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Mathematical Modeling of the Atmosphere–Cryolitic Zone Interaction

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Abstract—The main factors controlling the ground thermal regime characteristic of cold regions are analyzed with the use of a one-dimensional model of heat and moisture transport in the soil and its interaction with the atmosphere. The influence of these factors on the state of permafrost and the present-day climate as a whole is investigated on the basis of numerical experiments with a global model of general atmospheric circulation. It is shown that a decrease in the heat conductivity coefficient of the upper soil level, which can be interpreted as a layer of nondecomposed litter and moss, considerably increases the area occupied by permafrost. The introduction of the dependence of the heat conductivity coefficient on the phase state of water in the ground also increases the area occupied by permafrost and decreases the depth of the layer of its seasonal thawing in this territory. It is also established that the larger the relative amount of water which can be contained in the ground in a supercooled state is, the higher its temperature is, the lager the active layer depth is, and the smaller the area occupied by perennially frozen rocks is.

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1. INTRODUCTION

The cryosphere, i.e., the geographic shell of the Earth, which includes all objects related to water in the solid state, is one of the most important components of the climatic system. Its influence on near-surface fluxes of heat and moisture, cloudiness, precipitation, as well as atmospheric and oceanic circulations, produces various consequences and feedbacks which play a substantial role in the formation of climate, including its response to the anthropogenic impact. The cryolitic zone, as part of the cryosphere within the upper layer of the Earth's crust (in particular, permafrost), is one of the most sensitive components of the environment. The territory occupied by perennially frozen rocks currently amounts to about a quarter of the entire surface of the land [1] and about 80% (including regions of a deep seasonal freezing) of the land surface in Russia [2]. According to model estimates, it is high latitudes that will be most sensitive to global warming [3, 4], and observational data indicate that these predictions are now coming true [5]. Climatic changes suggest that the permafrost area will decrease [6] and the depth of the seasonal thawing of frozen ground will increase [7], as will the area of taliks, and shallow freezing will vanish in some areas [8]; in turn this will affect the thermal and hydrogeological processes on the land in these regions and those adjacent to them. The permafrost thawing and the temperature increase in upper soil layers will intensify the decomposition of organic matter and, as a consequence, increase the emission of carbon dioxide (or methane) from the soil into the atmosphere [9]; this in turn can increase the greenhouse effect. The consequences caused by permafrost degradation are most frequently estimated as unfavorable or even dangerous for humans [10, 11].

The aforesaid substantiates the necessity to investigate the interrelations between the cryolitic zone and other components of the climatic system in order to reveal the way they interact and to make it possible to predict the state of permafrost in conditions of a changing climate. For this purpose, one can use mathematical models, which are rather numerous now and are hierarchically ordered in their degree of complexity. The construction of different freezing indices [12] relating frozen ground parameters (for example, the temperature at a certain depth or at the depth of the seasonal layer of thawing) to atmospheric parameters (annual mean air temperature, sum of positive temperatures, and amount of precipitation), which are either taken from observations or are calculated from global climatic models, is one of the most simple methods. With the use of such semiempirical relations, one can determine the southern spreading boundary of perennially frozen rocks [13]. The geographic distinctions in properties of soil and vegetation can introduce a substantial uncertainty into the results of calculations with the use of these relations; however, because such results do not require large computational resources, the freezing indices can be considered as first approximations when the permafrost dynamics is estimated.

The construction of one-dimensional (along the vertical) mathematical models based on different balance relations is another popular and promising method. A great number of analytical solutions to the heat conduction equation [14] that make it possible to calculate the depth of the layer of seasonal thawing have been proposed, and numerical models with a simplified description of the thermodynamic processes happening in frozen grounds have been developed [15]. One special class includes models based on a numerical solution to complete equations of heat and moisture transport in soil (see, in particular, [16–18]) that take into account a large number of factors, such as the possibility of the existence of water in the supercooled state [19] and the low heat conductivity of mosses and litter forming the upper soil layer [20, 21]. This approach has gained wide acceptance in recent years, first of all because such models are of interest in their own right because they ensure a better understanding of heat and moisture exchange in the ground and, secondly, because this approach can be used as a tool for global diagnostics of the permafrost state, e.g., in an stand-alone regime. In this case atmospheric parameters obtained as a result of reanalysis or from calculations with the use of atmospheric general circulation models and coupled models of the atmosphere and ocean in the course of experiments on the reproduction of the present-day climate or scenario experiments on climate prediction for the 21st century are used (see, in particular, [22–26]).

The approaches described above take into account only the unidirectional action of the atmosphere on the state of permafrost, because its characteristics are calculated diagnostically. As a rule, the backward action of the cryosphere on the atmosphere is investigated with the use of relatively simple models that cannot stably maintain the existence of the model permafrost during a prolonged period of integration. There are only several studies devoted to the complete interactive modeling of the climatic system, which not only diagnoses the permafrost state but also describes its influence on the atmosphere. Thus, [27] studies the influence of accounting for the phase transitions of water in ground on the modeled present-day and predicted climates, whereas study [28] investigates the influence of organic matter in soil, whose properties differ substantially from those of the mineral skeleton in soil (high porosity, low heat conductivity, etc.), on the energy and hydrologic cycles in the soil column and in the global climatic system. Currently, such investigations are not numerous; however, they are highly needed as the main tool for correctly predicting the state of the cryolitic zone in the 21st century.

This study presents the results of interactive climate modeling with the use of the global climatic model developed at the Institute of Numerical Mathematics, Russian Academy of Sciences (INM RAS). The atmospheric part of this model is conjugate with the thermodynamic model of soil. Therefore, the reproduced heat and moisture regime of the model cryolitic zone adequately responds to changes in atmospheric characteristics and simultaneously affects the atmosphere. In particular, this work studies the factors responsible for the capability of the model to reproduce the thermal regime of perennially frozen soils and the backward influence of these factors on the modeled climatic system. Here, these factors include the fact that the heat conductivity of frozen ground is higher than unfrozen ground, the low heat conductivity of the moss–leach cover, and the existence of water in the supercooled state. We also consider some additional aspects related to the choice of the calculated domain depth for the solution of equations of heat and moisture transport in ground. This study is organized in the following way. First, we analyze the factors controlling the thermal regime of soils, which is characteristic of cold regions, with the use of the one-dimensional model of heat and moisture transport in soil and its interaction with the atmosphere (section 2). Then the influence of these factors on the permafrost state and the reproduction of the present-day climate as a whole is investigated on the basis of numerical experiments with the global model of atmospheric general circulation (section 3). The results are systematized and conclusions are made in section 4.

2. EXPERIMENTS WITH THE ONE-DIMENSIONAL MODEL OF THE SOIL–SNOW–VEGETATION–SURFACE ATMOSPHERIC LAYER SYSTEM

The results of validating the one-dimensional model of the active soil layer developed at the INM RAS are presented in [29, 30]. Data of multiyear regular instrumental observations at the Valdai Scientific Station and at different meteorological stations of Siberia were used for this purpose (data of the Valdai Station were used in the PILPS(2d) international experiment on comparing the parametrization schemes of the underlying surface [31]). It has been shown that this model is good at reproducing the interannual and interseasonal variabilities of the temperatures of the underlying surface and soil at different depths, as well as the snow cover height under different climatic conditions. Below, we give a brief description of this model.

2.1. Brief Description of the One-Dimensional Model

We use a widespread approach according to which the processes of heat and moisture transport in soil and snow cover, as well as the interactions of the atmosphere with the surfaces of soil, snow, and vegetation, are described on the basis of the one-dimensional (along the vertical) model of these processes, because vertical gradients of temperature and humidity in soil and snow are, in most cases, considerably larger than their horizontal gradients. In this model, soil and snow are multilayer media. The heat and moisture transport in soil is described by the equations of diffusion of these substances with nonzero sources in the right-side parts:

$$C\rho\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\lambda_T\frac{\partial T}{\partial z} + \rho(L_iF_i - L_vF_v), \quad (1)$$

$$\frac{\partial W}{\partial t} = \frac{\partial}{\partial z}\lambda_W\frac{\partial W}{\partial z} + \frac{\partial \gamma}{\partial z} - F_i - F_v - R_f - R_r, \quad (2)$$

$$\frac{\partial V}{\partial t} = \frac{\partial}{\partial z}\lambda_V\frac{\partial V}{\partial z} + F_v, \quad (3)$$

$$\frac{\partial I}{\partial t} = F_i. \quad (4)$$

Here, T is the ground temperature, W is the liquid water content, V is the water vapor content, I is the ice content, z is the vertical coordinate directed downward, C is the ground heat capacity, and ρ is the ground density. The terms ρL_iF_i and ρL_vF_v (L_i and L_v v are the specific heats of melting and evaporation, respectively), which are present in Eq. (1), characterize the rate of change in the heat content of the ground caused by water phase transitions. Equations (2)–(4) contain terms describing the rate of change in the moisture content due to water phase transitions (F_i and F_v), water infiltration under the action of gravity ($\frac{\partial \gamma}{\partial z}$, where γ is the hydraulic conductivity), water lateral runoff (R_f), and water suction by vegetable roots (R_r).

