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Simulation of Thermal and Hydrological Regimes of Siberian River Watersheds under Permafrost Conditions from Reanalysis Data

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Abstract—A one-dimensional dynamic model of heat and moisture transfer in the soil has been developed. The use of the ERA-40 reanalysis as input data makes it possible to compute characteristics of the soil thermal and hydrological regimes, including watershed runoff, from specified climatic characteristics of the atmosphere. Results are presented of numerical experiments on a comparison of the model estimates of the depths of seasonal thawing with observations at several Siberian stations. For the latter half of the 20th century, the depths of seasonal thawing are mapped and runoff from watersheds of the largest Siberian rivers is computed. The model reproduces observed runoff variations. For the Ob basin, the model-derived runoff estimates agree well with observational data if peat deposits in the upper 2-m layer are taken into account.

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INTRODUCTION

Simulation of land hydrological processes influenced by changing climatic factors is important for a diagnosis of runoff from large rivers. Analysis of the numerical simulations with the Oboukhov Institute of Atmospheric Physics of the Russia Academy of Sciences climate model (IAP RAS CM) under different scenarios of an increase in greenhouse gases demonstrated a general growth of the annual mean runoff from Siberian rivers (Ob, Yenisei, and Lena) and from the Volga and Ural rivers in the 21st century [1, 2]. The largest increase in river runoff and in the probability that the 20th-century maximum values of river runoff will be exceeded in the 21st century has been obtained for the Lena River [1, 2]. In [1, 2], the annual mean runoff was characterized by the difference of the annual means of precipitation and evaporation in river basins. In this case, some feedback between the processes occurring in the atmosphere and in the active layer of land could be taken into account incompletely. In [1, 2], for example, mention was made of specific features of the simulation of the hydrological regime characteristics in the permafrost regions, in particular, in the Lena basin. Watersheds of most large Siberian rivers include vast areas underlain by surface permafrost. Because of this, the determination and analysis of permafrost parameters such as the depth of seasonal thawing, temperature of the underlying surface, moisture content of a thawed layer, and thermal characteristics of permanently frozen soils are important in the context of possible changes in the thermal and hydrological regimes during the 21st century.

At present, a wide variety of mathematical models are being used that describe the interaction between the atmospheric climate and soil processes. The simplest models based on the frost indices of the soil thermal regime are able to diagnose the presence or absence of permafrost, to compute the position of permafrost boundaries [3–5], and to estimate the depth of seasonal thawing [6]. Additionally, for computation of the bedrock temperature beneath a seasonally thawed layer, models based on the method proposed by V. Kudryavtsev can be used in which the influence of snow cover, vegetation, and an organic layer is taken into account [4, 7, 8]. An advantage of the models listed above is that they require a relatively small number of input parameters. A major drawback is that these are integral models and variation in variables with time is missing there. In another class of models, various characteristics of permafrost are computed depending on climate change over time. These models have differences in their numerical techniques for the heat transfer equation with phase change interfaces [9–11] and in their methods of describing the processes occurring in a vegetation–snow system. In the model of [11, 12], for example, the equations of change in the water vapor content, liquid water content, and ice content in the soil are solved in conjunction with the heat transfer equation, and the heat and moisture transfer processes in the soil are described in detail.

However, models of intermediate complexity that describe heat and moisture transfer processes in the soil are few in number. The need for such is urged by the development of climate models of intermediate complexity [13, 14]. The soil block for such models, on the one hand, must have sufficient detail for describing seasonal and regional features of soil freezing or thawing and their response to climate change (with a realistic description of the corresponding feedback) and, on the other, they must not substantially complicate or slow down the computation process. It should be noted that all currently available climate models of intermediate complexity include only very simple soil blocks (http://www.pikpotsdam.de/~andrey/emics/toe_05-06-07.pdf) that fail to simulate processes in the permafrost regions realistically. In particular, the hydrological components of these blocks are represented by a one- or a two-layer numerical scheme which prohibits a description of moisture transformations because of soil water phase change.

