Temperature, thickness and discontinuity of permafrost
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PREFACE

This translation is the ninth arranged by the Permafrost Subcommittee of the Associate Committee on Soil and Snow Mechanics of the National Research Council of Part I of the Russian permafrost publication, "Principles of Geocryology".

The first translation in this group was of Chapter VI entitled "Heat and Moisture Transfer in Freezing and Thawing Soils" by G.A. Martynov (TT-1065). The second was of Chapter IV "General Mechanisms of the Formation and Development of Permafrost" by P.F. Shvetsov (TT-1117). The third was Chapter VII "Geographical Distribution of Seasonally Frozen Ground and Permafrost" by I.Ya. Baranov (TT-1121). The fourth was "Ground (Subsurface) Ice" by P.A. Shumskii (TT-1130). The fifth was "Ground Water in Permafrost" by V.M. Ponomarev and N.I. Tolstikhin (TT-1138). The sixth was "Cryogenic Physico-Geological Phenomena in Permafrost Regions" by S.P. Kachurin (TT-1157). The seventh was "Perennially Frozen Ground and Vegetation" by A.P. Tyrtikov (TT-1163) and the eight "On Physical Phenomena and Processes in Freezing, Frozen and Thawing Soils" by N.A. Tsytovich et al. (TT-1164). The remaining four chapters I, II, III and XIII will not be translated. The first three chapters describe the history of permafrost investigations in the U.S.S.R. and the last chapter presents a summary of the preceding chapters. It is hoped that selected chapters of Part II (Engineering) will be translated in future.

This translation of Chapter VIII by V.A. Kudryavtsev describes the temperature regime at the ground surface and the seasonal freezing and thawing of the upper layer of the earth's crust. This is followed by a discussion of ground temperatures at the bottom of the layer with seasonal temperature variations. The influence of soil and rock type, bodies of water, and ground water on the thickness of permafrost are reviewed. The chapter concludes with an account of the influence of climatic and terrain features on the boundary conditions at the permafrost table and lower surface.

This translation has been prepared by G. Belkov of the N.R.C. Translations Section and checked by R.J.E. Brown of the Northern Group of the Division of Building Research.

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TEMPERATURE, THICKNESS AND DISCONTINUITY OF PERMAFROST

Introduction

The formation and development of permafrost from the thermo-physical point of view are determined by the following factors:

(1) boundary conditions on the upper and lower surfaces of the permafrost;

(2) the composition of the permafrost and the processes taking place in it.

These conditions are connected with the physico-geographical and geological environment in which the permafrost originated and exists.

The boundary conditions at the permafrost table are the result of the interaction of a number of complex processes. In the first place, at the surface of the earth, radiant solar energy is transformed into heat energy and heat exchange develops between the lithosphere and the atmosphere. As a result a specific temperature regime is established in the atmosphere and in the upper soil horizons. The latter is of particular importance for geocryologists since it reflects the boundary conditions at the permafrost table. The boundary conditions beneath the permafrost will be considered in the section dealing with the thickness of the permafrost.

Initially we will deal with the thermo-physical aspect of the freezing of the upper horizons of soil and will subsequently show the effect of various conditions on the formation of the temperature regime and the freezing of the soil.

The problem of the freezing of a medium, formulated by Lamé and Clapeyron in 1831 (the so-called Stefan problem), had not been solved until recently when in 1947 a solution was suggested by L.I. Rubinshtein although in a very general form owing to the extreme complexity of the problem. At the present time a new solution of this problem has been obtained by V.G. Melamed (1957) by reducing the Stefan problem to a system of ordinary differential equations. This method makes it possible to obtain a specific solution with any degree of accuracy.

According to V.G. Melamed if a single zone is considered $0 < x < \xi(t)$ where $\xi = \xi(t)$ - the coordinate of the boundary of the interface and the effect of the second zone on their boundary is taken into account temporarily by some limited function $q_1 = q_1(t)$, the statement of the problem has the form

$$\frac{\partial u(x, t)}{\partial t} = a^2 \frac{\partial^2 u(x, t)}{\partial x^2}; \quad 0 < x < \xi(t);$$

$$u(0,t) = \Phi(t);\ u(x,0) = \varphi(x)^{*};\ 0 < x < \xi(0); \tag{8.1}$$

$$u[\xi(t),t] = 0;$$

$$\lambda \frac{\partial u(x,t)}{\partial t}\Big|_{x=\xi(t)} - q(t) = \xi'(t), \tag{8.2}$$

where
$$\lambda = \frac{\lambda_1}{Q},\ q(t) = \frac{q_1(t)}{Q};\ \xi(0) \neq 0.$$

Make the substitution: $u(x,t) = v(x,t) + \Phi(t)\frac{\xi(t) - x}{\xi(t)}$, then $v(0,t) = 0$, $v[\xi(t),t] = 0$.

In this solution the possibility is shown of expanding $v(x,t)$ into a Fourier series for the section $(0,\xi(t))$ with a fixed t.

Then, $v(x,t) = \frac{2}{\xi(t)}\sum_{i=1}^{\infty} A_i(t)\sin\frac{i\pi x}{\xi(t)}$, where $A_k(t) = \int_0^{\xi(t)} v(x,t)\sin\frac{k\pi x}{\xi(t)}dx.$

For $A_k(t)$ and $\xi(t)$ an infinite system of differential equations is formulated. Multiplying (8.1) and (8.2) by $\sin\frac{k\pi x}{\xi(t)}$ and integrating from 0 to $\xi(t)$, and also keeping in view

$$\int_0^{\xi(t)} \frac{\partial^2 u(x,t)}{\partial x^2}\sin\frac{k\pi x}{\xi(t)}dx = -\frac{k^2\pi^2}{\xi(t)^2}A_k(t);\ \int_0^{\xi(t)} \frac{\partial u(x,t)}{\partial t}\sin\frac{k\pi x}{\xi(t)}dx = A_k'(t) +$$

$$+\frac{\Phi'(t)\xi(t)}{k\pi} - \frac{\Phi(t)}{k\pi}\xi'(t)(-1)^k + \frac{2k}{\xi(t)}\xi'(t)\sum_{i=1}^{\infty} A_i(t)\,w_{ik};$$

$$w_{ik}\begin{cases}(-1)^{i+k+1}\dfrac{i}{i^2-k^2} & i \neq k\\[2mm] -\dfrac{1}{4}k\ \ i = k\end{cases}$$

We obtain

$$A_k'(t) = -\frac{a^2k^2\pi^2}{\xi(t)^2}A_k(t) - k\xi'(t)\left[\frac{2}{\xi(t)}\sum_{i=1}^{\infty}A_i(t)\,w_{ik} - \frac{\Phi(t)}{k^2\pi}(-1)^k\right] - \frac{\Phi'(t)}{k\pi}\xi(t) \tag{8.3}$$

$$\xi'(t) = \frac{2\lambda\pi}{\xi(t)^2}\sum_{i=1}^{\infty} iA_i(t)(-1)^i - \frac{\lambda\Phi(t)}{\xi(t)} - q(t). \tag{8.4}$$

In the paper of V.G. Melamed it is shown that the solution of the system (8.3), (8.4) can be obtained from the solution of a shortened system by passage to the limit when $n \to \infty$:

$$A_k^{(n)\prime}(t) = \frac{a^2k^2\pi^2}{\xi^{(n)}(t)^2}A_k^{(n)}(t) - k\xi^{(n)}(t)'\left[\frac{2}{\xi^{(n)}(t)}\sum_{i=1}^{n}A_i^n(t)\,w_{ik} - \frac{\Phi(t)}{k^2\pi}(-1)^k\right] -$$

$$-\frac{\Phi'(t)\,\xi^{(n)}(t)}{k\pi}, \tag{8.5}$$

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* It is assumed that $\varphi'(x)$ and $\Phi'(t)$ are continuous, $\varphi''(x)$ is integratable and $\varphi(0) = \Phi(0)$.

$$\xi^{(n)'}(t) = \frac{2\lambda\pi}{\xi^{(t)^2}} \sum_{i=1}^{n} i A_i^{(n)}(t)(-1)^i - \frac{\lambda\Phi(t)}{\xi^{(n)}(t)} - q(t). \tag{8.6}$$

Moreover the solution of the system (8.5) - (8.6) is considered to be the only one possible.

Considering in a completely analogous way the second zone $\xi(t) < x < l$, $u_2(x,t) \doteq \Phi_2(t)\,\frac{x - \xi(t)}{l - \xi(t)} + \frac{2}{\xi(t)} \sum_{i=1}^{n} B_i(t) \sin \frac{i\pi(l - x)}{l - \xi(t)}$, we obtain a complete system of differential equations reflecting the dynamics of the temperature field in the case of a two-phase system with a mobile boundary*.

$$A_k'(t) = -\frac{a_1^2 k^2 \pi^2}{\eta(t)^2} A_k(t) - k\eta'(t)\left[\frac{2}{\eta(t)} \sum_{i=1}^{n} A_i(t)\,\omega_{ik} - \frac{\Phi_1 t}{k^2}(-1)^k\right] - \frac{\Phi'(t)}{k}\eta(t), \tag{8.7}$$

$$B_k'(t) = -\frac{a_2^2 k^2 \pi^2}{[L-\eta(t)]^2} B_k - k\eta'(t)\left[\frac{2}{L-\eta(t)} \sum_{i=1}^{n} B_i(t)\,U_{ik} - \frac{\Phi_2(t)}{k^2}\right] -$$

$$-\frac{\Phi_2'(t)}{k}\eta(t)(-1)^k \tag{8.8}$$

$$\eta'(t) = \frac{\alpha_1}{L-\eta(t)}\left[\frac{1}{L-\eta(t)} \sum_{i=1}^{n} iB_i(t) + \Phi_2(t)\right] - \frac{\alpha_2}{\eta(t)}\left[\frac{1}{\eta(t)} \sum_{i=1}^{n} iA_i(t)(-1)^i - \Phi_1(t)\right]. \tag{8.9}$$

where

$$\eta(t) = \frac{1}{\pi}\xi(t);\quad L = \frac{1}{\pi}l;\quad \alpha_2 = \frac{\lambda_2}{Q\pi^2};\quad \alpha_1 = \frac{\lambda_1}{Q\pi^2};\quad U_{ik} = \begin{cases} \dfrac{i}{i^2 - k^2} & i \neq k. \\[2mm] \dfrac{1}{4k} & i = k \end{cases}$$

The solution of the system of differential equations obtained can be carried out by the Euler method with automatic selection of steps.

To obtain a solution with accuracy of the order of 0.01 the time step should be about 10 days (200 - 300 hours). Knowing at the outset the right-hand part of equation (8.9), we find η' for this instant and substituting the value obtained for η' in (8.7) and (8.8) we find $A_k(0 + \Delta t)$, $B_k(0 + \Delta t)$. Finally substituting these values in the expression for $u_1(x,t)$ and $u_2(x,t)$ we have:

$$u_1(x, t_i) = \frac{1}{\xi(t_i)}\left[\Phi_1(t_i)\,x + 2\sum_{k=1}^{n} A_k(t_i) \sin \frac{kx}{\eta(t_i)}\right] \text{when } 0 < x < \xi(t_i),$$

$$u_2(x, t_i) = \frac{1}{l - \xi(t_i)}\left\{\Phi_2(t_i)\left[x - \xi(t_i)\right] + 2\sum_{k=1}^{n} B_k(t_i) \sin \frac{k(l-x)}{L - \eta(t_i)}\right\}$$

$$\text{when } \xi(t_i) < x < l$$

and we obtain the temperature of the soil at any depth at the time $t = 0 + \Delta t$.

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* An increase in the number of phases results only in an increase in the number of equations.

It is simpler to find $A_k(t_{n+1})$ and $B_k(t_{n+1})$ using the formulae:

$$A_k(t_{n+1}) = A_k(t_n)\,e^{-\psi_k(t_n)h} + T_k(t_n)\cdot h;$$

$$B_k(t_{n+1}) = B_k(t_n)\,e^{-\theta_k(t_n)h} + R_k(t_n)\cdot h,$$

where h - a step with respect to time.

$$\psi_k(t_n) = \frac{a_1^2 k}{\eta(t_n)^2};$$

$$\theta_n(t_n) = \frac{a_2^2 k^2}{[L-\eta(t_n)]^2};$$

$T_k(t_n)$ and $R_k(t_n)$ - the values of the right-hand parts of (8.7) and (8.8) respectively when $t = t_n$ without $\psi_k(t_n)A_k(t_n)$ and $\theta_k(t_n)B_k(t_n)$.

Because in most cases in practice u(x,0) is given numerically (or graphically), $A_k(0)$ and $B_k(0)$ can be obtained at least by the Simpson method. If the numerical value of the boundary conditions ($\Phi_1'(t)$ and $\Phi'(t)$ are unknown), in equations (8.7) and (8.8) one must introduce a substitution of the type $z = A_k(t) + \frac{\Phi_1(t)\eta(t)}{k}$. The system (8.7), (8.8) and (8.9) when n = 1,2,3,4 and under the condition

$$u_1(0, t) = -3-7{,}5\sin\frac{2\pi}{T}t;\ T = 8760\ \text{hours},$$

$$u_2(l, t) = \text{const} = -3;\ l = 19{,}4\ \text{м},$$

was computed with the machine "Strela" at the computation centre of the Moscow State University. From a comparison of the results using various values for n it follows that the convergence of a series of approximate solutions is rather rapid. Beginning with n = 2 the solutions differ little from each other and therefore for specific cases can be limited to this number.

With this condition a solution may be obtained on a manually operated computer. A large number of the time consuming operations can be nomographed which simplifies the computations.

This solution of the Stefan problem permits the finding of the mechanisms of the seasonal freezing of soil as well as the formation of permafrost. Because of this we will use it for analyzing the development of seasonally freezing and perennially frozen zones.

Temperature Regime of the Surface of the Soil and the Seasonal Freezing and Thawing of the Upper Layer of the Earth's Crust

The consideration of boundary conditions at the surface of the soil will be based on the interrelationship between the temperature regime and the depth to which the soil freezes. This relationship is shown in a classification diagram (Fig. 27) of the conditions and depth of seasonal freezing and thawing of the soil. The idea behind this diagram is explained below.

The depth of seasonal freezing and thawing of the upper layer of the earth's crust as well as the temperature regime are determined by conditions of heat exchange between the soil and the atmosphere. Therefore the thickness of the seasonal freezing and thawing layers should be determined from data on the heat cycle and thermal balance of the soil. However, the quantitative aspect of the heat cycle in the soil is not well known and therefore the discussion must be restricted only to giving those simple mechanisms in the freezing and thawing of the earth's crust which have been established at the present time.

Boundary conditions of heat exchange at the surface of the soil are determined by geographic factors. Their influence is reduced to variations in the mean annual temperature and the temperature range at the surface of the soil. The nature of the temperature field and the propagation of thermal waves in the various soil horizons are governed by geological factors.

Under the same boundary conditions soils of different composition and structure freeze to different depths. On the other hand the same soil but under different boundary conditions also freezes to different depths.

Because of this the classification of layers of seasonal freezing and thawing should be constructed with respect to the use of two mutually related indicators: the temperature regime at the surface and the composition of the soil.

In the classification given the subdivisions, with respect to mean annual temperatures of the soil, are based on the considerations given below.

The periodic changes in the mean annual temperatures of the air and soil have a range of from 0 to ±1°C; variations of ±1 to ±2°C occur only for individual years. Because of this when there is a perennial mean annual temperature with a range of 0 to ±1°C there is a periodic transition of the temperature through the 0°C point and correspondingly the seasonally freezing layer transforms periodically to a seasonally thawing layer. Thus at temperatures from 0 to ±1°C the seasonally thawing and seasonally freezing layers are called transitional. In the temperature range of ±1 to ±2°C such a transition can occur only occasionally in individual years; correspondingly this type of seasonal freezing and seasonal thawing is called semi-transitional. In the range of ±2 to ±5°C the transition through 0°C can occur only as a result of prolonged changes in the conditions of heat exchange between the soil and the atmosphere; correspondingly this type of freezing and thawing is called prolonged stable. At temperatures above +5°C and below -5°C even prolonged changes in climatical conditions do not bring about a transition through th 0°C point and therefore this type of seasonal freezing and thawing is called stable. Temperatures below -10°C are characteristic of severe climatic conditions and therefore the type of layers of soil with changes of temperature reaching -15°C is called arctic and below -15°C - polar. Temperatures above 10°C occur only in southern regions, above 14°C in subtropic regions and above 18°C in the tropics. These layers are called southern, subtropical and tropical, respectively.

When the above 0°C mean annual temperature is equal to the temperature range of the soil, seasonal freezing does not occur since the minimum temperature of the soil barely reaches 0°C and will nevertheless remain above freezing. But taking into account the fact that the mean annual temperature varies from year to year by ±2°C as a maximum, the conditions excluding the possibility of a seasonally freezing layer can be written in the form

$$t_m > \frac{A}{2} + 2^\circ,$$

where A - the meteorological amplitude in temperature of the soil at the surface equal to twice the usual physical amplitude in the temperature of the soil.

The condition for the development of a stable seasonally frozen layer would correspondingly be

$$t_m < \frac{A}{2} - 2^\circ.$$

Hence it follows that in the interval between these two limits, i.e. under the condition

$$\frac{A}{2} + 2 > (t) > \frac{A}{2} - 2^\circ$$

there will be unstable types of seasonally freezing and seasonally thawing layers. Depending on the ratio of the mean annual temperatures and amplitudes of the soil temperatures it is convenient to subdivide the unstable types of seasonal freezing into the following subtypes:

when $\frac{A}{2} - 1 > (t) > \frac{A}{2} - 2$ – occasionally disappearing,

when $\frac{A}{2} > (t_m) > \frac{A}{2} - 1$ – periodically disappearing,

when $\frac{A}{2} + 1 > t_m > \frac{A}{2}$ –periodically developing,

when $\frac{A}{2} + 2 > t_m > \frac{A}{2} + 1$ –occasionally developing,

when $t_m > \frac{A}{2} + 2$ –daily freezing of the soil; seasonal freezing is completely absent

A similar subdivision can be made also for the classification of seasonally thawing soil.

A similar derivation of subtypes depending on the amplitude of soil temperature was based on the changes in depth of seasonal freezing and thawing of the soil corresponding to changes in the amplitude of soil temperatures. It is known for example that in the European part of the U.S.S.R., in Western Siberia and in Yakutia these depths vary. It is therefore convenient to introduce the following numerical gradations with respect to amplitudes corresponding to specific climatic regions and simultaneously derive types of seasonal freezing and thawing:

when $A < 15°C$ - sea coasts - maritime,

when $A = 15$ to $22°C$ - sea coasts - moderate maritime,

when $A = 22$ to $27°C$ - European part of the U.S.S.R. to the Urals - moderate continental,

when $A = 27$ to $34°C$ - Western Siberia - continental,

when $A = 30$ to $42°C$ - central Siberian Plateau - elevated

when $A = 42$ to $48°C$ - Eastern Siberia - severe continental,

when $A > 48$ - central part of the Siberian mainland - extreme continental.

The classification of seasonally freezing and thawing layers by geological indicators should be based on the lithological features and moisture content of the soils comprising these layers. With respect to lithological features and water permeability 9 types of soil are identified: (1) peat, (2) clay, (3) clay loam, (4) sandy loam, (5) silt, (6) fine sand, (7) medium sand, (8) coarse sand, (9) gravel and stones in the absence of fine-grained material.

With respect to moisture content it is convenient to identify the following three types of seasonal freezing and thawing.