The coefficients of heat and moisture conductivities nonlinearly depend on the water content in the layer. The heat conductivity coefficient is calculated by the formula proposed in [32]:

$$\lambda_T = 418.7\max(\exp(-\log_{10}(-\psi) - 2.7), 0.00041). \quad (5)$$

According to [33], the soil moisture potential ψ in this equation, the diffusion coefficient for clean water λ_W, and the hydraulic conductivity γ in Eq. (2) are calculated as

$$\psi = \psi_{\max}\left(\frac{W_{\max}}{W}\right)^b, \quad \lambda_W = \lambda_{\max}\left(\frac{W}{W_{\max}}\right)^{b+2},$$

$$\gamma = \gamma_{\max}\left(\frac{W}{W_{\max}}\right)^{2b+3},$$

where b is the Klapp–Hornberger dimensionless parameter and $W_{\max}$ is the maximal liquid water content in the soil layer, which is determined from the relation

$$W_{\max} = \frac{\Pi - \rho_{gr}I/\rho_{ice}}{\rho_{gr}}\rho_{wat},$$

where ρ_{ice}, ρ_{wat}, and ρ_{gr} are the densities of ice, water, and dry ground, respectively, and Π is the soil porosity depending on the soil type. The quantities $\psi_{\max}$, $\lambda_{\max}$, and $\gamma_{\max}$ represent the soil moisture potential, the moisture conductivity coefficient, and the hydraulic conductivity in the state of saturation, respectively; they also depend on the soil type.

The soil heat capacity is calculated with allowance for the contents of liquid water and ice.

$$C = C_{dry} + C_{wat}W + C_{ice}I,$$

where C_{dry} is the heat capacity of dry ground, C_{wat} is the heat capacity of water, and C_{ice} is the heat capacity of ice. The effect of hysteresis during the freezing and thawing of water in soil is incorporated into this model. This effect implies that, although ice thaws at 0°C, liquid water can be contained in soil in the supercooled state and freezes gradually during the temperature lowering. The amount of water W_{nf}, which can exist in the unfrozen state at the temperature T, is calculated by the formula

$$W_{nf} = W_\infty + (W_0 - W_\infty)e^{T/T_0}, \quad (6)$$

where $T_0 = 3$°C, W_0 is the amount of unfrozen water at the temperature 0°C, W_∞ is the amount of water which remains unfrozen (formally) at $T \longrightarrow -\infty$. The parameters W_0 and W_∞ also depend on the soil type.

The absence of heat and moisture fluxes through the lower boundary (which is usually assumed to be 10 m in climatic models) serves as the boundary conditions for the equations of heat and moisture conductivity. The upper boundary condition is specified (depending on the season) at the upper boundary of snow cover or at the soil surface, depending on the temperature of the underlying surface, which is determined from the equation of its heat balance:

$$R_{short}(1-\alpha) + R_{long} - \sigma T_s^4 + H + LE + B = 0,$$

where R_{short} and R_{long} are the intensities of short-wave and long-wave radiations coming to the surface, respectively; α is the albedo of the surface; H and LE are the turbulent fluxes of sensible and latent heat, respectively; and B is the heat flux into soil (snow).

All of the experiments discussed below are performed with a comparatively simple scheme of the snow-cover parametrization [16], which calculates only the redistribution of heat through heat conductivity and disregards the phase transitions of water in snow.

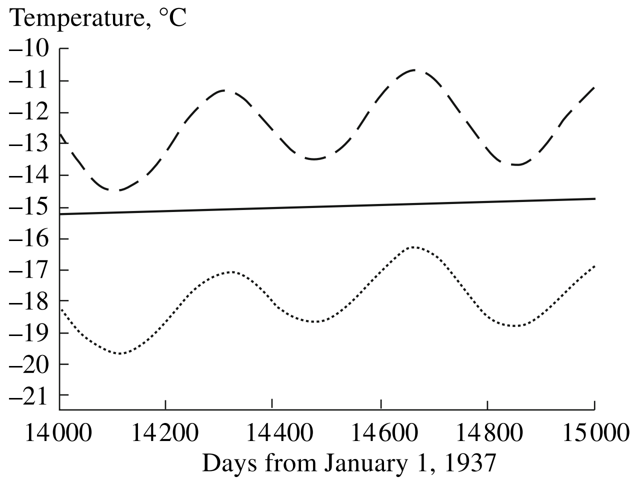

Fig. 1. Temporal trend of the soil temperature at a depth of 10 m. The solid, dashed, and dotted lines are for the CONTROL, 4Cond, and FrCond experiments, respectively.

2.2. Experiments

In order to reveal the influence of some factors (for example, the role of the heat conductivity coefficient) on the heat regime of grounds in regions with permafrost, we performed experiments on the continuous multiyear stand-alone numerical integration of the one-dimensional model. The data of regular measurements at the meteorological station in Yakutsk over 1937–1984 (temperature, humidity, and pressure of the surface air; wind velocity; amount of precipitation; and calculated intensities of short- and long-wave radiation) were used as the input information. The experiments include (1) a series of calculations for absolutely dry ground (under the assumption that there is no precipitation, snow cover, or water vapor in the soil or in the adjacent atmosphere) and (2) a series of calculations for real conditions with consideration for precipitation and the presence of snow cover. Methodological experiments with the absence of snow cover are performed in order to show that snow cover, by isolating the soil from its interaction with the atmosphere, noticeably upsets the symmetry between the cold and warm seasons of the year and can mask the sensitivity of the ground temperature profile to the factors under investigation. Similarly, the condition of the absence of moisture is set with the purpose of excluding water phase transitions in order to separate the effects of pure heat conductivity and phase transitions and to analyze their influence separately.

It is known (see, for example, [10]) that frozen ground has higher heat conductivity than ground with a positive temperature, and this difference can be fourfold. However, the widespread parametrization of the heat conductivity coefficient with the use of formula (5) disregards this jump-like dependence on temperature. In order to estimate its maximal effect, the following experiments for the dry ground conditions were conducted: (a) a CONTROL experiment, during which the heat conductivity coefficient was calculated by formula (5); (b) a 4Cond experiment, during which the heat conductivity coefficient determined by formula (5) was increased fourfold; and (c) a FrCond experiment, during which the heat conductivity coefficient was calculated by the formula

$$\lambda_T = \begin{cases} \lambda_T \text{ by formula (5) at } T \geq 0, \\ 4\lambda_T \text{ by formula (5) at } T < 0. \end{cases} \tag{7}$$

In addition, we performed experiment (d) with an increased (up to 8 cm) thickness of the moss–lichen cover. During this experiment, the heat conductivity coefficient for the upper 8 cm of ground was specified to be equal to the value 0.3 W/(m K) characteristic of vegetation. An analogous series of experiments were performed for the soil containing moisture, i.e., for the conditions actually observed in the nature.

2.3. Results

Figure 1 shows the interannual dynamics of the soil temperature at a depth of 10 m from the results of modeling with the use of dry soil.

We will first consider the influence of temperature-independent variations in heat conductivity (its uniform increase by a factor of 4) on the ground temperature and the active-layer depth. It can be seen from Fig. 1 that, at a depth of 10 m, not only the amplitude of variations increases (which would be expected) but the annual mean temperature also increases in the entire soil column (at an atmospheric-forcing level of 2 m). This can be explained by investigating the behavior of the heat-balance components at the soil surface. The balance of short-wave radiation is the same in both cases, and the latent-heat flux is zero, because there is no moisture in the soil or in the atmosphere. Figure 2 shows the turbulent flux of sensible heat and the heat fluxes into the soil and out of the soil (the latter has a positive sign) for summer (Figs. 2a, 2b) and winter (Figs. 2b, 2c) (in all of the three figures, the vertical scale is the same).

