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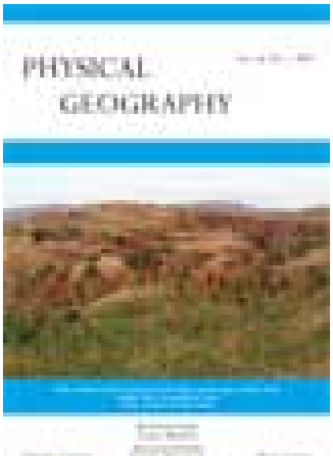
O. A. Anisimov
State hydrological institute, St.Peter…



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CHANGING CLIMATE AND PERMAFROST DISTRIBUTION IN THE SOVIET ARCTIC

Oleg A. Anisimov [a]

[a] State Hydrological Institute , 23 Second Line Leningrad 199053, USSR

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CHANGING CLIMATE AND PERMAFROST DISTRIBUTION IN THE SOVIET ARCTIC

Oleg A. Anisimov
State Hydrological Institute
23 Second Line
Leningrad 199053, USSR

Abstract: Anticipated global warming during the next century will produce many environmental changes, including widespread thawing of permafrost in the northern hemisphere. A nonstationary model of heat and water transport in a stratified medium was used in conjunction with results from a climate-change model to estimate the severity of permafrost degradation. Results suggest that the area of continuous permafrost in the USSR may be reduced by 15 to 20% over a 50-year period.

INTRODUCTION

The process of global warming induced by increasing concentrations of CO_2 and other trace gases raises many problems, among them changes in the temperature and distribution of permafrost. To address this problem, accurate estimates must be made of the sign and magnitude of changes in the permafrost, as well as in the timing of any increases of permafrost temperature. This paper takes a step in this direction through use of a one-dimensional nonstationary model of heat and water transport in a stratified medium. The model's structure, its main equations, a data base used for input parameters, and the main results form the basis of this discussion.

Among many remarkable results in the long history of investigations in the arctic regions, the most important for our purposes is the study of feedbacks between climate and permafrost, which allow application of certain empirically proven relationships between the corresponding parameters to the forecasting of permafrost distribution (Nelson, 1986; Nelson and Outcalt, 1987). From the view of these studies, parameters such as averaged annual air temperature sums, duration of warm and cold periods, and snow thickness may be applied to obtain solutions, provided that changes in these parameters are considered to result from minor deviations of the climate system from the quasi-stationary state. In such cases, it is possible to estimate the future configuration of the permafrost boundaries and the depth of seasonal thawing on the basis of scenarios of climate change.

At the same time, several principal arguments against the accuracy of such forecasts may be adduced. The methods cited above fail to take the timing of permafrost heating into account. Contemporary permafrost is known to be out of balance with the climate; the permafrost temperature profile in the upper layer

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(approximately the upper 100 m) has a significant negative curvature that reflects the permafrost's reaction to global warming (Lachenbruch and Marshall, 1986; Lachenbruch et al., 1988). Thus, such semi-empirical methods allow estimation of permafrost distribution only in the upper layer of the substrate, with thermal conditions at greater depths remaining unknown.

A second problem is that significant joint changes of air temperature, precipitation, and snow cover will cause a response in the permafrost that cannot be described by such relatively simple methods. Non-linearity appears in a wide range of the predictors' variation, and actual relationships between the climatic and permafrost parameters become extremely complicated. Nor do such methods provide precise translations of observed air temperatures to those at the soil surface, which play the main role in the formation of the permafrost's thermal regime. To solve this problem effectively, it is necessary to introduce many more parameters and, in some appropriate form, to take into account heat transport in the snow and in the upper layer of the soil.

When investigating permafrost distribution, it is necessary to recognize that soil temperature at the depth of zero annual amplitude differs from the mean annual soil surface temperature by as much as 1.0 to 1.5°C (Pavlov, 1979). Because of this temperature shift, permafrost may remain at sites where the mean annual temperature at the surface is slightly above zero. This temperature shift or "thermal offset" results from differences in the soil's thermal diffusivity in the frozen and thawed states. As the frozen soil's thermal diffusivity may be 1.3 to 1.5 times higher than in the thawed state, cooling of the soil in winter is more intensive than is its heating in summer, even with temperature gradients with the same absolute value.