1) $w_u < w_f < w_u + \frac{1}{3}(w_f - w_u)$ - deep freezing and thawing,

2) $w_u + \frac{2}{3}(w_f - w_u) > w > w_u + \frac{1}{3}(w_f - w_u)$ - medium freezing and thawing;

3) $w_f > w > w_u + \frac{2}{3}(w_f - w_u)$ - shallow freezing and thawing.

Here w - natural moisture content of seasonally freezing or seasonally thawing soil;

w_f - full moisture capacity of the layer;

w_u - quantity of unfrozen water at the given temperature.

The above classification scheme can be used to find the general mechanism in the latitudinal zone for the distribution of seasonal freezing and thawing layers with respect to depth.

The scheme takes in the basic natural features which determine the depth of seasonal freezing and thawing of the soil. The depth (accurate to centimetres) can be determined for each specific point only with the simultaneous accounting of the four fundamental features: the mean annual temperature, the amplitude in the variations of the mean monthly temperatures, the lithological composition of the soil and the moisture content.

A similar approach is applicable also for vertical zones with the one difference that changes in the distribution of seasonal freezing and seasonal thawing of the soil with respect to depth will be taken from sea level up to higher altitudes instead of from south to north. The effect of latitude and altitude is complicated by changes in lithological composition of the soil and its moisture content.

The derivation of types of seasonal freezing and thawing and knowledge of the laws governing their formation provides a new approach to the mapping of this phenomenon. The four main factors noted above should be plotted on the map (Table XVI).

The Temperature of the Soil at the Bottom of the Layer With Seasonal Variation in Temperature

Let us now consider the formation and development of the temperature regime of the seasonal freezing and thawing of the soil, i.e. the boundary conditions at the permafrost table. The general mechanisms in the formation and development of the temperature regime are considered in Chapter IV. We will therefore deal only with some particular mechanisms depending on specific conditions.

It is generally known (Voeikov, 1904; Kudryavtsev, 1954) that the mean annual temperature of the air and of the soil are not the same; the mean annual air temperature in the permafrost zone is usually lower than the mean annual soil temperature by several degrees. The difference between them is denoted by S_z, then

$$S_z = t_s - t_a, \tag{8.10}$$

where t_s is temperature of the soil and

t_a is temperature of the air.

The inequality of the soil and air temperatures is due to a variety of geological and geographic factors. It would be desirable to determine the contribution of each of the component factors. It would then be possible to determine the mean annual soil temperature from the mean annual air temperature and vice versa.

An important factor in forming the temperature regime of the soil is snow cover. This question is very complex and extensive and therefore we will deal only with the dependence of the mean annual temperatures and annual temperature amplitudes of the soil on snow cover. In most cases snow cover results in an increase in the mean annual temperature of the soil. Let us consider first a case when snow cover during the wintertime plays the role of a heat insulator.

We will use the emperically established relationship (Kudryatsev, 1954)

$$S_z = \frac{A_{mm}}{2}\left(1 - \frac{1}{f}\right), \tag{8.11}$$

where A_{mm} - the annual amplitude of the mean monthly air temperature, and

$$f = e^{+z\sqrt{\frac{\pi}{kT}}}$$

where z - thickness of snow cover,

k - heat conductivity coefficient for snow,

T - period of oscillation equal to one year.

Substituting the value of S_z from (8.11) we obtain:

$$t_s = t_a + \frac{A_{mm}}{2}\left(1 - \frac{1}{f}\right). \tag{8.12}$$

Introducing the correction for the change in the mean annual air temperature for height $\Delta t = 1/213 \frac{\text{degrees}}{\text{metres}}$, we obtain:

$$t_s = t_a + H\Delta t + \frac{A_{mm}}{2}\left(1 - \frac{1}{f}\right); \tag{8.13}$$

where t_a - the mean annual temperature of the air at sea-level;

H - the height of the place in metres.

From this relationship it follows that the mean annual temperature of the soil increases with the mean annual temperature of the air.

The value of S_z depends on the thickness and density of the snow cover and on the amplitude of the air temperature. The value of $(1 - \frac{1}{f})$ depending on thickness h and density ρ and the coefficient of heat conductivity k of the snow cover are given in Table XVII.

In using formula 8.13 one should take the calculation thickness of the snow cover h_{calc} to be

$$h_{calc} = k_c \cdot h_{max},$$

where h_{max} - is the maximum mean depth of snow cover taken over a 10 year period,

k_c - the coefficient for determining the calculation thickness of snow cover (Table XVIII).

The approximate dependence between the thickness and density of snow cover is given in Table XIX.

Using the data of Tables XVII, XVIII and XIX and knowing the annual mean monthly temperature amplitude of the air (according to meteorological handbooks or by direct measurement) one can calculate the value of S_z.

To facilitate computation, Table XVII gives the most frequently encountered values of $(1 - \frac{1}{r})$ depending on the thickness of the snow cover.

With formula 8.13 the dependence of temperature of the soil on the thickness and density of snow cover can be refined and a quantitative estimate can be given.

It is important to note the different insulating effect of snow cover under different climatic conditions. The greater the amplitude of the air temperature, under otherwise equal conditions, the greater the insulating effect of snow cover. With the same thickness and density of snow cover and with the same mean annual temperature of the air under conditions of continental climate, the mean annual temperature of the soil will be higher than in a maritime climate. Therefore the insulating effect of snow cover can be considered only with respect to climate. In this connection to increase the mean annual temperature of the soil by 1°C different amounts of increase in snow cover are required.

For illustration we present Tables XXI and XXII.

According to formula 8.13 the mean annual temperature of the soil is determined by the following factors: the mean annual temperature of the air at sea-level, the altitude of the location, the annual amplitude of the mean monthly air temperatures, thickness and density of snow cover. The value of each of these factors is shown in Table XXIII.

The remaining climatic factors are not considered in formula 8.13. This, however, does not mean that they do not exert an influence on the formation of mean annual temperatures of the soil. Some of them show up in the mean annual temperature of the air and its amplitude, whereas others are generally not taken into account since their values are as yet unknown. By virtue of this, formula 8.13 elucidates only one aspect of the phenomenon and does not express the general picture. The mean annual temperature depends also on all other factors of the geological and geographic medium which will be considered below.

In the comparison of results of calculations using formula 8.13 with a large number of data it has been established that the mean quadratic error is ±0.5°. This accuracy is completely sufficient for the practical application of formula 8.13 in calculating the mean annual temperature of the soil from meteorological data.

Using formula 8.13, calculations were made of the difference between the mean annual temperature of the air and of the soil (S_z) for the entire territory of the U.S.S.R. from which a map was drawn (Fig. 28). On this map it can be seen that the value of S_z increases with the thickness of the snow cover and with the extent to which the climate is continental. The maximum value of 7° for S_z is reached in the region of Igarka and in central Yakutia. In some regions the value of S_z reaches 11°C (Middle Kolyma), where heavy snow cover is observed (up to 1.1 m) with a relatively large annual amplitude in the mean monthly temperatures of the air (up to 55 - 60°C).

In the south, where the thickness of the snow cover may be 0, S_z will also be equal to 0. By insulating the soil, the snow results in a decrease of the annual mean monthly amplitude of the soil temperature. If one takes the value of the amplitude at the surface of the snow to be equal to the amplitude of the air temperature, under the snow at the surface of the vegetation cover the amplitude will be less by the value of S_z. Correspondingly, the meteorological amplitude will be decreased by $2S_z$. Thus using formula 8.13 and the map showing the isolines of S_z one can determine the decrease in amplitude of the temperature due to snow cover.

Using formula 8.13, a computation was made of the critical thickness of snow cover at which the mean annual temperature of the soil must be equal to 0 under otherwise equal conditions. The computation was carried out for a network of points distributed over the entire territory where permafrost occurs and from these data a map was drawn (Fig. 29). The values of the critical thickness of snow cover (h_k) have been converted to sea level. For computing these values taking into account absolute altitude, annual temperature amplitude of the air and density of snow cover, two nomograms (Fig. 30 and 31) are applied to this map. The isolines of h_k are drawn on the map at an interval of 0.25 m from 0 to 15 m; south of the 0°C line, isolines are given for the altitude of the place where the mean annual temperature is equal to 0°C. These isolines are dotted.

The second nomogram (Fig. 31) shows that the denser the snow the greater will be its critical thickness and vice versa. The map does not show isolines of the thickness of snow cover above 1.5 metres which is observed very rarely in some negative types of relief (ravines, gorges, etc.). Thus north of the isoline of 1.5 metres the 0°C temperature of the soil and the separation of the permafrost from the seasonally freezing layer cannot be explained by the thickness of snow cover and is the result of other causes. In the territory between the isolines of 1 and 1.5 m this phenomenon can be explained by the influence of snow only in exceptional cases. Finally, south of the 1 m isoline snow cover may be one of the fundamental causes for the separation of permafrost from the seasonally freezing layer.

These nomograms can be used also for determining t_s of the soil and S_z for various densities and thicknesses of snow cover which is particularly important for regions where density and thickness of snow cover are not uniform. This refers for example to tundra regions where as the result of snow mobility and snow transfer the density and thickness of the snow is not uniform over individual micro sections of the region. Great differences in thickness of snow cover is also observed in fold-mountain regions.

Using formula 8.13, computations were made of the mean annual temperature of the soil for a number of points (200 calculation points and 400 observation points) for the territory of the U.S.S.R. (Fig. 32). This map shows the temperature zones with respect to latitude and the soil temperatures have been converted to sea level; the map gives a good indication of the influence of the Atlantic, Arctic and Pacific Oceans on the formation of the geothermal zones. As a result the isolines protrude far south in the central part of the continent and rise to the north in the west and east. The distance between the isolines with a temperature spacing of 1°C, called the latitudinal temperature stage, varies from 70 - 80 up to 400 - 500 kilometres and has a mean value of 150 - 300 kilometres.

The minimum mean annual temperature of the soil at sea level reaches -11°C in the Taimyr Peninsula and in the delta of the Lena River.

The isotherms on this map were drawn only with consideration of climatic factors (snow, mean annual air temperature and its annual amplitudes). The influence of other factors should be considered in addition.

It should be noted that snow cover acts as an insulator and moreover reflects and absorbs radiant energy. Its influence on the temperature regime of the soil depends on the combination of the above-mentioned influences and will be different in different regions. In winter the snow reflects a considerable part of the incident solar energy and retards radiation from the soil to the atmosphere. In the spring a considerable part of the solar energy is spent on melting the snow. In this case snow is a factor acting to cool the soil.

In high latitudes where there is evaporation of snow at low air temperatures the thickness of snow cover is greatly reduced by the time the air temperature increases to above freezing values and thus relatively less solar energy is spent on melting the snow; however, the cooling effect of snow on the soil occurs much earlier than the advent of warmer air temperatures.

In the far south where the thickness of snow cover is measured in several centimetres its insulating effect is relatively small and can be neglected. The main effect here, is the reflection of radiant energy from the white surface of the snow. In this case the snow acts to cool the soil and the mean annual temperatures would therefore be lower than the mean annual temperature of the air. This effect has been noted for the northern Caucasus (Stotsenko, 1953).

No one has yet made a detailed study of the role of snow cover in the thermal exchange between the soil and the atmosphere and it is therefore difficult to give it any quantitative estimation. Qualitatively it can be characterized by the graph shown in Fig. 33.

When the snow cover is light (from 0 to h_1) the cooling effect of snow is predominant; its insulating effect is close to zero and snow acts to cool the soil. As the thickness of the snow cover increases its insulating effect increases with a relatively small increase in reflecting capability. When the snow cover reaches the value of h_2 the insulating effect of snow decreases owing to the expenditure of energy on melting the snow and the reflection of solar energy from the snow surface. Consequently, as the thickness of the snow cover increases its insulating effect gradually decreases and subsequently has a cooling effect*. When the thickness of snow cover reaches h_3 the snow does not have time to melt during the summer and firns and glaciers begin to be formed. From this time on the white surface of the firn or glacier reflects part of the solar energy all year round and thus cools the soil to the maximum degree.

Thus Fig. 33 illustrates a complex relationship between the changes in the temperature of a frozen soil stratum and the thickness of the snow or ice cover.

Snow cover also has a great influence on the depth of seasonal freezing of the soil. In a section where the snow cover has been removed the depth of freezing in some cases may be 50 - 60% greater than on a section with natural snow cover.

Let us consider this problem in more detail.

In the schematic diagram (Fig. 34) let us take two points - one corresponding to a plot with a soil temperature of t_1 under natural snow cover with an amplitude of A_2 and the second - a plot from which snow has been removed. The removal of snow results in a decrease in the mean annual temperature of the soil from t_1 to t_2. Correspondingly because of this change, at the same amplitude of A_2 there should be a change in the depth of seasonal freezing from h_1 to h_1'. However, the removal of snow cover increases the amplitude from A_2 to A_3 and consequently changes the depth of seasonal freezing of the soil from h_1' to h_2. Snow cover decreases the depth of seasonal freezing of the soil by increasing its mean annual temperature as well as by decreasing the amplitude.

For seasonal thawing a different mechanism is observed. Let us again consider two points in Fig. 35 - one for a section with natural snow cover and the other for a section from which the snow has been removed. The removal of

* Editor's note: The author does not take into account the duration of the existence of snow cover.

snow results in a decrease in temperature from t_1 to t_2. Correspondingly with unchanged amplitude A_2 there should be a decrease in the depth of seasonal thawing from h_1 to h_1'. However, the removal of snow cover leads to an increase in amplitude from A_2 to A_3. Correspondingly there should be an increase in the depth of seasonal thawing from h_1' to h_2.

Thus the absence of snow cover decreases the depth of seasonal thawing of the soil owing to a decrease in winter temperature and to an increase in the depth of seasonal thawing owing to an increase in amplitude. One compensates the other and the total effect of the snow cover is insignificant.

From the above one can see an important difference between the influence of snow cover on the formation of seasonal freezing and seasonal thawing of the soil.

In addition to the mechanisms considered above some regional climatic features connected with the formation of the mean annual temperature of the soil should be considered. One of these is the winter temperature inversion. In fold-mountain terrain and in regions of winter anticyclone the mean annual temperature of the air in the valleys is lower than on water divides and slopes. This phenomenon has not been studied to any extent and a quantitative characteristic cannot be determined. It is therefore difficult to connect it with mean annual temperatures of the soil. The changes in mean annual temperature and amplitude of the air with altitude has been quite fully determined for Yakutsk (Table XXIV).

From this table it is seen that for the first thousand metres the increase in temperature of the air is from -14.2°C to -10.9°C, i.e. almost 0.37°C per 100 metres. Approximately from the altitude of 1,000 metres upwards there is a decrease in temperature with altitude: from 1,000 to 2,000 metres - 0.27°C per 100 metres, from 2,000 to 3,000 metres - 0.47°C and from 3,000 to 4,000 metres - 0.50°C.

The annual mean monthly amplitude of air temperature in a hut was 63°C, at an altitude of 1,000 m - 48.1°C, at an altitude of 2,000 m - 39°C, at 3,000 m - 36°C and at 4,000 m - 35.9°C.

In connection with the winter temperature inversion the variation in mean annual temperature of the soil compared with the vertical temperature gradient can vary within the range of 0 to 6°C. This deviation is a maximum. It is impossible to give a general quantitative estimate.

<u>The position of the section with respect to relief</u>. Relief is an important factor and in many ways determines the temperature regime of the soil. An increase in altitude of 100 metres corresponds to a decrease in soil temperature of 0.5°C (formula 8.13). The change in the amplitude of air temperature

and in snow cover with altitude and also temperature inversion complicate this relationship and influence the change in vertical temperature gradient of the soil. Depending on the location of the gauge point the temperature of the soil may vary by 10 - 20°C and more.

It is known that the quantity of solar energy reaching the surface of the earth depends on the slope and exposure.

Correspondingly there is a difference in the temperature of the soil. As an illustration we give an example of variations in the temperature of the soil in eastern Transbaikal in Padi Mul'tsai (after P.I. Koloskov); the measurement was carried out on July 3-6 at 1:00 PM (Table XXV). The variation in mean annual temperature of the soil on a north-facing slope is governed by a decrease in annual amplitude and on a south-facing slope by an increase in amplitude.

The range between the mean annual temperature of the soil on north and south-facing slopes will vary depending on the steepness of the slopes. As the steepness of the slope increases this range will increase and vice versa. For slopes of up to 30° the difference may reach 1.5 - 2°C.

The exposure of slopes also influences the variation in the annual temperature amplitude of the soil. During the winter when the slope is covered with snow the temperature becomes approximately the same for all exposures. A large difference is observed in the summer when a change occurs in the temperature regime of the soil.

The temperature regime of the soil for north and south-facing slopes is shown schematically in Fig. 36 where the annual temperature variation with depth for a horizontal surface is plotted.

The effect of exposure may be complicated by other factors, for example, uneven distribution of snow cover. In this respect the region of Vorkuta is typical where many investigators have established the absence of any influence of exposure of slopes on soil temperature. Here the prevailing winds are from the south and southwest and snow cover blown from these slopes is deposited on the north and east-facing slopes which result in their relatively high temperature. In the summer, on the other hand, the south and west-facing slopes receive more heat than the north and east-facing slopes. As a result, these two factors compensate each other and the temperature of the soil on all exposures is equalized.

Micro, meso and macro relief affect the temperature of the soil primarily because of uneven distribution of the snow cover. Various types of positive forms of micro relief (for example mounds) usually have a relatively light snow cover and are frequently bare. On the other hand, all negative forms of micro relief are almost filled with snow. This one factor ensures a relatively high temperature for negative forms of relief as compared with positive forms. The influence of macro and meso relief on the temperature of the soil extends over considerable areas and depth; the effect of micro relief is limited to very small areas (frequently measured in fractions of a metre) and insignificant depth.

Many investigators have established that on the hummocky peat bogs of eastern Siberia the permafrost table is usually a mirror image of the surface relief. Under the mounds the permafrost table is deeper than between the mounds, resulting in a difference in the temperature regime of the soil particularly to the 0.5 m depth. At a depth of 2.5 m the temperature is equilized. For large mounds and depressions, up to several tens of metres in cross-section, the depth to which their influence is felt in the soil is correspondingly greater - it is equal approximately to the minimum cross-section. The amplitude of the temperature on the mounds increases because of winter cooling, as well as because of greater summer heating as compared with the depressions between the mounds where the amplitude is decreased.

In considering the role of relief in the formation of the temperature regime of the soil it is necessary to deal with the age of the relief and its importance with respect to the problem under consideration. Some investigators have noted that on relatively young relief the temperature of the soil is as a rule higher than on older forms of relief; it therefore becomes necessary to consider the process of formation of the permafrost in connection with the general history of the development of the relief.

It is quite natural that in considering permafrost under development one must also consider the dynamics of the medium in which it takes place. In this case one must keep in view the different rates of development of relief and permafrost. If the rate of permafrost development is comparable to the rate of change in relief, the history of the development of relief should be of primary importance; if these two rates differ substantially they cannot be in direct relationship to each other.

The thermal inertia of permafrost for various depths is different. For upper horizons within the range of about 50 metres it is measured from tens to several hundred years. According to the data of the Igarka Permafrost Station the thermal inertia for depths of 5 - 8 metres barely reaches 10 years. The same is observed for a number of points. As regards depths of several hundred metres, the thermal inertia is measured in thousands, tens of thousands and hundreds of thousands of years. The time of formation of river terraces is measured in thousands of years. Correspondingly the historical progress of the development of relief may be of decisive importance in the formation of only the lower horizons of the permafrost.

Since the mean annual temperature of the soil in the stratum with annual temperature variation is the result of present-day climatic conditions occurring during the past 10 - 20 years, the age of the relief cannot have any important effect.