It follows from Fig. 2 that, in summer, the increased heat flux from the soil during the experiment with the increased heat conductivity coefficient is compensated by smaller heat losses for the turbulent exchange with the atmosphere; therefore, the atmospheric forcing in both cases will be approximately the same. In winter, the differences between the heat exchanges due to heat conductivity remain, whereas the turbulent exchange with the atmosphere is suppressed because of its stable stratification. As a result, the soil, whose heat conductivity coefficient is larger, more effectively cools the atmosphere and thus such

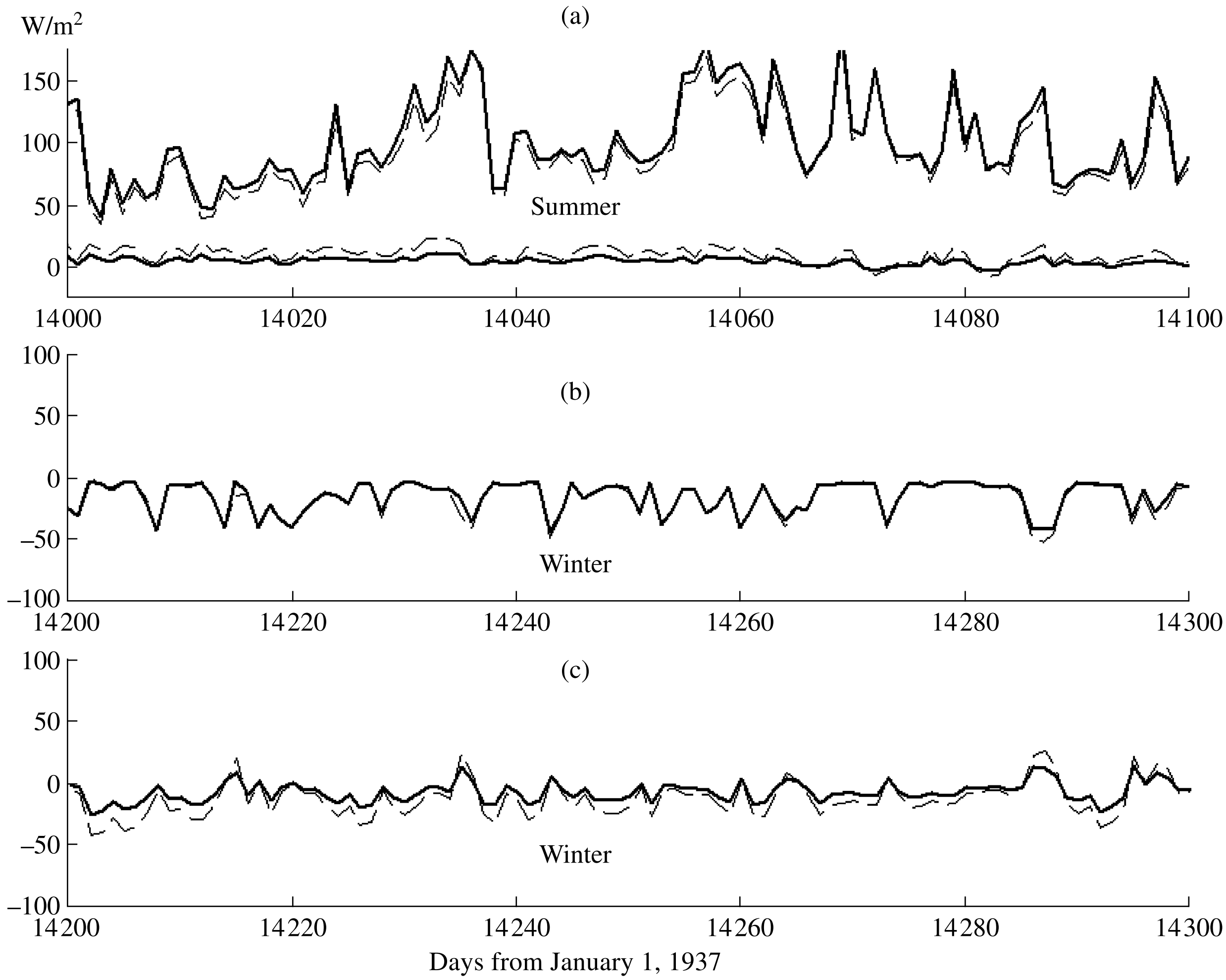

Fig. 2. (a, b) Sensible heat flux, and (a, c) heat flux from soil: (a) in summer and (b, c) in winter. The solid and dashed lines are for the CONTROL and 4Cond experiments, respectively.

soil is warmer than the soil which is more inert from the standpoint of heat conductivity.

The influence of temperature variations due to the mechanism described above on the active-layer depth (depth of the zero isotherm) is shown in Fig. 3. If this depth is about 0.9 m during the control experiment (recall that this idealized experiment does not take into account precipitation, i.e., it deals with the case of dry soil without winter thermal insulation by the snow cover), it increases to about 2.3 m when the ground heat conductivity increases by a factor of 4.

We will now consider the influence of a variable heat conductivity coefficient on the ground temperature (compared with the control experiment), i.e., the case when the heat conductivity coefficient is equal to its control value at a positive temperature and increases fourfold at a negative temperature. As is seen from Fig. 1, in this case the mean ground temperature becomes lower than even during the control experiment. This fact can be explained by examining Fig. 4, which shows the temperature trends for all three experiments at a depth of 2 m. As is seen from this figure, the temperature during the experiment with variable heat conductivity is close to the temper-

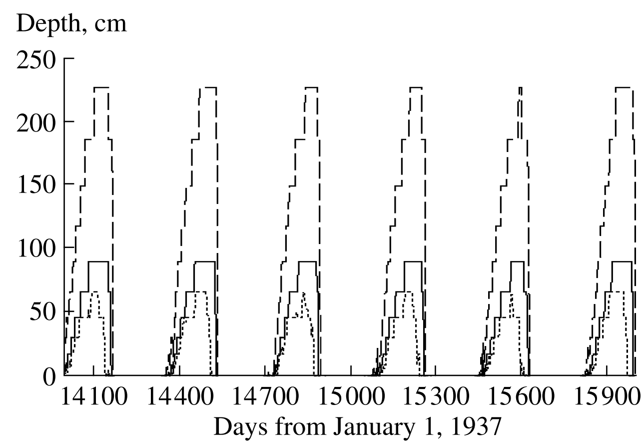

Fig. 3. Active layer depth. The solid, dashed, and dotted lines are for the CONTROL, 4Cond, and FrCond experiments, respectively.

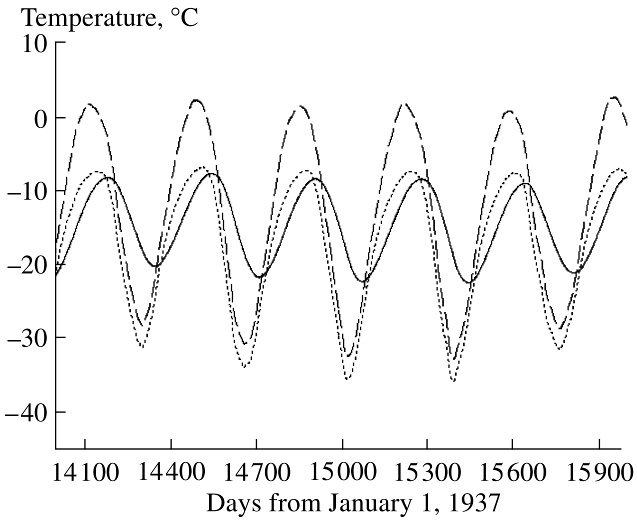

Fig. 4. Soil temperature at a depth of 2 m. The solid, dashed, and dotted lines are for the CONTROL, 4Cond, and FrCond experiments, respectively.

ature of the control experiment in summer because the temperature of upper soil layers is positive; in winter, the temperature is close to the temperature of the experiment with increased heat conductivity because the soil temperature is negative. As a consequence, the annual mean temperature during the last (third) experiment is lower than during the first and second experiments. Returning to Fig. 3, we can see that this lower annual mean temperature falls on the active-layer depth. This depth is about 0.7 m; i.e., it is smaller than during the first two experiments (0.9 and 2.3 m, respectively).

Since the ground heat conductivity depends on many factors, the formulation of the model heat conductivity coefficient also must depend on the factors taken into account in the model. The mechanism described above can be apparently regarded as fairly general. Thus, according to formula (5), the heat conductivity coefficient increases with increasing soil moisture. Consequently, the wetter the soil is, the warmer (on average) it will be (if the influence of water phase transitions is disregarded). In addition, it is widely known that the heat conductivity of mosses, litter, peat, and other organic matter in soil is low [34–36], as is the fact that the removal of such a thermal insulating layer from the soil surface increases the thickness of the layer of seasonal thawing in regions with permafrost [35, 37]. In order to elucidate whether the active layer thickness increase is related to the temperature increase in the entire soil column, we performed a MOSS numerical experiment analogous to the preceding experiments. During this experiment, the heat conductivity coefficient of the upper three soil layers (with a total thickness of 8 cm) was chosen to be equal to 0.3 W/(m K) (as an approximately mean

