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Balobaev, V. T.; National Research Council of Canada. Division of Building Research

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HEAT EXCHANGE BETWEEN PERMAFROST AND THE ATMOSPHERE IN THE PRESENCE OF A VEGETATION COVER

BY

V. T. BALOBAEV

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PREFACE

The heat exchange between the ground and the atmosphere in the North is extremely important in the formation and continued existence of permafrost. An understanding of this process and how it is influenced by surface terrain features such as vegetation is vital in furthering knowledge of the thermal regime and other characteristics of permafrost having important engineering implication.

Investigations in this field are underway in Canada and the U.S.S.R. This article was written by a senior worker at the Permafrost Institute, Academy of Sciences of the U.S.S.R., Yakutsk in Eastern Siberia. It reviews North American and Soviet studies and presents analytical material pertaining to the important role of vegetation in the heat exchange between the permafrost and the atmosphere.

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Title: Heat exchange between permafrost and the atmosphere in the presence of a vegetation cover

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HEAT EXCHANGE BETWEEN PERMAFROST AND THE ATMOSPHERE IN THE PRESENCE OF A VEGETATION COVER

Heat and moisture exchange in thawing soil and in the atmosphere are intimately connected. The rate of exchange depends not only on the properties of the air and mineral layers but also on the interface which has a great deal of influence on the total amount of accumulated radiation energy, evaporation and turbulent heat exchange in the air, and this in turn influences the heat flux into the permafrost and the thawing of permafrost in the summer.

Under natural conditions one can encounter the most varied surfaces, but in the permafrost zone the greater part of the surface is covered by vegetation that actively participates in heat and moisture exchange between the air and the soil. The vegetation cover creates a microclimate that differs from that of uncovered ground. Depending on the density, height, canopy density and the type of vegetation, the ground surface receives and loses various quantities of heat. Depending on heat and moisture exchange features, a vegetation cover can be divided into three basic groups: forest, shrub and meadow.

On bare ground the transformation of radiant energy into thermal energy and its distribution occurs in a very thin layer. On the other hand, inside a vegetation cover it is impossible to delimit such a layer. The absorption of radiation and emission of heat in a forest occurs primarily at the outer surface of the crowns involving a certain three-dimensional space. Only a small quantity of radiation can penetrate through the canopy, depending on the density and closeness of the crowns.

Kittredge (1951) measured radiation intensity in an open area and under a forest canopy. In the first case it was 1.5 and in the second 0.1 $cal/cm^2 \cdot min$. Tree crowns absorb more than 90% of the radiation. R. Geiger (1960), on the basis of Trapp's measurements indicates that in a beech forest, as a rule, 80% of the incoming radiation is intercepted by the crowns and only 5% reaches the forest litter. Most of the radiation is absorbed by the outer surface of the crowns.

A meadow vegetation, in contrast to a forest, does not have an outer surface and therefore the absorption of radiation occurs inside the entire grass stand. According to the data of Angstrom (Geiger, 1960), on a section overgrown by meadow vegetation only 0.19 $cal/cm^2 \cdot min$ of the incoming 1,08 $cal/cm^2 \cdot min$ of radiation heat reaches the ground surface. The radiation intensity at a height of 0.5 and 0.1 m was 1.04 and 0.28 $cal/cm^2 \cdot min$ respectively. Hence it can be seen that the absorption of radiation is gradual on approaching the ground surface.

Shrubbery occupies an intermediate position between a forest and a meadow.

A vegetation cover produces a fundamental change in water exchange between the soil and air. Evaporation from the surfaces of leaves is added to evaporation from the surface of the soil. The total evaporation from soil covered by vegetation may be 2-5 times greater that evaporation from uncovered soil (Geiger, 1960). Direct evaporation from the soil surface under a vegetation cover decreases with increase in density of the vegetation. When there is a continuous vegetation cover almost all the moisture is evaporated through the plants (Fundamentals of agrophysics, 1959). Even at night when there is no transpiration, evaporation from the surface of the soil is not large because of the high humidity in the air surrounded by vegetation and the small turbulent exchange coefficient.

Heat is transferred in air by convection; only near the ground surface in a very thin layer where the air is immobile is there any heat transferred by molecular means. Inside a vegetation cover convective motion of air is difficult and therefore the rate of turbulent heat exchange is less than in the free atmosphere. Even at high wind velocities the rate of air movement inside the vegetation cover is greatly attenuated. Air velocity is reduced primarily by the upper parts of vegetation. According to the data of Geiger (1960), in a pine forest the wind velocity is reduced by 2-3 times in the upper parts of the crowns; between the crowns and the forest litter there is a gradual attenuation in wind velocity down to the zero point.