Vegetation cover. The vegetation cover being on the boundary between the lithosphere and the atmosphere has an important effect on the temperature exchange between them and on the temperature of the soil. It has an influence on the amount of radiant energy absorbed and reflected by the surface; it protects the surface of the soil from the effect of winds and finally, to an important degree, it determines the moisture exchange between the air and the soil. The absorption of water by vegetation and transpiration, the precipitation of dew on vegetation, the decrease in evaporation from a surface covered with vegetation, etc., unquestionably have an important effect on the heat exchange between the upper soil horizons. The processes of moisture exchange occurring with the participation of vegetation cover are very complex and cannot in all respects be subjected to a quantitative estimation. The difficulty in taking them into account is complicated by the multiplicity in forms of the vegetation cover and its variation with respect to time. Forest, shrub, meadow and tundra vegetation have different influences on moisture exchange and heat exchange between the air and the soil.

All of this indicates that vegetation cover, in contrast to snow cover, cannot be regarded simply as some additional thermal resistance. Its influence on the temperature of the soil is varied and complex and requires further study.

The investigation of the influence of vegetation cover on the formation of the mean annual temperature of the soil was undertaken as a secondary study primarily by agroclimatologists. The only one who made a special study of this problem was P.I. Koloskov (1925) in connection with a general study of seasonal freezing but he had at his disposal only data published by agroclimatologists.

In recent times the V.A. Obruchev Permafrost Institute has made a new step towards the study of the part played by vegetation cover in the heat exchange between the soil and the atmosphere and the frozen subsoil (see Chapter XII).

Vegetation cover protects the soil from winter cooling and summer heating, thus reducing the amplitude of temperature variation. In the south the decrease in summer amplitudes will be greater than the decrease in winter amplitudes; therefore vegetation cover in the south has primarily a cooling effect on the soil and in the north a warming effect. In addition under different conditions the effect of vegetation cover will differ. For example, it has been established that for the Far East when there is a snow cover of up to 20 cm the removal of the forest results in a cooling of the soil but when the snow cover is deeper the removal of the forest results in warming of the soil. The increase in annual amplitude due to removal of the forest reaches 10 - 13°C.

After the cutting of the forest and removal of the vegetation cover in the Far East, either with a light snow cover or with a heavy snow cover, there is an increase in the mean annual temperature of the soil and the value of the increase depends on the thickness of the snow cover. It can be said that under conditions of a continental climate with a prolonged warm summer and severe winter, when there is a snow cover of 10 - 20 centimetres, the vegetation cover protects the soil from summer heat and from winter cold to approximately the same extent. With a heavier snow cover the warming effect of the vegetation cover during the winter will be greater than the cooling effect during the summer.

In northern regions the insulating effect of grass and shrub vegetation increases the mean annual temperature of the soil by not more than 1 - 2°C. Grass and shrub cover, like forest cover, decreases the annual temperature amplitude by approximately 15 - 25% and when taken together by 30 - 50%.

The role of moss cover in the formation of the temperature regime of the soil is determined by its low thermal conductivity, high moisture capacity and hygroscopic properties. The moisture capacity of Hypnum moss according to available data reaches 360% and for Sphagnums 1,300 - 5,000% (in relation to dry weight). To evaporate the moisture to produce 1 gram of dry moss requires up to 8,000 calories (Sumgin, 1937).

Under a moss cover there is a greatly reduced amplitude in temperature. During the winter the heat conductivity of frozen moss increases greatly. The resultant decrease in soil temperature causes a decrease in the mean annual temperature of the soil by 1 - 3°C. For regions with heavy snow cover there is an inverse relationship. The decrease in annual amplitude of the soil due to moss cover reaches 50 - 60% and when there is a thick moss cover it may be 80 and even 90%.

To determine the influence of vegetation cover on the depth of seasonal freezing and thawing we use the diagrams shown in Fig. 37.

Let us take a southern region where the vegetation cover has a cooling effect on the soil. Let us plot two points on the diagram: one for a section with natural vegetation cover and the second for a section from which the vegetation cover has been removed. In the first case the temperature will be lower and the amplitude less than in the second. The effect of vegetation cover in this case will be similar to the effect of slope exposure.

In regions of seasonal freezing the above-mentioned effect does not occur to any extent whereas in permafrost regions the removal of the vegetation cover results in a very rapid increase in the depth of seasonal thawing.

For the second case let us take a northern region where the vegetation cover has a warming effect on the soil. This case is diagrammatically represented in Fig. 38. Here the temperature of the section with natural cover will be higher and the amplitude lower than for a section from which the vegetation has been removed. The effect of the vegetation cover in this case will be similar to that of a snow cover. In a region where there is seasonal freezing the depth of freezing will increase greatly for the section that has been stripped of vegetation as compared with that retaining its natural cover. In the permafrost region there will be no important difference in the depth of freezing.

Lithological differences in soil composition have different thermo-physical properties and have different properties on freezing and thawing. These differences have an effect on the formation of the mean annual temperature and amplitude of the soil.

In this problem, moisture content and ice content of the soil are of great importance. They determine the expenditure of heat on freezing and thawing and have an important effect on the thermo-physical properties of the soil. Of greater importance is the filtration property since the infiltration of warm precipitation into deeper lying soil horizons bring in substantial quantities of heat. This quantity varies with climate and conditions of surface run-off and ground water. For example, in the Aldan Plateau the water divide is frequently comprised of highly weathered bedrock covered by coarse fragmented residual rock and talus. During the summer more than half of the annual precipitation occurs here which infiltrates and heats up the layer of residual rock and talus and also the fissured part of the weathered bedrock. As a result, in such regions permafrost is frequently absent and the temperature of the ground varies within the range of 1 to 2°C although by climatic data it should be of the order of -1 to -4°C. Such temperatures are in fact observed for areas composed of clay soil where there is no infiltration of warm precipitation.

Smaller differences in temperature resulting from this factor are observed by many investigators for highly varied regions. In the Selemdzha River region in the Far East where there are sandy gravelly soils and clay soils this difference reaches 2.8°C (Kudryavtsev, 1939). In western Siberia along the valleys of the Dakh, Elogui and Ob' Rivers the difference is 1.5 - 2.0°C (Popov, 1953_1). For the Vorkuta region approximately the same values are recorded.

Using this particular feature the research workers and mining engineers of Dalstroi developed a method of artificially thawing the upper horizons of the permafrost by passing water through the seasonally thawed layer. Using this method it is possible to thaw a layer of permafrost 4 - 8 metres deep during a summer. In addition to good infiltration an appropriate slope of the terrain is required which will ensure a satisfactory rate of filtration. Where the flow of ground water is blocked swampy areas are frequently formed within which the mean annual temperature of the soil is substantially lower than that of the surrounding area.

In the absence of extensive infiltration of surface water, the mean annual temperature may be 1 - 2°C lower than in areas where there is filtration if this difference is due entirely to the lithological character and moisture content of the soil. If the temperature of sandy loam is taken as an average the temperature of clay and peat soils will be 0.5 - 1°C lower and in sand and coarse gravel it may be 0.5 - 1°C higher. The infiltration of surface water and circulation of ground water may increase this difference to 3 - 5°C.

These data agree completely with the classification of different types of seasonal freezing and thawing of soils given above and the depths in both cases are determined by the temperature regime of the soil, its composition and moisture content. In this case it is interesting to note the difference depending on moisture content in the depth of seasonal freezing and thawing for sand and clay soils. For sand the depth of seasonal freezing decreases with increase in moisture content and for clay there is first an increase in depth of freezing with increase in moisture content up to a certain value of w_u corresponding to the maximum quantity of bound water at a given temperature but with a further increase in water content there is a similar effect to that observed for sand. These differences are explained as follows. With an increase in the moisture content there is an increase in the heat conductivity coefficient and also in the heat capacity of the soil as well as in the liberation of heat due to ice formation. The first factor leads to an increase in the depth of seasonal freezing and the second to a decrease.

Bound frozen water plays no part in the heat of crystallization and therefore its increase leads only to an increase in the coefficient of heat conductivity and an increase in the depth of seasonal freezing. For sand containing small quantities of frozen moisture the dominating role is played by the heat of crystallization. For clay within the range of 0 - w_u there is only an increase in the coefficient of heat conductivity. The heat of crystallization does not play a part in this case. When the moisture content is above w_u the dominating role is played by the heat of crystallization and the same effect is observed as indicated for sand.

Important factors in governing the depth of freezing and thawing of the soil are density, structure and conditions of occurrence of the soil. The heat conductivity coefficient increases with density which in general should result in an increase in depth of seasonal freezing and thawing; however a simultaneous variation in moisture content complicates this relationship.

For uniform soils and for laminar structures made up of different types of soils the depth of seasonal freezing and thawing will be different. In the first case the depth will be greater than in the second.

Salinization of the soil. There are no data in the literature on special studies devoted to the influence of salinization on the temperature regime of the soil. Some papers on permafrost indicate salinity as a factor favouring a decrease in temperature of permafrost.

On an experimental plot of the Anadyr' Permafrost Station the temperature of the upper horizons of highly saline clay soils is 1.0 - 1.6°C lower than for the surrounding areas (Shvetsov, 1938). In the region of the Kempendyai salt deposits (Yakut ASSR) the usual temperature of the soil is close to -3.5°C and in individual highly saline regions it reaches -8.0°C. Thus salinization results in a decrease in the temperature of permafrost since saline soils constitute a particular type of saline cooling mixture. The decrease in temperature of the soil due to this factor varies from 1 - 2°C (Anadyr') and from 4 - 5°C (Kempendyai).

Swampiness. The effect of swampiness on the temperature regime of the soil is generally known and has been noted frequently in the literature on permafrost. It is generally considered that swampy areas generally have temperatures of 0.5 - 1.0°C lower than on dry and well-drained areas. This regularity is observed in the Far East from the Amur-Yakutsk Road to Komsomol'sk*, in the Transbaikal (Kachurin, 1950) and in a number of other regions which have a light snow cover. For Vorkuta (Redozubov, 1946), western Siberia (Popov, 1953_1), Igarka (Meister, 1948) with heavy snow cover (0.8 - 1.0 m) an inverse effect is observed: in swampy areas the temperature is higher than where there is good drainage.

To explain the above-mentioned differences the following considerations may be advanced. In summer a large quantity of the heat is consumed by evaporation over swampy areas. As a result the heating of the soil is retarded and the temperature of the underlying horizons is decreased. In winter if there is a light snow cover there are sufficient heat losses for the heavily saturated soil to freeze and the underlying horizons undergo intensive cooling.

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* Report of the Baikal-Amur Road Expedition, holdings of the Permafrost Institute, Academy of Sciences, USSR, 1932-1935.

If there is heavy snow cover the heat transfer from the soil is retarded and as a result there is only partial freezing of the heavily saturated soil. During the summer this thin layer of frozen soil rapidly thaws and the soil heats up more than in areas that are well drained.

Thus, moisture saturation of soil leads basically to a decrease in mean annual temperatures but the effect is varied and, depending on the thickness of snow cover, may result in an inverse effect, i.e. an increase in the mean annual soil temperature.

Surface water run-off. The particular features of the temperature regime of the soils in valleys of large and medium size rivers have been noted by many geocryologists. It has been established that for the Yenisei River in the Igarka region there is a decrease in temperature along the cross-section of the valley from the river bed to the water divide by approximately 2.5°C (Meister, 1948). The same was noted for the Ob' River (Popov, 1953) and for the Lena River (Mel'nikov, 1951). The warming influence of these rivers is approximately of the same order, 2 - 3°C. A similar picture is observed in the valley of the Selemdzhi River (Kudryavtsev, 1939). Figure 39 shows a temperature profile along the route traversed indicating a transition of soil temperatures from above freezing to below freezing. As a result of the warming effect of the Selemdzhi River and its water flow, permafrost is absent in the flood plain and in the first terrace. The typical temperature regime for the region (-1.0°C) is observed only on the ancient terrace and the water divide. The same picture is observed also in northern regions (Ponomarev, 1952). Within the limits of the river bed the temperature of the underlying soil may differ from the temperature typical for the region by 10 or more degrees owing to the influence of large rivers. Along the terraces of the valley it usually does not exceed 2 - 3°C.

In conclusion it should be noted that large and medium sized rivers are one of the most important factors which affect the temperature regime of the upper soil horizons above the permafrost over large areas under any general temperature conditions.

In the literature (Kudryavtsev, 1954) there are indications that in some cases small streams have the reverse effect of cooling the permafrost under the stream bed. This effect frequently takes place near the sources of streams (the Selemdzhi River basin) or in places where there is intensive meandering over clay loam soil (between the Lena and Amga Rivers) and is explained as follows: the clay loam soils prevent infiltration of water into the adjacent horizons as a result of which summer precipitation flow along the surface without giving off heat to the underlying horizons. The source of these rivers and streams is located in regions of continuous permafrost where there is very little precipitation. The main source of water is from the summer thawing of the soil. Thus the temperature of the water at the source of the stream is close to 0°C and the temperature of the water increases only substantially further downstream. The flow of water thus has the effect of removing the summer heat from the horizon under the river bed and transports it downstream. Further downstream this effect is not observed and there is a general warming effect due to the river flow.

Lakes. The temperature regime of the bottom deposits of fresh-water, land-locked lakes depends on their depth. When the water is shallow the mean annual temperature in the bottom deposits may be either above or below freezing. In deep lakes the temperature is above the freezing-point and the soil is unfrozen, forming taliks which either partially or completely penetrate the permafrost.

To indicate the characteristics of the temperature regime of water and bottom deposits in deep, fresh-water lakes we cite the following two examples (Shvetsov, 1951_2): (1) Lake Pereval'noe (63°7'N, 139°82'E) located on the divide between the Kebyume and Khandyga Rivers at an altitude of 1418 metres; the width of the lake is 400 metres, the length 800 metres and the depth 17.9 metres; (2) Lake Mezhdurechnoe (63°21'N, 141°08'E) located between the Kebyume and Suntara Rivers at an altitude 1,060 metres; the width of the lake is 2,000 metres, the length 2,000 metres and the depth 12.85 metres. The results of measuring the temperature of the lake waters carried out in winter are given in Tables XXVI and XXVII. The region is characterized by a low mean annual temperature of the soil (to -10°C).

From the tables it can be seen that the temperature of the water adjacent to the bottom increases with depth and a temperature of 4.0°C occurs only at depths above 17 metres.

When the width of the lake is much less than the thickness of the permafrost it appears that a pseudo-talik is formed under the lake and underlain by permafrost. In this case the incomplete thawing of the permafrost is explained by lateral flow of heat into the surrounding frozen mass. If the width of the lake is greater than the thickness of the permafrost then in all probability a talik will form through the permafrost.

Surface water influences the depth of seasonal freezing and thawing of bottom deposits affecting the heat exchange at their surface which influences the mean annual temperature and amplitude. In the first approximation this variation can be presented in the form of the diagrams shown in Fig. 40. Seasonal freezing and thawing of bottom deposits can take place only when the depth of the lake H is less than the depth of seasonal freezing of open lakes h_f. Since under the most severe conditions this does not exceed 2 - 2.5 m, this situation can only occur in shallow lakes less than 2.5 m deep and usually much less. At these depths the temperature of the water in the summer can be considered uniform throughout its entire depth. Thus t_{max} - the maximum temperature of the water during the summer in this diagram is represented by a vertical line. The minimum temperature at the surface of the ice is determined by climatic data and is equal to t_{min}. At the lower surface of the ice this temperature is 0°C.

For simplicity let us assume that the minimum annual temperature in the ice varies according to the linear law from t_{min} at the upper surface of the ice to 0°C at the lower surface. Then at each point at depth h in the ice the mean annual temperature t_m will be

$$t_m = \frac{\frac{h_f - h_1}{h_f} t_{min} + t_{max}}{2}. \tag{8.14}$$

From the diagram (Fig. 40) and the formula it can be seen that an increase in the depth of the lake results in an increase in the mean annual temperature and a decrease in amplitude in the bottom deposits.

Let us suppose that at a certain depth h_1 the mean annual temperature will be 0°C. When the depth of the lake is less than h_1 the mean annual temperature of the bottom deposits will be below freezing and consequently in this case there will be a seasonally thawing layer underlain by permafrost. When the depth of the lake is greater than h_1 there will be a seasonally freezing layer underlain by unfrozen soil. The value of h_1 from formula 8.14 is:

$$h_1 = h_f\left(1 + \frac{t_{max}}{t_{min}}\right).$$

As the depth of the lake increases there will be a rapid decrease in the depth of seasonal freezing of the bottom deposits which is explained by an increase in the mean annual temperature in the deposits and a decrease in temperature amplitude.

In a region where permafrost prevails when there is a gradual increase in the depth of the lake, the seasonally thawing layer of bottom deposits will at first increase and then, having reached a maximum at a mean annual temperature of 0°C, will convert to a seasonally freezing layer, the thickness of which will gradually decrease. If the lake dries out gradually a reverse process will take place, i.e. there will be an increase in the depth of the seasonally freezing layer of bottom deposits with a subsequent conversion to a seasonally thawing layer at a mean annual temperature of 0°C. The thickness of this layer of seasonal thawing will decrease to the value of the depth of thawing typical of the soil in the given region.

The mechanism considered above is valid for fresh-water lakes but is not always extended to saline lakes. It has been shown that in saline lakes the bottom deposits may freeze even when the depth of the lake exceeds the depth of freezing of open lakes (Dzens-Litovskii, 1938). Investigations have shown that in winter and summer at the bottom of saline lakes the temperature of the brine is below 0°C. This is explained by the fact that in saline lakes the maximum density of the brine occurs at temperatures of -15, -20°C. Thus for saline lakes convection is excluded and heat exchange occurs basically through heat conduction. For such lakes in the summer a temperature of -5°C has been recorded for the bottom layer and in winter -20°C. However saline lakes may have a warming effect if over the brine there is a layer of fresh water or less saline water acting as a heat reservoir.

The temperature regime of the water adjacent to the bottom depends also on variations in the hydrochemical regime of the lake owing to salt accumulation. If the concentration of salt is in the process of increasing, the temperature of the water at the bottom of the lake will decrease and under it a layer of permafrost will form. On the other hand, if the concentration of salt in a lake is decreasing the temperature may increase and the permafrost at the bottom of the lake will be in the process of thawing. Consequently, the temperature regime of the deposits at the bottom of saline lakes depends not only on the concentration of mineral matter but also on whether the concentration is increasing or decreasing.

Seashores. The particular features of the temperature regime of permafrost along the shores of northern seas is governed by the interaction between ground water and saline sea-water.

Sea-water whose freezing point is below 0°C remains liquid in winter and at temperatures below 0°C may circulate along the fissures of permafrost formations. During the summer when the temperature of the sea-water is above 0°C its penetration into a permafrost formation may bring in heat. Directly along the shoreline permafrost is frequently absent although the sea freezes at the shore (Ponomarev, 1953_3). The presence of water-bearing horizons having contact with the sea and the circulation of water within the permafrost account for a convective heat flux which has an important influence on the temperature field of the permafrost. Moreover the dynamics of the saline state of the water, the periodic precipitation of salt and redissolving, accompanied by the liberation and consumption of heat also affect the temperature regime of the upper horizons of the sea bottom. As a result the deviation of the temperatures of the upper permafrost horizon from their usual values for the given region may be substantial.

Ground water. An important part is played by the thermal interaction between the permafrost and water-bearing strata or massifs. The temperature regime of the upper horizons of permafrost is greatly influenced by ground water and artesian water, particularly thermal water. Rapidly circulating ground water increases the temperature of the permafrost and frequently destroys it completely.

