus, shear modulus, deformation modulus, and the transverse deformation coefficient), the most promising techniques for investigating frozen ground are acoustic methods, which, in contrast to other methods, make it possible to obtain the quantitative indexes, which are physically interrelated with the elastic, strength, and rheological characteristics of the frozen ground.

When using ultrasonic well logging, the values of the acoustic characteristics of the ground are obtained along the walls of the wells. In the case of acoustic sounding, the emitter and receiver of the elastic vibrations are placed in the blast holes or wells and the mass of frozen ground lying between them is sounded.

It is possible to calculate the elastic parameters of the frozen ground (the elastic modulus, the shear modulus, and Poisson's ratio) from the speed of transmission of the longitudinal and transverse waves. The damping characteristics of these waves make it possible to determine the rheological properties. The set of acoustic characteristics makes it possible to determine the strength characteristics, the jointing, and the inclusions in the frozen ground.

We still have only a limited volume of statistical data on the acoustic characteristics of frozen ground. We do have at our disposal some comparative results with respect to the strength characteristics of frozen ground obtained by the standard procedure and we also have the requisite forces for its disaggregation.*

Therefore, in the first stage of the research† we shall determine the correlation between the strength and acoustic characteristics of the frozen ground in a massif. Comparing the latter with the technological parameters of the machines or methods used for its disaggregation, we obtain the quantitative relations between these indexes. For example, with a transmission rate of 2,800 m/s of the longitudinal ultrasonic wave in frozen clayey silt in the vicinity of Vorkuta the optimal specific consumption of explosive is 0.65 kg/m^3. This is the maximum velocity, according to the data of the Caterpillar Tractor Company, at which rippers can be used on the D8N tractor, which has a horsepower of 375 (for the specified type of ground).

When a sufficient quantity of statistical information on the acoustic characteristics of frozen ground and on the forces required for its qualitative disaggregation has been amassed, it will be possible to do away with the traditional methods of investigating their physico-mechanical characteristics.

Furthermore, the possible application of combination studies, using acoustic, electric, radiometric, and other techniques, for the most part within the soil massif cannot be ruled out. With each specific problem there may be its own combined technique. For example, in order to select the means of working frozen ground, it is possible to combine the ultrasonic method with studies of variations in the temperature and moisture content of the soil in the massif, using electrical methods, and with determination of the specific weight of the soil, using the radiometric method.

The acoustic characteristics of the frozen ground can be used as a basis for determining the logical field of application of various machines and methods for its disaggregation, and also the optimal consumption of the explosives, for selecting the appropriate excavation system, etc.

*A. N. Zelenin. Osnovy razrusheniya gruntov mekhanicheskimi sposobami (Principles of disaggregation of ground by mechanical methods). Moscow, Izd-vo Mashinostroyeniye, 1968.

†The work is being done by [Siberian Branch] Research Institute of Foundations and Underground Structures of USSR Gosstroy jointly with the SONIIOSP Institute, using the acoustic studies methodology worked out at the Mining Institute.

SANITARY ENGINEERING COMPLEXES IN SETTLEMENTS IN THE PERMAFROST REGION

L. K. CLARK Clark and Groff Engineering Company, United States

In May of last year this complex was visited by Professor Mel'nikov and his colleagues. It was recently completed at Wainwright (Alaska) and serves 350 people. The complex includes installations that provide the settlement with water for drinking, washing, and showering and affords the opportunities for laundering and drying. It also includes sauna baths, sewage disposal, and wastewater purification plants. The wastewater is purified by means of chemical processes and is reused. The household garbage and solid waste is concentrated and burned. Thus, pollution of the water, the ground, and the air is nonexistent here.

This complex comprises 12 individual sections. It was first assembled in an industrial center located in the temperate zone. The dimensions of the sections permit them to be transported in cargo aircraft to remote sites where they are assembled into a single building at an already prepared construction site.

SOME REMARKS ABOUT TERMINOLOGY

YE. V. PINNEKER Institute of the Earth's Crust, Siberian Division of the USSR Academy of Sciences

In any branch of knowledge, including geocryology, there has recently been a rapid increase in the number of terms, with the result that it is necessary to take a critical approach to usage, the formalization of terminology, and the refinement of the concept meaning. This pertains primarily to fundamental concepts that serve as the basis for new terms.

The repeated efforts of Soviet geocryologists to effect a refinement in existing terms and concepts have not yet met with success. As far as I know, there is also considerable disagreement and indeterminancy in the United States and Canada with respect to geocryological terminology. Moreover, the lack of complete synchronism between the Russian and English terms in geocryology is hindering the translation of scientific papers from one language to the other.

The absence of clarity is even to be seen in the naming of our science and in its translation from Russian to English. This is apparent even in the prospectuses for this conference. Indeed, "permafrost" is a rather inexact equivalent of "merzlotovedeniye"! Another such example is the translation of the word "pingos."

The terminology used in geocryology does not always meet the requirement for a one-to-one correspondence between the concepts and the terms reflecting the meaning inherent in corresponding concepts. By way of an illustration the recently proposed term "kriopegi" [cryopegs] can be cited. It is used to denote water with a negative temperature and the term is literally translated as "cold water." The question that arises is would it be more precise to call such water "negative-temperature water" or more simply "supercooled water"?

The inverse phenomenon is more frequently noted--the multiplicity of terms with identical meaning. One does not have to look far for examples. I shall mention, for example, the well known dispute as to which of the following terms is more correct: "vechnomerzlyy," "mnogoletnemerzlyy," or "besprestannomerzlyy." [Translator's note: All three of these terms are adjectives usually followed by the Russian word for ground or rock and translated together as permafrost.] Here are some more terms with identical meanings: "tolshcha merzlykh porod," "geokriozona," and "kriolitosfera." [Translators's note: Usually translated as "permafrost layers or permafrost region."] In both cases the three terms reflect one and the same concept. Which should be given preference?

The vast majority of the terminological disputes in geocryology [permafrost studies] stem from a lack of coordination between terms or from inexactness in the definition of concepts and are not of fundamental importance. Differences of this kind can therefore be brought into line comparatively easily: It is sufficient to reach agreement on the meaning of terms and their use. With this aim in view it is proposed that the participants in the Second International Conference on Permafrost should select a special terminological committee, which would immediately get to work on resolving terminological problems (in Russian and English) in order that questions of terminology can be submitted for examination and approval at the Third International Conference on Permafrost, which will be held in 5 yr time.

Of course, the means of bringing order into geocryological terminology must not be limited to the work of this committee. In geocryology there will naturally be questions on which it will not prove possible to reach agreement around the table, since the introduction of accuracy into certain definitions will give rise to fundamental differences of opinion. Consequently, a reexamination of the facts and principles forming the basis for the concepts is inevitable.

All of this poses a need for a far-reaching discussion, primarily in the United States, the USSR, and Canada. In my view, the discussion of geocryological (permafrost) terminology should take place simultaneously with preparations for the Third International Conference on Geocryology. Only by means of extensive and collective discussion of these problems will it be possible to make objective recommendations.