The above noted features of heat exchange in a vegetation cover result in a particular temperature distribution in the vegetation cover and in the air above it. Figure 1 shows the distribution of temperatures at midday in a flower bed and in a field of winter rye (the vegetation cover in all figures is shown schematically corresponding to its height, density and type) according to the data of R. Geiger (1960). In the flower bed the leaves retained radiation energy in the upper layers and therefore the maximum temperature is located near the upper boundary of the vegetation. Only a small part of the radiation reached the soil surface, and therefore the temperature was below that of the air temperature at a height of 1.5 m. In the rye stand the temperature distribution was different. Beginning with the boundary between the air and the vegetation the temperature increases on moving downwards reaching a maximum at a height of about 5 cm from the soil surface, or somewhat higher among the stems. The temperature of the soil surface is always higher than the air temperature at a height of 1.5 m. The temperature distribution above a cut section of the field is described by a typical logarithmic curve. Since in summer the advent of heat to the soil surface exceeds its loss, on the average the temperature curves in the summer period should be analogous to the midday curves that we are considering here.

In general terms this is the qualititive influence of various types of vegetation on heat and moisture exchange between soil surface and the atmosphere based on data in the literature (Obolenskii, 1944; Kittredge, 1951; and Geiger, 1960).

The influence of vegetation on the temperature regime and thawing of permafrost has received less attention. Many investigations have shown that under any vegetation cover the temperature of the soil in summer is lower and the depth of thawing less than in areas where there is no vegetation cover (Koloskov, 1918; Tyrtikov, 1956). Grass stands have less influence on the depth of thawing and soil temperature that shrubs and forests. A. P. Tyrtikov has indicated (Fundamentals of geocryology, 1959) that under a vegetation cover the depth of thawing is usually 1.5-3 times less than in areas without vegetation.

One could cite more evidence on the influence of vegetation on temperature and depth of thawing, but data of this type reduce the complex imfluence of vegetation to statements regarding the temperature of surfaces with vegetation cover and without it, or to some values of thawing depths of soil under various vegetation associations. Such an approach ignores the role of other important factors and their relative influence on heat exchange and thawing.

It is evident from the above that the influence of vegetation cover on heat exchange in the atmosphere-soil system and on thawing of permafrost is far from being fully elucidated. Some relationships obtained in the past, however, make it possible to obtain approximate theoretical solutions to this question. Let us consider jointly the processes of heat exchange in free air, vegetation cover and the thawing permafrost layer. Mathematically these processes are described by three heat exchange equations taking into account the thermophysical features of each medium.

As already stated above, heat exchange in the air results from turbulence, and its rate is determined by the coefficient of turbulent exchange k, which is thousands of times greater than the molecular heat conductivity of air. This determines the quasi-steady state temperature regime of air. The re-establishment of the temperature field after it has been disturbed occurs almost instantaneously (Budyko, 1946). The turbulent exchange coefficient in the air adjacent to the ground increases with height. M. I. Budyko (1956) and many other authors consider that it depends linearly on height:

$$k = k_1 z,$$

where k_1 is the turbulent exchange coefficient for a unit of height in free air.

Inside the vegetation cover turbulent heat exchange is retained since the wind, although weakened, nevertheless penetrates the vegetation cover. The turbulent exchange coefficient decreases but nevertheless remains substantially larger that the molecular heat conductivity and it is therefore permissible to consider the temperature field in a vegetation cover likewise to be in a quasi-steady state. There are no data whatsoever regarding the relationship of the turbulent heat exchange coefficient within the vegetation cover to height. We shall consider that, as in free air, it increases linearly with height, i.e.,

$$k = k_2 z,$$

where k_2 is the turbulent exchange coefficient inside a vegetation cover when $z = 1$.

At the boundary between the vegetation and the free air the two coefficients of turbulent exchange should be equal.

In solving problems dealing with the thawing of permafrost extensive use is made of the assumption that the temperature in the thawed layer is in a quasi-steady state. This position was first stated and used in 1931 by L. S. Leibenzon (1955). It is based on the fact that the rate of thawing is much less than the rate of heat propagation in the soil. Since vegetation retards thawing, in our case it can be assumed that the temperature field in the thawed layer is in a quasi-steady state.