As an example we give two cross-sections through river valleys where there are taliks (Fig. 41) within the range of water circulation. The warming influence of the flow of water is observed on steep slopes consisting of coarse residual stones and talus and also on bald peaks and water divides of fold-mountain regions in many areas (Transbaikal, Aldan, mountains of southern Siberia and the Far East).

In these areas the process increases the temperature of the ground by 3 - 5°C and in northern regions by 10°C and more.

Of greater importance in increasing the temperature of permafrost is artesian water, particularly thermal artesian water. For example, in the region of Khal'mer-Yu (Vorkuta) a cross-section of the permafrost has been established as shown in Fig. 42. A number of springs have been found in the talik with water temperatures in winter of 4 - 6°C. Approximately the same has been observed in the basin of the Indigirka River (Shvetsov, 1951_2), where permanent taliks are connected with the outflow of water from below the permafrost under conditions of extremely severe climate, where the mean annual temperature of the soil is about -8°C and the thickness of the permafrost reaches 300 - 400 metres. The temperature of the water of these springs does not go below 6 - 8°C in winter.

These examples graphically demonstrate that ground water is a decisive factor in determining the temperature regime of permafrost and in altering it over a wide range. One of the basic problems in permafrost research is the study of the overall geological structure and the hydrogeological features of a region.

Figures 43 and 44 showing soil profiles in the Chernovskoe coal region and the Khoronorskaya depression illustrate how artesian water in anticlinic and synclinic folds determine the thickness and temperature regime of the permafrost.

Ground ice. In a number of cases the particular conditions of the formation of the temperature regime of the soil and the upper horizons of permafrost are governed by the presence of ground ice. When the ground ice lies deep it has little effect on the formation of the temperature regime of the upper horizons of the permafrost. However, if the ground ice is within the range of seasonal thawing it may melt and begin a thermokarst process resulting in a negative form of relief which subsequently fills with water. As the thermokarst develops, the depth of the water body increases and it in turn results in a gradual increase in temperature at the bottom of the depression. On the termination of the thermokarst process the water body dries up and in its place permafrost is again formed. In regions where there is extensive ground ice lying near the surface of the soil one frequently observes a full range of transitional temperature regimes of the soil, as for example in the Far East (Kudryavtsev, 1939), central Yakutia (Solov'ev, 1948), etc.

Geochemical processes. In northeastern Siberia there are regions where the increase in the temperature of permafrost is explained apparently by geochemical processes, namely an extensively developed process of pyrite oxidation. A unique geochemical oxidation of the ore body has been established for the Ege-Khai sulphide deposits.

The usual mechanism of the oxidation of sulphide ore and the formation of the so-called "iron cap" changes somewhat under permafrost conditions. In this case there is the formation chiefly of sulphate compounds with a large liberation of heat. As a result the upper horizons of the permafrost have a mean annual temperature of -6°C when the surrounding permafrost has a temperature of -10°C.

In regions where the formations have a high concentration of radioactive elements one can probably also expect a certain increase in the mean annual temperature of the soil due to the flux of heat from below. It is also possible to have a local increase in temperature due to spontaneous combustion of coal in regions of coal deposits. The effect of volcanoes on the formation of the temperature regime of permafrost has not been elucidated in the literature but one can confidently say that the large flux of heat from the lower reaches of the earth's crust can bring about a substantial increase in the temperature of the upper horizons. An example of this is the nature of the distribution of permafrost on Kamchatka.

The effect of industrial activity. Industrial activity has a very important effect on the temperature regime of the soil and underlying horizons. In the process of industrial development of a region there are important changes in the physico-geographical and geological conditions of heat exchange between the earth's crust and the atmosphere.

In cities and large settlements the snow cover is compacted or frequently removed from large areas. The building up of an area results in a redistribution of the snow cover. In some places large snow drifts are formed measuring up to several metres deep and in other places the snow is almost completely blown off. Of great importance is the density of the snow, the variation of which can result in a change in the mean annual temperature of permafrost by

0.5 - 1.0°C. When the snow becomes blackened by dust particles there is a substantial increase in the absorption of solar radiation which results in rapid thawing in the spring. As a result the snow cover even when it is deep (1 - 1.5 m) melts very rapidly and exposes the soil to solar heat. This effect is particularly noticeable in the region of Vorkuta where the snow becomes covered by a layer of coal dust. In some populated areas there is sufficient accumulation of waste products on top of the snow that the snow is conserved and sometimes lasts throughout the summer. In such cases one observes a decrease in the temperature of the upper horizons of the permafrost.

In the process of industrial development, great changes take place in the nature of vegetation cover. The removal of forests and afforestation, ploughing and the destruction of vegetation cover, the sowing of crops and hay cutting all have some effect on the formation of the temperature regime of the soil.

In large settlements and towns a similar role is played by the so-called "cultural layer". As an example of a rapid change in the temperature regime of the permafrost under the influence of a "cultural layer", one can cite the city of Igarka (Meister, 1948) and Yakutsk (Mel'nikov, 1951). Table XXVIII shows that the decaying of sawdust in the region of Igarka not only raises the temperature of the sawdust to 80°C but also raises the temperature of the underlying ground to 15°C.

A year and a half after the temperatures were measured the process of decay resulted in spontaneous combustion of the sawdust pile. A similar thing took place at Vorkuta in coal piles. Repeated measurements in coal dumps indicated temperatures close to those mentioned above.

In some settlements there is also increased salinity of the soil due to industrial wastes and manure piles. As an example one can give the city of Yakutsk (Mel'nikov, 1951) where at the local leather works the effluence of highly mineralized water resulted in a considerable disturbance of the permafrost regime. In some places there was an increase in the cooling of the soil resulting in lower temperatures and in other places there was thawing of the permafrost. The temperature increased several degrees and where there was a decrease it reached 1 - 1.5°C.

A considerable change in the temperature regime takes place after the construction of dams and reservoirs. The building of dams and ponds, where the depth is greater than the depth of seasonal freezing, results in a separation between permafrost and the seasonally freezing layer of soil and results in the formation of taliks extending through the permafrost in southern regions and in pseudo-taliks in the north.

An example is the Skovorodino reservoir where the permafrost table dropped to a depth of 10.5 metres. The reservoir is not deep (up to 4 metres) and a large part of it freezes to the bottom. Large reservoir can result in the formation of pseudo-taliks even in northern regions. Permafrost under sufficiently large reservoirs can thaw under any temperature conditions of the air. The difference between the temperature of the soil under a reservoir and of that surrounding it may reach 10 - 15°C in the north and 2 - 3°C in the south.

Under heavy snow cover and water saturation of the soil or flooding there is an increase in the temperature of the upper soil horizons, whereas draining of the water results in a decrease in the temperature. When the snow cover is light, on the other hand, draining results in an increase in the temperature. Irrigation or drainage for agricultural purposes has less effect and the changes in the temperature due to them does not exceed 1 - 2°C.

The tendency of local saturation of the soil and flooding of the seasonally thawing horizons is usually observed in places where ground water is present. The tapping of springs and the installation of pipes for purposes of water supply have an important influence on the hydrogeological conditions of the surrounding territory and correspondingly results in a change of soil temperature. The tapping of springs may lead not only to an increase in soil temperature and thawing of permafrost but may also result in a decrease in temperature and the freezing of taliks due to draining of the soil (Vel'mina, 1952). The greatest effect of changing the temperature regime of the soil is produced by the tapping of thermal springs..

The activity of man has a very important influence on the landscape where ground ice is present. In central Yakutia (Solov'ev, 1948) the ploughing of virgin land, cutting of forests, hay cutting and other disturbances of natural conditions in areas where ground ice is near the surface, results in melting of the ice and the formation of thermokarst lakes.

Forest fires accelerate the melting of fossil ice. In the past during times of drought the local population frequently set fires to melt the ice thus providing water for the soil to improve their crops (Solov'ev, 1948). It has since been established that the destruction of ground ice results in excessive drying of the soil and is generally harmful to agriculture.

Considerable changes in the temperature regime of the upper horizons of the permafrost takes place also under structures in which large quantities of heat are produced, for example, the heat-producing industrial plants such as blast furnaces, stoke-holes, electric-power generating stations, etc. In such cases the permafrost thaws even under the most severe climatic conditions. As examples one can cite the old brick factory in Vorkuta, the glass factory in

Yakutia (Mel'nikov, 1951), the bakery in the settlement of Anadyr' (Kachurin, 1938_1), etc. The changes in the temperature of the subsoil in these cases reaches several tens of degrees.

When the permafrost thaws at the base of structures there may be deformation and even destruction of the structures. The most important condition for the stability of a structure is the conservation of the foundation soils in their initial state. In construction, using the principle of conservation of permafrost, one must prevent the flow of heat into the ground. This is done by providing a ventilated crawl space. In such cases the temperature decreases and the thickness of the seasonally thawing layer decreases. An example of a structure with a ventilated crawl space is the electric generating station in Yakutsk (Zhukov, 1958). In spite of great changes in the external conditions one can control the temperature regime of the permafrost, either conserving it or thawing it as required.

The influence of man on the formation of the temperature regime of the upper horizons of permafrost is increasing continually although in the Far North it is not yet large and is limited to small areas. Populated places in the permafrost region are subject to specific conditions in the formation of the temperature regime of the upper horizons of the permafrost and therefore they must be considered separately.

Thickness of the permafrost. The above-mentioned physico-geographical and geological conditions of heat exchange between the earth's crust and the atmosphere determine the depth of seasonal freezing and thawing of the soil, the temperature at the level where annual temperature variations are attenuated and consequently the thickness of the permafrost. It should be noted that there is no direct relationship between the thickness of the permafrost and the temperature at a depth of 10 - 15 metres, i.e. the level at which annual temperature variations are attenuated, since the thickness of the permafrost is determined not only by the temperature within the permafrost but also by the composition, structure and thermo-physical properties of the frozen strata and the underlying strata.

Monolithic crystalline compact microporous soils under otherwise equal conditions freeze to a greater depth than macroporous water-bearing soils. The difference in the depth of freezing of such lithologically different types of soils may reach tens and even hundreds of metres; for example, at the same soil temperature of -10°C at a depth of 10 - 15 metres, the thickness of the permafrost due to differences in composition may vary from 300 - 600 metres. Therefore one cannot compute the thickness of permafrost on the basis of the mean geothermal gradient alone without taking into account the composition of the permafrost and of the underlying unfrozen formations. The value of the geothermal gradient within the permafrost when there is a constant flow of heat is determined by its composition and may vary with the composition. For this reason one cannot characterize the development of permafrost with respect to latitudinal zones solely from the depth of the lower surface of the permafrost without taking into account the composition.

The lithological composition is one of the determining factors in the formation of the thickness of the permafrost. Moreover, as we will see below, the composition of the soil is of no less importance in the formation of boundary conditions at the permafrost table and at the lower surface of the permafrost. Soil composition must also be taken into account in considering the propagation of heat from heat sources located within the permafrost.

The mean annual temperature of the upper soil horizon, as already mentioned, is one of the basic conditions determining the thickness of the permafrost. The mean annual temperature of the earth's surface is itself formed by the influence of a number of physico-geographical factors according to the mechanisms considered above. In areas where the mean annual temperature of the upper soil horizon is high, the thickness of the permafrost must be, under otherwise equal conditions, less than in areas where the temperature is low.

By virtue of this factor, in western Siberia the thickness of the permafrost in the tundra and forest zones are different (Popov, 1953_1). In the tundra zone the temperature is approximately 1°C lower than in the forest zone and for this reason the thickness of the permafrost in the tundra zone is 30 - 60 metres greater than in the forest zone.

Of the surface boundary conditions surface water has the greatest influence on the thickness of permafrost.

If the depth of a lake is greater than the depth of its seasonal freezing the temperature of the bottom deposits remains above freezing throughout the year. Because of this permafrost can exist under a lake only because of lateral cooling. This can take place if the minimum width of the lake is not greater than the thickness of the permafrost at the shores of the lake. If the width of the lake is greater than the depth of the permafrost then a talik will form under the lake. However, in all cases where a lake or stream is formed it is associated with a decrease in the thickness of the permafrost underneath it.

In areas or belts with migrating lakes or rivers there is a very wide variation in the thickness of the permafrost and in the rates of change in thickness. Figure 45 shows a repeated displacement of the river bed in opposite directions. Such a migration of a river results in the formation of a laminar structure consisting of frozen and unfrozen deposits. In the diagram the Roman numerals and horizontal hatching denote newly formed perennially frozen strata in the place from which the river bed has receded. The diagonal hatching illustrates taliks formed as a result of the river bed advancing over areas that were previously frozen.

The lamination of frozen and unfrozen deposits is observed in the valley of the Lena, Ob' and other rivers. As an example we give a cross-section through the valley of the Ob' River in the vicinity of Salekhard (Fig. 46).

The thickness of the permafrost may vary substantially depending on the presence and nature of ground water currents which frequently carry large quantities of heat and have a warming effect on the permafrost. The warming effect of ground water currents has much in common with the effect of surface water bodies (see Fig. 40). On the appearance of ground water currents or basins of ground water the thickness of the permafrost decreases and taliks may form which penetrate either partially or completely through the permafrost. A decrease in the thickness of the permafrost may occur in the absence of taliks resulting from the single factor of an increase in the mean annual temperature of the upper permafrost horizons. The greatest influence of ground water on the thickness of permafrost is observed in river valleys and slopes of water divides.

In considering the boundary conditions at the permafrost table one must take into account the interaction between the temperature regimes of the atmosphere, the soil and the lithosphere. Variations in the temperature regime of the atmosphere adjacent to the earth and of the soil affect the temperature regime of the lithosphere. It is generally known that thermal waves penetrate into the earth's crust to different depths, depending on the length of the period and amplitude of the temperature oscillation at the surface. If the thermal wave of oscillations lasting for several years are attenuated before they reach the lower surface of the permafrost (Fig. 47) there is a change in the temperature and temperature gradient in the permafrost within the range of penetration of these waves. Usually the sign of the temperature gradient remains constant, although at relatively large amplitudes and at temperatures close to 0°C the temperature gradient may change its sign. This was noted by M.I. Sumgin. Temperature curves for permafrost having a minimum temperature beyond the range of the stratum with seasonal temperature variations were called degradational by M.I. Sumgin. Thermal wave oscillations which do not reach the lower surface of the permafrost do not have any effect on changes in the thickness of the permafrost.

Heat oscillations penetrating beyond the lower surfaces of the permafrost are attenuated at this surface and do not penetrate into the underlying strata. This is explained by the thermal effect of phase conversions which occur at the lower surface of the permafrost.

The distribution of temperatures in a permafrost stratum when there are short period and long period oscillations should be different. Figure 47 illustrates the former case and Fig. 48 the latter. In the latter case the temperature curves are straightened out as a result of phase transition and the lower surface of the permafrost will periodically rise and fall within the range of depth from Z' to Z". The depth of Z would be a mean value.

As a result of thawing and freezing at the lower surface of the permafrost there will be an inflection in the temperature curve; the sign of the angle of the inflection will indicate the direction in which the lower surface is being displaced.

The greater the ice content of the soil, the greater will be the inflection of the temperature curve connected with the heat of ice melting. The magnitude of inflection depends also on the rate of change in the temperature field resulting from thermal wave oscillations at the surface of the earth; as a result this factor should be more evident with short period oscillations than with long period oscillations.

It has been established that the maximum ice content of the soil occurs in the upper horizons of permafrost to a depth of 15 - 25 metres (Pchelintsev, 1948_2). Thus a high content at the lower surface of the permafrost is observed mainly near the southern boundary of the permafrost zone. Here one should observe the greatest inflection in the temperature curve. Friable deposits have a higher ice content and inflection of the temperature curve will be greater in them than in rock that is not highly fissured.

The maximum angle of inflection is formed when there is a zero temperature gradient at the lower horizons of the permafrost. It occurs when there are relatively rapid changes in the temperature field, flow rates of displacement of the lower surface of the permafrost and in the presence of phase transitions of the water within the permafrost at below freezing temperatures.

Thus changes in conditions of heat exchange at the surface of the earth due to lakes, rivers and ground water brings with it changes in the thickness of the permafrost, partially determines the temperature field and inflection of the temperature curve at the lower surface of the permafrost. Similar changes in the temperature field and thickness of the permafrost may be caused by changes in other boundary conditions.

In considering the mechanisms of the formation of the mean annual temperature of the soil it has already been mentioned that ice cover has an important influence on these mechanisms. It can exclude the possibility of the formation and existence of permafrost under the ice or it may result in the formation of the perennially frozen stratum where it had not existed before the ice formation. Therefore one cannot always consider ice cover as excluding the formation and existence of permafrost; under specific conditions, for example, where the ice is not thick, near the southern boundary of the permafrost region, ice formation must be regarded as a phenomenon facilitating the formation of a perennially frozen stratum.

Other processes also lead to a change in the temperature field of a perennially frozen stratum and to changes in its thickness. For example, sedimentation, denudation, epeirogenic oscillations, displacement of geobotanical zones, etc.; changes in the conditions of heat exchange between the lithosphere and the soil and atmosphere bring about changes in the temperature field and thickness of the permafrost.

The position of the lower surface of the permafrost is determined by the value of the geothermal gradient and hydrological conditions in addition to the above-mentioned causes. Above, it has already been noted that the inflection of the temperature curve at the lower surface of the permafrost and the position of this boundary are determined by the relationship between heat flow to the lower surface of the permafrost from the earth's interior and the flow of heat from it to the surface of the earth. The thickness of the permafrost, the formation of which is connected with long-term oscillations (tens and hundreds of thousands of years), varies in extent several fold depending on the magnitude of the geothermal gradient. This will become evident if one makes even a rough approximate calculation. During a period of 100,000 years at a temperature gradient of 1°C per 100 metres approximately 1,800 kcal per 1 cm² will reach the lower surface. When the gradient is 1°C per 5 metres this number increases to 36,000 kcal. When the thickness of the permafrost is not large, near its southern limit where short-period oscillations are decisive, this difference will not be important. However, where the permafrost is thick and its existence is measured over long periods this difference may have an important influence on its thermal regime.

Thus the value of the geothermal gradient will vary in importance in the formation of the thickness of the permafrost depending on the duration of its existence.

The situation is somewhat different with the hydrogeological conditions. Ground water, particularly artesian water under the permafrost, carries large quantities of heat and therefore its warming influence at the lower surface of the permafrost is always large. When artesian water is in direct contact with the lower surface of the permafrost its warming influence reaches maximum values. If it is separated from the permafrost by a water-impermeable horizon the influence is reduced.

The supply and circulation of ground water are important factors. If the ground water is rising it transports heat from the lower reaches to the frozen stratum and the warming effect increases with the depth from which the water rises. In this case the rate of circulation, the filtration properties and nature of the hydrogeological structure are important. The greater the rate of ground-water circulation the greater is the warming effect. In hydrogeological structures in which there is no ground water movement the warming effect is insignificant. The greatest warming effect is produced by thermal water circulating along tectonic fissures, faults and fractures. The waters of variegated-laminated and fissured-laminated structures have a smaller warming effect.

The chemical composition of ground water also has an important influence on the position of the lower surface of the permafrost. When the water has a high mineral concentration which reduces the freezing temperature the lower boundary of the permafrost does not coincide with the 0°C isotherm but lies somewhat above it. If the mineral concentration is low the difference may be expressed in tenths of a degree and consequently ten or twenty metres in the thickness of the permafrost. If the mineral concentration is high this difference may be increased to several degrees which is accompanied by a decrease in thickness of the permafrost by one hundred to two hundred metres.