REGULATION OF THE HEAT EXCHANGE OF BUILDINGS AND STRUCTURES ON VENTILATED FILLS WITH A FROZEN BASE

WAYNE TOBIASON Cold Regions Research and Engineering Laboratory, United States

The report of O. V. Snezhko and Ye. P. Kabanov discusses a method of preserving the permafrost condition of the ground in the foundations of buildings by means of porous ventilated fills. S. S. Vyalov also mentioned a ventilation system in his paper. Accordingly, I should like to direct the following questions to at least one of these authors:

Is this procedure an experiment or is it a proven method?
Has any quantitative information been obtained under the conditions during operation of the structures?
Is the floor insulated above the foundation and if so, to what extent?
Has the ventilation system been designed so that it is open in winter and closed in summer?
Is this natural ventilation or are fans used?
Has freezing of the pore spaces been observed?

In the figure to the report by Snezhko and Kabanov, various pathways for the passage of air are indicated: upward through the tube; from left to right; within the structure. Which of them is used and when?

RELATIONSHIP BETWEEN HEAVE AND THE MOISTURE-CONDUCTING PROPERTIES OF THE GROUND FROM ITS INITIAL STATE (DENSITY AND MOISTURE)

B. I. DALMATOV Leningrad Construction-Engineering Institute

Being in the city of Yakutsk in the summer in well-built accommodations assigned to the participants in the conference, it is difficult to believe that beneath us there is permafrost extending to hundreds of meters.

The residents of the city of Yakutsk and the organizing committee of the conference have had to work hard to ensure that we do not experience the difficulties associated with living in a permafrost region. We express our deep appreciation for their kind hospitality and for organizing the work of the conference.

As many of the speakers have noted, the study of the mechanisms of development of cryogenous processes is an urgent problem. Its solution is very complex, as it is necessary to take into consideration a large number of interrelated factors that together influence the course of the phenomena being investigated. Therefore, when studying frost heave of the ground, the experiments must be staged with great care, having regard to the previously discovered relationships.

Unfortunately, the information discussed in some of the papers indicates that these relationships are not always considered. Arising from this and the absence of complete data on the conditions under which the experiments were performed, it is difficult to judge the validity of the conclusions drawn from these experiments.

For instance, in the report by G. D. Voroshilov (Section IV), there is almost no discussion of the procedure followed in performing the laboratory experiments (the height of the samples, their depth of freezing, and so on). Of course, it is impossible to establish the intensity of the heaving from the total heave magnitude of the sample alone.

In addition, it is known that the soil moisture before the beginning of freezing is not the only factor on which ground heave depends. As indicated by V. D. Karlov's studies, which were performed at the Department of "Foundations, Footings and Soil Mechanics" of Leningrad Construction Engineering Institute, the intensity of frost heave also depends to a significant extent on the original density of the ground undergoing freezing.[2]

The "heave limit" moisture concept, which was discussed in the report by V. O. Orlov (Section IV), gives rise to major objections. The carefully staged experiments at the Leningrad Construction Engineering Institute demonstrated that with a "heave limit" moisture (according to V. O. Orlov) and even a lower moisture, heaving of suglinoks can be observed. In these experiments the intensity of heave of an unsaturated suglinok also proved to be density-dependent.[1]

The studies performed at the Leningrad Construction Engineering Institute and the Scientific Research Institute of Foundations indicate that the volume of water- and ice-free pores in a frozen clayey silt is almost twice what it is prior to freezing.[1]

It is interesting that the dependence of the intensity of ground heave on the density and moisture can be expressed, as was noted[3] in terms of the specific moisture by volume which for the specific moisture by weight in question increases as the density increases.

It has been found from these studies that in unsaturated suglinoks, which are in the semisolid and highly plastic state, the intensity of heave depends linearly on the specific moisture by volume. Moreover, migration of moisture from the unfrozen part of the sample to the freezing zone is observed here. The migration depends on the permeability of the unsaturated ground, which for each type of soil is a function of the density and moisture.[2]

Recently, judging from the reports published at this conference, a great deal of research has been conducted along these lines at the State University (Section IV). This research has confirmed the dependence of heave and the moisture-conducting properties of the ground on its initial state (density and moisture).

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GEOCRYOLOGICAL-ENGINEERING CONDITIONS AFFECTING CONSTRUCTION ACTIVITY IN THE MINING INDUSTRY OF THE SOVIET NORTHEAST

V. G. GOL'DTMAN All-Union Research Institute No. 1

The total number of negative degree-days in the area of the Soviet Northeast with a continental climate is as much as -6,500. In the coastal region it is -4,000. The total number of positive degree-days are 1,000 and 600, respectively. On the Chukchi littoral this total is 300. The mean annual ground temperatures range from 0°C to -9°C.

For the purpose of describing the construction conditions and the conditions governing mining operations, the following geocryological zones are distinguished: a zone with a ground temperature (at a depth of 20 m) ranging from 0°C to -1.5°C and islands of permafrost; a zone with a temperature between -1.5°C and -3.5°C where the permafrost is discontinuous and the competence of the rocks during underground mining operations is unsatisfactory; a zone with a temperature below -3.5°C and characterized by continuous permafrost occupying the greater part of the Soviet Northeast. In the continuous permafrost zone a major role is played by hydrothermal processes: On the mountain slopes the suprapermafrost groundwater affects the mean annual ground temperature, with the result that it rises from the peaks to the foothills and the thickness of the permafrost decreases correspondingly. In summer the concentration of the suprapermafrost runoff in the alluvial shingle of the gorges and valleys of the mountain streams is attended by sufficient convective heating of the rocks for the existence of taliks, that is, for thawing of the layer that freezes in the winter and for compensation for the heat losses to the permafrost zone. Above these taliks the mean annual surface temperature of the soil is negative. Along the periphery of the suprapermafrost taliks there are open taliks that are confined to places where the groundwater moves vertically through the jointed rock accompanied by sufficient transfer of heat to the surrounding permafrost.

In the lowlands, open taliks and deep suprapermafrost taliks exist beneath the unfrozen part of the area comprising lakes and rivers.

The existing information on the paleoclimate of the Pleistocene affords a basis for conjecture that during this period there were at least three occasions on which the temperature of the ground in the northeastern part of the USSR decreased by 60-80 degrees below the contemporary values. This was accompanied by freezing of the jointed rocks to depths attaining 400-600 m. In the interglacial epochs, with warming the depth of the permafrost base decreased. Simultaneously in the mountain regions tectonic uplifts occurred, denudation processes were operative, and the depth of erosional downcutting was approximately 200-400 m. In the lowlands, ice-rich sediments accumulated.

As applied to the region with alpine relief these interrelated processes lead to the following geocryology-engineering conclusions:

1. During cooling and deep freezing, many open taliks disappeared in cases where the convective influx of heat by the groundwater did not compensate for the increased losses to the surrounding permafrost. During the subsequent warming epochs, these frozen open taliks were not usually restored, inasmuch as the vertical movement of the groundwater between the suprapermafrost and the subpermafrost aquifers was interrupted by the persisting permafrost, the thickness of which had decreased. Consequently, under contemporary climatic and permafrost-hydrogeological conditions, there may be artificially restored open taliks that are heated by groundwater in places where they had previously existed but had become irreversibly frozen during the cooling period.

2. In the lowlands and in many areas of the region characterized by alpine relief, beneath the contemporary permafrost base there is a zone of thawed rocks that had previously been influenced by cryogenous processes--frost heave and development of fissures. The permeability of the rocks in this zone is higher. With the increase in vertical pressure, compaction of the soils is possible.