We limit the layer of air under investigation to a height h measured from the ground surface and denote the mean height of vegetation by l, and the depth of thawing by ξ. The value of l apparently should coincide with the "height of roughness" above which the wind profile is subject to the logarithmic law. Thus we shall solve the system of equations:

$$\frac{d}{dz}[\varkappa_0 + k_2 l - k_1(z+l)]\frac{dT_1}{dz} = 0,$$
$$-h < z < -l, \tag{1}$$

$$\frac{d}{dz}(\varkappa_0 - k_2 z)\frac{dT_2}{dz} + \frac{G(z)}{C_p \rho} = 0,$$
$$-l < z < 0, \tag{2}$$

$$\frac{d^2 T_3}{dz^2} = 0, \quad 0 < z < \xi, \tag{3}$$

where T_1, T_2, T_3 are the temperature of free air, the air within the vegetation cover, and in the thawed soil layer respectively;

$G(z)$ is the three-dimensional (homogeneous) constantly acting source of heat (for explanation see below);

C_p is the heat capacity of the air;

ρ is air density.

The origin of the coordinates is located on the surface, the positive axis z is directed down into the soil and the negative into the air. The system of equations (1), (2) and (3) assumes that the temperature of the permafrost is equal to the thawing temperature, or is very close to it. This is of no particular importance in principle, since the influence of the frozen layer during the process of thawing is not significant and can be readily accounted for by approximation (see foregoing article in this book).

The analysis of experimental data carried out at the beginning of this article shows that the accumulation of radiation heat occurs not only on the upper boundary of the vegetation and on the ground surface $z = 0$, but also within the vegetation cover. There is heat loss due to radiation and evaporation inside the vegetation cover. To calculate the difference between the inflow and loss of heat inside the vegetation cover an additional term is introduced in equation (2) that takes into account the influence of these internal sources $G(z)$.

Assume that the radiation balance of a plot with vegetation is B, and that heat V is consumed in the total evaporation. Then the quantity of heat passing from the atmosphere to the vegetation boundary and determining the exchange process will be $B - V = R$. Of the total quantity of heat R, part is absorbed by the surface of the vegetation $R - R_0$, the other part reaches the soil surface R_π and the third part is absorbed inside the vegetation cover $(R_0 - R_\pi)$. Here R_0 is the portion of the heat penetrating into the vegetation cover. It is not difficult to see that $R - R_0$ is equal to the radiation balance of the outer surface of the leaves minus their transpiration; $R_0 - R_\pi$ is equal to the radiation balance of the inner parts of the vegetation minus their transpiration, and R_π is equal to the radiation balance of the soil surface minus evaporation from the surface.

The value of $R_0 - R_\pi$, depending on the type of vegetation cover, can be distributed by height in any way at all. For any given distribution the type of function $G(z)$ will be singular. For determinacy we assume that this quantity of heat is absorbed uniformly by height. Then

$$G(z) = \frac{d}{dz}\left[R_\Pi + \frac{R_0 - R_\Pi}{l} z\right] = \frac{R_0 - R_\Pi}{l}, \tag{4}$$

and equation (2) will have the form:

$$\frac{d}{dz}(\varkappa_0 - k_2 z)\frac{dT_2}{dz} + \frac{R_0 - R_\Pi}{C_p \rho l} = 0; \quad -l < z < 0. \tag{2a}$$

The problem will be solved with constant boundary conditions:

$$\text{when } z = -h \quad T_1 = T_B, \tag{5}$$

$$\text{when } z = -l \quad T_1 = T_2, \quad \frac{dT_1}{dz} - \frac{dT_2}{dz} = \frac{R - R_0}{(\varkappa_0 + k_2 l) C_p \rho}, \tag{6}$$

$$\text{when } z = 0 \quad T_2 = T_3, \quad \lambda_2 \frac{dT_2}{dz} - \lambda_3 \frac{dT_3}{dz} = R_\Pi, \tag{7}$$

$$\text{when } z = \xi \quad T_3 = T_0, \quad -\lambda_3 \frac{dT_3}{dz} = q_0 \gamma_3 \frac{w}{1 + w}\frac{d\xi}{dt}, \tag{8}$$

where T_B - air temperature at height h;

T_0 - thawing temperature of frozen soil;

λ_2 - coefficient of molecular heat conductivity of the air;

λ_3 - coefficient of heat conductivity of thawed soil;

q_0 - latent heat of ice melting;

γ_3 - bulk density of the soil;

w - moisture content of soil by weight.

The second condition in equation (8) means that all of the heat moving to the boundary between the thawed and frozen soil is absorbed during the phase transition between ice and water, as a result of which the surface $z = \xi$ is in motion, i.e., $\xi = \xi(t)$.

