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TECHNICAL TRANSLATION 1358

PERMAFROST INVESTIGATIONS IN THE FIELD

PART ONE. GEOCRYOLOGICAL SURVEYS
CHAP. II, P. 44 - 88
CHAP. IV, P. 136 - 170, 189 - 201
CHAP. V, P. 211 - 222

PART THREE. LONG - TERM GEOCRYOLOGICAL
INVESTIGATIONS, P. 362 - 421

V. A. OBRUCHEV INSTITUTE OF PERMAFROST STUDIES
ACADEMY OF SCIENCES OF THE U. S. S. R

IZDATEL'STVO AKADEMII NAUK SSSR,
MOSCOW 1961

TRANSLATED BY

V. POPPE

THIS IS THE ONE HUNDRED AND SEVENTY-NINTH OF THE SERIES OF TRANSLATIONS
PREPARED FOR THE DIVISION OF BUILDING RESEARCH

OTTAWA

1969

PREFACE

This translation of the Soviet manual on site investigations in permafrost regions is of particular interest to the Division of Building Research in its studies of permafrost and building problems in northern Canada. The Russians have been involved in construction on permafrost for many years in Siberia and their experiences are of great interest to those who are involved in this activity.

This Russian manual was not translated in its entirety because of its considerable length and irrelevance of some chapters to Canadian conditions. This translation includes about two-thirds of the original text. Comments upon the contents of this translation from any who have had experience in permafrost areas will be welcomed by the Division. Such mutual exchange of information will be of great assistance to the Division in its task of providing essential information on permafrost in Canada.

The Division is grateful to Mr. V. Poppe of the Translations Section, National Research Council, for translating this document and to Dr. R.J.E. Brown of this Division who checked the translation.

Ottawa
May, 1969

R.F. Legget
Director

NATIONAL RESEARCH COUNCIL OF CANADA

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PERMAFROST INVESTIGATIONS IN THE FIELD

PART ONE. GEOCRYOLOGICAL SURVEYS

Chapter II. The nature of investigations in the course of geocryological surveys

2. The layer of seasonal freezing and thawing of soil (rock)

The processes of seasonal freezing and thawing of soil in time, the thickness of the layer of seasonal freezing and thawing, the regime of the layer with annual temperature fluctuations, and the thickness of this layer are just as varied as the natural conditions themselves. In their development they are influenced by the latitudinal geographical zonality and the local altitude.

It is essential to distinguish between "the layer of seasonal freezing" and "the layer of seasonal thawing". The first is the upper layer of soil which freezes in the cold season and is underlain by constant unfrozen soil.

By the layer of seasonal thawing is understood the upper layer of soil which thaws in the warm season and is underlain by permafrost.

Their similarity lies in the fact that in both instances seasonal freezing is followed by seasonal thawing. The difference between them is that the layer of seasonal freezing is developed in the absence of permafrost or when the latter is at a considerable depth, while the layer of seasonal thawing is present only when permafrost is near the surface. The chief characteristics of the layer of seasonal freezing are: possitive mean annual temperature of the soil, the fact that freezing takes place only from above, and thawing from above and from below.

The layer of seasonal thawing is characterized by: negative mean annual temperature of soil, and the fact that freezing takes place from above and from below, but thawing only from above. In the permafrost region, the layer of seasonal freezing is found only on section where permafrost is absent or at a depth exceeding that of seasonal freezing.

The layer of seasonal freezing reaches its maximum thickness in the second half of winter (Fig. 2). The layer of seasonal thawing is thickest in the fall and occasionally at the beginning of winter.

The thickness of layers of seasonal freezing and thawing may vary greatly and depends on the combination of conditions determining the processes of freezing and thawing. The changes in the thickness of the

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layer of seasonal thawing raise or lower the permafrost table. If the soil temperature in this layer rises for several years, the layer of seasonal thawing is transformed into the layer of seasonal freezing. The criterion of stability of the layer of seasonal thawing or freezing is the mean annual temperature of the ground (soil).

During geocryological field studies the following factors must be investigated on each section: a) composition and cryogenic morphology of soils in the layers of seasonal freezing and thawing; b) physical and other properties of soils; c) relation between the depth of freezing and thawing of soils and various physico-geographical conditions (relief forms, surface exposure, depth of snow, type and properties of vegetation, etc.; d) the depth of seasonal freezing and thawing of ground under prevailing conditions.

1. <u>The study of composition and cryogenic morphology of soil in the layers of seasonal freezing and thawing</u>

On studying the composition and cryogenic morphology of soils, it is essential to know their genesis, i.e. the method and conditions of their formation. This problem is solved by geomorphological methods. In the examination of deposits undergoing seasonal freezing, special attention should be paid to the peculiarities of their cryogenic texture.

The cryogenic structure of layers subject to seasonal freezing is still not well understood. On comparing available data it becomes obvious that there are marked differences between the cryogenic structure of soils in the layers of seasonal freezing and thawing. In the layer of seasonal freezing, inclusions, layers and small lenses of ice predominate in its upper part due to unilateral freezing of the soil. In the layer of seasonal thawing the distribution of ice inclusions is uneven, since freezing takes place here both from above and below.

For example, a relatively fine layered texture of frozen soil is characteristic of the layer of seasonal freezing in Western Siberia, while coarse layered texture is characteristic of permafrost.

The following aspects of cryogenic soil texture are characteristic of many areas in the permafrost region where the layer of seasonal thawing freezes both from above and below:

a) the cryogenic texture is best defined in the lower part of the layer near the contact with permafrost, where apart from thin there are also large ice lenses (flat and horizontal in alluvial and peaty deposits, slightly wavy in talus deposits);

b) small ice inclusions of various shapes (Table I) are characteristic of the upper part of the layer;

c) the middle part of the layer is relatively dry owing to migration of moisture from top to bottom during freezing, and therefore the cryogenic textures here are not well defined.

On sections with a low moisture content (steep slopes, dry terraces above the flood plain, etc.), no large ice inclusions are formed in the layer of seasonal thawing.

The cryogenic textures shown in Table I are characteristic of soils undergoing seasonal thawing. In some cases it is relatively simple to recognize alluvial and talus formations from cryogenic textures and other indicators, and to explain the reasons for the varying ice content of the permafrost.

Seasonal cryogenic textures formed in areas of deposition are retained in contemporaneous permafrost. The study of cryogenic textures of the layer of seasonal thawing helps to determine the maximum depth of thawing.

The composition and cryogenic morphology of deposits undergoing seasonal freezing are studied by means of drill cores and by examining the walls of pits and trenches. Of greatest value are pits and trenches where the distribution and the mode of occurrence of ice inclusions can be traced over considerable distances. The depth of pits, trenches and bore-holes must exceed the thickness of the layer of seasonal freezing or thawing by at least 30 - 50 cm.

The tasks of an investigator studying the composition and cryogenic morphology of soil are:

a) to describe the location of the excavation, the elements of relief, the exposure of the section, the degree of swampiness of the surface, the thickness of the moss-lichen cover, etc.;

b) to determine the lithological type of soil (clay loam, sandy loam, sand) and to make a detailed record of the cross-section;

c) to study the primary layering of soil if it is present (type of layering: horizontal, wavy, lenticular, inclined; thickness and composition of each layer);

d) to determine the presence of carbonates, iron and humus;

e) to determine the genetic type of soil and the facies to which they belong;

f) to determine the presence of vegetative remains or peat (for example, the type of peat: allochthonous or authochthonous);

g) to find any signs of gleyzation of soil, note the colour of the layer, determine the depth and horizons when gleyzation is present, find the causes of gleyzation, and determine the outlines and boundaries of gleyed horizons;

h) to determine the moisture content in the layer of seasonal freezing and thawing (for each layer and the total moisture content)*;

i) to investigate thoroughly the shapes, dimensions and distribution of ice inclusions (lenses, crusts, pockets); to describe, sketch and photograph cryogenic textures - the systems of ice inclusions in different genetic horizons; to determine the systems of fractures to which the ice inclusions are related and the processes responsible for the presence of fractures or voids containing ice.

Each genetic type and variety (facies) of soil is characterized by a definite cryogenic morphology, due to the effect of heterogeneous thermal processes taking place in them, the direction of the heat flux and the rate of freezing. Therefore the identification of lithological and genetic types of deposits undergoing seasonal freezing and their geocryological analysis is one of the main tasks in the study of the layer of seasonal freezing and thawing.

2. Relation of depth of freezing and thawing to the natural environment

The thickness of the layers of seasonal freezing and thawing depends on the natural environment of a given region or section. Therefore it is essential to study the physico-geographical and geological characteristics of a given territory.

1. Soil composition. Soil composition affects the depth of freezing and thawing, because different lithological types of deposits differ in their physical and thermophysical properties. Of great importance here is the ice (moisture) content, which determines the main consumption of heat on freezing and thawing and has a considerable bearing on the thermophysical properties of soil. The permeability of thawing and unfrozen soil is of equal importance. The seepage of atmospheric precipitation contributes a great deal of heat to the thawing process.

The depth of seasonal freezing and thawing is greatly affected also by the density, structure and mode of occurrence of the soil or rock. An increase in density raises the coefficient of heat conductivity, which in

* It is important to find the moisture content of soil both in summer and winter.

turn increases the depth of seasonal thawing (freezing).

2. The relief affects the depth of seasonal freezing and thawing because of uneven distribution of snow, different amounts of moisture and peat on the surface, different moisture (ice) contents of surface deposits, etc. In the permafrost region, the thickness of the layer of seasonal thawing on better drained and better heated elevated sections is greater than in depressions.

Outside the permafrost region, the depth of seasonal freezing is greater on windswept elevated sections.

3. Exposure and steepness of slopes. The depth of seasonal thawing on slopes facing south is on the average 50 - 60% greater than on slopes of northern exposure. This difference often exceeds 100% in the southern parts of the permafrost region and steadily diminishes from south to north.

The effect of insolation on southern slopes becomes more pronounced as they become steeper. With all other conditions remaining the same, the difference in the depth of seasonal thawing may reach one metre on slopes dipping at angles of up to 30°.

4. Snow cover, its depth, density, growth and dynamic properties play an important role in the freezing and thawing of unconsolidated deposits.

When the snow cover is thin (2 - 10 cm), the reflecting properties of snow are relatively greater than its insulating characteristics. With increase in the depth of snow, its insulating properties become more pronounced. On sections free of snow, the depth of seasonal freezing is at times 40 - 60% greater than on sections with an undisturbed snow cover.

In certain continental areas, the effect of snow is relatively uniform. The depth of thawing of soil is little affected by snow, since it often disappears prior to the onset of warm weather.

5. Vegetation (living and dead) has a diverse effect on freezing and thawing of soil. Vegetation and organic remains in and on the soil (matting, moss cover, peat) favour an increase in the moisture (ice) content in its upper horizons, lower the temperature and reduce its amplitude, and reduce the depth of freezing and thawing. The intensity of solar radiation is reduced (as much as 50 times and more) under a forest cover and this limits the extent of soil heating. By lowering the force of the wind, forest vegetation favours a uniform deposition of snow and its retention in a loose state, which reduces the amount of heat emitted by the soil. Therefore, the depth of thaw beneath a forest cover is two or three times less than on sections devoid of trees. The greater the density, height and compactness of the vegetation, and the higher the content of vegetative

remains in the soil and on its surface, the greater its effect on the freezing and thawing of soil.

Determination of relationships between the depth of seasonal freezing or thawing and the lithological composition, relief, snow, vegetation, etc., is one of the tasks of field investigations. Quantitative relationships can be determined only by special observations at field stations.

The depth of thawing and freezing must be investigated separately for each type of terrain.

The following factors must be investigated in the field:

a) the thickness of the layer of seasonal freezing and thawing in unconsolidated deposits of different lithological types and under uniform or complex environmental conditions;

b) the effect of different forms of relief on freezing and thawing under uniform and complex environmental conditions;

c) the thawing of soils on slopes of different exposure with similar and different angles of dip, similar lithological types of deposits, similar vegetation, etc.;

d) the relation between the depth of soil freezing and the depth and the density of the snow cover;

e) the effect of different types of vegetation on seasonal freezing and thawing of unconsolidated deposits in different types of terrain.

The observations are recorded as described below (see appendix to this section).

3. Determination of the depth of seasonal freezing or thawing of soil

The following methods are used in the field to determine the depth of freezing or thawing of soil: a) direct measurement, b) temperature measurement, c) extrapolation, d) textural method, and e) calculations.

a) Direct measurement of the depth of seasonal freezing is based on the determination of the depth of freezing and thawing with the help of excavations and boreholes. The layer of seasonal freezing is studied on a cleaned wall of a testpit or in the core from a borehole. The depth of freezing is determined from crystals and layers of ice seen with the naked eye, from changes in the strength of soil and its colour (frozen soil is usually somewhat lighter in colour than unfrozen soil), and by observing the thawing of a sample.

The pits are excavated in dense soil, while boreholes are usually drilled in sandy soil and clay loam. A probe (a pointed steel rod 5 - 6 mm in diameter provided with a handle and gradations every 5 cm) is often used when measuring small depths of seasonal thawing (1 - 1.20 m). However, this method is not always reliable.

The measurements are repreated two or three times at each given point and reliable results are recorded in a log-book. The depth of thawing determined at the end of autumn is the maximum depth for a given year and a given section. The depth of seasonal freezing is determined in the second half of winter or early spring. The measured thickness of this layer indicates the conditions of freezing in a given year and a given section.

b) The temperature method is used when processing the temperature measurements obtained in the field and at research stations. A temperature curve is plotted, and the depth at which the curve crosses the zero ordinate will correspond to the depth of thawing or freezing. This method is less accurate than direct determinations. It should be remembered that the cooling of certain soils to $0^{\circ}C$ may not always coincide with the depth of actual freezing of soil. Such a discrepancy is observed in fine-grained soils in which water crystallizes at temperatures below zero. If the mineral content of water is high, the crystallization sets in at a still lower temperature.

A zero soil temperature is often observed over a relatively large interval of depth, which points to the inadequacy of this method.

c) The method of extrapolation is based on observations of the rate and depth of thawing of soil carried out at the nearest meteorological station. It is known that seasonal thawing of soil is a function of time. Therefore for any period of time, thawing may be expressed as a percentage of the maximum thawing in the course of a season. Such data for each time interval serve as a standard in the calculation of maximum thawing based on observations obtained earlier. This method can be used in the case of uniform environmental conditions.

Table II contains the data on the rate of seasonal thawing near Igarka (Tumel, 1941) and near Yakutsk (Mel'nikov, 1952) expressed as a percentage of the maximum depth of seasonal thawing.

In both localities the rate of thawing has been determined over a period of many years at grass-covered sites with water content of soils not exceeding saturation of the latter. The rate of thawing is shown graphically on Figure 3. The calculation of the depth of thaw based on the data in Table II can be done by means of the following formula:

$$H = \frac{h \ 100}{n},$$

where H - the maximum depth of thaw, in cm or m;

h - depth of thaw at the time of the investigation, in cm or m;

n - coefficient of thawing at the time of the investigation, in %.

Example. At the end of August, the silty clay loam near Igarka thawed to a depth of 1.20 m (h). From Table II we find that for this region the depth of thaw in August (the coefficient of thawing n) is 60 or 80% of the maximum depth, depending on the weather, lithological composition of soil, and other factors. Let us assume that the summer was relatively dry. Hence the maximum depth of thaw in October (H) will be 1.50 m:

$$H = \frac{h \cdot 100}{n} = \frac{1.20 \cdot 100}{80} = 1.5 \text{ [m]} \qquad a)$$

Below we give the intensity of seasonal thawing of unconsolidated deposits near Noril'sk (in % of maximum depth of seasonal thawing) calculated by A.V. Leont'ev, and the possible error in the determination of the depth of thaw at different times of the year:

	1.VI	15.VI	1.VII	15.VII	1.VIII	15.VIII	1.IX	15.IX
Depth of thaw, %	5	25	45	65	75	90	95	100
Possible error, %	±50	±25	±20	±15	±10	±5	±3	0

d) The method of determining the long-term maximum depth of seasonal thawing of soil from its cryogenic texture is as follows.

The changes in ice content with depth are observed at a chosen site in a testpit or borehole. Where the ice content gradually diminishes with depth, reaches a minimum value (ice ceases to be noticeable) and then increases sharply, the boundary of this transition often indicates the depth of maximum thaw. To ascertain that this is the maximum depth, it is essential to deepen the pit or the borehole by at least one metre. If the thickness of ice layers increases or the soil is more or less uniformly saturated with ice, then the aforementioned boundary does coincide with the maximum depth of thaw. In the presence of alternating massive and layered textures, the maximum depth of thaw will be indicated by the lowest boundary of the transition from massive to layered texture. The uppermost boundary of this transition corresponds to the depth of seasonal thawing in one of the years following the maximum thawing.

In such determinations it is recommended to use the core lifter designed by A.M. Pchelintsev (1951) et al. The extracted cores are sawn along the axis, cleaned with a knife, and photographed (or carefully sketched and described). The core photographs are glued together to form one column and this is used to complete the analysis of the cryogenic texture of a given soil.

The advantage of testpits over boreholes lies in the fact that in a pit it is possible to trace the boundary of thawing for a considerable distance. However, the excavation of pits requires more labour and is made difficult in the summer by caving and flooding. Therfore it is not always possible to excavate a pit.

Let us examine some examples of the cryogenic texture of some typical cross-sections of frozen soil.

Table III contains the characteristics and sketched examples of various cryogenic textures. Tables IV to IX contain typical cross-sections of frozen soil and determinations of the depth of seasonal thawing from cryogenic textures.

A section of epigenetic frozen soil typical of Western Siberia is shown in Table IV (section I), where the lower boundaries of contemporaneous and maximum seasonal thawing coincide and lie between layers 5 and 6, characterized by medium- and coarse-layered cryogenic textures. Permafrost occurs above the upper boundary of the layer with a coarse-layered texture.

The medium-layered texture of layer 5 was formed as a result of freezing of the thawed layer from below. The fine-layered texture in layer 1 is due to rapid freezing of the soil, which eliminated the possibility of formation of large ice lenses. The textures of layers 3 and 4 are due to a low moisture content of freezing soil. Layer 2 contained a considerable amount of moisture but it froze slowly.

Section II (Table V) differs from Section I by the fact that the boundary of the layer of contemporaneous seasonal thawing in it is above the boundary of maximum thawing (which is between layers 7 and 8) and lies at the base of layer 5 formed as a result of freezing from below. This is indicated by the size and distribution of ice lenses. The considerable thickness of ice lenses points to a high water content of the soil.

In section III (Table VI) the lower boundaries of contemporaneous and maximum seasonal thawing coincide and lie at the base of layer 8. The texture of layer 6 resembles that of layer 5 in section II, although they were formed in a different way. Layer 6 in section III froze from above, while layer 5 in section II from below.

In section IV (Table VII) the lower boundary of the layer of maximum seasonal thawing lies at the base of layer 13. The lower boundary of the layer of contemporaneous thawing is not well defined.

In section V (Table VIII) it is impossible to pinpoint the lower boundaries of contemporaneous and maximum thawing. We can only assume that they lie in the sandy layer 5.

The permafrost layer (5) in section VI (Table IX) is characterized by an even distribution of ice layers. Therefore it is impossible to pinpoint the lower boundary of maximum thawing.

The textural method of determining the maximum depth of thaw may be used under following conditions: 1) when the layer of seasonal freezing is consistently merging with permafrost; 2) then the lower boundary of the layer of seasonal thawing lies in clayey soil (clay, clay loam and sandy loam).

Let us examine two examples pertaining to the first condition.

Example 1. The layer of seasonal freezing has merged with permafrost. It is required to find the boundary between them. According to all that was said above, this boundary corresponds to the depth at which soil with massive or medium-layered cryogenic texture assumes a coarse-layered texture characteristic of permafrost.

Example 2. The layer of seasonal freezing does not merge with permafrost. The depth of seasonal thawing cannot be determined, since the depth of seasonal freezing is less than the possible depth of seasonal thawing. We can speak only of the maximum depth of seasonal freezing as indicated by the pereletoks.

As far as the second condition is concerned, the textural method can be used only in the case of clay soils in which lenticular ice inclusions can be formed. Ice lenses are not formed in sandy and coarse soils, irrespective of their moisture content*.

e) Determination of the depths of seasonal freezing and thawing by the thawing by the method developed by V.A. Kudryavtsev is based on the examination of the effect of a combination of factors determining the seasonal freezing of soils, and on the study of heat exchange. The heat exchange in the soil can be determined by way of the temperature regime at the surface; attention should be paid to soil composition, lithology, texture, moisture content, ice content, and thermophysical properties. V.A. Kudryavtsev (1959a) distinguishes four main indicators, which characterize the processes of seasonal freezing and thawing of soil: a) lithological composition, soil structure, b) moisture and ice contents; c) mean annual temperature and d) temperature amplitude on the surface. The combination of these factors determines the type and depth of seasonal freezing and thawing of soil.

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* When investigating the layer of seasonal freezing and thawing, and particularly when determining the depth of seasonal thawing, it is essential to note the colour of soil. The layer of seasonal thawing is characterized by a rusty-brown colour resulting from the presence of ferric oxides. Permafrost is characterized by a greyish-green, dirty light blue, or blue colour resulting from the presence of ferrous oxides (Russian editor).

In contrast to the usual method of determining the depths of seasonal freezing and thawing at the time of the investigation, the above procedure makes it possible to establish the patterns of formation of the layer of seasonal freezing and thawing.

The mean annual temperature and the annual temperature amplitude at the ground surface reflect the complex heat exchange process in the surface layer of the lithosphere. The lithological composition of soils, their structure, moisture content, and ice content characterize the conditions under which the soils freeze and thaw. It is a relatively simple matter to determine each one of these conditions in the field.

The aforementioned characterisitcs form the basis of the determination of the depths of seasonal freezing and thawing by means of formulae given below. The data obtained by these equations may be used to predict changes in the depth of seasonal freezing and thawing with disturbance of the natural conditions during development of the territory.

Let us examine the sequence of determining the types of seasonal freezing and thawing of soils, using the four indicators mentioned earlier.

a) The lithological composition and structure of soils are determined by means of boreholes and testpits, by examining outcrops, etc. Samples for laboratory tests are taken from typical layers. The distribution of different lithological soil types is traced on a field map.

b) The moisture content is determined in pits and trenches. Frozen soil is obtained by means of trench sampling, while unfrozen samples are taken from every lithological soil type and at least every 50 cm. The data is traced on a field map which is then divided into sections showing different average moisture contents of soils. In each section the differences in the moisture content must not exceed 5 - 10%, but larger differences are possible on small scale maps.

The maps showing the types and moisture contents of soils are compiled by conventional methods used in engineering geology.

The compositions of typical lithological soil types are determined in the laboratory. The number of analyses depends on the scale of the survey and the complexity of soil composition and structure. The unit weight (γ), the relative ice content (i), the specific heat (C), and the coefficient of heat conductivity (λ) of typical lithological soil types are also determined. These properties are determined in both disturbed and undisturbed samples.

c) The mean annual soil temperature (t_{av}) at the base of the layer where the temperature fluctuates throughout the year can be determined by a series of observations and by individual measurements at different depths (Kudryavtsev, 1959c). In the latter case, use may be made of a table showing the relation between the depth of the layer with annual temperature fluctuations (z), the mean annual soil temperature (t_{av}), and the coefficient of heat conductivity (K) (Table X). If the value of K is known, it is possible to find on the curve illustrating the changes in temperature with depth, a point where the values of z and t_{av} coincide with those in the table. These are the required values of z and t_{av}. If K is unknown, then there is only one point on the curve which coincides with the values of z, K and t_{av} in the table. If there are several points, the most likely value of K is chosen. It should be noted that the error in the determination of the depth of the layer of seasonal freezing or thawing does not normally exceed a few centimetres and may be ignored in the determination of z and t_{av}.

By using such individual temperature measurement in stabilized boreholes, it is possible to determine the mean annual temperature of the soil. These data are also traced on a field map and contours are drawn through sections with similar temperatures. Within these sections, the range of temperature fluctuations on large- and medium-scale maps must not exceed $0.5 - 1.0^{\circ}C$.

d) The annual temperature amplitude at the ground surface is determined from the amplitude of mean monthly air temperatures by subtracting the difference which arises owing to the effect of snow and vegetation. Let us examine the method of determining the effect of snow and vegetation on the amplitude of temperature fluctuations (Kudryavtsev, 1954).

The decrease in the temperature amplitude owing to the effect of snow is equal to:

$$S_z = A_a - A_v \qquad (1)$$

where A_a - physical air temperature amplitude*

A_v - physical temperature amplitude beneath the snow (at the surface of the vegetation cover).

From the equation of harmonic fluctuations we obtain:

$$A_v = A_a e^{-z\sqrt{\frac{\pi}{KT}}} \qquad (2)$$

where z - depth of snow in m;

T - period of fluctuations in hours;

K - coefficient of diffusivity of snow in m^2/hr.

* Half the meteorological temperature amplitude (Russian editor).

On substituting the expression for A_v into equation (1) and on denoting $e^{z\sqrt{\frac{\pi}{KT}}}$ by f, we obtain $S_z = A_a\ (1 - \frac{1}{f})$.

When determining the value of $\left(1 - \frac{1}{f}\right)$ it is essential to take the maximum depth of snow with a known coefficient K. The values of K are given below:

Depth of snow, m	0-0.2	0.3-0.4	0.5-0.6	0.7-1.0
K	1.0	0.9	0.8	0.7

The value of $\left(1 - \frac{1}{f}\right)$ is determined from Table XI.

If the depth and density of snow are known and having determined S_z, it is possible to calculate the annual temperature amplitude beneath the snow, i.e. at the surface of the vegetation cover.

The decrease in the temperature amplitude beneath the vegetation cover, including the moss, the grass and the forest litter, is determined from observations at sites prepared in characteristic sections with different types of vegetation. Measurements are made of the daily temperature amplitude at the surface of the vegetation cover (A_v) and in the soil beneath it (A_s). This is done by placing the maximum and the minimum thermometers on the surface of the plant cover and in the soil and keeping them there for 24 hours. The minimum thermometers are heated, while the maximum thermometers are cooled prior to insertion. Such measurements are repeated at least three times on the same site. This gives the values of $A_{v\ day}$ and $A_{s\ day}$, which makes it possible to calculate K of the vegetation cover:

$$A_{s\ day} = A_{v\ day}e^{-z\sqrt{\frac{\pi}{KT}}}, \qquad (3)$$

where z is the thickness of the vegetation cover in m; T is the time in hours (per day).

On establishing the same relationship for the yearly cycle:

$$A_{s\ year} = A_{v\ year}e^{-z\sqrt{\frac{\pi}{KT}}}, \qquad (4)$$

it is possible to determine A_s, since $A_{v\ year}$ is known (see the aforementioned method of determining the effect of snow): A_v is the temperature amplitude beneath the snow; K has been determined on the site.

There is a better way to relate the decrease in daily temperature fluctuations to the decrease in annual fluctuations. This is done as follows.

On taking the log forms of equations (3) and (4), we obtain:

$$\ln \frac{A_{v\ year}}{A_{s\ year}} = z\sqrt{\frac{\pi}{KT365}}, \tag{5}$$

where T is the time in hours per day;

$$\ln \frac{A_{v\ day}}{A_{s\ day}} = \sqrt{\frac{\pi}{KT}}. \tag{6}$$

It follows from equations (5) and (6) that

$$\ln \frac{A_{v\ year}}{A_{s\ year}} = \ln \frac{A_{v\ day}}{A_{s\ day}} \Big/ 19, \tag{7}$$

where $19 = \sqrt{365}$.

Hence it is possible to omit the determination of K and to proceed directly to the determination of A_s (i.e. the annual temperature amplitude on the ground surface) from $A_{v\ day}$ and $A_{s\ day}$. The quotient $A_{v\ day} : A_{s\ day}$ is determined in the field. The difference between repeated determinations must not exceed 0.05.

The annual amplitudes of temperature fluctuations on the ground surface are also traced on a field map.

A final map showing the types of seasonal freezing and thawing of soils in a given region (Fig. 4) is then compiled on the basis of all data obtained. The recorded soil types, within which each of the given factors varies in area within limits imposed by the scale of the survey, are characterized by definite depths of seasonal freezing and thawing. The nomograms in Figures 5 and 6 show the functional relationship between each of these factors and the depth of seasonal freezing or thawing. The nomograms are based on the determinations of freezing and thawing of soils by means of the V.S. Luk'yanov's formula (1957). By using the nomograms it is possible to calculate the effect of each of the mentioned factors and conditions on the process of seasonal freezing or thawing, and consequently to forecast the changes in the depths of seasonal freezing or thawing on disturbing the natural conditions, for example, on changing the moisture content of the soil, removing or creating artificial covers, etc.

The Luk'yanov formula is as follows:

$$\tau = \left(Q + \frac{CQ}{2}\right)\left(\frac{\lambda\theta}{q^2} \ln \frac{\lambda\theta - qS}{\lambda\theta - q(h + S)} - \frac{h}{q}\right), \tag{8}$$

where τ - the duration of the winter or summer season, hours;

θ - mean air temperature in the winter or summer season;

q - average heat flux to the freezing plane from the underlying soil layers in winter, $kcal/m^2/hr$;

S - thickness of the soil layer equivalent to the average thermal resistance of insulation on the ground surface in winter, m;

C - unit specific heat of frozen soil, kcal/m³/deg;

λ - coefficient of heat conductivity of frozen soil, kcal/m/deg/hr;

Q - latent heat of melting of ice in a unit volume of soil, kcal/m³;

h - depth of seasonal freezing (thawing), m.

If we determine the effect of snow and vegetation by means of V.A. Kudryavtsev's formulae and consider the conditions prevailing on the surface of the ground, then S will become equal to zero.

The factors τ and θ are determined from A_s and T_{av} recorded on the field map by assuming that the change in the temperature at the ground surface follows a sinusoidal curve. Then $\tau = \tau_2 - \tau_1$, where τ_1 is the time which has elapsed since the onset of freezing in hours; τ_2 is the time up to the end of freezing in hours.

$$\tau_1 = \frac{T}{2\pi} \arc \sin \frac{t_{av}}{A}. \qquad (9)$$

θ is found from $\theta = t_{av} - \frac{AT}{2\pi\tau}\left(\cos\frac{2\pi}{T}\tau_2 - \cos\frac{2\pi}{T}\tau_1\right)$.

The values of τ and θ may be found by plotting.

The heat flux q from the underlying unfrozen (or frozen) soil is determined by the method suggested by V.S. Luk'yanov and M.D. Golovko (1957).

After certain modifications, Luk'yanov's formula assumes the following form:

$$\frac{\tau q^2}{\left(Q+\frac{C\theta}{2}\right)\lambda\theta} + 1 = \ln\frac{\lambda\theta}{\lambda\theta - qh} + \frac{\lambda\theta - qh}{\lambda\theta}.$$

By substituting v for $\frac{\tau q^2}{\left(Q+\frac{C\theta}{2}\right)\lambda\theta}+1$ and u for a $\frac{\lambda\theta}{\lambda\theta - qh}$ we obtain: $v - \ln u + \frac{1}{u}$.

Given the values of u, we can draw the curve $v = f(u)$. Having solved the left half of the equation and having determined the values of u from the curve, we find the depth of seasonal freezing (thawing) of soil:

$$h = \frac{\lambda\theta(u-1)}{uq}.$$

The nomograms for the maps of seasonal freezing and thawing of clay loams and sandy soils are constructed separately. A nomogram for any lithological type of soil may be constructed if necessary. The initial data for the nomograms (Figures 5 and 6) are as follows:

1. The mean annual temperatures of soil at the base of the layer of annual fluctuations (t_1) are equal to: ±0.1, ±1, ±2, ±3, ±4, ±5°C.

2. The annual temperature amplitudes at the ground surface A (physical)

are 11, 17 and 24^{o}.

3. The moisture contents of soils in the layer of seasonal freezing (thawing) W in % of dry weight are: for sand - 5, 10, 20, 30, 40; for clay loam - 15, 20, 30, 40, 50.

4. The heat conductivity of soils (kcal/m/deg/hr): for sand $\lambda_{unfrozen} = 1.26$; $\lambda_{frozen} = 1.64$; for clay loam $\lambda_{unfrozen} = 0.92$, $\lambda_{frozen} = 1.2$.

5. The unit weight of the soil skeleton (mineral soil) (kg/m³): for sand $\gamma = 1600$; for clay loam $\gamma = 1700$.

6. The relative ice content: for sand $i = 1.0$; for clay loam $i = 0.8$.

7. The unit specific heat of soil (C) is determined using the following formula:

$$C = \gamma_{sk} C_{sk} + \gamma_{sk} \frac{W}{100} i C_{ice} + \gamma_{sk} \frac{W}{100} (1 - i) C_w,$$

where γ - unit weight of the soil skeleton (kg/m³);

C_{sk} - specific heat of the soil skeleton (0.2) in kcal/m³/deg.;

C_w - specific heat of water (1) in kcal/m³/deg.;

i - relative ice content;

C_{ice} - specific heat of ice (0.5).

The amount of heat Q required for phase transformation is determined as follows:

$$Q = \gamma_{sk} \frac{W}{100} i \, 80,$$

where 80 is the amount of heat in calories required to transform one gram of ice into water.

Let us give an example of how to use the nomograms.

Let us assume that it is required to determine the depth of seasonal freezing of clay loam when its moisture content is 20%, the temperature amplitude on the ground surface is 17^{o}C, and the mean annual temperature of the soil is 1^{o}C.

In this case we make use of the nomogram for clay loam (see Fig. 5). The curve corresponding to the moisture content of 20% is selected from the family of curves denoting the physical temperature amplitude of 17^{o}C. A line is drawn from a point on the abscissa corresponding to the mean annual soil temperature of 1^{o}C. The ordinate of the point of intersection of this line with the amplitude curve corresponds to the depth of seasonal freezing. In the given case it is 2.4 m.

A nomogram may be constructed using any formula. "The sole requirement in this case is that the depth of seasonal freezing and thawing of soil must be expressed through the mean annual temperature of the latter and the temperature amplitude at the surface" (Kudryavtsev, 1959).

The use of the aforementioned parameters in calculations eliminates the possibility of human error. The calculated values may be checked by comparing them with actual depths of seasonal freezing and thawing on the site.

A map of seasonal freezing and thawing of soils compiled by this method is in good agreement with general geocryological conditions in a given region and renders it possible to determine the characteristics of this region.

4. Mapping of the layer subject to seasonal freezing and thawing using the landscape method

A map showing the following data on the layer subject to seasonal freezing and thawing is compiled in the course of the geocryological survey:

a) the lithological composition of soils and their structure; the presence of peat, humus and gleyzation; the facies to which the soils belong;

b) the characteristics of the cryogenic morphology of soils, i.e. the cryogenic textures of soils from different facies;

c) the ice content of soils (in winter);

d) the depth of seasonal thawing (freezing) within each element of mesorelief and corresponding soil facies;

e) the moisture content of unfrozen soils (in summer).

This information is mapped by the landscape method with emphasis on the identification of cryolithological soil types comprising the layer subject to seasonal freezing and thawing. The landscape method of mapping this layer is based on the identification of landscapes, i.e. the types of terrain characterized by different physico-geographical and geological conditions, and consequently by different regimes of seasonal freezing and thawing corresponding to these conditions.

The mapping by this method is done as follows.

a) The most typical types of terrain are singled out during field work and their boundaries are traced on a map.

b) the depth of thaw (freezing) is determined several times in each type of terrain during maximum seasonal thawing (or during maximum seasonal freezing in the spring). The average depth and the most typical depth of thaw (freezing) are traced on the map or mentioned in the footnotes.

The schematic map compiled by E.G. Katasonova for a section of a lowland (Fig. 7) may serve as an example of a landscape map. It is a morphological map since it does not reflect the main soil characteristics in the layer subject to seasonal freezing (thawing). It can be improved considerably by including the data on the cryolithological soil types in this layer. The study of cryolithological soil varieties is based on the identification of their genetic types.

The compilation of such a map or a section of a more complex map* is done as follows.

a) In the warm season, work is done on the study of the soil composition within various elements of the mesorelief, the identification of the most characteristic lithological soil types in the layer undergoing seasonal thawing, and the determination of their moisture content.

b) In winter and spring (and in the North even during the first half of summer), investigations are carried out on the cryogenic morphology of soils, i.e. the cryogenic texture (distribution, shape and thickness of ice inclusions) and on the identification of cryolithological types of deposits undergoing seasonal freezing. Their boundaries are traced on the map. In the majority of cases they coincide with the boundaries of the elements of mesorelief.

c) The average depth of thaw (or freezing) for each cryolithological type of deposits undergoing seasonal freezing or thawing is determined in the period of maximum thawing or freezing. The depth of thaw or freezing is measured by drilling, excavating or testing with a metal probe.

The main taxonomical factors identified in the field and traced on the map are the cryolithological soil types (shown by different colours or different types of shading). The data on the map are supplemented by information concerning the composition (shown by symbols), the moisture content and the ice content of soils, as well as the depth of seasonal thawing or freezing (in numbers for each soil type).

Figure 8 shows an example of such a map (compiled by E.G. Katasonova).

It is expedient to supplement the map showing the cryolithological soil types in the layer undergoing seasonal freezing with nomograms illustrating the relationship between the extent of thawing (freezing) and the changes in environmental conditions. The nomograms are constructed according to instructions given in the preceding section.

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* Combination maps showing the types of terrain and the cryolithological soil types (Russian editor).

Appendix 1

Recording of data on seasonal freezing and thawing of soils

Depth of freezing and thawing, m	Date, year	Element of relief (water divide, terrace, flood-plain, etc.)	Element of meso- or microrelief (levee, swampy depres. channel of an intermitten stream, etc.)
1	2	3	4

Comp. of deposits and their moist. cont. (in % of dry or wet wt.)			Degree of swampiness of section	Depth of snow cover, cm
Depth, m	Comp.	Moist cont. %		
5	6	7	8	9

Exposure and steepness of slope	Type of vegetation (forest, meadow, etc.)	Comp. and depth of top soil, cm	Comp. and depth of subsoil and peaty horizon
10	11	12	13

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3. Composition and Cryogenic Structure of Permafrost

The study of composition and structure of permafrost is one of the most important tasks of geocryological surveying. It is known that rock and soil of any origin may be present in a frozen state. Their structural, physical and engineering characteristics depend to a great extent on their ice content. It has been established that the amount of ice, as well as the shape, size and distribution of ice inclusions in unconsolidated rocks depend on the genesis of the latter (the lithology and the characteristics of facies to which they belong), and in consolidated and semiconsolidated rocks on the extent of their fracturing and weathering.

A field investigation of the composition and morphology of permafrost consists of the study of their lithological composition and cryogenic textures.

By cryogenic texture we mean the morphological characteristics of permafrost as defined by the shape, size and distribution of lenses, layers, veinlets, crusts and pockets of ice. The ice in these inclusions is termed texture forming, and that which holds the individual soil particles together is known as ice cement.

The morphology of any soil, unfrozen or frozen, is determined by its structure and texture; the ice obscures the primary structure and texture, which reflect the conditions of soil formation. The term "cryogenic structure" and "cryogenic texture" describe the characteristics of composition and morphology of frozen ground resulting from the presence and distribution of ice in it.

Four types of structure are distinguished in the case of unfrozen unconsolidated soil: coarse (psephitic), sandy (psammitic), powdery (aleuritic), and clayey (pelitic). The structural characteristics of the soil are reflected in its name. Therefore it is most important to know how to determine the size of soil-forming mineral particles and to identify the soil accordingly. This is not difficult in the case of soils with coarse and sandy structures (rock debris, gravel, coarse-grained sand, etc.). It is much more difficult to identify the aleurite, clay loam, and sandy loam, and these are the soils met with most frequently. As a practical aid, we may recommend the field soil classification compiled by V.V. Okhotin (1940) or Table XII based on this classification.

It is known that the morphology of sedimentary rocks depends largely on their texture. The textural characteristics are the lamination and the alternation of layers and lenses which differ in their granulometric and petrographic compositions, with accumulations of clay particles, mica scales and plant detritus on lamination planes. There are four main types of lamination: horizontal, wavelike, lenticular, and cross lamination (Fig. 9).

In field investigations it is essential to distinguish clearly between the primary structure and texture of sedimentary rocks (as this is done by geologists) and the cryogenic texture.

1. The study of frozen consolidated and semiconsolidated rocks*

The shape and distribution of ice inclusions in these types of rock correspond to the shape of fractures and voids of various origins. In granites, diabasis and similar rocks, the ice forms pockets and veinlets (Fig. 10). In bedded rock (limestone, argillite, marl, sandstone, etc.) the ice layers and lenses are found in fractures and voids on the bedding planes. Therefore the examination of fracturing of consolidated and semiconsolidated rocks is most important in the study of their cryogenic structure.

In hydrogeology a distinction is made between fracture and fracture-stratal (fracture-pore) water. Accordingly we may distinguish between fracture cryogenic textures found in consolidated rock and fracture-stratal (fracture-pore) cryogenic textures characteristic of sedimentary rocks (Table XIII).

For the purposes of a geocryological survey it is not always required to study the rocks in great detail. At times it is quite sufficient to mention the following points only:

a) the correct name of rock;

b) the structure of the entire rock body (homogeneous, bedded, compact, etc.);

c) the nature and extent of fracturing; the shape of fractures and voids and their distribution, whether or not all of them are filled with ice (and illustrate this with sketches), and whether there are any fresh fractures;

d) the extent of modification brought about by weathering (whether or not the rocks are leached, oxidized, etc.); and whether or not there are deposits of oxides, salts, etc., on the fracture planes;

e) the structure of ice inclusions and the orientation of limonite crusts, zeolite, calcite, air bubbles, etc.

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* Translator's note: Terms frequently used by Russian engineering geologists. Consolidated rocks (skal'nye porody) are hard rocks with rigid crystalline or elastic amorphic bonds. They include massive-crystalline extrusives, metamorphic and certain hard sedimentary rocks, such as quartz, sandstones, conglomerates, breccia with strong cement, quartzites, etc.

Semi-consolidated rocks (poluskal'nye porody) are rocks with elastic crystalline or amorphic bonds, and plastic colloidal bonds. When subjected to a load, they undergo elastic deformation up to a certain limit, beyond which they behave as unconsolidated rocks. Semiconsolidated rocks include marls, clay, shales, aleurites, mudstones, etc.

More attention should be paid to bedrock, where the latter forms the uppermost horizon of permafrost (to a depth of 10 - 12 m) and reveals considerable changes resulting from weathering and cryogenic processes within and outside the zones of tectonic disturbances.

2. <u>The study of perennially frozen Quaternary deposits</u>

There are several methods of studying Quaternary deposits. The most important are the lithological and the mineralogical methods described in appropriate handbooks.

A special study of the stratigraphy of Quaternary deposits is not one of the tasks of the geocryological survey. However, it should be noted that an expert investigating these deposits has the opportunity not only of collecting the fauna, gathering plant remains, or studying the boulders, but checking the validity of available stratigraphical data as well.

The study of permafrost is done by means of the permafrost-facies analysis. The basic principles of this method are as follows:

a) The perennially frozen Quaternary deposits differ in their ice content depending on their genesis and the facies to which they belong. The last two factors affect the local migration of water and its crystallization on freezing.

b) The shape, size and distribution of ice inclusions depend on the lithological characteristics of the deposits containing the ice (or rather the characteristics of facies, to which these deposits belong), and on the morphogenetic conditions prevailing at the time of freezing of the deposits, i.e. the element of relief where the freezing was taking place (flood plain, slope, depression at the foot of a terrace, etc.). The rocks of each facies display a specific cryogenic morphology, i.e. a specific cryogenic texture (Tables XIV and XV).

c) The ice inclusions determine the cryogenic morphology of soils and serve as a reliable genetic indicator of permafrost (flood plain, deluvial, oxbow lake, and lucustrine deposits, their facies and subfacies).

The permafrost-facies method renders it possible to investigate the characteristics of composition and cryogenic structure of perennially frozen Quaternary deposits, and to identify their facies and the main genetic types with the help of the following two groups of factors: 1) the grain size composition, bedding, fauna, plant remains, etc., and 2) the ice content, shape, size and distribution of ice inclusions. These two groups of factors are not simply superimposed one upon the other but are closely interrelated.

At present the classification of Quaternary deposits complied by E.V. Shantser is considered to be the best and is recommended for use in geocryological surveys (Table XVI).

It is relatively simple to distinguish between the genetic types of Quaternary deposits. It is, however, essential to have a fair knowledge of geomorphology and to know how to subdivide the relief into genetic elements which are closely related to various types of Quaternary deposits. It is also essential to know how to interpret correctly the genetic characteristics of the materials (grain size composition, sorting, bedding, plant remain, etc.) and how to use them for the reconstruction of environmental conditions at the time of deposition. In this respect the study of cryogenic textures of frozen soil also offers considerable help.

Below we give the characteristics of perenially frozen deposits of various origins.

a) Cryogenic characteristics of eluvial deposits. It is known that within the permafrost region the destruction of bedrock takes place more rapidly than outside this region. Because of the effect of freezing and thawing, the rocks here are subjected not only to a more intensive breakdown but also displacement and even mixing. Therefore we may refer to cryogenic eluvium as a variety of eluvium in general.

The composition and morphology of eluvial deposits depend on the petrographic characteristics of the parent rock. There are eluvial varieties with a predominance of boulder, rubble, and fine-grained (clay loam) material formed as a result of disintegration of consolidated and semiconsolidated rocks, both resistant and non-resistant to frost.

A description of eluvium should be started by noting its petrographic and grain size compositions, the proportion of fine-grained and coarse-grained material, the size and shape of rock fragments, and the extent of their weathering, roundness and sorting. Main attention should be given to the cryogenic texture of eluvium. It has been found by observations that eluvium displays three types of cryogenic textures*: crust-like (Fig. 11), basal (Fig. 12) and fissured-branching (see Table XIII).

During a geocryological survey one should attempt to study all characteristics of eluvial deposits which may indicate the conditions of their formation and freezing. Special attention should be paid to the shape of ice crusts which are less clearly defined in eluvial deposits than, for example, in the talus on steep slopes (see Table XIV).

When describing eluvium with a basal texture in which the ice is the main component and serves as an infilling material of a sort (basal cement), attention should be paid to the presence of mineral particles and their

* The terms given to cryogenic textures are provisory. Some of them will be made more precise in the course of future work.

aggregates "suspended" in the ice.

Eluvial deposits often include ice veins which at times extend to a depth of up to 6 m.

b) Cryogenic characteristics of talus deposits. Talus deposits are sometimes over 20 m in thickness. They contain a great deal of heterogeneous clay loam and sandy loam, often of the loess-type, with admixtures of waste, fragments and boulders of parent rock. It is the presence of these materials which usually helps to classify the formation as a talus deposit. In the absence of rock fragments and waste it is rather difficult to identify clay loams as a talus material. The overlying "mantle clay loams" are often regarded as problematic as far as their genesis is concerned*.

The perennially frozen talus deposits of the loess type display characteristic ice inclusions the study of which makes it a relatively simple matter to identify various genetic types.

These genetic types are given in Table XIV.

The clay loams and sandy loams in the talus deposits on relatively dry slopes are characterized by lenticular ice formations (Table XIV, sketches 1 and 2), while on swampy slopes they display reticulate cryogenic textures (Table XIV, sketch 3) and "bands", i.e. gently waving ice layers. The "bands" are a most important textural and genetic indicator. They are formed on the boundary of the layer subject to seasonal thawing due to the presence of water accumulated on the surface of the frozen material. The study of "bands" makes it possible to evaluate the changes in the morphology of slopes with time and to interpret the conditions prevailing at the time of formation on the talus deposits.