Thus an increase in the mineral concentration of ground water results in a decrease in thickness of the permafrost without disturbing its temperature regime. However, if one takes into consideration the precipitation of salts and their subsequent redissolving accompanied by the liberation and absorption of heat one must also consider substantial changes in the temperature field. In general it must be considered that ground water plays an important part in determining the position of the lower surface of the permafrost.

An important influence on the thickness of permafrost are internal sources of heat. This problem has received very little attention and therefore in considering it one can only express some general considerations.

Geochemical processes within the permafrost can be exothermic or endothermic and thus their influence may be to decrease or to increase the thickness of the permafrost. The important factor is the material composition of the permafrost itself.

In concluding this section we give as an illustration some data on the thickness of permafrost for a number of regions (Table XXIX).

From the table it can be seen that the thickness of the permafrost varies within a range of several metres to 600 - 800 metres and these latter values are apparently not the maximum. In high mountainous regions where the mean annual temperature of the upper horizons of the monolithic rock is below -15°C the probable thickness of the permafrost exceeds 1,000 metres.

Characteristics of the Permafrost Depending on Changes in Boundary Conditions at the Permafrost Table and at the Lower Surface

The boundary conditions at the permafrost table and at the lower level are determined by a number of geological and geographic conditions and processes. By the nature of their influence on the temperature of the soil, geological and physico-geographical processes and conditions can be divided into three groups:

(1) those leading to periodic variations in the temperature regime of the soil;

(2) those leading to changes in the temperature regime of the soil in one direction;

(3) those leading to single spasmodic rapid changes in the temperature regime of the soil.

From the point of view of thermal physics it is very important in all cases to know the position of the upper boundary of the semispace (soil) through which heat exchange takes place between the lithosphere and the atmosphere. Here also one must differentiate between three cases: a constant position of the surface, its elevation due to sedimentation and its drop resulting from denudation.

It is customary to differentiate between periodic, daily, annual and perennial changes in the temperature regime of the soil. Daily and annual variations are close to being regular sinusoidal; perennial variations frequently do not coincide with the regular sinusoidal, their periodicity is not regular but for all of these oscillations a change in the sign of the temperature increment with respect to time is characteristic, i.e. as is customary to say, periods of temperature increase are displaced by periods of temperature decrease. Daily and annual oscillations are observed directly in nature. Perennial short-period oscillations within the range of 150 to 200 years are fixed by meteorological observations. Long-period oscillations lasting several thousands, tens of thousands or hundreds of thousands of years, are determined from geological data. Oscillations of even longer period are determined for various geological periods lasting tens, hundreds of millions and even billions of years.

Periodic oscillation should not be associated solely with astronomical causes. To a large extent they are determined by processes occurring on the earth. Thus periodic changes in the temperature regime of the soil is caused by displacement of the axis of rotation of the earth, changes in sea and air currents, changes in the ratio of area occupied by land and by sea, the configuration of continents, marine transgressions and regressions, mountain-forming processes, processes of denudation and sedimentation, etc.

Each of these factors results in a change in the condition of heat exchange at the surface of the earth and oscillations in the temperature of the soil and underlying horizons. The complexity of heat exchange and the combination of similar and opposite influences to some extent is reflected in the strange peak-shaped curve of variations in the temperature regime of the air and soil with respect to time. At the present time one cannot isolate the meaning of each of the component factors contributing to heat exchange and the formation of the temperature regime of the soil. However, this statement of the problem makes it possible to draw some conclusions regarding the development of permafrost which are important for permafrost scientists.

As is known, M.I. Sumgin in 1931 suggested the theory of "degradation of permafrost". According to this theory the permafrost was regarded as continuously changing and not established once and for all (Sumgin, 1937). M.I. Sumgin considered that the existence of permafrost is in disharmony with present-day climatic conditions and consequently with present-day heat exchange at the surface of the soil and is therefore to some extent a relic of the Ice Age. At the present time this problem must be given a much wider consideration (Kudryavtsev, 1954). An increase in temperature and thawing of the permafrost (degradation) must be regarded as applicable to a specific period of time. The present-day state of the permafrost as compared with the Ice Age is characterized by an increase in reserves of heat but when compared with the post-glacial thermal maximum there is a decrease in reserves of heat. Moreover in the same place and at the same time, in different horizons of the permafrost, one can observe an increase as well as a decrease in the temperature of the lithosphere and consequently degradation and aggregation of permafrost. The combination of these two opposite trends represents the general process of the development of permafrost. Changes in boundary conditions at the permafrost table is also the result of a complex combination of processes of warming and cooling with respect to time. The various oscillations in temperature have different rates of propagation in the upper strata of the lithosphere and therefore their separation is of great importance for the study of permafrost. The determination of the rates of propagation of various oscillations within the soil in conjunction with a study of the present-day temperature field of the permafrost, its composition and thickness makes it possible to estimate its development.

In analyzing the boundary thermal conditions at the permafrost table one must determine the duration of changes in temperature, the magnitude of the change in time and its initial value. All of these data have frequently been obtained with some degree of accuracy by palaeogeographic methods. Trends and temperature changes of the soil going in only one direction, i.e. only in the direction of warming or only in the direction of cooling are unlikely to take place. Any long-period oscillation with respect to short-period oscillations is always a trend in one direction. In view of this one can consider any trend in a change of the soil temperature regime in one direction as being a long-term oscillation. For periodic oscillations, as well as for trends in one direction, an important factor is the continuity of the change in boundary conditions at the permafrost table. At every subsequent moment of time commensurate with the period of change in boundary conditions the temperature of the soil will differ from the previous moment since the duration of various changes in boundary conditions varies from a day to hundreds of thousands and even millions of years. Hence it follows that one cannot speak of any steady state in the temperature field of the permafrost, its composition or its thickness.

For example, when a lake dries out or a new river bed is formed during spring flood the temperature regime at the surface of the bottom deposits undergoes a rapid change along with which there should be a change in the temperature field of the bottom deposits. It would seem that after a certain period of time one should expect the establishment of a new steady state temperature regime in the bottom deposits corresponding to the new conditions at the surface. In fact one can only speak of the advent of a new dynamic equilibrium state corresponding to a new boundary condition.

A diagram of changes in the temperature field and thickness of the permafrost in this case will be somewhat different from that of periodic oscillations at the surface of the soil.

In considering the temperature of formations at the lower surface of the permafrost one should differentiate between conditions determining conductive and convective heat exchange at this surface. Conductive heat exchange is determined by the composition of the formation underlying the permafrost and by the geothermal gradient. Convective heat exchange is determined by hydrogeological conditions (particularly by the nature of the circulation of ground water) and also by vapour migration, its condensation and evaporation of water under the permafrost. Boundary temperature conditions at the lower surface of the permafrost are less dynamic than at the permafrost table; variations of the former are to a large extent due to the effect of the latter. The extent of this influence depends on the thickness of the permafrost since changes in the temperature field of the lithosphere are determined by the period of temperature oscillation and amplitude.

Thus boundary temperature conditions at the lower surface of the permafrost are not in a steady state. The absolute value of the change in temperature of the ground at the lower surface of the permafrost are naturally much lower than at the permafrost table.

In the formation of permafrost much importance is attached to the composition and occurrence of exothermic and endothermic processes. Particular attention should be drawn to the ice content, the result of which is mainly a change in heat conductivity of the permafrost due to differences in ice content.

With the temperature field in a steady state at the permafrost table and at the lower surface of the permafrost, the heat of formation and melting of ice do not have any influence on the thickness of the permafrost, but when the conditions at the permafrost table and at the lower surface are dynamic the value of the heat of ice formation increase rapidly. To a large extent they determine the temperature field and thickness of the permafrost. Hence the assumption of steady state in the boundary conditions leads to a completely erroneous conclusion that the composition, texture and structure of permafrost is not directly connected with the temperature field and thickness and that the influence of these factors is reduced only to the establishment of a temperature gradient. In fact the structure, texture and composition of permafrost is so closely connected with its temperature field and thickness that there is some basis for considering that in the near future we will be able to study the history of the development of permafrost by the ice content. In connection with the dynamic nature of boundary conditions and the temperature field of the permafrost, its composition and particularly its ice content should also change continuously.

The physico-chemical processes taking place within permafrost is also of some importance. Oxidation and reduction of various minerals, dispersion and coagulation of colloidal particles of soil, the dissolving and precipitation of salts, etc., have some effect on the formation of the temperature field in permafrost.

An important conclusion can be derived from the above discussions. It is that there is a continuous change going on within the permafrost itself, in its temperature regime and thickness. The mineral composition and the lower and upper boundaries conditions of the permafrost determine the depth of freezing, the temperature field at both zones (frozen and unfrozen) and the heat flow through any cross-section. The rate of change in boundary conditions determines also the rate of change of these characteristics of the permafrost.

In studying the laws governing the formation and the history of the development of permafrost one must establish correctly the changes in the geological and geographic conditions of the past. Ignoring these conditions can lead to erroneous conclusions. As an example one can indicate the work of D.V. Redozubov (1946) who considered that the conditions at the surface were in a steady state and then subsequently went through a spasmodic transition to other boundary conditions. The author determined the time of the establishment of the new steady state temperature field of the permafrost and its thickness and arrived at the conclusion that the new steady state is established in the permafrost very rapidly.

The development of permafrost is regarded by D.V. Redozubov as a disturbance of a state of rest at specific moments of time. Hence he arrives at the incorrect conclusion that the theory of degradation suggested by M.I. Sumgin is valid and the temperature field of the permafrost is in a steady state and is explained only by space distribution of heat flow. The conditions considered above for the freezing of the soil indicates that the simplified conclusions derived by D.V. Redozubov are valid and that his error consists in solving the thermo-physical problem only for steady state geological and geographic conditions, i.e. for a particular case of upper boundary conditions.

As proof we will give some computations of the duration of freezing of the lithosphere under various boundary conditions at the surface. We will consider two cases: the first for a periodic change in the upper boundary conditions, and the second for a steady state in conditions at the surface with a spasmodic establishment of new steady state conditions.

The computations were carried out with the IG-1 hydrointegrator for oscillations with a period of a hundred thousand years and an amplitude of 6°C with a mean temperature at the surface of the soil for the entire period of the temperature oscillation of 0°C*. The results of the calculations are shown in Table XXX and Fig. 49. The data in the table show that for the assumed boundary conditions and thermo-physical characteristics of the soil 33,000 years are required for complete freezing of a stratum 190 metres thick.

The thickness of the permafrost increases if the length of the period or the amplitude of temperature oscillations of the surface are increased. The time of freezing is determined only by the length of the period and duration of the existence of variable temperatures below freezing at the surface of the earth. With large periods of oscillation in temperature the time required to freeze the lithosphere will be correspondingly greater.

Depending on the composition and particularly the moisture content, i.e. the ice content of the soil, the thickness of the permafrost may vary under the same boundary conditions by a factor of 1.5 - 2. The same effect can be produced by different amplitudes. Hence it follows that the freezing of a stratum 100 metres thick with a periodic change in the upper boundary conditions may in some cases last longer than the freezing of a stratum 200 metres thick, i.e. permafrost strata of less thickness may be older than those of greater thickness.

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* Calculations carried out by L.N. Kritsuk, V.G. Melamed and M.D. Golovko in 1956 at the Permafrost Section, Department of Geology, Moscow State Univ.

When there are spasmodic changes of boundary conditions (rapid oscillations in temperature at the surface), the rate of freezing of the soil will differ greatly. In such a case the freezing of a stratum 196 metres thick can occur in 5,000 years rather than 33,000. This example graphically illustrates the importance of correctly selecting the boundary conditions and the rates of variation of these conditions for the historical period of time under consideration.

The composition of the permafrost will also be different for the two cases cited above: in the case of periodic oscillations the heat flow attenuates with depth and at the surface of the earth it, as is known (Obolenskii, 1944), is determined by the relationship

$$Q = \frac{A}{2}\sqrt{\frac{C\lambda\rho T}{\pi}}, \tag{8.15}$$

where Q - heat flow through the surface of the earth propagating into the permafrost;

C - heat capacity of the soil;

λ - heat conductivity of the soil;

ρ - density of the soil;

T - period of temperature oscillation at the surface of the earth;

A - amplitude of temperature oscillation at the surface of the earth.

Without the freezing of the soil, heat cycles decrease with depth according to the law of geometric progression. Phase transition of the water greatly accelerate the attenuation of the heat cycles with depth. Hence one can derive one very important conclusion concerning the structure of permafrost: with temperature oscillations at the surface of the earth the ice content of the epigenetic permafrost should decrease with depth which has in fact been observed by a number of authors (Pchelintsev, 1948_2, Bakulin, 1958). This is also connected with the quantitative estimation of heat cycles in the permafrost. Heat cycles increase in direct proportion to the square root of the length of the period. Frozen strata formed during spasmodic changes in temperature conditions at the surface will not have a regular variation in ice content with depth. The thickness of the ice lamina will be less owing to the rapid rate of freezing.

Thus from the composition of the frozen stratum and by its ice content one can estimate the boundary conditions at which the permafrost was generated and exists. Of great importance in the formation of permafrost is the situation prevailing at the surface of the earth.

With simultaneous accumulation and freezing of deposits a syngenetic type of permafrost is formed. The thickness and temperature regime of the permafrost depends on the rate of sedimentation, the thawing of the permafrost from below and changes in the temperature conditions at the surface. The structure and composition of syngenetic strata differs greatly from epigenetic strata.

Without considering all the complexities of the formation of syngenetic permafrost we note that under otherwise equal conditions sedimentation during a cold period results in greater thicknesses of permafrost, whereas warm periods result in lesser thicknesses. Frequently in the same river valley in places where deposits are accumulating and erosion terraces there are different thicknesses of permafrost because of a difference in lithological composition and structure. During the formation of syngenetic permafrost, hydrogeological conditions and variations in boundary temperature conditions at the surface of the earth are also of importance.

We pass now to a consideration of the temperature field within the permafrost itself. Here also a basic condition is the continuous change in the temperature field. The equations given at the beginning of this chapter indicate that the nature of the temperature field within the permafrost and its variation with respect to time are determined by the composition of the permafrost and the boundary conditions at the lower surface and at the permafrost table. Continuous changes in boundary conditions determine continuous changes in the temperature field of the permafrost. Calculations carried out on the computer "Strela" using the above-mentioned equations* showed that during freezing with periodic changes in temperature at the surface the distribution of temperatures in the permafrost is close to linear. This is particularly evident when the thickness of the permafrost is not large. With depth, the distribution of temperatures becomes curved. When short-period oscillations corresponding to the formation of permafrost are applied to long-period oscillations, the short-period oscillations are attenuated in the permafrost and do not reach its lower surface. In this case periodic changes in the temperature field connected with the short-period oscillations are imposed on the temperature curves that are almost linear and the temperature curves take on a form corresponding to an increase in permafrost and degradation. The rate of change of these curves is determined by the length of the period of short-period oscillations. When the thickness of the permafrost reaches 100 metres the increase and degradation temperature curves may be due to approximately 300-year or more short-term temperature oscillations at the surface, in thicknesses reaching 300 metres - due to 2,500-year and more short-period oscillations and in thicknesses of 800 metres - due to approximately 20,000-year and more short-period oscillations.

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* The computations were carried out by V.G. Melamed at the Permafrost Section, Department of Geology, Moscow State University.

In considering the temperature field of permafrost it is important that this field be linear. Only in this case can one speak of degradational or aggregational temperature fields and establish the history of the development of permafrost from these curves. If the temperature field is not linear the temperature curves may be either aggregational or degradational and do not in fact reflect the course of development of the permafrost.

As an example of the correct interpretation of temperature variations, one can point to the temperature observations in the shaft at the Skovorodino Permafrost Station and at the Shergin Mine in Yakutsk. At the Skovorodino Station observations have established a forty-year oscillation in the temperature of the permafrost to a depth of 28 metres (Kapterev, 1946). In the Shergin Mine, temperature measurements of the permafrost to a depth of 116 - 130 metres have established a 300-year oscillation in temperature (Kudryavtsev, 1954). Temperature measurements in this case were carried on two occasions with an interval of 100 years.

We will deal with one more mechanism in the formation of permafrost. The system of equations given above for determining the depth of freezing of soil shows that in addition to other conditions it is determined by the length of the oscillation period or the duration of below-freezing temperatures at the surface of the earth. During a diurnal period the ground freezes during the night. The duration of the existence of frozen horizons in this case will be determined in hours. Correspondingly the depth of this freezing will be determined in centimetres. With an annual period of oscillation the ground freezes only during the cold period of the year. The depth of seasonal freezing in this case is measured from several centimetres to several metres and the duration of the frozen state varies from several days to half a year. With perennial oscillations the frozen state may last from several years up to hundreds of thousands of years. In the presence of a perennially frozen stratum the upper soil horizons will thaw periodically. In the case of diurnal oscillations at the surface of the earth, when the temperature goes above 0°C, there will be diurnal thawing lasting for several hours to a depth of several centimetres or fractions of a centimetre. With annual oscillations there will be seasonal or summer thawing of the soil. By analogy there is perennial thawing of the soil. In this case the duration of the unfrozen state above the permafrost table will be determined by the duration of the oscillation or the duration of the above-freezing temperatures of the surface of the earth. The depth of perennial thawing will be determined by the composition of the soil and by a number of other factors in addition to time. This type of perennial thawing occurs in the western Siberian lowlands (Salekhard, Narykary, Kondinskoe) and in a number of other regions. By virtue of these considerations and also practical purposes one should consider seasonal and perennial freezing and thawing of the soil separately.

In finishing the examination of general preliminary considerations regarding the formation and development of permafrost and its temperature regime we will deal with the consideration of individual particular mechanisms connected with these problems.

It is generally known that the mean annual temperature of the soil and air over a number of years varies greatly, resulting in changes in the temperature of the upper horizons of the permafrost reaching several degrees.

The mean perennial values of air temperature do not reflect the dynamics of the climate but rather mask it. Therefore the statistical method used extensively in meteorology in studying the dynamics of permafrost requires special deciphering and in meteorology this method is not always used without reservation.

V.N. Obolenskii (1944) indicated that climate is not a constant factor which at one time had been formulated and undergoes only slight changes from time to time. The process of climate formation is going on all the time and each year produces new conditions comprising the perennial regime.

E.S. Rubinshtein (1946) confirms and illustrates this hypothesis. She gives two curves for the moving ten-year averages of annual air temperatures, one (Fig. 50) for a permafrost region (from the data of the Salekhard Meteorological Station for a period from 1881 to 1941) and the other (Fig. 51) for Leningrad for a period from 1801 to 1941.

Both curves illustrate the variability.of mean annual air temperatures. At Salekhard during 55 years of observations there is a clearly evident increase in the mean 10-year air temperatures by almost 2°C. This warming trend follows a complex curve with a sloped axis. A similar mechanism is observed for Leningrad but in this case the axis of the temperature curve has a more gradual slope.

The mean perennial temperature may have highly varied values depending on the section of time for which it is determined. Moreover the use of mean perennial values give an incorrect representation of air temperature for any given period of time and does not give any indication of the trend. The temperature of the upper horizons of the permafrost should also be regarded as characteristic of a specific time period.

Naturally the geoisotherms constructed for each ten-year period may differ somewhat from the geoisotherms plotted on our map (Fig. 52). Changes in areas of the temperature zones, displacement of latitude and altitude boundaries of the permafrost zone correspond to the displacement of the geoisotherms. If one regards the displacement of geoisotherms for a cycle of short-period oscillations it will take place continuously from north to south with respect to some comparatively constant axis, the displacement of which would be governed by long-period oscillations.