The greater the difference in thermal diffusivity between the soil's frozen and thawed states, the higher the value of the thermal offset. Figure 1 shows thermal diffusivity for three types of soil as a function of water content, expressed as a percentage of the soil's dry weight. The greatest differences between the frozen and thawed values are in sandy soils, while the minimum differences occur in clay soils; in both cases the difference increases with the degree of soil wetness. If one accepts that the effect of the temperature shift always lowers the temperature at the level of zero annual amplitude, and assuming equal temperature increases, permafrost will degrade first in regions dominated by clay and loamy soil with low water content, while sandy soils with high moisture are relatively favorable for preservation of permafrost. This preliminary analysis suggests that to obtain correct physically-based estimates of permafrost evolution the annual cycle of soil wetness must be considered.

To estimate permafrost parameters and their probable future changes it is necessary to acknowledge that the permafrost thermal regime is everywhere influenced by local factors such as vegetation, surface exposure, albedo, surface roughness, and so on. In our model equations we include only the main parameters such as soil moisture, snow depth, and albedo; all others are assumed constant. Estimates suggest that changes in these parameters may produce the same effect on the permafrost thermal regime as a 1.5 to 2.0°C change in mean annual air temperature (Smith and Riseborough, 1983).

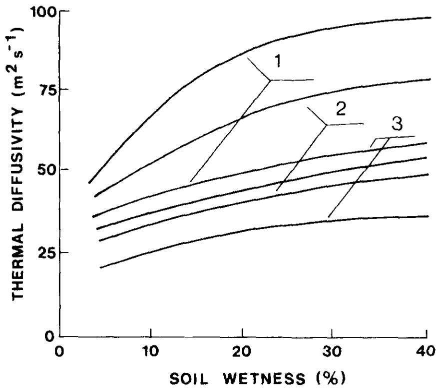

Fig. 1. Thermal diffusivity of soils used as input parameters. Upper curves correspond to frozen soil, lower curves to thawed soil: 1) sand; 2) loam; 3) clay.

STRUCTURE OF THE MODEL

The model of heat and water transport in soil, used to investigate the reaction of permafrost to climatic change, consists of two relatively independent blocks. In the first, which may be termed the hydrometeorological block, equilibrium soil-surface (or snow-surface) temperature, water content of the upper layer of the soil, and the components of the surface energy budget (radiation balance, evaporation, turbulent heat flux, heat flux to the soil) are calculated. Figure 2 depicts the model's structure schematically.

To calculate the equilibrium soil-surface temperature, an energy budget equation is used:

$$Q(1 - \alpha) - E_{ef}(T_s) - P(T_s) - B(T_s) - LE(T_s) = 0. \tag{1}$$

In this equation, Q is net solar radiation, α is surface albedo, $E_{ef}(T_s)$ is effective radiation of the surface, $P(T_s)$ is turbulent heat flux, $B(T_s)$ is heat flux to the soil, $LE(T_s)$ is heat of evaporation, and T_s is surface temperature. Surface albedo was considered to be 0.18 when snow is absent, and varied from 0.18 to 0.8 according to snow thickness.

To calculate the effective radiation of the surface, a semi-empirical formula was used (Budyko, 1974):

$$E_{ef}(T_s) = E^{\circ}_{ef}(T_s) \cdot (1 - cn) + 4\,\delta\,\sigma\,T^3(T_s - T) \tag{2}$$

where n is sky cloudiness (proportion of unity), c is an empirical constant (for 60-70° N latitude Budyko sets c = 0.79), δ is the surface irradiative ability (emissivity, taken to be 0.95), σ is the Stefan-Boltzmann constant (5.67×10^{-8} W m^{-2} °C), T is air temperature at the height of 2 m, and E°_{ef} is the effective radiation from a clear sky, defined as:

$$E^{\circ}_{ef} = \delta\,\sigma\,T^4\,(0.39 - 0.058\,e^{1/2}), \tag{3}$$

where e represents the air's humidity.

Turbulent heat flux is calculated by means of the parameterized formula:

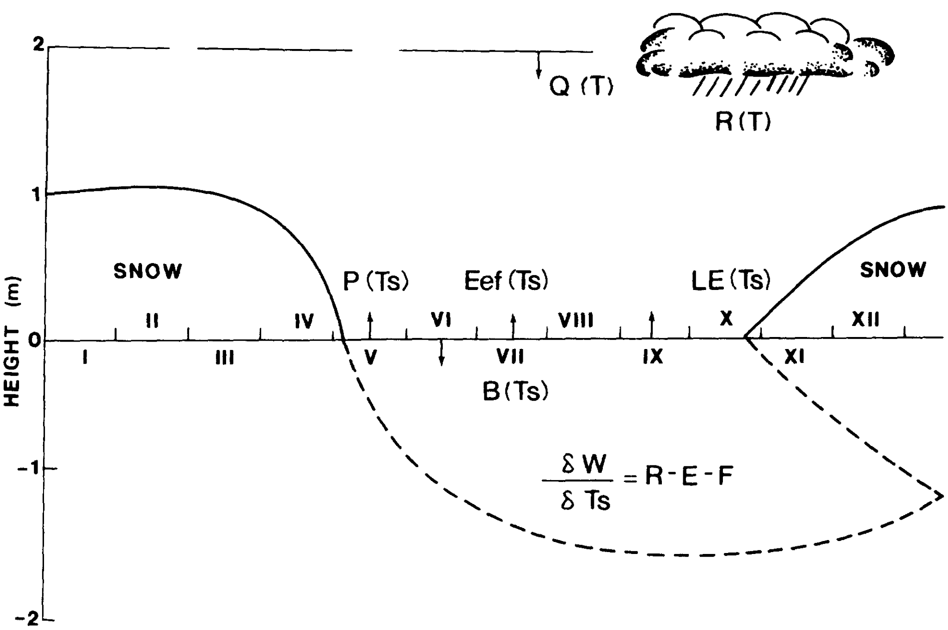

Fig. 2. Scheme of the hydrometeorological block of the model. Notations are explained in the text.

$$P(T_s) = \rho_a c_p D(T_s - T), \tag{4}$$

in which ρ_a, c_p, and D are air density (kg m^{-3}), specific heat capacity (J kg^{-1} °C^{-1}), and the integral coefficient of turbulent diffusion (6.3 X 10^{-3} m s^{-1}, Budyko, 1974), respectively.

The heat flux to the soil (or to the snow) was calculated via the temperature gradient, which in turn was calculated in the thermal block of the model by

$$B(T_s) = -\lambda\frac{\partial T_s}{\partial Z}, \tag{5}$$

where λ is the thermal conductivity (W m^{-1} °C^{-1}) of the soil (or snow). Snow thermal conductivity is assumed constant at a value of 0.23 W m^{-1} °C^{-2}, corresponding to snow density of 300 kg m^{-3}. The parameterized formula for soil thermal conductivity is given by

$$\lambda_i = k_i\,(0.001\,\rho_{dry} + 10W - 1.1) - 11.6W, \qquad i=1,2 \tag{6}$$

where ρ_{dry} is density (kg m^{-2}) of dry soil, W is the water content of the soil, expressed as a proportion of the soil's dry weight, and k_i is a coefficient defining different types of frozen (k_1) and thawed (k_2) soils (Table 1).

Evaporation and water content of the upper layer of the soil (approximately 1 m) was calculated by means of the complex method based on the solution of the water balance equation:

Table 1. Coefficient Values for Different Soil Types

	sand	loam	clay
k_1	1.95	1.75	1.60
k_2	1.75	1.60	1.50

$$\frac{\partial W}{\partial t} = R - E - f \tag{7}$$

where R is precipitation, E is evaporation, W is the soil's water content and f is surface runoff.

Equation (7) was supplemented with the correlation between runoff and precipitation:

$$F = \begin{cases} R \dfrac{W}{W_k} \sqrt{\mu^2 \left[1 - \left(1 - \dfrac{E_o}{R}\right)^2\right] + \left(1 - \dfrac{E_o}{R}\right)^2}, & R > E_o \\ \mu R \dfrac{W}{W_k}, & R \leq E_o \end{cases} \tag{7a}$$

where E_o designates evaporativity, defined by

$$E_o = \rho D (e_o - e). \tag{8}$$

Here e_o is saturated air humidity (mb), given by Magnus's formula:

$$e_o = 6.11 \exp((17.57\, T_s)/(241.9 + T_s)). \tag{9}$$

In this equation T_s designates surface temperature in °C.

It is also assumed that evaporation is equal to evaporativity E_o when soil water content is higher than the critical value W_k = 200 mm mon^{-1}, otherwise evaporation decreases linearly with water content according to the relation

$$E = \begin{cases} E_o, & W > W_k \\ E \dfrac{W}{W_k}, & W \leq W_k \end{cases} \tag{10}$$

In numerical experiments with the model it was assumed that soil water content during winter remained constant and that evaporation from the snow surface was negligible. Snow-cover thickness was calculated as a function of winter precipitation, with snow density increasing during the winter period.