In some cases over 50% of talus on very swampy slopes consist of ice and represent a sort of ice breccia. It includes relatively thick (up to one metre) layers of texture-forming ice containing small lumps of clay loam, peat, rock waste and even small rock fragments. The ice layers have a banded texture (Table XIV, sketch 4) and consist of superimposed ice layers or bands up to 5 or 6 cm in thickness. Similar deposits representing a conglomerate of clay loam, peat, rubble and boulders cemented together by ice have been described by S.P. Kachurin (1950), who thought that ice in such deposits was due to solifluction.

As a rule, the ice veins in the talus deposits are small.

c) Cryogenic characteristics of alluvial deposits. Perennially frozen alluvial deposits form the terraces and the river valley floors, as well as

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* In this case the mantle clay loams represent a product of diagenesis, and of cryogenesis of frozen materials of various origins in particular. (Russian editor).

vast plains in Northern Siberia. In these regions formation and freezing of a major part of alluvial deposits took place simultaneously. As is known, the alluvium may be subdivided into the following facies: river-bed, oxbow lake, and flood-plain, each of which is further subdivided into subfacies. From the point of view of geocryology, the identification of genetic varieties of perennially frozen alluvium is extremely important, since the same soils (sand, gravel, clay loam, etc.) belonging to different facies often reveal different ice contents and cryogenic morphology. Below we give a brief description of alluvium which froze in the course of its formation, i.e. syngenetically.

River bed deposits may be represented by pebbly-gravelly, or sandy soils, and occasionally clay loam, which belong mainly to the two most common facies of the river bed alluvium: the facies adjacent to the river race and the facies of the river shallows.

The perennially frozen soils belonging to the facies of the river shallows sometimes contain pores not filled with ice, and this is true of gravel, sand and sandy loam. The sand and sandy loam often reveal a low ice content if they were deposited in the part of the shallows which remains above water and dries up by the time freezing sets in in the fall. Water contained in these soils crystallizes and forms ice grains distributed along the bedding planes, which serves as an indicator of the primary bedding of sedimentary soils (Fig. 13). This is the way the inherited* cryogenic textures are formed (Table XV, sketch 1).

Apart from such textures and voids not filled with ice, the morphology of these deposits is characterized by considerable sorting of material and primary, fine cross-bedding.

Facies of alluvium adjacent to the river race is represented by relatively poorly sorted deposits of sand and pebbles, and displays an irregular, lenticular cross-bedding of layers consisting of material of different grain size composition. Such deposits are characterized by a high ice content mainly due to the ice cement. Even pebble beds are at times oversaturated with ice. The silty sand contains irregular broken lenses of ice which give rise to a lenticular, cross-laminated cryogenic texture (Table XV, sketch 3).

Thus the cryogenic morphology of both facies mentioned above is different, even if their grain size composition is the same.

Oxbow lake deposits reveal a number of compositional characteristics: 1) the predominance of silty clay loam, sandy loam and fine-grained sand; 2) the presence of fine horizontal bedding; 3) the presence of single pebbles, shells of molluscs, or even pieces of wood; 4) the gleyzation of

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* The majority of cryogenic textures are superimposed (secondary) structures.

soil (revealed by bluish-grey colouration).

As a rule, perennially frozen oxbow lake deposits have a high ice content. The ice forms characteristic broken lenses from fractions of a millimetre to 2.5 - 3 cm in thickness, which are usually inclined and give rise to cross-lenticular (Fig. 14) or cross-laminated (Fig. 15) cryogenic textures (Table XV, sketches 3 and 4). Ice inclusions on bedding planes are found relatively seldom.

The facies of flood plain hollows represents a transition stage between the oxbow lake and the flood plain deposits. It is represented by clay loams and sandy loams containing humus, wood fragments and branches of shrubs. Together with plant remains brought in from elsewhere there are also grass blades and roots buried in situ.

The ice content of soils of this facies is high. The ice inclusions are represented by two varieties: a) broken cross-orientated lenses resembling those in the oxbow lake deposits, and b) horizontal layers formed at the lower boundary of the layer subject to seasonal thawing (Fig. 16). Thses ice inclusions alternate throughout the cross-section and give rise to mixed (feathery) cryogenic textures, which morphologically and genetically occupy and intermediate position between the cryogenic textures of oxbow lake and flood plain deposits (Table XV, sketch 5).

Perennially frozen flood plain deposits are very common. They sometimes form thick (up to 40 - 50 m) beds composed mainly of clay loam and sandy loam which may be gleyed, humic, or peaty to a greater or lesser extent. Occasionally they contain buried peat.

Most flood plain deposits are enriched by plant remains, mainly roots and blades of grass (cotton grass, sedge, etc.). The grass roots often interweave and form an intricate network ("felt") which results in a peculiar soil structure preventing the growth of ice inclusions.

The flood plain deposits may be subdivided into several genetic varieties. The most common of these are: a) the facies of the middle flood plain, and b) the facies of the high flood plain.

The facies of the middle flood plain is represented mainly by brownish-gray clay loams and sandy loams of the loess type often found in thin layers penetrated by thread-like roots of grass. The ice content of these soils is insignificant. The ice forms small lenses less than one millimetre thick, and against this background there are uniform ice layers 3 to 5 and occasionally 8 to 10 mm thick which give rise to a horizontal, parallel, laminated cryogenic texture (Table XV, sketch 6). Some varieties of these deposits (subfacies) are characterized by an ice lattice (reticulate texture) superimposed by horizontal parallel layers (Fig. 17).

The facies of the high flood plain is represented by deposits in small hollows (up to 50 m in diameter) known as concave polygons. The predominant deposits here are clay loams and sandy loams with various amounts of peat. The soils contain a large number of small (up to 1 mm) lenses of ice and continuous ice layers from 0.5 to 2.5 cm in thickness. Together with lenticular ice inclusions, these layers give rise to a concavo-parallel-laminated, lenticular cryogenic texture (Table XV, sketch 7).

There are two main subfacies among the high flood plain deposits:

1) The subfacies of flat, at times dry, polygons represented by clay loam of the loess type very similar to deposits of the middle flood plain. The peat content of the clay loam is not uniform which results in a variation of their ice content, owing mainly to the ice cement and small (fractions of a millimetre) ice lenses. Slightly concave ice layers are characteristic of this subfacies.

2) The subfacies of concave, constantly wet polygons represented by very peaty clay loams containing a large amount of plant remains. The latter (mainly remains of bog plants) accumulate in the form of "mounds" and "peaty pot-holes" with extensions of humus beneath them. There are also occasional occurrences of silty peat up to 3 m in thickness.

The ice content of these deposits is high. Apart from ice cement and a large number of small ice lenses, there are also clearly defined, continuous, strongly concave ice layers.

The flood plain deposits also include the wet meadow facies (which approximately corresponds to the facies of secondary water basins in Shantser's classification). This facies is represented by peat-free, weakly gleyed clay loams. sandy loams and some fine-grained silty sands. The characteristic feature of perennially frozen wet meadow deposits is a dense network of merged ice inclusions superimposed by both slightly and strongly concave ice layers or bands which give rise to concavo-parallel-laminated and reticulate cryogenic textures (Table XV, sketch 8).

It should be remembered that ice "bands" and layers were originally formed at the lower boundary of what was then the layer subject to seasonal thawing, which now forms part of permafrost. They reflect the roughness of the perennially frozen substratum within the limits of the flood plain. This explains the variations in their concavity.

The perennially frozen alluvial deposits contain a large number of syngenetic ice veins. The ice "bands" and layers are fused directly to these veins.

d) The cryogenic characteristic of alas deposits. Melting of buried ice in permafrost regions leads to formation of depressions ("alas" in Yakut).

Deposits formed during the development and disappearance of these depressions are known as alas deposits.

Alas deposits are formed under peculiar conditions from materials which were subject to thermokarst processes and redeposition. As far as their origin is concerned, they differ from all other geological formations. Therefore they should perhaps be regarded as a separate genetic type of deposits.

According to numerous investigators, the thermokarst depressions ranging from tens of metres to many kilometers in size are especially common in the central and northern regions of Siberia and to some extent in the North of the European part of the U.S.S.R. It is relatively simple to recognize these depressions by geomorphological methods and to single out the alas deposits found in association with them. The latter may be subdivided into the following genetic varieties.

1. The facies formed in the alas water basins represented by bluish-grey, silty clay loam and sand loam as well as silty sand. There are shells of fresh water molluscs, bits of wood, and lumps of peat giving these deposits a spotted appearance. Their cryogenic texture is very similar to that of oxbow lake deposits and is characterized by the presence of irregular, broken lenses and layers of ice (Table XV, sketches 3 and 4).

2. The facies formed on wet meadows and in "grassy creeks" represented by greenish light grey and dark grey clay loam and sandy loam free of peat. These deposits are characterized by reticulate cryogenic textures.

3. The swamp facies represented by peat and clay loam containing numerous remnants of plants which grew in the place of deposition. The peat formations reveal a fairly high ice content due to the presence of ice cement; ice inclusions in the form of irregular lenses and pockets are relatively rare. Peaty clay loam and sandy loam contain fine lenses (up to 1 mm) and layers (1 - 1.5 cm) of ice.

The alas facies may be further divided into subfacies. When investigating and describing these deposits, attention should be paid to cryogenic textures, which to date have not been adequately studied. The composition of the alas deposits varies with climatic and physico-geographical conditions. For example, the deposits formed in the alas water basins predominate in the central and southern parts of the Yakut A.S.S.R., while peat formations and peaty clay loams predominate in the northern areas. Ice veins are widely present in the alas deposits in the North. The soils accommodating these veins are characterized by concavo-parallel-laminated cryogenic textures. The alas deposits differ from alluvial formations mainly in the distribution of facies in cross-section and area.

e) Cryogenic characteristics of marine deposits. According to available information, the marine deposits in the West Siberian lowland are represented mainly by dark grey sandy clay and silt containing occasional boulders and remnants of marine molluscs. The cryogenic structure of perennially frozen marine deposits (Fig. 18) is characterized by a network of broken and usually inclined ice layers and lenses 2 to 10 cm and occasionally 20 to 30 cm thick. According to A.I. Popov and A.M. Pchelintsev, these ice inclusions form "an ice lattice thinning out in depth", which points to the epigenetic freezing of unconsolidated deposits.

3. Methods of studying frozen soils and the sequence to be followed in describing them

When studying permafrost and soils subject to seasonal freezing, it is essential to investigate all characteristics of their composition and structure which should be examined in relation to their origin and mode of formation.

The genetic approach to the study of composition and structure of frozen soils has been introduced fairly recently. Not all genetic types and cryogenic textures of perennially frozen Quaternary deposits have been identified and studied. The aforementioned data on the cryogenic structure of the more common types and varieties of deposits may be used in field investigations.

Quarternary deposits give rise to characteristic elements of relief. Therefore the first step is to study the geomorphology of terrain and to establish which elements of relief (flood plain, slope, alas floor, etc.) are formed by deposits of interest to us. The geomorphology of a given section is noted when describing a natural outcrop or an excavation.

A cross-section through permafrost is usually described by layers beginning with the uppermost layer. The layers with more or less homogeneous composition and cryogenic structure are noted.

The following features are recorded for each layer:

a) thickness;

b) type of soil, its compositon, which particle fractions predominate and which form the admixture, nature of occurrence and amount of animal and plant remains; in the case of formations derived from organic material (e.g. peat), notes are made on their relationship to the mineral part of the soil and their mode of occurrence;

c) colour of soil, humus content, gleyzation, iron content, salt content, etc.;

d) textural characteristics of soil: primary type of bedding, mottling, porosity, cavitation, etc.;

e) nature (shape, size) of ice inclusions and characteristics of cryogenic textures (see Tables XIII, XIV, and XV).

Main attention is paid to cryogenic textures. It is essential to distinguish clearly between uniform, continuous ice layers and lenses (Fig. 17) and broken layers and lenses (Fig. 16). It is necessary to note their thickness and mode of occurrence (horizontal, inclined, wavy, etc.), as well as other ice inclusions found in association with them (crusts, thin lenses, etc.). Special attention is paid to vein ice and its relation to cryogenic textures of accomodating soils.

It is desirable to complement the description of cryogenic textures with sketches and photographs. Photographs are made of well polished specimens (in winter) or of soils in situ (in summer). In the latter case the outcrop or part of it must be cleared of thawed soils and cut by a sharp spade or a kitchen knife. When the camera shutter is about to be released, the clean surface is wetted by pouring water over it, which clarifies its structure for photographing purposes.

The composition and cryogenic structure of perennially frozen Quaternary deposits in northern regions are best studied in natural outcrops found along the river banks, the lake shores and on the sea coast. It is advisable to examine them in summer when the soils cave in on gradual thawing making it possible to study fresh cross-sections without special clearing operations. Excavations in winter require a great deal of labour and do not give the desired results.

Permafrost in deep trenches and excavations is usually examined in winter or early spring (February - April). The cryogenic structure of permafrost is studied in drill cores. It is best to do so after a preliminary examination of similar soils in situ. Specimens for photographic purposes may be prepared from the core.

Descriptions of natural outcrops and excavations may prove very valuable for the study of stratigraphy of Quaternary deposits. For this purpose the studies must also include the analyses of pollen, spores and diatomes, examination of the fauna, and age determinations.

When studying the cryogenic structure of permafrost, it is essential to look for cavities, describe them and explain whether they form part of contemporary cryogenic texture or represent traces of former textures.

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Chapter IV. Field Investigation Methods in Geocryological Surveys

1. Sampling Methods

Geocryological investigations are carried out by testpits, trenches, cleared sections, and boreholes, which make it possible to study the composition, structure and properties of frozen soils. The type of excavation selected depends on the tasks facing the investigator, the environmental conditions, and the availability of equipment.

Testpits offer a good opportunity to study undisturbed permafrost as well as the soils subject to seasonal freezing, and to observe their textural characteristics which are often disturbed on drilling. However, the depth to which the soils can be observed in pits is limited and this is a serious drawback.

Trenches make it possible to investigate the morphology of cryogenic relief, the structure and distribution of large ice bodies, the structure of the layer subject to seasonal freezing, and that of the permafrost table over long distances but only at shallow depth. Clearing operations are usually carried out while examining natural outcrops.

Many properties of frozen ground may be studies by using one method, for example boreholes. They make it possible to investigate the distribution of temperature with depth, soil structure, etc., even if the frozen mass is hundreds of metres deep. They may be drilled under difficult conditions, for example under the sea (in shallow places), lakes, rivers and in other places where any other kind of investigation would be difficult if not impossible.

In geocryological investigations the major part of excavation and drilling is done in Quaternary materials, which change their properties not only on thawing but also when the temperature rises while still remaining below zero. When drilling test holes is essential to disturb the soil temperature as little as possible and to obtain frozen cores as required for the study of physical and physico-mechanical properties of undisturbed permafrost.

Mechanical drilling involving the use of water does not meet these requirements. Recently a new method of mechanical drilling of deep holes has been developed involving the use of compressed air for cleaning the holes. Experience shows that this method may be applied successfully in frozen ground and that such holes can be used throughout the geocryological investigations.

Testpits

To keep pits in wet ground as dry as possible, it is expedient to plan the main excavation work for the periods of the year when the layer subject to seasonal thawing is least saturated with water, or is frozen.

If field work is undertaken in summer, the excavation site should be surrounded by a ditch to drain surface and groundwater.

Frozen ground is excavated with picks, crowbars, wedges, pneumatic and electric drills, and explosives. In regional investigations pits and trenches are usually excavated by hand.

Frozen ground cannot be thawed by fire, hot stones or steam jets, since these considerably distort the natural state of the frozen rock and soil.

It is recommended to excavate pits in frozen ground which is sufficiently stable. The soils in the layer subject to seasonal thawing about 1.5 - 3 m thick may turn out to be unstable. If the depth of a pit does not exceed 3 - 5 m and will be filled again after investigations and sampling have been completed, it is not obligatory to brace the walls. To prevent a cave-in of soils in the layer subject to seasonal thawing, the mouth of the pit within this layer is made somewhat wider, or the walls are braced with temporary supports.

Deep pits which will have to be maintained for long periods of time must be supported at least in the upper part. It is best to use a continuous crown support and to fill the space around it with well-packed clayey material throughout the entire depth of the layer subject to seasonal thawing.

In the summer it is recommended to protect the pit from atmospheric precipitation and sunlight by means of an improvised cover made of any material, such as planks, fibreboard, sackcloth, etc. The walls should be disturbed as little as possible. Descriptions should be made and samples taken of the north facing wall.

In the cold seasona, when the danger of thawing is no longer present, a cover is no longer necessary and the studies are made of the south facing wall since it is better illuminated. At the end of the shift, the pit is covered by planks or logs, and a layer of dry moss or twigs.

The descriptions are made and the samples are taken as the pit gets deeper. All that was said above applied to a considerable extent to trenches as well.

Manual drilling

Manual drilling is often used in regional and engineering-geocryological investigations. In spite of the low efficiency of manual operations and the shallow depth of boreholes drilled in this way (up to 30 m), they may be used for comprehensive geocryological studies.

In comparison with mechanical drilling involving the use of water, manual drilling has certain advantages: 1) it makes it possible to obtain undisturbed samples of frozen soil by means of a core lifter, and 2) the temperature may be measured immediately after drilling.

Drilling operations in the permafrost region have a number of specific characteristics owing to the frozen state of the soils, increased water content of the soils in the layer subject to seasonal thawing, etc. The great strength of frozen ground makes drilling much more difficult. The bits and the methods used in drilling unfrozen soils are of little use in drilling frozen ground of the same lithological composition.

Frozen clays, clay loams, sandy loams, and sands without boulders or pebbles are best drilled by the rotary method. Use is made of spoon bits and core pipes, the cutting elements of which are mounted somewhat differently than in a standard drilling assembly. In places where fine-grained frozen soils alternate with pebbly horizons, or in the presence of boulders, the holes are drilled alternatively with conventional bits and spoon bits or bailers. Small and medium pebbles found in fine-grained frozen deposits are best drilled with core tubing up to one metre in length. Large pebbles are crushed without lifting the drill by hitting the pivoted collar mounted on rods with a metallic ram. In this case use is made of drill rods at least 42 mm in diameter and two pivoted collars mounted one on top of the other. The pivoted collar is hit as the drill is being rotated.

When drilling frozen soils it is essential to use special gear, e.g. the KP-1 corelifter designed by A.M. Pchelintsev (Pchelintsev, 1951). It consists of a bit, core tubing and an adapter for joining the bit to the drill rod (Fig. 29). The bit is a hollow steel cylinder with an outside diameter of 75 mm and an inside diameter of 63 mm. The lower edge of the bit has 8 pobedit* teeth. This corelifter makes it possible to obtain undisturbed frozen specimens 12 to 15 cm in height by lowering the drill only once.

A similar bit but with teeth not made of hard alloys is used in Canada (Potzger, 1955). Its internal diameter is also 63 mm (Fig. 30).

It is also possible to utilize standard pobedit bits used in core drilling. The worn out pobedit teeth are replaced by electric welding using T-600 electrodes (Fig. 31). Such modern bits have been used successfully for drilling frozen morainic clay loam and sandy loam with pebbles up to 8 - 10 cm in diameter. It is possible to use conventional bits by providing them with teeth of pobedit plates used for the core-lifter. The plates must

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* Translator's note: A tool alloy.

be joined to the head of the bit in such a way as to obtain a cutting angle of about 80° (Fig. 32).

The main difficulty lies in detaching the core from the face of the hole. This is done by way of the friction between the frozen core and the inside wall of the bit. However, if the strength of the frozen soil is very high, this frction is not sufficient, especially if the cross-section of the core is greater than 10 cm^2 (diameter greater than 60 mm). The lower the soil temperature, the greater the hardness and resistance to failure of frozen soil, and this is especially true of ice-saturated fine-grained soils with temperatures below $-5^{\circ}C$. Therefore the drilling of such soils with a core lifter and bits is slow and often impossible.

When drilling low temperature sandy loam, clay loam or sandy soil, use is made of a spoon bit the cutting part of which is reinforced with hard-alloy plates and consists of four wings (Fig. 33). Two of these wings are intended for cutting and crushing the soil along the periphery of the bore. On penetrating the soil, they form a core within the bore. The other two wings (collectors) undercut and crush the soil left behind by the cutting wings and move the crushed soil particles into the bit. On lifting the drill gear, these wings support the soil within the bore. On changing the size and location of collecting wings relative to the axial plane of the bore, it is possible to obtain a core 30 - 40 mm in diameter which can be detached in spite of the great strength of the soil.

Hard-alloy plates are mounted on all cutting edges of the bit. Two plates measuring 10 x 10 x 3 mm are inserted in each cutting wing of the spoon bit 3" in diameter, one behind the other at a distance of 20 - 25 mm. The collecting wings are supplied with one plate measuring 30 x 20 x 2.5 mm. All plates are welded to the bit head with copper or brass.

To obtain the core, the size of collecting wings is reduced and somewhat smaller plates are selected. The shape and size of plates are chosen from handbooks.

When drilling fine-grained soils without sand and with temperatures not lower than -2° or $-3^{\circ}C$, it is possible to use spoon bits made of steel pipes, with the cutting part similar to that described above but without reinforcing hard-alloy plates on the wings. A diagram of such a bit is shown in Figure 34. It can be produced in a workshop from hot rolled steel (type 45) tubing measuring 108 x 4.25 mm. The working end of the spoon is tempered after assembly. It is not recommended to make the bits from pipes with thicker walls, since this would reduce the penetration rate. These bits render it possible to obtain frozen soil cores good enough for laboratory investigations.

Use of coiled tubing is not recommended. It usually wedges in and this makes it impossible to obtain undisturbed soil samples.

The most convenient tools for field investigations are compound percussion rotary hand drills 2" and 3" in diameter. In some instances it is more convenient to use drills with larger diameters (4 1/2" and 6").

The drilling of frozen, ice-saturated sand, and especially that with a high content of quartz, results in considerable wearing of bits. The drill rig must contain a larger than usual number of spoon bits (4 to 5 bits of various diameters for the summer season) and at least 15 bits for drilling clay loam with pebbles.

When using spoon bits, core lifters and conventional bits, it is essential to equip the drilling rig with necessary adapters and tubing of various diameters (1 - 2 pipes of each diameter), as well as additional pivoted collars (at least three).

It is categorically forbidden to use bits heated in a fire or to pour hot water down the hole, since this completely alters the moisture (ice) content of the soil, its structure and the thermal regime of the hole. It is absolutely essential to enforce this rule at all times.

Manual drilling of frozen soils has its own characteristics, depending on the time of the year when the geocryological investigations are carried out. The drilling of holes in the warm season is at times more difficult owing to high water content of soils in the layer of seasonal thawing. On penetrating this layer, casing is lowered down the hole and anchored in the permafrost. In winter, casing is no longer necessary, providing the hole is not intended for long-term temperature observations. The casing must be anchored in the permafrost in such a way as to eliminate the possibility of the soil thawing around them as a result of heat transfer through the casing walls, otherwise water would flood the hole. The length of casing for complete penetration of the layer of seasonal thawing depends on its thickness and the temperature of the permafrost table. When the temperature is $-3^{\circ}C$ or lower and the layer of seasonal thawing does not exceed one metre in thickness, it is sufficient to extend the pipes into the permafrost to a depth of not more than one metre. When the soil temperatures is $-1^{\circ}C$ or close to $0^{\circ}C$, while the layer of seasonal thawing is over 2 m thick, the casing must be extended into the permafrost to a depth of at least 1.5 or 2 m.

Casing must also be provided for all thawed layers encountered in permafrost.

If the hole is stopped in thawed soil and must be kept dry, casing is provided for the entire length of the hole and is carefully plugged at the bottom. This is accomplished by layers of clay or viscous clay loam alternating with oil-impregnated fibre. The casing is pressed into this plug to a depth of a least 0.5 m. When the water content of the soil is high and use is made of a column of casing, the last section should have a welded bottom. When casing is complete, the upper column is pulled out.

Drilling in summer is made more difficult by adfreezing of the drill to the walls of the hole. Therefore it is essential not to leave the drill in the hole when not in use and to pull it out completely even during brief interruptions in drilling. It is recommended to examine a shallow hole periodically by means of sunlight directed into the hole by a small concave mirror. This makes it possible to see whether water is percolating from underneath the casing, which may lead to the formation of an ice plug.

Frozen drills are freed by pouring a small amount of saturated salt solution down the hole. This is done when the hole is cased.

A hole is maintained for systematic temperature observations as follows: a) by casing and plugging in an appropriate way, and b) by protecting the protruding end of the casing from heating and cooling. To satisfy the second requirement, the mouth of the casing is covered with a tight-fitting wooden plug or by a screw-on metal lid. The mouth of the hole is protected by a strong, covered, wooden box measuring at least 40 x 40 cm. The space between the walls of the box and the casing is filled with insulating material, such as slag, sawdust, moss or dry peat. The number of the hole, its depth, the year of drilling, and the name of the drilling firm are recorded on the box or the wooden plug. The same information should be recorded on the casing itself.

The casing is taken out by the same lifting methods and equipment as under normal conditions. To make this easier, the casing is heated by hot salty water or by inserting a hot drill immediately prior to lifting.

The depth of a hole depends on the task in hand. If it is required to find the depth of seasonal thawing and obtain data on the layer subject to seasonal thawing, it is sufficient to drill to a depth of 1.5 - 3 m, and occasionally 5 - 6 m. On surveying building sites, the holes are drilled to a depth of 10 - 15 m but when designing large and important structures the holes may reach a considerable depth. Mechanical drilling is used in this case. During regional geocryological investigations, when it is not possible to drill across the entire permafrost body the holes are drilled to just below the base of the layer of annual temperature fluctuations, i.e. to a depth of 15 - 30 m.

Additional equipment required for manual drilling

1.	3" KP-1 corelifter	1
2.	4" KP-1 corelifter	1
3.	4" spoon bit set to operate with four cutting edges	2
4.	4" spoon bit set to operate with four cutting edges and reinforced with pobedit plates	1
5.	3" spoon bit set to operate with four cutting edges and reinforced with pobedit plates	3
6.	3" spoon bit set to operate with four cutting edges but without pobedit plates	3
7.	92 mm bits reinforced with pobedit plates	3
8.	75 mm bits reinforced with plates	3
9.	92 mm bits with welded teeth of high-grade electrodes	10
10.	75 mm bits with welded teeth of high-grade electrodes	5
11.	Core barrel for 92 mm bits shortened to 1 m	2
12.	Core barrel for 75 mm bits shortened to 1 m	2
13.	Adapter for the 92 mm core barrel	1
14.	Adapter for the 75 mm core barrel	1
15.	Hinge clamp	3

Bits (1 - 10) are interchangeable and therefore their number must be selected with reference to the soil composition and temperature.

Mechanical drilling

Mechanical drilling in the permafrost region differs from that outside this region.

Drilling methods and equipment are determined by the following characteristics of permafrost: temperature, rapid and considerable change in strength on disturbing the thermal regime, presence of ice, occasional presence of mineralized water in the liquid phase, alternation of soil layers with positive and negative tempratures in the same geological cross-section, etc.

The following requirements must be satisfied when drilling in permafrost:

1) the temperature of the drilling solution must be close to that of the soil;
2) the drilling solution must be prevented from freezing in the hole during drilling and during unforseen interruptions in its circulation;
3) the casing must be protected from damage;
4) the mouth of the hole must be protected from the effects of the drilling solutions.

The ease with which frozen soils can be penetrated by a bit depends directly on their physico-mechanical properties, which in turn depend on the mineral composition, structure, water (ice) content, and the depth of occurrence of frozen soil.

The presence of ice in pores and fractures, as well as free ice in the form of lenses, make the soil more plastic. On drilling, it is necessary to combine high axial pressures on the face of the hole with high rotation rate of the drill. When using a drilling solution for cleaning the holes, it is found that the least stable soils are those with a low ice content and a coarse structure (coarse sand, pebbles and gravel), which have not been firmly cemented with ice.

Let us examine drilling with the help of hard-alloy bits and washing solutions.

The drilling solution freezes in the hole if there is a lengthy interruption in its circulation (3 - 10 hours).

If a hole is cleaned with a high-temperature solution, this may disturb the thermal regime and lead to thawing of soil, collapse of the walls, destruction of the core, and other difficulties. Frozen soil is destroyed during drilling owing to the effect of the washing solution, as well as a result of mechanical detachment of individual soil particles due to friction between the solution and the walls of the hole.

Water (both fresh and salty) or a clay solution may be used as a drilling solution depending on the time of the year, the temperature and the strength of the soil.

If frozen soil consists of hard varieties (sandstone, limestone, compact clay shale, hard argillite, etc.), it is possible to use water with a positive temperature, since the hole will not be damaged to any considerable extent by a slight heating of such soils.

If drilling is done in a frozen sedimentary mass without inclusions of dry loose soils or soils saturated with highly mineralized water, the holes may be washed with salty water with a negative temperature.

Table salt is added to water to prevent it from freezing. The amount of salt used (P_s) depends on the temperature of the soil (Table XXI) and is calculated by means of the following formula:

$$P_s = \frac{n \cdot Q_w}{100 - n}$$

where Q_w is weight of water in kg;

n is salt concentration in %.

Example. It is required to obtain an aqueous salt solution with a freezing point of -4.4^oC. From Table XXI we find that such a solution must have a concentration of 7% (n = 7). To obtain this concentration, we must add the following amount of salt to each 100 kg (100 litres) of water:

$$\frac{7 \cdot 100}{100 - 7} = 7.5 \text{ kg or } 75 \text{ kg/m}^3.$$

It should be taken into consideration that a drilling solution containing a considerable amount of salt destroys the ice which cements the soil particles. It is known that the destructive effect of salt increases with rising concentration of the solution (Fig. 35).

On drilling frozen sedimentary soils or those cooled to a temperature below 0°C (e.g. dry loose soils, or unstable and unconsolidated deposits containing highly mineralized water) it is best to use the clay solution only.

Data for the clay solution are given in Table XXII.

The temperature of the clay solution must be negative and if possible below that of the soil. The solution is cooled in summer in special holes drilled in frozen soils to 2 - 3 m below the layer of seasonal thawing.

To retain the clay solution in operating condition (i.e. to retain its negative temperature) and to protect it from freezing, an optimum amount of table salt is added to it, which depends on the type of clay used. The required amount of salt is determined as follows.

Example. It is required to obtain a clay solution with a freezing-point of -4.4°C. From Table XXI we find that the salt concentration of such a solution must be 7%. Therefore the following amount of dry salt must be added to one cubic metre of clay solution, the specific weight of which is 1.2 gm/cm³:

$$P_s = \frac{n \cdot Q_1}{100-n} = \frac{7 \cdot 1200}{100-7} = 90.3 \text{ kg}$$

where n = 7% - salt concentration,

Q_1 = 1200 kg - weight of one cubic metre of clay solution.

Salt can be added to the clay solution only in the form of brine. The amount of water required to prepare the latter is calculated as follows:

$$Q_2 = \frac{P_s}{0.36} = \frac{90.3}{0.36} = 251 \text{ litres}$$

where 0.36 gm is the amount of salt required to saturate one litre of water.

To retain the salt concentration at 7%, it is necessary to add more salt (owing to the addition of 251 litres of water). This extra amount of salt is calculated as follows:

$$P_{s_1} = \frac{n \cdot Q_2}{100-n} = \frac{7.251}{93} \approx 19 \text{ kg}.$$

The total amount of salt P which must be added to obtain the required concentration of the washing solution is as follows:

$$P = P_s + P_{s_1} = 90.3 + 19 = 109.3 \text{ kg}.$$

The total amount of water Q required to obtain a saturated salt solution is equal to:

$$Q = \frac{P}{0.36} = \frac{109.3}{0.36} = 304 \text{ litres}.$$

Therefore to obtain a concentration of 7% it is necessary to take 109.3 kg of dry table salt for each cubic metre of clay solution with specific weight of 1.2 gm/cm^3 and dissolve the salt in 304 litres of water. The saturated salt solution is gradually added to the clay solution with constant stirring.

The choice of the design of hard-alloy bits depends on structural and abrasive properties of the given soil. It is essential to select bits which will ensure a maximum drilling rate.

When drilling low-strength soils with increased ice content, and especially when "compression" of the pump is observed, it is recommended to use bits with ribs, as well as bits with a small number of cutters (6 - 9) protruding above the head and the sides of the bit.

To reduce the resistance to circulation of the drilling solution, the bit rings must be provided with large slits (coolers). When drilling abrasive soils, it is best to use self-sharpening bits, while in strongly fractured rock it is expedient to use bits reinforced with large cutters which do not protrude too much above the head and the sides of the bit.

The handbook of increased estimated drilling rate standards for geological survey work (1954) contains hardness categories of frozen soils. For example, pure ice is in category III; frozen fine and medium sand, silt and peat are in category IV; frozen coarse sand, gravel, compacted silt and sandy clay in category V; frozen compacted clay cemented with clayey material, and pebbles with ice layers are in category VI. The following drilling rates have been accepted for these categories:

Soil category	III	IV	V	VI
Clean drilling, m/hr	2.3	1.6	1.1	0.75

The specific types and diameters of bits (Table XXIII) are chosen in relation to soil type and design of the hole.

Table XXIV contains the parameters of an efficient drilling regime. To determine the total load on the bit, the load on one cutter is multiplied by the number of cutters in the bit. The rotational speeds of the drilling rig stem are reduced to the peripheral speeds of the bit (m/sec). The total consumption of drilling solution is found by multiplying the specific consumption of the solution per centimetre of bit diameter by the outer diameter (in cm) of the selected bit. To protect the soil around the mouth of the hole from the effect of the washing solution, it is recommended to start drilling without using solution to a depth of 10 - 15 m. The drilled section of the hole is then reinforced with casing. A T-piece is attached to the upper end of the pipe to serve as a runoff for the solution. All further drilling can be done by using the solution for cleaning the hole.

The difficulty of maintaining the temperature of the solution close to that of the soil, collapse of the hole walls, destruction of the core resulting from the disturbance of the thermal regime of the soil, unhealthy working conditions for the drill crew (especially in winter), all makes it necessary to find such methods of cleaning the holes which will not change the low temperature properties of the soil and destroy the core and the walls of the hole.

In recent years the Yakut geological administration used diesel oil to clean the holes. However, this type of cleaning solution turned out to be unacceptable because large volumes of it were absorbed by the hole. Furthermore various rubber components of the pumps were severely damaged by oil. A much better effect has been achieved by replacing the liquid cleaning solution with compressed air*.

Let us now examine this drilling method.

With compressed air, the mechanical rate of drilling increases by a factor of 4 to 6 in hard soils and 2 to 3 in softer soils, as compared with rates achieved on using drilling solutions. This is due to a reduction in hydrostatic pressure on the face of the hole, more efficient removal of slurry, etc. This drilling method eliminates the repeated crushing of soil particles on the face and the slipping of the bit along the face in the absence of lubrication, the role of which was filled by the washing solution. As a result of all this and also due to a better cooling of bits by a more powerful stream of previously compressed air, the life of the bits increases by factors of 1.5 to 10 and in some cases even more. The core is obtained more rapidly and its quality is improved. Furthermore, the core is not contaminated or damaged by water. This simplifies and improves the quantitative and qualitative analyses of soils at a given depth.

Compressed air drilling has the following advantages over other drilling methods:

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* Compressed air was first used for removing the slurry from a hole in 1918 in the Borislav oil field.

1) the costs of supplying water and preparing salt or clay solutions are eliminated;

2) the breakdowns resulting from freezing of the solution are also eliminated;

3) the conditions for studying the temperature of permafrost are improved, since there are practically no variations in the temperature regime of the soil;

4) the working conditions of the drilling crews are greatly improved.

Apart from the aforementioned advantages, this drilling method has a number of disadvantages which limit its use. The greatest difficulties are encountered on drilling water-saturated horizons (especially those containing water under high pressure), as well as plastic, sticky, ice-saturated, or loose-soil horizons.

Compressed air drilling can be done with the followings rigs: ZIF-1200A, ZIF-650A, ZIF-300, V-3, KAM-500, KA-2M-300, URB-ZAM, and SBU-150ZIV.

This drilling method has been described by A.V. Maramzin (1958), B.S. Filatov (1958), and I.P. Elmanov (1958).

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2. Geophysical Investigations

In contrast to thawed soils, the physical properties of frozen ground are greatly influenced by temperature and ice content. The present section deals with those methods of geophysical investigations without which it is difficult to carry out geocryological surveys in the permafrost region.

There is a sharp difference in the values of the true specific electric resistivity ρ and the dielectric constant ε of frozen and thawed soils, and in the rates of propagation of longitudinal waves in them. Of the commonly accepted geophysical methods, the following are justifiable from the point of view of physics for the study of frozen soil formations: geothermal, direct current, unimetric, seismic and acoustical methods. There are practically no differences between the gravitational and magnetic fields of frozen and thawed soils. Let us examine the aforementioned methods more closely.

a) Seismic methods (using refracted and reflected waves) have been repeatedly applied to geocryological investigations in the U.S.S.R. and the U.S.A. and proved quite successful, but for various reasons have not been used in practice (e.g. because of lack of portable equipment for investigations at shallow depths, certain difficulties in interpreting the data, and a relatively high cost of investigations).

b) Unimetric methods have also been repeatedly tested in the field but the insufficient depth of penetration of electromagnetic waves and the presence of various factors distorting the results have prevented a further development of this method. However, the present level of development of radioelectronics, and radar in particular, permits us to assume that the unimetric methods and especially a modification of "X-raying" from the holes will find some application in geocryology.

c) The radar method may prove useful in solving the problems of delineating various taliks, determining the thickness of the frozen (ice saturated) layer around freezing boreholes, locating polygonal earth wedges and voids in them, etc.

d) The natural field technique is used in the U.S.A. to locate the corrosion of pipes and cables in permafrost (Bull. Geol. Soc. America, 1956).

e) Electromagnetic logging and the ultra-acoustical method are at present in the development stage. The first may be useful in differentiating between layers with very similar physical characteristics, determining the depth of the lower surface of the permafrost, and locating the water-bearing horizons below the permafrost. It will probably be possible to use the ultra-acoustical field method to find the shape and the depth of thaw basins beneath buildings, the depth of the permafrost table, the thickness of the layer of seasonal freezing (in winter), the thickness of permafrost, the depth of taliks, the thickness of glaciers, etc. Ultra-acoustical logging can be used to determine the relative ice content of frozen soils, locate the boundaries of lithological soil varieties, and measure the elastic soil constants in the field.

f) Electrometric (resistivity) and thermometric methods are the most widely used techniques in the study of permafrost.

Depending on the temperature, the water (ice) content of frozen soils undergoes quantitative changes in horizontal and vertical directions. In a general case, there is a deviation from a linear relationhip between resistivity and temperature owing to an uneven distribution of ice in the soil. It is known that the amount of unfrozen water in the soil is a decreasing function of cooling.

It has been established that different soils have different regions of phase changes. Clays and clay loams, in which water begins to freeze at a temperature below $-0.2^{\circ}C$, may be frozen under natural conditions at a temperature between $0^{\circ}C$ and $-0.2^{\circ}C$. The presence of ice cement and ice layers in such soils in this temperature interval is due to the difference between freezing and thawing rates, i.e. with a temperature rise ice melts more slowly than it forms when the temperature drops. Therefore residual phenomena, a peculiar "cryogenic hysteresis", may be observed both during freezing and melting.

Sharp differences in physical parameters near the boundary between frozen and thawed soils may be observed already at the beginning of pronounced phase transformations. Because of this, the true specific resistivity of the permafrost table differs fairly sharply from that of the overlying thawed soils. This differentiation is less clear near the lower permafrost surface, where it is evidently dependent on the lithological composition and the water content of the surrounding soil. In frozen clay and clay loam with temperatures ranging from 0° to $-1^{\circ}C$, the resistivities remain sufficiently heterogeneous, and therefore it is possible to attempt a determination of the depth of the lower permafrost surface on the basis of VEL curves. This is possible in the case where the transition from frozen to thawed soils takes place in the absence of sharp differences in their lithological composition.

The aforementioned is confirmed by the log of hole 19 located near the southern boundary of the permafrost region (Fig. 36). On drilling it was not possible to determine the physical state of the dense heavy clay loam near the lower surface of high-temperature permafrost and it was assumed that the thickness of the frozen formation exceeded the depth of the hole. Temperature measurements did not help either since water was encountered at the depth of 7 m. The temperature log supplied the information on the physical state of the clay loam to a depth of 7 m (Fig. 36, curve 4). The location of the lower boundary of permafrost was accurately determined by means of electric logging.

Curve 3 (Fig. 36) shows that the current remains practically unchanged (about 6 ma) to a depth of 8 m and only at a depth of 8.75 m does it rise sharply to 40 ma. At the same depth, the apparent specific resistivity drops sharply from $\rho_a > 30{,}000$ ohm.m (curve 1) and $\rho_a > 5{,}000$ ohm·m (curve 2) to $\rho_a = 8 - 10$ ohm·m. The point of divergence of the curves denoting the current (curve 3) and the apparent resistivities (curve 2) indicates the depth of the lower permafrost surface (7.75 m).

The given diagram is a clear example of the fact that electric logging can be used to determine the depth of the lower permafrost surface.

It is known that the point of the phase equilibrium ice-water is not stable in the region of negative soil temperatures. It is essential to bear this in mind during vertical electrical logging for control purposes and when interpreting the electrical survey data.

The differences in the lithological composition of frozen soils have a noticeable effect on their resistivity, especially in the temperature range from 0°C to -2°C. Coarse-grained frozen soils containing mainly gravitational water even at temperatures close to 0°C have the highest resistivity. Fine-grained soils containing a large amount of bound water which does not freeze at very low temperatures have a low resistivity. The resistivity of soils of intermediate grain size composition occupies the middle part of a wide range of resistivities of frozen soils.

Frozen bedrock (argillite, siltstone, clay shale) has a lower resistivity than unconsolidated soils. At temperatures below -2°C, the lithological differences do not result in a significant differentiation of soil resistivities.

The relatively low resistivities (200 - 1000 ohm·m) indicate that frozen soils not only contain liquid water but that there are continuous, current conducting paths in the lattice system of the ice skeleton.

Frozen soils differ from thawed ground in that the current conducting paths in them are longer due to the presence of ice partitions, while the electrolyte has a different electric conducting paths in them are longer due to the presence of ice partitions, while the electrolyte has a different electric conductivity as a result of changes in the quantity and concentration of salts in soil solutions, temperature and pressure.

Hence, in a general case, the resistivity of soil depends on its mineral composition, grain size, chemical composition, concentration of solutions and the extent to which these saturate the soil, pressure and temperature. The latter is the deciding factor in the permafrost region. In soils with a natural water content, the resistivity is an increasing function of their cooling.

The resistivity of frozen soils varies over a wide range of values. The best studied resistivities are those of frozen unconsolidated deposits. In the Soviet Far East, the resistivities of frozen overburden (mainly sand) vary between 35,000 and 1,500,000 ohm·m; in Central Yakut Region the resistivities range from tens of ohm·m (in soils with a high salt content) to 18,000 ohm·m and more; in the Far North (the Yana-Indigirka lowland), from 4,000 to 40,000 ohm·m and more. In Southern Yakut Region the resistivity of bedrock varies between 1,000 and tens of thousands of ohm·m; in the European North and in Northwestern Siberia, between 200 and 8,000 ohm·m. In some northern coastal regions, soils containing sea water cooled to below $0^{o}C$ have resistivities of several ohm·m, or a fraction of one ohm·m (Fig. 37). The interpretation of the curve in Figure 37 gave the following information: the thickness of the first frozen layer was 6 m, which has been confirmed by drilling; the thickness of the underlying layer with a negative temperature was over 13.3 m.

Different combinations of temperature and lithological composition of soils result in a wide range of resistivities, but in spite of this the changes follow a regular pattern. For example, in unconsolidated deposits the resistivity is highest in coarse-grained soils (sand with gravel and pebbles) and lowest in fine-grained deposits (heavy clay loam and clay). The characteristic feature of bedrock is that the resistivity is highest in dense coarse-grained rocks (conglomerates, sandstones, crystalline rocks) and lowest in fine-grained varieties (argillites, siltstones, clay shales). The highest resistivities are found in frozen low-temperature conglomerates, sandstones and sand. The lowest resistivities are characteristic of frozen high-temperature argillites, siltstones, shales, and clays.

Apart from a distinct regional change in the resistivity of frozen soils, there is also a change with depth. Where the frozen mass does not contain large accumulations of subterranean ice and ice layers, its resistivity decreases with depth depending on the normal increase in temperature.

Soils with the highest ice content are found in the upper horizon of the frozen formation (from its upper surface to the base of the layer of seasonal temperature fluctuations). Below this layer the ice content decreases and often does not change much with depth. Therefore, on interpreting the curves obtained by vertical electrical logging in regions where the temperature of the frozen soil mass varies between $0^{o}C$ and $-2^{o}C$, the soil is divided into two horizons: the first with a resistivity $\rho_2' > 1{,}000$ ohm·m, characteristic of the layer of seasonal temperature fluctuations h_2';

the second with $\rho_2'' < 1{,}000$ ohm.m, where $h_2'' = H - h_2^2$; H is the thickness of the entire frozen soil mass.

The interpretation of the upper curve (Fig. 38) makes it possible to determine the thickness of the layer of seasonal thawing, which in this case is 1.7 m. The thickness of the first horizon of the frozen mass with the highest ice content is 15 m; the thickness of the second horizon is 85 m. The total thickness of the frozen mass reaches 100 m.

The separation of the frozen mass into two horizons makes it easier to interpret the curves. The determination of the first horizon is of great practical importance, since it is often saturated with ice. Its thickness is equal to the thickness of the layer of annual temperature fluctuations in the period of formation of the frozen soil mass. This layer is characterized by high values of ρ_a at the maximum of VEL curves obtained with small separation of current electrodes.

It is interesting that the maximum values of ρ_a are found not only in low-lying areas with thick unconsolidated deposits, but also in mountainous regions where bedrock consisting of metamorphic, sedimentary, or crystalline rocks is very close to the surface. Figures 38 and 39 show curves, one of which refers to a frozen mass consisting of unconsolidated soils and the other to frozen consolidated ground. In both cases the thickness of the frozen mass is approximately the same. A separation of the frozen mass into two horizons on the basis of the aforementioned curves proves useful in regions where the soil temperature ranges from 0°C to -2°C and the ratio $\frac{h_2'}{H} \geqslant 0.1$ is valid.

In northern regions, where the temperature of the frozen soil is below -2°C (sometimes -9°C or even -11°C), the electro-geocryological cross-section becomes much more complex. The first horizon is formed by the layer of seasonal thawing. Its thickness reaches 1.5 - 3 m and its resistivity is several hundred ohm·m. The second horizon is 2 to 3 m thick. It is represented by frozen unconsolidated deposits and has been referred to as the intermediate horizon by B.S. Yakupov (1960). The soil temperature in this horizon changes with depth from 0°C to -2°C, but the resistivity increases. The third horizon in frozen overburden has a temperature below -2°C and a resistivity from several thousand to several hundred thousands ohm·m. The fourth horizon corresponding to frozen bedrock has a much lower resistivity (3,000 - 5,000 ohm·m). Next comes thawed bedrock with a resistivity which ranges from 100 to 1,000 ohm·m. At times the resistivity of frozen soils reaches such values for which there are no theoretical master curves. Therefore V.S. Yakupov suggests that in order to find the thickness of frozen overburden, a four-layer cross-section should be replaced by an equivalent three-layer cross-section with resistivity ratios (moduli) available in handbooks of conventional three-layer master curves.

Figure 40 shows one of the curves obtained by vertical electrical logging on the Kolyma River. Its maximum indicates the resistivity of the uppermost permafrost horizon which has the highest ice content and includes the second intermediate horizon and part of the third (the zone of weathered bedrock).