If the mean perennial below-freezing temperature is not less than the amplitude of short-period oscillations (Fig. 52), the variations will occur within a range below 0°C, and with a general increase in temperature the permafrost will not separate from the seasonally freezing layer; if it is less, then in the warm semi-period the temperature of the permafrost will pass through the 0° mark and thawing of the permafrost will begin separating it from the seasonally freezing layer. In a limited boundary case when the mean perennial temperature coincides with 0°C or is a little higher, the perennial frozen strata will periodically form and disappear. This is characteristic of the southern boundary of the permafrost zone and reflects its periodic displacement.

As an example we used the meridional cross-section along the line Salekhard-Berezovo-Tyumen'. Let us plot along this line the perennial mean annual air temperatures and the corresponding position of the southern boundary of the permafrost zone (Fig. 53). The difference between the mean annual temperature of the air and of the soil will be determined by physico-geographical and geological conditions. For the region of Berezovo it is 3.6°C.

Changes in air temperature reflect complex and continuous superpositions of a number of harmonic oscillations in climatic conditions with different amplitudes and different periods. Each of these oscillations governs the displacement of the critical temperature of the formation of permafrost and should bring about a displacement of the southern boundary of the permafrost zone.

Let us consider oscillations with a period T. In Fig. 52 a diagram of the displacement of the critical temperature for the formation of permafrost has been constructed to correspond to this oscillation period. The solid line represents the axis of these oscillations, and the dotted line the extreme temperature values. The distance between the lines is 2a, twice the amplitude of the oscillation. The variation in the critical temperature is periodic and passes from point A to point B. Correspondingly the displacement of the southern boundary of the permafrost zone also takes place periodically with the same period and within the same range from A to point B. Hence the rate of displacement of the southern boundary is equal to the distance between A and B divided by the half-period of oscillation $\left(\frac{T}{2}\right)$. It should be considered also that the permafrost has a heat inertia which will result in some delay in the formation of the thawing of permafrost with respect to displacement of the critical temperature.

During the formation and thawing of the permafrost not all characteristics change to the same extent and therefore even with regular sinusoidal oscillations the formation and thawing of permafrost do not always exactly correspond to each other. If one takes into account the fact that temperature oscillations at the surface are far from being regular sinusoidal naturally the permafrost formed during the first semi-period may partially survive until the following cold half-period.

We have given an example of the displacement of the southern boundary of the permafrost zone under the influence of only one type of oscillation with a period T, but there are several types and all of them to some extent result in the displacement of the southern boundary of the permafrost zone. The amount of this displacement will in each specific case be determined by the amplitude and period of temperature oscillation. However, since the amplitude of oscillation with different periods changes within narrow limits (1 - 2°C) and only for many-year periods does it reach several degrees, the amount of displacement of the southern boundary of the permafrost zone will be of the same order for ten-year as for century or many-century oscillations. Correspondingly the rate of displacement of the southern boundary will be almost in inverse proportion to the length of the period. The total effect of all of these long-period oscillations at any given moment will depend on the relation between periods and amplitudes of each type of oscillation and will be equal to their algebraic sum.

In Fig. 52 and 53 it can be seen that the mean annual air temperature varies with latitude approximately by 1°C per 150 - 200 kilometres. For brevity we will call this distance the latitudinal temperature stage; at the southern boundary this stage is approximately 180 kilometres. For the European North of the U.S.S.R. it is 120 kilometres and for eastern Siberia it is 200 kilometres. If one takes the amplitude of short-period oscillations at 2°C the amount of displacement of the southern boundary for the western Siberian lowlands will be 350 kilometres. Hence the rate of displacement of the southern boundary of the permafrost can be readily calculated: for a period of 40 years it will be 17 kilometres for western Siberia, 12 kilometres for the European North of the U.S.S.R. and 20 kilometres for eastern Siberia. For a hundred-year period in these regions it will be approximately 5 and 8 kilometres per year.

As can be seen from these amounts the rate of displacement of the southern boundary of the permafrost zone during short-period oscillations is large and could scarcely go unnoticed.

Changes in the physico-geographic conditions at a given region are accompanied by displacement of the isotherm of critical temperatures and a corresponding deviation in air temperature from the critical which even at values of the order of 1°C in the direction of either cooling or warming produces formation or thawing of permafrost in the corresponding region. Adjacent to the southern boundary of the permafrost region is a zone of residual frozen soil lasting one or more years. Within the limits of this zone the residual frost formations may change to islands of perennially frozen soil or the frost may disappear completely. Therefore the displacement of the southern boundary of the permafrost zone does not have a clearly defined nature and can be determined only after special investigations. The southern boundary cannot be regarded as a line separating the region of permafrost from that of seasonally freezing soil since it is a wide band of 200 - 400 kilometres and possibly greater.

According to the diagram (Fig. 54) of M.I. Sumgin (1937), which is presented by us in a somewhat different form (Fig. 55), the displacement of the southern boundary of the permafrost takes place only from south to north since this displacement is regarded by M.I. Sumgin as the result solely of degradation of permafrost. In fact the displacement of the southern boundary of the permafrost zone does not follow a smooth curve but a more complex curve representing the results of a number of harmonic oscillations of corresponding oscillations in climate having periods from many thousands of years to decades. Taking the diagram of M.I. Sumgin as a basis, the displacement of the southern boundary of the permafrost should be represented in a more complex form (Fig. 56). In comparing this diagram with that of M.I. Sumgin it can be seen that in principle they are the same since in both cases the basic historical trends of the development of permafrost remain the same, degradational, but at the same time these diagrams differ greatly. In Fig. 56 with general degradation of the permafrost the southern boundary may be displaced from south to north or from north to south, whereas in the diagram of M.I. Sumgin the direction of displacement of the southern boundary from north to south must correspond to a general advance of the permafrost.

Thus for the development of permafrost in the direction of degradation the diagram of the displacement of the southern boundary was given correctly by M.I. Sumgin; but not taking into account a number of other oscillations it is insufficiently accurate for practical utilization in studying the dynamics of the permafrost region. In fact we have shown above that the rate of the displacement of the southern boundary during short-period oscillations as compared with long-period oscillations is very large, and therefore the latter are lost and become completely unnoticed in the general overall effect of all types of oscillations.

The basic considerations applied to the example of displacement of the southern boundary of the permafrost region are fully value for any region along the southern boundary.

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Table XVI

Depth of seasonal freezing and thawing of the soil for various regions

Location	Type of seasonal freezing and thawing of the soil				Depth of seasonal freezing and thawing of the soil	
	By amplitude of soil temp.	By mean annual soil temperature	By litho-logical composition	By moisture content, %	Mean in metres	$\frac{\text{Max}}{\text{Min}}$
Leningrad	Moderate maritime, A = 21°C	Stable, t_m = 6°C	Clay loam	20	0.6	$\frac{0.9}{0.3}$
Moscow	Moderate continental, A = 23°C	Stable, t_m = 7°C	Clay loam	20	0.4	$\frac{1.2}{0.2}$
Stalingrad	Continental, A = 31°C	Southern, t_m = 10.5°C	Sandy loam	31	1.6	$\frac{2.0}{1.2}$
Sukhumi	Moderate maritime, A = 18°C	Subtropical, t_m = 16°C, periodically disappearing	-	-	0.2	-
Khibiny	Moderate maritime, A = 18°C	Prolonged stable, t_m = 3°C	Sandy loam	24	1.5	-
Omsk	Continental, A = 31°C	Prolonged stable, t_m = 3.6°C	Clay loam	31	2.2	$\frac{2.5}{1.8}$
Sverdlovsk	Moderate continental, A = 24°C	Stable, t_m = 5.5°C	Clay loam with stone fragments	21	1.1	$\frac{2.0}{0.4}$
Salekhard	Continental, A = 28°C	Transitional, t_m = 0.5°C	Sandy loam	25	2.5	-
Alma-Ata	Continental, A = 27.5°C	Southern, t_m = 11°C	-	28	0.7	$\frac{1.2}{0.3}$

continued

Table XVI - continued

Location	Type of seasonal freezing and thawing of the soil				Depth of seasonal freezing and thawing of the soil	
	By amplitude of soil temp.	By mean annual soil temperature	By litho-logical composition	By moisture content, %	Mean in metres	$\frac{\text{Max}}{\text{Min}}$
Barnaul	Elevated continental, A = 34°C	Prolonged stable, t_m = 4.8°C	Sandy loam	36	2.5	$\frac{2.8}{2.0}$
Irkutsk	Continental, A = 32°C	Prolonged stable, t_m = 2.5°C	Sandy loam	36	2.2	$\frac{2.6}{2.0}$
Igarka	Elevated continental, A = 37°C	Transitional, t_m = 0.5°C	Clay loam	30	2.8	$\frac{3.0}{2.0}$
Chita	Severe continental, A = 43°C	Semi-transitional, t_m = 1.3°C	Sandy loam	36	3.8	-
Yakutsk	Extreme continental, A = 51°C	Prolonged stable, t_m = 2.7°C	Sandy loam	—	1.9	-
Verkhoyansk	Extreme continental, A = 52°C	Stable, t_m = 9°C	Sandy loam	—	1.6	-
Anadyr'	Moderate continental, A = 27°C	Prolonged stable, t_m = 5°C	Clay loam	35	0.7	-
Bukhta-Nagaevo	Moderate maritime, A = 25°C	Transitional, t_m = 0°C	Sand and gravel	18	3.2	-
Blagoveshchensk	Elevated continental, A = 38°C	Prolonged stable, t_m = 4°C	Sand	22	2.2	-
Vladivostok	Continental, A = 28°C	Stable, t_m = 6°C	Clay loam	25	1.2	$\frac{1.5}{1.0}$

Table XVII

The value of $(1 - \frac{1}{f})$ depending on the thickness (h), density (ρ) and coefficients of diffusivity (k) of snow cover

Density of snow cover, ρ	Coeff. of diffusivity, k	Thickness of snow cover, m									
		0,1	0,2	0,3	0,4	0,5	0,6	0,7	0,8	0,9	1,0
0,075	0,010	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	0,288	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	0,081	0,124	0,161	0,197	0,232	0,265	0,297	0,327	0,356

Table XVIII

The value of the coefficient k_c for determining the calculation thickness of snow cover

Thickness of snow cover, m	Coefficient, k_c
0,0—0,2	1,0
0,3—0,4	0,9
0,5—0,6	0,8
0,7—1,0	0,7

Table XIX

Relationship between density and thickness of snow cover

Thickness of snow cover, m	Density of snow, ρ	Thickness of snow cover, m	Density of snow cover, ρ
0,1	0,07	0,6	0,27
0,2	0,12	0,7	0,29
0,3	0,17	0,8	0,30
0,4	0,21	0,9	0,31
0,5	0,24	1,0	0,32

Table XX

Value of $(1 - \frac{1}{f})$ depending on the thickness of snow cover

Thickness of snow cover, m	Value of $1 - \frac{1}{f}$	Thickness of snow cover, m	Value of $1 - \frac{1}{f}$
0,1	0,080	0,6	0,220
0,2	0,120	0,7	0,240
0,3	0,155	0,8	0,265
0,4	0,180	0,9	0,290
0,5	0,200	1,0	0,310

Table XXI

Increase in thickness of snow cover to increase the mean annual temperature of the soil by 1°C depending on the annual amplitude of mean monthly air temperatures

Annual amplitude of mean monthly air temperatures, °C	Increase in thickness of snow cover, cm	Annual amplitude of mean monthly air temperature, °C	Increase in thickness of snow cover, cm
20	38—45	50	13—20
30	24—30	60	11—16
40	17—25		

Table XXII

Increase in the annual amplitude of mean monthly air temperatures required to increase the temperature of the soil by 1°C depending on thickness of snow cover

Thickness of snow cover, cm	Increase in annual amplitude of mean monthly air temperatures, °C	Thickness of snow cover, cm	Increase in annual amplitude of mean monthly air temperatures, °C
12	20,0	75	8,0
29	13,0	95	6,5
50	10,0		

Table XXIII

Increase in t_a, H, A_{mm} $(1 - \frac{1}{f})$ required to change the temperature of the soil by 1°C

Δt_a	ΔH	ΔA_{mm}	$\Delta\left(1-\frac{1}{f}\right)$	Δh
1°	213 м	6—20°	0,035—0,1	10—45 см

Table XXIV

Data of aerological observations (after K.I. Kashin) in Yakutsk on air temperatures for 1940 - 1941

Month	In a shack		Altitude in metres			
	Perennial mean values	For 1940-1941	1000	2000	3000	4000
January	—43,2	—45,4	—32,1	—27,9	—32,6	—37,7
February......	—36,6	—43,5	—31,8	—29,4	—32,2	—37,3
March	—22,4	—30,2	—26,1	—27,3	—30,3	—35,1
April	—8,1	—9,9	—8,7	—13,7	—19,7	—25,7
May..........	+5,5	+4,5	0,7	—6,0	—14,6	—17,0
June	15,5	15,6	12,0	6,3	—1,5	—7,5
July.........	18,9	17,9	16,0	9,6	3,4	—1,8
August	14,7	11,0	8,4	2,8	—2,7	—7,4
September...	6,1	1,5	0,1	—4,8	—9,7	—15,6
October	—7,9	—12,0	—12,9	—16,7	—21,1	—27,3
November....	—27,5	—37,0	—26,2	—26,0	—29,1	—33,6
December....	—39,5	—41,3	—30,1	—28,2	—31,8	—37,0
Mean annual..	—10,3	—14,2	—10,9	—13,6	—18,3	—23,6
Amplitude ...	62,3	63,0	48,1	39,0	36,0	35,9

Table XXV

Variations of soil temperature in Padi Mul'tsai depending on exposure

Depth, cm	Soil temperature, °C		
	Northern slope, 5°	Northern slope, 3°	Southern slope, 35°
Up to 10	10,5	12,0	22,0
10—20	—	—	18,0
20—30	9,0	10,5	16,5
30—40	—	—	9,5
40—50	—	—	9,0
50—60	6,5	7,0	—
60—70	2,0	—	—
70—80	1,0	—	8,0
80—90	0,5	5,5	—
90—100	—	3,5	—
100—110	—	—	6,0
130—140	permafrost	—1,0	—
140—150	—	—	—
150—160	—	—	—
160—170	—	permafrost	—
170—200	—	—	4,0

Table XXVI

Water temperature in Pereval'noe Lake

Depth from lower surface of the ice, m	Water temperature °C 28/4/1948	Depth from lower surface of the ice, m	Water temperature °C 28/4/1948
Ice (1,7 m)			
1,0	1.2	8,0	3,1
2,0	1,3	10,0	3,2
3,0	1,9	12,0	3,4
4,0	2,7	14,0	3,5
5,0	3,0	16,0	3,6
		Bottom	4,0

Table XXVII

Water temperature in Mezhdurechone Lake

Depth from lower surface of the ice, m	Water temp. °C 28/4/1948	Depth from lower surface of the ice, m	Water temp. °C 28/4/1948
Ice (1,75m)		7,0	1,7
1,0	1,2	8,0	1,8
2,0	1,3	9,0	1,8
3,0	1,4	10,0	2,1
4,0	1,5	11,0	2,2
5,0	1,5	12,0	2,4
6,0	1,6	Bottom	2,6

Table XXVIII

Temperature of sawdust in a pile and that of the underlying soil in the territory of the Igarka forest products combine in December 1939*

Depth, m	Temperature °C	Depth, m	Temperature °C
At the surface	—0,4	6,0	72,0
0,5	58,0	6,5	70,0
1,0	76,0	7,0	68,0
1,5	81,5	8,0	46,0
2,0	85,0	9,0	29,0
2,5	87,5	10,0	24,0
3,0	85,0	11,0	20,0
3,5	82,0	12,0	17,0
4,0	81,5	13,0	15,0
4,5	79,0	14,0	14,0
5,0	78,0	15,0	12,0
5,5	78,0		

* To a depth of 9.5 m - sawdust below 9.5m - mineral soil

Table XXIX

Data on the thickness of permafrost for several regions of the U.S.S.R.

Location	Temp. at bottom of layer with annual oscillations, °C	Thickness of permafrost, m	Age and composition of permafrost	Remarks
Vorkuta	-1 to -1.5	40 - 130	Moraine, sandstone and clay shale	Water divide
Salekhard	-1	20 - 80	Alluvial sand	Floodplain
Salekhard	-2 to -4	200 - 380	Quaternary and Cretaceous deposits (clay loam, sand)	Water divide
Yakutsk	-3 to -7	230	Quaternary deposits Jurassic sandstone and clay shale	First terrace
Vorkuta-Vom	about 0	10 - 30	Quaternary deposits (moraine, sand loam)	First terrace
Verkhoyansk	-6 to -7	180	Quaternary deposits (sand, sand loam)	Second terrace
Vilyui	-7	800		Water divide
Kozhevnikova Bay	-12	600		Edge of valley
Bodaibo	0 to -1	20 - 100	Quaternary deposits (sand, sand loam)	First terrace
Skovorodino	-1	50	Same as above	Same as above
Igarka	0 to -1	20 - 50	Same as above	Same as above
Chita	about 0	8 - 20	Same as above	Same as above
Khasyn, alt.340 m	-1.5 at 30 m	115	Cretaceous andesite and sandstone	On a hill
Omsukchan, Galimyi Section, borehole No.1, alt.827 m (near Lake Sol'veik)	-1.6 at 10 m	57	Triassic - Jurassic, shale, hornfels, sandstone	In river valley
Dzhebariki-Khaya, borehole No.19, alt.215 m	-6.6 at 10 m -6.1 at 15 m -5.5 at 30 m -3.7 at 170 m	Not penetrated at 261 m; calc. to be 416 m	Palaeozoic; sandstone, sandy limestone, silt stone, argillites, coal	On the bank of Aldan River, 30 m terrace
Arkagala, at Znatnaya R., borehole No.27, alt.780 m	-6.5 at 10 m -6.0 at 15 m -3.0 at 125 m	Not penetrated at 125 m; calc. to be 170 m	Upper Cretaceous sandstone with clay laminae	On the second terrace of the Armagelly River
Butugychaya, alt. 1400 m	-7 at 70 m	More than 230 m	Cretaceous, granite	On a hill
El'gen-Ugol'nyi, borehole No.1, shaft alt. 603 m	-1.3 at 15 m -1.7 at 30 m -0.8 at 55 m	Not penetrated at 55 m; calc. to be 77 m	Neogenic, conglomerates, sandstone and clay	
Valley of Verina R. at Kan'on spr., borehole No.5, alt. 779 m	-3.8 at 15 m	Not penetrated at 227 m; calc. to be 263 m	Cretaceous, granite	On a slope of the Verina River
Sretnikan (Kotel village), borehole No.18, alt. 320 m	+1.8 at 10 m +0.4 at 15 m -0.9 at 30 m -1.2 at 100 m	Not penetrated at 100 m; calc. to be 183 m	Clay shale with inerstrata of tuff	In river bed
Alyaskitovyi, borehole No.1, alt. 932 m	-6.4 at 10 m -5.2 at 15 m -3.6 at 30 m -2.2 at 90 m	166 m	Triassic, sandstone, limestone and hornfels	In floodplain of Angara R.
Artyk, borehole No.1, alt. about 800 m	-7.0 at 10 m -6.0 at 30 m -3.9 at 120 m	Not penetrated at 128 m; calc. to be 293 m	Tertiary sandstone, conglomerates	-
N. Zyryanka at the mouth of Yasachnaya R., borehole No.2	-3.2 at 15 m -2.9 at 30 m -2.7 at 79 m	Not penetrated at 114 m; calc. to be 200 m	Tertiary sandstone, sand loam, stones	-
Pevek-Val'kumei, borehole No.1, alt. about 10 m	-5.2 at 50 m	230 m	Cretaceous, granite	On the seashore