The goal of this study is validation of the hydrological part of the model of heat and moisture transfer in the soil designed for inclusion in the climate model of intermediate complexity of the IAP RAS CM. It should be specially stressed that the proposed model contains multilevel schemes of thermal and hydrological processes in the soil. The validation of the thermal part of the model was performed in [15]. A distinguishing feature of the dynamic model [15, 17] to be used in our study is a combination of the details in description of the heat and mass transfer processes in the soil and a relatively short computation time. The high-accuracy algorithm used for computing the vertical temperature profile and phase change boundaries [10] was supplemented with a differential scheme for computation of soil moisture [18] in a seasonally thawed layer and with a more detailed parametrization of runoff processes [11]. The algorithm that has been developed reproduces the dynamics of the formation and degradation of surface and relict permafrost. In the numerical experiments performed, the annual mean runoff from watersheds of large Siberian rivers for the period 1960–2000 has been computed using the proposed model of soil heat and moisture transfer.

DESCRIPTION OF A MODEL OF THERMAL AND HYDROLOGICAL PROCESSES IN THE ACTIVE LAYER OF LAND

A one-dimensional dynamic model is based on the soil heat and moisture transfer equations [11, 18]

$$\rho C\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda\frac{\partial T}{\partial z}\right), \qquad (1)$$

$$\frac{\partial w}{\partial t} = \frac{\partial}{\partial z}\left(\beta\frac{\partial w}{\partial z}\right) + \gamma\frac{\partial w}{\partial z} - Rf, \qquad (2)$$

where the following symbols are introduced for soil parameters: T is temperature, K; ρ is density, kg/m^3; C is the heat capacity, J/(kg K); λ is the thermal conductivity, W/(m K); β is diffusivity, m^2/s; γ is the moisture capacity coefficient, m/s; w is the moisture content, m/m; Rf is the moisture content change due to runoff, l/s; z is the space coordinate directed downward; and t is time, s. The soil thermal characteristics were computed depending on moisture content from semiempirical formulas [19]. The diffusivity and soil moisture capacity were found from the formulas given in [18, 20]. For boundary conditions at the surface for Eq. (1), the surface temperature T_s of snow or of the soil if snow was absent was specified at $z = z_s$. A geothermal heat flux of 0.06 W/m^2 was set at the lower boundary [21]. The depth of the computational domain was taken equal to 100 m [22].

The soil (snow) surface temperature T_s is determined from the heat balance equation

$$R - P - B - LE = 0, \qquad (3)$$

where R is the radiative balance, P is the eddy heat flux, B is the heat flux into the soil, and LE is heat losses due to evaporation. The heat balance components are calculated from the equation

$$R = Q(1-\alpha) - E_{ef}, \qquad (4)$$

where Q is the total incident solar radiation; α is albedo; and E_{ef} is the effective radiation of the underlying surface, which was determined from a semiempirical formula [23]:

$$E_{ef} = E_{ef}^0(1-0.79n) + 4\delta\sigma T_a^3(T_s - T_a). \qquad (5)$$

Here, n is the cloud amount, δ is the emissive capacity of the underlying surface, σ is the Stefan–Boltzman constant, T_a is the 2-m air temperature, and E_{ef}^0 is the clear-air effective radiation:

$$E_{ef}^0 = \sigma\delta T^4(0.39 - 0.058\sqrt{e}), \qquad (6)$$

where e is the partial pressure of vapor, mb.

The eddy heat flux is determined from the equation

$$P = \rho_a c v D(T_s - T_a), \qquad (7)$$

where ρ_a is the air density, kg/m^3; c is the specific heat capacity, J/(kg K); v is wind speed, m/s; and D is the diffusion coefficient, m/s.

The heat flux into the soil at the upper boundary is

$$B = \lambda\frac{dT}{dz} \quad (z = z_s), \qquad (8)$$

where z_s is the position of the upper boundary of soil (or snow).

The computation of evaporation heat losses took into account the integral moisture content in the layer of seasonal thawing; in this case, $E = E_0$ if w exceeds the critical value and $E = E_0(w/w_k)$ if $w > w_k$. The potential evaporation E_0, the maximum possible evaporation under given meteorological conditions, is determined by

$$E_0 = \rho_a v D(e_0 - e), \tag{9}$$

where e_0 is the partial pressure of vapor.