In the Verkhoyansk-Kolyma mountain folding region, the resistivity of frozen unconsolidated deposits exceeds that of underlying frozen bedrock by factors ranging from 10 to 100. In Central Yakut region, the resistivities of frozen Quaternary deposits and the underlying bedrock are almost the same (Figs. 41 and 42). The maximum resistivity ρ_a on the curve in Figure 41 has a smaller abscissa than ρ_a in Figure 39, although the frozen soil mass in the first example is thicker. The value and the location of the maximum in Figure 41 are determined by the resistivity of the frozen unconsolidated deposits. In this case the resistivities depend on the changes in the soil temperature with depth. Therefore the value and the location of the maximum on the curve in Figure 41 may be used to find the thickness of the permafrost. Such curves may be interpreted as three-layer curves without an artificial division of the frozen mass into separate horizons.

The changes in resistivity in a horizontal direction in individual regions (local variations), with all other conditions remaining the same (similar lithology, absence of centres of chemical activity involving emission or absorption of heat, absence of distortions owing to hydrogeological conditions, etc.), are determined mainly by the heterogeneity of the surface relief and vegetation cover.

The zonal changes in resistivity follow a regular pattern: generally speaking the resistivity increases from south to north (which is due mainly to changes in the climate) and on passing from mountain regions to lowlands and plains.

The extent to which the resistivities change in latitudinal and meridianol directions differs at different points in the Soviet North. For example, in the case of the Bol'shezemel'skaya tundra, noticeable changes occur over a distance of 40 km from north to south and of more than 300 km from east to west.

There is little unfrozen water in the upper horizon of a low-temperature frozen soil mass and therefore it has the highest resistivity ($\rho > 1{,}000$ ohm·m). High-temperature permafrost contains a larger amount of unfrozen water and its resistivity is relatively low (400 - 1,000 ohm·m). When the layer of seasonal freezing does not merge with permafrost, the unfrozen water content is higher still. Therefore the resistivity of soils in this case rarely exceeds 400 ohm·m).

The same resistivity may be found in low temperature frozen soils if they are fine-grained and do not contain large accumulations of ice.

Among unconsolidated sedimentary deposits, the highest resistivities are found in frozen sand (especially in low-temperature frozen sand), as well as unfrozen sand with a low water content. In some cases the resistivity of unfrozen sand with a low water content may exceed that of frozen sandy loam, clayey loam, and clay. The fact that sand is found in association with definite forms of relief (river valleys) makes it possible to identify it without fail from the resistivity curves.

Thawed sands and horizons containing gravel, pebbles and boulders often contain water and therefore have practically the same resistivities as other types of Quaternary deposits. Only thawed but dry alluvial sands found on the slopes of river valleys have a high resistivity. The resistivities of thawed clay, clay loam, and sandy loam resting on permafrost are usually below 100 ohm·m.

Let us examine the methods of geophysical field investigations.

1. For a correct organization of electrical prospecting in permafrost regions, it is essential to consider the geocryology of a given area. A geophysicist must be able to recognize the relationships between individual landscape elements (particularly relief and vegetation) and permafrost. The success of electrical prospecting depends on the correct choice of measuring points, which can be made only by considering the effect of relief, microrelief and vegetation on the distribution of permafrost.

2. It is essential to consider the local climate. For example, in many northern areas seasonal thawing of soil extends to the normal depth of installation of ground electrodes only in the second half of June, while seasonal freezing sets in usually in early September. However, considerable deviations from this time table can occur in some years. Hence the geophysical field season is three, at most three and a half months. The best way to prolong the field season is to extend the work into the fall. In the first 10 or 20 days of seasonal freezing (October), movement across swamps becomes much easier, while the electrical contacts between the electrodes and the soil remain good.

Electrical prospecting should be started when the depth of seasonal thawing reaches the depth of installation of ground electrodes. Another reason for the late start of field work in northern regions is the desire to simplify the electrical prospecting on sections where permafrost occurs at great depths. When seasonal thawing is complete, a five-layer electrogeocryological cross-section of the type $\rho_1 < \rho_1' > \rho_1'' < \rho_2 > \rho_3$ is simplified and reduced to a three-layer cross-section of the type $\rho_1 < \rho_2 > \rho_3$ (where ρ_1 is the resistivity of soils in the layer of seasonal thawing; ρ_1' is the resistivity of frozen soils in the pereletok; $\rho_1'' = \rho_1$ is the resistivity of unfrozen soils between the pereletok and permafrost; ρ_2 is the resistivity of permafrost; ρ_3 is the resistivity of thawed soils beneath permafrost).

The field working season in the forest zone is somewhat longer than in the tundra.

The geophysical work should be carried out by considering the meteorological characteristics of each month of the field season.

In changeable weather, when rain is periodically interrupted by clear periods, work can go on without interruption. Electrical sounding can be done under a tent or any other cover, while electrical profiling may be carried out with the help of appropriate cover for the instruments, for example canvas with an opening on top protected by a plexiglas plate and water-insulating material. The size of the canvas must be such as to enable the operator to perform the necessary measurements. It is also possible to use special sleeves sewn into the canvas cover.

It is not recommended to leave reels with wet wire in the field for long periods of time, or overnight when there is a danger of night frost. In swampy and generally damp places, it is expedient to use steel-copper wire with a vinyl chloride insulation (PVR-0.35 or PVR-0.26). It should be taken into consideration that insulation on this wire cracks at low temperatures and loses its insulating properties.

In winter electrical prospecting is very difficult and expensive: it is possible to carry out only deep soundings (with the distance between current electrodes $AB > 6{,}000$ m) with the help of heated and movable huts, small power stations, oscillographic records, tractors, and trailers on skis.

3. When organizing a geophysical party, it is essential to consider the availability of different means of transportation which would make it possible to move quickly throughout a given area.

When working in the tundra, it is essential to have an adequate supply of sighting rods and survey pegs.

To save weight and ensure maximum mobility of the crew, the geophysical field batteries with dry elements of the B-72 type should be replaced by cold-resistant or anode batteries. Prior to field work, the outlets from the batteries must be connected to a panel with a change-over switch for the most commonly used voltages.

The difficulties associated with the field work as well as a wide range of changes in voltage differences Δv and current I make it advisable to use the electronic potentiometers of the EKS-1 or KSR type.

Since small electrode separations AB result in small currents, the shunt with a resistance of 0.01 ohm in the switch of an old potentiometer should be replaced by a shunt with a resistance of one ohm. This makes it possible to measure the current not only in linear units (centimetres) but also in milliamperes.

4. In the first months of field work and in the fall, the equipment and power supply should be put on a rubber mat placed on pegs, rods or twigs. A more permanent wooden floor is required when working in swampy places. The batteries for the accumulators or dry elements cannot be left in the field without a reliable protection from the rain. Moss cannot be used as a protection either above or below because of its hydrophylic nature.

Large electrode separations (MN >100 m) result in the appearance of large non-stationary natural electric fields which are absent when the separations in the circuit are smaller. These fields are due to the presence of earth currents which are especially strong at high latitudes. They are strongest in the evening and at night and therefore in the absence of a pulsator geophysical investigations should be carried out in the morning or during the day. The effect of these currents must be determined in situ.

The investigations are hampered by strong and unstable electromotive forces arising from the interaction between the electrodes and soil solutions. The following steps should be taken to reduce the polarization phenomena in the receiving circuit: 1) use electrodes made of red, chemically pure copper; 2) prior to measurements wipe the electrodes until absolutely dry; 3) reduce the contact surface of the electrodes or apply the principle of duality, i.e. occasionally change the output and input circuits.

The true electric resistivity ρ, dielectric permeability ε, and rate of propagation of longitudinal waves v_e for frozen and thawed soils should be measured in excavated sections, in shallow pits, on outcrops, spot medallions and in places where solifluction has occurred. The dimensions of experimental sites and pits must exceed the maximum separations of electrodes.

The parameters ρ, ε and v_e of frozen and thawed soils must be measured for all the most commonly occurring soils and all main types of landscape: It is expedient to make small probes for the measurement of ρ with three or four electrode separations ranging from $\frac{AB}{2} = 0.3$ m to $\frac{AB}{2} = 1.2$ m.

In summer the resistivity of thawed soils in the active layer usually decreases, while the dielectric permeability and the velocity of propagation of elastic longitudinal waves increase from top to bottom. Therefore when taking measurements in two or three pits, it is better to carry out several soundings which are repeated as the pit gets deeper (to a depth of 0.5 - 0.7 m).

To reduce the volume of excavation, it is more convenient to investigate permafrost on sections with a thick moss and peat cover, where frozen soils are usually found at shallow depths.

The transition from qualitative estimates of the depth of the permafrost table based on equal resistivity maps to quantitative estimates is accomplished by using drilling, ultra-acoustical and VEL data, a minimum amount of which must be obtained on sections intended for investigation by the electrical profiling method.

It is difficult to obtain undistorted VEL curves because of the horizontal heterogeneity of permafrost. At times these curves are so distorted by the screening and streamlining effect that half of them cannot be used for quantitative interpretations.

To reduce the number of unusable curves, the centre of the logging system must be located in the middle of a selected uniform section of relief, the contour radius of which must exceed 80 - 100 m. The periphery closest to the centre must be uniform and located, if possible, within a radius equal to the half-separation of current electrodes which gives a maximum on the VEL curves.

When the soil cover and meso-relief are very varied and complex, i.e. in conditions where it is impossible to satisfy the aforementioned requirements, one should strive to obtain an adequate VEL curve by using an arrangement with one current electrode in infinity. In this case, the logging should be orientated in the direction of least distortions caused by relief and heterogeneity of permafrost. Under difficult environmental conditions, three or four additional loggings across the main direction of logging are unavoidable. Electrical logging carried out near the boundaries of sections where the layer of seasonal freezing reaches or does not reach permafrost, as well in the vicinity of deep and open taliks, results in curves which cannot be interpreted.

To avoid errors it is absolutely essential to follow the aforementioned rigid requirements concerning the selection of points in a given area. If it is impossible to obtain a completely undistorted curve, one should strive to measure as precisely as possible the lower part of the descending branch of the curve, the inflexion point on transition to the right asymptote and finally the asymptote itself, which should be determined from three points.

The field notebook must contain a thorough description of relief, vegetation and hydrography, at least within a radius equal to a half-separation of current electrodes. Furthermore, there should be a description of the section closest to the electrodes which when moved produced distortions on the VEL curve.

A profile of the earth's surface showing the types of landscape is sketched for every geophysical profile. The characteristics of the landscape are given for the points of installation of geophysical pickets. The profiles of the surface are sketched on a relative scale along the horizontal and on an approximate scale along the vertical. Figure 43 contains an example of how to shoe the characteristics of the landscape.

The electrical survey crews conducting long-term investigations of large geological structures over a wide area must include in their plan of work investigations of geocryological conditions. The latter should incorporate a small number of electrical profiles for all main types of local landscape, two or three vertical electrical logs, and two manually drilled holes up to 8 - 10 m in depth with logs of temperature T and apparent specific electric resistivity ρ_a.

A series of these investigations is carried out every 50 km along the main traverse if it runs along a meridian. This makes it possible to obtain a general idea of geocryological conditions in the area of future investigations. On going from east to west, the investigations are conducted every 100 or 150 km. Electrical loggings and drilling must be carried out on similar elements of relief and landscape. In the tundra, the survey points are marked by earth pyramids and in the forest zone by wooden bench marks. The points and traverses are surveyed by a topographer using special instructions worked out for geophysical crews, or by a geophysicist himself.

5. The electrical profiling and logging must satisfy the following requirements.

If the roof of the basement rock is above or just below the lower permafrost surface, it is difficult to locate it without an adequate amount of reliable logging data on the resistivity of the lower horizons of frozen unconsolidated soils and rocks (frozen and thawed). In this case attempts to locate the lower permafrost surface by electrical logging will not be expedient.

The mapping of the permafrost table is done with the help of a diagram of paired electrical profiles AA'MNB'B, where the distance between large separation electrodes (in the outer current circuit) is 40 m (AB = 40 m), the distance between small separation electrodes (in the inner circuit) is 16 m (A'B' = 16 m), and the distance between the electrodes in the receiving circuit is 2 m (MN = 2 m). If it is required to obtain a detailed description of geocryological conditions on small sections, the spacing of measuring points must not exceed 8 - 10 m, and that of profiles 25 - 50 m. In preliminary geocryological surveys use may be made of a symmetrical profile AMNB, which is both simpler and cheaper. In this case the unit spacing is increased to 20 m and on sections with uniform landscape to 40 - 80 m. In areas with very diversified landscape and in the vicinity of test holes and VEL points the spacing is reduced to 8 m, and to 1 or 2 m when surveying repeated sequences of ice veins.

When surveying large areas it is expedient to adopt a different method of investigations, a so-called method of reference sites. The sites are selected after certain intervals along a given route and a series of investigations consisting of drilling, electrical profiling, electrical logging, and landscape studies is carried out there. On such sites drilling and logging are done on the same type of terrain. It is expedient to obtain electrical profiles which cross at the point where drilling is done. The length of profiles depends on whether or not the survey is intended to include all adjacent types of terrain.

If the site selected is not truly representative of all types of local terrain, individual profiles must extend beyond the limits of the site in order to obtain all the information required.

If there are boreholes in the area under investigation, they should be logged if possible.

6. The electrical and ultra-acoustical logging of boreholes is a sufficiently accurate method of determining the thickness of permafrost, its ice content, and the location of its upper and lower surfaces.

Basically there are the following methods of electrical logging: apparent resistivity method, electric current method, concentrated emf method, electrode potential method, dielectric method, and temperature method.

Lateral logging commonly used in the oil industry is also quite promising as far as permafrost studies are concerned (Doll, 1956; Uinn, 1956). The value of lateral logging, which is done by means of a microprobe, lies in the fact that it makes it possible to measure values close to those of true specific electric resistivity. The results can be made more accurate by measuring the resistivity with the help of sliding contacts. In this case it is a good idea to wet the walls of the hole with water prior to logging (for example with a cleaning rag).

The more simple arrangement required for logging with a protecting electrode (screened grounding) has certain advantages over conventional logging of resistivities, because it eliminates to a considerable extent the distorting effect of low resistivity of the drilling solution, or water in the case of high resistivity soils. This method can also be applied with the help of a microprobe.

The resistivity logs provide comparable data on the apparent resistivities of permafrost in different regions (required for the interpretation of VEL curves) and make it possible to determine more accurately the depth of the permafrost table. Furthermore, by considering the wide range of apparent resistivities in relation to ice content, it is possible to determine the ice content of soils in the cross-section and locate the boundaries of different lithological horizons.

The differences in the hydrochemical composition of aqueous solutions on the boundary between frozen and thawed soils result in the appearance of noticeable concentrated emf. The logging of concentrated potentials in combination with resistivity logs may serve as an additional method of determining more accurately the depth of the permafrost table, and locating the taliks and the pereletok bodies within permafrost.

Electric logging is simple but gives only a rough idea of changes in the ice content throughout a cross-section. Grounding on the surface is done close to a borehole; the sleeve of the casing cannot be used for grounding purposes.

The circuit is powered by dry batteries only for resistivity and electric current logging. The efficiency of these methods is clarified by the example of determining the depth of the lower permafrost surface (Fig. 36).

In the case of electric logging it is absolutely essential to satisfy the following requirements: the number of grounding electrodes on the surface must be constant; they must be located at the same depths and the same distances from the top of the borehole; the voltage must remain constant, i.e. one should strive to achieve uniform grounding conditions.

Electrode potentials will be produced in a system with two electrodes made of two different metals and moved in a borehole drilled in a heterogeneous frozen soil mass. The electrode potential logging may prove to be an auxiliary method of investigating permafrost.

The difference in the dielectric permeability of water ($\varepsilon_w = 81$) and ice ($\varepsilon_i = 3$) is the underlying principle of dielectric logging. The advantage of this method is that it is no longer necessary to create reliable contacts between the sensing elements in the hole and the frozen soil. It makes it possible to distinguish the phase composition of water in the soil and is promising as an additional method of investigating a permafrost cross-section more accurately. It is the most easily applied method of geophysical investigations of frozen soils.

The electric logging of hand-drilled boreholes can be done by the method of sliding contacts in dry holes, or with the help of a conventional probe in holes containing a drilling solution. The first method can be applied only in the southern part of the permafrost region, where the soil temperatures are close to $0^{\circ}C$; the second method may be applied in all other regions. The measurements may be carried out continuously or at intervals of 15 to 20 cm. The voltage Δv and the current I are read from field potentiometers of EP, EKS-1, or KSR types without recorders. The measurements can be carried out with the help of a milliampere metre. In all types of logging the points of measurement must be located at the same depth.

The holes intended for logging are drilled dry; casing is inserted to the depth of the thawed layer only. Use is made of standard borehole diameters, i.e. 2", 3" and 4". The electrical or the ultra-acoustical logs are obtained when the temperature measurements have been completed. During logging of the upper part of the cross-section, the casing is gradually removed.

The distance MN between the receiving electrodes in resistivity measurements and the electrodes measuring the concentrative emf is equal to 5 - 15 cm.

It has been shown experimentally that due to high voltages the distance AO between the current electrodes and the centre of measuring electrodes must be at least 75 cm. Otherwise, due to very small values of coefficient K and large values of apparent resistivities of frozen soils, the voltages become so high that it is no longer possible to measure them with a field potentiometer.

The charge is lowered to the bottom of the hole by means of drill rods or a weight.

The brushes for logging dry holes by the method of sliding contacts must be available for all borehole diameters (2", 3" and 4"). To increase the strength and improve the shock absorption on moving along the hole, the brushes must be protected from all sides by thick rubber rings.

On logging the electrode potentials, it is possible to use brushes made of the following metal pairs: iron-copper, iron-brass, and iron-zinc.

On logging the holes filled with a drilling solution, it is essential to use water with the highest possible resistivity (atmospheric waters, swamp water).

The ultra-acoustical logging may be done in dry holes or holes filled with a drilling solution. On logging dry, uncased holes, the acoustical contact between the piezometers and the wall of the hole is achieved by means of small rubber balloons filled with air and joined to the surface pump by a rubber tube. The ultra-acoustical logging by the sliding contact method can also be done by using an elastic base for the microprobe in the form of ρc-rubber (or other types of rubber with fillers which provide it with the required insulation and acoustical characteristics) as a contact and probe-accommodating material.

The logging of holes filled with a drilling solution is done by means of an ultrasonic, hermetic probe developed by the Institute of Earth Physics, Academy of Sciences of the U.S.S.R.

In regions with large temperature gradients in permafrost, the logging of resistivities, concentrated emf, and electrode potentials is done immediately after the end of the drilling. The temperature is measured 6 to 12 hours later. The measurements are repeated several times until the results begin to coincide.

In regions with high-temperature permafrost, the holes must be left for periods ranging from several days to several months prior to geophysical investigations. The period of settlement is considered to be over when new measurements coincide with the previous results throughout the entire depth of the hole within the given degree of error (up to $0.1^{\circ}C$).

The temperature logs have been described earlier and we shall not discuss them here.

In order to check the results of quantitative interpretation of VEL curves when determining the mode of occurrence of permafrost and the thickness of its upper and most dynamic horizon, it is required to measure the temperature on the so-called cooling surface (at a depth of 4 to 6 m) close to VEL points. These data are used to check by way of calculations the VEL results with the help of the geothermal method developed by D.V. Redozubov (1955). The temperature measurement points must correspond to the general calculation pattern accepted for the given area and the shape of the frozen soil body.

At least one temperature value for a depth of 300 - 500 m (below permafrost) must be obtained from temperature logs of deep holes. Alternatively one should have some data on the geothermal gradient in the given region.

7. Up to now field investigations of permafrost have been carried out mainly by means of the resistivity method. Electrical profiling proved its worth in mapping the permafrost table, and in locating vertical or inclined taliks, sections with the highest ice content of soils, and repeated sequences of ice veins. The curves in Figure 44 are a good example of this.

Vertical electrical logging can be used to determine the depths of the upper and lower permafrost surfaces, as well as the depth and the thickness of horizontal taliks which are as thick as the frozen layer above them. Figures 45, 46 and 47 show VEL curves of discontinuous and continuous permafrost.

Curve 2 in Figure 45 has a smaller modulus $\mu_1 = \frac{\rho_2}{\rho_1}$ than curve 1, and its left asymptote can be traced to the abscissa AB/2 ≈ 9 m. By using a two-layer measuring grid, it is easy to see that curve 2 refers to a section where the layer of seasonal freezing does not reach the permafrost which is at a depth of 5 m. The curves in Figure 46 reflect the presence of one thawed horizon enclosed in permafrost. Curve 1 indicates that this thawed horizon is 11 m thick and that its top is at a depth of 13.4 m. The corresponding values for curve 2 are 14 and 17 m.

The curve in Figure 47 was obtained on a section with two horizontal layers of thawed soil enclosed in permafrost. A quantitative interpretation of the six-layer electrogeocryological cross-section was not possible because of weak differentitation of resistivities. The presence of two thawed horizons was confirmed by drilling.

The permafrost table is usually mapped by means of an electrophilic arrangement with small separations of input electrodes. To reduce the distortions due to frequent heterogeneities near the electrodes, it is expedient to use an arrangement with input electrodes referred to infinity, i.e. the arrangement AA'MNB' $\rightarrow\infty$, B $\rightarrow\infty$.

It is known from experience that to satisfy the condition of infinity without exceeding the given limits of error, it is sufficient to make the distance between the electrode B or B' and the centre of the arrangement O five times that of the distance OA or OA'.

In cases where it is difficult to locate the permafrost table, a paired dipole electroprofiling arrangement has proved to be superior to other arrangements. It gives a very clear picture of various small details in the relief of the permafrost table.

The advantages of a dipole arrangement are eliminated in the following cases: a) when the permafrost table is relatively flat and b) when the horizontal measurements of small individual structures on an extensively dissected permafrost surface are smaller than the dimensions of the receiving dipole.

Since the cost of investigations involving a dipole arrangement is relatively high, the latter may be recommended only for detailed surveys of individual sections.

As a rule, the results obtained by the middle gradient method are not better than those obtained by other methods.

The most complete and reliable information on the composition and structure of permafrost is obtained by a combination of drilling with electrical surveys, geothermal studies and electrical logging of boreholes.

On mapping the upper and lower surfaces of permafrost and its ice content, use is made of a larger number of separations of electrical profiles and electrical logging, the curves of which are extended to the right asymptote corresponding to thawed soils beneath the permafrost.

For a reliable determination of the depth of the lower permafrost surface from VEL curves, the following parameters must be known: h_1 - depth of the permafrost table, ρ_1 - resistivity of the thawed soils above permafrost and ρ_2 - resistivity of the permafrost.

During an engineering survey of small sites, geocryological investigations may be carried out by using the same basic modifications of electrical, microseismological and ultra-acoustical surveys. It is recommended to make the distance between the profiles 25 - 50 m and to take measurements every 8 - 10 m. The number of VEL points should amount to 2 - 3% of the total number of electroprofiling points and to not more than 5%, if points of intersection are also taken into consideration. The number of manually drilled boreholes in this case should amount to not more than 20% of the given number of VEL points (without taking into consideration intersecting and circular loggings). An efficient distribution of boreholes and points of ultra-acoustical measurements is achieved on the basis of maps of equal resistivities or electrical profiles (Akimov, 1960).

In preliminary surveys, the geophysical results are mapped on a scale of up to 1 : 10,000, and in detailed engineering surveys on a scale of up to 1 : 5,000.

During surveys along a predetermined route or on large sites, the geophysical investigations are carried out in the following order. Electrical profiling is the first step. The sites for electrical logging or ultra-acoustical and microseismic measurements are then selected on the basis of maps of profiles or equal apparent restivities.

The taliks are delineated by means of several separations of current electrodes, which depend on the depth of the taliks and the thickness of the permafrost.

Repeated sequences of ice veins can be located by an electrical survey with the help of circular electrical profiles. The directions of main ice veins in the polygonal network (first generation veins) are determined first. The resistivities are measured along three profiles intersecting at one point at angles of 60°. The length of these profiles must exceed the maximum diameter of the polygon by a factor of 4, and the spacing of the measuring arrangement must be equal to 2 or 3 MN.

If circular electrical profiles do not supply the necessary data for the identification of the main elements of the polygonal system, the profile lines are orientated parallel or perpendicular to the direction of a river (or ravines, ancient terraces, etc.), since veins are usually orientated in this way.

In an electrical survey of ice veins, the two-layer cross-section with ρ_1 (the resistivity of thawed soils above permafrost) and ρ_2 (the resistivity of permafrost) is complicated by the vertically orientated vein, the resistivity of which $\rho_3 > \rho_2$. There is a complex relationship between the apparent resistivity above such a vein, which extends below the base of the first layer, the resistivities ρ_1, ρ_2 and ρ_3, the thickness of the first layer

h_1, the width of the ice vein m, and the separation of electrodes in the input circuit. However, if all parameters, except h_1 and m, are constant and the screening effect of the first layer is small $\left(\frac{\rho_1}{\rho_3} > 0.005\right)$ the difference between anomalous and normal ρ_a in the first approximation will depend on the ration m/h. Under actual conditions, at $\frac{\rho_1}{\rho_3} > 0.005$, there is a limit (at $m/h_1 < 0.17$) beyond which such a dependence on the ratio m/h_1 is no longer valid and the identification of ice veins becomes difficult, because the latter no longer have any marked effect on the behaviour of the ρ_a curves. This means that at h_1 = 3 m (the maximum depth of thaw in summer in the regions where ice veins are common), the ultimate thickness of an ice vein which can be determined by electrical prospecting is 0.5 m. At this depth, thinner ice veins give rise to barely noticeable ρ_a maxima which are difficult to define because of the variable nature of the ρ_a curve.

The following system of electrical profiles may be recommended for a detailed survey of ice veins: the distance between the profiles must be equal to the diameter of the smallest polygon; the spacing of the measuring arrangement must be equal to the separation of measuring electrodes MN; the length of the profiles must depend on the dimensions of the given area. To reduce false ρ_a maxima, it is expedient to use an arrangement with one input electrode in infinity, or a dipole electrical profiling which defines the ice veins more clearly.

The results of the electrical survey of ice veins are represented in the form of a map of electrical profiles. Different values of ρ_a maxima are correlated with the sequence of generations of fractures. In this way it is possible to identify the axes of ice veins of different generations.

Electrical, microseismic and ultra-acoustical methods can also be used during long-term geocryological studies at one particular place. They make it possible to determine the rate of thawing or freezing of soils in different types of terrain, and to observe qualitative changes in the permafrost surface in populated areas or beneath buildings while these are being used. This is done by periodic measurements along the same profiles.

To avoid the interfering effects of stray or telluric currents, when such effects are especially strong and the application of a pulsator does not eliminate them, it is recommended to use the IAK-1 device which does eliminate the interferences.

It is often possible to avoid strong and unstable polarization effects on sites where they are especially pronounced by reducing the contact surface between the circuit electrodes and the ground to a minimum, or by transferring throughly dried pins to a different place.

Exploration of thawed sections beneath buildings or detailed surveys of important areas can be done by means of an ultrasonic ground-tester (modified PEL-1 or PEL-2 echo sounders), or by microseismic methods. In the first case it is best to use a magnetostriction vibrator as an emitter and a piezoelectric receiver. However, best results are obtained at frequencies which are below ultrasonic (f = 6 - 21 kc).

Geophysical investigations cannot be done without due considerations being given to geocryological ocnditions. A geophysicist working on surveying and geostructural problems must have a good knowledge of geocryological and physical properties of the permafrost in a given region.

In some cases permafrost distorts and complicates the results of geophysical investigations, but in others it may have a beneficial effect.

On tracing geological structures at shallow depths beneath the overburden, a small separation of a paired electroprofiling arrangement, unless fully justified, may be unduly affected by the permafrost. A small separation of a paired profile on high-resistivity structures (for example, on limestone anticlines) results in changes in the apparent resistivity which reflect the changes in the frozen mass better than those in the morphology of high-resistivity structures. Because of this, the ρ_a curves obtained with small separations of current electrodes are of little help in the interpretation of electroprofiling data.

The separation of current electrodes must be based on at least three results of electrical logging: on the axis, the limbs, and beyond the limits of the structure. All three VEL curves must be obtained on sections where the temperature of the permafrost is at its lowest.

When tracing large structures extending over long distances, a check should be made every few kilometres to see whether the selected separation of current electrodes is correct.

When planning geophysical investigations, use should be made of regional permafrost maps based on geocryological data.

A geophysical survey of coal deposits in a permafrost region revealed anomalous electric resistivities which were thought to be due to the coal seams but were in fact due to accumulations of ground ice. In this case permafrost had a detrimental effect on the interpretations.

In some regions the resistivity of coal seams located in low-temperature permafrost under a thin blanket of overburden differs sharply from the resistivity of country rock. The water content of coal seams prior to freezing was higher than that of country rock. Because of this, the resistivity of the coal seams rose sharply after freezing due to their high ice content. Therefore the frozen state of coal seams and country rock is a factor which favours a geophysical survey because it brings out the differences in the resistivities. However, the resistivity of coal seams found in thawed rock differs little from that of the latter, because in this case the coal seams become saturated with water.

A reliable correlation of "coal" peaks on apparent resistivity curves for a series of profiles can be done only after a detailed study of geocryological conditions in the region. If these conditions are not known, the field electrometric system should contain a third separation of current electrodes, which should be such as to ensure mapping of the permafrost table.

When the separation of current electrodes is small (AB 100 m), use is made of light ground pins which can be easily inserted into the soil. With the exception of a few cases (work conducted in early spring or at the beginning of seasonal freezing, areas with outcrops of basement rocks where seasonal thawing does not extend to any considerable depth), there is no need for reinforced ground pins if logging is conducted at shallow depths. The latter are, however, essential if the separations of electrodes are large (AB > 100 m). When determining the cost of field work, use is made of a coefficient from the existing standards which takes into consideration the difficulties involved in providing a contact between the electrodes and the soil. The same coefficient is used when working in the period of incomplete seasonal thawing (at the beginning of the field season) and at the beginning of winter freezing (at the end of the summer season).

When the electrode separations are very large (AB > 10,000 m); the geocryological conditions do not have to be taken into consideration when interpreting the geophysical data.

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5. Investigations of Groundwater and Surface Water

The formation of water-bearing horizons in the permafrost region depends on climatic, geological and geocryological conditions. The main task in the study of groundwater in this region is the determination of occurrence of water-bearing horizons, layers, beds, etc., and their interrelationship with permafrost containing water in a solid phase.*

The types of occurrence of groundwater in the permafrost region are given in various classification schemes (Table XXX), in which the importance of the location of water-bearing horizons and permafrost with respect to each other is given different meaning. N.I. Tolstikhin (1933) suggested that there are three types of water: suprapermafrost, intrapermafrost, and subpermafrost water. However, already the first geocryological studies have shown that in the permafrost region use should be made of a classification based on the genetic characteristics of groundwater and its relationship with permafrost. Such a classification was proposed by I.Ya. Baranov (1940).

Comprehensive studies in different parts of the permafrost region led to the discovery of new characteristics of the genesis, regime and occurrence of groundwater.

Classifications of groundwater reflect two main schools of thought.

The supporters of one school strive to retain the division into three types of water (supra-, intra-, and subpermafrost), and propose further subdivisions of each type. The supporters of the other school underline the mutual interrelationship of groundwater types (artesian, stratal, stratal-fissure, fissure, and karst water) and their relationship to surface water. The location of water-bearing soils and permafrost with respect to each other is regarded as an additional characteristic of hydrogeological conditions only.

When identifying the genetic types and varieties of groundwater in a given region, it is required to determine its main characteristics both from the point of view of general hydrogeology and its occurrence with respect to permafrost. The investigator should strive to determine the effect of permafrost on the supply, circulation and discharge of groundwater, as well as the effect of the latter on the development of permafrost.

Hydrogeological investigations required for permafrost studies are examined below. Descriptions are given of both overall and special investigations. The volume of work to be performed will depend on the task in hand and the amount of details required.

* Under groundwater in this section, the author understands free water.

The study of water basins and rivers

Water basins and rivers in the permafrost region are of special interest in the study of geocryological and hydrogeological conditions in a given area. They are connected with subsurface water through taliks in the permafrost.

a) Water level. A change in the water level in water basins may serve as an indicator of the existence of a link between groundwater and surface water. It is required to establish the nature of these changes throughout the year, determine in what period of the year the water level rises or falls, the extent of fluctuations and the pattern of their occurrence over a period of many years. Special attention should be paid to the changes in water level on complete freezing and thawing of soil on sections adjacent to the water basin and at the start of ice formation, since during these periods the part played by surface, ground and artesian water in supplying the water basins, and conversely the effect of the water basin on the water supply, can be seen especially clearly. Preliminary information concerning all this may be obtained from local inhabitants, by observations at hydrological stations, by special observations, and indirectly from other sources.

The sources of indirect information include the traces of former water and ice levels (salt films, deposits of silt) left on trees, shrubs and rocks. Changes in the water level may be established by examining the morphology and dynamics of the ice crust in the middle or in the second half of winter (i.e. by studying the subsidence, heaving and fracturing of ice).

Hydrogeological observations of a general nature are conducted in accordance with existing instructions. The data obtained are analyzed to establish the relationship between surface and groundwater. Heaving in the central part of the ice crust on a water basin points to freezing of subsurface drainage channels and the presence of a subaqueous inflow of water. Subsidence of the ice cover and concentric fractures in the ice indicate complete freezing of groundwater streams which fed the lake when the drainage was uninhibited.

b) Water temperature. The temperature changes in time and space are determined by observations. The temperature is measured in several places in a grid-like fashion and across the lake at different depths, depending on the depth of the lake and the pattern of temperature distribution. In the rivers the temperature is measured in several places across the channel. When selecting the measuring points, attention should be paid to the geological structure of the river valley and especially to signs of tectonic activity. The observations make it possible to find local changes in the water temperature abnormal for the given basin or river and resulting from the presence of subaqueous inflows of subsurface water (along a water-bearing layer, tectonic fractures, a channel of karst origin, etc.). Such a section of the valley must be investigated more thoroughly.

It is desirable to measure the water temperature in basins and rivers periodically throughout the year. Systematic observations are especially important in winter, when the effect of groundwater is much more evident due to a reduction of or interruption in the supply of surface or groundwater. The investigations are carried out in a river section or a water basin with most typical geological structures and characteristic relief, or where the regime differs from that of other rivers and basins. Systematic observations make it possible to determine the temperature regime of water bodies and should be performed at least twice a month.

Observations carried out on lakes in winter make it possible to calculate approximately the heat fluxes through the ice resulting from the heat exchange in the system lithosphere (material composing the bottom of the lake) - water - ice - atmosphere. The annual heat cycle without the latent heat of thawing (freezing) and evaporation (condensation) may be calculated by means of the simplified equation proposed by M.M. Krylov (1952):

$$\Sigma_q = \int C\gamma(t_1 - t_2)dh,$$

where Σ_q - total heat cycle, kcal/m^2;

C - specific heat of water (1.0);

γ - unit weight of snow, ice or water, kg/m^3;

$t_1 - t_2$ - difference between the highest and the lowest temperature in the snow layer, ice or water throughout the year, $^{\circ}$C;

h - depth of the basin, m.

Similar calculations help to determine the nature of interactions between water and permafrost.

c) Geocryological examination of the bottom of a water basin. It is important to determine the temperature and the state of the ground composing the bottom of a water basin or a river bed, which is difficult and requires a great deal of work. It involves drilling (preferably in winter) and casing. The composition of soils beneath the bottom of the basin and the water temperature are then determined. The borehole must be made watertight to prevent surface water from entering it. If this is impossible, the measurements should be made all the same, because they provide some indirect information concerning the presence of a talik, or indicate whether the soils beneath the bottom are frozen and their properties. The observations are carried out according to instructions used in hydrogeology. Geophysical methods should be used also (electrical and temperature logging).

On flood plains the outlines of taliks beneath water basins, rivers and creeks may be quite complex (in plan and in depth) and may change somewhat with time in accordance with annual variations of hydrological conditions. The investigations are carried out across the taliks within and outside the basins and river channels. Boreholes and testpits are located at different distances from the water edge.

In coastal areas and shallow sections of northern seas, studies are made of the soil temperature and phase composition of the water. Here one may find frozen soils or soils with a negative temperature but containing mineralized water.

d) Chemical composition of water. Simplified hydrochemical analyses of water are performed in field laboratories. Complete analyses of water and gas (free and dissolved) are carried out at the main laboratory attached to the base camp of the expedition.

The results of chemical analyses make it possible to investigate the interrelationship between surface and groundwater and to determine more precisely the subaqueous exits of the latter. Water samples are taken by conventional methods. The best results may be obtained by hydrochemical investigations carried out at different times of the year (in the "low-water" period, on freezing, during maximum freezing, on ice break up, etc.).

e) Characteristics of the ice crust. Certain aspects of formation and structure of the ice crust on water basins and rivers serve as indicators of hydrogeological characteristics of a given area. Therefore the ice crust should be examined several times during on season. Attention should be paid to the presence of river icings, ice mounds, and air holes. The first two are discussed in Part II of this publication. The air holes persisting throughout the winter or a greater part of it are often formed as a result of subaqueous exits of groundwater, or a considerable increase in the rate of flow, for example in shallow sections. Measurements are made of the size of the air hole, the water depth, the ice thickness (by drilling or piercing the ice around the air hole and both upstream and downstream from it), water temperature, rate of flow, and the discharge within the air hole as well as upstream and downstream from it (beneath the ice). Hydrochemical and gas analyses are carried out also. The points of individual measurements are marked on the map.

The time of formation, the pattern of recurrence, and the changes in the dimensions of the air hole are determined by repeated observations or from local inhabitants. By summarizing the collected material and comparing it with the geological structure of the area, it is possible to determine the genesis of the air hole and certain characteristics of the taliks.

Investigation of springs

By studying groundwater springs in the permafrost region it is possible to establish the variations in their regime in the course of a year. Seasonal fluctuations in their yield, temperature and chemical composition are often much greater here than outside the permafrost region. This is due to the freezing of water-bearing horizons supplying the springs and to local concentration of groundwater in winter.

The following investigations are required in addition to conventional hydrogeological studies.

a) Observations at the source of a spring. It is established whether the source of a spring is stationary or whether it migrates in the course of a year (and if so, when). The number of springs varies greatly and they often disappear for a variety of reasons. Hence it is essential to determine the pattern of migration over a period of many years. This can be done by periodic observations, questioning local inhabitants, and studying various indirect signs which indicate that the source of the spring changes its location in winter. Such signs are: the remnants of an icing, poorly developed vegetation with small crooked trees, delayed beginning of the vegetation stage in shrubs and grass as compared with their normal development on adjacent sections, and bare rocky patches of ground. The last two factors indicate the presence of a thick icing cover.

It is most important to establish the relationship between the source of a spring and the form of relief, its exposure, and the geological structure of the area (lithology and tectonics). When studying the geological structure, special attention should be paid to the ice content of frozen soils (the presence of ice layers and lenses). It is also important to determine the outline and the hydrogeological properties of the talik surrounding the spring.

b) Composition and properties of water. The chemical composition of spring water, its gas content and contamination with bacteria are determined for a period of one year. Such investigations are especially important in the permafrost region, because due to freezing and thawing of some water-bearing horizons supplying the spring, the seasonal changes in the chemical composition of groundwater are much more pronounced here than outside this region.

Water samples for chemical and gas analyses must be obtained several times a year during the most typical periods in the regime of a given spring, i.e. at the time of maximum seasonal thawing (in late fall) and maximum seasonal freezing (at the end of winter). In the "critical period" (at the end of winter), the chemical composition of spring water, its gas content and the temperature give an especially clear indication of the genesis of the water-bearing horizon constantly supplying the spring. Therefore this period is especially favourable for the identification of "indigenous flows" of groundwater.

Salt films and crusts which precipitated out during the winter are often observed on the soil and rocks near a spring, especially in early spring before the first rain. In winter salt films can be found on the surface of the icings. These precipitates are carefully collected and analysed.

In winter the bottles with water samples are carried in boxes lined with cotton wool or felt. The bottles are protected by felt or cloth jackets and are packed in moss or hay. One or two hot water bottles are placed in the box as well. In this way water samples can be transported for several hours.

c) Temperature of spring water. The temperature being one of important characteristics of groundwater in the permafrost region should be studied periodically throughout the year. The temperature changes during the critical period in the groundwater regime are especially significant. If there are several exits of water, the measurements are carried out in different places to find the exit of the main stream, for example the place where water emerges from a crack in the rocks and enters the overburden.

d) The yield of a spring. The yield is measured in different seasons, and especially in the critical period. In some cases the yields in the summer are smaller than those in winter (for example when rising water streams are absorbed by thawed alluvial deposits). The measurements are done by conventional methods. The "head" of a spring is often covered with ice and can be found only by clearing the ice away, which may be quite difficult.

The yield of a spring in winter is sometimes determined by periodic measurements of the volume of ice in a growing icing. However, this method does not always reveal the true yield, since some water may disappear in thawed soils (hidden yield) or beneath the icing, which sould be taken into consideration.

The methods of studying the icings, these characteristic cryogenic formations revealing a great deal of information on the hydrogeological conditions in a given area, are discussed in Part II.

The investigation of taliks

In the study of hydrogeological conditions in a given area it is extremely important to know the nature of occurrence, composition and structure of frozen and thawed (unfrozen) soils. Thawed soils in low-lying areas are usually saturated with water. They act as collectors of groundwater or serve as paths for its migration.

The term talik is commonly given to soils with positive temperatures enclosed in frozen ground.

It should be noted that there is still no universally accepted genetic classification embracing all types of taliks. If the area occupied by unfrozen soils is greater than that of permafrost, distinctions should be made between taliks located within permafrost, and unfrozen soils surrounding permafrost "islands" (in the southern zone of the permafrost region). Unfrozen soils and taliks are studied in accordance with instructions given in the manuals of engineering geology and hydrogeology. Main attention is paid to the hydrogeological characteristics of taliks. The task of finding and especially of outlining the shape of a talik is very difficult.

Stable taliks are usually indicated by the following factors: a) large water basins and stream which do not freeze solid (taliks beneath lakes and river beds and on flood plains); b) powerful and constantly acting ground water springs, indicated by seasonal and perennial frost mounds, seasonal icings, and perennial icings (hydrothermal taliks); c) local concentrations of heat-loving plants represented by deciduous trees (poplar, aspen, birch) and evergreens (spruce, pine) (insolation and infiltration taliks on slopes and flood plains), dense willow groves (infiltration-sublacustrine taliks and taliks beneath river beds), or a dense cover of mixed grasses (taliks on flood plains), which are easily identified against a background of tundra, taiga (larch) or steppe vegetation characteristic of a given area*.

The delineation of taliks may be done with the help of excavations combined with geophysical exploration (vertical electrical logging and profiling). Delineation with the help of excavations only is costly and requires a great deal of labour.

The sudy of the water content, dynamics, temperature regime, and origin of a talik can be done only by a combination of geocryological and hydrogeological methods. These investigations are extremely important in engineering-geocryological and hydrogeological exploration.

The effect of structures

The erection of various structures often leads to considerable changes in the groundwater regime. In the permafrost region these changes give rise to specific phenomena. Let us examine some of them.

a) Artificial icings. Construction of roads, excavation of ditches, removal of vegetation, compaction of snow, etc., often lead to a local increase in the depth of seasonal freezing, and, if groundwater is at a shallow depth, to formation of frost mounds and icings. As a rule, these

* Vegetation may serve as an indicator of taliks only if the geocryological characteristics of a given area are known (Russian editor).

are located above the point, where the natural groundwater regime had been disturbed. Special attention should be paid to the effect of an icing on the engineering structures, the causes and the nature of changes in the gounidwater regime, and the mechanism of formation of frost mounds and icings. This information is required to develop the protection measures.

The icings or the frost mounds are marked on the map and hydrogeological investigations are carried out in the area around them. The depth, the direction of flow and the yield of groundwater are determined together with seepage properties of soils in the water-bearing horizon, its regime, its sources of supply, etc. The sections with frost mounds and icings are examined in the summer and at the end of winter. This renders it possible to evaluate correctly the hydrogeological and geocryological conditions and hence develop suitable protective measures.

b) Increase in the level and the pressure of groundwater beneath structures. If groundwater is near the surface and permafrost is at a shallow depth beneath the water-bearing horizon, the groundwater level may rise considerably in winter or early spring in taliks containing permeable soils formed beneath heated buildings. This is due to pressure exerted by the expanding layer of seasonally frozen soil on groundwater outside the building and the fact that the groundwater flow tends to become narrower. This leads to flooding of basements and a reduction in the strength of foundation soils.

Observations are carried out to determine the onset and the end of the increase in the water level, the rate and the extend of maximum increase, the direction and the yield of groundwater flow (in special boreholes or test-pits), the composition and the seepage properties of soils in the water-bearing horizon, the rate and the depth of seasonal freezing of soils, etc.

In the period of maximum seasonal freezing, the groundwater stream is sometimes divided into several "flows", the direction of which changes and often does not coincide with the direction of the groundwater stream in the period of maximum seasonal thawing. The hydrostatic and the hydrodynamic pressures increase with the freezing of soil. Under natural conditions this leads to formation of frost mounds and icings.

In studies of a more general nature it is necessary to determine the area over which the rise in the groundwater level is taking place and find the relationship between the depth, the rate of freezing, and the pressure of the water-bearing horizon on one hand, and the air temperature and atmospheric pressure on the other. This information is required to develop methods of avoiding undesirable increases in the groundwater level beneath buildings. The regime of suprapermafrost groundwater is studied by means of testpits and boreholes. The walls of excavations are braced. The testpits and boreholes should be covered.

Investigations at mining sites

A number of observations is carried out during hydrogeological investigations of mining sites in the permafrost region.

a) Cryogenic structure of rock and soil. The general methods of studying the structure of frozen ground have been described elsewhere. When investigating soils containing ice it is required to establish the type of ice-filled voids, i.e. whether ice is enclosed in tectonic fractures, karst voids, pores, unconsolidated soils, etc., and to determine whether there are any regular changes in the ice content at different depths and in different sections.

Having determined the nature of ice (its structure, the chemical composition of melt water, etc.), it is required to establish the type of water from which it was formed (stratal, fracture, karst, atmospheric, condensation, etc.).

b) Icing of mine workings. Ice is often formed in mine workings, making mining operations difficult. It is essential to establish the type of water which formed this ice (groundwater, seepage water, etc.) and to determine the location of ice, its volumes, rate of growth, etc. It is also necessary to find the reasons for and the conditions of ice formation, whether this takes place because of the cooling effect of the permafrost, or owing to the circulation of air in winter. Therefore apart from soil temperature measurements in places with largest ice accumulations, it is also necessary to determine the temperature of the air, the rate of its movement in the workings, seasonal variations of these factors, etc.

c) Mineralized water in cooled ground. In the workings mined out in the cold zone of the earth's crust with temperatures below zero, there may be an occasional horizon containing highly mineralized liquid water with a negative temperature. There may be cases where water will move along fractures in the permafrost and on gradual cooling and freezing will increase its mineral content. Such water is found on the coast of northern seas in mine workings situated below sea level. Water-bearing soils cooled to a temperature below zero can be found also in inland regions on mining salt-bearing deposits characteristic of vast areas of the Middle Siberian Upland, where they form the lower part of the zone of cooling of the earth's crust.

Apart from conventional hydrogeological investigations, in all these cases it is also required to measure the temperature of rocks, determine the chemical composition and the temperature of groundwater at different depths, in different parts of the deposit and at different times of the year, establish the relationship between the changes in the temperature and the chemical composition of groundwater on one hand and its yield on the other (for example on removing the mine water by pumping), and determine whether there are any variations in the yield and the chemical composition of water during mining. It is also required to measure the freezing-point of mineralized water, examine the distribution of ice inclusions in the overlying permafrost, and determine the characteristics of transition from permafrost to the cooled water horizon.