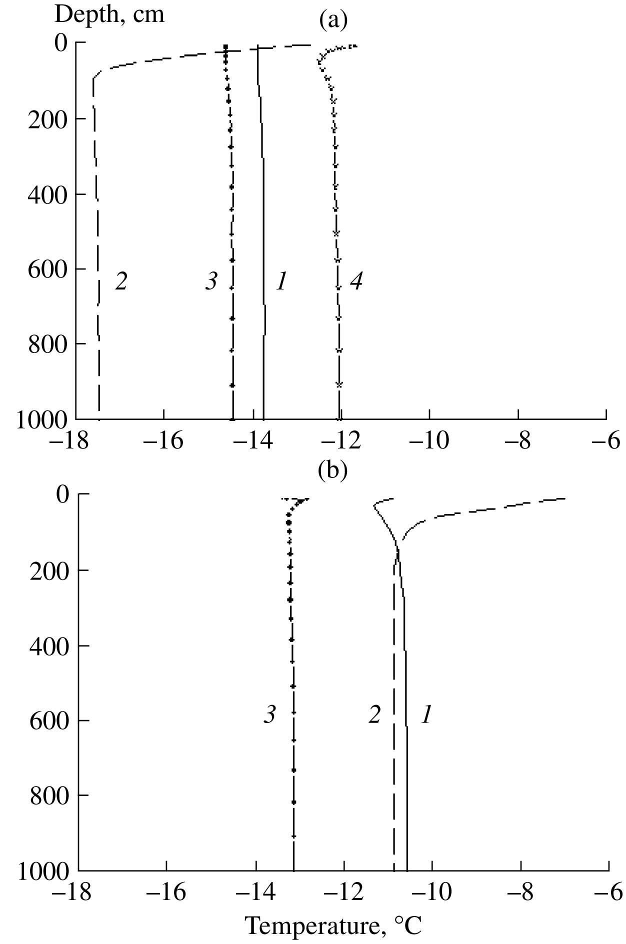

Fig. 5. Vertical profile of the annual mean ground temperature. (a) Dry ground: (*1*) CONTROL experiment, (*2*) FrCond experiment, (*3*) MOSS experiment, (*4*) Control experiment for wet soil (without snow cover). (b) Wet ground + snow cover: (*1*) CONTROL experiment, (*2*) FrCond experiment, (*3*) MOSS experiment.

value from the set of experimental data (see, for example, [10, 34, 38, 39])).

Figure 5a shows the annual mean (averaged over 48 years of integration) vertical temperature profiles of the ground obtained from the results of experiments with dry soil (meteorological forcing with zero precipitation and without snow cover) performed for the following conditions of the specification of the heat conductivity coefficient: (*1*) by formula (5), (*2*) by formula (7), and (*3*) by formula (5) with the heat conductivity coefficient equal to 0.3 W/(m K) for the upper 8-cm soil layer. The same figure presents the results of an experiment for wet soil, during which the heat conductivity coefficient was calculated by formula (7), whereas the soil moisture was calculated with the use of the model in the standard regime by Eq. (2) and was found to be equal to about 0.2 kg/kg. It is seen from Fig. 5a that the distinctions in the ground temperatures caused only by the difference between the heat conductivity coefficients can attain several degrees and, as a rule, a lower heat conductivity coefficient leads to a lower temperature in the soil column. Case (2), when the heat conductivity coefficient depends on the temperature sign (negative or positive) in the soil layer, is an exception. In this experiment (broken line), the equilibrium ground temperature is by 4°C lower than in the control experiment (continuous line). A decrease in the heat conductivity coefficient of the upper 8-cm soil layer (line with rhombs) lowers the equilibrium temperature by about 1°C. If water is present in soil and this increases the heat conductivity coefficient in accordance with law (5), the annual mean ground temperature increases by about 2°C (line with crosses).

The presence of snow cover in winter can somewhat mitigate these distinctions, because the heat-insulating properties of snow and the relatively small temperature gradient in the soil column decrease (in absolute value) the heat flux into the soil. The annual mean vertical profiles of the ground temperature constructed from the results of analogous experiments described above, but with allowance for the snow cover and actual precipitation, are presented in Fig. 5b. It follows from comparing the curve with crosses in Fig. 5a with the continuous line in Fig. 5b (these experiments differ only in the presence or absence of snow cover) that snow has a warming effect on the soil equal to about 1.5°C for the annual mean temperature in the soil column at depths below the active-layer base. Since the snow cover in Yakutsk is relatively thin throughout the entire winter, this estimate can be regarded as close to the minimal possible estimate.

Further, comparing the continuous and broken curves in Fig. 5b (CONTROL and FrCond experiments, respectively), it is possible to conclude that, in conditions of wet ground and in the presence of seasonal snow cover, the sensitivity to the method of calculating the heat conductivity coefficient (its increase by a factor of 4 at temperatures below 0°C) substantially decreases compared with dry ground (0.3 and 4°C, respectively). However, it is necessary to remember that this insignificant difference (0.3°C) relates only to the regime of modeling with specified meteorological conditions. In order to estimate the sensitivity of the thermodynamic regime of the ground to the uncertainties of calculating the heat conductivity coefficient depending on the ground-layer temperature, we should invoke the results of the complete three-dimensional modeling of the climate. It is also seen from Fig. 5b that a decrease in the thermal conductivity coefficient for the upper 8-cm layer (line with rhombs) decreases the ground temperature in depth by 2.5°C, i.e., by a larger value than in the case of dry ground.

3. EXPERIMENTS WITH THE GLOBAL MODEL OF ATMOSPHERIC GENERAL CIRCULATION

In this study we used the climatic model of atmospheric general circulation developed at the INM RAS and comprehensively described in [40]. The one-dimensional soil–vegetation–snow cover model described in the previous section is used as a module for the interaction between the atmosphere and the underlying surface and is included into the atmospheric model, i.e., when this module must describe properties of the underlying surface of the entire grid box of the model, a mosaic approach is taken in it. In each grid box, vegetation, snow cover, water surface, and bare soil occupy their own fractions of the area. The surface fluxes of heat and temperature are calculated separately for each surface type at the lower model level in the atmosphere. In addition, the area occupied by vegetation is, in turn, divided into fractions occupied by its different types (in all, 13 types of vegetation may exist), and the effects from vegetation, which are taken into account when calculating the atmosphere–underlying surface interaction (for example, resistance), are the weighted means over all vegetation types, which are assigned to a given grid box.

3.1. Description of Experiments

All of the experiments, the results of which are described below, are analogous to experiments performed within the framework of the AMIP international project on comparing atmospheric models. The specified monthly means of the ocean surface temperature and the sea ice distributions for 1979–1995 were used as external parameters. The fields of atmospheric characteristics, which were adapted to each other and obtained as a result of numerous repetitions of the 17-year cycle (1979–1995) of external conditions for the standard version of the model corresponding to the CONTROL experiment (see below), were assumed to be the initial conditions for the atmosphere. For the soil, the initial conditions for all experiments were as follows: southward of latitude 30° N, all necessary initial fields (temperature and contents of liquid water, ice, and water vapor in the soil) were also equal to the fields for the standard version, which were adapted to each other. To the north from the specified latitude, the soil column temperature for all grid boxes was assumed to be –16°C, and the entire soil moisture, whose total amount was also equal to the adjusted values, was regarded as ice. Further, the 17-year cycle of external conditions was repeated until the model climate reaches a quasi-stationary level (the state of the model permafrost, i.e., the most slowly changing components of the model, was controlled). From six to eight repetitions of this cycle were needed for different experiments, i.e., an additional integration of the model for each concrete experiment was performed for 100–150 years. The results of the last 17 years were averaged and were regarded as the model climate.

The experiments with the global model of atmospheric general circulation were done to reveal the influence of the factors which were investigated in the previous section for the (one-dimensional) case that was not interactive, on the modeled heat and moisture regimes of perennially frozen rocks in the case when the cryosphere is included into various feedbacks with other components of the climatic system. Thus, it is reasonable to study the influence of disturbances on the heat conductivity coefficient of the upper soil layer in regions abundant in mosses, on the geographic distribution of permafrost, and on the modeled climate. The same is also true of the heat conductivity coefficient of frozen soils compared with thawed ones. Additionally, the depth of the lower model level is one more parameter currently attracting the attention of researchers [41]. One of the experiments discussed below analyzes the role of this factor.

It is noted in [42] that there is an active boy formation of the territory and its overgrowth with mosses in the taiga zone of western Siberia. The same work reports that, in conditions of intense boy formation, the temperature of peat can additionally lower by 0.5°C over ten years due only to the surface overgrowing with mosses. Thus, if the existence of mosses ensures a lower temperature of the underlying ground, the intense boy formation can prove to be a favorable factor for the preservation of permafrost in Western Siberia. In connection with this, it is important to determine, at least by a simplified method, how the area occupied by permafrost will change if the territory now occupied by the taiga is swamped and overgrown with mosses.

The possibility of the existence of supercooled liquid water at a negative ground temperature is another important factor attracting the attention of researchers. Work [43] investigates the influence of unfreezing water within the framework of modeling the processes of heat and moisture transport in ground and shows that the presence of unfreezing water leads to a higher soil temperature; this difference can sometimes reach 9°C. This effect of gradual water freezing with a temperature decrease is implemented in the INM RAS model (see formula (6), section 2); however, the uncertainties associated with the geographic distributions of the parameters W_∞ and W_0 remain. Therefore, it would be reasonable to elucidate to what degree the modeled permafrost is sensitive to these uncertainties.