The moisture content w in the layer of seasonal thawing is found from (2). For a parametrization of horizontal runoff, a scheme used in the IAP RAS model was applied [11], according to which the surface Rf_1 and subsurface Rf_2 runoff components were found from the semiempirical formulas

$$Rf_1 = p + w\Delta h - \left[1 - \max\left(\left[\left(1 - \frac{w}{w_{\max}}\right)^{\frac{1}{1+a}} - \frac{p}{(1+a)w_{\max}}\right]^{1+a}, 0\right)\right] w_{\max}, \tag{10}$$

where p is the precipitation amount, m; Δh is the layer thickness, m; and $w_{\max}$ is the maximum value of soil moisture determined by soil porosity, $a = 0.01$. The subsurface runoff component is

$$Rf_2 = \Delta t \Delta z \rho d_{\min} \frac{w}{w_{\max}} \tag{11}$$

if $w < w_h$, and

$$Rf_2 = \Delta t \Delta z \rho d_{\min} \frac{w}{w_{\max}} + (d_{\max} - d_{\min})\left(\frac{w - w_h}{w_{\max} - w_h}\right)^d \tag{12}$$

if $w > w_h$, where $w_h = 0.75 w_{\max}$, $d_{\min} = 2.8 \times 10^{-10}$ s^{-1}, $d_{\max} = 2.8 \times 10^{-8}$ s^{-1}, and $d = 1.5$.

If an increasing soil surface temperature passes through zero and becomes positive, all heat is assumed to be spent or snowmelt.

The numerical algorithm is based on an implicit computational scheme. The time step in the model was 1 day. The depth step was varied from 0.05 m in the upper 10-m layer to 0.8 m at the lower boundary of the domain, the depth of which in the experiments was taken to be 100 m. The soil type in the model was specified by the corresponding values of density, heat capacity, and thermal conductivity. The upper 10 cm of the soil are occupied by an organic layer, the thermal properties of which have an important impact on permafrost parameters [24]. To simulate the regime of bogs, the thickness of the organic layer can be increased up to a few meters.

SIMULATION OF SOIL THERMAL REGIME CHANGES INDUCED BY ATMOSPHERIC FORCING FROM REANALYSIS DATA FOR THE LATTER HALF OF THE 20TH CENTURY

Input parameters of atmospheric forcing included monthly means of air temperature, humidity, precipitation, and shortwave solar radiation from the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-40 reanalysis [25]. In preliminary computations, it was found that the use of the downward thermal radiation as input data in the heat balance equation leads to significant differences of the model estimates from direct measurements. This is due to the absence in such computations of a feedback between the downward thermal radiation and the model-derived surface temperature. Because of this, the downward thermal radiation in the computations in our study was specified not from the reanalysis data but from (5) and (6). With the heat and moisture transfer model included as an interactive block in the climate model, it is planned to use the downward thermal radiation flux in the heat balance equation.

The simulated values of the depth of seasonal thawing were compared with observational data for the four stations: Yakutsk, Tiksi, Marre-Sale, and Vorkuta [26]. The simulated depths of seasonal thawing for Yakutsk (1.8 m) match well the observational data (1.6 ± 0.4 m). For Marre-Sale, the simulated depth value (1.1 m) is also very close to observations (1.2 ± 0.6 m). For Tiksi, the model estimates of the depth (0.6 m) are overstated relative to observations (0.4 ± 0.2 m). The largest discrepancy between the modeled and observed values is obtained for Vorkuta: the numerical result is 1.4 m, and the observed value, 0.8 ± 0.4. A computation of the depths of seasonal thawing was also performed for the land in the Northern Hemisphere free of ice for the period from 1960 to 2000. The depth of seasonal thawing/freezing was computed in each cell of a model grid from 20°N to 80°N with a step of 2.5×2.5 from the ERA-40 reanalysis. Identical initial conditions for soil temperature of –1°C and moisture content of 0.2 m/m were specified throughout the whole thickness of the model domain. The first 20 years of computation were run with a constant external forcing (corresponding to 1960) for all grid cells and were eliminated from further analysis. The resulting estimate of the area on which seasonal thawing occurs or the soil surface temperature is always negative amounts to 12.9×10^6 km^2. This value lies in a range between the estimates of the total area extent of permafrost and the area of continuous permafrost that are computed from the temperature indices [3, 5] with air temperatures specified from the reanalysis data (22.8 and 8.6×10^6 km^2, respectively). The estimate of the permafrost area produced by the