It is essential to establish the origin of mineralized water and its relationship to certain soils or tectonic phenomena, and develop methods of eliminating or utilizing this water (for example for obtaining salt solutions, for medicinal mineral baths, etc.).

d) Groundwater in taliks in mined deposits. Mine workings in permafrost are usually dry. In the presence of water-bearing horizons outside the permafrost, especially if these are under pressure, it is essential to carry out thorough geocryological investigations to be able to predict the appearance of water in the mine. It should be taken into consideration that water-bearing soils may occur not only beneath but also beside the permafrost and may dissect the latter into blocks. Therefore it is essential to conduct systematic checks of temperature changes in the face of stopes advancing in the direction of a possible talik. Systematic observations with the help of special boreholes drilled from within a stope render it possible to detect the proximity of a talik or a large water-bearing fault zone. When drilling such testholes, steps should be taken to prevent sudden outbursts of high pressure water.

Investigations in boreholes and wells

Hydrogeological investigations in boreholes drilled in permafrost are of great interest.

Drilling with water, brine or clay solutions, supplies very poor data on permafrost since it greatly disturbs the natural temperature of soil around the holes. Air drilling is, therefore, much more promising, because in this case the hole remains dry, and the temperature regime of soil is altered to a much lesser extent. The cryogenic soil structure is not disturbed and can be studied in the core.

In wet drilling, conventional hydrogeological investigations must be supplemented by systematic temperature measurements of the solution entering and leaving the hole. The data are presented in the form of a diagram (Fig. 49). The measurements are conducted at least twice every shift and the results are recorded in a log book. These data provide only general information on the changes in the hydrogeological and the temperature regimes of soil. When the hole reaches the base of the permafrost layer, there is usually a change in the relative positions of soil temperature curves.

A good idea of the water content of thawed soils may be obtained by systematic measurements of absorption of the washing solution which can be done by conventional methods.

It is essential to examine the core immediately after extraction, since ice may be present in cracks and voids indicating that the soil is indeed frozen. The ice inclusions should be sketched or photographed and subjected to chemical analysis.

During water surveys and while investigating sites for hydrotechnical structures, attempts are made to pump water either after certain intervals of depth (for example, every 50 or 100 m), or in accordance with information obtained by examining the core and by measuring the absorption and the temperature of the drilling solution. Special attention should be paid to soils in the lower part of the permafrost layer and below it, because they may provide some information concerning the interrelationship between permafrost and the water-bearing horizons. However, since water can be pumped usually only for short periods of time, the data obtained may be used as a general guide only, while some properties are invariably distorted (e.g. temperature and chemical composition of the water).

The operating wells in permafrost are studied by conventional methods. The main task is to prevent water from freezing. Therefore pumping should be done quite vigorously with as few interruptions as possible. With prolonged pumping, a talik may form around the well. In some cases forced pumping will prevent water from freezing even if frazil ice is beginning to form in the well. It is recommended to heat the well prior to pumping.

In operating wells (both free flowing and pumped), it is essential to investigate the seasonal fluctuations in the yield and the possibility of a talik forming around the casing.

On completing the drilling or pumping of wells in permafrost, the piezometric level of fresh water in the well comes to rest in the permafrost zone and the water may freeze. Under such conditions, systematic measurements of the water level become very difficult. Freezing of fresh water is possible even in a flowing well located in a thick layer of low temperature permafrost, or if the yield and the water pressure are low. Therefore when studying a flowing well, especially in winter, it is essential to clean it periodically and to prevent the formation of an icing at the top of the well hindering the flow of water.

The freezing of water in a non-flowing well is prevented as follows.

1) A concentrated aqueous solution of table salt is poured down the well making certain that in the upper part of the water column, which is

usually the coldest because of the effect of the permafrost, the concentration of the solution will be high enough to prevent the water from freezing. It should be borne in mind that the upper layer of water gradually loses its salt content and more salt should therefore be added from time to time, as indicated by temperature and mineralization of the water at different depths. The addition of salt distorts the natural chemical composition of the water and the natural water level in the well. Nevertheless, this is the most commonly used method, since it is simple, reliable and makes it possible to observe the relative changes in the water level.

2) The well is periodically heated with hot water, steam or electrical devices. A single application of hot water is of little use because its heating effect is short lived.

Prolonged application of steam is more effective because it results in the formation of a talik around the well. The heating is done by means of a steam boiler and a tube (3/4 - 1" in diameter), the open end of which is lowered into the well. It is best to place the tube between two columns of pipes of suitable diameters (e.g. between the casing and the water pipe). This method disturbs the temperature regime of the soil. It may be used in specially equipped wells intended for experimental pumping of the water.

Another fairly effective method is the use of an electric heater similar to a domestic electric kettle, which can be made on site bearing in mind the diameter of the well, the depth at which it will be used, and the available voltage.

Freezing can sometimes be prevented by periodically pouring warm water down the well for long periods of time. The temperature of this water does not have to be high (5 - 10^{o}C) and it can be introduced at a rate of 1 - 2 m^3/sec.

3) Non-freezing liquids (petroleum, solar oil, used lubricating oil) are sometimes poured down the well to displace water from the low-temperature zone. This method may be used if the subsurface water pressure is not high. However, it has many disadvantages (pollution, waste of valuable oil products).

It is very difficult to measure the water level in a well in permafrost, since the measuring device (electric level meter, etc.), the cables and especially the twine used for lowering the meter freeze to the wall of the well. To avoid this, all equipment must be kept dry and lubricated with some substance which will not freeze to the pipes (e.g. solidol). The device is lowered slowly and the cable to which it is attached is pulled up and down to prevent the probe from freezing to the walls. The probe cannot be left in the well. A probe frozen to the side of the well can be released with the help of a concentrated solution of table salt.

All steps taken to prevent the well from freezing must be recorded in the logbook.

Closed or abandoned exploration wells containing frozen water can be used for certain studies of hydrogeological properties of soils. If the well is empty above the ice plug, it can be used for measuring the temperature. In some cases the depth of the ice plug gives an indication of the depth of the static water level prior to freezing. However, an ice plug may be formed or may increase in size as a result of water flowing in from above. At times the plug may occur above the true water level due to lowering of the latter.

It is more expedient to drill through the ice plug rather than drill a new well. When drilling through a plug attention should be paid to its position in the well. It is not recommended to drill right through a plug in one operation. The presence of a plug is indicated by increased resistance to drilling.

The hydrogeological studies in shallow testpits and boreholes containing groundwater are also specific. The link with the water-bearing horizon may be interrupted by the freezing of walls of a testpit, a well, or a borehole in the seasonally frozen layer or permafrost. In the warm season, surface of vadose water may accumulate in testpits and boreholes which freeze solid in winter or do not contain groundwater. The water level in such pits and holes cannot serve as an indication of the groundwater table. Therefore, the results of investigaions in shallow testpits and boreholes should be examined with great caution, and the best method of establishing whether there is an actual link with a water-bearing horizon is to pump water. Such pits and holes can be protected from freezing by means of a heat-insulating cover.

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Chapter V. Geocryological Mapping

2. Standard Symbols for Geocryological Maps, Profiles and Cross-Sections

The purpose of symbols is to represent on a map the geocryological characteristics of a given area. The symbols used depend on the type of map and its scale. In the case of aggregate geocryological maps, it is required to use complex symbols; in other instances where it is necessary to show only one phenomenon and related features, the symbols used may be relatively simple.

Geocryological maps reproduce the most important characteristics of frozen soils and related phenomena: a real extent and depth (thickness), composition and structure, genetic types and their boundaries, types of cryogenic and postcryogenic formations, age and stage of development of frozen beds, etc.

The aforementioned symbols are divided into groups characterizing groups of related phenomena. Maps, plans, profiles and diagrams can be both coloured or black and white. The colours on coloured maps represent the main characteristics of objects under investigation, while the geographical base is shown in black. On black and white maps, various phenomena are shown by thicker lines, hatching, scale-adjusted symbols (in the case of large-scale maps), and conventional symbols (on small-scale maps).

The most important factors are composition, structure, and genesis of frozen soils. The genetic types should be represented by a variety of colours or hatching.

For clarity, colours may be used in the preparation of special maps also.

Profiles through typical sections are added to the maps to make them three-dimensional and more detailed. The profiles incorporate symbols used on maps of larger scales and are placed either beyond the margins of the map or on the map itself.

To add more detail to maps, plans and diagrams, use may be made of inserts (small-scale maps), tables, etc.

Table XXXII contains the symbols recommended for black and white geocryological maps, plans, cross-sections, and profiles.

Structure, composition and properties of frozen soils, their location and age gradation with depth are shown by profiles or block-diagrams (block maps). All factual data are given in tables. The methods of graphic representation of these data vary. Diagrams are constructed to show temperature, moisture content, certain properties, and structure of soils, etc.

Figure 54 shows a combined diagram of air temperature, snow depth and soil isotherms. Figure 55 shows a diagram of temperature and composition of frozen soils. Another example of a combined diagram is Figure 56, which shows composition, temperature, ice content, and moisture content of soil.

Individual areas are described by means of profiles showing the geological structure, the depth of the upper and lower surfaces of frozen soils, the location of taliks, the location of groundwater, ice inclusions, boreholes, and other data.

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PART THREE. LONG-TERM GEOCRYOLOGICAL INVESTIGATIONS

Aims and Methods of Field Studies

Long-term geocryological investigations are carried out when it is required to study over a long period of time the dynamics of processes taking place in permafrost on seasonal freezing and thawing in areas affected by land development, construction, etc. Such investigations form an integral part of a whole series of exploratory type studies and are initiated to solve various theoretical and practical problems.

The objects of obligatory long-term studies are: a) thermal regime of soil; b) water regime of soil; c) seasonal freezing and thawing of soil; d) heaving and subsidence of soil; and e) snow and ice covers.

Long-term geocryological studies intended to solve theoretical problems are included in or are based on investigations of physico-geographical, geological, geophysical, geomorphological, and other characteristics of a region. Studies of a practical nature are carried out on given or typical construction sites and agricultural land to determine their suitability for construction, industrial and agricultural purposes. In particular, they help to determine the methods of construction and land development, as well as establish the condition of existing engineering structures, and the effect of the latter on the foundation soils (and vice versa), etc. The nature and volume of observations and methods to be adopted are decided separately in each case and depend on the task in hand. In certain types of investigations, for example in the study of heat exchange between the ground and the atmosphere, it is required to carry out actinometric, meteorological and gradient observations in the surface layer of air.

Recently, instrumental measurements of the thermophysical properties of soils and of heat fluxes in the soils have been included in long-term geocryological investigations. These studies are of great importance in the investigation of the formation of seasonally and perennially frozen soils in conjunction with the design of structures. Special engineering observations are conducted when investigating the conditions of construction (Principles of Geocryology, Part II, 1959).

Prior to long-term studies, investigations are made to determine the environmental conditions at experimental sites using the appropriate detailed geocryological survey as a reference. The knowledge of environmental conditions, particularly of the composition and properties of soils, is absolutely essential for the success of any long-term investigations.

The experimental site must be located in an area where natural conditions, i.e. characteristic relief, vegetation and lithological soil composition have been retained. Additional sites are prepared, if required, on sections with similar relief but different lithological soil composition and vegetation.

When determining certain changes in the freezing and thawing processes, the sites are cleared of turf and snow which gives rise to somewhat different conditions of heat and moisture exchanges between the soil and the atmosphere - an important factor in road and airport construction. These bare sections are essential in the investigation of building sites, because as far as the development of cryogenic phenomena are concerned, they will resemble actual conditions of construction much closer than the sites where the vegetation and snow have been retained.

The dimensions and shape of additional sites are determined in such a way as to make the distance from the boundaries of each site to its centre not less than twice the maximum depth of the instruments placed in the ground. The area of the site must be at least 100 m^2. If temperature measurements are not required, the dimensions of the main site must be the same as those of the additional site. The sites for temperature measurements are selected to avoid the effect of other sites with similar or different environmental conditions.

Fresh snow is immediately removed from a stripped experimental site with wooden shovels and is dumped at a distance of at least 100 m. The site must be swept after that. The sections accommodating the instruments are cleared of snow first. The plants are cut as soon as they appear. A stripped section must be prepared at least a year prior to the beginning of investigations. In the first year or even in the first two years, the thermal and water regimes of the soil will be in the process of adjusting to the new conditions of heat and moisture exchanges between the soil and the atmosphere. For certain purposes, studies are conducted on a stripped section immediately after its preparation and in this case the analysis of data obtained is made with allowances for changed natural conditions.

Quite often the object of the study is the actual process of attaining an equilibrium in the thermal and moisture regimes of soil on a stripped section (for example when investigating the freezing and thawing processes in the soil, the lowering of the permafrost table, heaving, subsidence, etc.). In this case the observations are conducted for one year prior to preparation of the site and are continued after the natural conditions have been disturbed.

The advantage of electric frost gauges over other methods of measuring the depths of freezing and thawing lies in the fact that it is not necessary to take them out of the hole to obtain a reading, and this considerably increases the accuracy of observations.

Tables are made up during the investigations to show both the primary data and the summarized material required for subsequent analyses and conclusions.

Conclusions after one year of investigations are recorded as shown in Forms 2 and 3. Columns 1, 2 and 3 pertaining to ten day periods may be replaced by any dates, while the intervals between the dates of observations may be increased or reduced depending on the given conditions.

It is rather difficult to determine the duration of stable freezing and thawing (see Form 2). The date after which the soil no longer thaws at the surface is taken as the beginning of stable freezing, and the date when the depth of freezing becomes stable is taken as its end. If there are several periods of freezing during one winter (the southern part of the zone of seasonal freezing) and each period lasts more than 30 days, then the onset of the first period of freezing is taken as the beginning of stable freezing, and as its end - the onset of stabilization of freezing in the last period. In the study of seasonal freezing of soil in the permafrost region it is essential to find the depth of the permafrost table or indicate that permafrost is absent (there is a special column for this in Form 2). If the investigations of freezing and thawing of soils were not systematic, this is noted in the table and the dates and durations of the various periods are not recorded.

The dates and duration of stable thawing are determined as follows. The onset of thawing which is no longer interrupted by frost resulting in the freeze up of the thawed layer, is taken as the beginning of stable thawing. The final thawing of the seasonally frozen layer is taken as the end of the period of thawing. In the permafrost region this date coincides with the onset of seasonal freezing of soil from below or with stabilization of the depth of thaw.

The processing of annual observations makes it possible to obtain average results and data for a period of many years. The average values for a period of many years are calculated from the average depths of freezing (thawing) for each year. The average data concerning the onset of stable freezing of the soil, the date when the thawing has been completed, and the duration of stable freezing (thawing) of soil is calculated from appropriate data for each year.

Objects and Methods of Measurement

a) Thermal regime of soils

The thermal regime is studied by measuring the rate of change of soil temperatures, heat capacity of soil, thermal conductivity, diffusivity, and the magnitude and direction of heat fluxes at different depths and different times of the year.

The soil temperature is studied under natural conditions and under conditions which have changed owing to the effect of structures, land development and agricultural activities.

The temperature at depths exceeding 3 m is measured in specially equipped boreholes once every 5 or 10 days, once a day at depths of 1 - 3 metres, and three times a day at depths of 0.4 - 1.0 m at 7 a.m., 1 p.m., and 7 p.m. (sometimes four times a day with an additional reading at 1 a.m.). The measurements at depths of up to 0.4 m are carried out every 3 or 6 hours (at 1 a.m., 7 a.m., 1 p.m. and 7 p.m., or 1 a.m., 7 a.m., 10 a.m., 1 p.m., 4 p.m. and 7 p.m.) with Savinov thermometers (in the summer), thermocouples, thermistors, and resistance thermometers. The temperature on the ground surface is measured with rapid reading thermometers and thermal spiders.

Automatic temperature measurements are of great importance in long-term observations. Resistance thermometers and thermoelectric thermometers (thermocouples) may be used as temperature-sensitive elements. Electronic potentiometers, electronic bridges and photo-recorders may be used as recording devices.*

The automatic recorders consist of the required number of probes located at known depths. The temperature of the snow is measured with thermocouples. Devices in which the temperature of the thermocouple is determined by resistance thermometers may be recommended for various long-term geocryological investigations which call for a degree of accuracy to within 0.05 - 0.1°C.

A basic design of a thermocouple assembly is shown in Figure 114. The following depths of installation of thermocouples may be recommended for investigations of a general type: 0.05, 0.10, 0.15, 0.20, 0.30, 0.50, 0.75, 1.0, 1.5, 2.0, 2.5, and 3.0 m. In such a system all thermocouples usually operate on the same constantan wire and the reference thermocouple

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* The design of temperature-sensitive elements and methods of temperature recording are given in Appendix 1.

is also common to all thermocouples. It is installed at a depth of 3 m or deeper, depending on the depth of freezing and thawing of the soil. The copper wire from the reference thermocouple leads directly to one of the clamps on the galvanometer or the potentiometer. The copper wires from all other thermocouples lead directly to the corresponding couples in the switch, which makes it possible to connect any thermocouple to the recorder.

All thermocouples and cables are placed in a vinyl tube provided with a tray and the tube is filled with a cable filler to make it water tight.

The temperature of the reference thermocouple is measured by a resistance thermometer installed at the same depth as the thermocouple.*

For better control, use is often made of an additional resistance thermometer. The temperature of the reference thermocouple and the calibration scale of the thermocouple are checked by measuring two given temperatures of the upper reference thermocouple, which is located outside the soil.

The thermal spider is intended for measuring the temperature of the soil surface to within ±0.1°C. It represents (Fig. 115) a copper-constantan thermal battery of 5 - 10 thermocouples connected in series. The lower (reference) thermocouples are at a depth where daily temperature fluctuations are no longer felt. The upper thermocouples are located in the surface layer (at a depth of 2 - 3 mm) in such a way as to include all noticeable depressions and elevations of the surface. If the surface is sufficiently flat, the thermocouples are located in a circle.

The cables of the battery and the lower thermocouples are placed in a vinyl, textolite or carbolite tube in a tray. The tube is filled with a cable filler. The upper thermocouples are enclosed in casings of plexiglas or AKR-7 plastic. The surface cables are protected by rubber, vinyl or vinyl chloride tubes.

Resistance thermometers are used to measure the temperature of the reference thermocouples in thermoelectric systems. These measurements are made with the help of a three-wire system and a Wheatstone bridge.

The accuracy of measurements by resistance thermometers (±0.05°C) is quite adequate for the processing of data supplied by temperature-sensitive elements in the course of periodic measurements and photoregistration. During photoregistration the temperatures are measured with thermocuples to within ±0.1°C.

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* In determinations of soil temperatures to within ±0.1°C, the temperature of the reference thermocouple is measured by a pull-out soil thermometer.

For convenience the instruments for measuring the temperature of soil and snow are assembled in a single system. It combines the soil thermocouple system, the thermal spider, the thermocouple system for the snow, and the resistance thermometers. The wires and the thermocouples are placed in a vinyl tube with an inside diameter of 3 cm. The tube must be provided with a metal tray. The free space within the tube is filled with a cable filler. The tube with wires is lowered into an uncased borehole and the space between the walls of the hole and the tube is filled with soil obtained from drilling. The tube must protrude above the surface of the ground to the height of the snow or vegetation.

The temperature regime of the soils at the base of the layer of annual temperature fluctuations is studied in boreholes drilled to a depth of up to 25 m.

The holes are drilled dry and cased with thin-walled metal piples. If the hole is stopped in permafrost, the bottom end of the pipe is closed with a wooden plug. If it is stopped in thawed ground, a cover is welded on. The pipe must protrude 40 - 50 cm above ground. The protruding part is protected by a wooden box measuring 35 x 35 x 60 cm. The space between the box and the pipe is filled with heat insulating material (sawdust, slag, moss, etc.). The top of the pipe is covered with a lid and corked.

Systematic measurements are started after the natural temperature regime of soil has been re-established. Use is made of slow-reading (inertia) thermometers, resistance thermometers or thermistors made up into bundles.

The depth of installation of individual thermometers depends on the task at hand. The first thermometer is usually installed at a depth of 2 - 3 m. Subsequent thermometers should beat depths of 5, 7, 10, 15, 20, and 25 m. The use of slow thermometers has some disadvantages and the accuracy of measurements does not exceed ±0.1°C. For more precise determinations, systems are used consisting of several metal or semiconductor resistance thermometers (see Appendix 1). At depths of up to 5 m, measurements are made once every 10 days, and once a month at lower depths, if there is no water migration in the soil.

Precise measurements are done by remote-control automatic recorders consisting of temperature-sensitive elements (thermocouples and thermal spiders), multichannel program switch, recording block, supply block, and control block.*

* Devices with optical recorders have been used at the field station of the Permafrost Institute near Moscow since 1957.

Figure 116 shows an example of a temperature record obtained by an automatic photo recorder. The temperature determinations by means of thermophotograms are carried out as follows.

The zero curve of the galvanometer corresponding to the soil temperature at the reference point (at a depth of 3 m) is found on the thermophotogram and on processing the latter, all curves on it are referred to this curve: their deviations from the zero curve of the galvanometer correspond to the differences in the temperatures at the reference point and all other points of the measuring system. Since the temperature at the reference point changes slowly and smoothly, the temperature at any point and any period of time may be found with sufficient accuracy by interpolation.

For convenience, an isometric line corresponding to a whole number of degrees is traced on the thermophotogram. For example, on the thermophotogram shown in Figure 123 this line can be traced by using the given temperature scale either above or below the zero line of the galvanometer. In the first case it will correspond to a temperature of 5°C and 6°C in the second.

On tracing the isometric line, allowances should be made for the shape of the zero line. All fluctuations of the latter must be reflected on the isothermic line. In the intervals where the curvature of the zero line is at its maximum, the isothermic line must be based on the largest possible number of temperature readings obtained at the reference point.

A thermophotogram may be processed in the following way:

1) note the date and time when the photorecorder was switched on;
2) establish a time scale (in days); the start of the day is marked on the scale line corresponding to 24 hours;
3) note the temperature at the reference point and record the readings on scale lines corresponding to the time of measurements;
4) trace the isothermal line;
5) find the temperature for all depths.

The temperature is determined with a thermophotogram in the following way. A transfer temperature scale is established along the corresponding line of the time scale. The divisions of the transfer scale and the corresponding temperature on the isothermal line are correlated in such a way as to make the increase in the temperature on the thermophotogram coincide with its increase on the transfer scale. The temperatures at all points of the thermometric system are then determined from the temperature scale.

The thermophotograms makes it possible to find the temperature as well as to express the changes in degree-hours. For this the area enclosed between the temperature curves on one side and the isothermal line on the other is measured with a planimeter. To convert the readings on the planiméter to degree-hours, use is made of a conversion factor based on the given time scale and the temperature scale of the given photorecorder.

The thermophotograms of snow and the photograms of thermal spiders are processes in the same way.

The heat fluxes in thawed and frozen soils at depths of 0.1 to 2 - 3 m can be measured with devices developed at the Leningrad Technological Institute of the Refrigeration Industry.* The device consists of a 6 mm thick rubber disk with an iron-constantan thermal battery in the centre. The hot and the cold thermocouples are on the opposite sides of the disk. Both sides of the disk are covered with a rubber coating 2 mm thick to protect the thermocouples from damage and water. The protective coatings are strongly bonded to the main disk by vulcanization. The measuring device is 300 mm in diameter, and the working zone in the centre is 200 mm in diameter. It contains the thermocouples on a paired Archimedean coil and is surrounded by a circular, protective zone 50 mm wide. The heat meter has 700 - 800 thermocouples connected in series to the thermal battery.

The meter is calibrated in the presence of a steady heat flow with a flat device with an electric heater. A note is made of the conversion factor to relate the emf of the battery expressed in micro-volts to the heat flow in kcal/m^2/hr. The heat flow is measured to within $0.1 \cdot 10^{-3}$ kcal/cm^2/min. or 0.05 kcal/m^2/hr.

Prior to placing the meter in the soil, the exit clamps are isolated from moisture by raw rubber which is then vulcanized. Raw rubber extends over the protective ring and the rubber jacket of the cable.

The depth of installation of the meter depends on the task at hand (0.1, 0.5, 1 and 2 m).

If the meter is to be installed in thawed soil, a pit is excavated rapidly to a depth exceeding that of the installation. A recess with a flat floor is provided in the wall of the pit. The soil taken from the recess is somewhat thinned out and spread on the floor to provide a base for the meter. The latter is then mounted firmly on this base; the recess and then the pit itself are filled with carefully compacted soil. The

* Devices of this type are intended for measuring conductive heat exchanges. On freezing or thawing of water-saturated soils, their readings are sometimes distorted by heat convection, seepage and migration of water. These devices do not last very long.

measurement of heat flow may be started 2 or 3 weeks after the installation of the meter, when disturbed thermal and water regimes of the soil have been practically restored.

The heat flow at a depth of over one metre is measured once or twice a day by a PPTV-1 potentiometer. At depths of up to one metre, the heat flow is measured at least four times a day or continuously recorded with photorecording devices, pyrometric millivolt meters or automatic potentiometers. The recording of heat flow is checked by periodic measurements with PPTN-1 or PPTV-1 potentiometers. Automatic meters can be switched over to periodic measurements with three-polar switches (tumblers).

Figure 117 shows a diagram of a system to measure heat flow. Owing to considerable convection of water and the effect of solar radiation, such meters cannot be used in water basins. Heat flow can also be measured with meters of different designs (Kolesnikov and Speranskaya, 1958; Deacon, 1950).

Let us now examine the methods of determining heat flow with the help of automatic photorecorders. Figure 118 shows an example of a photorecord obtained a depths of 0.1 and 0.5 m.

The zero line of the galvanometer corresponds to heat flow equal to zero. The deviation of the heat flow curve from the zero line indicates on a certain scale the magnitudes of positive or negative heat flow. The scale factor K_1 is found as follows:

$$K_1 = \frac{E}{A} \text{ mv/cm}, \qquad (1)$$

where E - emf of the meter determined with a potentiometer, mv;

A - deviation of the heat flow curve on the photogram from the zero line of the galvanometer, cm.

In the determination of K_1, E and A must refer to the same moment of time. A more precise value of K_1 is found as the arithmetic mean of several values of K_1 obtained for different moments of time. E and A are determined in the periods when the rate of change of the heat flow is low.

The photograms are processed in the following way:

1) a note is made of the date and time when the photorecorder became operational;

2) the time scale (in days) is traced in coloured ink; the beginning of a day is marked on the scale line corresponding to 24 hours;

3) having established the time scale, the periodic potentiometer measurements of the emf are plotted on the photogram;

4) K_1 is found and is used to prepare a scale rule;

5) the zero line of the galvanometer is traced in black ink and the lines denoting heat flow in coloured ink;

6) the heat flow is found for the given periods; the zero on the scale rule is made to coincide with the zero line of the galvanometer and the magnitude of the heat flow is determined.

In this way the heat flow in the ground can be determined for any period of time.

The photograms make it possible to determine the total heat flow in the soil (ground) in any interval of time. To do this, a planimeter is used to measure the areas enclosed between the lines denoting the heat flow and the zero line of the galvanometer. The total heat flow for the period is found by multiplying the planimeter reading by its coefficient.

This coefficient is determined by moving the planimeter over a rectangle formed by two lines L (cm) and two perpendicular lines B (hours) three times. The coefficient K_p is found from the formula:

$$Kp = \frac{K_1 K_2 LB}{n} \text{ kcal/m}^2, \qquad (2)$$

where K_1 - scale of the photogram;

K_2 - constant of the heat measuring device;

n - planimeter reading (the average of three readings).

Within the CGS system, the coefficient of the planimeter is:

$$Kp = 0.1 \frac{K_1 K_2 LB}{n} \text{ kcal/sec}^2 \qquad (3)$$

Let us examine the methods of measuring the heat flow in the soil (ground) with a meter developed at the Leningrad Technological Institute of Refrigeration Industry.

In periodice measurements of emf of the meter (E) with a potentiometer, the heat flow (Q) is found from the equation:

$$Q = EK_2, \qquad (4)$$

where K_2 - constant of the heat flow meter.

Daily, monthly and yearly totals of heat flow can be easily determined with the help of Q. In the case of daily totals of heat flow at a depth of 0.1 - 0.2 m, Q must be measured at least every two hours, otherwise the systematic error may exceed 10%.

Let us now examine the method of determining the heat flow at a depth exceeding 2 m. The change in the temperature at this depth takes place slowly and may be represented with sufficient accuracy by a straight line. Then from equation (1):

$$\frac{dT}{dz} = \frac{T_2 - T_1}{z_2 - z_1}, \tag{5}$$

where $T_2 - T_1$ is the difference in the temperature at depths z_2 and z_1.

The coefficient of thermal conductivity λ is calculated by the method proposed by Kersten (Kersten, 1949). In the formulae below, λ is in kcal/m/hr/deg, the unit weight of soil γ is in kg/m^3, and the moisture content W is in percent.

a) for thawed silty and clayey soils:

$$\lambda = 0,124\,[0,9\,\lg W - 0,2]\,(10)^{0,01\frac{\gamma}{16}}; \tag{6}$$

b) for frozen silty and clayey soils:

$$\lambda = 0,124\left[0,01\,(10)^{0,0114\frac{\gamma}{16}} + 0,025\right](10)^{0,014}; \tag{7}$$

c) for thawed sandy soils:

$$\lambda = 0,124\,[0,7\,\lg W + 0,4]\,(10)^{0,01\frac{\gamma}{16}}; \tag{8}$$

d) for frozen sandy soils:

$$\lambda = 0,124\left[0,011\,(10)^{0,0214\frac{\gamma}{16}} + 0,0256\right](10)^{0,0146\frac{\gamma}{16}}\,W. \tag{9}$$

Nomograms have been established for equation (7) to (10) simplifying the calculation of λ.

In the permafrost region the saturation of frozen soils with water may exceed the lowest moisture capacity. In this case, λ of soil may be taken as approximately equal to that of ice (2 kcal/m/hr/deg).

Let us examine the calculation of the heat exchange between the soil (ground and the atmosphere, if the heat flow at a certain depth is known.

Let the total heat flow at depth h and in the period of time τ be equal to q. The heat content in the layer dx in this period of time will be:

$$c\,(T_1 - T_2)\,dx, \tag{10}$$

where c - unit heat capacity of soil in the layer dx;

$T_1 - T_2$ - change in the temperature in the layer dx in time τ.

Hence the heat exchange between the soil and the atmosphere in time τ will be equal to:

$$B = \int_0^h c\,(T_1 - T_2)\,dx + q. \tag{11}$$

In practice the following formula is used:

$$B = \sum_{1}^{h} c(T_1 - T_2)\,\Delta x - q. \qquad (12)$$

The sequence of calculations by means of equation (12) is as follows:

1) A table is constructed showing the soil temperature in the initial period τ_0 and the final period τ_1 to depth h where the heat flow has been measured; h must not exceed 0.2 - 0.3 m. The difference between the depths where the temperature has been measured is taken as Δx. The average temperature for all layers Δx in the initial period τ (T_1) and the final period τ_1 (T_2) is determined.

2) The unit heat capacity c is calculated for each layer Δx.

3) q is measured with a planimeter or by calculations.

4) The heat exchange B in time τ is then determined.

The thermophysical properties of soils (thermal conductivity and diffusivity) are found by various methods but most of these can be applied in a laboratory only. Under field conditions these properties are determined by calculations from temperature, moisture content, etc.

In the field the coefficient of thermal conductivity of thawed and frozen soils is determined with a spherical probe designed by the Agrophysical Institute. This is a metal sphere enclosing an electric heater and one of the junctions of a differential thermocouple. The coefficient of thermal conductivity can be found from the change in the temperature of the outer surface of the sphere which depends on the extent of heating of the sphere and the thermal properties of soil.

When studying the freezing and thawing of soil, it is best to install the spherical probes at depths of 0.1, 0.2, 0.3, 0.5, 0.75, 1.0, 1.5 and 2.0 m. The sphere together with the thermocouple junctions are placed in a niche made at the required depth in the wall of a pit. The distance between the sphere and the cold junction is set at 0.3 m. The niche is then filled with soil from the same horizon. When all probes are installed, the pit is gradually filled with excavated soil which is compacted layer by layer.

The cables from the probes extend to the surface and are connected to a switch which connects the probe to the circuit of the measuring device. The latter can be switched on either directly in the field or by remote control from the measuring station. The switch is connected with the switchboard at the station by means of a ShRPS cable.

In accordance with specifications, the coefficient of thermal conductivity is determined once a month.

A serious disadvantage of this method is the long time required to obtain each measurement (about one hour) and the rapid deterioration of the instruments. Other designs of probes with different types of the heating element (plate, cylinder, thin wire) are also used in long-term investigations (Deryabin, 1957).

b) Water regime of soils

The study of the water regime is essential in all long-term geocryological investigations, since it explains the development of cryogenic and numerous other phenomena and processes taking place in the soil and is required for engineering calculations. It involves the investigation of dynamics of the hydro-physical properties of soil, i.e. water permeability, water capacity, total water content of frozen soil, residual water content of mineral layers after freezing, and ice content.

The studies also involve investigations of the unit weight, water regime and ice regime of thawed and frozen soil, and observations of precipitation, evaporation and condensation.

The water content is determined in samples taken at identical sites away from experimental apparatus. The total water (ice) content of soils subject to seasonal freezing is measured in samples taken at 10 cm intervals for the first 100 cm (sometimes at 5 cm intervals for the first 10 or 20 cm), and then every 20 or 50 cm. The total water content of frozen soils is determined once a month. In the study of heat fluxes and thermophysical properties, the water content is examined at least once every ten days.

In some cases the remote-control methods of measuring the water content of seasonally-frozen soils with hygrometers* give good results.

In the permafrost region, the total water content is best determined by the method of obtaining an average sample (Pchelintsev, 1954). The water content of mineral layers is found from samples taken in these layers. As an approximation, it can be made equal to the lower Atterberg limit of plasticity (Bakulin, 1958)

In the permafrost region, the measurement of the water content of seasonally thawed soils with remote control hygrometers is difficult, because the soils are saturated with water and the hygrometers give reliable results only if the water content is not high. These methods are still in the experimental stage.

* The most common are the IVP-53 hygrometers with carbon electrodes (Danilin and Razumova, 1956).

The ice content of frozen soil due to the presence of ice cement is calculated from the water content and the relation between the amount of unfrozen water and temperature (Nersesova, 1954).

The wilting point of soil is 1.2 to 2.0 of its maximum hygroscopicity. It is determined by the method of using seedlings (Ventskevich, 1952).

In regions where the natural water content of soil is high, it is not possible to find the minimum field water capacity. A method of determining it has been described by S.I. Dolgov (1957).

c) Seasonal freezing and thawing of soil

Data on seasonal freezing and thawing of soil in different physico-geographical regions are essential for the solution of various practical and theoretical problems.

The depth of seasonal freezing and thawing in relation to the properties of thawed and frozen soils are determined during long-term investigations by various methods. Visual, temperature and frost gauge methods are the most commonly used. The first two have been described in Chapter II of Part I. We shall now examine the use of frost gauges of various designs.

The most simple devices are the frost gauges designed by Danilin and Ratomskii.

The Danilin gauge (Fig. 119) consists of an ebonite or carbolite tube, the length of which depends on the depth of freezing or thawing of the soil. The inside diameter of the tube is 20 mm, the outside 26 mm. The main part of the frost gauge is a rubber tube filled with pure (distilled) water. The inside diameter of the rubber tube is about 8 mm and its wall thickness is 1 mm. A scale is provided on its surface. The upper end of the rubber tube and the zero mark on the scale must coincide with the surface of the ground. The tube is fixed to a wooden rod. The other end of the rod is joined to a lid which covers the protecting carbolite tube. The lower end of the carbolite tube is buried in the soil below the probable depth of freezing; its upper end must extend above the snow cover on the ground.

The observations are started as soon as cold weather sets in and are carried out periodically every 5 or 10 days. The depth of freezing is found from the location of the lower edge of the frozen water column in the rubber tube. A frost gauge of this type makes it possible to determine not only the depth of freezing but also the depth of thaw (in the permafrost region). The depth of thaw is found from the location of the top of the frozen water column.

The frost gauge designed by Ratomskii consists of a thin-walled metal tube provided with slits and filled with moist soil from the hole drilled for the installation of the gauge. The tube is shown in Figure 120. The lower end of the tube is plugged with a wooden plug and a wooden rod is inserted in its upper end. The metal tube is then placed in an ebonite or a phenol-plastic casing 2 to 5 m in length, depending on the maximum depth of freezing or thawing. The internal diameter of casing is 25 mm. A mark is made on the casing some distance from its stop to indicate the position of the zero mark on the scale on the metal tube. The lower end of the casing is provided with a cone-shaped bottom and a ring to make the end stronger and watertight. The hole to accommodate the frost gauge is drilled with a bore 37 mm in diameter. Soil samples are taken during drilling to determine the natural water content and the grain size composition.

The hole is cased in such a way as to make the mark on the tube coincide with the ground surface. The space between the wall of the hole and the casing is packed tightly with soil. The metal tube filled with soil saturated to full water capacity is lowered into the cased hole. The time of installation of the gauge is noted in the log book. The depth of freezing or thawing is determined by lifting the frost gauge out of the hole. The boundary of the frozen layer is found from the resistance to some mechanical action such as poking with a blunt awl or needle, or from the presence of visible ice crystals.

In the study of seasonal freezing in the permafrost region, measurements are made of the upper and lower limits of freezing. In the study of thawing in the zone of seasonal freezing of soil, measurements are made of the upper and lower limits of thawing.

In the course of these studies it is essential to watch for the vertical shifting of casing resulting from frost heaving. In the presence of heaving, the frost gauge readings must be corrected or the casing put back into place.

If the frost gauge is not long enough, the soil will freeze throughout its entire length. In this case the gauge is refilled with unfrozen soil, the wooden rod is lengthened, and the gauge is installed at a lower depth.

The frost gauge readings are recorded as shown in Form 1.

Form 1

Freezing and thawing of soil

Site .... Frost gauge ....

Date of installation ....

Date	Frost gauge reading, cm	Correction, cm	Freezing (thawing) after correction, cm		Depth of snow cm	Remarks
			from above	from below		

In the zone of seasonal soil freezing the observations are started with the arrival of first frost and continued until the thawing of the soil is complete. In the permafrost region the observations are started when the soil begins to thaw and continued until thawing is complete; later, when freezing sets in, the observations are continued until the seasonally frozen layer has reached the permafrost.

All changes noted during the observations are recorded in the column marked "Remarks", for example: appearance of water in the casing, removal of soil which fell out of the gauge from the hole, vertical shifting of casing, etc.

The electric frost gauges are more complex devices. They are based on the fact that the electric resistivity of frozen soil is many times higher than that of thawed soil. This is the criterion used to determine the depth of freezing or thawing.

The design of an electric frost gauge is simple. It consists of a receiver, a device measuring the resistivity, and input batteries. One type of electric frost gauge is shown in Figure 121. It consists of a M-1101 megohmmeter and a rod with electrodes. The rod consists of a tube (1) the outside diameter of which is 8 cm. It is made of vinyl plastic which has good insulating properties, is strong, and is highly resistant to the corrosive action of water. The ring electrodes (2) are made of brass and are flush-fitted into the rod at equal distances from each other. A conductor passing through the tube connects each electrode to the contact points on the commutator (3).

With electrodes of the same width and the same distances between them, the resistivity measured by the megohmmeter will depend on the conducting properties of the medium surrounding the electrodes. Hence the megohmmeter will indicate the position of the boundary between frozen and thawed soils.

The advantage of electric frost gauges over other methods of measuring the depths of freezing and thawing lies in the fact that it is not necessary to take them out of the hole to obtain a reading, and this considerably increases the accuracy of observations.

Tables are made up during the investigations to show both the primary data and the summarized material required for subsequent analyses and conclusions.

Conclusions after one year of investigations are recorded as shown in Forms 2 and 3. Columns 1, 2 and 3 pertaining to ten day periods may be replaced by any dates, while the intervals between the dates of observations may be increased or reduced depending on the given conditions.

Form 2

Conclusions after one year of investigations of seasonal freezing of soils

Area ____________ Site __________ Year __________

Experimental plots					Depth of freezing, cm months																											Date of onset of stable freezing	Date of onset of stable thawing	Duration of stable freezing, days	Time of maximum freezing	Depth of perma-frost table	Remarks
Observation point	Relief, microrelief	Vegetation	Lithological soil comp.	Methods of determin. freezing boundaries	IX			X			XI			XII			I			II			III			IV			V								
					1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3						

Form 3

Conclusions after one year of investigations of seasonal thawing of soils

Area ____________ Site __________ Year __________

Experimental plots					Depth of thaw cm months																								Date of onset of thawing	Date of end of thawing	Duration of thawing, days	Time of maximum thaw	Total depth of thaw	Remarks
Observation point	Relief, microrelief	Vegetation	Lithological soil composition	Method of determining the extent of thawing	IV			V			VI			VII			VIII			IX			X			XI								
					1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3						

For individual periods the average, maximum and minimum depths of freezing (thawing) of soil over a period of many years are calculated from data obtained in the course of ten-day observation periods.

The results obtained are used to calculate the rate of freezing in the course of a season, average rates over a period of many years, etc.

All aforementioned observations are carried out on undisturbed sections, as well as in places where snow and vegetation have been removed.

All data concerning the seasonal freezing or thawing are analyzed in relation to lithological composition, temperature and water content of soils, and heat flow observed at the site.

d) Heaving and settlement of soil

Heaving results from the redistribution of water on freezing of soil and formation of lenses of segregation ice. Settlement on thawing results from the reduction in the soil volume on melting of ice lenses and thinning of soil.

Vertical displacement of the soil surface on freezing and thawing is determined by means of systematic levelling. The experimental site is divided into equal squares the measurements of which depend on the task at hand. The points of intersection of the sides of the squares are marked by pegs which are later replaced by cement plates (20 x 20 x 5 cm). The plates are installed at a depth of 5 cm. On levelling, the rods are placed on the plates.

Vertical displacement of soil can be measured also by conventional and remote-control linear displacement gauges, which may be connected to a recorder. To assemble the gauges it is necessary to have one or two reference rods which are rigidly connected to the measuring device.

The reference rod shown in Figure 122a may be used in the permafrost region. The rod shown in Figure 122b is recommended for use in the zone of seasonal soil freezing.

Heaving and settlement of soil at different depths is measured with a differential heaving gauge. The design of the gauge used at Skovorodino permafrost station is shown in Figure 123.

The device consists of a number of metal disks (1) and steel rods (2) attached to these disks. The diameter of the rods is about 6 mm. The device is installed in a testpit or a borehole. On installing the device in a pit, the disks are pushed into one of its walls to a required depth.

If the gauge is to be installed in the ground, boreholes are drilled to the required depths. The boreholes are spaced every 7 - 10 cm along a straight line.

The disks of outermost rods are placed at a depth where heaving and settlement are absent. The outermost rods are attached to the ends of a metal strip (5), which serves as the zero of vertical displacement of the indicator rods. To prevent the rods from adfreezing to the soil, they are protected by rubber tubes (3) filled with solidol. The bottom of the tube is fixed to the rod. To make the ends of indicator rods located outside the ground less elastic, they are passed through holes in a wooden headpiece (4).

The disks of the indicator rods are placed at different depths, such as 0.05, 0.20, 0.40, 0.60, 0.80, 1.00, 1.25, 1.50, 1.75, 2.00, and 2.25 m. The disks of reference rods are placed at a depth which depends on the depth of seasonal thaw.

The vertical displacement of the indicator rods is measured once every 5 days, but the displacement of the rods close to the boundary of seasonal freezing or thawing is measured every day.

If great accuracy is not required (to within ±1 mm), the measurements are made with a metal ruler placed along the rod. The zero end of the ruler is mounted on the strip (5) and the readings are taken from the position of the top of the indicator rod. For more accurate measurements of vertical displacements of the indicator rods, use is made of dial indicators.

e) Snow and ice covers

During long-term geocryological investigations it is absolutely essential to examine the snow cover. Studies are made of its depth, temperature regime, physical properties (structure, stratigraphy, density), water content, thermal properties, radiational characteristics, physical processes which take place in the snow (migration of water vapour, recrystallization), and its role as an insulator during seasonal freezing of the soil. The amount of investigations depends on the task at hand.

The depth of snow is investigated according to instructions issued by the hydrometeorological service of the USSR.

The temperature regime is studied with rapid-reading and minimum thermometers, thermocouples, and resistance thermometers (both metal and semi-conductor types). The choice of instrument depends on local conditions and the task at hand.

In a snow layer with daily temperature fluctuations, use is made of temperature-sensitive elements with low thermal inertia (copper-constantan thermocouples). If it is required to record the temperature continuously, the thermocouples are connected to a recording device (electronic potentiometer or photorecorder).

In a snow layer with low or negligible daily temperature fluctuations use may be made of high inertia instruments (resistance and rapid-reading thermometers). Minimum thermometers may be used when other types are not available. Rapid-réading thermometers are accurate but distort the natural structure of snow and are not recommended for long-term investigations.

At the field stations of the Permafrost Institute, Academy of Sciences of the U.S.S.R., use is made of thermocuple assemblies. The general outlays of such assemblies for snow and soil are similar (Fig. 124) but there are significant differences in design.

Thermocouples for use in the snow (Fig. 125) are placed in thin vinyl plastic tubes (3). The tubes are fixed to a wooden rod at 10 cm intervals, and to make them more secure they are also attached to an inclined wooden plank (2). The base (4) of the thermocouple assembly consists of a wooden rod measuring 4 x 4 x 120 cm and provided with a slot for the wires. The assembly must be somewhat higher than the maximum depth of the snow.

The plastic tubes of various lengths are inserted into holes in the rod. The constantan wire, one end of which is in the thermometric borehole at a depth of 3 m, is passed through the groove in the rod and into the highest tube in the assembly. From there it is passed through all the remaining tubes. The wire will emerge from the rod at the end closest to the smallest tube. A copper wire is then soldered to the wire at a distance of 10 cm from the end of the tube. This joint serves as a temperature-sensitive element and is placed at the boundary between the vegetation and the snow (at undisturbed sites), or between the ice and the snow (on water bodies). The remaining elements are assembled by soldering copper wires to the bends in the constantan wire at the end of each tube. All copper wires are inserted into the plastic tubes and passed through the wooden base. The wires in the base and the tubes are protected from water by an insulating solution.

A copper wire is then soldered to the other end of the constantan wire. The copper wire from the reference junction is connected to the recorder and the copper wire from other junctions is joined to a switch in the registration channel. The thermocouples are covered by a water-insulating lacquer, while the thermocouple measuring the temperature at the boundary between the vegetation and snow is protected from moisture by organic glass or AKR-7 plastic. The wires from thermocouples recording the temperature on the snow surface are enclosed in a long, thin, chloro-vinyl tube. This thermocouple must be always on the surface of the snow.

If a resistance thermometer is used for continuous measurement of snow temperature, there may be errors due to the emission of heat by the thermometer.

The ice temperature is measured in the same way. The structure, stratigraphy and density of the snow are studied throughout the entire period of formation of the snow cover. Studies of the water regime, thermal and radiation properties of snow have been described by P.P. Kuz'min (1957).

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Appendix 1

Methods of Measuring Soil Temperature*

Drilling and Excavation of Thermometric Boreholes and Test Pits and Their Maintenance

The determination of the true temperature of the soil and especially permafrost by means of boreholes is a fairly complicated task which involves special techniques and careful observation methods. Temperature measurement in boreholes is made difficult by the fact that drilling disturbs the natural temperature field. To eliminate the effect of these disturbances, it is necessary to prepare the holes in a special way and maintain them for a long time (conservation), which is not always feasible.

There may be two types of disturbances of the natural temperature field: those which can be eliminated and those which cannot. The first include disturbances resulting from the effect of drilling (the effect of the drill and the solution circulating in the hole). These become weaker with time and the temperature of the soil surrounding the hole returns to its true value.