In the algorithm used to describe the snowmelt period, spring begins when calculated surface temperature changes from negative to positive. The snow surface temperature throughout the snowmelt period is assumed constant at 0°C. An imbalance in the surface energy-budget equation, which appears in such a case, was equal to the heat expended on snowmelt. The end of the snowmelt period was defined as the moment when calculated snow-cover thickness reaches zero.

In contrast to traditional algorithms used to calculate evaporation, in which activity is presumed to a depth of 1 m, this model assumes water exchange to be possible only within the thawed layer, the thickness of which is calculated for every time step. Numerical experiments showed that in this case calculated values of evaporation during the first 1.5 to 2 months after the end of snowmelt are 10-15% less than that obtained using the more usual methods.

The temperature profile in the soil is calculated in the second block of the model using the heat-transport equation for a stratified medium (snow and/or a combination of the frozen and thawed soil layers) with variable thermophysical properties and moving phase-change boundaries. In the general case, this system is a four-layer medium consisting of the snow cover, two layers of frozen soil, and an intermediate unfrozen layer.

The heat-transport equation is given by

$$\rho c \frac{\partial T_z}{\partial t} = \frac{\partial}{\partial Z}\left(\lambda \frac{\partial T_z}{\partial Z}\right). \tag{11}$$

At the lower permafrost boundary (Z_*), temperature is considered constant at 0°C. Soil or snow surface temperature (T_s) is a function of time, and is calculated at every step in the hydrometeorological block. The initial vertical profile of soil temperature (T_z) is an input parameter of the model, and must be obtained from experimental data. Thus, the initial and boundary conditions for equation (11) may be formulated as:

$$T_z\Big|_{Z=0} = T_s(t); \qquad T_z\Big|_{Z=Z_*} = 0; \qquad T_z\Big|_{t=0} = T_z^0(Z)\,. \tag{12}$$

Additional equations can be used to describe the velocity of the phase boundaries, which is proportional to the difference between the heat fluxes from the frozen and thawed layers:

$$\frac{\partial T_j}{\partial t} = (-1)^{j+1}\frac{1}{W L}\left(\lambda_m \frac{\partial T_z}{\partial Z}\Big|_{Z=Z_j+0} - \lambda_f \frac{\partial T_z}{\partial Z}\Big|_{Z=Z_j-0}\right) \tag{13}$$

Here, λ_m and λ_f are the thermal conductivity of the thawed and frozen soil, respectively, and Z_j is the vertical coordinate of phase front j.

At the boundary between snow and soil, the equation of heat-flux continuity is:

$$\lambda_{sn} \frac{\partial T_z}{\partial Z}\Big|_{Z=Z_0+0} = \lambda_f \frac{\partial T_z}{\partial Z}\Big|_{Z_0=Z-0}\,, \tag{14}$$

where Z_0 is the coordinate of the soil surface, and λ_{sn} is the snow's thermal conductivity. Similar equations may be formulated on the boundaries between frozen and thawed soil. To solve equations (11) through (14), the modified method of registering the front to the grid point was used with a time step of ten days and a vertical step of 5 cm.

Input to the model of heat and water transport in a soil-atmosphere system therefore includes five meteorological parameters: net solar radiation, air temperature, air humidity, precipitation, and cloudiness. In the numerical experiments we used mean monthly climatic data for the territory of the USSR, averaged over a period of 30 years. Permafrost parameters and soil composition were taken from permafrost and soil maps of the USSR.

RESULTS

The model verification showed good agreement of the numerical results with the experimental data. The divergence between calculated and measured depths of seasonal thawing did not exceed 10 cm, an accuracy within 5%. A complete description of model verification is given in Anisomov (1989).

The model has been applied to the problem of the effect of global warming on permafrost. In this study we used the scenario of regional climatic change achieved through paleoreconstructions. This method allows us to reconstruct the regional distribution of the air temperature and precipitation of the winter and summer seasons for the previous warm epochs, which may be regarded as analogous to future climates (Budyko and Izrael, 1987).

Numerical experiments showed two types of permafrost reaction to global warming. When mean annual soil-surface temperature remained negative despite the warming, only quantitative changes of the permafrost thermal regime occurred. In this case the model allowed us to estimate the timing and vertical distribution of this process.

If the mean annual soil-surface temperature rose above zero as a result of warming, the model revealed the appearance and growth of a thawed layer of soil between the upper permafrost layer and the active layer above. Since this development is of great practical importance, it will be discussed below in detail.