We consider that in the vegetation cover heat is transferred only by means of air; plant stems do not conduct heat. This follows from the fact that convective heat conductivity of air is greater than the heat conductivity of vegetation, the volume of air is as a rule much greater than the volume of vegetation, and moreover the movement of cold liquids in vegetation from the roots hinders the transfer of heat through the stems downwards (Geiger, 1960).

The solution of equations (1), (2a) and (3) does not involve any difficulty (Stepanov, 1959);

$$T_1 = C_1 \ln [\varkappa_0 + k_2 l - k_1 (z + l)] + C_2; \quad z < 0, \tag{9}$$

$$T_2 = C_3 \ln (\varkappa_0 - k_2 z) + C_4 - \frac{R_0 - R_п}{C_p \rho l k_2^2} (\varkappa_0 - k_2 z); \; z < 0, \tag{10}$$

$$T_3 = C_5 + C_6 \frac{z}{\xi}; \quad z > 0. \tag{11}$$

The continuous integrations of C_i are found after reducing the solutions obtained in correspondence with the boundary conditions (5) - (8). Leaving out these computations we write the final form of the solution;

$$T_1 = T_в + \alpha_1 \frac{\ln \frac{y'}{y_1}}{\ln \frac{y_1}{y_2}} \times$$

$$\times \left[\frac{T_в - T_0 - R_п \frac{\xi}{\lambda_3} + (R - R_0) \frac{\alpha_1}{\lambda_3} + (R_0 - R_п) \beta \left(\frac{\alpha_1 \lambda_2 y_2}{\varkappa_0 \lambda_3} + \xi \frac{\lambda_2}{\lambda_3} + l \right)}{\alpha_1 + \alpha_2 + \xi} - \right.$$

$$\left. - (R_0 - R_п) \frac{y_2}{k_2 l \lambda_3} - (R - R_0) \frac{1}{\lambda_3} \right]; \tag{12}$$

$$T_2 = T_в + (R - R_0) \frac{\alpha_1}{\lambda_3} + (R_0 - R_п) \beta \left(\frac{y_2 \lambda_2 \alpha_1}{\varkappa_0 \lambda_3} + l + z \right) +$$

$$+ \left(\alpha_2 \frac{\ln \frac{y''}{y_2}}{\ln \frac{y_2}{\varkappa_0}} - \alpha_1 \right) \times$$

$$\times \frac{T_в - T_0 - R_п \frac{\xi}{\lambda_3} + (R - R_0) \frac{\alpha_1}{\lambda_3} + (R_0 - R_п) \beta \left(\frac{\alpha_1 \lambda_2 y_2}{\varkappa_0 \lambda_3} + \xi \frac{\lambda_2}{\lambda_3} + l \right)}{\alpha_1 + \alpha_2 + \xi} \tag{13}$$

$$T_3 = T_0 + (\xi - z) \times$$
$$\times \left[\frac{T_в - T_0 + R_п \frac{\alpha_1 + \alpha_2}{\lambda_3} + (R - R_0) \frac{\alpha_1}{\lambda_3} - (R_0 - R_п)\, \beta \left(\frac{\lambda_2}{\lambda_3} \alpha_2 - l - \frac{\alpha_1}{\beta \lambda_3} \right)}{\alpha_1 + \alpha_2 + \xi} \right], \tag{14}$$

where

$$\left.\begin{aligned} y_1 &= \varkappa_0 + k_2 l + k_1 (h - l); & y_2 &= \varkappa_0 + k_2 l; \\ y' &= \varkappa_0 + k_2 l - k_1 (z + l); & y'' &= \varkappa_0 - k_2 z; \\ \alpha_1 &= \frac{\varkappa_0 \lambda_3}{\lambda_2 k_1} \ln \frac{y_1}{y_2}; \quad \alpha_2 = \frac{\varkappa_0 \lambda_3}{\lambda_2 k_2} \ln \frac{y_2}{\varkappa_0}; & \beta &= \frac{\varkappa_0}{l k_2 \lambda_2}. \end{aligned}\right\} \tag{15}$$

The individual components of the formulae obtained express the influence of those heat fluxes which are retained by the soil surface R_π, vegetation cover $R_0 - R_\pi$ and the surface layer of vegetation $R - R_0$. Each formula combines the more important parameters influencing temperature.