The disturbances which cannot be eliminated result from the heat exchange due to convection in the hole and in the space between the casing and the walls. This heat exchange is the outcome of water exchange between various water-bearing horizons. Such disturbances do not weaken with time and may become even stronger. Hence it is not always possible to obtain true temperature readings. However, certain problems can still be solved in spite of disturbances; for example the position of the zero geoisotherm (the lower permafrost surface), stratification of the permafrost, geothermal gradients in the positive temperature zone at great depths, soil temperature (to within ± 1 - $2^{\circ}C$), etc., can still be found.

Geothermal observations are best done in dry holes which do not penetrate water-bearing horizons and have been drilled "dry", i.e. without the use of solutions and water.**

Holes which have penetrated water-bearing horizons or have been drilled wet must be dried, at least in the permafrost. If it is impossible to dry a hole completely, it must be cleared of the drilling solution and flushed prior to the temperature measurement.

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* Appendix to Section 4, Ch. II, Part I.

** Still more favourable conditions for temperature measurements in permafrost are created with forced air drilling (Russian editor).

In the case of long-term studies which are to be carried out over a period of many years, it is necessary to modify the top of the hole in a special way. The pit at the top is filled with a clay material; part of the hole is cased with pipes which must be free from defects. The spaces between the lengths of casing and between the casing and the wall of the hole near the top are filled with cement or packed with clay to prevent water from entering the hole. The casing protruding above the surface of the ground should be protected by a housing made of boards or other materials. The housing is filled with soil, sawdust, moss, etc. Its diameter must be three to four times larger than that of the casing. The top of the hole is plugged with a well-fitting wooden plug provided with a felt lining. A metal cover is put (screwed on) on top of the plug.

If the hole is filled with water (solution) and no measures are taken to prevent it from freezing at the end of drilling, the duration of observations will be restricted. The start of freezing in a water-filled hole left unattended depends on a n mber of factors (temperature of the permafrost, geological and hydrogeological conditions, design of the hole, method and duration of drilling) and may vary considerably within the same area. Table XXXV contains some general data concerning the start of freezing in boreholes.

If nothing is done to prevent water from freezing, it is necessary to inspect the hole continuously and at the first sign of slush ice add salt or warm water.

If it is expected that drilling will be interrupted long enough for water in the hole to begin freezing, salt should be added at the start of the interruption. A concentrated aqueous solution of table salt is used. The solution is cooled as much as possible and is then poured down the hole. Care is taken to prevent the solution from freezing at the level of the coldest soil horizon. After the addition of the solution, water must be mixed, which is done by repeatedly lowering a bailer, or some other tool, until it reaches the bottom of the hole.

In the presence of a water-bearing horizon which has not been isolated, the salt solution is usually washed out of the hole at different rates, depending on existing conditions. The salt content of the solution should be checked at least every three months and more salt should be added if necessary. This method of protecting the hold from freezing requires a great deal of salt (about 4 tons for a 100 m water column in a hole 35 cm in diameter, at a permafrost temperature of -5° to $-7^{\circ}C$). The application of this method is not always expedient and profitable.

On maintaining a hole for 6 to 12 months, the salt solution is added regularly in small amounts which substantially lowers the salt consumption. The method is designed to create an uneven salt concentration which would decrease from top to bottom in relation to the permafrost temperature.

Let us cite an example of an area where permafrost was 200 - 250 m thick, the soil temperature at the level of water in the hole was $-4^{O}C$, the water column was 420 m high, and the hole was 25 - 30 cm in diameter. The first addition consisted of 800 litres of solution prepared from 180 kg of salt. At first, additions had to be repeated every 20 - 25 days and later every 40 - 60 days. Five to fifteen days were required to restore the temperature regime disturbed by the additions of solutions. Under different geological and geocryological conditions, the possibility of adding small amounts of solution and the methods of doing so must be determined experimentally.

If on maintaining a hole for long-term studies it is not possible to drain it or lower the water level to the horizon with a positive temperature, use may be made of a method based on the displacement of water by solar oil which is added until it completely fills the hole in the permafrost horizon. The oil is added gradually until the entire estimated volume has been used up. To speed this up, it is advisable to lower the water level in the hole by pumping prior to the addition of oil.

The weight of oil is equal to the weight of the water column from the static level to the lower permafrost surface. This method may be used if

$$(H-h)\left(\frac{1}{m}-1\right)<h,$$

where h - depth of the static water level, m;

H - depth of the lower permafrost surface, m;

m - specific weight of oil.

The addition of oil has its disadvantages. For example, the temperature can be measured with electric thermometers only if oil-resistant cables are available.

To reduce the disturbances in the natural temperature field in the soil and to restore the temperature in the hole to that of the surrounding soil, it is essential to stabilize the hole. By stabilization of a hole we understand a state of rest of the air, water or drilling solution in it, which is not disturbed by forced circulation or movement of some tool or instrument. The stabilization period is the time interval which is required under given conditions to restore the temperature in the hole to that of the surrounding soil.

The stabilization period depends on a number of factors:

a) diameter of the "blanket" of soil round the hole, within which the temperature of soil or the phase composition of water in the soil have been changed on drilling, and the extent of these changes;

b) soil temperature outside the "blanket" and the difference between this this temperature and that of the drilling solution;

c) thermophysical properties and water permeability of the soil;

d) discharge into the hole of ground water, oil or gas, their yield and temperature;

e) duration of drilling;

f) method and regime of drilling, and composition, consumption and temperature of drilling solution;

g) design of hole, its diameter and depth, casing, surface structures, presence of water or drilling solution, cementation of space between casing and wall of hole, etc.

There have been no systematic studies of the required periods of return to undisturbed conditions in boreholes drilled in various parts of the permafrost region. Approximate stabilization periods in relation to the aim of temperature measurements, depth of holes and drilling methods are given in Table XXXVI, which may be used as a guide in planning the investigations and for a very rough evaluation of results obtained. It should be borne in mind that with all other conditions remaining the same, the lower the temperature of the permafrost, the shorter the stabilization period.

The stabilization periods are determined by repeated measurements of temperature in the holes.

The following relationships between the stabilization period (in days) and the difference between the temperatures of soil and the drilling fluid (in % of the initial difference) may be used for an approximate evaluation of the temperature of the soil in the positive temperature zone in holes filled with a drilling fluid:

Stabilization period	1	2	3	4	5	6	7	8	9	10	20	30	40	50	100
Temperature difference...	50	35	30	28	25	20	19	18	17	16	9.5	6.5	5	4	0

This difference may be calculated as follows:

$$P = 100 \frac{t_1 - t_s}{t_o - t_s},$$

where P - stabilization period,

t_1 - temperature of the drilling fluid when undisturbed conditions have been restored,

t_s - natural temperature of the soil,

t_o - initial temperature of the drilling fluid.

The stabilization period in the permafrost is considerably longer. In one rotary drilled hole, the temperature of the drilling fluid remained at about $0^{o}C$ for 48 days at a soil temperature of about $-4^{o}C$.

The disturbances of the natural temperature field of soils which can or cannot be eliminated are especially great in water-filled holes, and the deeper the holes, the greater these disturbances. In a deep hole which has penetrated water-bearing horizons, it is practically impossible to find the natural temperature of the soil in the upper part of the cross-section above a certain depth. This depth varies and depends on a number of factors. To obtain a reading close to the natural temperature, it is desirable to supplement the studies in a deep hole with measurements in a shallower control hole, or carry out separate measurements in frozen and unfrozen soils.

The control hole should be drilled in the same way as the main hole at a distance of 25 - 50 m from the latter and outside the zone of influence of surface structures. The following depths may be recommended for control holes: 10 - 20 m if the main hole is up to 200 m in depth; 20 - 50 m if the main hole is 200 - 500 m deep; 50 m if the depth of the main hole exceeds 500 m. The bottom of the control hole must be located in permafrost and the hole must be kept dry. On measuring the temperature in deep holes drilled dry and which did not penetrate any water-bearing horizons, the depth of the control hole may be kept at 15 - 30 m.

Separate temperature measurements in frozen and unfrozen zones in the same hole are taken when the hole penetrates a water-bearing horizon. By using the available data on the thickness of permafrost in the given area, it is best to stop the hole 5 to 10 m above the probable depth of the lower permafrost surface and measure the temperature of the permafrost. Drilling is then continued to the required depth and the soil temperature is measured below the permafrost. This method is recommended for dry holes drilled without drilling fluid. In wet drilling this method may be used in holes with a small initial diameter, as well as in the case of a prolonged interruption in drilling for one reason or another.

In boreholes drilled with the help of drilling fluids, it is expedient to determine the temperature of the thawed soil first and then the temperature of the frozen soil, having isolated the upper part of the hole from its lower part. This is best done by casing the hole. The socket of the pipe should be cemented and the cement plug should be left at some distance above the lower surface of permafrost. The section of the hole above the plug should then be completely dried.

On maintaining a hole for a long time it is possible to apply the freezing method. The clay solution is completely removed and the hole is washed out until the water in it becomes absolutely clear. The temperature is then measured in the zone of thawed soil and the hole is plugged. After a certain interval of time the column of water will freeze. The time required for natural freezing of water in the permafrost varies and is often measured in months (about 3 to 6 months in rotary-drilled holes at a permafrost temperature of up to -5°C). The ice is then drilled out but an ice plug is left 10 to 15 m above the assumed lower surface of the permafrost and the temperature is measured in the dry hole.

In boreholes drilled through frozen, fine-grained deposits without using any, or virtually no, water, it is possible to measure the temperature in the process of drilling in the section free of water. In this case drilling is interrupted briefly for periods ranging from one to ten days (the method of successive build up of the geothermal curve). The temperature is measured when the hole is 50 to 100 m deep (this must not exceed one half of the planned final depth of the hole), two hours after the removal of the drill and again at the end of the idle period. If drilling is interrupted regularly, the temperature is measured every 10 - 20 m and during idle periods. A series of successive thermograms reveals the degree of agreement between the measured and natural temperature in different horizons. It is essential to measure the temperature at least 4 to 6 times at each point. Five to ten days after the end of drilling, the temperature is measured throughout the entire depth of the hole.

In long-term temperature studies of frozen and underlying thawed soils, use may be made of electric thermometers with probes located at different depths. A probe is placed in a borehole which has been dried throughout the permafrost section. The hole is then filled with sand. (In the presence of water, electric thermometers cannot be frozen into it, because their insulation would soon deteriorate.) However, this type of equipment cannot be used for long-term studies lasting several years, because electric thermometers and thermistors available at present soon become unreliable or fail altogether and cannot be repaired or replaced in situ.

In the absence of continuous records in the form of a thermogram, the temperature must be measured in a borehole in a definite sequence and at definite intervals depending on the depth of the hole, the mode of occurrence of permafrost, etc.

The most complete information must be obtained on the layer of annual temperature fluctuations (with the help of a control hole), the central part of the permafrost mass, and the borehole section next to the lower permafrost surface. The standard depths of temperature measurements given in Table XXXVII should be regarded as a guide which may be used when the geothermal curve is a straight line.

Additional readings are taken if necessary, especially at the boundary of the zone of annual temperature fluctuations, in places of tectonic disturbances, and at the contacts of soils of different composition.*

The depth of temperature readings must be determined to within ±1 m if the hole is 150 m deep (±2 m for deeper holes). The depths of readings must be clearly marked on the cables used for lowering the thermometers and the thermistors. The markings are secured by means of string, insulation tape, soft wire or brass rings. The cables must be made of materials which do not contract on wetting or stretch on loading. The markings should be periodically checked in the course of the measurements and all corrections noted in the log book.

The depth of readings is measured from a fixed point decided upon prior to drilling and recorded in the geological account of the hole, with corrections for its location relative to the surface of the ground (zero depth).

If automatic depth meters or semi-automatic recorders are available, they should be used in accordance with the "Logging and shooting instructions" (1952).

In testpits the soil temperature is measured with slow-reading mercury thermometers as the pit gets deeper. When the pit reaches a certain depth, a hole is drilled at the foot of the north wall facing. This hole (a blast hole)** is drilled at an angle to the horizontal of 45°, is 0.44 m long, and is somewhat larger in diameter than the thermometer. A template is used to give the drill the necessary angle. The template is a wooden right-angled isosceles triangle provided with guide pins (Fig. 126).

When drilling the hole in a pit and also during temperature measurements, it is important to prevent water accumulating at the bottom of the pit from entering the hole. To achieve this the top of the hole must be a little above the bottom of the pit. The thermometer is covered with vaseline or wrapped in paper to prevent it from adfreezing to the walls of the hole. It must not be lowered to the bottom of the hole. The top of the hole is plugged with moss.

Pits intended for temperature measurements are excavated without lighting fires to remove the vegetation. At night and during long interruptions in the work, the bottom of the pit is covered with hay, snow or dry moss.

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* It is assumed that the cross-section of the borehole is known.

** The hole should be drilled as soon as possible, so that if the pit is excavated slowly, the temperature distortion will not be great (Russian editor).

The pit itself is covered with boards, rods or brush, which in turn are covered with tar paper or tarpaulin and surrounded by shallow drainage ditches.

The stabilization period of the thermometer in testpits depends on its design and inertia, the difference between its temperature and that of the surrounding medium, thermal conductivity of the medium, conditions in the pit or borehole, etc.

Methods of and Instruments for Temperature Measurements

In geocryological investigations use is made of three methods of temperature measurement based on three different principles: thermal expansion of liquids, thermoelectric effect, and changes in electrical resistivity of metals and semiconductors. Each of these methods has its advantages and disadvantages.

The instruments used in boreholes include various mercury thermometers, resistance thermometers, semiconductor and electronic thermistors, and thermoelectric devices. The type of thermometers and devices used depends on the task at hand, diameter of the borehole, its depth and content. The characteristics of various instruments and their application are given in Table XXXVIII.

Mercury thermometers are sufficiently accurate for the majority of geocryological investigations and are simpler to use than other instruments. However, they cannot be used for remote control measurements, have to be lowered into the holes and taken out at frequent intervals, and are easily damaged.

The advantages of thermocouples as well as metal and semiconductor devices include the fact that they can be used for remote control measurements and automatic readings, and can be scaled for any temperature interval with great accuracy. They have some disadvantages which will be mentioned later.

Slow-reading mercury thermometers are used for measuring the soil temperature in boreholes and excavations. An increase in the thermal inertia of a thermometer is required to prevent any change in the readings on taking it out of the hole. For this it is sufficient to slow down its heat receiver, which is accomplished by enclosing the latter (a ball of mercury, alcohol) in a jacket filled with a material of low heat conductivity or high heat capacity, or both. This material may be a mixture of small bits of cork and soot, fine copper filings mixed with soot, etc. There must be enough soot to fill the spaces between the cork particles or filings.

Thermometers with a different heat inertia are required for temperature measurements in boreholes of different depths. The greater the depth of the measurement, the higher must be the inertia of the thermometer.

Normally, use is made of psychrometric and soil thermometers. Both types have divisions for every 0.2°C and therefore measure the temperature to within ±0.1°C.

The use of maximum thermometers speeds up the measurements only in the presence of special jackets with reduced inertia. Maximum thermometers may be used only if the temperature of the outside air is below that in the hole. Otherwise arrangements must be made for cooling the thermometer prior to lowering into the hole and have some reserve of the cooling agent (ice, snow, table salt, etc.). Maximum thermometers are sensitive to jolts on lifting them to the surface and therefore often give distorted readings.*

The slowing down of a thermometer is done as follows. It is placed in a metal (brass or aluminum) or plastic tube, the inside diameter of which is 1.5 - 2 mm larger than the diameter and the length is 50 mm greater than the length of the thermometer. The walls of the metal and plastic tubes are 1 - 1.5 mm and 1.5 - 2 mm thick, respectively. Diagrams of slow-reading thermometers are shown in Figure 127.

The thermometer is wrapped in several layers of paper to prevent it from touching the walls of the tube. The paper facing the inspection window in the tube is cut out with a knife. Rubber lining or insulation tape may be used instead of paper. Any free space between the thermometer and the tube is filled with cotton wool. Having placed the thermometer in the tube, the space between the heat receiver (sensing element) and the tube is well packed with an inert mass. The slits between the tube and the body of the thermometer along the inspection window and above the thermometer are well packed with iron oxide or Mendeleev cement.

The tube must provide reliable protection for the mercury chamber of the thermometer from external pressure exerted by the water column, because any compression of the tube will damage the thermometer. A special jacket (Fig. 128) provides the necessary airtightness, which is quite important for the application of mercury thermometers in deep holes filled with water.

When processing the readings of mercury thermometers, it should be determined whether the thermometer is sufficiently slow, and whether there are any distortions in the reading resulting from pressure, and any

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* P.G. Surikov ("Kolyma", No. 10, 1960) has suggested a method of measuring the temperature in deep holes (over 1000 m deep) with maximum thermometers at any time of the year. The method consists in lowering a slightly slowed-down thermometer upside down (with the mercury drop facing upwards).

fluctuations in the height of the mercury column in the first few seconds after unscrewing the jacket. The limits of such fluctuations must be recorded in the log-book.

The action of thermoelectric thermometers (thermocouples) is based on the thermoelectric effect. The principal component of a thermocouple is the junction between wires of different metals or alloys (usually copper and constantan wires). A thermocouple has two junctions: a measuring junction and a reference junction. The copper wires soldered to both ends of the constantan wire are connected to a measuring device. The reading obtained corresponds to a certain difference in the temperature between the measuring and the reference junctions.

The emf of a thermocouple depends on the selected metals and is proportional to the difference in the temperature between the measuring and the reference junctions, but does not depend on the length and diameter of wire used. If the temperature of the reference junction is kept constant, the reading on the measuring device will indicate the temperature of the measuring couple. Usually the reference couple is kept at $0^{o}C$ by placing it in melting ice or snow. The copper-constantan thermocouples are the most commonly used because copper makes it possible to eliminate additional junctions which would be parasitic in a given measuring system.

A copper-constantan thermocouple has almost linear characteristics within a wide range of temperature measurements, which is more than adequate for the temperature field of the soil. In the interval from $0^{o}C$ to $100^{o}C$, the thermal emf of a copper-constantan thermocouple is 40 microvolt/deg.

It is simple to make such a thermocouple because copper and constantan are easily soldered together. Use is made of PShDK constantan wire 0.5 - 0.8 mm in diameter and copper wire of the same diameter covered with PVM chlorovinyl insulation. The ends of the wires are cleaned and tightly joined together for a length of 3 to 4 mm. The joint is soldered with pure tin.

Prior to making a thermocouple, the constantan wire should be heat treated because certain types of constantan are not sufficiently homogeneous and this may affect the properties of the thermocouple. By tempering the constantan wire, the internal stresses in the alloy are reduced and its structure becomes less sensitive to changes in temperature. Tempering is best done by electric current. The ends of the wire are connected to an electric circuit through a rheostat. The current in the circuit must be such as to heat the wire until it becomes dark-red within 15 or 20 minutes. The thermoelectric homogeneity of the wire is then checked. For this the ends of the wire are connected to a galvanometer and the wire is passed through a heater. The needle of the galvanometer is watched continuously, and the wire sections corresponding to strong deflections of the needle are removed because they are unsuitable for use in the thermocouples.

Copper has a high temperature coefficient of resistance and therefore the resistance of wire will vary in relation to temperature. If the thermal emf are measured by the compensation method (by means of a potentiometer), the changes in resistance of the wire will not affect the accuracy of the readings. In direct measurements of the thermocurrent with a galvanometer, which is normally done in practice, these changes will have the greatest effect when the internal resistance of the galvanometer is low. When using a high-resistance galvanometer, the resistance of the thermocouple and the input leads may be ignored.

When measuring the thermocurrent with a galvanometer, the accuracy of temperature readings is affected by changes in the internal resistance of the galvanometer. As is known, the sensitivity of a galvanometer varies and depends on the temperature. It is essential to control the temperature of the glavanometer in the field and to introduce a variable additional resistance (a temperature compensator) which will render it possible to maintain the sensitivity of the galvanometer on the same level. The maximum additional resistance must be equal to changes in the internal resistance of the galvanometer at extreme temperatures. The resistances corresponding to intermediate temperatures of the galvanometer are calculated from the internal resistance of the galvanometer and the coefficient of thermal resistance of copper. The scale of the compensator is calibrated with a Wheatstone bridge. With an increase in temperature of the galvanometer the compensation resistance is decreased and at a given maximum temperature is equal to zero. The total resistance (internal plus compensational) of the galvanometer must be constant throughout the entire given temperature range.

The measuring circuit (Fig. 129) consists of a thermocouple, connecting wires, galvanometer and temperature compensator. The compensator is usually made of some material with a low temperature coefficient of electric resistance. At the contact with copper this material must not generate a thermal emf exceeding the accuracy of the measurements. The most suitable material in this respect is manganese.

The electric circuit of the galvanometer consists of copper, brass and phosphor-bronze. Different metals and temperature gradients present in the measuring circuit are a source of parasitic thermocurrents. These are not large, but at small differences in the temperature of the thermocuple the relative error in the reading may exceed unity. To eliminate errors, all measurements should be carried out twice (the second time by reversing the current). Figure 130 shows the circuits where the parasitic thermocurrent

I_p will save the same direction in both cases, while the thermocouple current I_T will reverse its direction in the measuring device on reversing the current.

The current generated in the thermocouple is found from the following formula:

$$I_{\tau} = \frac{I_1 - I_2}{2}. \tag{1}$$

Possible errors in the determination of I_{τ} and its sign may be avoided with the help of Table XXXIX. Double readings eliminate systematic errors resulting from incorrect positioning of the galvanometer needle at the zero mark on the scale. Use should be made of galvanometers whose needles can be deflected to the centre of the scale. If the scale of the galvanometer has the zero mark in the centre, then all cases given in Table XXXIX are possible. If the scale is unilateral, the thermocurrent and its sign can be determined as in case 3 in Table XXXIX.

In the measurement of thermocurrents, the contacts between the measuring device and the thermocouple must be protected from oxidation and contamination.

A schematic diagram of a thermocouple arrangement is shown in Figure 131. There are two reference couples with equal signs (an upper and a lower). Copper wires are soldered to the constantan wire of required length at all measuring points. The free ends of the copper wires are soldered to the clamps on the switch board. Use may be made of plugs, disconnecting joints, and PMT multipoint switches. All measuring thermocouples and their wires are combined into one bunched conductor. If necessary, this conductor can be placed into a vinyl plastic tube filled with a cable filler. It should be remembered that this will increase the inertia of the thermal elements.

The galvanometer must correspond to the given accuracy of the assembly and the given temperature interval.

To correct the errors resulting from changes in the electrical resistance of the input leads and the sensitivity of the galvanometer, and to eliminate the effect of transitional resistances, use may be made of PP or PPTN-1 potentiometers in the connections between the thermocouple and the measuring device. In this way it is possible to obtain direct measurements of thermal emf rather than of thermal current.

By using a PPTN-1 potentiometer in conjunction with a copper-constantan thermocouple, the temperature can be measured to within $\pm 0.25^{\circ}C$, and to within $\pm 0.5^{\circ}C$ with a PP potentiometer. To increase the accuracy of the measurements, it is necessary to use a multipoint thermocuple (a thermal battery) rather than a single thermocouple.

The potentiometric, or the compensation method of measuring the emf is the most accurate. Its accuracy depends only on the uniformity of the emf of the normal element and the resistances incorporated into the compensating circuit.

The normal element is easily affected by loads and therefore in practice the emf is measured by means of a circuit incorporating an auxiliary battery (Fig. 131). By using the normal element, which plays an auxiliary role, and variable resistances connected in series in the circuit of the auxiliary battery, the current in the compensating circuit is regulated when the switch is in position I. The regulation is based on the compensating method. On putting the galvanometer switch into position II, the emf is measured by the method described earlier.

The temperature probes change their properties with time, i.e. they "age", and this alters their calibration. Hence it is important to keep this in mind and to check the calibration before each series of measurements. The probes which can be removed are easy to check but those intended for continuous measurements cannot be tested at all.

If two given temperatures are measured at the reference (upper) junction and two corresponding galvanometer readings are obtained, it is possible to determine the temperature of the measuring junction without a calibration curve from the following formula:

$$t = t_1 - n_1 \frac{t_1 - t_2}{n_1 - n_2}, \qquad (2)$$

where t - temperature of the lower thermocouple;

t_1 and t_2 - given temperatures of the reference thermocouple;

n_1 and n_2 - corresponding readings on the measuring device.

This method makes it possible to place the thermocouples in the hole without preliminary calibration.

This method may be used together with another method using the thermocouples to check the performance of the resistance thermometers.

To find the temperature by means of the lower junction of the thermocouple, it is necessary to carry out two consecutive measurements. In this way the temperature coefficient of the measuring device and changes in the resistance of the input leads will not affect the accuracy of the measurements to any appreciable extent. Temperature measurements with a given accuracy may be carried out by means of thermocouples using different measuring devices, the sensitivity of which is approximately the same.

Resistance thermometers. The performance of metal and semiconductor resistance thermometers is based on the ability of conductors and semiconductors to change the electric resistance in relation to temperature. If a conductor is heated, its resistance increases and it reduces on cooling. The reverse is true of semiconductors.

Metal resistance thermometers usually contain platinum, copper, nickel or iron. Most thermometers used in geocryological studies contain platinum or copper wire.

For pure metals the relation between electric resistance and temperature in the first approximation is expressed by the following linear equation:

$$R_t = R_0(1 + \alpha t). \qquad (3)$$

where R_t - resistance at temperature t;

R_0 - resistance at $0^{\circ}C$;

α - thermal coefficient of resistance of a given metal.

In the case of platinum, $\alpha = 3.94 \cdot 10^{-3}$. In the case of copper, $\alpha = 4.29 \cdot 10^{-3}$. Impurities usually lower the thermal coefficient of resistance.

It is most important to provide the metal resistance thermometers with reliable protection against moisture. The thermometers used in dry holes are enclosed in AKR-7 plastic, and those intended for use at high pressures are provided with hermetically sealed containers.

One common disadvantage of all resistance thermometers is the fact that they measure not the temperature but current or electrical resistance, which may vary irrespective of temperature, and this is not always noticeable on taking measurements. Therefore it is necessary to check the calibration of all thermometers prior to measurements and then check the readings with a mercury thermometer at least once, at the depth of the initial measurements with an electric thermometer. It is desirable to carry out such checks at the beginning, in the middle and at the end of measurements and then introduce the necessary corrections ("spread" the experimental errors).

Factory-produced electric thermometers are intended for the measurement of positive temperatures. Their use at temperatures below zero entails additional calibration and the necessity of carrying out checks with mercury thermometers. A great disadvantage of temperature measurements at great depths by means of electric thermometers is the weight of the equipment and the difficulties of transporting it.*

* The development of semiconductor thermometers (thermistors) and electronic thermometers (radiothermometers) is highly promising as far as the production of compact equipment is concerned.

In temperature measurements with the help of resistance thermometers use is made of measuring devices based on the Wheatstone bridge principle. These devices may be divided into two groups: those arranged in form of a balanced bridge and those based on the principle of an unbalanced bridge.

A schematic diagram of a balanced bridge is represented in Figure 133, where R_1 and R_2 are constant resistances, R_t is the resistance of the thermometer, R_a is the resistance of the input leads, R_3 is a variable resistance which serves to bring the bridge into balance. The latter occurs when

$$\frac{R_1}{R_2} = \frac{R_t + R_a}{R_3},$$

or (4)

$$R_t = R_3 \frac{R_1}{R_2} - R_a.$$

Since the ratio $\frac{R_1}{R_2}$ is constant, when the bridge is balanced, for every value of $R_t + R_a$ there is a definite value of R_3.

The resistance R_3 may be in the form of a resistance box or a linear resistance with a sliding block. In such cases R_3 may be found from the resistance box or from the scale with the sliding block. If R_t as a function of the temperature is known, the scale may be calibrated directly in degrees.

Figure 134 shows a balanced bridge in which the the controllable resistance is connected to two arms of the bridge. On sliding the block along the slide wire, the resistance of one arm increases and that of the other arm decreases. It is necessary to find the precise position where the connection or disconnection of knob K does not deflect the galvanometer needle. The resistance of the electric thermometer connected to the balanced bridge is found by means of the slide wire on the resistance box.

The reading obtained in this way and corresponding to the balanced state of the bridge will characterize for the electric thermometer a definite temperature of a given body.

For this type of balanced bridge, a change in the voltage of the source is of no consequence and this is an important advantage of this method. The only requirement is to have a sufficient difference in the potentials at points a and b in order to determine the balanced position with the required degree of accuracy.

In the case of the assembly shown in Figure 133, the slightest change in the variable resistance at the sliding contact will distrub the balance state of the bridge and will result in erroneous readings. This is a shortcoming of such a system. Figure 135 shows an arrangement which is free of such defects.

If we fix the sliding block in a balanced bridge of the type shown in Figure 134, we will obtain an unbalanced bridge. In a balanced bridge, a change in the temperature of the thermometer will lead to a deflection of the galvanometer needle through an angle proportional to the current passing through the galvanometer, i.e.

$$\varphi = CI_g$$

where φ - angle of reflection of the galvanometer needle;

C - coefficient of proportionality which is constant for a given instrument;

I_g - current.

The current in the diagonal line of the bridge connected to the measuring device is found from the following expression:

$$I_g = I\frac{R_1R_t - R_2R_3}{R_D(R_1 + R_2 + R_t + R_3) + (R_1 + R_t)(R_2 + R_3)},$$

where I - current in the input circuit of the bridge;

R_D - resistance of the diagonal line of the bridge;

R_t - resistance of the thermometer;

R_1, R_2, R_3 - constant resistances.

If R_D, R_1, R_2, and R_3 are constant, the current passing through the galvanometer will depend on the current I flowing into the bridge and on the resistance of the thermometer R_t. If I is constant, I_g will depend on R_t only. This makes it possible to use an unbalanced bridge for temperature measurements and at the same time for continuous observations of temperature changes. The latter factor is an advantage of this type of a system.

To check the constancy of I, it is customary to introduce a control resistance R_c into the measuring system. The indicator is set at a given spot on the scale by means of variable resistance R. This spot is then marked red. The temperature can now be measured by means of thermometer R_t. The bridge is regulated in a similar way prior to the calibration of electric resistance thermometers.

The disadvantages of unbalanced bridges include the necessity to control the current during the measurements and to check the position of the needle prior to the adjustment of the apparatus and before taking the measurments. The input current must not exceed a certain value, otherwise the Joule heat produced in the electric thermometer would distort the true temperature reading. Any excess in current is easily noticed in the course of the measurements. If the thermometer has reached the temperature of the surrounding medium, the measurements with a current exceeding the optimum value make it impossible to obtain rapid temperature readings. Once the measuring device has been connected, its needle will soon reach the reading corresponding to the given temperature but will not remain there and will slowly move in the direction of the temperature increase. The place where the needle stops corresponds to the exaggerated temperature. If the current flowing tnto the bridge does not exceed the optimum value, the needle will soon stop and the temperature reading will be close to the true temperature.

The unbalanced bridges are not linear. During the measurements the current flowing into such a bridge must be switched on and this results in a continuous production of Joule heat in the electric thermometer, which distorts the temperature readings. To prevent this, it is necessary to determine experimentally the optimum current and then make certain that it does not increase during the measurements.

The shortcoming of an unbalanced bridge is the fact that on using it, it is impossible to account for the parasitic thermoelectric currents arising in the circuit of the electric thermometer. In balanced bridges this does not occur.

A two-wire system of connecting the temperature probe with the measuring device is the simplest. Here the input leads connecting the electric thermometer with the measuring device form a part of the bridge-arm resistance which is affected by the temperature. Since the resistance of input leads is a variable which depends on the temperature of the wires, it cannot be accurately accounted for in a two-wire system. Therefore such arrangements are rarely used for temperature measurements in the field.

It is customary to use a three-wire system which makes it possible to account for the resistance of the input leads (see Fig. 135). The input leads a and b are connected not to one but to two arms of the bridge (R_3 and R_t). The balance of the bridge is found from the following formula:

$$\frac{R_1}{R_2} = \frac{R_3 + R_b}{R_t + R_a}$$

When $R_a = R_b$, the ratio R_3/R_t will remain constant and therefore the measurements may be carried out with sufficient accuracy. When $R_a \neq R_b$, the measurements will contain an error, and the greater the difference between R_a and R_b, the greater the error.

Another design of an electric thermometer consisting of two resistances R_t and R_3 is shown in Figure 134. R_t is affected by the temperature, while R_3 is not. The bridge is balanced by changing the ratio R_1/R_2 by shifting the sliding block.

To find R_t in a four-wire system (Fig. 136), it is necessary to carry out four calculations:

$$R_t + r_1 + r_3 = A;$$
$$R_t + r_2 + r_4 = B;$$
$$r_1 + r_2 = C;$$
$$r_3 + r_4 = D.$$

By solving these equations with respect to R_t, we obtain:

$$R_t = \frac{A+B}{2} - \frac{C+D}{2}.$$

A four-wire system makes it possible to omit measuring the resistance of the input leads and therefore the latter do not have to satisfy any special requirements.

The most commonly used field system is a three-wire system. A four-wire system is rarely used, in spite of its advantages, which is due to a large number of measurements and input leads required.

A thermometric system may be divided into three parts: a temperature probe, input leads, and a measuring device; each one of these may be a source of errors.

An electric thermometer has a tendency to change its characteristics ("age"). Hence, its calibration must be checked systematically and this is usually done before each series of measurements. Errors may arise particularly as a result of an increased electric load on the thermometer.

The measuring device is affected by changes in environmental conditions. The device is used to measure the air temperature from +30°C and to -40°C or lower, and must retain its parameters in this temperature range, otherwise the readings will not be accurate. Since manganese used in the constant resistances of the bridge has a temperature coefficient of electric resistance not equal to zero, each measuring device has its own temperature coefficient.

It is important to make certain that the measuring device used in the field will have a very low temperature coefficient, and that the experimental error throughout the entire temperature range will not exceed a given value. This may be achieved either by using a two-arm thermometer (see Fig. 134) or by introducing a temperature compensator into the system.

Basic designs of measuring devices recommended for temperature measurements in the field are also shown in Figure 137.

Diagram a represents a balanced bridge. The measuring device contains two-arm resistances of the bridge; two other arm resistances of the bridge are in the electric thermometer. This system does not require a temperature compensation.

Diagram b is a modification of diagram a. The measuring device includes three-arm resistances and the temperature compensator R_c. The electric thermometer is the fourth arm resistance of the bridge. The temperature compensator is made of copper wire. R_c is found as follows:

$$R_c = \frac{\alpha \text{ manganese}}{\alpha \text{ copper}} R_3,$$

where α is the temperature coefficient of resistance.

The variable resistance connected to the input circuit of the bridge makes it possible to change the current in the circuit of the bridge. At the start of the measurements, the variable resistance is switched full on to reduce the heating of the electric thermometer and is switched off after the final reading. The bridge is fed with a 4.5 volt battery.

In boreholes up to 20 m in depth, the temperature is measured with single electric thermometers and thermometric assemblies, which include several temperature probes located at different depths.

Let us examine some thermometric assemblies.

There are several, fundamentally different methods of connecting the temperature probes with the measuring device. In a two-wire connection (Fig. 133) the resistances of input leads form a part of one arm of the Wheatstone bridge, which also includes the resistance of the temperature probe. In such a system it is impossible to estimate accurately the resistance of the input leads, therefore it may be used for rough measurements only. By using thermistors as temperature-sensitive elements, this system may be used for more accurate measurements, but only if the following condition is fulfilled: the change in resistance of the input leads at extreme temperature values must be smaller than the change in the resistance of the thermistors corresponding to the given degree of accuracy of the measurement.

In another design of a two-wire system (Fig. 138), the measuring system contains the additional wire B, which makes it possible to determine the resistance and temperature distribution along the input leads.

A thermometric assembly represents a multistrand cable, in which the wires are joined together at the bottom. Each lead wire in the cable is the same length as the auxiliary wire B, and has the same temperature and the same temperature distribution along its length. One of the wires (A) is common for the entire measuring system. Metal or semiconductor elements are joined to breaks in the wire at the desired depth.

During the measurements one of the terminals in the Wheatstone bridge is constantly connected to wire A. The other terminal may be connected to a given probe by means of a switch.

If R_θ is the resistance of the thermometer, R_M is the common resistance of the measuring circuit, R_t is the resistance of the test line (A + B), R_{LC} and R_{TC} are the resistances of the input leads of the thermometer and the testing line, respectively, measured at a temperature t^oC prior to connecting the probes to the cable, then the resistance of the thermometer can be found from the following formula:

$$R_\theta = R_M - \frac{R_{LC}}{R_{TC}} R_T$$

If $\frac{R_{LC}}{R_{TC}} = 1$, then $R_\theta = R_M - R_T$.

Wheatstone bridges of the UMV or the MVU-49 type may be used as measuring devices in a two-wire system.

Two different three-wire systems may be used, depending on the design of the temperature probe.

Figure 138 shows a thermometric assembly for one-arm temperature probes. Wire A is common to the entire assembly and during the measurements is connected by means of a switch to wires B and C coming from the probe selected for the measurement.

The thermometric assembly used with a two-arm temperature probe is shown in Figure 137. The middle points of all probes are connected to wire 0, which is in turn connected to one of the terminals of the measuring device. Wires II coming from the ends of the temperature-sensitive arms of thermometers R_t, and wires I joined to the non-sensitive arms R are connected in pairs through a switch to two other terminals of the measuring device.

In the case of a three-wire system in is best to use linear Wheatstone bridges of the type shown in Figure 138.

The accuracy of temperature readings obtained with a three-wire system and the system shown in Fig. 138 is equal to ±0.05°C. This accuracy is achieved if the following conditions are satisfied:

a) the measuring device, the temperature probe and the input leads must be in working order;

b) the intermediate resistances in the places where the measuring device is connected to the input leads must correspond to the given accuracy of measurements (it is recommended to use conical connections);

c) the temperature probe and the input leads must be well protected from moisture;

d) the temperature probe must not be overheated during the measurements.

The input leads in thermometric assemblies may be made of 0.75 mm wires protected with a PMOV or a PMVG chlorovinyl insulation. The flexible PMVG wire is more convenient for portable assemblies. The individual lines of input leads should be made of wire without any soldered joints between the thermometer and the switch. This ensures a reliable protection against moisture and is important during measurements under field conditions. It is best to insulate the joint between the input leads and the temperature probe with AKR-7 plastic, which expands on polymerization and forms a tight joint with the chlorovinyl insulation of the wire.

Conical connections result in a better contact between the measuring device and the thermometric assembly (Fig. 139). On using a switch without a block and a protecting spring, the soldered joints would soon break due to frequent bending of the wire, or the contacts would become unreliable, and this would affect the accuracy of the measurements.

The thermometer assumes the ambient temperature only after a certain period of time, which depends on the difference in the temperature of the thermometer itself and that of the surrounding medium. The thermal inertia of the thermometer depends on its design, materials of construction, and thermal properties of the surrounding medium.

The heating and cooling of the thermometer follow the exponential law. In practice the rate of reaching the ambient temperature depends on the thermal inertia of the thermometer, which is numerically equal to the time of reaching 0.63 of the difference between its initial temperature t_o and the temperature of the given medium t_1.

In geothermal measurements it is best to use low-inertia thermometers. The thermal inertia of a thermometer can be lowered by reducing its heat capacity per unit length, which is done by increasing the coefficient of heat transfer and reducing the radius of the thermometer. The heat transfer coefficient, which depends mainly on the thermal insulation of the sensitive arm of the thermometer, is a factor which affects its inertia most.

The thermal inertia of a thermometer is found as follows. Two media with temperatures t_o and t_1 are prepared in Dewar flasks. The media are represented by substances the temperature of which is to be measured (air, brine, etc.). The reading corresponding to 0.63 $(t_o - t_1)$ is found from a calibration chart. The measuring device is then set at this reading. At the same time the thermometer is transferred from the medium with the temperature t_o to that with the temperature t_1 and the time required for the measuring system to be balanced is found by means of a stopwatch. This time interval will equal the thermal inertia of the thermometer. More accurate data may be obtained by carrying out several experiments at different temperature drops.

If the thermal inertia of a thermometer is known, it is possible to estimate possible errors for different temperature drops.

The calibration of electric thermometers is done to obtain a curve showing the relation between the recorded readings and the temperature of the probes. The calibration is carried out in a Dewer flask filled with melting snow or ice, water, or a cryohydrate. The electric thermometer is placed just above the bottom of the flask. The bulb of the control mercury thermometer is placed at the same depth to measure the temperature of the substance in the flask. The connection between the mercury and the electric thermometers should be such as to bring them as close as possible together during calibration. The divisions on the scale of the mercury thermometer must not exceed 0.1°C.

Three or four temperature readings are sufficient to construct a calibration curve. It is essential to measure the temperature at 0°C (melting snow or ice). The temperature above 0°C is given, bearing in mind that the reading must be taken near the end of the slide wire of the measuring device. The negative temperature is given to -30°C (Table XL). Water is used for calibrating the thermometers at temperatures above 0°C. During calibration, the substance in the flask is thoroughly mixed with the thermometers or a rod.

Prior to reading the temperature, it must be confirmed that the electric thermometer has attained the ambient temperature. The criterion for this is the stability of readings obtained with the measuring device after 5 minute intervals.

If calibration is carried out in this way, the readings corresponding to reference temperatures must form a straight line. If one of the readings does not lie on the straight line, it should be repeated by giving the calibration medium a temperature close to that of the reading.

If it is impossible to obtain a medium with a sub-zero temperature, calibration within the range of above-zero temperatures makes it possible to extrapolate the calibration curve into the negative temperature region, because the resistance of the copper wire in the electric thermometer is a linear function of the temperature, while a certain non-linearity of the entire measuring system in the temperature range of ±10°C results in an experimental error of less than ±0.1°C. If the measuring device contains provisions for a wider range of measurements (a set of additional resistances), a separate calibration must be carried out for each range.

Solid cryohydrates for the cooling medium are prepared as follows. The salt solution of given composition is placed in a Dewar flask. A thin-walled test tube (2) 26 cm in length and 2 cm in diameter is immersed into the solution and is held in place by a cork (Fig. 140). A test tube of a smaller diameter is then inserted into the first tube and is occasionally filled with solid carbonic acid. The cryohydrate solution cools down and crystallizes around the outer test tube. The crystallization process should be interrupted when there is still some liquid left, otherwise the Dewar flask may break.

In calibration, use is made of the outer test tube with a cryohydrate icicle frozen to its wall. The tube is filled with cooled alcohol or mercury. The thermometer to be calibrated and a control thermometer are then placed in it.

When calibrating a thermometer with cooling mixtures, use is made of a brass vessel with a capacity of about three litres resembling a rain gauge receiver. The upper part of the vessel is filled with a cooling mixture. A test tube containing mercury (or alcohol) is inserted into the mixture. The thermometer to be calibrated and the control thermometer are then placed in the tube. The salt solution formed when snow begins to melt runs down into the lower part of the calibrating vessel.

For the sake of convenience the calibration data is presented in the form of a table or a calibration curve (Fig. 141). The accuracy of thermometric devices is checked by placing their temperature-sensitive elements into a medium with a known temperature. The temperature of the medium is checked by a mercury thermometer in the same way as in the calibration of temperature probes. A disparity between the readings of control and test thermometers, which exceeds the given accuracy of the measurements, points to some changes in the measuring system. These must be detected and eliminated, after which the device must be calibrated again. If the defect is found in the measuring device, all temperature probes connected to it must be calibrated again.

A balanced Wheatstone bridge is checked by incorporating a standard resistance into its circuit. When packing electric devices and input leads for transportation purposes, care should be taken not to break the wires. The galvanometers should be securely locked.

Thermistors. The use of semiconductors renders it possible to reduce considerably the weight of instruments for temperature measurements in deep holes and ensures adequately accurate results.

The temperature may be measured in wet holes (i.e. containing brine, oil, slurry, etc.) and dry holes (dry drilled holes or those from which the slurry has been removed).

Measurements in wet holes are usually carried out by means of STT resistance thermometers (thermistors). When measuring the temperature in dry holes, care should be taken to prevent any excessive heat from entering the hole on installing the thermometer. The shape of the latter should facilitate a rapid heat exchange with the surrounding medium.

There are several thermistors of Soviet manufacture. The best thermistor for the geothermal measurements is the MMT-4 type with a nominal resistance of 3 - 10 kohm. Under normal circumstances the stability of thermistors and their service life are unlimited. They should not be overheated when soldering the electrodes and must be protected from falls, pressure effects and strong currents. Recommendations concerning artificial "ageing" of thermistors are ill-founded, since thermistors are "aged" during manufacture. Like other measuring devices, thermistors must be subjected to an annual inspection.

Calibration of thermistors. For each thermistor there is a specific relation between the ohmic resistance and temperature. The purpose of calibration is to find this relation within the limits of given accuracy.

The calibration is carried out in ultrathermostats which maintain a constant temperature over a long period of time.

The thermistors in groups of 30 to 50 are mounted on a wire frame measuring 120 x 70 x 60 mm (Fig. 142). The upper outlets of the thermistors are soldered to the frame. Therefore all thermistors have one common cable leading from the frame to the measuring bridge. The lower outlets of the thermistors (extending from glass insulation) are carefully covered with lacquer or enamel and connected to a switch. The latter provides an alternate connection to the resistance bridge. The number of the calibration group is engraved on the wire frame and every thermistor is assigned its own number. After calibration, the thermistors should be kept on the wire frame and taken off as required. During calibration the wire frame must be kept in the middle of the bath, 5 to 7 cm above the blades of the stirrer.

The liquid in the calibration bath must have a uniform temperature throughout. To check this, the temperature in the upper and the lower layers of transformer oil in the bath is measured with a mercury thermometer. Use should be made of special, precise mercury thermometers with 0.01 - 0.02°C divisions, or laboratory thermometers with 0.1°C divisions. In the latter case, attempts should be made to read the temperature to within 0.01°C with a magnifying glass or field glasses. Field glasses mounted on a tripod serve the purpose very well.

The thermistors are calibrated with a MVL-47 resistance bridge or an equally accurate device (in the 0.01 - 0.02 class). The MVL-47 bridge requires a galvanometer with a sensitivity of $1 \cdot 10^{-8} - 1 \cdot 10^{-9}$ amp per unit scale and a source in the form of dry 1 KS-L-3 elements.

The temperature in the room where the calibration is carried out should be the same as in the field. Any deviations of the temperature should be recorded and the calibration data corrected accordingly.

The best device for measuring ohmic resistances under field conditions is the MVU-49 DC resistance bridge or a similar bridge. By using the bridge correctly and introducing the necessary corrections, the experimental error may be reduced to 0.1% which corresponds to about 0.03°C. Before using the bridge the temperature corrections for the range of geothermal measurements should be determined experimentally.

The resistance bridge is connected to the cable by means of a flexible wire through a socket mounted on a winch (cable lock, plug, etc.). The joint between the flexible wire and the resistance bridge must be provided with a switch for alternate connection of the first and then the second thermistor when these are connected in series. The switch should clearly indicate which thermistor has been connected.

The selected type and size of the winch depend on the type of cable used and the depth of the borehole. The winches intended for geophysical work are heavy and not always suitable for measurements in deep dry holes (up to 500 m in depth). A lighter and simpler winch for a single geothermal measurement may be made if required. For holes depper than 500 m it is necessary to use a mechanically operated winch.

In dry holes it is recommended to use a three-wire cable (for two thermistors), the wires being of PVR, PTF-7 or PSM-0 type. All three wires are enclosed in a textile braiding which may be prepared in any electro-mechanical workshop. Having done this, the cable should be boiled in natural drying oil or linseed oil. While in use, the cable rubs against the walls of the hole which damages the braiding and the insulation. Therefore a cable can be lowered and lifted only 25 to 30 times. It is uneconomical and often impossible to try and repair the cable insulation.