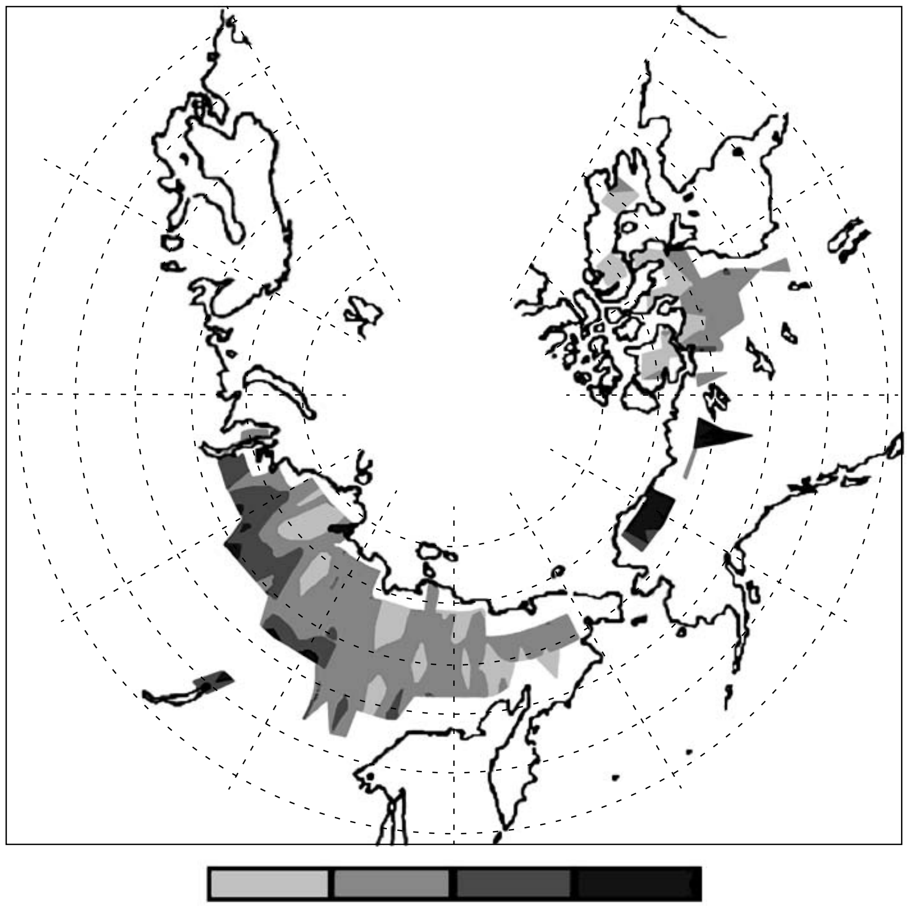

Fig. 1. Change in the model-derived decadal mean depth of seasonal thawing (m) from 1960–1969 to 1990–1999 in the permafrost regions.

model is close to that of continuous permafrost from direct measurements (11.7×10^6 km^2) [27]. The computed boundary of the region where seasonal thawing occurs is close to the boundary of continuous permafrost given in [27].

The changes in the seasonal-thawing depth H_a that are computed as the difference of the mean depths for the 1990s and 1960s are shown in Fig. 1. Positive values correspond to an increase in the depth of seasonal thawing and negative ones are associated with decreasing depths. The most significant increase in the depth of seasonal thawing is in Alaska, in agreement with measurements of the vertical temperature profile in this region [28]. However, despite a general rise in air temperature in the subpolar latitudes of the Northern Hemisphere, a small decrease in the simulated depth of seasonal thawing (up to a few centimeters) is noted in some regions of northern Asia and North America, a circumstance that was pointed out in [29]. To our knowledge, for the second half of the 20th century, similar estimates of the time changes in the depth of seasonal thawing with spatial detail have not been made before.