In obtaining formulae (12), (13) and (14) the last condition of formula (8) was not used. Let us substitute in it the value of the temperature gradient

$$\left. \frac{\partial T_3}{dz} \right|_{z=\xi} = - \frac{T_п - T_0 + R_п \frac{\alpha_1 + \alpha_2}{\lambda_3} + (R - R_0) \frac{\alpha_1}{\lambda_3} - (R_0 - R_п)\, \beta \left(\frac{\lambda_2}{\lambda_3} \alpha_2 - l - \frac{\alpha_1}{\beta \lambda_3} \right)}{\alpha_1 + \alpha_2 + \xi},$$

after which we obtain the ordinary differential equation for finding ξ, in which the variables are divided. The solution leads to the formula

$$\xi = -(\alpha_1 + \alpha_2) + $$
$$+ \sqrt{(\alpha_1 + \alpha_2)^2 + \frac{2\lambda_3 t}{Q} \left[T_в - T_0 + R_п \frac{\alpha_1 + \alpha_2}{\lambda_3} + (R - R_0) \frac{\alpha_1}{\lambda_3} + \right.}\rightarrow$$
$$\rightarrow \left. + (R_0 - R_п)\, \beta \left(l + \frac{\alpha_1}{\beta \lambda_3} - \frac{\lambda_2}{\lambda_3} \alpha_2 \right) \right], \tag{16}$$

where $Q = q_0 \gamma_3 \frac{w}{1 + w}$.

With this formula one can compute the depth of thaw of soil covered by any vegetation and also predict the changes in depth of thaw resulting from changes or destruction of the vegetation cover. For computing the temperature fields and depths of thaw it is necessary to know k_1, k_2 and R, R_0 and R_π which are usually measured only in special investigations, and therefore mass calculations using the formulae obtained present some difficulties at the present time. However, with the formulae we have derived one can analyse the physical process of heat exchange between the atmosphere and the soil in the presence of a vegetation cover.

Let us analyse the changes in temperature of the layers under consideration and the depth of thaw in relationship to the properties of the vegetation cover and the soil, conditions of radiation exchange, and heat exchange in the air and in the vegetation cover. Let us also consider the temperature profile. For these calculations with formulae (12) and (13) it was assumed that $T_B = 15^o$, $T_0 = 0$, $\lambda_2 = 0.021$ kcal/m · hour · degree, $\lambda_3 = 1.0$ kcal/m · · hour · degree, $\kappa_0 = 0.067$ m^2/hour, $\xi = 1.0$ m, $R = 60$ kcal/m^2 · hour, $k_1 = 300$ m/hour, $k_2 = 30$ m/hour. The other parameters were variable and therefore their vlaues will be presented below.

Figure 2 shows the distribution of temperatures in the air and in the grass stand. Section a gives the variation in temperature distribution with the growth and development of leaves or for varying degrees of interlocking of the canopy in the upper parts. It is assumed that in this case the quantity of heat reaching the soil surface remains constant ($R_\pi = 6$ kcal/m^2 · hour), and the quantity of heat absorbed in the upper layer of the leaves, $R - R_0$, takes on the values of 0, 6, 20, 40, 54 kcal/m^2 · hour (in Figure 2, curves 1, 2, 3, 4 and 5 respectively). Since the total quantity of heat R remains constant, an increase in $R - R_0$ leads to a corresponding decrease in heat absorbed inside the grass cover $R_0 - R_\pi$, i.e., there is a redistribution of heat absorbed by various parts of the grass stand. In the absence of horizontal leaves (curve 1 in section a) when all of the heat (with the exception of R_π) is absorbed inside the grass stand the temperature maximum lies in the lower part of the grass stand near the ground surface. With increase in interlocking of the canopy more and more heat is absorbed in the upper layer, with the result that the maximum is shifted upwards and with almost complete interlocking it coincides with the grass surface (curve 5 in section a). Here there is a decrease in the temperature of the ground surface by 3.5^{o}C although R_π remains constant and conditions for turbulent heat exchange do not change.

A comparison of Figures 1 and 2 shows full agreement in the types of temperature distributions. Curve 1, section a, corresponds to temperature distribution in a grass stand and curve 5, section a, corresponds to that of a flowering plant stand. Thus the position of the temperature maximum is determined by the absorption of radiation heat in the vegetation cover. On grass meadows the temperature maximum must always be close to the ground surface whereas in a flowering plant stand, where there are well developed horizontal leaves, it is closer to the surface of the stand. With the same total amount of incoming heat R and heat reaching the soil surface R_π, the temperature of the ground surface in a grass stand, under otherwise equal conditions, is much higher than in a mixed stand, particularly in flowering plant stands.