3. The rocks in the contemporary permafrost were affected by the cryogenous processes of heave and fragmentation, which are exhibited in the slow perennial freezing. However, this is only encountered in regions that were situated below the groundwater during the perennial freezing epoch (below the local erosion base level) and were saturated with water. The higher zones, which froze in the unsaturated state, evidently did not undergo such appreciable cryogenous changes, although their fissures and pores were partially filled with sublimation ice.

Information on the settlement of perennially frozen jointed andesites, tectonically shattered schists and sandstones, and also granites indicates that the slow perennial freezing (epigenetic freezing) under the conditions of a shallow geothermal gradient was accompanied by the disrupting effect of the enclosed volumes of water and by widening of the subhorizon fissures, the vertical component of the rock pressure being predominant on account of the weight of the overlying rock. In the granites there are horizontal ice veins of up to 0.4 m thick that are confined to the partings. Experience gained in prethawing indicates that at a depth of 10.30 m the settlement of thawing perennially frozen andesites, for example, is about 1 cm/m. The specific settlement depends on the vertical pressure on the rocks and is readily accomplished simultaneously with the thawing. For prethawing purposes, hydraulic pressure-percolation thawing is used, sometimes with artificial heating of the water.

In open-cast mining operations, during the exploitation of perennially frozen placer deposits extensive use is made of layered thawing by solar radiation and atmospheric heat during the period from May to September. The daily or periodic removal every 2 or 3 days of the thawed layer by bulldozers, and also the drainage of the melt water, afford a capability of working 3-8 m of permafrost throughout the season, depending on the iciness and the area. The duration of the solar radiation thawing season is extended by 15-20 days when using transparent film covers, which are especially effective in April and May.

In the dredge and excavator areas, the well-point method of hydraulically thawing permafrost is used. This procedure was proposed by an engineer by the name of Miles in Alaska in 1921 and has been greatly improved upon in the USSR as a result of scientific research. The well-points are installed with vibratory rotary drilling rigs, and equipment for heating the river water has been specially developed. In some cases, when there is high permeability of the ground being thawed, the percolation-drainage method is used with gutter or rainwater irrigation.

In the case of underground mining of permafrost placers, the high rigidity of the permafrost deposits is used. At rock temperatures below -3.5°C, the prepared and excavation workings remain stable for some time with almost no timbering. Way of protecting the rock from winter cooling or summer thawing are being sought, in particular, panels of polystyrene foam, etc.

These are by no means all of the problems affecting construction in the mining industry of the Northeast, the solution of which is being sought on the basis of general and engineering geological research.

THE CURRENT STATUS AND PROSPECTS FOR DEVELOPMENT OF CONTROL MEASURES FOR NALEDS ASSOCIATED WITH HUMAN ACTIVITY IN PERMAFROST REGIONS AND IN REGIONS WITH A SEVERE CLIMATE (AS EXEMPLIFIED BY TRANSPORT CONSTRUCTION)

YE. V. SHUSHAKOV Petrozavodsk University

The prospects for industrial construction in the regions of eastern Siberia, the North, and the Far East, with their innumerable natural resources, depend on the development of railroad and highway networks.

At the present time work is under way in these areas on the building of numerous railroads and highways. The construction of railroads and highways in many parts of Siberia, the North, and the Far East is a formidable and complex engineering problem. In addition, beginning with the commissioning of the finished projects, at first on a temporary basis and later for permanent use, many difficulties arise in arranging for their maintenance.

Many instances are known in which, as a result of the extreme difficulties associated with their operation, it is necessary to abandon recently built structures and even entire sections of roads and build new ones.[11,22]

The complexities of construction and the difficulties of using roads in these areas are primarily the result of the harshness of the climate and, in consequence, of clearly defined permafrost phenomena.

Naleds are numbered among the typical permafrost phenomena, which originate after the building of railroads and highways.

Whereas the formation of naleds on roads in the southern parts of Siberia and the Far East gives cause for great anxiety on the part of the personnel using them, in the northern regions they pose a serious threat to the safety of trains on railroads and can completely paralyze traffic on the highways.

There are large numbers of icy sections on the following railroads: Tayshet-Lena, Khrebtovaya-Ust'-Ilimskaya, Izvestkovaya-Urgal, Reshety-Voguchany, Abakan-Tayshet. The same can be said of the highways connecting Skovorodino-(Never)-Tommot-Yakutsk, Irkutsk-Listvennichnoye, Bratsk-Ust'-Ilimsk, etc.

The expansion of the network of railroads and highways and also the strengthening, reinforcement and completion of existing ones in order to reduce operating costs is forcing us to devote much more attention to problems of naled formation than previously. It is apparently for this reason that during the last 10 yr many more research studies have been devoted to naleds than in all of the preceding years.

At the same time, a characteristic feature of the recent studies has been their descriptive content, the discussion centering primarily on problems pertaining to the genesis of naled processes. On the other hand, problems pertaining to naled control are inadequately treated and not correlated at all with the other types of permafrost-hydrogeological phenomena affecting transporation structures in general and earth roadbeds and artificial structures in particular. This, in turn, is the reason why naled control measures do not always achieve the desired result, even though they are recommended by many standard documents.

Following the traditional separation of the various methods of controlling naleds into "active" and "passive,"[8] depending on the extent to which they influence the naled processes, on the basis of a long-term study of these problems it can be stated that the most prevalent methods, both in the USSR and abroad, are of the "passive" kind. They are used not only in the construction and temporary operation of the structures, but also during their permanent operation.

By "active" naled control measures we mean those that either completely eliminate naleds or cancel out their harmful effect on the structures. By "passive" we mean measures that do not prevent the emergence of naleds, but merely weaken the effect of the naled on the installations being built or in use or measures that partly eliminate the consequences of naled processes.

Active methods of naled control are various types of drainage and catchment, regulating the runoff of small rivers, creeks, and springs; the use of permanent-type frost belts and earth embankments with shields and intakes; and also the elimination of various types of errors that arose during the planning stage. Examples such as replacing one type of culvert by another, the rerouting of a road or raising the shoulder of an earth roadbed, etc., call for very heavy capital investments, which often amount to hundreds of thousands of rubles for a single project, the long-term efforts of the large collective of builders, the use of powerful construction machinery, etc.

Passive methods of naled control--snow banks, temporary picket and trellis-type fences, seasonal frost belts, and also all possible methods of removing the ice formed, by fragmenting, melting and blasting it, the drainage and passage of naled feed water through special heated ditches, board weirs, and chutes, etc.--entail enormous operating expenses, reaching many millions of rubles each winter in our country, as well as the enlistment of scarce manpower and heavy consumption of materials.

Thus, according to the long-term observations by the author, on the approximately 700-km-long Tayshet-Lena line, which is the main section of the BAM (Baykal-Amur) trunk railroad, every year from 20,000 to 50,000 rubles and more than 5,000 man-days are expended on naled control measures.

It should also be noted that despite these heavy costs, passive methods of controlling naleds rarely achieve the desired result with respect to protecting the structures from the effects of icing, nor do they insure traffic safety on railroads or highways situated in regions characterized by a harsh climate and permafrost.