SIMULATION OF THE RUNOFF OF THE LARGEST SIBERIAN RIVERS

The proposed model of heat and moisture transfer in the soil is able to compute the runoff of the rivers whose watersheds include patches of near-surface permanently frozen ground, or permafrost. The model permits a description of the transition of the active soil layer from a state with near-surface permafrost (thawed layer) to the state with no surface permafrost (winter freezing layer).

The annual runoff of the largest Siberian rivers (the Ob, Yenisei, and Lena) was estimated with precipitation variations specified from the CRU data [30] and variations in the other atmospheric characteristics prescribed from the ERA-40 reanalysis. Additionally, computations were carried out with variations in atmospheric characteristics that were specified only

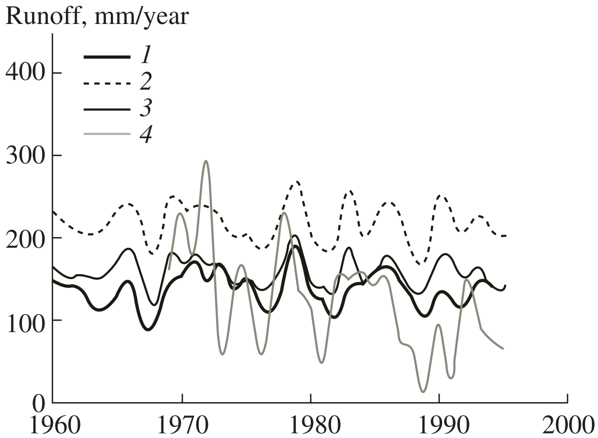

Fig. 2. Time variation of the annual runoff in the Ob basin from (*1*) observations and model estimates (with CRU precipitation) for (*2*) loam and (*3*) peat. (*4*) Precipitation minus evaporation (*P–E*) from the reanalysis data.

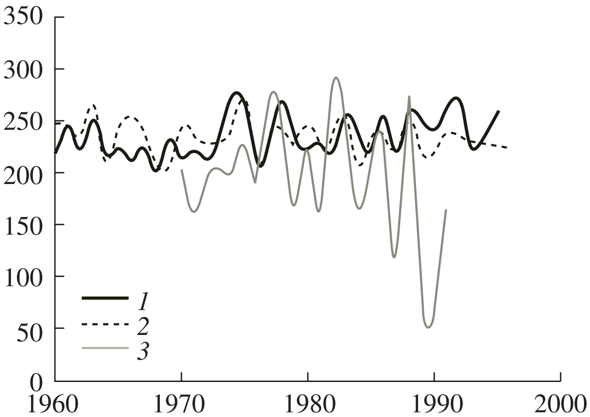

Fig. 3. Time variation of the annual runoff in the Yenisei basin from (*1*) observations and (*2*) model estimates (with CRU precipitation) for loam. (*3*) Precipitation minus evaporation (*P–E*) from the reanalysis data.

from the reanalysis, including precipitation. The total river runoff was determined as a result of spatial averaging (over watersheds of the corresponding rivers) and temporal averaging (over the calendar year) of the total surface and subsurface runoff estimated by the proposed model of soil processes. For the watersheds of these rivers, the annual mean precipitation minus evaporation was estimated from the reanalysis data and compared with direct runoff measurements at river sections [31].

The long-term annual runoff changes in the Ob, Yenisei, and Lena basins estimated from different precipitation data for two soil types are shown in Figs. 2–4. For the Ob (Fig. 2), the model-derived annual runoff differs for different precipitation data, but to a lesser extent than for the Lena watershed (see below). There is a strong connection between the calculated runoff and the soil type. For example, for loam (representative of all mineral types of soil), the annual runoff (218 ± 22 mm/year) is substantially overestimated relative to the observed data (140 ± 21 mm/year). If the peat is specified in the upper 2 m of soil and loam is set in the deeper layers (Fig. 2), the model runoff (159 ± 18 mm/year) coincides more closely with observations. The coefficient of correlation between the time series of the modeled and observed runoff is 0.7. A better agreement between the modeling results and observed data in the case where the upper organic-rich soil layer is taken into account in comparison with a version with mineral soil can be attributed to a higher percentage of peatlands in the Ob basin. The difference between precipitation and evaporation from the watershed from the ERA-40 reanalysis is shown on the plot for comparison. The interannual variability of this parameter far exceeds the interannual variations in observed runoff values, a result also common to the Yenisei and Lena basins.