Section *c*, Figure 2 shows changes in temperature on a grass meadow in relation to the density of the stand. With change in density there is a change in the quantity of heat reaching the ground surface R_π. With decrease in grass density, if the total quantity of heat remains unchanged, R_π increases owing to the decrease in absorption within the grass stand.

From the figure it can be seen that a change in R_π affects only the temperature of the ground surface and the adjacent thin layer of air (in our case 10 cm). In Figure 2, curves 1, 2 and 3 (section *b*) were plotted for the values of R_π, equal to 0, 6 and 20 kcal/m^2 · hour respectively. An increase in R_π from 0 to 20 kcal/m^2 · hour under the conditions selected results in an increase in temperature of the ground surface by more than 5°C. The surface temperature on a meadow with a sparse grass stand can be quite high, even higher than on bare ground (Figure 2, section *c*). Consequently, in summer the temperature of the ground surface under a vegetation cover is not always lower that on bare ground as indicated in many reports. Evidently such cases in nature are not rare.

In a shrub stand the temperature distribution is the same as in a flowering plant stand but the absolute temperature values are lower. In a shrub stand an important role in heat exchange is played by the shading of the ground surface (Figure 3, section *a*, where curves 1, 2 and 3 are plotted for a $R - R_0$ value of 60, 40, 0 kcal/m^2 · hour respectively). Dense leaves, particularly in the upper layer, retain a large quantity of heat and therefore the temperature maximum is in the upper part of the shrub stand, whereas the soil surface remains cool. In our case, when the ground surface is fully shaded its temperature was only 9.8°C (Figure 3, section *a*, curve 1) and even when 1/3 of the total incoming heat passes through the canopy and is absorbed inside the shrub stand and on the ground surface the temperature of the ground surface remains below that of the air temperature at a height of 2 m and is substantially lower that that in a meadow (Figure 3, section *a*, curve 2). Only in the absence of leaves in the spring and fall, when practically all of the heat is accumulated on the ground surface and the turbulent flow of heat in the air is lower than in an open place, does the temperature of the ground surface greatly increase and may be higher than on a bare surface (Figure 3, section *a*, curve 3). Such a situation should be observed and is in fact observed in small openings among shrubs.

Figure 3 (section *b*) shows a change in temperature on increasing the height of the shrub stand as well as any other vegetation; curves 1, 2 and 3 correspond to heights of 0.3, 1.0 and 1.5 m. In a higher shrub stand, under otherwise equal conditions, the temperature of the ground surface and of the air is always lower than in a low shrub stand.

An extensive tree stand differs from shrub cover in being much higher and having separate crowns. In a forest, heat is absorbed in summer by the crowns and by the soil surface; the trunks of trees absorb practically no heat. The relationship between R_π and $R - R_0$ is determined by the interlocking of crowns. Figure 4 shows the temperature field in a forest with varying degrees of crown interlocking. Curves 1, 2 and 3 correspond to R_π values of 6, 20 and 40 kcal/m^2 · hour; curve 4 was plotted for an open plot.

Dense tree crowns, absorbing almost all of the incoming heat, warm up the air in the upper part of the forest. The temperature in this case decreases towards the ground surface (Figure 4, curve 1). With a decrease in interlocking of crowns the air temperature inside the forest stand and ground surface increases and exceed the temperature of the crowns and air above the forest. At some relationship of R_π and $R - R_0$ one may observe in a forest an isothermic distribution of temperatures even in a strong wind. For example, when R_π = 15.4 kcal/m^2 · hour, $R - R_0$ = 44.6 kcal/m^2 · hour, the air temperature above the forest is 15°C, and the values of other parameters are those adopted here, the temperature between the ground surface and crown tops would be constant at 15.4°C

It follows from formula (13) that an isothermal state is possible when the conditions

$$T_в - T_0 - R_п \frac{\xi}{\lambda_3} + (R - R_0)\frac{\alpha_1}{\lambda_3} = 0, \qquad (17)$$
$$и \ R_0 - R_п = 0,$$

are fulfilled, i.e., there is practically no heat absorption in a forest below the crowns.

In a sparse forest with a relatively small crown area a considerable amount of heat may reach the ground surface. In this case the temperature of the ground surface may be higher than in open areas (Figure 4, curves 3 and 4).

Thus, the temperature in a forest depends on the type of tree stand. Sparse light-coloured pine forests, widely distributed throughout the permafrost zone, should have a higher ground surface temperature than larch and particularly spruce forests.