The "passive" naled control measures can be recommended as temporary measures only suitable for use in the course of constructing or finishing the work on a linear installation during the early stages of its temporary operation.[1,6,7]

It is important to note that "passive" naled control measures are themselves essential, for without them the operation of transportation and other structures would be altogether impossible. In addition, when implementing them damage is frequently caused to the installations being protected from the naleds.[2]

Because of the specific manner in which the naled formation processes develop, and in particular the tendency of naleds to originate and grow in newly-built sections, it is almost impossible to make advance provision for the requisite naled control measures and to arrange for their timely inclusion in plans and estimates.

It has been found in practice that by far the greatest number of the iced sections originate only after the building of the earth roadbed and the installing of bridges and piping on highways and railroads.

When carrying out surveys relating to the drawing up of specifications and the development of working drawings for the building of a railroad, insufficient attention is devoted to problems of hydrogeology. This leads to improper designation of the type of structures, to inaccurate selection of their designs, and may go far to explain the sharp increase in the number of ice-covered installations of many railroad lines. The effect of building an earth roadbed and installing bridges and piping on the formation of naleds is enormous, and failure to consider icing phenomena when drawing up plans leads to unjustifiable errors. On the other hand, contributing factors in the formation of naleds and their increase in size are shortcomings in the standard designs for the earth roadbed and ancillary structures, which make no distinction between temperature zones and hydrogeological conditions.

Thus, for example, in building a connecting section of the BAM--the Izvestkovaya-Urgal railroad--the number of naleds, according to N. A. Rogozin and V. N. Makarov, increased from 102 to 247 by the time construction was completed.[5,8] On the Tayshet-Lena railroad the number of naleds, according to the author, had increased from 31 to 151 by the end of construction.

Approximately the same conditions are observed in foreign countries. Thus, on the North American continent, when building the Alaska Highway (Alaska to Canada), according to V. L. Eger and V. T. Prior, 200 new icing sections appeared.[21]

This happens because the designs of the naled control structures are usually developed by the planning organizations and are issued several years after the earth roadbed and ancillary structures have been built, when the construction gang either have been relocated from the sites where the naled control structures are required or they are not foreseen at all by the planning and authorizing agencies, who tend to regard naled control problems as being of secondary importance, even though naleds do great harm to the earth roadbeds and structures.

Thus, a more precise definition of the previously adopted design solutions for the construction of principal naled control structures, which almost wholly embody "active" methods of naled control, becomes necessary to one degree or another at the very moment in the construction program when the budgetary allocations have been largely used up and the construction gangs have been relocated to other sections and projects. Such was the case, for example, when completing the Tayshet-Lena railroad, the Abaken-Tayshet railroad, and so on.

In this event, the implementing of plans for naled control installations drags on for years and entails handing over responsibility for such control by passive methods, initially to the temporary operating staff, and, after commissioning of the road, to the personnel responsible for its permanent operation.[1,3,5,6,12,13]

The long delay in the issuance of plans for naled control structures is chiefly a consequence of inadequate investigation of the naled formation processes. In some cases the planners, when issuing the working drawings for an earth roadbed and ancillary structures, are unable to determine whether naleds will form after the installation has been built.

At the same time, the current status of research on the problems of naled formation is such that, with a known degree of accuracy and in a form suitable for practical use, the basic parameters of the naled formation processes can be determined as a function of a variation in the local hydrologic, hydrogeological, geological, engineering, permafrost, and other conditions resulting from human economic activity.[6,10-17]

From the foregoing it is evident that a possibility exists of forecasting naled formation in the course of carrying out surveys and geological-engineering investigations associated with the planning of the main construction projects. For this purpose, however, in some cases it is necessary to carry out experimental field work in order to identify the sections where naleds may form, so that, with the coming of winter, it will be possible to simulate on the spot the disturbances of the permafrost, hydrogeological, and other conditions that will be caused by the planned structures.

Here it must be kept in mind that the implementing of programs relating to the prediction of naled formations requires heavy expenditures of time and also the efforts of a large and highly qualified staff of surveyor-geocryologists who are capable of working at the research level. An example of this is the planning and construction of the Khrebtovaya-Ust'-Ilimskaya railroad adjoining a main section of the Baykal-Amur trunk line the Tayshet-Lena railraod, where, by a specially formed expedition of the Tomgk State Planning and Surveying Institute for Transport Construction and the Department of Permafrost Studies of Moscow State University, it proved possible to perform a large volume of work, enabling the planners to determine in advance and with a sufficient degree of accuracy the main sections of naled formation and to design the requisite control structures in such a way that their cost was included in the overall construction estimate, and enabling the builders to implement them in the course of building the main components of this railroad.

In analyzing the effectiveness of the modern naled control facilities existing in the arsenal of planners and builders, both in our country and abroad, with few exceptions they are time-consuming and call for manual labor, for they are almost not amenable to mechanization. Unfortunately, most of the standard naled control measures and structures recommended by the technical specifications and the conditions under which construction takes place in permafrost regions with a harsh climate do not give favorable results. For the most part they rapidly become unserviceable, and 1 or 2 yr after construction they need to be repaired.

Therefore, in our view, the priority task of the scientists and engineers engaged on naled control problems is to work out new principles of controlling them, based on the specific genetic type and characteristics of the icing process and also on the peculiarities of the interaction of the structure being planned with the surface or goundwater.

Considering that the facilities under construction and in use disturb local hydrogeological, hydrologic, and geological-engineering conditions that have evolved over a long period of time, the cardinal principle of controlling naled formation processes is that there should be the least possible disturbance of the local conditions or even an improvement in them and that this should be achieved by means of the design characteristics of the structures being built themselves. Although these facilities and structures will not be naled control structures in the direct sense, they incorporate components of naled control structures and there is therefore good reason for regarding them as such.

In the future such structures must find wide application, since their cost must be appreciably less than that of ordinary bridges, culverts, or embankments and hollows in an earth roadbed, including the separate naled control projects. The same can be said of the appreciably lower expenditures on the upkeep of such structures. Economic calculations indicate that these universal structures are more efficient than ordinary ones. In addition, the calculations show that they are more reliable and must result in greater operating safety. They must include the new designs proposed by S. I. Gapeyev[2] for a culvert on a percolating rock bed and for pipes with an increased height to pass the naleds.

Also deserving of careful study and experimental testing is A. F. Matveyev's design for a culvert on a pile foundation with inset panels and also a number of new designs for reinforced concrete bridges recently developed by the Leningrad State Planning and Surveying Institute for Transport Construction (Lengiprotrans) especially to permit the obstacle-free passage of naleds along the beds of water courses and dry valleys.[2] These universal measures must also include a number of new designs for culverts and bridges developed at Petrozavodsk State University under the author's supervision.

At the present time the builders are starting to make wide use of industrial drainage structures, which are longer lasting and, above all, permit mechanization of the work required for their installation.

In this connection, it is first of all necessary to mention the important design developments at the Central Scientific Research Institute of the Ministry of Railroads, the S. Ya. Zhuk All-Union Planning Surveying and Scientific Research Institute (Gidroproyekt), the Rostov-on-Donu Construction Engineering Institute (RISI), and elsewhere.[3,10,13,16-19,21]

Mention must also be made of the prefabricated drainage structures composed of porous "shungizite" concrete developed under the author's supervision at Petrozavodsk State University. These closely correspond to the work by many times and result in major financial savings.[20]

The main characteristic of the proposed and developed designs for the drainage structures is the relative thinness of the prefabricated elements, amounting to no more than 0.15-0.25 m. In spite of this, the porous drainage components completely absorb and drain off the groundwater over a long period of time without significant silting up of the pores in the drainage elements. It is necessary to emphasize that in this respect these designs are much more efficient than the ordinary covered ditch type.