To simplify transportation and storage problems, the temperature probe (with a section of the cable) may be made detachable. This may be achieved by means of a geophysical cable lock (Fig. 143). The cable must be protected from water. A moist cable may freeze to the wall of the hole, while a wet cable may lead to errors in the temperature measurements.

Prior to measurements, it is necessary to note the electric resistance of the cable.

A coupling for suspending a weight is fixed to the cable at a distance of 0.5 - 0.7 m from the cable lock. The weight serves to lower the cable into the hole and to stretch it. It is attached by means of two steel cables (also PVR or PTF-7 wires) and is suspended to a depth of 2 to 2.5 m below the thermistor (Fig. 144).

A balancing block from a hydrological winch is a convenient tool for use in the holes when drilling has been completed and the ends of the casing are at different depths. The use of a balancing block intended for logging operations is difficult and requires a great deal of preparation.

The geothermal observations are carried out as follows. Having transported the equipment to the experimental site, all apparatus should be examined and corrected if necessary. The sites for the winch, resistance bridge, etc. are then selected, the equipment is installed, the winch and the tripod are secured (Fig. 151)*, and the cable is passed through the balancing block. The temperature probe and the weight are attached to the cable, the block is set at zero (the position when the protecting coating of the thermistors is touching the ground), and the cable is lowered to the depth where the first measurement is to be carried out. The cable is secured, the depth is checked with a depth meter, and all pertinent data are recorded in a log book.

For a certain period of time, a thermistor lowered into a borehole acquires the ambient temperature. In practice, the time of exposure of thermistors with protective coatings can be determined in the course of measurements. At the first depth, the thermistor is given a preliminary exposure of 1.5 hours. If the measurements are carried out in winter, the thermistor, the cable and the weight are first lowered to a depth where the temperature is considerably lower than that at all subsequent depths. Condensation takes place and ice crystals are formed on the weight, the cable and the thermistor. Hence it is essential to keep the thermistor at the first point of measurement for a long time. At all subsequent points the exposure is reduced, and at the third point (if the distances between the points are not less than 10 m) usually does not exceed 10 - 20 minutes, depending on the diameter of the hole and the time constant of the thermistor.

After the required exposure of the thermistor at a given depth, the resistance bridge is connected to the winch socket and the galvanometer is set at zero. The measurement is started by selecting a resistance on the fourth decade of the bridge ("1000"). If the decade is fully switched on but the resistance turns out to be high (the needle of the bridge galvanometer is deflected), the selected resistance should be discarded, the ratio of the bridge arms reset to "1/10", and another resistance selected on the fourth decade. When one of the resistances is smaller and the next larger than the given resistance, the smaller resistance should be left on this decade and the other resistances selected on the third decade (marked "100"). The correct choice of resistances should be first checked with the switch marked "approximate" and then with the switch marked "precise". A measurement is regarded as complete if on connecting the galvanometer its needle is not deflected from the zero mark.

* Editor's note: No Figure 151 in text - presumably Figure 144.

Three measurements should be carried out at each point to find the following: the resistance of the first thermistor, the resistance of the second thermistor, and the total resistance of both thermistors. The air temperature at the time of the measurements should be recorded also. The winch and the resistance bridge should then be disconnected and the thermistors lowered to the next depth.

The most common defects of a thermistor are as follows:

a) The given resistance is continuously changing. The main reason for this may be a gap between the protective cover and the outer surface of the thermistor. In this case a new protective covering should be provided or the thermistor should be tightly wrapped in metal foil and inserted (into a 4 mm opening) in such a way as to eliminate the gap completely. The same detrimental effects may be due to moisture in the cables. In this case the cable and the thermistor should be dried and their insulation checked.

b) The temperature readings are obviously wrong. In some instances this is due to dirt (clay, slurry, etc.) on the protective covering of the thermistor, which should be examined and cleaned. In other instances wrong readings result from the thermistor being damaged (exposure to very strong current, etc.). In this case the thermistor should be replaced.

c) Two thermistors at the same depth give different temperature readings. There may be several reasons for this: inadequate exposure, defects in one of the thermistors, presence of water on the walls of the hole, etc.

The following data should be recorded in the logbook: the number of the borehole or testpit, type of cable used and its total length, numbers of thermistors used in the assembly, type of resistance bridge and its number, depth of measurements, time (date, hours and minutes), air temperature at time of measurements, resistance of first thermistor, resistance of second thermistor, and total resistance of both thermistors. All necessary corrections are also recorded (for the cable resistance, the temperature of thermistors, etc.). The temperature at a given point is then determined from the calibration curve.

Table XLI shows an example of records obtained in the field.

General Instructions for Temperature Measurements

Prior to measurements it is essential to check the measuring system, investigate the condition of the hole, and examine it for the presence of water or slurry. Not more than 4 to 6 slow-reading mercury thermometers are lowered on the same cable to different depths. Not more than two thermometers on the same cable are lowered into a deep hole. It is not recommended to lower the sets of thermometers on different cables, because the latter may become entangled.

A string made of hemp or flax and boiled in linseed oil is often used as a cable. The thermometers are attached to it at required distanced from each other. The lowest thermometer is firmly attached but it should be possible to remove all other thermometers quickly or move them to another position. Fig. 145 shows how the thermometers should be attached to the cable.

The temperature measurements in holes up to 15 - 20 m in depth are usually carried out not with one but several resistance thermometers connected up as shown in Figure 144. Single thermometers are normally used at greater depths.

Bundles of slow-reading thermometers or strings of electrical thermometers are prepared in such a way as to make sure that once in the hole all thermometers would be at the required depths. During the initial stabilization period of thermometers, the hole should be covered with a plug.

Great care should be taken to arrange the cables properly at the top of the hole. To prevent the cables from breaking and to protect their insulation, a soft pad should be provided at the point where the cables come out of the casing.

Because use is made of thermometers of varied and non-standard designs, the minimum stabilization period for each of them must be determined experimentally.

The stabilization period for remote-control apparatus is determined by a series of readings to a set degree of accuracy. The first reading is obtained after a minimum of exposure under given conditions and the following readings after 5 to 10 minute intervals. The stabilization period is considered to be adequate when the differences between successive readings are less than $\pm 0.05^{\circ}C$.

The approximate stabilization periods for thermometers are given in Table XLII.

The stabilization period of a thermometer in a borehole must be 5 to 10 times longer than the time of its thermal inertia. On moving the thermometer to a new depth, the stabilization period will always be shorter than the previous one.

In continuous temperature recording, a moving electric thermometer is not fully stabilized at the points of measurement. The rate of lowering the thermometer is based on its thermal inertia. Changes in the temperature gradient occur during continuous recording. The true soil temperature is determined after long stabilization periods of the thermometer at three different depths.

The rate of lowering the ES-16 and ES-17 electric thermometers for a continuous temperature recording may be as fast as 100 m/hr; for ES-SB, ES-SB-1, and ES-SB-2 thermometers - 150 m/hr; for ESO-2 and STT thermometers - 400 m/hr; and for ETMI-55 - 2500 m/hr.

When using only one thermometer, it is best to start the measurements from above to avoid any undesirable mixing of air at the point where the thermometer is located and to speed up the attainment of a thermal equilibrium between the thermometer and the surrounding medium. The measurements are repeated on lifting the thermometer. The above methods are recommended for all dry holes. The holes are usually cased with steel pipes. Because of a high heat conductivity of metal, the pipe distorts the natural temperature field to a certain extent, which is especially noticeable at depths of up to 2 - 3 m. Hence at such depths it is recommended to measure the temperature in holes cased with plastic pipe.

The operational procedures for resistance thermometers are as follows: During the preliminary balancing of the bridge it is essential to switch on the rheostat to its full capacity to avoid any heating of the thermometer. Having balanced the bridge approximately, the rheostat is switched off and the bridge is balanced more precisely by momentarily switching in a battery from time to time. After about half a minute, the balance is checked once more and the reading is taken from the slide wire scale. When reading the scale, the eye should be positioned in such a way as to make the two index lines on the opposite sides of the scale window coincide. The division of the scale which coincides with the above index lines corresponds to the true reading.

It is by no means essential to make the needle of the zero galvanometer coincide with the zero on the scale prior to and during the measurements when the bridge power supply is switched on. The balanced condition of the bridge in this case corresponds to arbitrary zero. In this position the switching on of the bridge power supply (by pressing a button) does not cause the needle to move relative to the scale. It is important to attain this position while balancing the bridge.

When using unbalanced bridges, the reading corresponding to a given temperature is taken directly from the scale of the galvanometer. Prior to this it is essential to check the position of the galvanometer needle relative to the scale (on switching on the bridge power supply the needle must be made to coincide with the zero mark on the scale with the help of a corrector). The bridge must be regulated by means of a control resistance.

The position of the needle of the measuring device relative to the scale is determined when it is seen that the point of the needle coincides with its reflection in the scale mirror.

For temperature measurements by means of thermocouples, it is essential to know the temperature of the reference junction. This is usually determined with a mercury (control) thermometer with divisions for every 0.1°C or less. Occasionally it is found with a resistance thermometer. The following should be borne in mind when using thermocouples: a) the temperature of the reference couple and the thermometer bulb must coincide; b) the control thermometer must be read very carefully, since any error will be included in the temperature changes; c) when the galvanometer is switched on, its needle must point to zero. To achieve the given accuracy of measurements, the temperature of the reference couple must be determined just as precisely.

On calibrating a thermocouple, the reference couple is usually placed in melting ice with a temperature of 0°C. During the measurements it is often necessary to keep the reference couple at a constant temperature t_2 not equal to 0°C. If the calibration curve has a linear dependency, the true temperature of the reference couple t is derived from the formula

$$t = t_1 \pm t_2,$$

where t_1 - temperature determined with the measuring device and the calibration curve;

t_2 - actual temperature of the reference couple not equal to 0°C.

If t_1 and t_2 have different signs, a plus sign is placed between them, and a minus sign if their signs are the same. The readings obtained in arbitrary units (in divisions of the scale of the measuring device, in ohms, or in millivolts) are converted to temperatures by means of calibration tables or curves, bearing in mind the corrections obtained with the mercury thermometers.

The results are presented in the form of a thermogram which reflects the shape of the temperature curve. In the absence of a semi-automatic recorder for a continuous temperature record, the measurements are carried out in the following order. In deep holes the apparatus is lowered to a given depth and the readings are recorded. If there are sharp differences in the temperature between two adjacent points, an additional reading is obtained at a point 10 to 20 m lower down the hole. Subsequent readings are taken at standard depths. If there are no anomalies in the distribution of soil temperatures throughout the depth and no deviations from the readings obtained on lowering the thermometers, control measurements are also taken on lifting the apparatus but only at every second or third standard depth.

If the temperature curve has a sharp inflection, i.e. there is a sharp change in the temperature gradient, the depth at which this inflection occurs is checked by measuring the temperature at intermediate depths.

If there are anomalies, their upper and lower limits and the maximum temperature deviations are determined in the same way.

If the differences between the readings obtained on lowering and lifting the thermometers exceed ±0.2°C, the measurements should be repeated after one or two standard intervals of depth and certainly at the points where the discrepancies were noticed.

On drilling deep holes it is not always possible to measure the soil temperature after a long stabilization period of the hole. In this case the recorded temperatures will solve only some of the problems. The observations are carried out gradually; the temperatures are recorded during the drilling or on pumping the water out of the hole. The temperature of the water column in the hole is measured throughout the time interval available to the investigator.

Several thermograms obtained in this way are compared and attempts are made to arrive at possible conclusions. In particular, this method may be used to find the thickness of the permafrost.

In dry drilled holes, the measurements are carried out during drilling, either during idle periods between shifts or during special stoppages which may be as long as a day. This method makes it possible to eliminate any unexpected phenomena which may arise as a result of penetration of water-bearing horizons with water under pressure. When drilling has been completed and a certain amount of time has been allowed to elapse, the temperature is measured throughout the depth of the hole.

Presentation of Temperature Measurements

The initial soil temperature data must contain the following basic information.

a) Location of the borehole (administrative region, settlement, river basin, local name of the area, etc.), geographical coordinates, absolute elevation of the top of the hole and its elevation relative to the bottom of a depression or the water level in a river or lake; the element of relief, its exposure, distance to the water edge in a river or a lake or to the crest of a steep slope, the height and steepness of slopes, the depth and width of a river or a lake; vegetation and degree of swampiness of the section, etc. The schematic plan of the hole must be on a scale 1 : 10,000 - 1 : 2,000.

b) Type of hole, method of drilling, regime and duration of drilling; work performed in the hole (casing, cementation, elimination of breakdowns, idle periods, experimental pumping of water, testing, etc.); methods of washing out the hole, temperature of fresh and used water, etc.

c) Design, depth, and diameter of the hole at various depths; various diameters of casing; number of cement rings, bridges and their depths; presence of filters; shooting of casing, i.e. use of explosives in hole to remove obstructions to allow casing to be driven, etc.

d) Geological, lithological and geocryological profiles of the hole based on core and soil samples.

e) Depth of occurrence and thickness of water-bearing and oil- and gas-bearing layers; rate of absorption or flow of fluid or gas; specific yield; pressure, static (piezometric) layer; type of water (layer, fissure, etc.); temperature on pumping; chemical composition of fluid and gas.

f) Effect of permafrost on drilling; freezing of slurry - formation of an ice plug, adfreezing of instruments.

g) Preparation of the hole for temperature measurements. The times of the following operations should be recorded: end of drilling, lowering of instruments, washing, casing, cementation, test pumping, injection, logging, etc.; duration and methods of pumping, injection and washing, stabilization period; amount of fluid poured into the hole or extracted from it; temperature of fluid entering or leaving the hole; methods of equipping and preserving the hole for temperature measurements (description of equipment and procedures used).

h) Condition of the hole immediately prior to and during measurements; total stabilization period; presence of a liquid column; level of fluid at the start and end of measurements; mineral content of fluid, presence of natural flow, gas emission, etc.; depth to which the hole is empty (whether to water level, plug, bridge, or bottom).

The measurements are recorded in a log book as shown in Table XLIII.

The data is used to construct the thermograms and the latter are then analyzed. The main purpose of detailed interpretation is the discovery of anomalies and their causes, and the direction of displacement of the thermogram relative to the geothermal curve corresponding to natural conditions. The nature of anomalies related to the peculiarities of the geological structure, emission of gas, circulation of water in the space behind the pipes, effects of recent cementation, etc., is established. The methods of investigating such anomalies are described in handbooks of geophysics.

The next step is to investigate the characteristics of the geothermal curve related to the location of the hole. For example, small and negative gradients in the upper part of the curve often result from the proximity of a talik below a river bed or a lake, or conversely from the proximity of a frozen soil mass.

It is sometimes found that in the holes filled with salinized water there is a sharp drop in the temperature gradient of the water column in the permafrost, resulting from convection and levelling of the temperature of the solution near its freezing-point. At great depths the readings of electric thermometers are often distorted owing to poor insulation. If the stabilization period of a borehole in the permafrost is short, there is either no gradient or a low gradient in the distribution of temperatures close to 0°C. Should there be a rise in temperature, the depth corresponding to this section of the geothermal curve would indicate the approximate position of the lower surface of the permafrost. If the stabilization period is brief (the temperature in the hole is above or equal to 0°C), the depth of the lower permafrost surface is reduced, but if the stabilization period is long (the temperature is below 0°C) the depth is increased. A relatively accurate position of the zero geoisotherm may be determined only by interpolation between two points of measurement. Throughout the depths where no measurements have been carried out, extrapolation should be done most carefully and only if measurements with a stable average temperature gradient over the preceding 50 - 100 m are available and the geological conditions are uniform.

If the depth of the zero isotherm and the soil temperature at the depth of 10 - 15 m are known, graphic interpolation would give the approximate average gradient and temperatures at different depths in the permafrost. These data may be used to determine roughly the extent to which the measured temperature in the hole resembles the temperature of the surrounding soil.

If a series of temperature measurements taken at a given depth at different times during the stabilization period of the hole is available, the approximate soil temperature at this depth can be found by calculations or by plotting.

The report based on geothermal observations includes the following settions.

1) A general section containing information on the method, regime and duration of drilling, stabilization period and nature of investigations in the borehole, data on the design of the hole and its condition, reasons for drilling, aim of geothermal observations, etc.

2) The main part of the report with a description of the geology and geography of the borehole site and a plan of the site at a scale 1 : 10,000 - 1 : 2,000; a generalized geological and lithological cross-section of the hole; information on the content of water, oil and gas, description of permafrost in the region and as revealed by drilling; information on the preparation of the hole for temperature measurements and its condition during the investigations.

The report must also contain detailed information on the methods of geothermal investigations: instruments used, curves showing the stabilization of thermometers, and suggestions concerning possible experimental errors. The results of temperature measurements are presented in the form of tables and curves.

The main attention in the report is devoted to the analysis of the data obtained. The thickness and continuity of the permafrost are compared with the results of measurements, the presence of and the reasons for natural and artificial temperature anomalies are explained, the degree of resemblance between measured and true temperatures is established, the geothermal gradient throughout various depth intervals is calculated and the reasons for its variations are analyzed, the gradients of frozen and unfrozen soils are established.

If the results are accurate and make it possible to make conclusions concerning the dynamics of permafrost, an appropriate analysis of the effect of changes in natural conditions in the region is then carried out.

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Appendix 2

Permafrost Record

(for testpits and boreholes)

Region....................................area.......................

Nearest settlement...

Plane table nomenclature...

Basin of river................................stream......................

Latitude......^{o}.....';Longitude......^{o}.....'. Absolute elevation (m).......

Relative elevation (with respect to benchmark, water basin, etc.)..........

Depth of exposure...........No............ Type (borehole, testpit, cleared section, natural exposure)..

...................................diameter.................................

When exposed (start, finish)..

Method used (wet drilling, forced air drilling, burning of vegetation, etc.)...............................Date of examination..................

Relief (main forms, mesorelief, microrelief)...............................

....

Exposure (north, south, etc.)...

Surface drainage (swampiness)..

....

Vegetation (forest, meadow, moss, density of vegetation, absence of vegetation)..

Snow cover, depth (cm), density and uniformity of distribution (in winter).

....

Depth of permafrost table and lower permafrost surface (m)................

....

Type of groundwater encountered..

Depth of water (m)	Steady water level (m)	Base of water-bearing horizon (m)	Thickness of water-bearing horizon (m)	Depth of temp. meas. (m)	Temp. ^{o}C	Specific yield litres/sec

Chemical composition of water...

....

Cross-section

Layer No.	Depth intervals (m)	Lithological composition, facies, cryogenic structures - continuous, layered reticular, basal	Depth of temp. meas. (m)

Temp. ^{o}C	Date of temp. meas.	Water cont. (ice cont. in % of dry or wet sample)	Depth of soil sampling (m)

Remarks. Methods of temperature measurements and soil sampling to determine water content..

Name of investigating department...

....

Date.................................... Signature.......................

Additional information
(for deep holes)

1. Design of hole..
....
2. State of hole..
....
3. Hydrogeological and geothermal observations in the holes (sampling, stabilization period, etc.)..
....
4. Other information...
....
5. Additional information concerning the cross-section.................
....

Date.................................... Signature.......................

Table I. Cryogenic texture of alluvial, eluvial and talus deposits undergoing seasonal freezing
(Compiled by E.G. Katasonov on the basis of the survey carried out in the northwestern part of the Yakut A.S.S.R.)

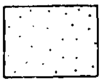
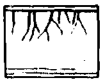
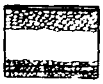
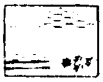
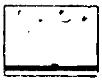
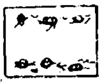
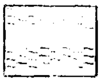
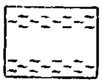
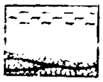
Cryogenic structure of the layer of seasonal freezing			Meso-relief	Genetic Soil type	Soil composition	Ice content in % of dry weight	Moisture characteristics	Thickness of moss-lichen cover
Schematic drawings of layers of seasonal freezing and thawing	Thickness of ice inclusions	Cryogenic textures observed in cross-sections						
	-	Massive	Channel banks on a low flood plain and embankments on high flood plain	Alluvial	Fine and medium sand, sandy loam without humus	5-20	Dry	Usually absent
	From fractions of mm to 1-2 mm	Streaky with thin horizontal layers in the lower part of the layer of seasonal freezing	Depressions between ridges and flat levelled sections on low flood plain	Alluvial	Sandy loam, clay loam (occasionally containing sand and some humus)	20-30	Swampy in rainy season only	Thin (1-2 cm) absent on some sections
	From fractions of mm to 0.5 cm From 0.5 to 3 cm	Reticulate (network), with horizontal ice layers in the lower part of the layer of seasonal freezing	Flat levelled sections on high flood plain	Alluvial	Clay loam, occasionally sandy loam and fine sand silty, and sometimes peaty	50-60	Permanently swampy	Continuous, 4 to 6 cm thick
	From fractions of mm to 1 mm From 0.5 to 3 cm	Fine lenticular, with even ice layers in lower part of layer of seasonal freezing (in peaty deposits), and mixed (reticulate with slanting ice lenses in gleyed clay loam	Depressions on high flood plain and terraces above flood plain	Alluvial	Peaty clay loam, silty peat, occasionally underlain by dark gleyed clay loam, containing few organic remains	60 - 100	Very swampy	Continuous, 6-8 cm thick
	From 1-2 mm to 3-5 cm From 1-5 cm	Pockets and around vegetative remains, large horizontal ice layers in the lower part of layer of seasonal freezing	Filled with oxbow lake sediments	Peaty-swampy	Peat	Over 100	Permanently wet	Continuous, 8-10 cm thick and more
	From fractions of mm to one mm	Crusty	Steep slopes composed of coarse material	Eluvial	Grass, rock waste with some silty clay loam	10-20	Dry	Not continuous
	From fractions of mm to 1 mm From fractions mm to 2 mm	Fine lenticular on top, crusty at the bottom	Fairly dry slopes of medium steepness with bedrock close to surface	Eluvial-talus	Clay loam and sandy loam with num. rock frag., at times sandy	20-40	-	Continuous, 2-3 cm thick
	From fractions of mm to 1 mm	Fine lenticular	Dry slopes with bedrock at great depth; talus cones at the foot of slopes facing south	Talus	Clay loam and sandy loam containing rock fragments	30-50	Dry	Continuous, 2-3 cm thick
	From fractions of mm to 0.5 cm	Coarse lenticular	Relatively dry slopes with bedrock at great depth; talus cones at the foot of slopes facing north	Talus	Same	40-60	Swampy	Continuous, 3-5 cm thick
	From fractions of mm to 0.5 cm From 0.5 - 5 cm	Lenticular at top of layer; reticulate with slightly wavy layers of ice at the bottom of layer	Gentle swampy slopes with bedrock at great depth	Talus	Clay loam and sandy loam with low humus content and small amount of rock fragments	60, often over 100	Very swampy	Continuous, over 5 cm in thickness

Table II

Seasonal thawing of sandy loam and clay loam deposits
in Yakutsk and silty clay loam in Igarka

Yakutsk area		Igarka area	
Date	Thawing, %	Date	Thawing, %
May 1 - 5	10	2nd ten days in May	Beginning of thawing
May 5 - 10	20	End of May	10 - 15
May 10 - 15	35	End of June	30 - 40
May 15 - June 1	45	End of August	50 - 60
June 1 - 15	60	2nd ten days in September	60 - 80
June 15 - July 1	80	Beginning of October	95 - 100
July 1 - 15	85	2nd ten days in October	100
July 15 - August 1	90		
August 1 - September 15	95		
September 15 - October 1	100		

Table III

Characteristics and schematic drawings of cryogenic textures

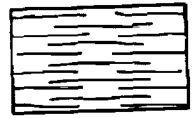
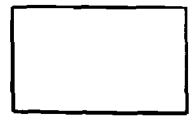
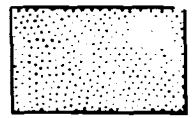
Soil	Texture	Characteristics	Schematic drawing
Clayey	Fine-layered	Thickness of ice layers: from fractions of mm to 1 mm. Thickness of mineral layers: 1 - 2 mm	
"	Medium-layered	Thickness of ice layers: 2 - 10 mm. Thickness of mineral layers: 2 - 5 mm	
"	Coarse-layered	Thickness of ice layers: 10 mm (in the upper part) to 400 mm (in the lower part of the layer). Thickness of mineral layers: 20 - 30 mm (in the upper part) to 1000 mm (in the lower part of the layer)	
"	Massive	Visible ice inclusions absent	
Sandy	"	Same	

Table IV

Cross-section I

No. of layer	Soil	Thickness of layer, cm	Cryogenic texture	Lower boundary of layer of seasonal thawing	
				Present	Maximum
1	Clayey	3-5	Fine-layered	At the base of layer 5	At the base of layer 5
2	"	10-20	Medium-layered		
3	"	35-50	Fine-layered		
4	"	100-120	Massive		
5	"	10-20	Medium-layered		
6	"	1,000-1,400	Coarse-layered		
7	"	1,000-1,500	Massive		

Table V

Cross-section II

No. of layer	Soil	Cryogenic texture	Lower boundary of layer of seasonal thawing	
			Present	Maximum
1	Clayey	Fine-layered	At the base of layer 5	At the base of layer 7
2	"	Medium-layered		
3	"	Fine-layered		
4	"	Massive		
5	"	Medium-layered		
6	"	Massive		
7	"	Medium-layered		
8	"	Coarse-layered		

Table VI

Cross-section III

No. of layer	Soil	Cryogenic texture	Lower boundary of layer of seasonal thawing	
			Present	Maximum
1	Clayey	Fine-layered		
2	"	Medium-layered		
3	"	Fine-layered		
4	"	Massive		
5	Sandy. Over 5 cm in thickness	Massive		
6	Clayey	Medium-layered		
7	"	Massive		
8	"	Medium-layered	At the base of layer 8	At the base of layer 8
9	"	Coarse-layered		

Table VII

Cross-section IV

No. of layer	Soil	Cryogenic texture	Lower boundary of layer of seasonal thawing	
			Present	Maximum
1	Clayey	Fine-layered		
2	"	Medium-layered		
3	"	Fine-layered		
4	"	Massive		
5	Sandy. Over 5 cm in thickness	Massive	In the sand layer?	
6	Clayey	Medium-layered		
7	"	Fine-layered		
8	"	Massive		
9	Sandy. Over 5 cm in thickness	Massive	In the sand layer?	
10	Clayey	Medium-layered		
11	"	Fine-layered		
12	"	Massive		
13	"	Medium-layered	At the base of layer 13?	At the base of layer 13
14	"	Coarse-layered		

Table VIII

Cross-section V

No. of layer	Soil	Cryogenic texture	Lower boundary of layer of seasonal thawing	
			Present	Maximum
1	Clayey	Fine-layered		
2	"	Medium-layered		
3	"	Fine-layered		
4	"	Massive		
5	Sandy	Massive	In layer 5	In layer 5
6	Clayey	Coarse-layered		

Table IX

Cross-section VI

No. of layer	Soil	Cryogenic texture	Lower boundary of layer of seasonal thawing	
			Present	Maximum
1	Clayey	Fine-layered		
2	"	Medium-layered		
3	"	Fine-layered		
4	"	Massive	In the upper part of layer 5 or on the boundary between layers 4 and 5	
5	"	Medium-layered		

Table X

Depth (in m) of layer of annual fluctuations of temperature (z) in relation to mean annual soil temperature (t_m) and coefficient of heat conductivity (K)
(after Kudryavtsev)

Mean annual soil temp. °C	Coefficient of heat conductivity (K)								
	0,12	0,01	0,0081	0,0064	0,0049	0,0036	0,0025	0,0016	0,0009
0,5	9,4	8,5	7,6	6,8	5,9	5,1	4,2	3,4	2,5
1,0	13,4	12,1	10,9	9,7	8,5	7,3	6,1	4,9	3,6
1,5	15,7	14,3	12,9	11,4	10,0	8,6	7,2	5,8	4,3
2,0	17,4	15,8	14,2	12,7	11,1	9,5	7,9	6,4	4,7
3,0	19,7	18,0	16,9	14,4	12,6	10,8	9,0	7,2	5,4
4,0	21,4	19,5	17,5	15,6	13,6	11,7	9,7	7,8	5,9
5,0	22,7	20,7	18,6	16,5	14,5	12,4	10,3	8,3	6,2
6,0	23,8	21,6	19,5	17,3	15,1	13,0	10,8	8,7	6,5
7,0	24,7	22,4	20,2	18,0	15,7	13,5	11,2	9,0	6,7
8,0	25,4	23,1	20,8	18,5	16,2	13,9	11,6	9,3	6,9
9,0	26,1	23,8	21,4	19,0	16,6	14,2	11,9	9,5	7,1

Table XI

Factor $(1 - \frac{1}{f})$ in relation to the depth (h), the density (p) and the coefficient of heat conductivity (K) of the snow cover

ρ	K	Depth of snow cover (h) in m									
		0,1	0,2	0,3	0,4	0,5	0,6	0,7	0,8	0,9	1,0
0,075	0,0010	0,094	0,181	0,259	0,329	0,398	0,451	0,503	0,551	0,597	0,632
0,110	0,0015	0,081	0,155	0,224	0,228	0,345	0,400	0,447	0,491	0,532	0,572
0,150	0,0020	0,071	0,136	0,197	0,253	0,306	0,355	0,400	0,442	0,482	0,518
0,190	0,0025	0,064	0,123	0,178	0,230	0,279	0,324	0,367	0,407	0,445	0,480
0,225	0,0030	0,058	0,113	0,164	0,213	0,259	0,302	0,343	0,381	0,416	0,450
0,250	0,0035	0,054	0,105	0,153	0,198	0,242	0,282	0,321	0,357	0,392	0,425
0,300	0,0040	0,051	0,098	0,143	0,186	0,227	0,267	0,303	0,338	0,371	0,403
0,340	0,0045	0,048	0,093	0,136	0,178	0,216	0,254	0,289	0,323	0,356	0,386
0,380	0,0050	0,045	0,088	0,130	0,169	0,206	0,242	0,277	0,309	0,341	0,371
0,415	0,0055	0,043	0,084	0,124	0,161	0,197	0,232	0,265	0,297	0,327	0,356

Table XII

Field classification of soils (after V.V. Okhotin)

Soil type	Feeling on rubbing soil between fingers	As seen by naked eye and through lens	State when dry	State when wet	On rolling a wet sample	On cutting a wet sample with a knife	Other characteristics
Clayey	Sand particles not felt; small lumps crushed with difficulty	Sand grains not visible	Hard and lumpy	Viscous, plastic, sticky (can be spread)	Formation of long thin (0.5 mm) strings	Has a smooth surface without visible sand grains	Leaves a bright fine mark when dry
Clay loam	Sand particles felt; small lumps crushed more easily	Sand grains clearly visible against a background of fine powder	Lumps and pices not hard. On hitting with a hammer, disintegrate into small fragments	Not very plastic or sticky	The strings thicker and shorter than above	Noticeable presence of sand grains	Leaves a dull mark when dry. The mark is deeper and wider
Silty clay loam	Sand rarely felt, lumps are easily crushed	Very little sand, fine particles of silt	As above	Not very plastic or sticky	Long strings cannot be obtained, they break on rolling	Rough surface	As above
Silty	"Dry powder feeling"	Little sand, numerous particles of silt	Lumps disintegrate very easily	Floating soil	As above	—	—
Sandy loam	Sandy particles predominate; lumps are easily crushed	Sand grains predominante over clay particles	Lumps easily disintegrate on hitting	Not plastic	Almost impossible to roll a string	—	—
Sandy	Clay particles not felt	Sand grains only	No cementation	Not plastic	Strings cannot be rolled	—	—
Gravelly	Numerous particles > 2mm. If more than 50% of specimen consists of such particles, soil is termed gravel	—	Loose	—	—	—	—

Table XIII

Cryogenic textures of consolidated, semiconsolidated and eluvial deposits
(Compiled by E.M. Katasonov)

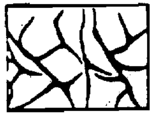
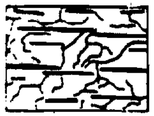
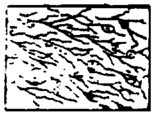
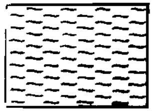
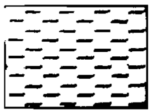
No.	Schematic drawing of cryogenic texture	Thickness of ice inclusions, mm	Cryogenic texture	Deposits displaying this texture	Genetic type, variety	Moisture cont. % dry weight
1		Up to 80, sometimes thicker	Fissured	Deposits without bedding, mainly consolidated rocks	Not identifiable	3-5 to 30
2		Up to 10, sometimes 20	Sheet-fissured	Bedded sedimentary deposits	As above	As above
3		1 - 50	Crusty	Badly weathered deposits	Cryogenic eluvium	25 - 50
4		Ice predominates	Basal	Badly weathered deposits of what used to be a layer subject to seasonal thawing	Cryogenic surface eluvium	70 - 100 or more
5		0.1 - 2 or 3	Fissured branching	Water saturated deposits from beneath river beds, etc.	Cryogenic eluvium from bottom horizons	20 - 40

Table XIV

Cryogenic textures of talus deposits
(Compiled by E.M. Katasonov)

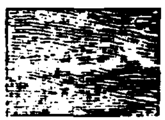
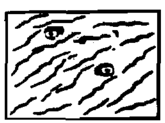
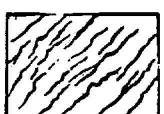
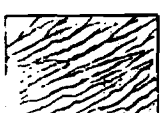
No.	Schematic drawing of cryogenic texture	Thickness of ice inclusions, mm	Cryogenic texture	Material in which this texture is formed and under what conditions	Facies displaying this texture*	Moisture content, % of dry wt
1		Up to 1	Fine, lenticular	In clay loam with low or uniform moisture content	Steep, relatively dry slopes covered with sod	20 - 40
2		5 - 6	Coarse, lenticular	In sufficiently wet clay loam and sandy loam	Gentle, relatively dry slopes covered with sod	30 - 60
3		30 - 35	Gentle wavy - reticulate (network)	In peat-free supersaturated clay loam, sandy loam and silty sand	Gentle, very wet slopes with or without sod	60 - 140
4		Up to 60	Banded (streaky)	In supersaturated clay loam containing rock fragments	Swampy slopes	100 - 500 and higher
5		Up to 20	Crusty	In the products of weathering of bedrock (ice covers)	Steep and gentle slopes composed of consolidated rock	15 - 40

* The contents of this column require further study (Russian editor).

Table XV

Cryogenic textures of perennially frozen alluvial deposits

(Compiled by E.M. Katasonov)

No.	Schematic drawings of cryogenic textures	Thickness of ice inclusions, mm	Cryogenic texture	Deposits in which the given texture is formed	Genetic type of deposit	Facies in which the given texture is formed	Moisture content, % of dry wt
1		Up to 1 - 1.5	Wavy, inherited	Silty, thin-layered, relatively dry sandy loam	River bed alluvium	Deposits on river shallows	20 - 30
2		Up to 5 - 8	Cross-lenticular	Silty sand occasionally containing gravel and pebbles		Deposits adjacent to the river channel	40 - 60
3		Up to 1.0	Cross-lenticular	Silty clay loam and sandy loam (bottom horizons)	Oxbow lake alluvium	Deposits in oxbow lakes constantly filled with water	60 - 80
4		Up to 20 - 25	Cross-laminated	As above		As above	70 - 80

...continued

Table XV continued

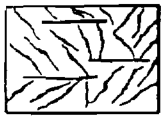
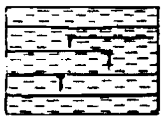
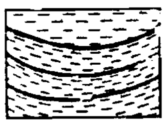
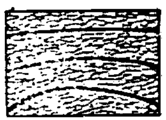
No.	Schematic drawings of cryogenic textures	Thickness of ice inclusions, mm	Cryogenic texture	Deposits in which the given texture is formed	Genetic type of deposit	Facies in which the given texture is formed	Moisture content, % of dry wt.
5		Up to 15	Mixed (feathery)	Clay loam and sandy loam which froze as bottom deposits and as a layer subject to seasonal thawing	Flood plain alluvium	Periodically drying up floodplain depressions	45 - 80
6		Up to 10	Horizontal, laminated, lenticular	Loess type of clay loam in the active layer, almost free of peat		Middle floodplain	35 - 55
7		Up to 20	Concavo-parallel-laminated lenticular	Peaty clay loam and sandy loam in the active layer		High floodplain with polygonal microrelief	50 - 70
8		Up to 20	Concavo-parallel-laminated lenticular	Peat-free clay loam, sandy loam and fine sand at the bottom of the active layer		Wet meadow and secondary water basins	70 - 100 and higher

Table XVI

Genetic classification of continental Quaternary deposits
(After E.V. Shantser)

Group and paragenetic series	Genetic type
Eluvial series	Eluvium Soil
Slope series	Talus accumulations Accumulations due to soil creep Accumulations due to solifluction Talus
River bed deposits	Alluvium Proluvium
Lacustrine deposits	Lacustrine deposits in general Chemical precipitates (salts)
Organic swamp deposits	Peat formations
Glacial series	Glacial deposits (moraines) Fluvio-glacial deposits Limno-glacial deposits
Eolian series	Eolian sand Eolian loess

Table XXI

Specific weight and freezing temperature of aqueous solutions of table salt of different concentrations

Salt conc.		Spec. wt. of soln. kg/litre at 15°C	Freezing temp. of soln. °C	Salt conc.		Spec. wt. of soln. kg/litre at 15°C	Freezing temp. of soln. °C
%	kg/m^3			%	kg/m^3		
0.1	1	1.00	0	14.9	175	1.11	-11.0
1.5	15	1.01	-0.9	16.2	193	1.12	-12.2
2.9	29	1.02	-1.8	17.5	206	1.13	-13.6
4.3	45	1.03	-2.6	18.8	230	1.14	-15.1
5.6	59	1.04	-3.5	20.0	250	1.16	-18.2
7.0	75	1.05	-4.4	22.4	290	1.17	-20.0
8.3	90	1.06	-5.4	23.1	300	1.17	-21.2
9.6	106	1.07	-6.4	23.7	320	1.18	-17.2
11.0	124	1.08	-7.5	24.9	331	1.19	- 9.5
12.3	141	1.09	-8.5	26.1	352	1.20	- 1.7
13.6	157	1.10	-9.8	26.6	360	1.20	0

Table XXII

Characteristics of clay solution for drilling frozen soils

Type of frozen soil	Viscosity, sec	Ease of giving up water cm^3/30 min	Thickness of clay crust, mm	Shear strength mg/cm^2	Stability	Residue per day, %	Sand content %	Specific wt gm/cm
Continuous frozen beds with layers or lenses of ice:								
Silty sands	40	⩽20	3-4	⩽30	0.03	⩽4	⩽5	1.30
Sand with pebbles	40	⩽20	3-4	⩽30	0.04	⩽4	⩽4	1.30
Accum. of pebbles	50	⩽20	3-4	⩽40	0.04	⩽4	⩽4	1.40
Dry sandy soils	40	⩽10	2-3	⩽30	0.03	⩽3	⩽5	1.30
Sand with liquid water	50	⩽10	2-3	⩽40	0.03	⩽3	⩽4	1.40
Coarse-grained sand, clay, soft sandstone, limestone	20	⩽20	2-3	⩽30	0.02	⩽3	⩽4	1.20

Table XXIII

Bit type for different soils

Bit	(Soil category (with respect to ease of drilling)					
	III	IV	V	VI	VII	VIII
Rib bits:						
KR-1	+	+	+	-	-	-
KR-2	+	+	+	-	-	-
Multicutter bits:						
MR-2	-	-	+	+	+	+
MR-2-III	-	-	-	+	+	+
MR6-1	-	-	-	+	+	+
MR6-16	-	-	+	+	+	-
TsKB	-	-	-	+	+	+
BK-8M	-	-	-	+	+	+
TP-3	-	-	-	+	+	+

Table XXIV

Parameters of efficient drilling regime

Bit	Soil category	Load on one cutter, kg	Peripherical speed of bit, m/sec	Consumption of soln. for each cm of bit diam. litres/sec
KR-1; KR-2	III IV V	20-25 25-30 30-35	0.7-1.2 1.9-1.4 1.0-1.6	8-10 10-12 10-14
MR-2	V VI VII	50-75 75-100 100-175	0.8-1.2 0.8-1.0 0.7-0.9	>14 >12 >10
MR-2-NP	VI VII VIII	75-100 100-125 125-135	0.8-1.0 0.7-0.9 0.6-0.8	>12 >10 >10
MR 6-1	VI VII VIII	50-60 55-65 55-65	0.8-1.0 0.8-0.9 0.7-0.9	>12 >10 >10
MR 6-16	V VI VII	40-55 50-60 55-65	0.8-1.2 0.8-1.0 0.7-0.9	>14 >12 >10
TsKB; BK-8M; TP-3	VI VII VIII	120-150 130-160 150-170	1.2-1.4 1.0-1.2 0.9-1.1	>12 >10 >10

Table XXX

Comparison of main classification schemes of liquid groundwater in the permafrost region

N.I. Tolstikhin 1941		
Supra-permafrost water	Water in the active layer	
	Intermediate water	Water beneath river beds
	Water in perennial taliks	Water beneath lakes, in alluvial fans, etc.
Intra-permafrost water	Water supplied by supra-permafrost horizons	
	Water supplied by sub-permafrost horizons	
Sub-permafrost water	Water close to frozen zone: stratal, fracture, karst water, with low or even negative temperature	
	Deep-seated water: stratal, fracture, karst, fracture-vein water	

Table XXX (cont.)

I. Ya. Baranov 1940			
Water in unconsolidated Quaternary deposits	Alluvial water	Common water	Water in local taliks or in valleys in the presence of permafrost islands; free-flowing or under slight pressure
		Water related to permafrost	Supra-permafrost water, temporarily (seasonally) under pressure. Supra-permafrost taliks are fed by atmospheric water or water from great depths
			Intra-permafrost water, under pressure. Water in horizontal or cross-cutting taliks (rising flows)
			Sub-permafrost water beneath valleys, under pressure. Connected by taliks with other types of water
	Talus and eluvial water		Seasonal water fed by seepage or inflows from bedrock. Very similar to supra-permafrost, alluvial water. Free-flowing occasionally under pressure
Water in bedrock: common and sub-permafrost water	Stratal water, usually in sedimentary rocks	Free water and water under pressure	
	Stratal-fracture water in sedimentary and metamorphic rocks	Free water and water under pressure; concentrates in the form of horizons	
	Karst water in carbonate rocks	Circulates along fissures and various cavities	
	Fracture-vein water	Free water and water under pressure, circulates along fissures of various types	

Table XXX (cont.)

<table>
<tr><th colspan="2">A.M. Ovchinnikov
1954</th></tr>
<tr><td>Vadose water</td><td>Water in the active layer</td></tr>
<tr><td>Groundwater</td><td>Supra-permafrost water</td></tr>
<tr><td colspan="2">Intra-permafrost water</td></tr>
<tr><td>Artesian water</td><td>Sub-permafrost water</td></tr>
</table>

<table>
<tr><th colspan="5">V.M. Ponomarev
1953</th></tr>
<tr><td rowspan="4">Ground-water</td><td rowspan="4">Water interacting with the hydrographic system</td><td colspan="2">Water in the seasonally thawed layer (vadose water)</td><td rowspan="2">Above permafrost</td></tr>
<tr><td colspan="2">Water above the local base level of erosion</td></tr>
<tr><td rowspan="2">Water in sediments on flood plain terraces; water beneath river beds</td><td>Does not freeze solid throughout the year</td><td>Above and below permafrost, and in open taliks</td></tr>
<tr><td>Freezes solid in winter</td><td>Above permafrost</td></tr>
<tr><td colspan="4">Water of marine origin</td><td>Above permafrost</td></tr>
<tr><td rowspan="3">Artesian (stratal and stratal fracture water</td><td colspan="3">Water interacting with the hydrographic system</td><td>Below permafrost and in open taliks</td></tr>
<tr><td colspan="3">Water interacting with the sea</td><td>In permafrost and below it</td></tr>
<tr><td colspan="3">Water interacting with the hydrographic system</td><td>Below permafrost and in open taliks</td></tr>
<tr><td>Fracture and karst water</td><td colspan="3">Water interacting with the sea</td><td>In permafrost and below it</td></tr>
</table>

Table XXX (cont.)

A.I. Kalabin (for the northeastern part of the USSR) 1957	
Supra-permafrost water (ground and soil water)	Water in the active layer in interfluves in mountain areas
	Water in the active layer on plains and lowlands
	Confined taliks in river valleys fed by supra-permafrost horizons
	Confined taliks in river valleys fed by various sources
	Confined taliks in alluvial fans and in terraces in alluvial valleys
	Sub-lacustrine taliks
Intra-permafrost (intrastratal) water	Open taliks: stratal-pore, fracture, fracture-karst water
	Water in stratal river valleys
Sub-permafrost (artesian, intrastratal) water	Water below and near the lower permafrost surface: stratal-pore, stratal-fracture, fracture-vein, and vein-karst water
	Deep-seated, stratal-fracture, fracture, and fracture-karst water
	Hot and warm water; coastal zone; continental water

Table XXXII

Symbols for black and white geocryological maps, cross-sections, and profiles

I. Types of frozen soils

1) Genetic origin

sg	Syngenetic
eg	Epigenetic
pg	Polygenetic

2) Composition (shown in colour on large-scale maps)

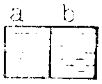
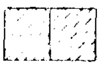
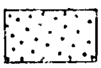
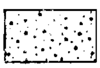
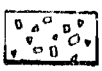
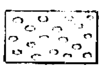
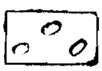
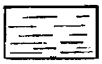
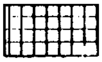
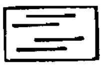
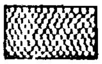
a b 	Clay: a-heavy; b-light
a b 	Clay loam: a-heavy; b-light
a b 	Sandy loam: a-heavy; b-light
	Coarse sand
	Fine sand
	Mixed sand
	Gravel
	Rock waste
	Pebbles
	Boulders
	Heaps of rock
	Peat

Table XXXII (cont.)

	Ice in soil
	Water (shown in the profile)
	Bedrock (composition shown by symbols, age by numbers)

3) Cryogenic structure and texture of frozen soils (for cross-sections, columns and profiles)

	Massive (monolithic) structure
	Horizontal bedding
	Cross-bedding
	Reticulate (network) texture
	Crust texture
	Scaly texture

II. <u>Occurrence of frozen soils</u>

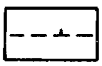
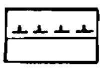
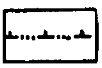
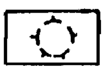
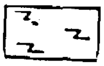
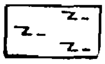
	Boundary of zone of seasonally frozen soils
	Boundary of permafrost region
	Boundaries of latitudinal geocryological zones
	Boundaries of geocryological areas
	Islands of unfrozen and thawed soil in frozen soil
	Islands of frozen soil in unfrozen and thawed soil
	Pereletoks

Table XXXII (cont.)

	Sections where the layer of seasonally frozen soil reaches the permafrost
	Sections where the layer of seasonally frozen soils does not reach the permafrost
	Permafrost table occurring at great depths
(0,5-1,5)	Depth of seasonal freezing (minimum, average, maximum, m)
(100)	Thickness of permafrost or the depth of its lower surface, m

III. Physical properties and the state of permafrost

-1°	Isotherms (-1°C, etc.)
• -1°	Temperature at a single point
$W_0 \cdot 20\%$	Total moisture content (ice content) of soil in % of weight of soil with a natural moisture content.
$W_H = 4,0\%$	Amount of unfrozen water
$i = 40\%$	Ice content by weight
$i' \cdot 10\%$	Unit ice content

IV. Cryogenic and postcryogenic formations

1) Fossil ice

	Congelation* ice of all types
	Segregation ice
	Ice wedges and re-current ice veins

* From P.A. Shumskii's classification, 1955

Table XXXII (cont.)