Numerical modelling experiments extended over 50 years. For the first 10 years the model was focused on mean annual values of climatic parameters; during this step the contemporary quasi-stationary state of the hydrometeorological regime was achieved. Over the next 40 years air temperature and precipitation were increased linearly.

Because it is very difficult to describe the regional distribution of this process, we used a generalized criterion of climatic change: deviation of the mean global annual air temperature from its present-day value. In the latitudinal zone 55-70° N, a 2°C increase in mean annual air temperature results in a 7-9°C temperature increase in winter and 4-6°C in summer. Annual precipitation increases range from 50 to 100 mm (Budyko and Izrael, 1987).

Figure 3 shows regions within the USSR in which a 2°C global warming will cause degradation of permafrost. The contemporary location of the continuous permafrost border is shown with a solid line, and its location after the 2°C heating (approximately 2030) by a dashed line. In both cases, quasi-latitudinal zonation of the continuous permafrost is disturbed by the influence of many factors. In southern Yakutia, orographic factors contribute to the fragmentation, while in nearshore regions of western Siberia, despite a relatively mild climate, degradation of the permafrost is prevented by the specific annual precipitation cycle and the very long snowmelt period.

Figure 3 suggests that in 50 years the area occupied by continuous permafrost will be 15-20% smaller than at present. This zone, which now is characterized by a mean annual soil surface temperature ranging from -1°C to -5°C, will become the zone of permafrost degradation, and will develop a subsurface layer of thawed soil. The vertical movement of the thawed layer is expected to be 0.5 to 0.8 m yr^{-1} during

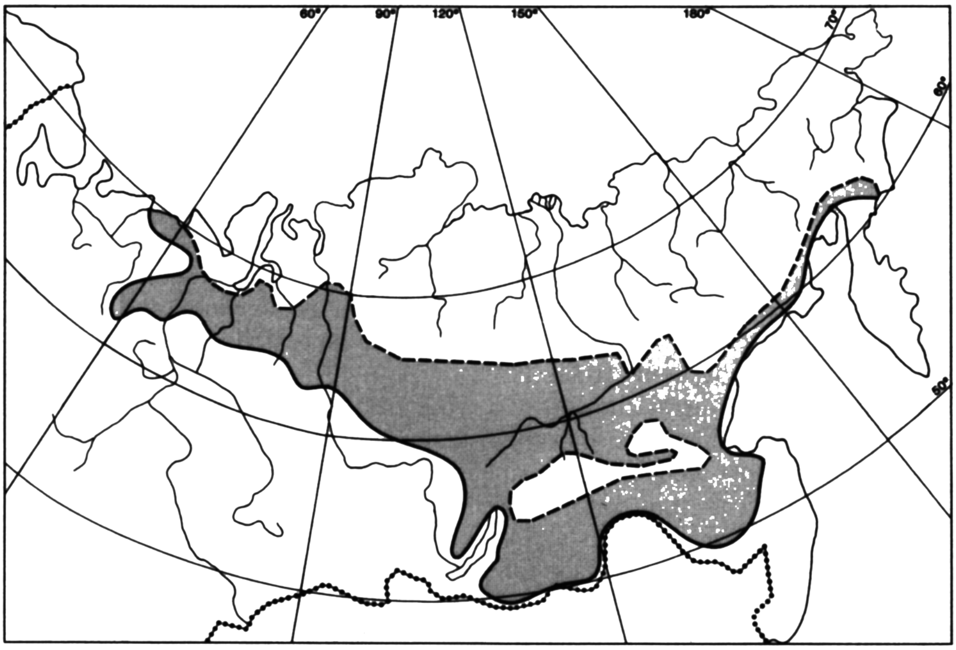

Fig. 3. Reduction of continuous permafrost in the USSR induced by 2°C global warming.

the first year after its appearance, and to decrease exponentially over time. The maximum depth of the thawed layer after the 50 years of the model run was 7 m. The speed of the (horizontal) permafrost boundary's retreat varies from 10-15 km yr^{-1} in western Siberia to 20-25 km yr^{-1} in eastern Siberia and the Far East.

CONCLUSIONS

This investigation was aimed at estimating the reaction of permafrost to changes of input parameters at the climatic scale. Many important problems relating to the sensitivity of permafrost to more localized factors therefore remain unsolved. Comprehensive study of these underdetermined processes must be based on analysis of weather records, and should become the goal of future investigations. The magnitude of changes in permafrost distribution suggested by this initial modelling effort demonstrates the urgency of the problem and the need for further effort along these lines.

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