Using the new drainage machines developed under the author's supervision for laying these prefabricated drainage elements (inventor's certificates Nos. 335341, 319681, and others), the heavy and time-consuming manual work involved in installing drainage structures for naled control has been almost completely eliminated. In addition, the author has developed several new naled control structures and devices (some of which are now undergoing experimental testing on railroads).

Based on the design characteristics, they can be lumped together into three independent types:

1. Naled control structures and devices used on railroads and highways which are built and in operation.
2. Naled control structures and devices used

during the construction and operation of highways that are built or are in the process of being completed.

3. Naled control structures used only when building new railroads and highways.

Their development was based on the view of many researchers that the formation of naleds in bypass structures on water courses with subchannel groundwater streams is basically due to their becoming blocked up by sheets of metal piling left by the builders when excavating trenches and also by the footings of bridges and pipes. On escaping at the surface in the form of dispersion fans, the groundwater quickly freezes in the winter and forms naleds that fill up the opening of the ancillary structures. As a rule, all of these ice formations originate after the construction of the road projects has been completed, and they are not observed prior to its commencement.

In order to control these types of naleds, it is necessary to reproduce the hydrologic conditions that existed prior to the construction of the road.

For these purposes the naled control structure developed by the author can be recommended, as it makes it possible to control the naleds on roads that have been built and are in operation, the routes of which pass through mountainous areas or where the beds of the water courses have steep slopes.

The operating principle of this structure is the capture and removal of the alluvial groundwater through a heated drain suspended on brackets inside a culvert. This type of structure is also recommended for roads under construction, which will pass through flat areas.

The chief difference between it and the preceding one is that the drain is buried. When trenchless laying of this drain is possible under a very high fill, it can also be recommended for use on roads in operation.

In addition, on the basis of the new principles for naled control proposed by the authors, other designs for naled control structures have recently been developed at Petrozavodsk University. They can be separated into four main groups:

1. Structures for controlling ground icings;
2. Structures for controlling channel icings;
3. Structures for controlling runoff icings;
4. Structures for controlling mixed icings.

To provide a general description of these designs under the author's supervision, an "Album for Naled Control Structures and Devices" has been prepared by the author, which includes a program classification of naled formations and also the naled control measures developed and recommended for use under specific geological-engineering conditions.

The discussion and experimental testing of new principles to be used in the planning of naled control measures, and also the new designs for the naled control structures and devices, will provide a basis for identifying the most efficient and for receommending those that are best fitted for wide use.

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METHODS OF SIMULATING AND CALCULATING NATURAL RESOURCES AND RESERVES OF SUBPERMAFROST GROUNDWATER

YA. V. NEIZVESTNOV *Association Research Institute of Arctic Geology, Sevmorgeo*

This paper is concerned with the problems discussed in part V, entitled "The Groundwater of the Permafrost Region."

1. I shall answer Ye. V. Pinneker's question: "Why was the term cryopeg introduced instead of supercooled water?"

The term cryopeg was proposed by Prof. N. I. Tolstikhin in 1970. A cryopeg is [a body of] water at a temperature below 0°C but not supercooled. In the majority of cases this is water with a freezing point below zero as a result of its high content of dissolved salts. The natural negative temperature of the cryopeg is higher than or equal to its freezing point. Sometimes the cryopeg temperature can be below its freezing point--in the event of its being supercooled.

The term cryopegs, like the term "therms-supertherms," reflects their temperature. A peg is cold water with a temperature above 0°C (the term was proposed by O. K. Lange in 1933) and a cryopeg-water with a temperature below 0°C.

2. Concerning methods of simulating and calculating the natural resources and reserves of subpermafrost groundwater.

There is evidence that recharge of the subpermafrost aquiferous layers can take place not only through taliks, but also under certain conditions through the frozen zone. Favorable conditions for the migration of water through the frozen ground are created at a temperature close to 0°. In the vicinity of Vorkuta, for example, where such conditions exist, the hydrogeologists N. G. Oberman and V. A. Kal'm have concluded that the frozen zone is not an obstacle for the percolation recharge, that is, for movement of gravity water.

Theoretically any other mechanism of the moisture transfer through the frozen zone is possible. Theoretical developments in the physics and physical chemistry of frozen ground presented at this conference in the form of published reports and speeches can serve as a key to the simulation of moisture transfer through the frozen zone and to the estimation of subpermafrost water recharge.

EXPERIENCE IN FOUNDATION CONSTRUCTION IN WESTERN YAKUTIA

G. L. GOMELAURI, I. YE. GUR'YANOV Scientific Research Institute of Foundations and Underground Structures

The area comprising western Yakutia lies within the continuous permafrost zone. It encompasses about half of the Vilyuy River basin in the Lena Olenek interfluve. The relief consists of low hills and ridges with absolute altitudes of 300-600 m above sea level. The climate is sharply continental. The mean annual air temperature ranges from -6° (south of Mirnyy) to -13° (in the upper reaches of the Markha River).

Geologically speaking, being situated at the junction of the Tunguss and Vilyuy synclinal regions and abutting upon the Anabar anticline from the south, the area encompasses a fracture zone and a zone of intensive trap magmatism (Figure 1). Therefore, the sedimentary rocks here are frequently breached by trap dikes or overlain by thick formations of the sill type. As a result of the denudation processes involved in the development of the relief the lower Paleozoic, sedimentary rocks are almost at the surface under a 2-3-m series of residuum. On the watersheds there are often traces of ancient erosion processes in the form of sandy-pebbly deposits of Mesozoic age.

The trap intrusions were formed by rocks of great strength that had been broken up, as a result of their high degree of jointing, into individual blocks. The sedimentary rocks of lower Paleozoic age consist of an argillaceous-carbonaceous complex, often including irregularly alternating beds of limestone, sandstones, dolomites, marls, and clays.

Figure 1 is a schematic geocryological map of the region, in the compiling of which reference was made to the All-Union map incorporated in the Construction Norms and Regulations (SNiP)[1] and the regional map of P. I. Mel'nikov.[2] It will be seen from the figure that on the whole the permafrost thickness follows the contours of the area's geological structure, and within the region in question it is 300-600 m. The areas marked by abrupt variations in the permafrost thickness coincide with the structural boundaries of the fracture zone and the Paleozoic cover.

The permafrost temperature at the depth of the zero annual amplitude ranges from -1° to -7°, depending on the geographic latitude, the morphological peculiarities of the terrain, and other factors. In some of the watersheds talik areas have been recorded in the sandy-pebbly deposits. The mean annual temperature of the trap rock is normally 1°-3° below the temperature of the sedimentary country rock.

The thickness of the layer with annual temperature variations is 15-18 m. The depth of seasonal thawing ranges from 0.4 to 3 m, depending on the lithological composition of the rocks, their moisture regime, and other factors.