	Injection ice
	Sedimentary (buried) ice
	Condensation ice

2) Soil subsidence

	Of thermokarst origin (local, polygonal)
	Of mixed origin

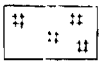
3) Soil heaving

	Seasonal frost mounds (mineral)
	Seasonal frost mounds (peaty)
	Perennial frost mounds
	Areas of seasonal heaving
	Areas of perennial heaving
	Heaved section
	Bump on the road (due to heaving)
	"Drunken forest"
	Peat bog with flat-top hummocks
	Peat bog with large hummocks

Table XXXII (cont.)

4) Polygonal formations

 Spot medallions, networks, etc.

 Polygonal ridges

 Cryptopolygonal formations

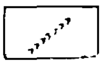
 Sinkholes and hummocks

 Baidzherakhi (cemetary mounds)

5) Solifluction formations

 Solifluction slopes

 Solifluction tracts

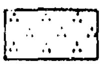
 Solifluction ridges

 Solifluction altiplanation terraces

 Landslides

 Dells

 Rock streams

 6) Nivation

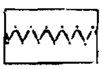
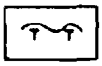
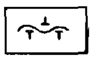
 7) Thermal abrasion on river banks and the shores of water basins

Table XXXII (cont.)

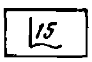
V. Hydrogeological indicators in the permafrost

1) Subsurface water

 Supra-permafrost water

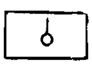
 Intra-permafrost water

 Sub-permafrost water

 Depth of occurrence of groundwater, m

 Groundwater level, m

2) Sources of groundwater

 Constant

 Seasonal (in summer)

 Seasonal (in winter)

 Migrating

3) Icings

 River icings

 Groundwater (ground) icings

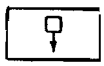
 Mixed

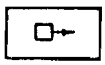
 Ice sills (large perennial icings)

Table XXXII (cont.)

VI. Special symbols

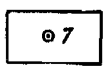
 Heaving of structures

 Settlement of structures

 Sliding of structures

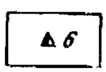
 Cracks in structures

VII. Other symbols



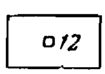
Borehole

Testpit



Excavation

 Outcrop with ice weins



Point of sampling

Table XXXV

Approximate duration of the period prior to freezing of fresh water or drilling mud in a hole which is left completely undisturbed (in days)

Temperature of soils penetrated by the hole	Method of drilling and drilling regime			
	Core drilling			Rotary drilling depths exceeding 500 m
	Drilled without fluid	Drilled using small quantities of fluid for short periods of time	Drilled using large quantities of fluid for long periods of time	
0 - -1	1/2 - 2	1 - 10	2 - 15	20 - 30
-1 - 5	1/12 - 1/4	1/2 - 1	1/2 - 22	5 - 20
-5 - -15	1/48 - 1/12	1/48 - 1/6	1/12 - 1/4	1/2 - 2

Table XXXVI

Approximate stabilization period of boreholes (in days) depending on the aim of investigations and local conditions (based on experimental data)

Drilling method	Use of drilling fluid	Depth of hole, m	Aims of investigations									
			Permafrost thickness		Alternation of warm and cooled horizons		Approx. gradients and natural temp. of soils to within ± 1-2°C (in zone where temp. above 0°C)		Natural soil temp. to within ±0.5°C in the zone where the temp. are			
									Above 0°C		Below 0°C	
			State of hole during stabilization period									
			dry	wet	dry	wet	dry	wet	dry	wet	dry	wet
Hand	Without fluid	50	1/12	1/12—1/4	1/4	1/2—1	1/4—1/2	1—2	1/2—1	2—5	1—2	10—15
Core	Same	200	1/12	1/12—1/4	1/4	1/2—1	1/2—1	2—3	3—5	5—10	5—10	10—30
	Using water or small quantities of slurry at temp. below 0°C	200	1/12	1/4—1/2	1/4—1/2	1—2	1—2	3—5	3—5	10—15	10—20	20—90
	Using slurry at temp. above 0°C	200	1/12—1/4	1	1/2—1	3—5	5—10	10—15	10—20	15—20	15—30	30—100
	Same	500	1/2—1	1—2	2—3	5—10	5—10	15—20	10—20	15—30	20—60	40—180
Rotary	Same	500	1/2—1	1—2	3—5	5—15	10—15	15—20	10—30	20—40	30—180	60—700
	Same	3000	1/2—2	2—10	5—10	5—15	10—15	15—20	15—30	30—100	30—180	180—1200 and higher

Table XXXVII

Standard depth of temperature measurements in boreholes of different depths, in m

Depth, m	Interval between points of measurement, m	Depth of hole, m		
		50	50 - 500	Over 500
0 - 10	1 - 5	1,2,3,4.5,6, 7,8,9,10	1,3,5,7	1,5,10
10 - 20	2 - 5	12,14,16,18,20	10,15,20	15,20
20 - 40	5 - 10	25,30,35,40	25,30,35,40	30,40
40 - 60	10	50	50	50
60 - 200	20	-	60,80,100,120,140,160,180,200	
200 - 1000	50	-	250,300,350 400,450,500	250,300,350,400,450,500, 550,600,650,700,750,800, 850,900,950,1000
Over 1000	100	-	-	1100 etc.

Table XXXVIII

Properties and uses of various thermometers

Thermometer	Diam. mm	Weight of assembly[1] kg	Inertia (duration of constant reading)	Interval of depths (m) at which thermometer may be used	
				Dry holes	Wet holes
Mercury-rapid-reading	32 - 100	Up to 2	20 - 80 sec	Cannot be used	300[2]
Mercury slow-reading: in open casing	19 - 75	Up to 12	25 - 70 sec	30-200	200[2]
Mercury slow-reading: in closed casing	75 - 100	40 - 150	Up to 10-20 min	700-800	500-700
Mercury maximum: in open casing	19 - 25	To 10[6]	Constant reading	2200-5000[3]	5000[2]
in closed casing	25 - 110	35 - 100[6]	Same	2200-5500[3]	500-1500
Mercury photothermometers	100	140 and over	None	To 5500[3]	1500-2500
Resistance thermometers: standard	65 - 75	Same	"	To 5500[4]	3000-5500
non-standard (self-made)	19 - 75	20 and over[7]	"	To 5500[4]	300-5500[5]
thermistors	10 - 40	5 and over[7]	"	To 5500[4]	300-5500[5]
Electronic	40 - 75	Same	"	To 5500[4]	3000-5500
Thermoelectric	50 - 80	Same	"	20-300[5]	20-300[5]

1) Weight of assembly includes that of thermometer, measuring device, conductor and cable.
2) Not more than 5 m from the surface of water.
3) Depending on soil temperature, range of scale and cable strength.
4) Depth of measurements depends on cable strength.
5) Depending on the design of thermometer.
6) Excluding the weight of cooling equipment.
7) At depths exceeding 600 - 1000 m, additional mechanical drive is required for the winch, depending on equipment and type of cable used.

Table XXXIX

Determination of thermal current and its sign
by the double reading method

Case	Sign			Current	Case	Sign			Current
	I_1	I_2	Resultant			I_1	I_2	Resultant	
1	+	−	+	$\frac{I_1+I_2}{2}$	5	0	+	−	$\frac{I_2}{2}$
2	−	+	−	$\frac{I_1+I_2}{2}$	6	+	0	+	$\frac{I_1}{2}$
3	+	+	$\begin{cases} +(I_1>I_2) \\ -(I_1<I_2) \\ 0(I_1=I_2) \end{cases}$	$\frac{I_1-I_2}{2}$	7	0	−	+	$\frac{I_2}{2}$
4	−	−	$\begin{cases} +(I_2>I_1) \\ -(I_2<I_1) \\ 0(I_1=I_2) \end{cases}$	$\frac{I_2-I_1}{2}$	8	−	0	−	$\frac{I_1}{2}$

Table XL

Composition of cooling mixtures and cryohydrates

Cooling mixtures			Cryohydrates		
Salt	Amount of salt per 100 g of ice of compacted snow, g	Temp. of mixture, °C	Salt	Anhydrous salt content in aqueous soln., %	Melting point of cryohydrate, °C
Potassium sulphate	10	- 1.9	Magnesium sulphate	19.0	- 3.9
Potassium nitrate	13	- 2.9	Zinc sulphate	27.2	- 6.5
Potassium chloride	30	-10.6	Potassium chloride	19.7	-11.1
Ammonium chloride	25	-15.0	Ammonium chloride	18.7	-15.8
Sodium chloride	33	-21.2	Sodium nitrate	36.9	-18.5
Calcium chloride	200	-35.0	Sodium chloride	22.4	-21.2
			Magnesium chloride	21.6	-33.6

Table XLI

Record of field observations using thermistors
Geothermal unit No. 3.
Cable consists of three PTF-7 wires.
Thermistors No. 3 and 4 from the 13th calibration group.
R_3 at +22.12°C = 9,080 ohm.
R_1 at +22.12°C = 10,020 ohm.
Direct current resistance bridge No. 11,232.
Depth measurements by means of marks on the cable.

No. of hole	Depth of meas. m	Time of meas.		Air temp. °C	Resistance, ohm			Signature of investigator
		Date	Hr-Min		Thermistor No.3	Thermistor No.4	Total (Nos.3 and 4)	
32	10	8.IX	11-00	+16	22,290	23,590	45,870	
	20		12-40	+18	22,370	23,660	46,080	
	30		13-30	+22	22,360	23,650	46,020	

Table XLII

Approximate stabilization periods of thermometers in boreholes

Thermometers	Diam. of thermometer mm	Init. diff. in temp. of thermometer and surrounding medium, °C	Stabilization period			
			Approx. (final diff. in the temp. of thermometer and medium 0.2 - 0.3°C)		Full (final diff. in temp. less than 0.05°C)	
			in air	in water	in air	in water
Mercury, rapid-reading*	32-50	2-20	-	10-20 min	-	20-50 min
Mercury, rapid-reading	60-100	2-10	-	10-20 min	-	30-40 min
Mercury, rapid-reading	60-100	10-21	-	15-30 min	-	40-60 min
Mercury, rapid-reading	60-100	21-23	-	20-40 min	-	60-120 min
Slow-reading, in open casing**	19-25	1-30	2-3 hr	-	3-4 hr	-
Slow-reading, in closed casing	32-75	1-30	3-4 hr	-	4-6 hr	-
	75-110	1-30	6-8 hr	4-6 hr	10-15 hr	6-8 hr
Mercury, maximum:						
in open casing	19-25	1-30	20-40 min	10-30 min	30-60 min	20-50 min
in special casing	25-60	1-30	1-2 hr	20-60 min	2-3 hr	30-120 min
in closed casing	75-110	1-30	6-8 hr	4-6 hr	10-15 hr	6-8 hr
Mercury thermometer	100	1-30	2-3 hr	1-2 hr	3-4 hr	2-3 hr
Resistance thermometers:						
standard (factory made)	60-75	1-30	2-5 min	1-2 min	20-120 min	10-60 min
non-standard (self-made)	40-75	1-30	10-60 min	1-20 min	30-120 min	10-60 min
Semiconductor thermistors:						
standard	70	1-30	2-5 min	1-2 min	20-120 min	20-60 min
non-standard	10-40	0-2	20-30 min	10-20 min	40-60 min	20-30 min
non-standard	10-40	2-20	60-90 min	20-40 min	80-130 min	40-60 min
Electronic	40-75	1-30	2-5 min	1-2 min	30-120 min	10-60 min
Thermoelectric	50-80	1-30	2-5 min	1-2 min	5-10 min	2-5 min

* Not used for air temperature measurements.
** Not used for water temperature measurements.

Table XLIII

Records of temperature measurements obtained by various types of equipment

Commercial electric resistance thermometers

Borehole No.	Depth, m			Time of meas.		Reading, microvolt			t_o-t*, °C	Reading, °C	Corrected reading, °C	Remarks
	Cable	Meter	Actual	Initial	Final	Compensator	Potentiometer	Total				

Thermistors

Borehole No.	Depth, m			Time of meas.		Resistance, ohm			Reading, °C	Corrected reading, °C	Remarks
	Cable	Meter	Actual	Initial	Final	Thermistor + cable	Cable	Thermistor			

Mercury thermometers

Borehole No.	Depth, m			Time of meas.		Thermometer no.	Reading, °C	Correction, °C	Corrected reading, °C	Remarks
	Cable	Meter	Actual	Initial	Final					

* Difference between the initial temperature of the thermometer and the temperature at the point of measurement.

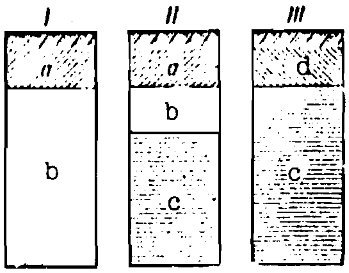


Fig. 2

Seasonal freezing and thawing:
I - layer of seasonal freezing (a) and unfrozen gound (b);
II - same (a) with permafrost at great depth (c);
III - layer of seasonal thawing (d) and permafrost (c)

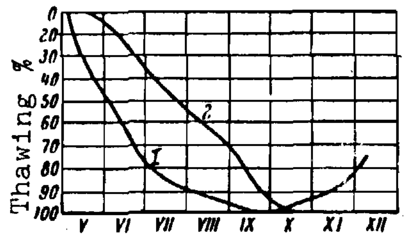


Fig. 3

Rate of seasonal thawing of soils based on data obtained over a period of several years in % of maximum depth of seasonal thawing
1 - at the Yakut permafrost station of the Academy of Sciences of the U.S.S.R. (P.I. Mel'nikov and P.A. Solov'ev, 1952); 2 - in the Igarka area (Tumel, 1941)

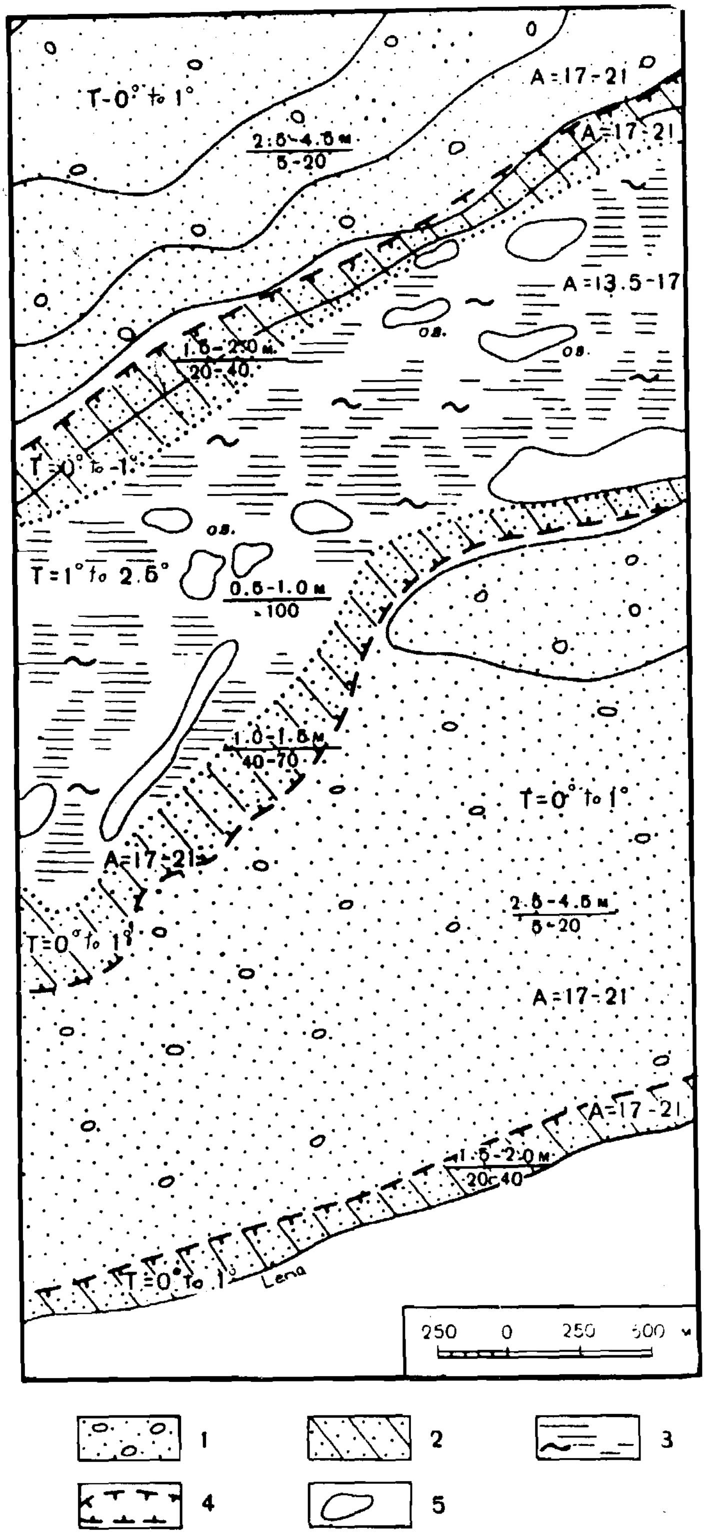


Fig. 4

Map showing the types of seasonal freezing and thawing on sections I and II on terraces above the flood plain. 1 - mixed sand with inclusions of gravel and pebbles; 2 - clay loam and sandy loam; 3 - peaty clay loam; 4 - boundaries of seasonal freezing and thawing; 5 - lakes; T - mean annual temperature of soil. Fractions - in the numerator: depths of seasonal freezing and thawing in m; in the denominator: moisture content of soil in %. A - temperature amplitude on the ground surface

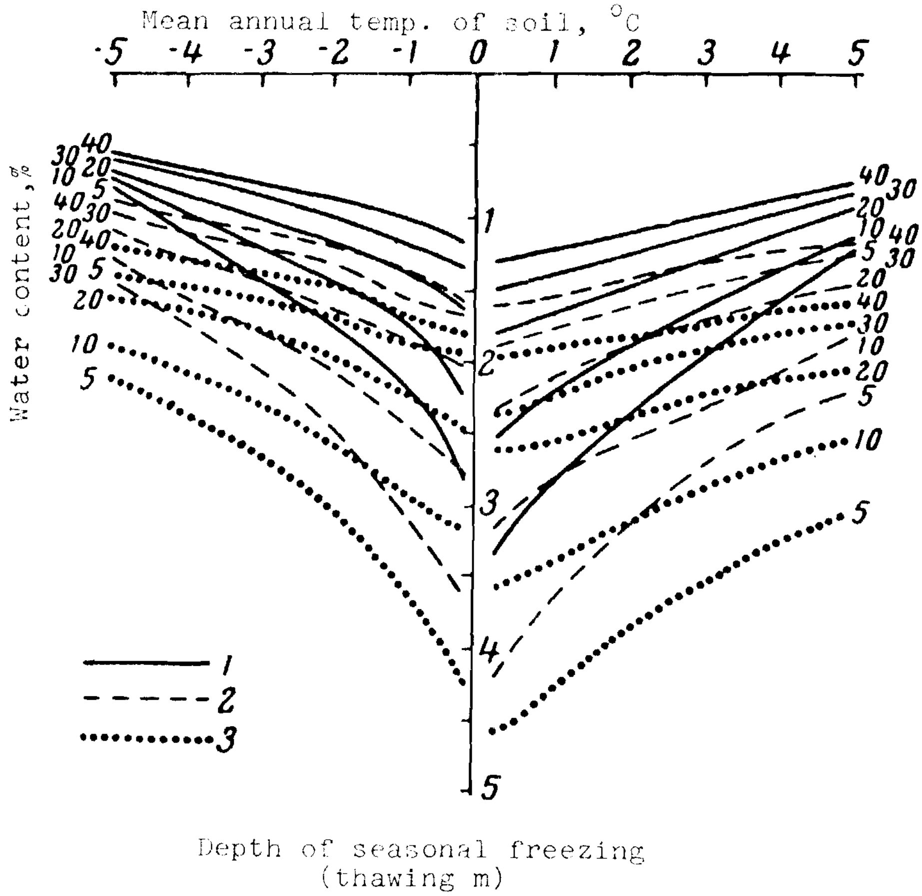


Fig. 5

Nomogram for the determination of depths of seasonal freezing or thawing of clay loam (constructed by N. Kh. Kufman and L.N. Maksimova) 1 - temperature amplitude on the surface of the ground equal to 11°C; 2 - same equal to 17°C; 3 - same equal to 24°C

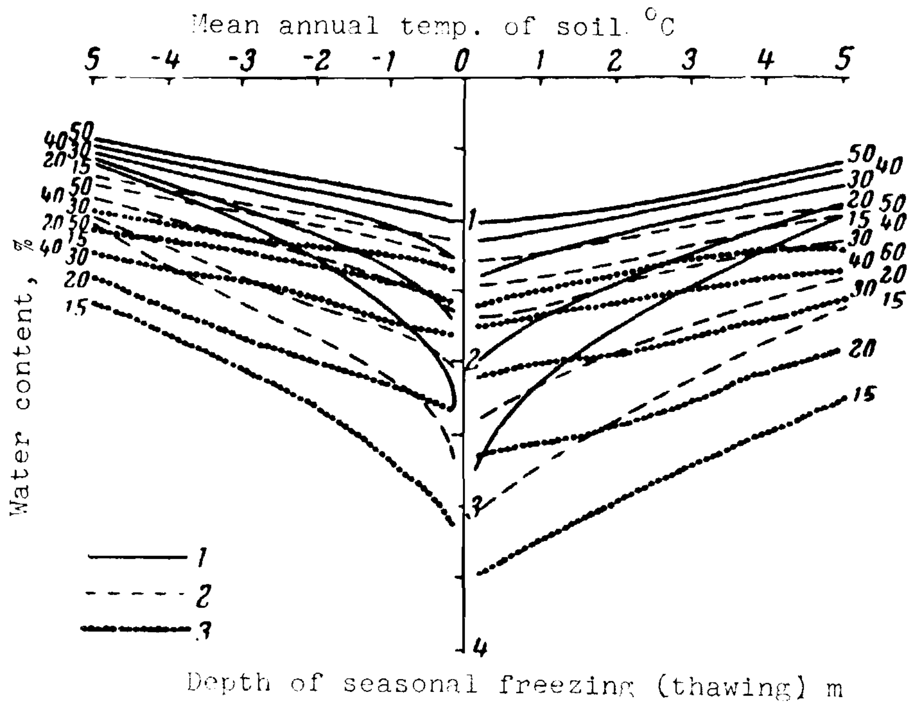


Fig. 6

Nomogram for the determination of depths of seasonal freezing or thawing of sandy soil (constructed by N. Kh. Kufman and L.N. Maksimova) 1 - temperature amplitude on the surface of the ground equal to 11°C; 2 - same equal to 17°C; 3 - same equal to 24°C

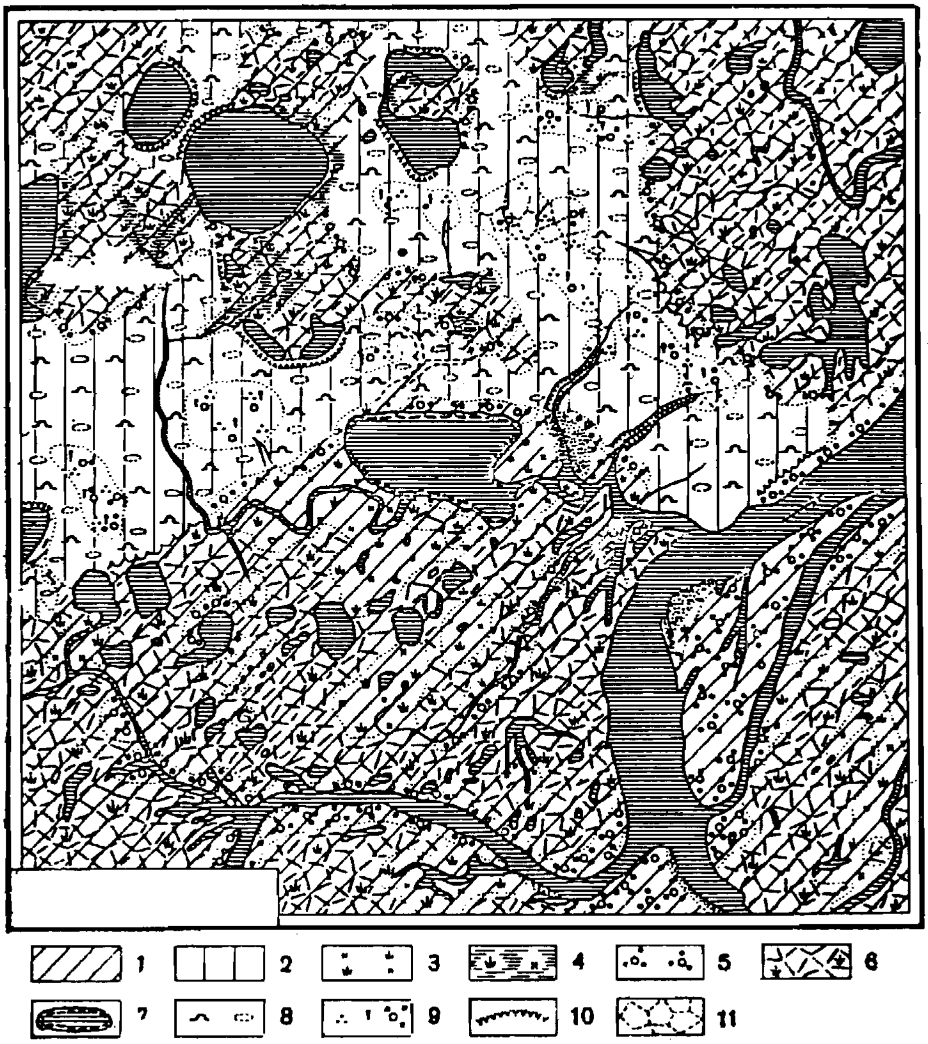


Fig. 7

Schematic map showing the types of terrain on one of the sections of a lowland: Types of terrain: 1,2 - ancient alluvial plain; floodplain and alasy. Subtypes of terrain: 3 - river embankments and lake shores covered by vegetation of the meadow type (grasses and horsetails), depth of thaw 0.8 - 1.0 m; 4 - sedge and cotton grass bogs, depth of thaw 0.3 - 0.4 m; 5 - brush tundra, depth of thaw 0.5 - 0.7 m; 6 - level sections of a flood plain with polygonal relief, depth of thaw on dry polygons 0.3 - 0.4 m, in swampy polygons 0.4 - 0.5 m; 7 - sections of laida* which dry up periodically, depth of thaw over 1 m; 8 - level sections of an ancient alluvial plain covered by hummocky moss-lichen tundra, depth of thaw 0.4 - 0.5 m; 9 - low-lying sections of an ancient alluvial plain covered by hummocky tundra, depth of thaw 0.3 - 0.4; 10 - escarpments of an ancient alluvial plain with baidzharakhi; 11 - collapsed polygons

* Yakut term for small tundra boglake

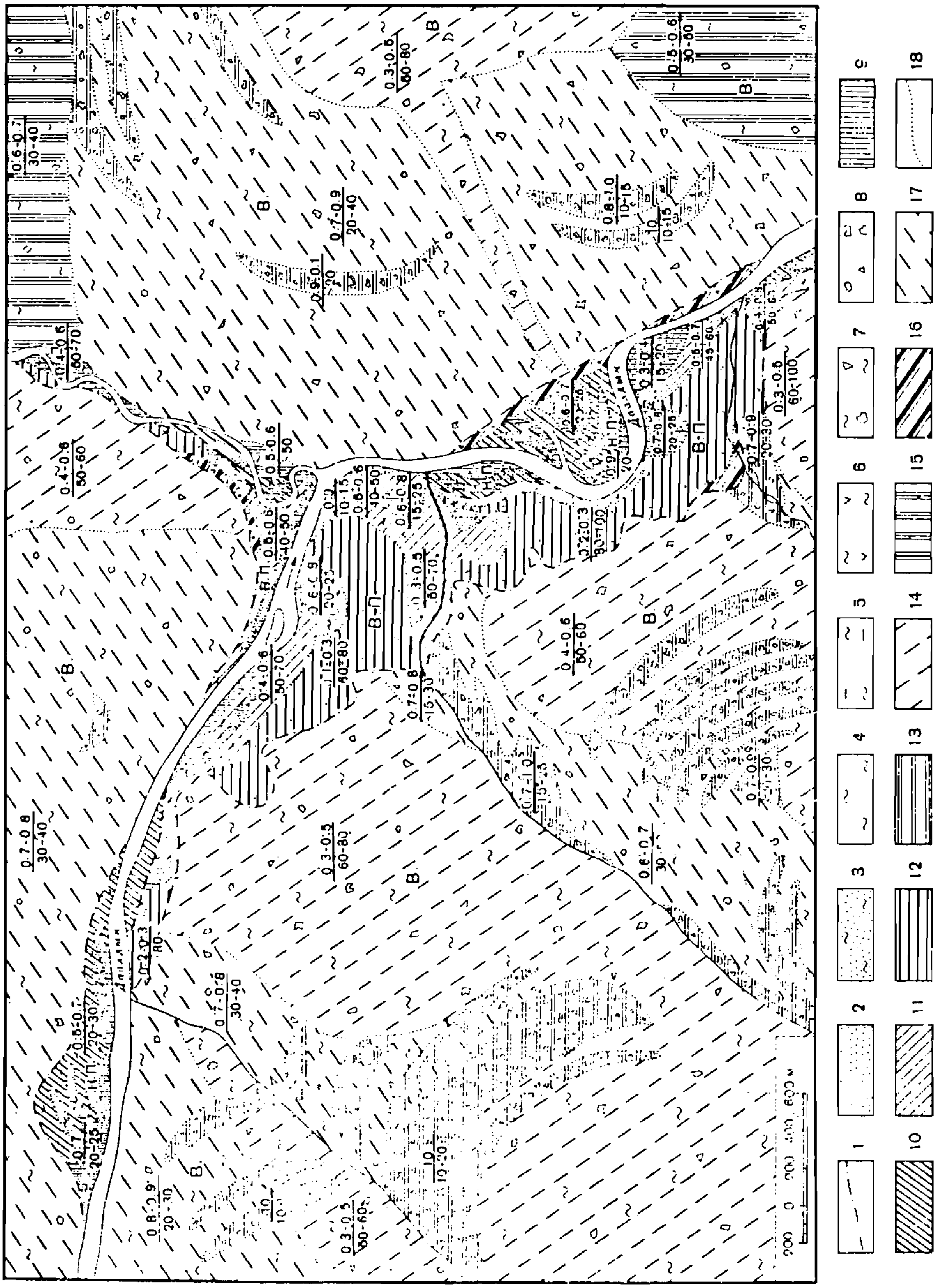


Fig. 8

Fig. 8 Schematic map showing the cryolithological soil types in the layer of seasonal freezing and thawing (the Daaldyn River)

Elements of relief: 1 - boundaries of geomorphological levels; НП - low flood-plain; ВП- high floodplain; В - water divide. Lithological composition of soil: 2 - sand; 3 - sandy loam and sandy clay loam; 4 - clay loam with some humus; 5 - silty clay loam; 6 - peaty clay loam; 7 - clay loam with rock fragments; 8 - small and large rock fragments; 9 - deposits on embankments and flood-plain crests represented by sandy varieties and characterized by massive cryogenic texture; 10 - somewhat swampy depressions between crests (low flood-plain) represented by clay loam and sandy loam with some humus or silt, characterized by streaky cryogenic structure; 11 - high flood plain represented by clay loam without peat but with occasional silt characterized by reticulate (network) cryogenic structure; 12 - overgrown secondary water reservoirs and old river channels represented by peat and strongly peaty clay loam characterized by fine lenticular cryogenic texture; 13 - steep slopes represented by coarse fragmentary material, characterized by bread-crust cryogenic texture; 14 - dry slopes (with bedrock close to the surface), represented by clay loam with a high content of fragmentary material. Cryogenic textures: in the upper part of the layer of seasonal freezing - fine lenticular; in the lower part - bread-crust texture; 15 - dry or relatively dry slopes (with bedrock at great depth) represented by clay loam with a low content of fragmentary material characterized by lenticular (coarse and fine) cryogenic texture; 16 - talus cones represented by clay loam characterized by lenticular cryogenic texture; 17 - gentle swampy slopes represented by clay loam with lenticular cryogenic texture in the upper part of layer subject to seasonal freezing and reticulate or cryptobanded texture in the lower part; 18 - boundaries of cryolithological soil types; fractions: in the numerator depth of thaw in m, in the denominator - ice content in % of dry weight

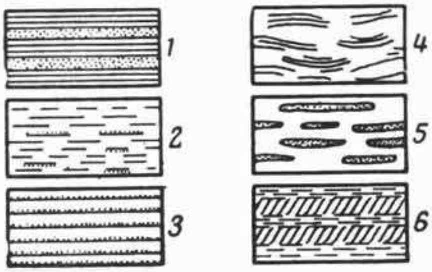


Fig. 9

Types of primary bedding in sedimentary rocks (after E.P. Bruns) 1 - horizontal-banded bedding; 2 - horizontal-discontinuous bedding; 3 - horizontal ribbon lamination; 4 - wave-like bedding; 5 - lenticular bedding; 6 - cross-bedding

Fig. 10

Fissured cryogenic texture of kimberlite (Table XIII, No. 1). Ice (white) fills cracks which dissect the rock into isometric blocks

Fig. 11

Crusty cryogenic texture of eluvium formed as a result of weathering of kimberlite (Table XIII, No. 3). Ice (marked with a cross) surrounds rock fragments

Fig. 12

Basal cryogenic texture of eluvium from kimberlite (Table XIII, No. 4) Ice forms the main part of rock (marked with a cross)

Fig. 13

Wave-like inherited cryogenic texture of deposits in river shallows (Table XV, No. 1)

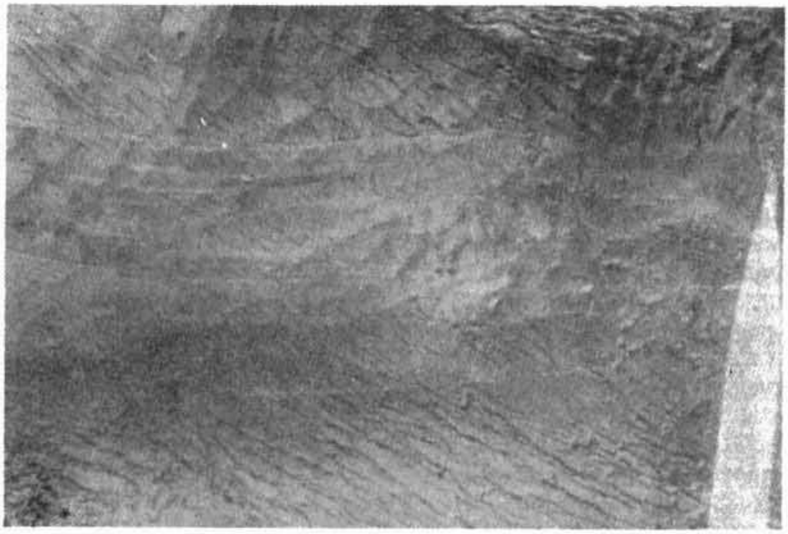


Fig. 14

Cross-lenticular cryogenic texture
of oxbow lake deposits (Table XV, No. 3)
Photograph shows slanted ice lenses
(dark lines) and sedimentary bedding
(light horizontal lines). A 4 cm wide
handle of an ice pick is used as a scale

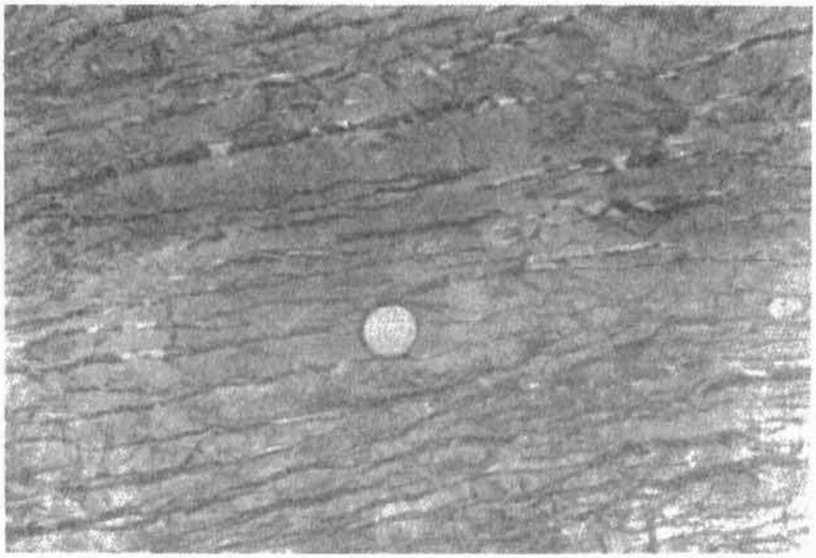


Fig. 15

Cross-laminated cryogenic texture of oxbow lake deposits. (Table XV, No. 4) Ice layers (dark lines) up to 2 cm in thckness. Diam. of circle 4.5 cm

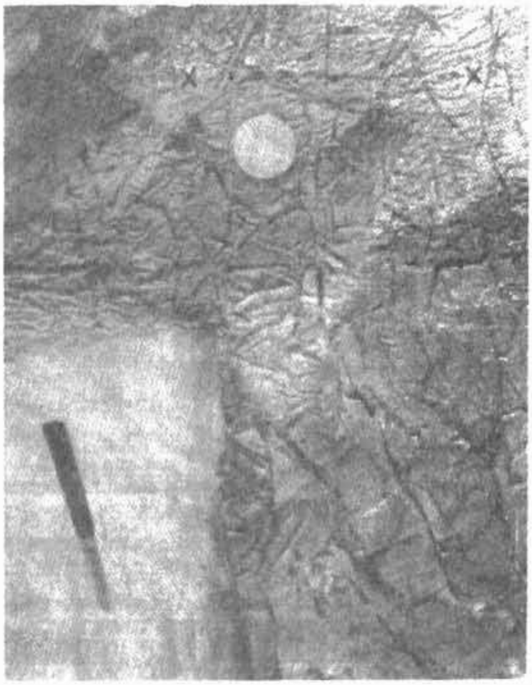


Fig. 16

Mixed cryogenic texture (Table XV, No. 5). Photograph shows slanted ice lenses, as well as dark and horizontal ice layers (marked with a cross). An ice vein may be seen in the lower left corner. Diameter of circle 3.8 cm

Fig. 17

Horizontal, parallel-layered, reticulate cryogenic texture (Table XV, No. 6)

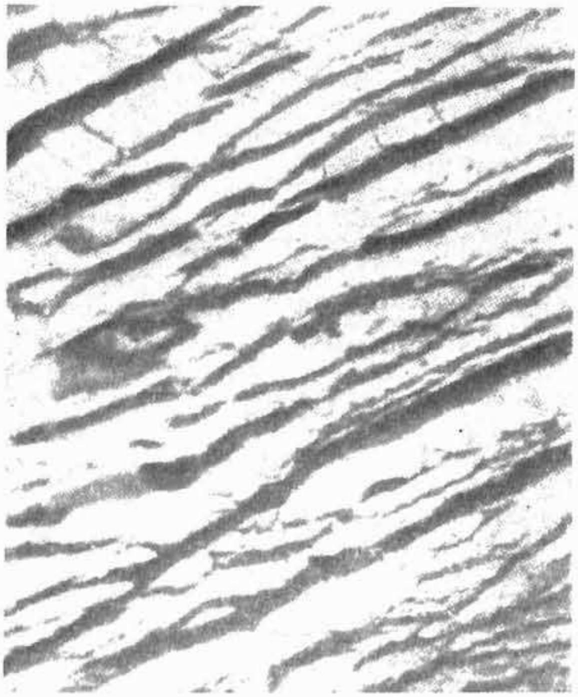


Fig. 18

Coarse cross-laminated cryogenic texture of marine deposits. Black ice layers up to 3 or 4 cm thick. Photo by A.G. Brodskaya

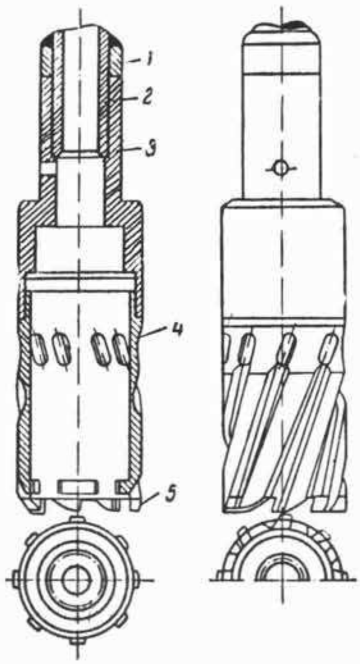


Fig. 29

Core lifter designed by
A.M. Pchelintsev
1 - ring; 2 - rod;
3 - adapter; 4 - bit;
5 - pobedit plate

Fig. 30

General view of bit
designed by I.E. Potzger
for sampling in permafrost

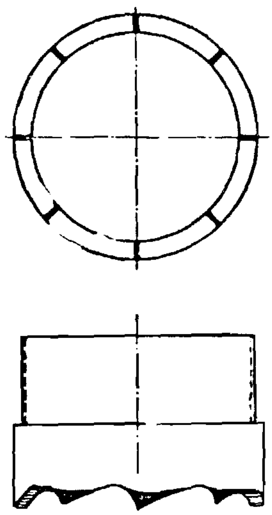


Fig. 31

Diagram of bit with welded teeth for drilling frozen soil containing boulders and pebbles

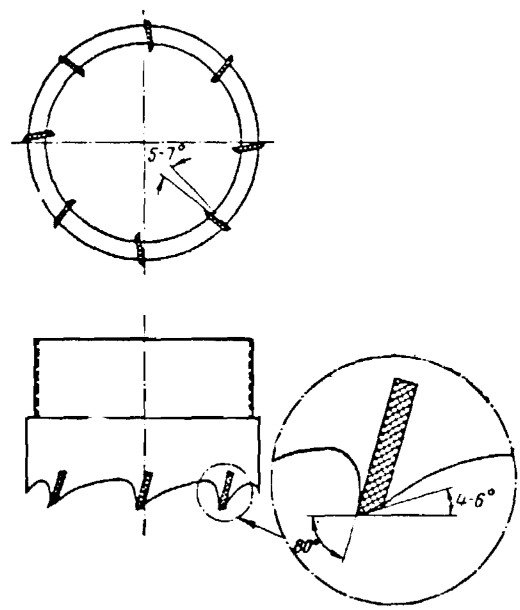


Fig. 32

Diagram of a bit with podebit teeth for drilling frozen soil without boulders and pebbles

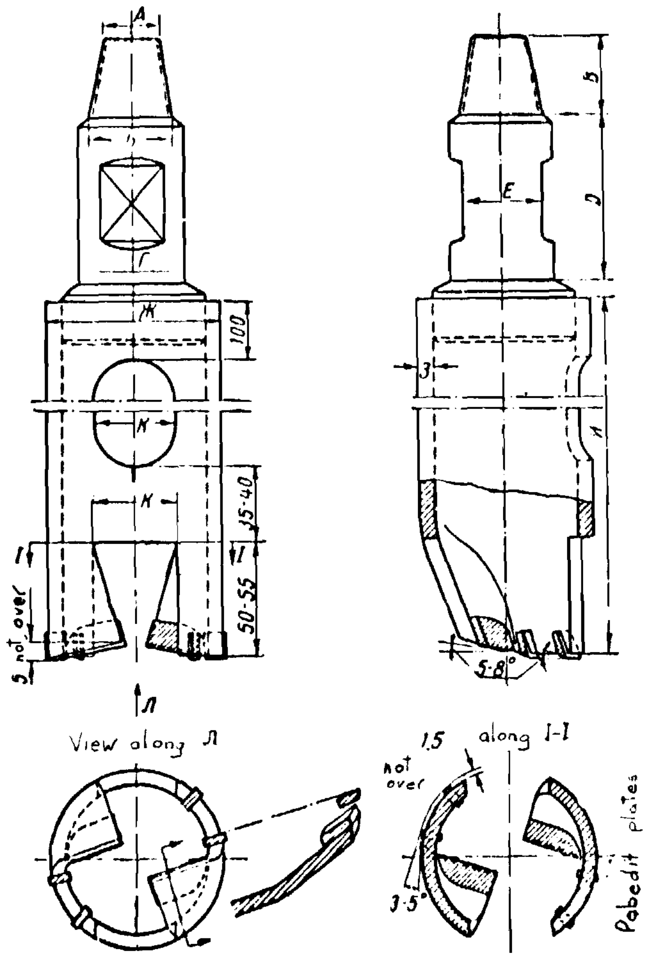


Fig. 33

Spoon bit with podebit plates which are joined to the bit by autogenous welding using copper or brass

Parameter	Symbol on drawing	Internal diam. of casing, mm	
		78	115
Diam. of upper base	A	22.75	315
Diam. of lower base	Б	32.5	45
Height	B	39	54
Number of threads per 1"	—	8	8
Diam. of spoon neck	Г	42	60
Height of spoon neck	Д	70	70
Width of recess for wrench	E	36	50
Outer diam. of spoon pipe	Ж	70	102
Thickness of spoon wall	З	6	6.5
Length of spoon pipe	И	700	700
Width of cut	К	38	44

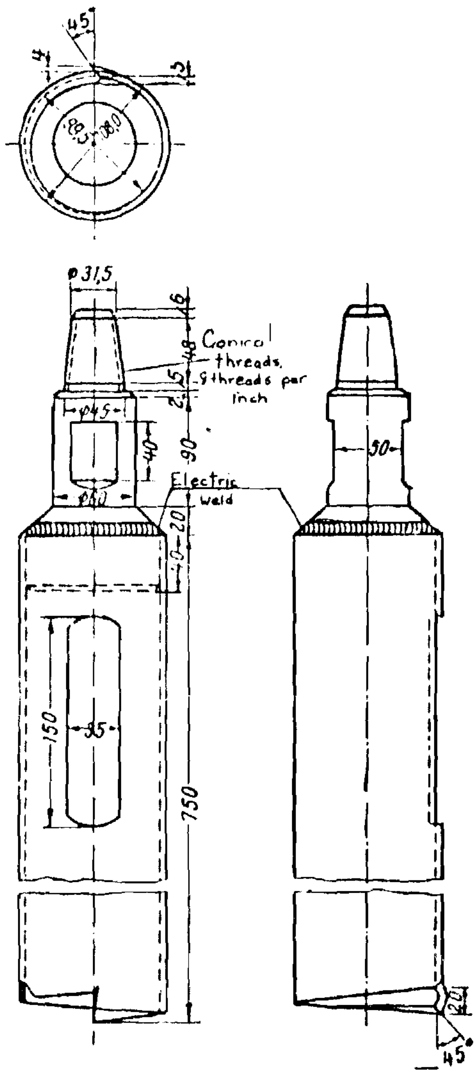


Fig. 34

Spoon bit for drilling frozen soil designed by "Sevmorproekt"

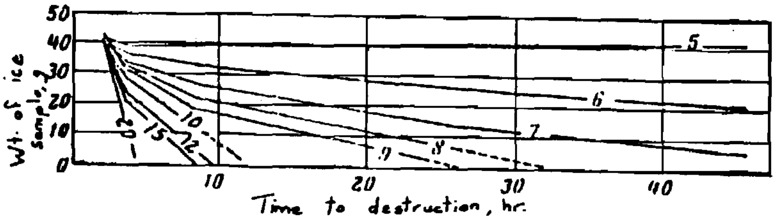


Fig. 35

Diagram showing the destruction of ice by aqueous solutions of table salt of different concentrations, in % (after A.V. Maramzin)