Table XXX

Freezing and thawing of the soil over a 100,000 year period at A = 6°C, t_m 0°C, computed by simulation on the hydrointegrator IG-1

Moisture content w, %	Coeff. of heat conductivity of frozen soil η kcal/m·hr·deg	Coeff. of heat conductivity of unfrozen soil, η_1, kcal/m·hr·deg	Coeff. of diffusivity of frozen soil a_2, m²·hr	Maximum depth of freezing, m	Duration of freezing process, years	Preservation of permafrost, years
10	2,2	1,7	0,00403	180	33 000	52 000
15	2,5	1,88	0,00422	196	33 000	54 200
20	2,8	2,1	0,00439	210	33 000	55 300

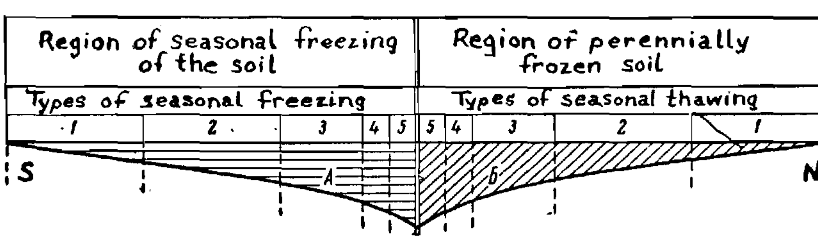

Fig. 27

Latitudinal zonation of the basic types of seasonal freezing and seasonal thawing of the soil from south to north

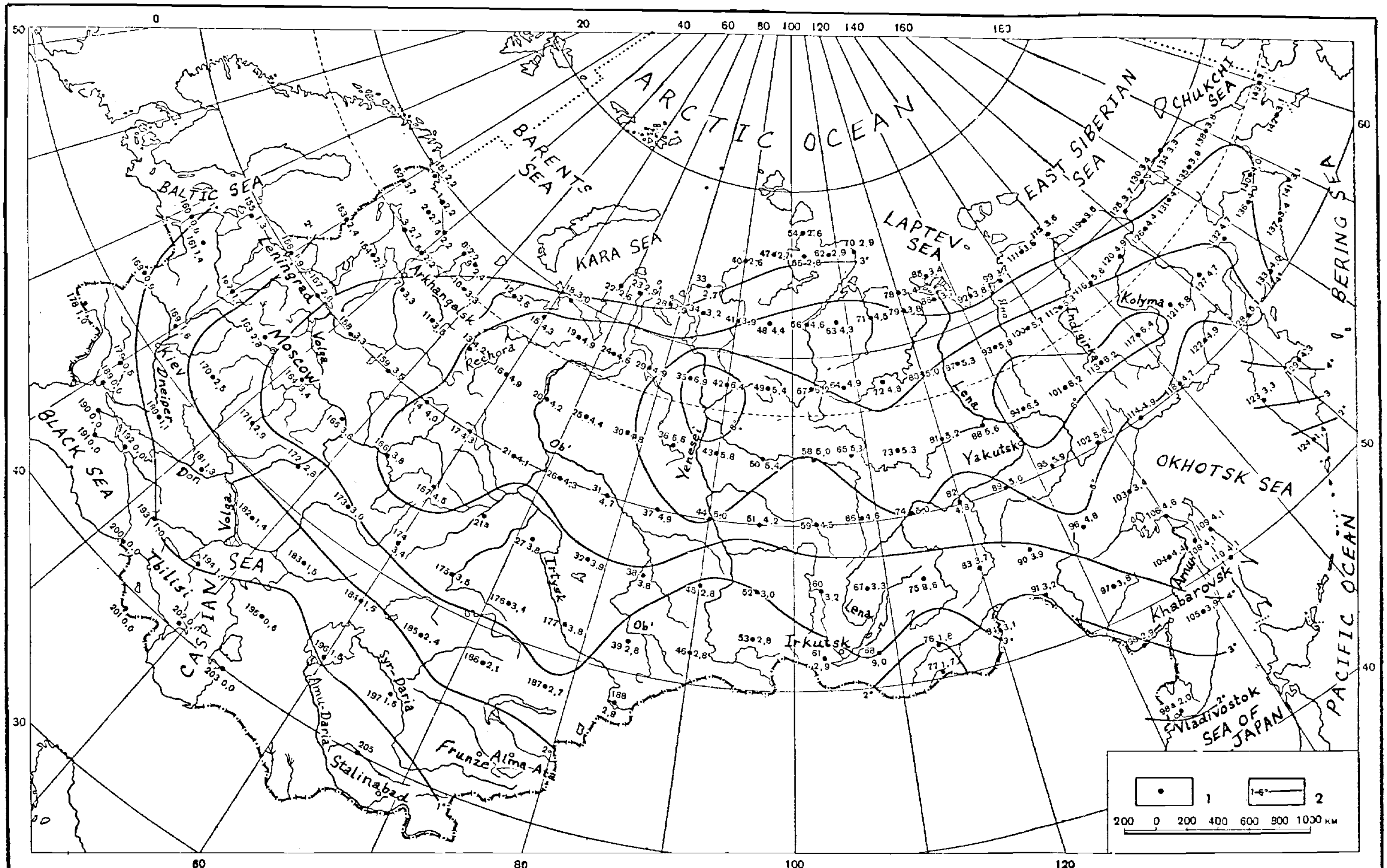

Fig. 28

A map showing differences between the temperature of the soil and mean annual air temperatures (S_z). 1 - calculation points; 2 - isolines of the value of S_z

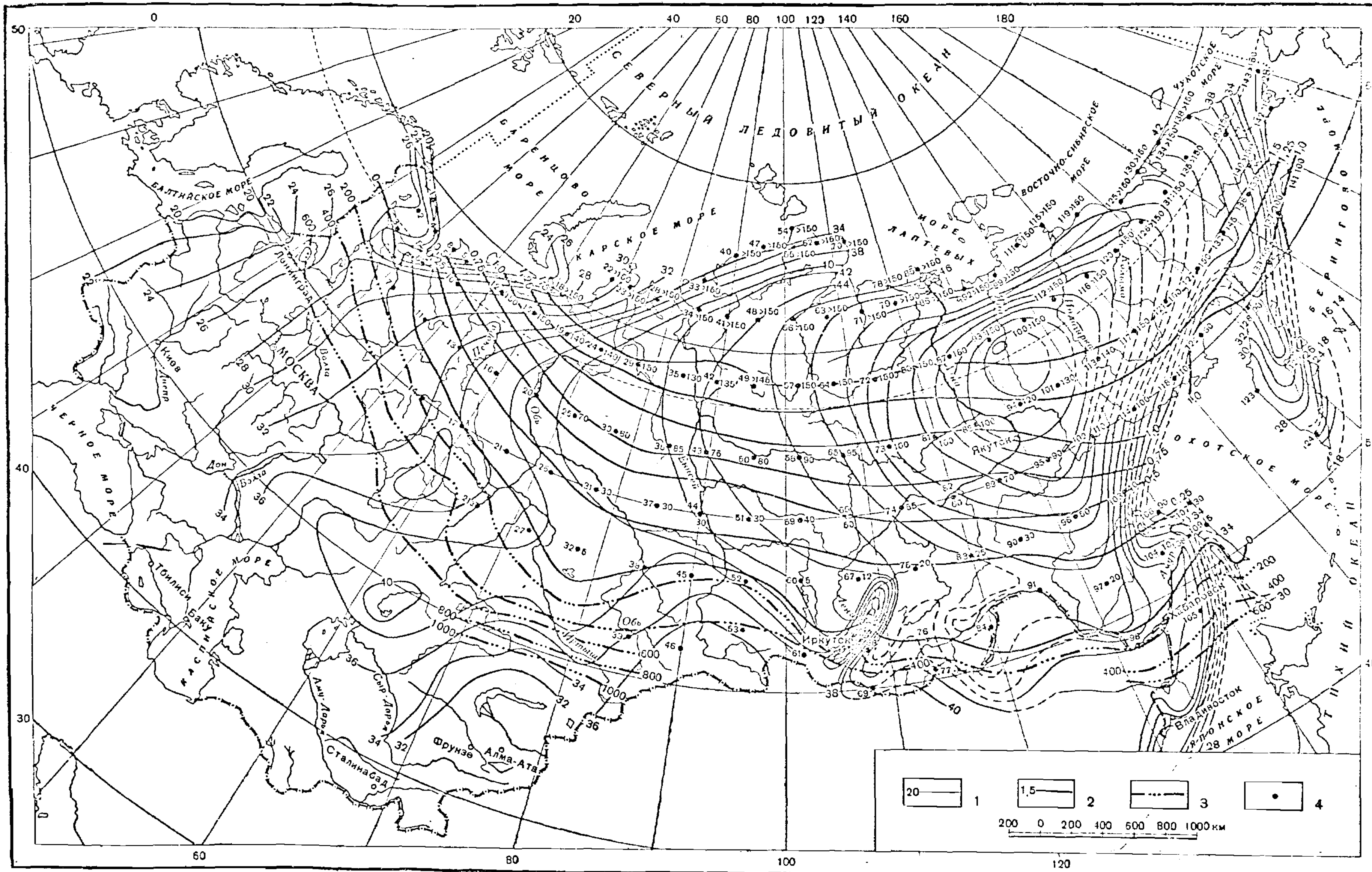

Fig. 29

A map of the critical thickness of snow cover converted to sea level. 1 - amplitude of air temperature, 2 - critical thickness of snow cover in metres, 3 - isolines of absolute altitudes corresponding to zero mean annual air temperature; 4 - calculation points

Note: See Fig. 28 for place names

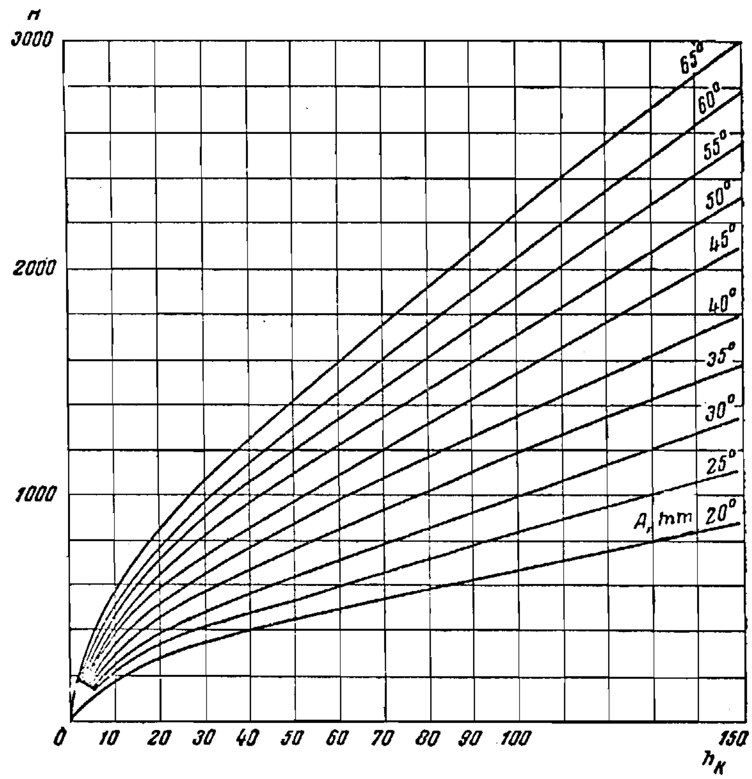

Fig. 30

Nomogram No. 1 for calculation the critical thickness of snow cover (h_k) depending on the altitude (H) and the mean annual amplitude of the air temperature (A, mm)

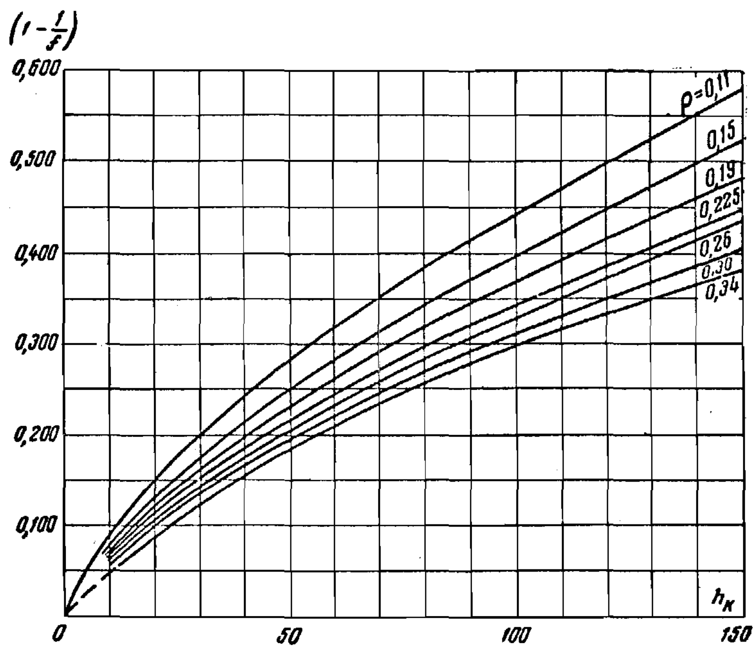

Fig. 31

Nomogram No. 2 for calculating the critical thickness of snow cover (h_k) depending on density of the snow (ρ)

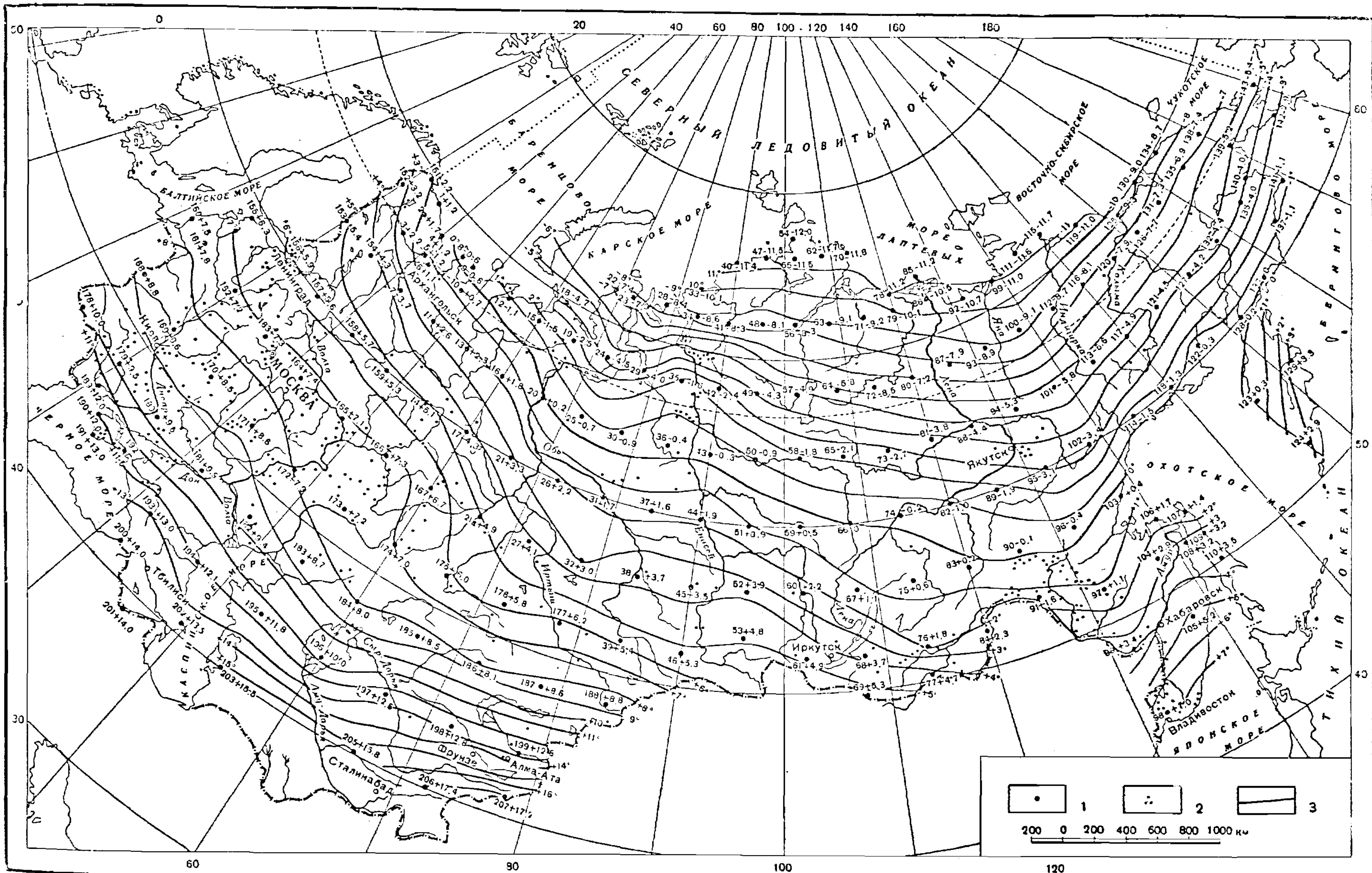

Fig. 32

The map of the mean annual soil temperature converted to sea level
1 - calculation points, 2 - control points, 3 - soil isotherms

Note: See Fig. 28 for place names

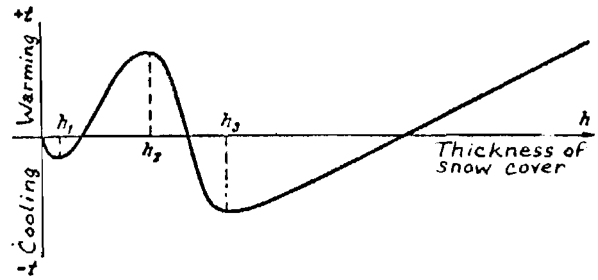

Fig. 33

Influence of thickness of snow cover on increasing and decreasing soil temperatures

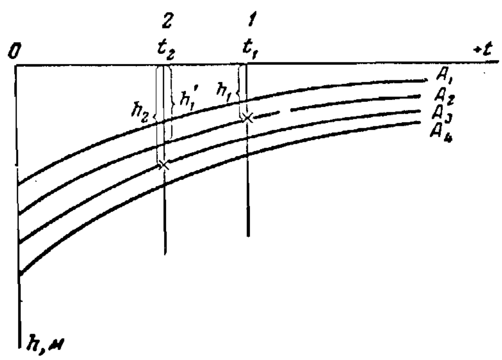

Fig. 34

Depth of seasonal freezing of the soil with snow cover (1) and in a section from which the snow has been removed (2)

t - soil temperature, A - temperature amplitude at soil surface, h - depth of soil freezing

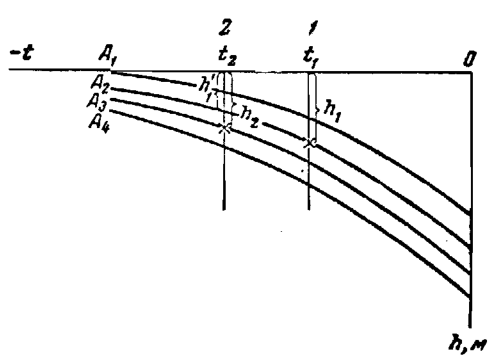

Fig. 35

Depth of seasonal thawing of the soil on a section with snow cover (1) and on a section from which snow cover has been removed (2). Legend same as in Fig. 34