The composition of the lower Paleozoic argillaceous-carbonaceous complex is such that to a depth of 30-50 m loan and strongly weathered marls make up 50 percent and more of the total thickness of the section. Within the first 10

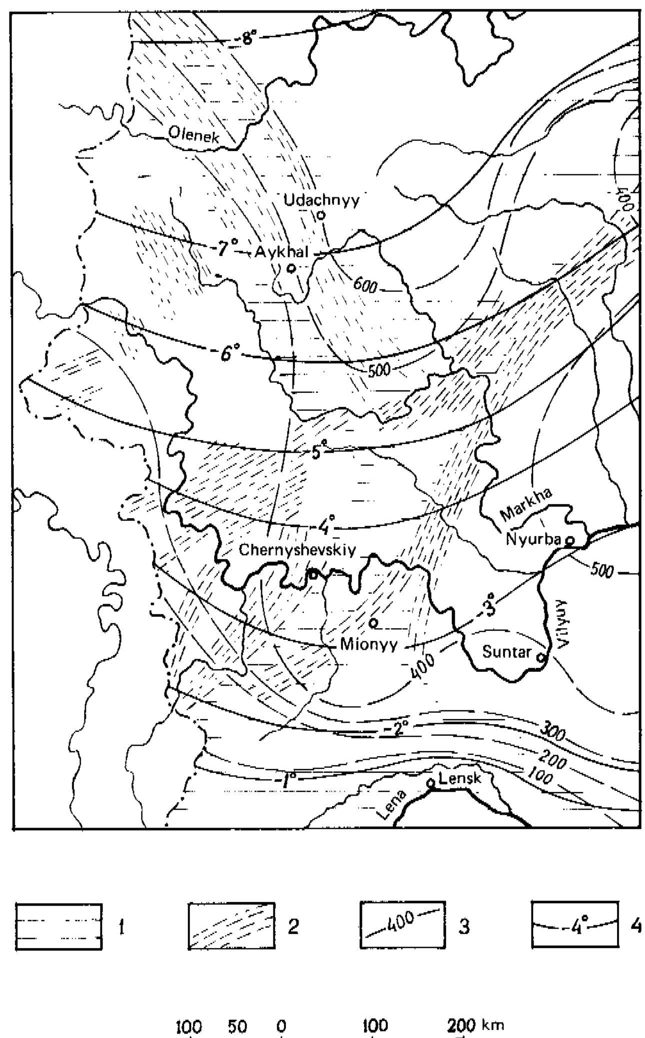

FIGURE 1 Schematic permafrost-geological map of western Yakutia: 1--regions characterized by rocks of the lower and middle Paleozoic; 2--zones of fractures and trap intrusions; 3--isolines of permafrost thickness; 4--isolines of the rock temperature at a depth of zero annual amplitude.

m from the surface these rocks have a total moisture content of 20-30 percent; with thawing they display a tendency to consolidate. With depth the total moisture content of the frozen argillaceous rocks gradually decreases, and below 15-20 m it reaches a value of the order of 10 percent, whereupon these rocks become almost non thaw-consolidating.

The features of the geological-engineering and permafrost setting predetermine the principles governing the use of permafrost as foundations for structures.

Industrial enterprises characterized by damp operating conditions and heavy floor loadings (garages and repair shops) are mostly erected on a rocky base consisting of trap dikes or dense limestone. This base remains stable even after the ground thaws. As a rule, general-purpose industrial buildings and civil buildings are erected on columnar foundations buried in a rocky base to a depth of 0.5 m (Figure 2b). Under the foundations there is a sand or concrete fill. The floors of the first story are laid at ground level with replacement of the surface layer of subsiding or consolidating ground by a gravel-sand fill.

The rocks of the argillaceous-carbonaceous complex are used in foundation for structures only when in the frozen state. During the initial years of building the city of Mirnyy, the buildings were erected on columnar foundations sunk to a depth of 3-5 m. Experience has shown that the major disadvantages of foundations of this type are the labor-intensiveness and high cost of the work involved in digging the trenches in the winter, thawing of the foundation soil when the work is being done in the summer, and also the possibility of frost heave of the foundations when loading them at an inopportune time. With the adoption of the impact-cable drilling techniques by construction organizations, the columnar foundations were completely replaced by pile foundations.

The reinforced concrete piles up to 14 m long are installed in previously drilled holes filled with a sandy-clayey slurry of preset composition.

The pile grillages and tie beams are made of monolithic reinforced concrete. In order to keep the foundation soil in a frozen state, ventilated crawl spaces from 0.5 to 2.0 m high are constructed under the buildings. The design of the standard pile foundation is illustrated in Figure 2a.

In the absence of a rock base, foundations taking the horizontal and dynamic loads are constructed in monolithic reinforced concrete in accordance with the estimated stresses. The foundation is made in the form of a grillage tying together the pile cluster. This is how the anchor supports of a suspension bridge across the Irelyakh River were made (Figure 3). Many of the foundations supporting heavy unbalanced equipment are also made in this way.

In areas with high-temperature frozen ground or taliks, artificial cooling and freezing of the ground is practiced. For this purpose the so-called "cold" piles incorporating solid-cast liquid cooling devices are used. The heat exchanges, which are a closed system of two coaxial pipes, were designed at the Yakutniiproalmaz (approximate expansion: Yakutia Research Institute for the Planning of Enterprises in the Diamond Industry) by engineer V. I. Makarov.[3]

The permanent service mains in the populated places of western Yakutia are laid in two versions. In the settlements of Aykhal and Udachnyy provision has been made for installing the mains in the walkways connecting the individual buildings. At Mirnyy the water supply and sewage systems, electric cables, and in some cases the heating systems as well are laid in an underground reinforced concrete utilidor. The depth at which this main is laid ranges from 2.5 to 5.0 m. With a constant width of the utilidor along the outer edge (2.4 m), three types of transverse cross-sections are used, which measure 1.8, 2.5, and 3.3 m in height. The

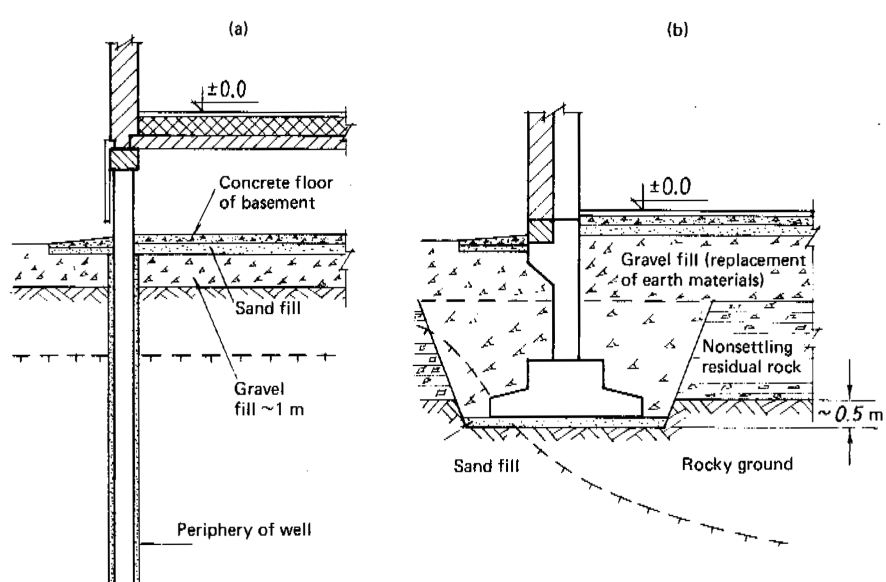

FIGURE 2 Standard designs of foundations: a--pile type (preservation in the frozen state of the earth materials constituting the foundation); b--columnar (allows for thawing of the foundation rock in the course of construction and use of the building).