We have considered the influence of vegetation from the point of view of differences in heat absorption. Vegetation however also reduces the coefficient of turbulent heat exchange k_2. Calculations using the formulae derived and analysis of these calculations show that, for example, in a grass stand the lower the value of k_2 the higher the temperature of the ground surface and air. Air temperature in a dense grass stand can reach $30^{\circ}C$ at an ambient temperature of $15^{\circ}C$ and k_2 = 3.8 m/hour.

As can be seen from the figures given, inside the vegetation cover, particularly close to the ground surface, there is an inverse temperature distribution. Such a stable thermophysical state is characterized by a low turbulent exchange coefficient. Therefore, the intensity of heat exchange between the ground surface and the air inside the vegetation cover should be small.

A change in the turbulent exchange coefficient in free air has an insignificant influence on the temperature field in the vegetation cover. As on a plot without vegetation, the greater k_1, the lower the air temperature above the vegetation cover. In any reasonable range of variation in k_1 the distribution of temperatures by height remains the same.

The temperature in all three media clearly depends on the depth of thaw ξ, and consequently on time, even when all other relevant parameters are constant. There is an inversely proportional relationship between temperature and depth of thaw. Consequently, under otherwise equal conditions, in the spring and at the beginning of summer the temperature of the ground surface under any vegetation cover will be lower than at the end of summer and in the fall. With the values that we have used in the formulae at a thaw depth of 0.5 m the temperature of the ground surface in a grass meadow will be $12.4^{\circ}C$, and $20.5^{\circ}C$ at a thaw depth of 2.0 m. It is evident that there is a considerable difference in the temperatures, which indicates the important influence of the depth of thaw on the temperature regime of the vegetation cover.

Formula (16) indicates inversely that a vegetation cover has a great influence on the depth of thaw of the permafrost.

The depth of thaw has the same relationship to the type and properties of the vegetation cover as the temperature of the ground surface. Therefore everything that has been said about the temperature of the ground surface refers also to the depth of thaw.

As the amount of heat absorbed by the upper layer of grass increases (Figure 2, section *a*) the depth of thaw decreases. For the condition of curve 1, Figure 2 (section *a*) the calculated depth of thaw after 2000 hours is 1.79 m, and for the condition of curve 5 it decreases to 1.54 m, i.e., by 14%. The density of the vegetation stand has a particularly strong influence on the depth of thaw since it determines the quantity of heat reaching the ground surface. Whereas in a dense grass cover (curve 1, Figure 2, section *b*) after 2000 hours the depth of thaw is 1.68 m, in a sparse stand (curve 3, Figure 2, section *b*) the depth of thaw reaches 2.02 m, and on a plot free of vegetation (Figure 2, section *c*) the depth is 1.94 m.

Under a shrub stand the depth of thaw is as a rule less than on a meadow. Calculations for the conditions shown in Figure 3 (section *a*) give the following values for a 2000 hour period: 1.26, 1.52 and 2.13 m. If the shrub stand is without leaves the depth of thaw is greater than on the meadow. An increase in the height of the shrub stand (Figure 3, section *b*) in our case results in a decrease in the depth of thaw from 1.63 m at $l = 0.3$ to 1.49 m at $l = 1.5$ m.

In a forest, as in a shrub stand, the depth of thaw can vary over a wide range depending on the interlocking of the canopy and the density of the stand. Under otherwise equal conditions for R_{π} values of 6, 20 and 40 kcal/m^2 · hour (Figure 4), the calculated depth of thaw will be 1.35, 1.80 and 2.22 m respectively over a period of 2000 hours. The presence of any vegetation cover in most cases results in a decrease in the depth of thaw as compared with a bare plot. However, there can be exceptions to this rule, for example, deep thawing may occur in a sparse forest.

The calculated values given above give only a comparative order of values and do not in any way pretend to encompass all possible cases when several parameters may vary simultaneously. In each specific case it is better to use the appropriate formulae.

Up to the present time there has been no proper statement or theoretical solution of the problems of heat exchange between a surface covered by vegetation and the air. and the influence of a vegetation cover on the depth of thaw. The available experimental data permit only a schematic approximation of taking into account some features of heat exchange under the conditions mentioned above, for example where the heat absorption inside a vegetation cover is constant with respect to height, or where there is an evident delineation between the air and the vegetation cover.

In order to carry out calculations with the formulae suggested in this paper a great deal of attention should be devoted to studying the processes of transpiration of various plant associations, the absorption of radiation heat on the surface and inside a vegetation cover and turbulent heat exchange within the vegetation cover.