The aforesaid suggests the performance the following experiments: (i) the CONTROL experiment, when the soil heat conductivity coefficient is calculated by formula (5); (ii) the MOSS experiment, when the soil heat conductivity coefficient is assumed to be 0.3 W/(m K) for the upper 8-cm soil layer grid boxes, where the tundra vegetation type occupies a certain fraction of the area; (iii) the MOSS 60 m experiment, which has the conditions of experiment (2), but the lower boundary of the calculated domain is located at a depth of 60 m; (iv) the TAIGA experiment, which is analogous to experiment (3), but the heat conductivity of the upper soil levels is decreased not only in grid boxes with the tundra but also in grid boxes where evergreen coniferous forests are present; (v) the FrCond experiment, which is analogous to experiment (2), but the soil heat conductivity coefficient is calculated by formula (7); and (vi) the UnfrW experiment, which is analogous to experiment (2), but, in formula (6) for calculating the unfrozen water amount for all grid boxes, the parameters W_0 and W_∞ are assumed to have equal values for the clay-type soil, i.e., the maximum possible values ($W_\infty = 0.2$ kg/kg and $W_0 = 0.4$ kg/kg).

3.2. Results

The present-day distribution of permafrost according to observational data from [44] is shown in Fig. 6, and the mean August distribution of ice in soil from the results of the CONTROL experiment is presented in Fig. 7a. It can be seen that the model in this formulation does not "retain" permafrost, and throughout the whole year, ice is preserved in soil only in Arctic and subarctic regions. The mean August distribution of ice in soil from the results of the MOSS experiment is shown in Fig. 7b. It follows from a comparison of Figs. 7a and 7b that a decrease in the heat conductivity coefficient of a relatively thin upper soil layer in regions abundant in mosses has a decisive effect on the existence of the model permafrost. In turn, the presence of permafrost affects the climate. The difference between the climatic temperatures of surface air during the CONTROL and MOSS experiments averaged over three summer months is shown in Fig. 8a. In the regions where permafrost exists in the MOSS model version but is not reproduced during the CONTROL experiment, the surface temperature during the MOSS experiment is noticeably higher (by up to 5°C).

Two factors characteristic of frozen grounds are possibly responsible for this result. First, the permeation of thawing water in the period of spring snow thawing is very difficult, because most of it is spent on the runoff formation. Second, in summer, an active layer is developed in the perennially frozen ground and, as the moisture evaporates, its moisture content decreases because there is no possibility for water transportation from deeper soil layers like in the case when the ground is not frozen through its entire depth. All this leads to the fact that, in summer, the moisture of the upper soil level during the MOSS experiment is smaller than during the CONTROL experiment (Fig. 8b). Further, since the soil moisture during the MOSS experiment is smaller, the contribution of the sensible heat flux at the surface to its heat balance increases (compared with latent heat), and the surface tempera-

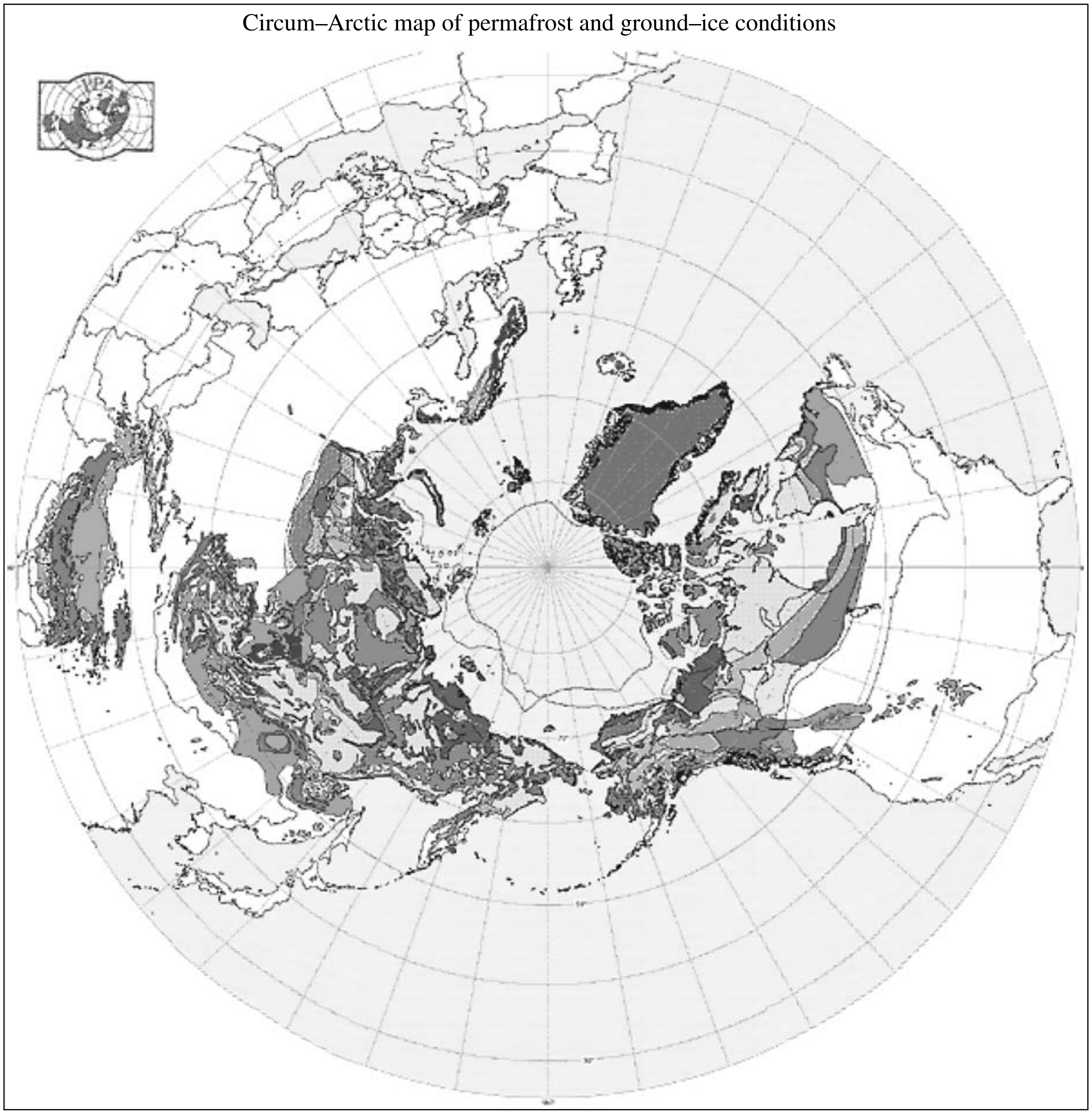

Fig. 6. The observed present-day geographic distribution of perennially frozen rocks [44].

ture becomes higher. In addition, the feedback arises between the evaporation decrease; the cloudiness decrease (not shown); and the incoming short-wave radiation increase, which in turn increases the surface temperature even to a greater degree.

It is interesting to note that similar regularities were detected in [2,], when the results of the FROST and NOFROST, i.e., the experiments with and without consideration for moisture phase transitions, were compared. In the regions where permafrost was reproduced in accordance with the FROST experiment (during the NOFROST experiment, water in the ground was always in the liquid phase), the modeled summer temperature of the surface was higher and the soil moisture smaller. The formulation of this model did not include a description of water permeation downward under the action of gravity; therefore, the redistribution of water took place due to diffusion alone. In the NOFROST experiment, the water transport from the underlying soil layers upward was overestimated and, as a consequence, the effect of water phase transitions of water was amplified. The results presented in this study indicate that, even if the water permeation downward under the action of gravity is taken into account, the conclusions about the influence of permafrost on the energy and hydrologic cycles at the land surface will remain qualitatively analogous to the conclusions made in [27].