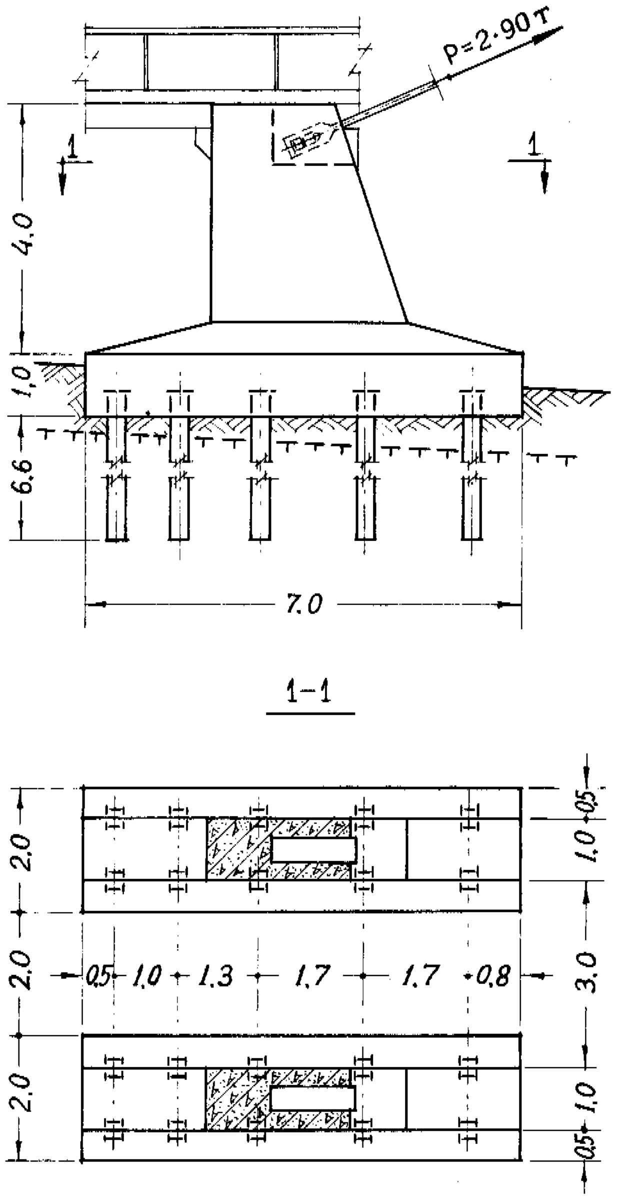

FIGURE 3 Anchor foundations of a suspension bridge across the Irelyakh River.

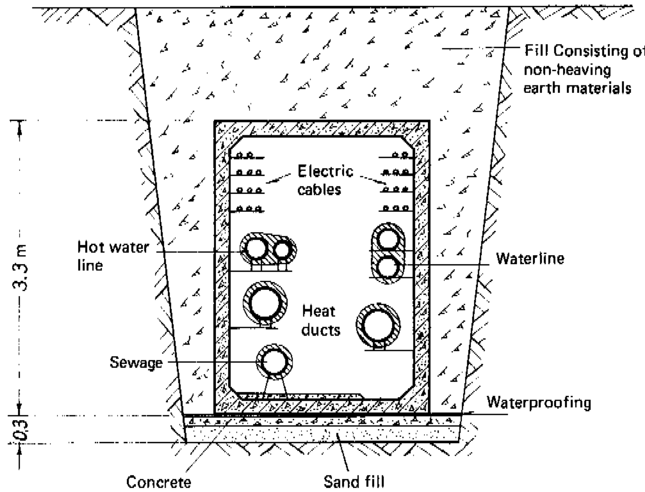

FIGURE 4 Section of the main utilidor.

heating systems are laid jointly with the rest but only in the tallest of the utilidors (Figure 4). The removal of excess heat is insured in winter by natural ventilation through inspection manholes.

The initial plan provided for replacement of the soil in the utilidor foundation by crushed rock to a depth of up to 1 m below the concrete course. Detailed studies of the heat sensitivity of thawing ground demonstrated that this replacement is unnecessary as the possible subsidence of the foundation is insignificant.[4] Moreover, the presence of the drainage course leads to intense flooding of the foundation in case of an emergency in the system with the result that the depth of thawing of the utilidor foundation increases abruptly, reaching 3 m in some areas. Therefore, at the present time replacement of the soil in this foundation is not being done (Figure 4). A common feature of the foundation building techniques used at Mirnyy and other sites in western Yakutia is the preliminary dumping of a crushed rock fill to a depth of 1.0-1.5 m in the zone of activity of all of the construction machinery. On completion of the pile placement operations, the fill is leveled in conformity with the planned elevations and the concrete floor for the future ventilated crawl space is poured so that it slopes toward the outside of the building (Figure 3a).

The leveling of the area being built upon is done both by filling and cutting. If with the new vertical control points an increase in the depth of seasonal thawing is possible and the ground is characterized by thaw-subsidence properties, the plan provided for replacement of the natural soil by nonthaw subsident soil over the entire depth of thawing.

With road construction, provision is made for the preparatory layer of coarsely-clastic nonsubsident soil to be thick enough to ensure that the seasonal thawing will not extend beyond it.

Perennial observations have shown that the development of an area with high-temperature ground leads to a lowering of the soil temperature at the dpeth of zero annual amplitude and, correspondingly, to a drop in the temperature of the permafrost forming the foundation.

At the same time, this process is not universal. Having regard to the direction of the overall variation in ground temperatures, the builders attempted to accelerate the process leading to cooling of the high-temperature soil in the Mirnyy area. For this purpose, bulldozers removed the layer of soil and vegetation over an area of several hectares. This measure, however, was effected without taking into consideration the peculiarities of the relief and the sequence being followed in developing the site, neither was it accompanied by any preliminary calculation. As a result, the cleared area was converted to a zone where atmospheric and melt water accumulated, and the soil-cooling process typical of the territory as a whole was suspended. The soil temperature at the depth of the zero annual amplitudes has remained at the -0.5° level for 3 yr, whereas in the adjacent area it dropped during the same period to -1° and lower. It follows from this experiment that measures aimed at improving the thermal regime of foundations that are not carefully thought out can lead to the opposite results.

The bearing capacity of the pile foundations used at Mirnyy and at other construction sites in western Yakutia is calculated in accordance with the Construction Norms currently in force,[1] beginning with the adfreezing bond strengths of the lateral surface of the piles to the slurry. In marls with a high content of rock inclusions, the bearing capacity of the pile is determined not only by the adfreezing bond strength of its lateral surface to the slurry, but also by the limiting equilibrium of the annular layer of slurry, which necessitates taking into consideration the strength of the country rock.

Experience in the use of industrial structures erected on trap-rock dikes indicates that as a result of the high degree of jointing of the rock, seepage of excess process water takes place into the mass, which results in a marked accelerating of the foundation thawing process. Moreover, there is a danger of thawing of the sedimentary rock underlying the trap rocks. The performance of a jointed trap-rock sill in subsiding thawing ground is the subject of special research.