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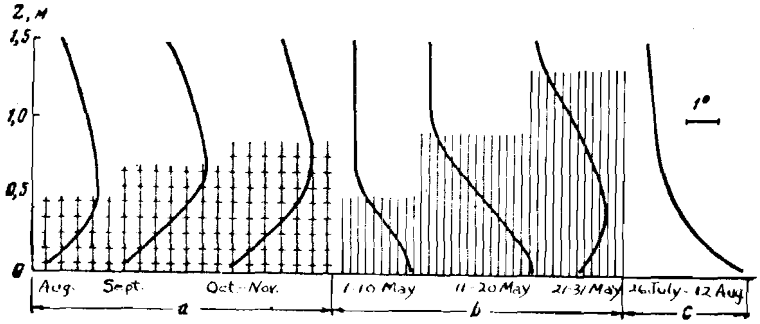

Fig. 1

Distribution of midday air temperature at various times of the year (after Geiger)
a - on flower bed; *b* - on winter rye field;
c - on mowed field

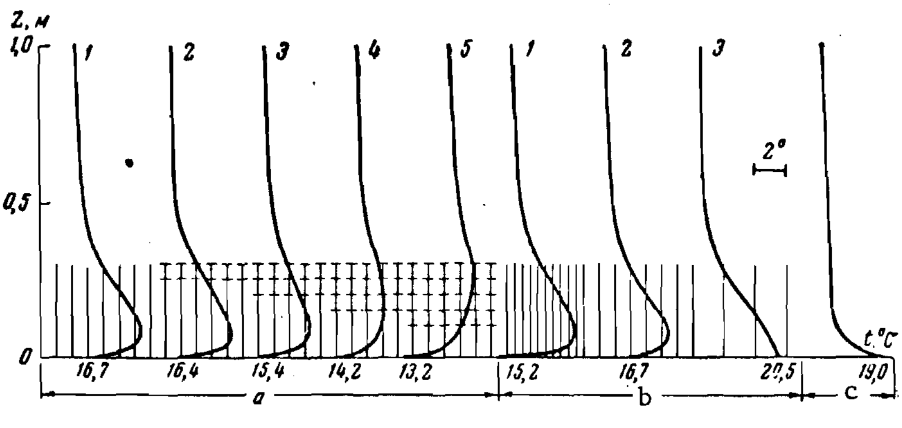

Fig. 2

Distribution of air temperature in grass stand

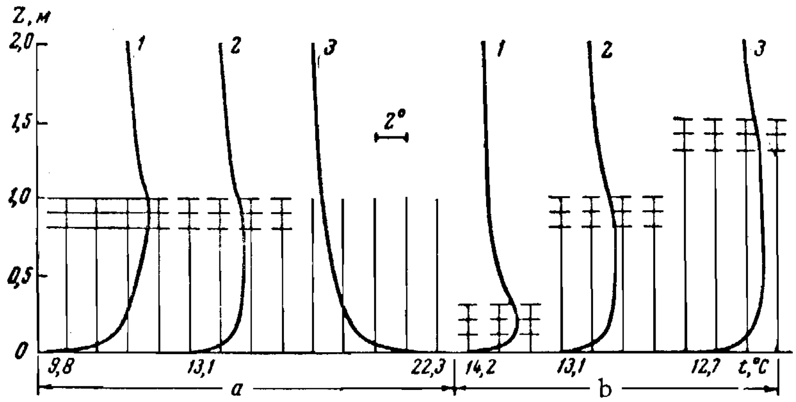

Fig. 3

Distribution of air temperature in air
and in shrub stand

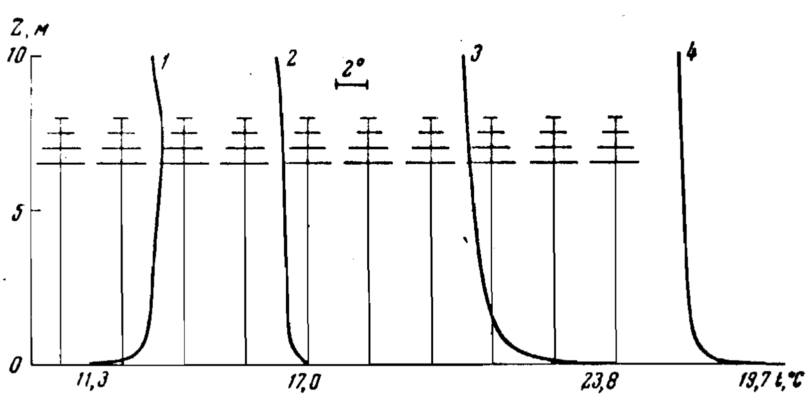

Fig. 4

Distribution of temperature in forest as related
to interlocking of canopy and on open plot