For the Yenisei, with precipitation specified from the CRU data and mineral soil prescribed, the calculated mean runoff over 1958–1996 is 235 ± 17 mm/year (Fig. 3). This value agrees well with runoff estimates from observational data (236 ± 19 mm/year) and with the precipitation minus evaporation estimates (237–244 mm/year). The interannual runoff variations are also reproduced well. An exception is the period from the mid-1960s to the mid-1970s, when the variability of the model-derived runoff is much larger than that from observational data, with a correlation coefficient of 0.4.

A strong dependence of the modeling results on the choice of precipitation is obtained for the Lena (Fig. 4). If precipitation is specified from the CRU data, the model runoff is substantially underestimated (about 150 ± 20 mm/year) and its trend is statistically insignificant. If precipitation data are taken from the reanalysis up to the

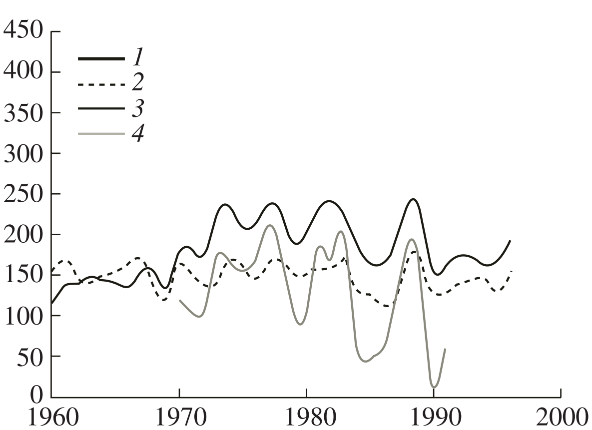

Fig. 4. Time variation of the annual runoff in the Lena basin from (*1*) observations and model estimates with precipitation from (*2*) CRU and (*3*) ERA-40 for loam. (*4*) Precipitation minus evaporation (*P – E*) from the reanalysis data.

late 1960s, the model runoff is also substantially underestimated, but then increases rapidly by about 25%, so that its mean for 1975–1996 becomes 180 ± 30 mm/year. The latter value agrees better with observed runoff data (220 ± 30 mm/year) and with its estimates as the difference between precipitation and evaporation from the watershed (190–214 mm/year). The correlation coefficient in this case is 0.8.

CONCLUSIONS

A model of heat and moisture transfer in the active soil layer has been developed for inclusion in the IAP RAS climate model of intermediate complexity. Verification of the model has been performed with atmospheric forcings specified from the ERA-40 reanalysis data for the latter half of the 20th century. The model adequately reproduces both the spatial distribution of permafrost soils and the seasonal-thawing depths measured at stationary sites. In the performed model simulations, a significant increase in the depth of seasonal thawing is noted for Alaska in recent decades. The model can be included as an interactive block in a global climate model for numerical simulations of possible climate changes with an assessment of regional consequences.

For the second half of the 20th century, numerical simulations of changes in the runoff of the largest Siberian rivers under the influence of climate change have been conducted with consideration for the influence of permafrost degradation on the integral moisture capacity and on the storage of snowmelt in the soil. Comparison of the numerical results with observational and reanalysis data has shown that the model is generally able to reproduce the observed variations in the runoff of the Ob, Yenisei, and Lena in recent decades. It is noted that an adequate simulation of the Ob runoff requires that the area of vast swamped lands in western Siberia be taken into consideration. For the Lena, the best agreement between the numerical results and the observed data has been obtained using precipitation from the ERA-40 reanalysis as compared with the CRU data.

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