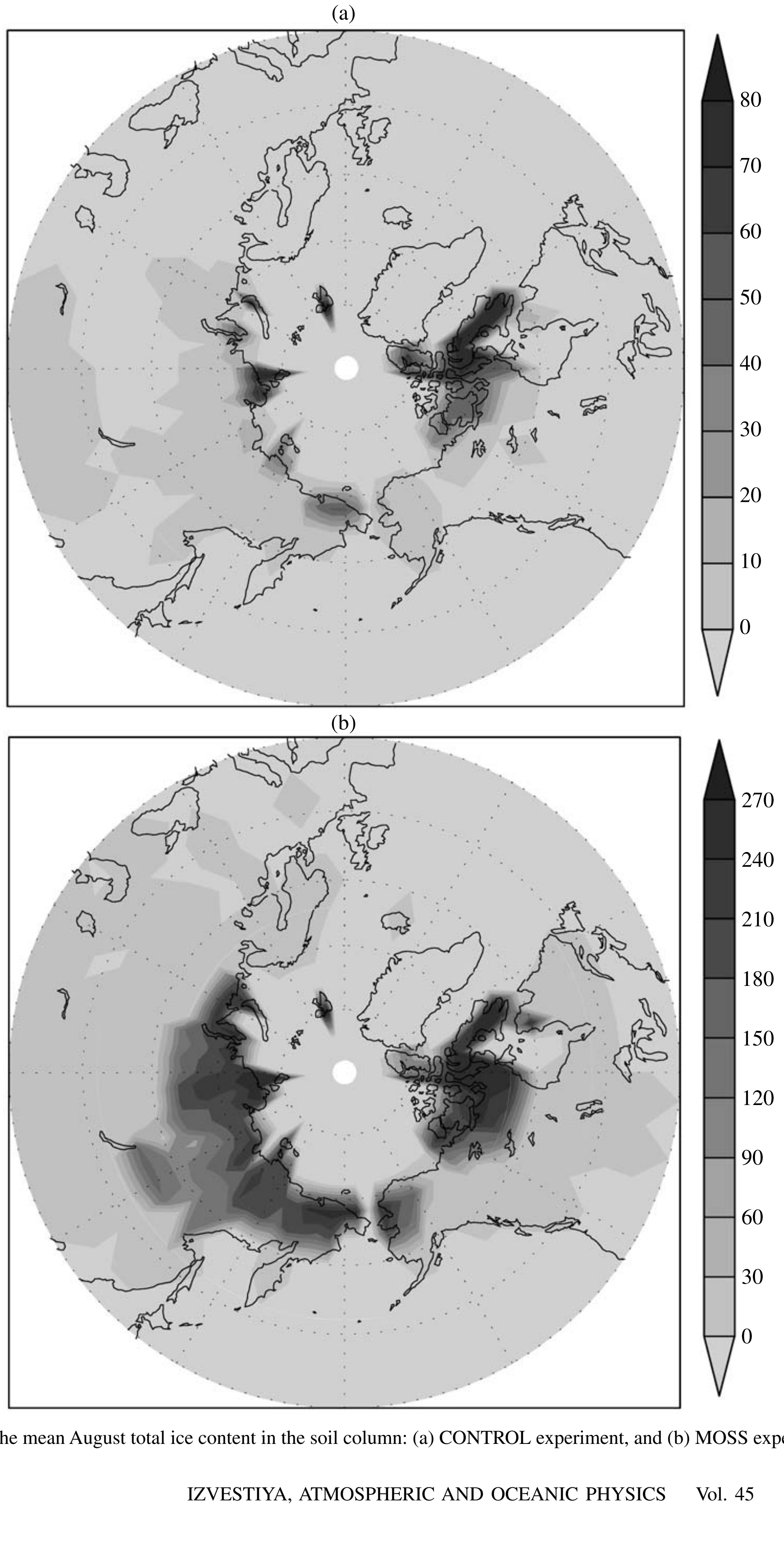

Fig. 7. The mean August total ice content in the soil column: (a) CONTROL experiment, and (b) MOSS experiment.

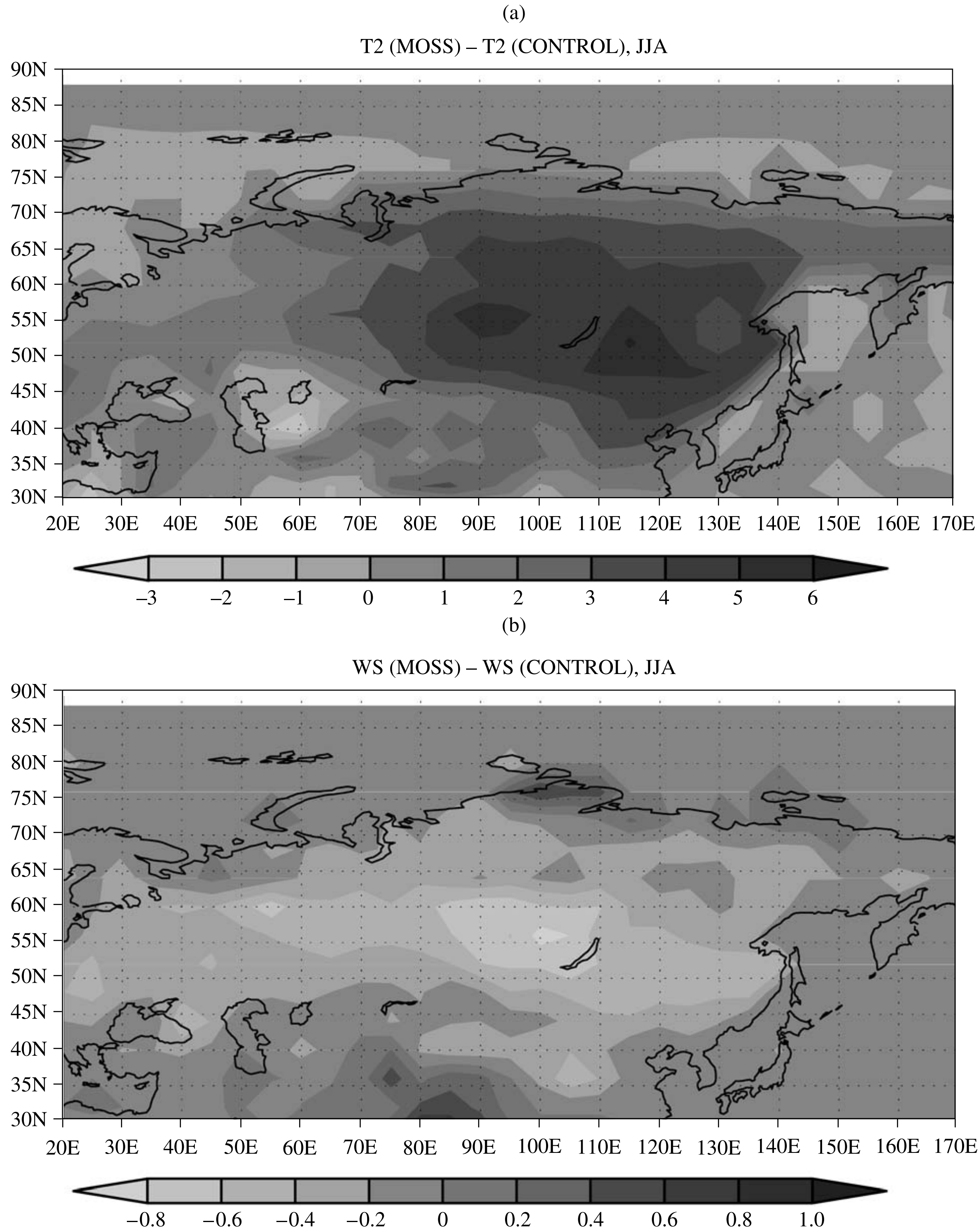

Fig. 8. (a) The difference between the climatic temperatures of surface air averaged over three summer months from the results of the MOSS and CONTROL experiments; (b) the same for the moisture content.