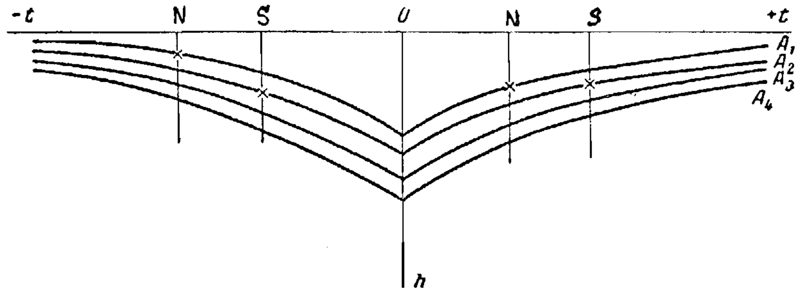

Fig. 36

Depth of seasonal freezing and thawing of the soil on slopes of north (N) and south (S) facing slopes

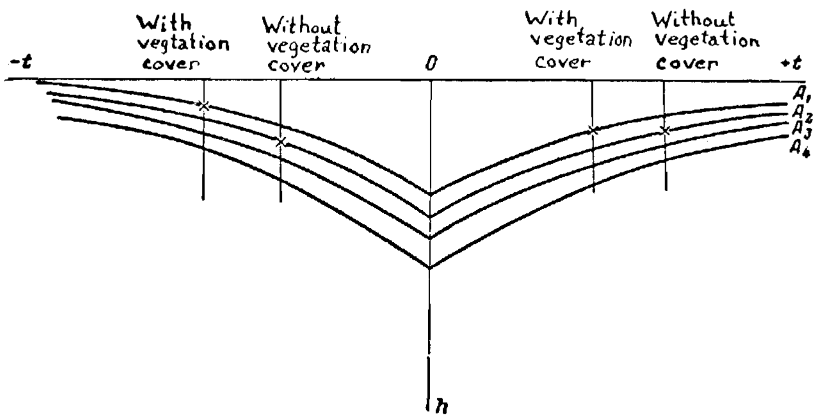

Fig. 37

Depth of seasonal freezing and thawing of the soil in sections with and without vegetation cover in southern regions

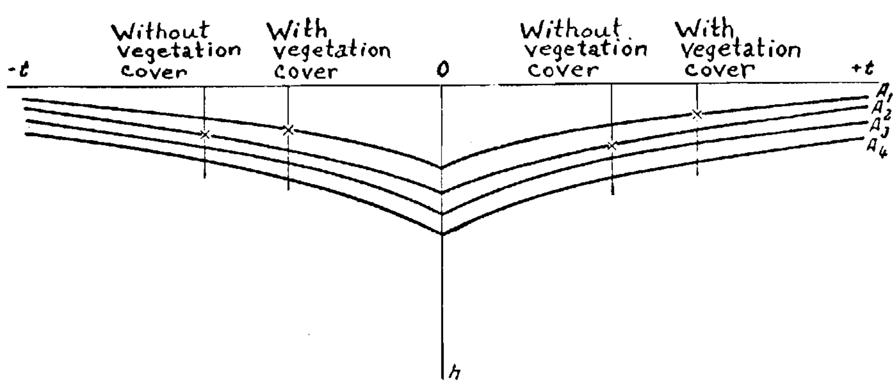

Fig. 38

Depth of seasonal freezing and thawing of the soil in sections with and without vegetation cover in northern regions

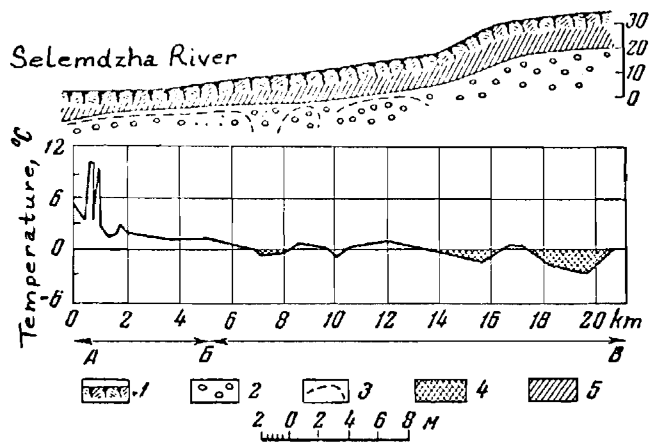

Fig. 39

Temperature profile between Selemdzha and Byssa Rivers

1 - vegetation cover, 2 - coarse gravel,
3 - permafrost table, 4 - frozen soil,
5 - clay loam cover

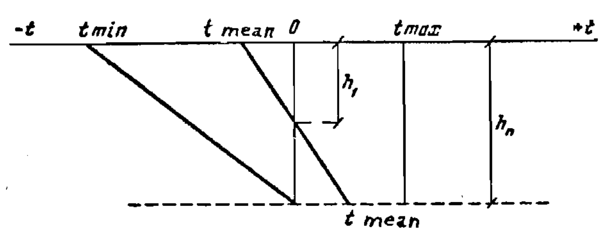

Fig. 40

Diagram of distribution of temperatures in ice cover
t - temperature, h_f - depth of freezing of the
water (thickness of ice), h_1- depth at which
the mean annual temperature is 0°C

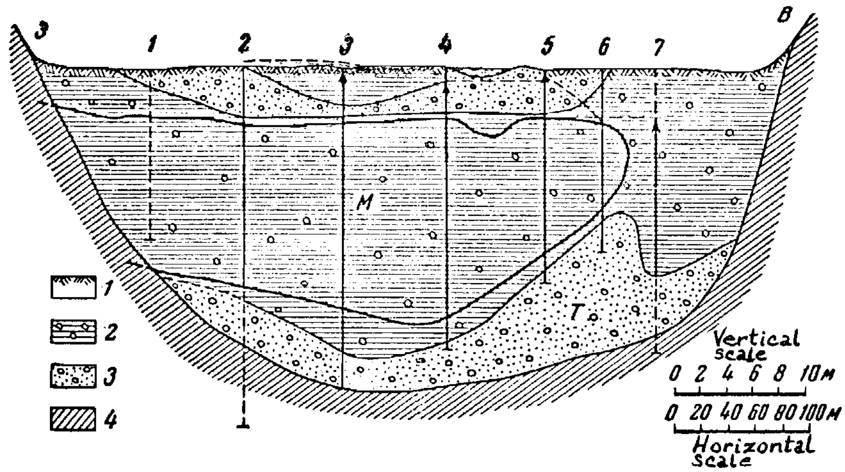

Fig. 41

Cross-section through a valley after N.I. Tolstikhin

1 - top soil, 2 - clay with stones and sand,
3 - stones with sand, 4 - clay shale and sandstone

ck = borehole

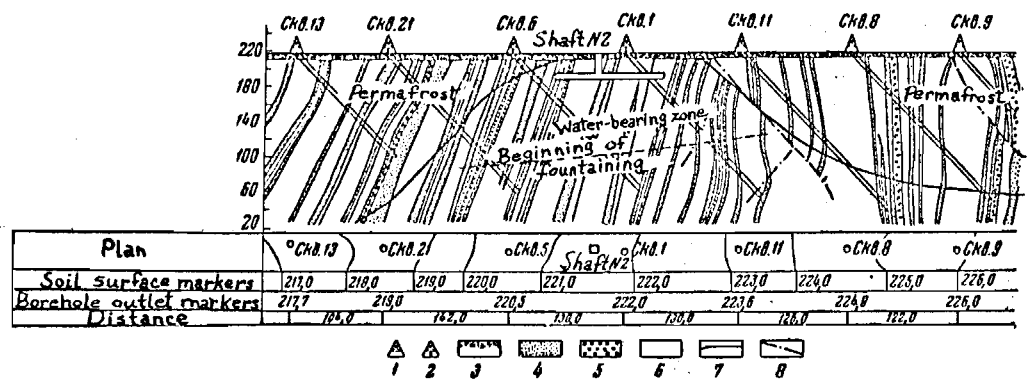

Fig. 42

Talik in the region of Khal'mer-Yu (Vorkuta)
according to V.M. Barygin

1 - boreholes, 2 - fountaining wells, 3 - Quaternary deposits,
4 - sandstone, 5 - conglomerates, 6 - siltstones and argillites,
7 - lower boundary of the permafrost, 8 - lines of tectonic fractures

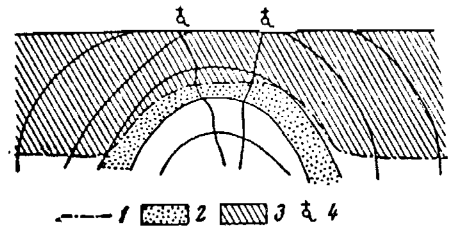

Fig. 43

Diagram of lower surface of permafrost
in an anticlinic formation

1 - lower surface of the permafrost, 2 - water-bearing horizon,
3 - permafrost, 4 - spring

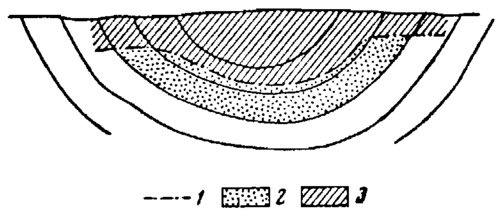

Fig. 44

Diagram of lower surface of the permafrost
with a synclinic formation

Legend same as in Fig. 43

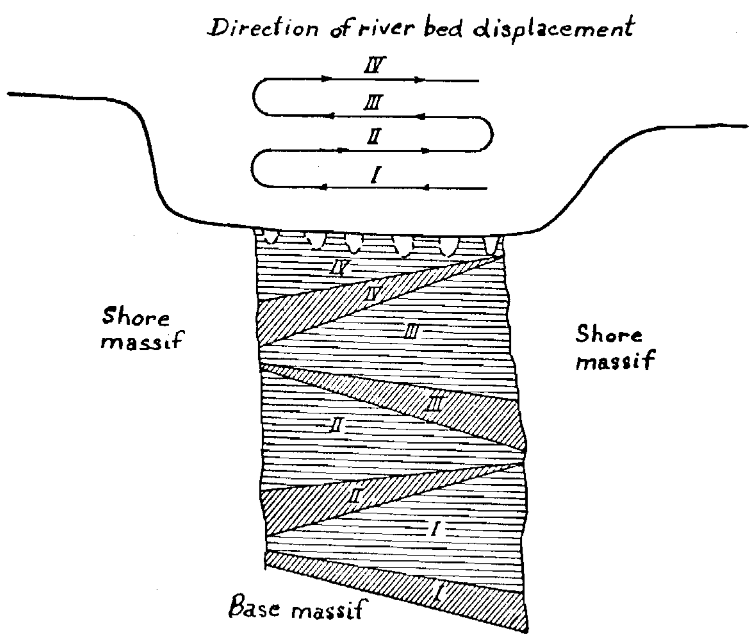

Fig. 45

Diagram of permafrost occurrence in a floodplain
resulting from repeated river bed displacement

Horizontal hatching - permafrost
diagonal hatching - taliks

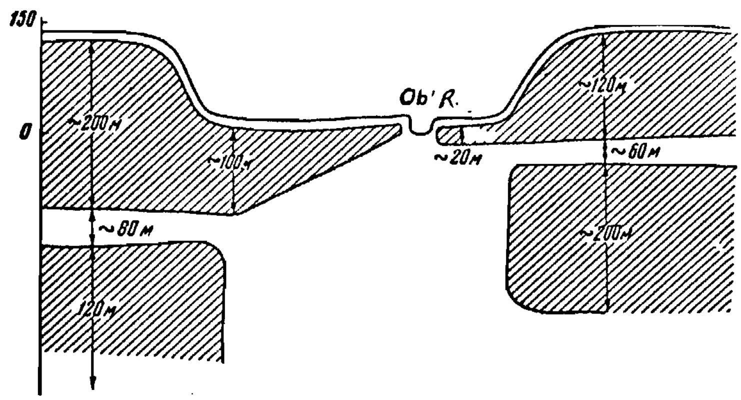

Fig. 46

Diagram of permafrost occurrence in the Salekhard region

Diagonal hatching - permafrost

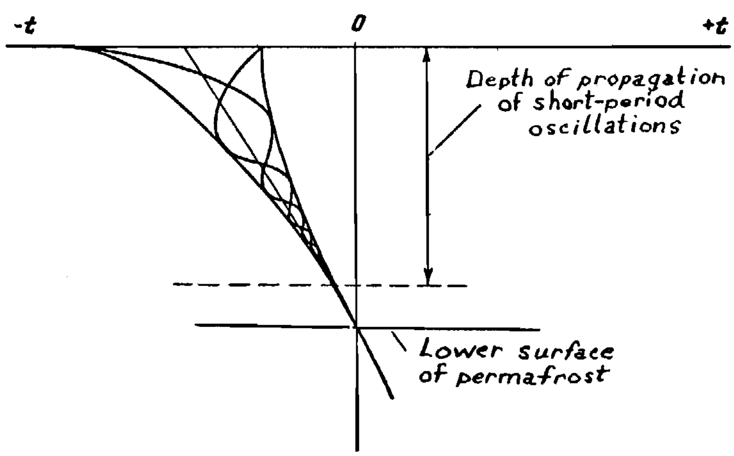

Fig. 47

Diagram of propagation of short-period temperature oscillations in permafrost

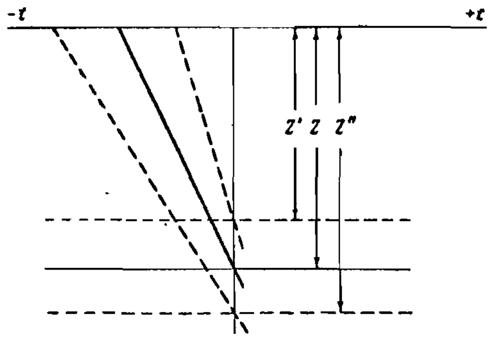

Fig. 48

Diagram of changes in the temperature field and thickness of permafrost resulting from long-period temperature oscillations at the soil surface

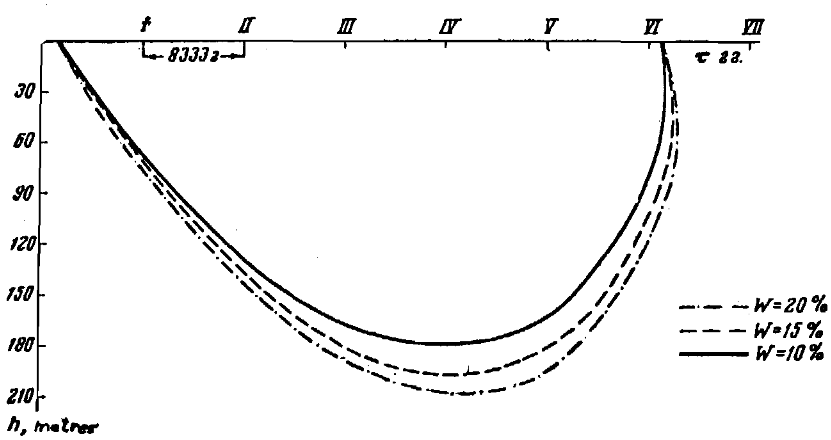

Fig. 49

Graphs of changes in depth of freezing of the ground with respect to time during a 100,000-year temperature oscillation at the surface

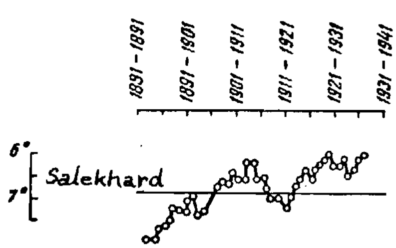

Fig. 50

Curve of moving ten-year average annual air temperatures at Salekhard for the period from 1881 to 1941, after E.S. Rubinshtein

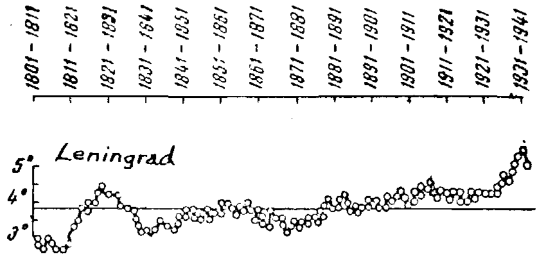

Fig. 51

A curve of the moving ten-year mean annual air temperatures at Leningrad from 1801 to 1911 after E.S. Rubinshtein

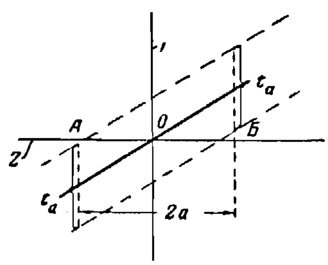

Fig. 52

Diagram of periodic displacement of the southern boundary of permafrost region along the meridian

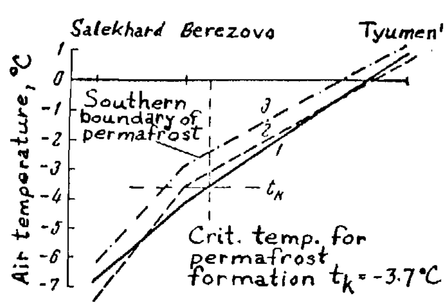

Fig. 53

Changes in the moving ten year mean annual air temperatures along the line Salekhard - Berezovo - Tyumen' for 1895 (curve 2) and 1905 (curve 3) and the mean perennial air temperature (curve 1)

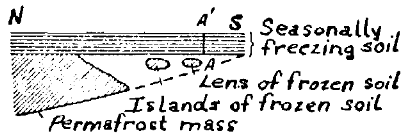

Fig. 54

Diagram of displacement of southern boundary of permafrost region during degradation, after M.I. Sumgin

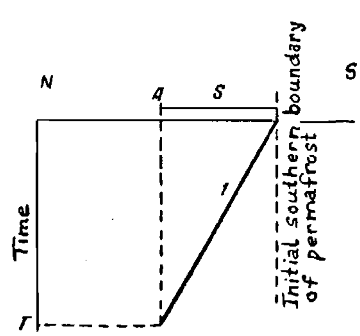

Fig. 55

Simplified diagram of M.I. Sumgin illustrating the displacement from south to north of the southern boundary of permafrost region during degradation depending on time

S - amount of displacement of the souther boundary from south to north

1 - line of displacement of the souther boundary during time T

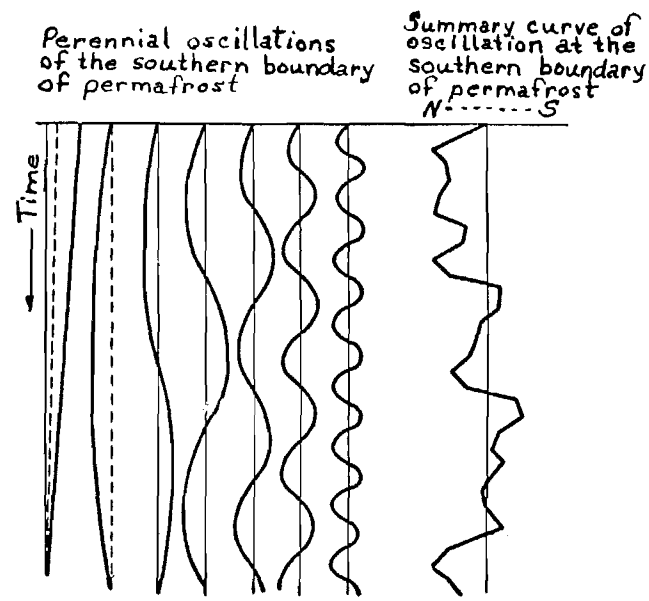

Fig. 56

Diagram of displacement along the meridian of the southern boundary of permafrost considered as a summary effect of a number of harmonic oscillations