The following can be numbered among the positive results of foundation construction in western Yakutia.

- The introduction of pile foundations in construction practice as the most industrially-suitable type;
- the design solutions for ventilated crawl spaces ensuring ease of access to the service mains and sure removal of emergency floodwater (height of the crawl space, vertical leveling, and concreting the floor in the crawl space);
- the working out of the designs and mastering of the techniques used in building an underground single-stage utilidor, allowing for greater freedom in the positioning of the piping and for better distribution of the main itself in comparison with two-stage ones;
- the practical use of various methods of artificially cooling the foundation soil.

At the same time, the experience accumulated poses a number of new construction problems. Foremost, among these is how to utilize the reserve bearing capacity of the piles. We are faced with the need to develop pile foundations in which the lower part of the hole is filled with coarsely clastic material of preset composition or with concrete, and also foundations tightly in contact with the soil in the borehole. Construction at the new sites must be accompanied by predictions of the thermal regime of the foundations and developing methods of achieving a directional variation in it.

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CONCLUDING REMARKS BY CORRESPONDING MEMBER OF THE USSR ACADEMY OF SCIENCES

P. I. MEL'NIKOV

Our conference has today concluded the first part of its program. Ahead are interesting tours on which you will see the famous Mamontova Mountains, an outcrop of thick ground ice along the banks of the Lena and Aldan rivers, and classical alas landforms resembling the surface of the moon. Those who are bound for the Northwest will visit the construction site of the first large hydroelectric power station to be built on permafrost. They will see the new city of Mirnyy, which was built under extraordinarily complex geocryological conditions on weak carbonaceous rocks. Those participating in the tour to the Northeast will examine unique naleds near the North Pole and feel their cold breath.

The reports and discussions that we have heard make it clear that in the last 10 yr major advances have been made in all areas of geocryology. The rapid growth of our science is a consequence of the general pattern of scientific and technological progress and the intensive economic development of the northern territories of Eurasia and America.

Geocryology is becoming increasingly complex and multifaceted and calls for closer contact with allied sciences; the use of the most advanced methods; achievements in mathematics, physics, mechanics, chemistry, biology, economics and other sciences; and improvements in the efficiency of scientific work through the automation of experiments, the application of computer technology, and the development of more sophisticated instrumentation.

Until quite recently it was widely held that the frozen strata of the Earth's crust are stable formations, little affected by the passage of time. Yet recent studies have shown that in a number of cases quite pronounced changes in the permafrost thickness and temperature are observed, which points to the necessity of expanding research into the dynamics of permafrost and for predicting the changes that will occur as a result of man's activities on a global scale. It therefore seems an opportune time to embark upon the task of working out the theoretical principles of the laws governing the existence and evolution of the Earth's cryosphere as a whole.

Man's increasing encroachment upon the northern areas, the nature of which, strange as it may seem, is especially fragile and vulnerable, confronts us with the urgent problem of developing as quickly as possible research connected with environmental protection and the efficient use of the natural resources of these areas.

Far too little attention is being devoted to problems pertaining to the relation of the cryosphere to the biosphere. Economic activity in regions with exceedingly harsh natural conditions urgently calls for a fundamental solution of this problem as well.

We are happy to note the rapid growth of cryogeology in the last decade in the United States and Canada. This is particularly evident in the papers presented by the scientists of these countries at the conference. In addition, I personally became convinced of this during my visit to the scientific centers of the United States and Canada. The contacts between Soviet and North American scientists are developing successfully. We have long been engaged upon joint studies with the scientists of Mongolia and Czechoslovakia.

We hope that our conference will be conducive to even closer scientific collaboration between the geocryologists of the various countries in solving the most urgent and complex problems of science.

In view of the widespread interest in the conference and the service it has rendered to the scientists of various countries, we consider that such forums should be held not once every 10 yr but much more often, in fact every 4 to 5 yr. We are therefore proposing that the next international conference be held in 1978 and that international symposia be held to discuss certain of the more urgent problems.

I wish to thank all of the participants in the conference, especially the chairmen and members of the Preparatory Committees of the United States and Canada for the important work they have done. I hope that our scientific tours will be no less successful and that you will see much that is new and interesting. In addition, during the conference we have organized several tours that had not been foreseen in the program and have familiarized our foreign colleagues with the methods used in laying gas pipelines and with the planning and building of installations under northern conditions. We hope that this conference will serve to strengthen the friendship and make for closer elaboration between the scientists of the various countries.

CONCLUDING REMARKS BY THE LEADER OF THE CANADIAN DELEGATION

J. R. MACKAY

In the name of the Canadian delegation and also the National Research Council of Canada, which is in charge of the Canadian Organizing Committee, I thank you, Mr. Chairman, the members of the Organizing Committee of the Second International Conference on Permafrost, and our hosts in Yakutsk, who have worked long and hard to ensure that this conference should be successful internationally. We value your efforts highly.

The president of the National Research Council of Canada, Dr. W. G. Schneider, has also asked me to thank you on his behalf for the cooperative spirit displayed by Soviet scientists over a period of many years towards the scientists of the National Research Council of Canada.

In addition, I should like to thank Dr. Péwé, the Chairman and member of the United States preparatory committee, the co-workers of the U.S. Army Cold Regions Research and Engineering Laboratory, the United States Academy of Sciences, and the National Research Council for their great help in the preparation and publication of the North American contribution to this conference.

Ten years have passed since the First International Conference on Permafrost was held in the United States at Purdue University, Lafayette, Indiana. We have now met at Yakutsk in the USSR for the Second International Conference on Permafrost. Undoubtedly, many are now wondering whether a Third International Conference on Permafrost should be held and, if so, when and where. The question of a Third International Conference on Permafrost was discussed at a number of scientific meetings in Canada, and it was proposed that it be held in Canada if this meets with this approval of those present here today. Therefore, in the name of the National Research Council of Canada I am authorized to extend a preliminary invitation for the holding of a Third International Conference on Permafrost in Canada in approximately 4 to 6 yr time. In extending this invitation I am sure that no difficulties will arise in making the necessary arrangements and that the proposal will be approved by the Canadian government. If the Third International Conference on Permafrost is held in Canada, then let us hope that we shall continue to work in the atmosphere of friendly international cooperation that was successfully engendered at the time of the First Conference in 1963 and has again been manifested at this conference in 1973.

CONCLUDING REMARKS BY THE LEADER OF THE UNITED STATES DELEGATION

T. L. PÉWÉ

In the name of the Board of Directors of the University of Alaska, I express the hope that of the Second International Conference on Permafrost will be a complete success and propose that the next international conference on permafrost be held in Alaska at Fairbanks University.

The Organizing Committee of the United States and the American participants in the conference propose the following resolution: In view of the fact that the Second International Conference on Permafrost in Yakutsk (USSR) required an enormous amount of work relating to its planning and preparation and a variety of arrangements to provide meeting halls, transportation facilities, publications, communications and translation services as well as many other useful arrangements to ensure the success of the conference and the fruitful activity of several hundred participants from many countries, and in view of the enormous contribution it has made to scientific progress and mutual understanding among peoples, by this resolution the participants in the conference express their profound gratitude and appreciation to Prof. P. I. Mel'nikov, the Organizing Committee, the Yakut Republic, the USSR Academy of Sciences, to all who have participated in the work of the conference and to the many others who have contributed to its successful completion.