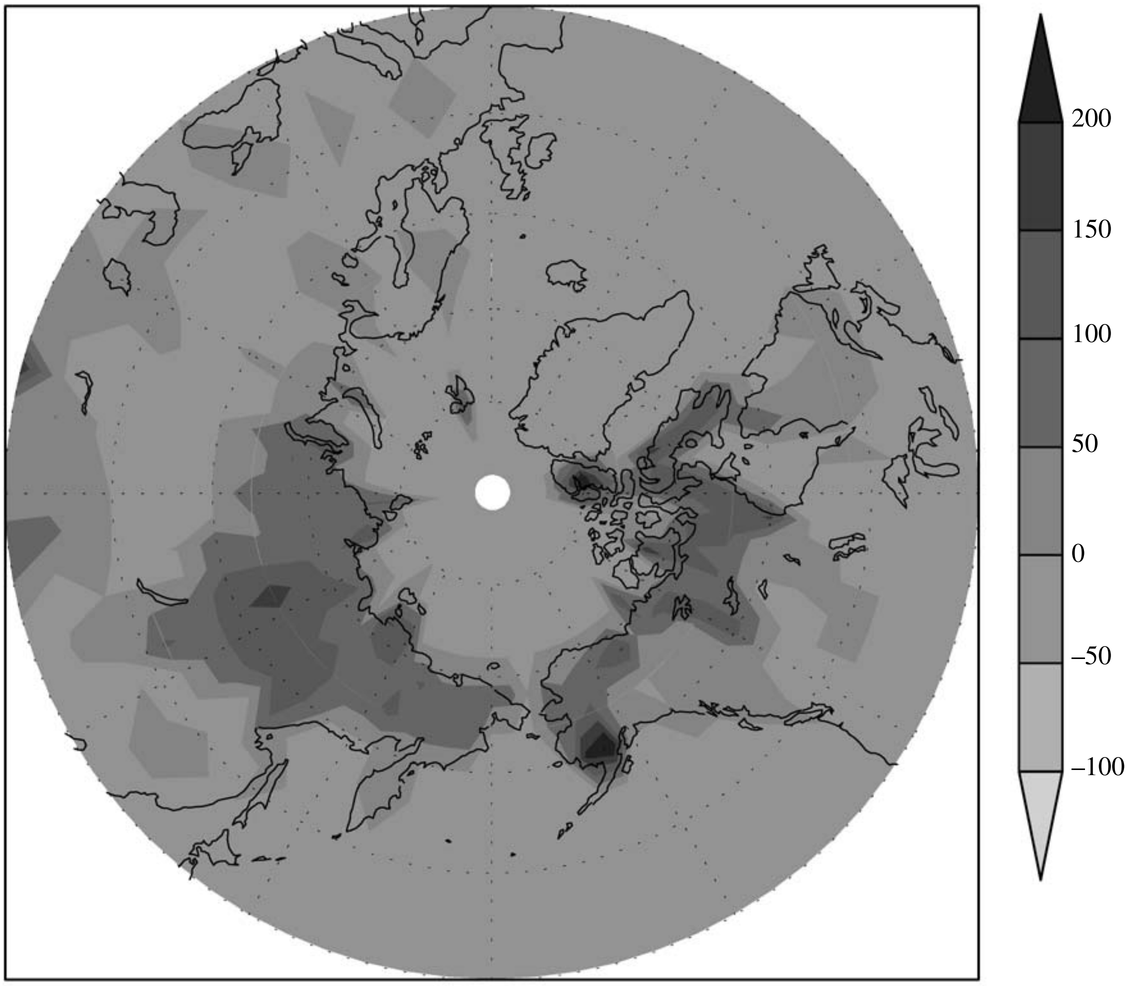

Fig. 9. The mean August difference between the total ice content in the soil column resulting from the MOSS and MOSS 60m experiments.

Figure 9 shows the mean August difference between the total ice contents in ground during the experiments, which differed only in the depths of the lower model level (60 and 10 m). It can be seen from this figure that the depth of the lower boundary of the calculated domain has a noticeable influence on the thickness of the layer of seasonal thawing. During the experiment with the calculated domain depth of 60 m, this quantity is smaller than during the experiment with a depth of 10 m. For some territories the calculated domain depth is a critical characteristic which determines the very possibility of the model permafrost existence in this territory. It should be noted that the ground temperature and moisture, on the one hand, and atmospheric characteristics, on the other, attained in both cases mutually balanced the values; however, for experiments with the depth of 60 m, this process continued much longer (by 200 model years).

The mean August difference between the total ice content in the soil columns during the TAIGA and MOSS 60m experiments is presented in Fig. 10. As would be expected, the results of the TAIGA experiment indicate that the ice amount in soil of the taiga zone is slightly larger than during the experiment in which heat-insulating properties of moss were assigned only to grid boxes with the "tundra" vegetation type.

The mean August difference between the total ice contents in ground resulting from the FrCond and MOSS experiments is presented in Fig. 11. The positive difference is observed almost along the entire southern boundary of the spreading of perennially frozen ground in Eurasia and Alaska. This indicates that the effect of the ground temperature decrease caused by the inclusion of the dependence of the heat conductivity coefficient on the temperature sign is stable and does not vanish when feedbacks of the climatic system turned on. By comparing Fig. 6 (the observed distribution of perennially frozen rocks) and Fig. 7b (the MOSS experiment), it can be concluded that the cor-

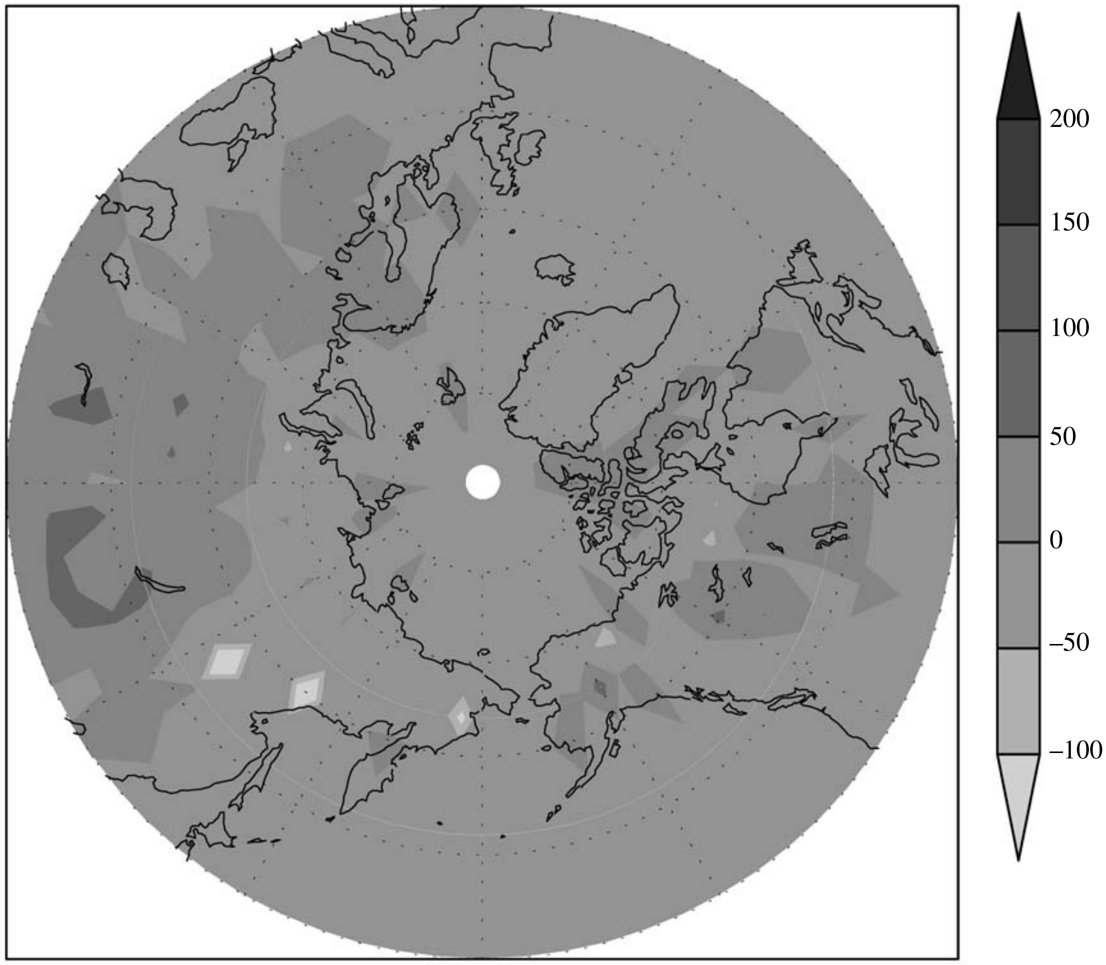

Fig. 10. The mean August difference between the total ice content in ground resulting from the MOSS 60m and TAIGA experiments.

rection for the water phase state must be introduced into the calculation of the heat conductivity coefficient of ground in order to adequately reproduce the area occupied by permafrost.

The mean August total ice content in ground resulting from the UnfrW experiment is shown in Fig. 12. It was assumed during this experiment that the temperature (below 0°C) dependence of the frozen water amount corresponded to the "clay" soil type, so that the unfrozen water amount at the given temperature was maximal among all soil types. All other physical parameters of soil remained undisturbed. It is seen from Fig. 12 that the area occupied by permafrost substantially decreases as the unfrozen water amount increases. This inference is in compliance with those made in [43]. If the possibility of the liquid moisture existence in the supercooled state is disregarded, the ground temperature will be underestimated compared with the measured temperature; otherwise it agrees better with observational data. Thus, it can be inferred that the equilibrium temperature in the soil column monotonically depends on the unfreezing water amount, beginning from its total absence; namely, the larger this amount is, the higher the temperature is. This is true of both stand-alone experiments with fixed meteorological forcing and global interactive modeling.

4. DISCUSSION OF THE RESULTS AND CONCLUSIONS

In this study, the factors controlling the heat regime of soils characteristic of cold regions are analyzed with the use of the one-dimensional model of heat and moisture transport in soil and its interaction with the atmosphere. The influence of these factors on the state of permafrost and the present-day climate as a whole is investigated on the basis of numerical experiments with the global model of atmospheric general circulation.