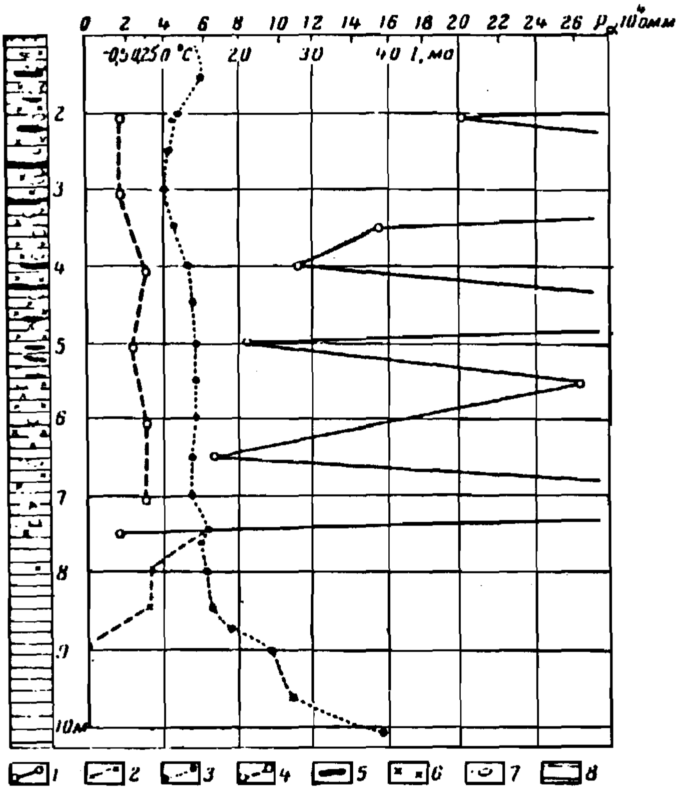


Fig. 36

Logging diagram of borehole No. 19
1 - apparent resistivity curve (ρ_a, ohm·m) obtained by means of a two-inch probe;
2 - same for a three-inch probe;
3 - current curve (I, ma); 4 - temperature curve (t, ^{0}C); 5 - ice lenses; 6 - ice crystals; 7 - boulders; 8 - clay loam moraine

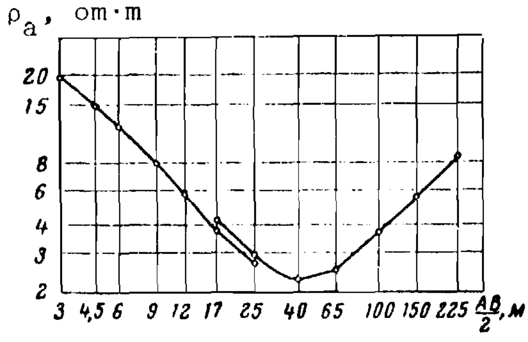


Fig. 37

Three-layer VEL curve
First layer - frozen marine deposits; second layer - thawed marine deposits saturated with highly mineralized water with negative temperature; third layer - frozen marine deposits

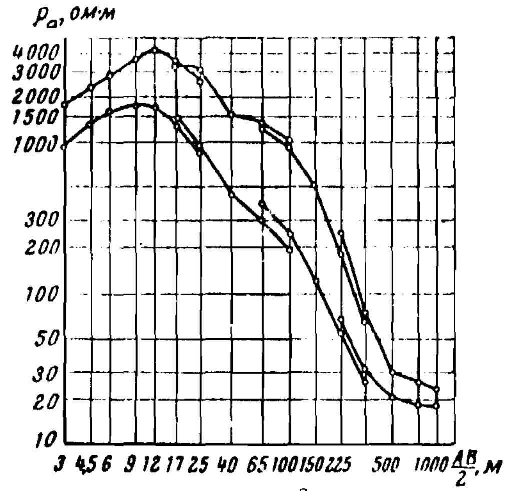


Fig. 38

VEL curves obtained above permafrost consisting of two horizons with different resistivities

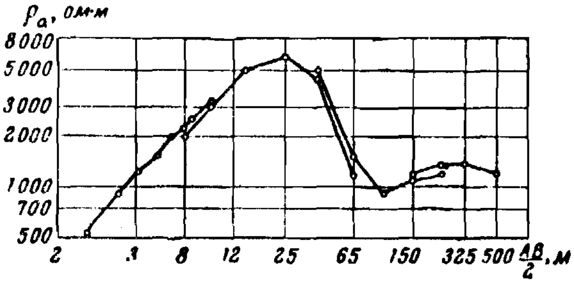


Fig. 39

VEL curve obtained above frozen bedrock

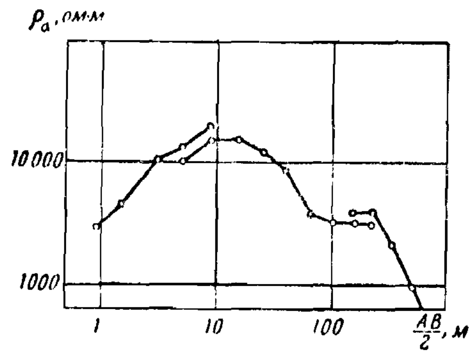


Fig. 40

VEL curve obtained above a thin (5 m) layer of frozen overburden resting on bedrock. The lower permafrost surface occurs at a depth of about 120 m

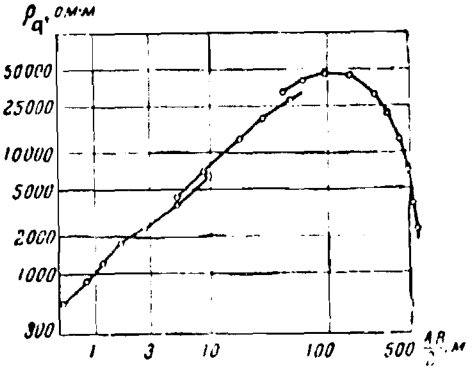


Fig. 41

VEL curve obtained above thick permafrost (about 300 m in thickness)

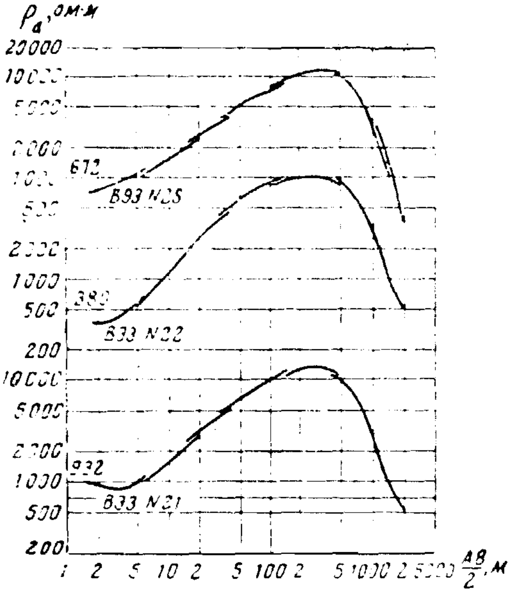


Fig. 42

VEL curves obtained above frozen unconsolidated bedrock. Temperature of frozen rock - 2.5 °C; thickness 250 m

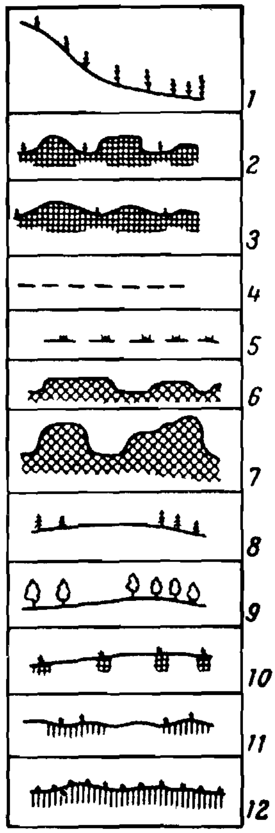


Fig. 43

Approximate representation of terrain as used in geophysical investigations
1 - tundra with willow stands; 2 - hummocky-bushy tundra (hummocky microrelief); 3 - same as above but with lower hummocks; 4 - swamps; 5 - sedge swamp; 6 - flat peaty tundra; 7 - peat hummocks; 8 - individual spruce trees and groups of spruce trees; 9 - individual birch trees and groups of birch trees, 10 - spotted tundra; 11 - degraded spotted tundra; 12 - "carpet" tundra

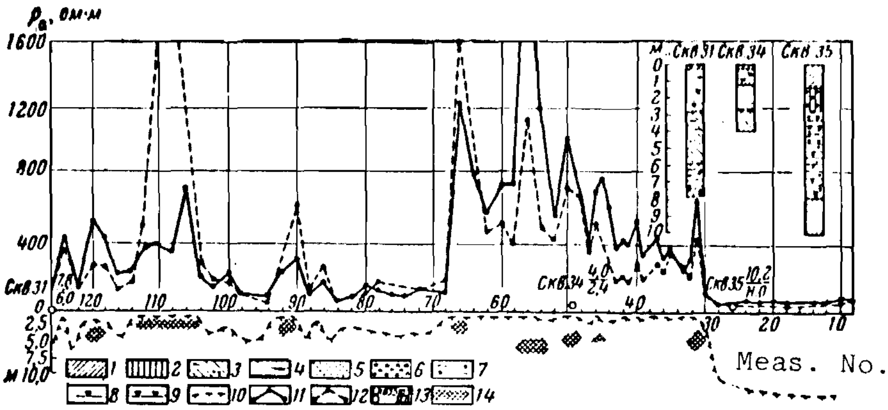


Fig. 44

Apparent resistivity curves
1 - talus clay loam; 2 - morainic clay loam; 3 - glacio-lacustrine clay loam; 4 - glacio-lacustrine clays; 5 - sand; 6 - pebbles; 7 - lenses and crystals of ice; 8 - depth of appearance of ground water; 9 - steady gound water level; 10 - permafrost table; 11 - ρ_a curve at AB = 40 m; 12 - ρ_a curve at AB = 16 m; 13 - borehole and its number (the borehole number is indicated by the numerator; the denominator indicates the depth of the permafrost table); 14 - permafrost horizons with high ice content.
Spacing of measurements 4 m

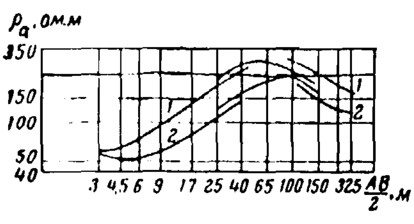


Fig. 45

VEL curves obtained on a section where the layer of seasonal freezing merges with permafrost (curve 1) and where this does not take place (curve 2). The displacement of curve 2 to the right indicates that permafrost occurs at great depths

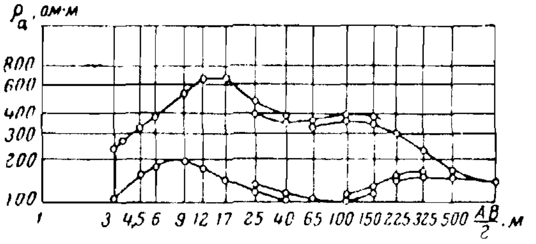


Fig. 46

VEL curves obtained above a horizontal talik enclosed in permafrost. The minimum in the centre of the curve indicates the presence of the talik

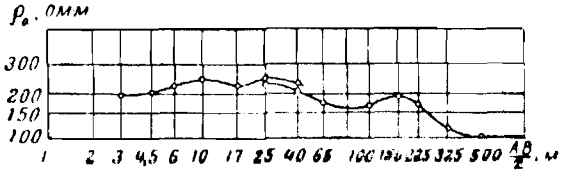


Fig. 47

VEL curve obtained above two horizontal taliks enclosed in permafrost. The two slight depressions in the centre of the curve are due to two thawed horizons

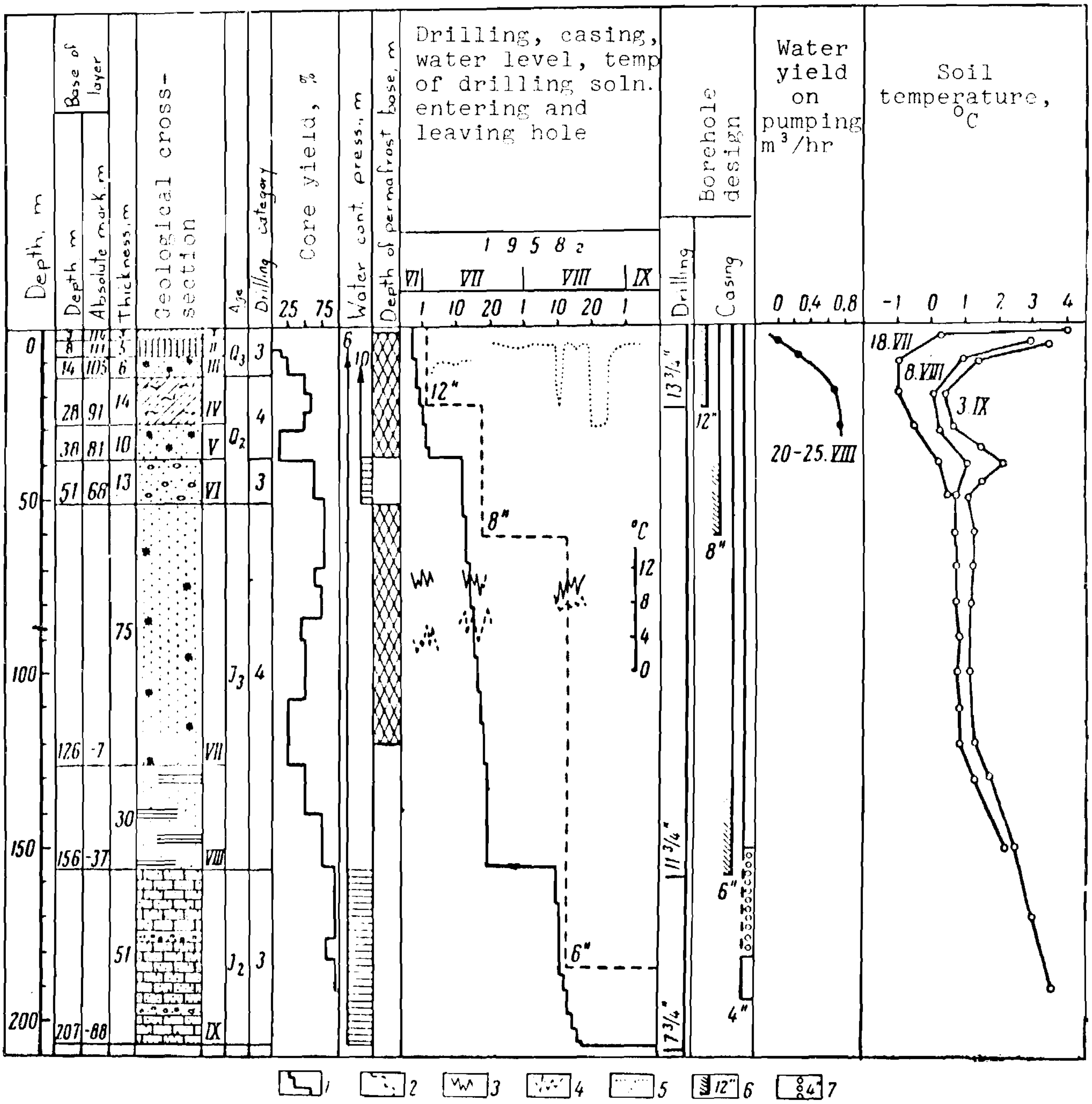


Fig. 49

I - compact clay loam, thawed to a depth of 2 m, frozen lower down; II - streaky fossil ice; III - fine-grained, frozen, monolithic sand; IV - alternating layers of sand and light clay loam, some thin layers of ice; V - frozen sand; VI - sand with pebbles; VII - fine-grained sand frozen to a depth of 120 m; VIII - sand with clay layers; IX - fine-grained water-bearing sandstone with layers of conglomerate;

1 - drilling curve; 2 - casing; 3 - temperature of drilling solution entering the hole; 4 - same for the solution leaving the hole; 5 - water level; 6 - cementation of borehole and pipe diameters; 7 - filter.

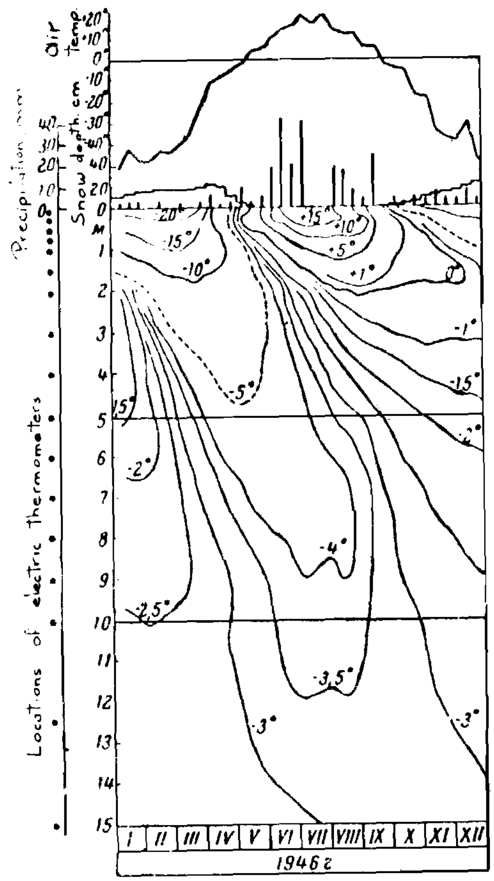


Fig. 54

Combined diagram showing air temperature, snow cover and soil temperature

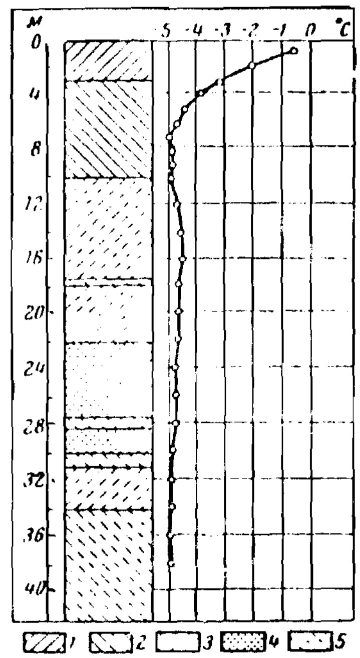


Fig. 55

Diagram showing the temperature and composition of soil
1 - heavy clay loam;
2 - heavy sandy loam;
3 - light clay loam;
4 - fine sand; 5 - light sandy loam

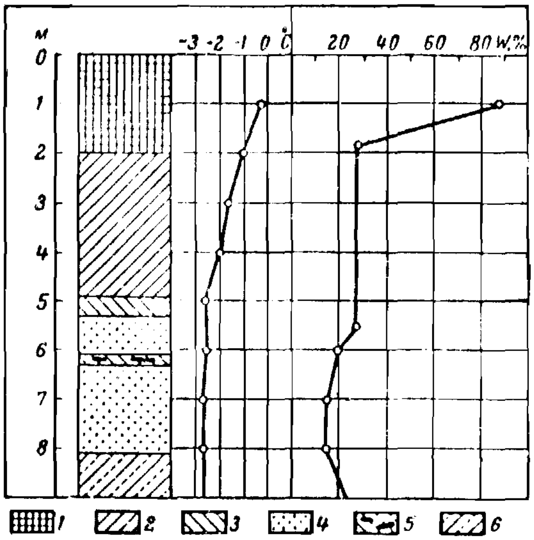


Fig. 56

Combined diagram showing temperature, moisture content and composition of soil
1 - peat; 2 - heavy clay loam; 3 - sandy loam; 4 - sand; 5 - clay loam with ice layers; 6 - light clay loam

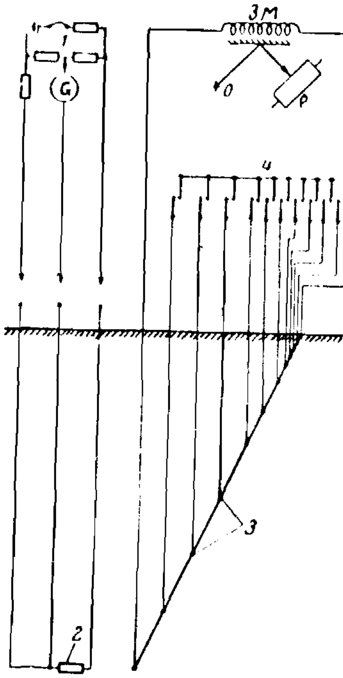


Fig. 114

Schematic diagram of an assembly for measuring soil temperature
1 - Wheatstone bridge; 2 - electric thermometer; 3 - thermocouples; 4 - contact couples of program switch; G - galvanometer; M - mirror galvanometer; O - illuminator of the mirror galvanometer; P - photorecorder

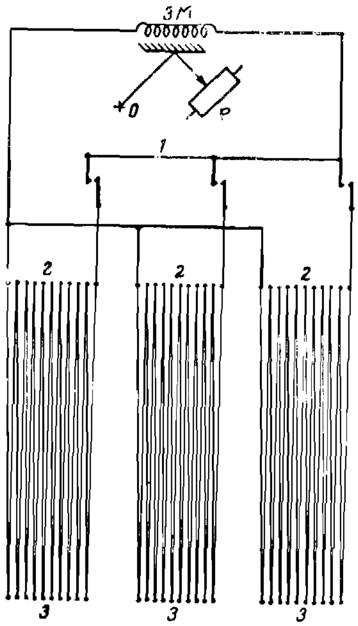


Fig. 115

Schematic diagram of three thermal spiders connected to an automatic recording channel
1 - contact couples of program switch; 2 - measuring thermocouples of thermal spiders; 3 - reference thermocouples; M - mirror galvanometer; O - illuminator of the galvanometer; p - photorecorder

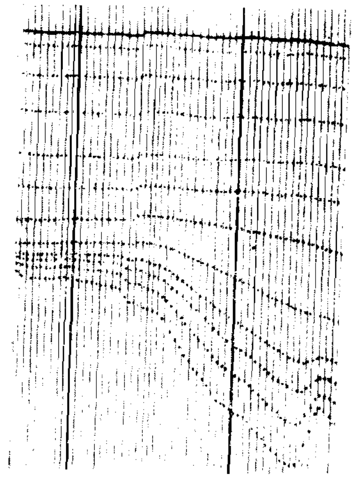


Fig. 116

Soil temperature record obtained at eleven different depths

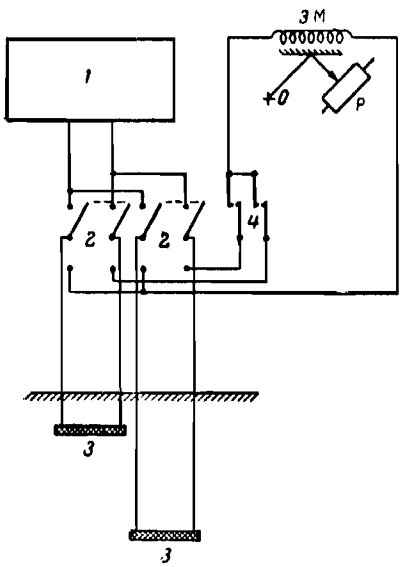


Fig. 117

Schematic diagram of a system for recording heat flow
1 - potentiometer; 2 - two-pole switches; 3 - heat flow meters; 4 - contact couples of program switch; M - mirror galvanometer; O - illuminator of the galvanometer; p - photorecorder

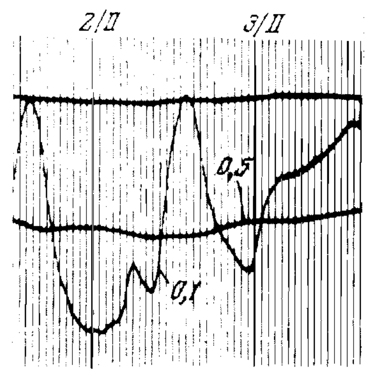


Fig. 118

A record of heat flow at depths of 0.1 and 0.5 m

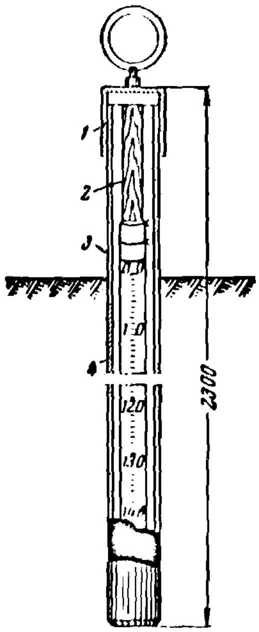


Fig. 119

Danilin frost gauge for measuring the depth of seasonal freezing and thawing of soil
1 - metal cap with ring;
2 - wooden rod; 3 - carbolite tube; 4 - rubber tube filled with water (the scale on the tube indicates depth)

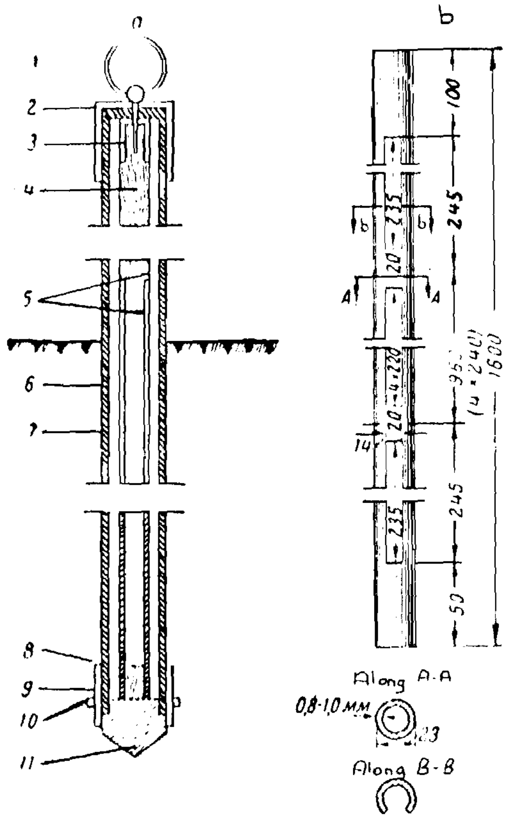


Fig. 120

Ratomskii frost gauge for measuring the depth of seasonal freezing and thawing of soil
a - longitudinal cross-section;
b - metal tube of the gauge
1 - screw with a ring;
2 - metal cap; 3 - metal rim of the wooden rod; 4 - wooden rod; 5 - wood screws (3);
6 - phenol plastic casing;
7 - metal tube filled with loam; 8 - wooden plug;
9 - metal rim; 10 - pin;
11 - conical steel bottom

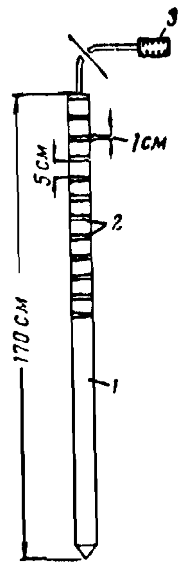


Fig. 121

Electric frost gauge
1 - rod; 2 - electrodes;
3 - commutator

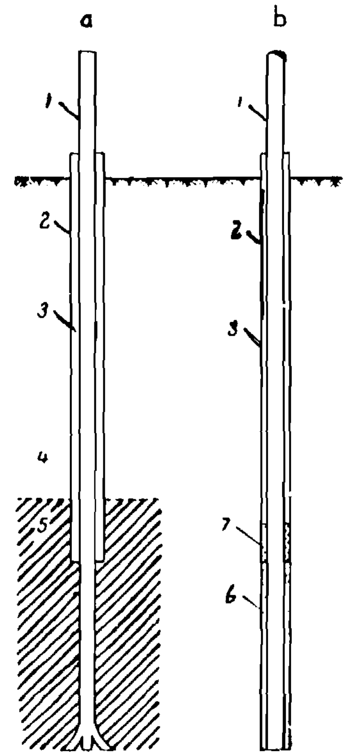


Fig. 122

Reference rods for use in permafrost (a) and seasonally frozen soil (b)
1 - inner tube; 2 - outer tube; 3 - solar oil; 4 - seasonally frozen soil;
5 - permafrost; 6 - cement
7 - grease or motor oil

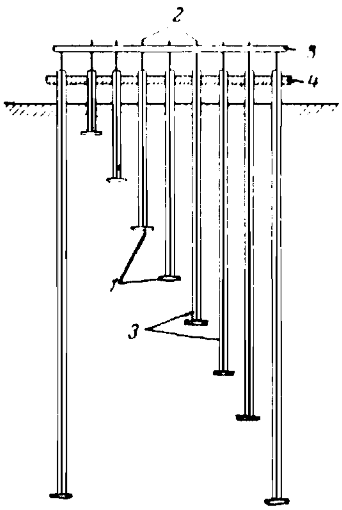


Fig. 123

Device for measuring heaving and settlement
1 - metal disks; 2 - indicator rods; 3 - rubber tubes; 4 - wooden headpiece; 5 - metal plank

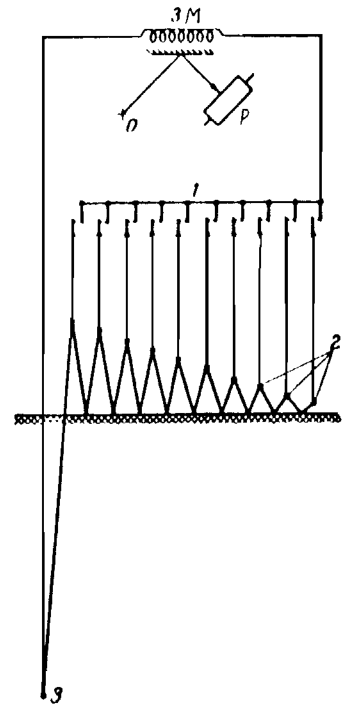


Fig. 124

Schematic diagram of a thermocouple assembly for use in the snow
1 - contact couples of program switch; 2 - measuring thermocouples; 3 - reference thermocouple; M - mirror galvanometer; O - illuminator of the galvanometer; p - photorecorder

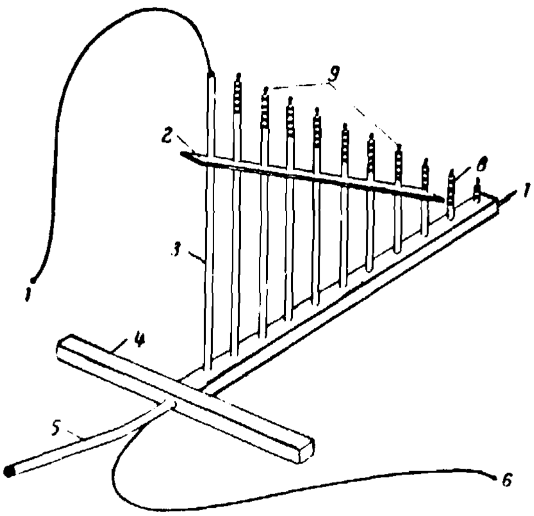


Fig. 125

Thermocouple assembly for use in the snow
1 - movable thermocouple for measuring the snow surface temperature; 2 - wooden plank; 3 - vinyl plastic tubes; 4 - base of assembly; 5 - cable; 6 - reference thermocouple; 7 - thermocouple for measuring the snow surface temperature; 8 - measuring thermocouples; 9 - temperature-sensitive elements

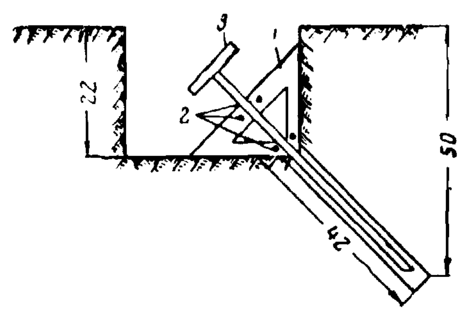


Fig. 126

Device for drilling a hole for temperature measurements in a testpit
1 - wooden template; 2 - guide pins; 3 - bit (dimensions in cm)

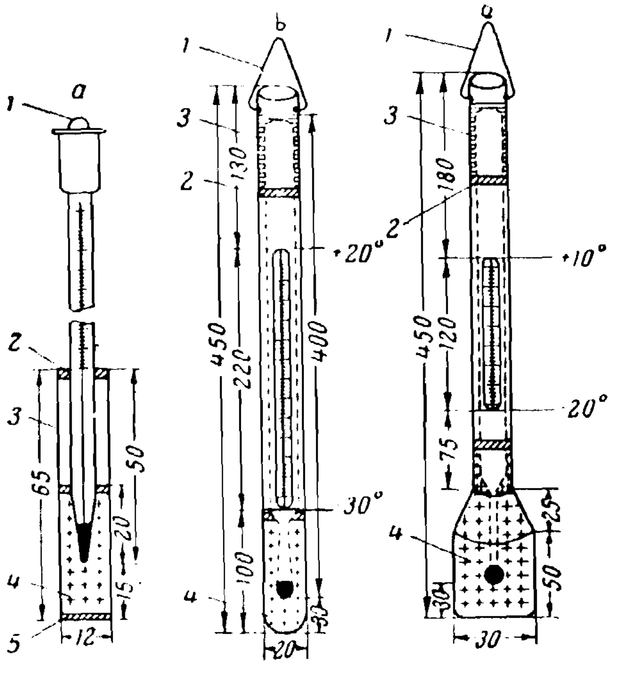


Fig. 127

Slow-reading thermometers
a - for temperature measurements to a depth of 2 m; b - to a depth of 10 m; c - to a depth of over 10 m; 1 - ring for suspending thermometer; 2 - ring for insulation tape; 3 - protecting tube; 4 - heat insulator; 5 - lid

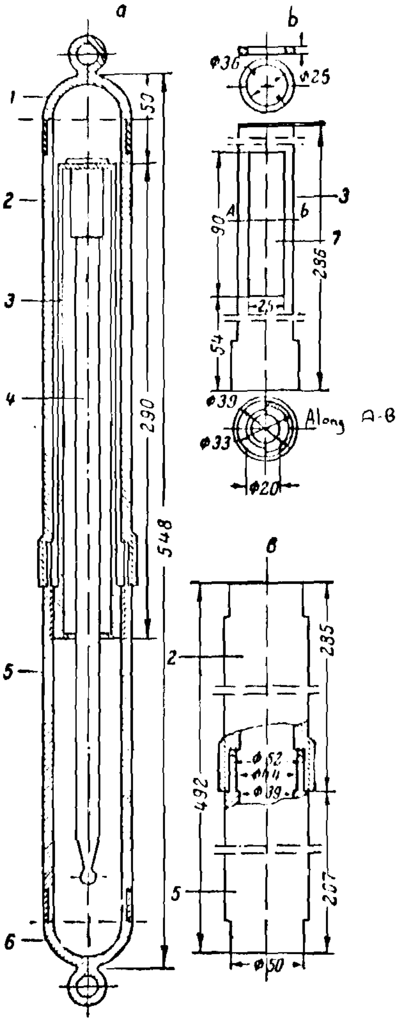


Fig. 128

Thermometer containers for temperature measurements to a depth of 500 m in waterlogged boreholes
a - cross-section of container; b - inner casing; c - outer casing
1 - upper lid with rings; 2 - upper part of outer casing; 3 - inner casing; 4 - thermometer; 5 - lower part of outer casing; 6 - lower lid with rings; 7 - observation window in inner casing

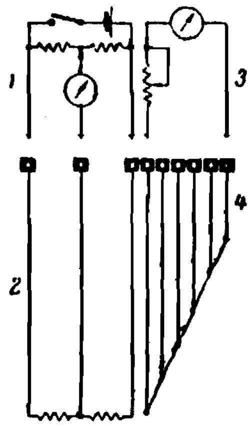


Fig. 129

Diagram of a thermometric assembly. 1 - Wheatstone bridge; 2 - resistance thermometer; 3 - thermal current meter; 4 - thermocouples

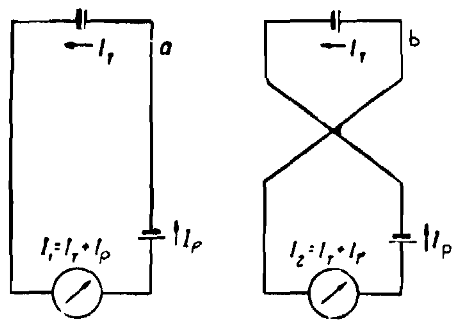


Fig. 130

Diagram of direct (a) and reversed (b) switching of measuring device into the thermocouple circuit. I_T - thermal current being measured; I_P - parasitic thermal current; I_1 and I_2 - total current on direct and reversed switching on of the measuring device

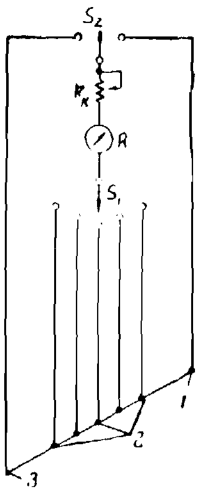


Fig. 131

Basic diagram of a thermocouple assembly
1 - upper reference thermocouple junction; 2 - measuring thermocouple junctions; 3 - lower reference thermocouple junction; S_1 - switch for measuring thermocouple junctions; G - galvanometer; r_K - temperature compensator; S_2 - switch for reference thermocouple junctions

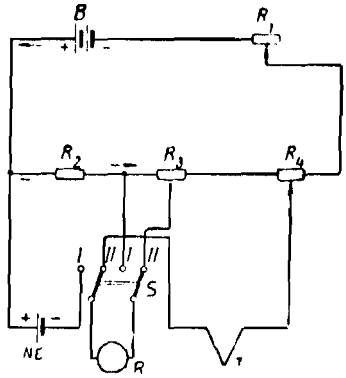


Fig. 132

Basic diagram of an assembly for measuring thermal emf by the compensation method
B - auxiliary battery; R_1 - control resistance; R_2 - constant resistance; R_3 - decade resistance box; R_4 - slide wire; NE - normal element; S - switch (position I - I - control; position II - II - measurement); G - galvanometer; T - thermocouple

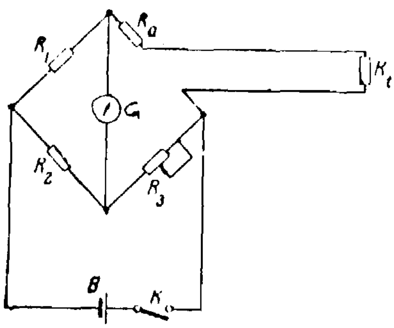


Fig. 133

Basic diagram of a balanced Wheatstone bridge. The electric thermometer is connected to the bridge by means of two wires: R_1 and R_2 - constant resistances; R_a - resistance of input leads; R_3 - alternating resistance; R_t - resistance thermometer; G - galvanometer; B - battery; K - key (button)

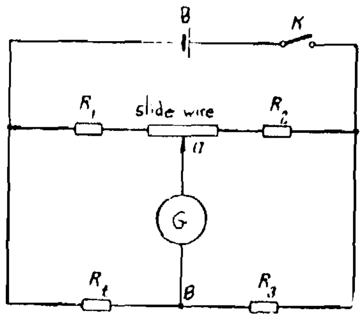


Fig. 134

Basic diagram of a balanced Wheatstone bridge. The two-leg resistance thermometer (R_t, R_3) is connected to the bridge by three wires. (For explanation see Fig. 133)

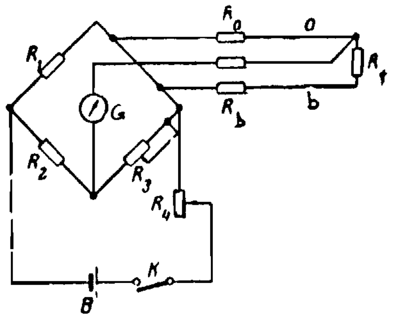


Fig. 135

Three-wire connection of single-leg resistance thermometer

R_a - resistance of wire a;
R_b - resistance of wire b;

R_4 - control resistance; other symbols as in Fig. 133

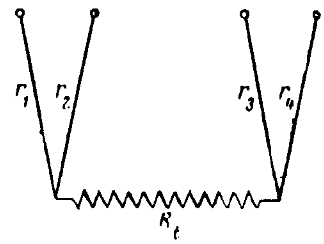


Fig. 136

Four-wire connection of resistance thermometer R_t

r_1, r_2, r_3 and r_4 - resistances of input leads

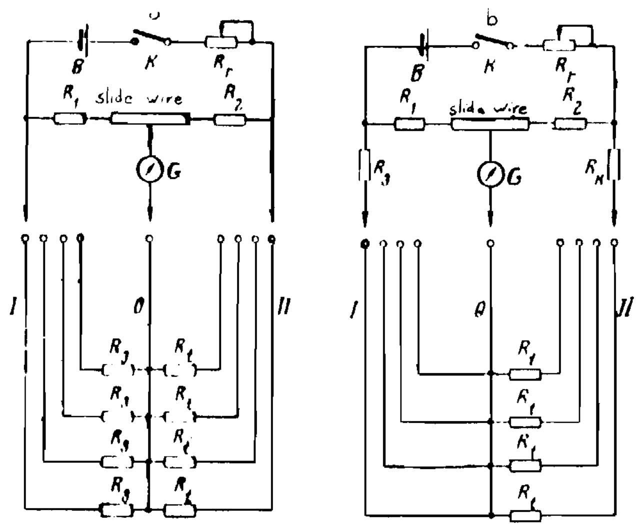


Fig. 137

Basic diagrams of measuring devices and thermometric assemblies consisting of resistance thermometers

a - two-leg thermometer; b - single-leg thermometer; R_1, R_2, R_3 - constant resistances; R_r - control resistance; B - battery; K - switch (button); R_t - temperature sensitive resistance; R_K - temperature compensator; G - zero deviation galvanometer; 0, I, II - input leads

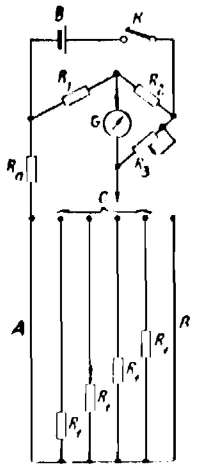


Fig. 138

A thermometric assembly with allowances for the resistance of input leads
A - common wire; B - auxiliary wire; C - input leads; the rest as in Fig. 140

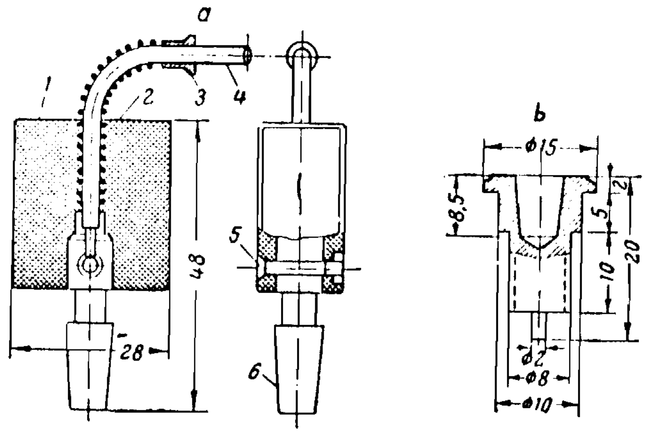


Fig. 139

Design of a measuring plug (a) and socket (b)
1 - ebonite block; 2 - coiled spring; 3 - bushing; 4 - conductor; 5 - pin; 6 - conical plug

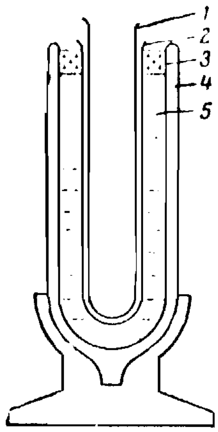


Fig. 140

Apparatus for the production of cryohydrates
1, 2 - test tubes; 3 - cork; 4 - Dewar flask; 5 - salt solution

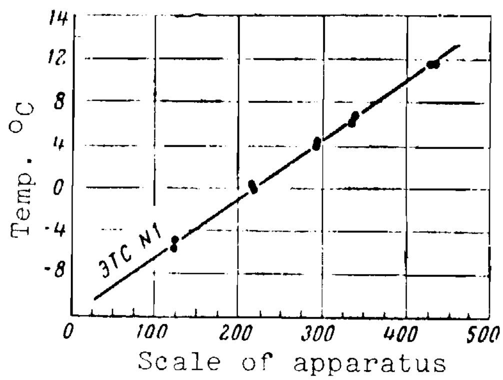


Fig. 141

Example of a calibration curve

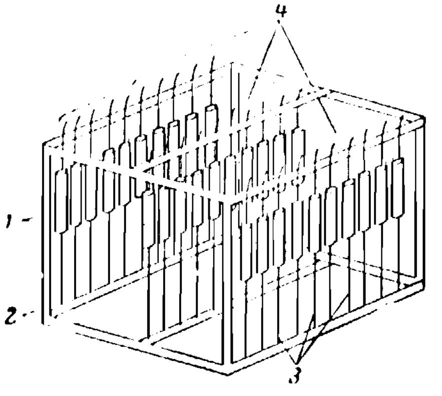


Fig. 142

Device for calibrating thermistors. 1 - thermistors; 2 - wire frame; 3 - soldered terminals of thermistors; 4 - terminals for connecting wire

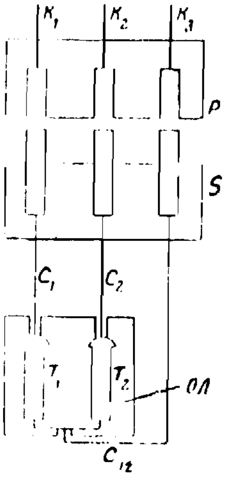


Fig. 143

Thermistors connected to a cable lock
K_1, K_2, K_3 - cable strands;
S - plug for cable lock;
P - socket of the cable lock;
T_1, T_2 thermistors;
C_1, C_2, C_{12} - connecting wires

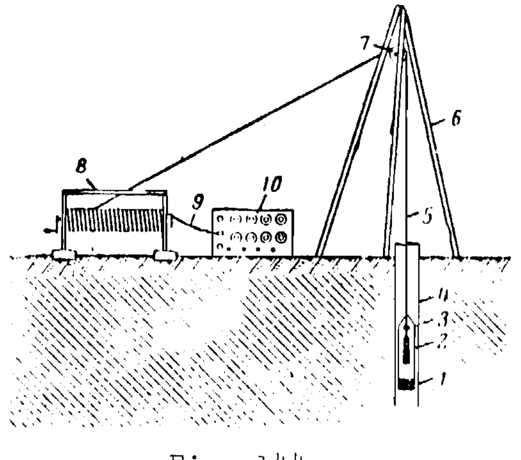


Fig. 144

Location of equipment during measurements
1 - lead weight; 2 - probe; 3 - cable joint; 4 - casing; 5 - cable; 6 - rig; 7 - balancing block and counter; 8 - cable reel; 9 - flexible cable; 10 - resistance bridge

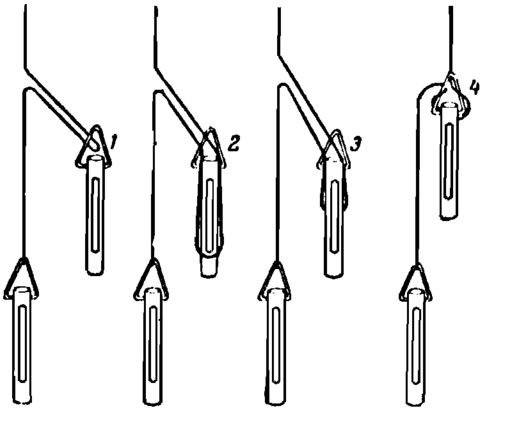


Fig. 145

Sequence of operations on attaching thermometer to cord for temperature measurements