The one-dimensional experiments with the prescribed atmospheric action showed that the heat conductivity (in particular, the character of distribution

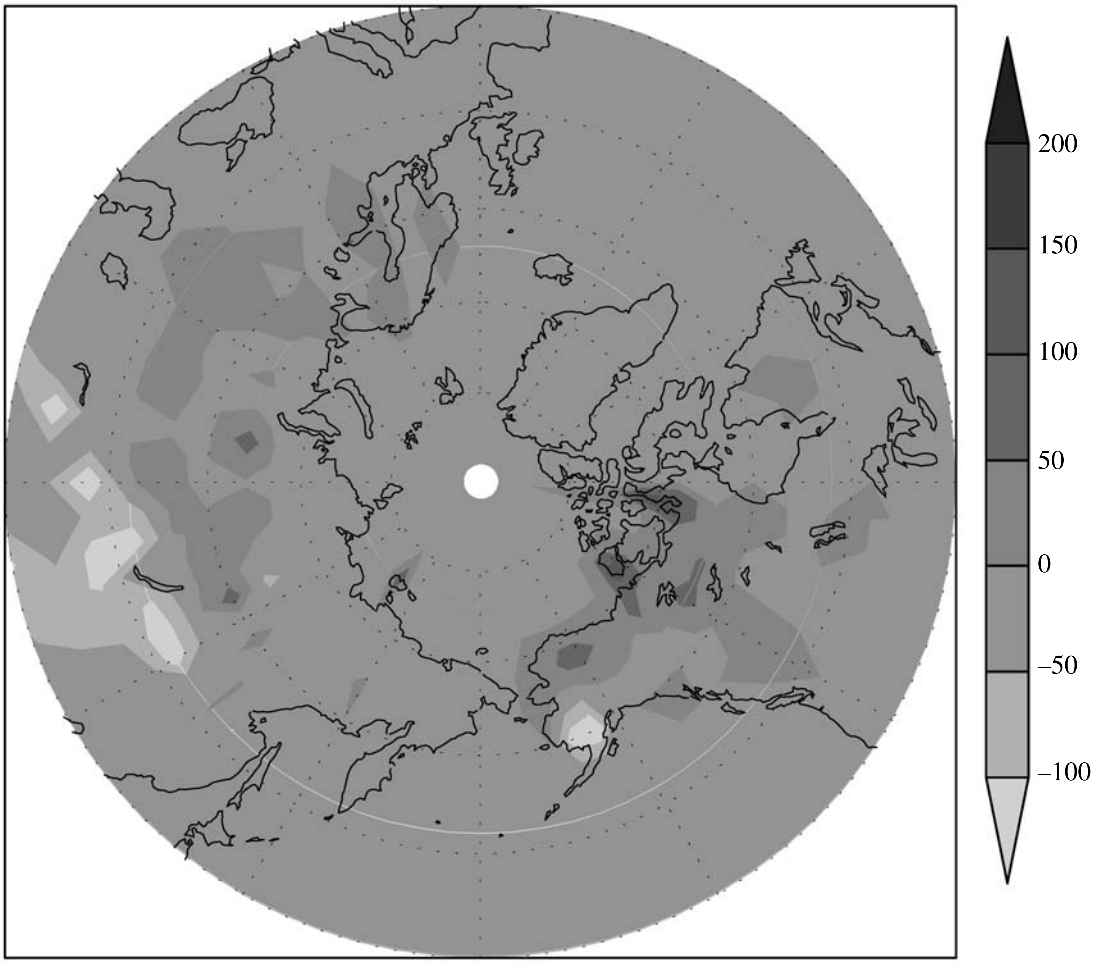

Fig. 11. The mean August difference between the total ice content in the ground resulting from the MOSS and FrCond experiments.

over depth) and the presence of liquid moisture in the supercooled state are important factors controlling the annual mean temperature in the soil column. The heat conductivity coefficient can be controlled both by the physical properties of dry ground and by the amount of moisture in ground and its phase state. In this case, the faster the heat exchange in the ground is (due to heat conductivity), the higher the annual mean temperature of the ground is.

The conclusions drawn from these one-dimensional experiments may change in switching to the complete three-dimensional modeling, because in this case the atmosphere will also respond to heat regime changes in the soil column and, in turn, will affect the active soil layer. However, the results of experiments with the global climatic model described in this study indicate that these conclusions remain qualitatively unchanged. A decrease in the heat conductivity coefficient of the upper soil layer, which can be integrated as the layer of undecomposed litter and moss, leads to a considerable increase in the area occupied by permafrost. The introduction of the dependence of the heat conductivity coefficient on the water phase state in ground (jump-like heat conductivity increase in transition from positive to negative ground temperatures) also increases the area occupied by permafrost and decreases the depth of the layer of seasonal thawing in this territory.

The conclusions associated with the existence of supercooled water in the ground are also the same both for experiments with the specified atmospheric forcing and for interactive experiments. It has been found that the larger the relative amount of water is that can be contained in ground in the supercooled state, the higher its temperature is, the larger the active layer depth is, and the smaller the area occupied by perennially frozen rocks is. Since the results of global modeling are rather sensitive to the specification of the relative amount of water that can remain unfrozen at a low ground temperature and because, in order for a correct diagnosis of climate changes, it is equally important not to overestimate or underestimate the modeled ground temperature (especially if feedbacks between the underlying surface and the atmosphere

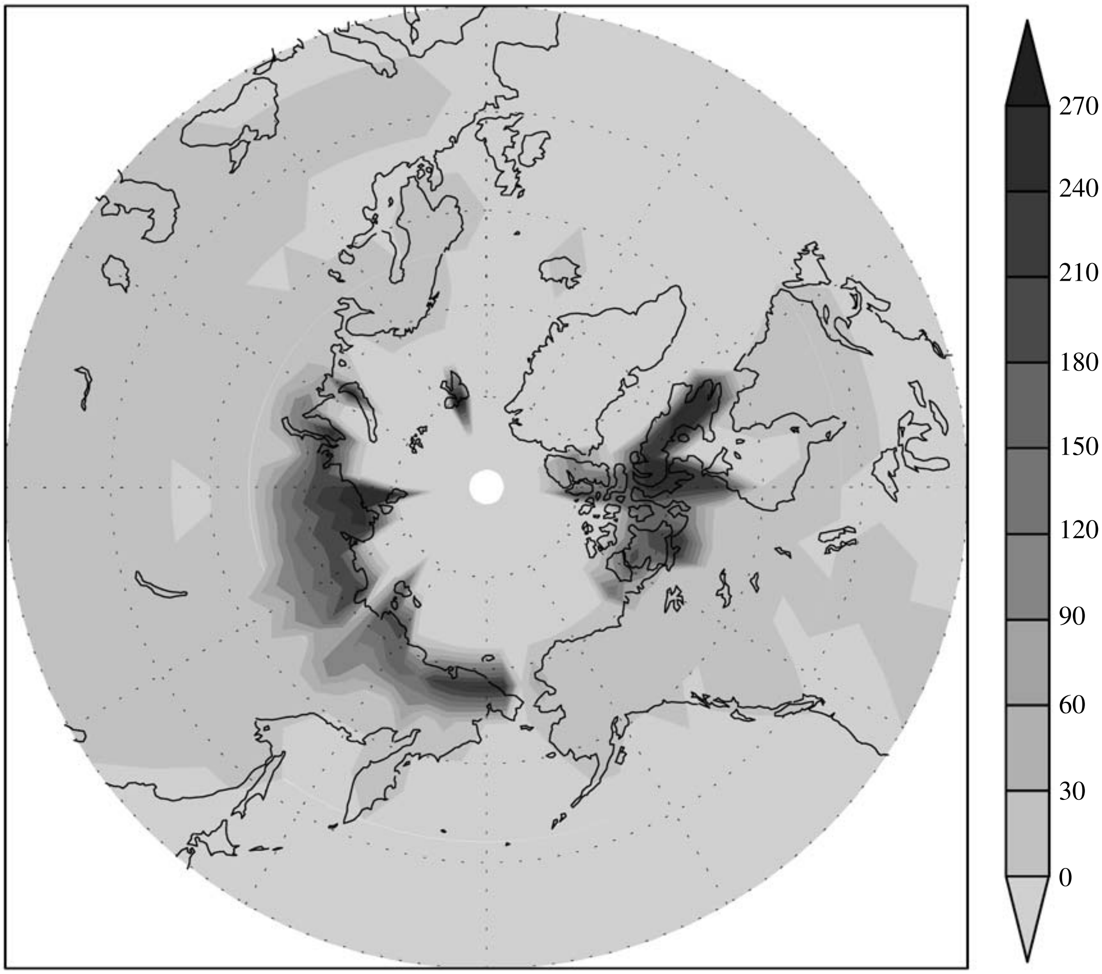

Fig. 12. The mean August total ice content in the soil column according to the UnfrW experiment.

are taken into account), it is necessary to give proper attention to the parametrization of this effect when conducting scenario experiments on climate prediction.

Since the heat conductivity of mosses is low, when a territory is swamped and overgrown with mosses, there is a decrease in the ground temperature in the depth of the soil. The TAIGA experiment, during which the heat conductivity of the upper soil layers was decreased not only in grid boxes with the tundra but also in grid boxes with evergreen coniferous forests, showed that the active layer thickness actually decreases in the territory occupied by the taiga. Within the framework of the experiment performed, the influence of the climatic system on soil does not eliminate the influence of mosses. In this case it is interesting which process (the ground temperature increase as a result of climate warming or the ground temperature decrease due to changes in the species composition of vegetation) will prevail in the future in the territory of, for example, western Siberia, and what the quantitative relationship between these oppositely directed tendencies will be? This question can be answered by performing analogous experiments with the global climatic model on the future climate prediction in accordance with the scenarios of anthropogenic emissions of greenhouse gases.

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