Modelling hydrological consequences of climate change in the permafrost region and assessment of their uncertainty

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Abstract A physically-based, distributed model of runoff generation in the permafrost regions is presented. The model describes processes of snow cover formation, taking into account blowing snow sublimation, snowmelt, freezing and thawing of the ground, water detention by a basin storage, infiltration, evaporation, overland, subsurface and channel flow. An important feature of the model is the detailed description of water and heat transfer within the active layer of soil during its seasonal thawing and freezing. A case study has been carried out for the Pravaya Hetta River basin (the catchment area is 1200 km^2) of Western Siberia within the Lower Ob River basin. The basin is located in tundra and forest-tundra vegetation zones. It has been shown that after precipitation, melt of ground ice is the second largest input to the basin water balance and accounts for about 70% of annual precipitation. Seasonal snow losses due to sublimation during blowing snow transport can reach almost 30% of the maximum snow accumulation. The model has been applied to assess the impact of climate change on hydrological processes in the permafrost basin. Uncertainty of the simulated hydrological consequences of climate change has been assessed by the multi-scenario approach. Simulated runoff response to the projected climate change varies significantly as a result of the uncertainty of the climate change scenario.

Key words permafrost hydrology; cold region modelling; climate change; uncertainty

INTRODUCTION

The permafrost region covers approximately one-quarter of the Earth's land surface, more than 60% of Russia and one-half of Canada. It constitutes a unique environment with an important role in the dynamics and evolution of the Earth system. Economic growth in the permafrost regions, the necessity of protection of the northern environment and maintenance of biodiversity, as well as increasing attention to global climate processes, make the investigations of permafrost hydrological processes, especially runoff generation, very important and urgent. It is well known that the permafrost region plays an important role in climate dynamics and represents potentially important sources and/or sinks of greenhouse gases. A changing climate could in turn lead to visible physical changes that could augment or retard global climate change and significantly impact ecosystems (Rouse *et al*., 1997). There is increasing evidence that environmental change has reached an unprecedented degree. Many of these changes are related to the hydrological cycle and can result from both the direct and indirect impacts of human activities (Vörösmarty *et al.*, 2001).

The importance of the aforementioned issues requires advancing the current understanding of permafrost hydrology and developing, on this basis, adequate methods for their solution. Such an advance has to be associated with field and modelling studies. In recent decades, extensive experimental work in permafrost regions of different physiographic conditions has been carried out and summarized by Woo (1990), Quinton & Marsh (1999) and Carey *et al.* (2010). The accumulated knowledge of the main specific features of permafrost hydrology, which make it clearly distinguishable from the hydrology of temperate regions (Woo, 1990), has created a foundation for the development of process-based hydrological models representing these features, such as the models presented by Kuchment *et al.* (2000), Zhang *et al.* (2000), Quinton *et al.* (2004), Pomeroy *et al*. (2007), Dornes *et al*. (2008) and Zhang *et al*. (2008). Parameters of such models have clear physical meaning and can be related to measurable characteristics of river basins, such as topography, soil, vegetation, etc. Combined with, and resulting from the physical background of the models, this feature provides fresh opportunities for *a priori* assessment of the parameters and model validation in poorly gauged basins (Gelfan, 2005), which are typical for permafrost regions. In addition, these models allow one to obtain reasonable results in the case of environmental changes, particularly to assess hydrological consequences of the projected climate change.

The main objectives of this paper are to present the development of a physically-based, distributed hydrological model of permafrost watersheds located in the tundra vegetation zone, as well as applications of the model for assessing hydrological consequences of climate change and their uncertainty. In comparison with the model presented by Kuchment *et al*. (2000), more sophisticated modelling of heat and moisture transfer in the active layer has been developed in this study; it also uses the model of blowing snow transport (Essery *et al*., 1999).

MODEL OF RUNOFF GENERATION: STRUCTURE, PARAMETER ASSESSMENT AND VALIDATION

The low-gradient Pravaya Hetta catchment is located in the north of Western Siberia, it drains west into the Nadym River (a tributary of the Ob River). The study area of approx. 1200 km^2 (up to the outlet in the Pangody town, 61°53′N, 147°43′E) is situated at the boundary of tundra and forest-tundra vegetation zones. Wetland and peatland occupy about 79% of the area covered by moss, lichen, and shrubs. The river basin is situated in the zone of continuous permafrost interrupted only by taliks under the river beds. The dominant soils are tundra podzols. On slopes of northern exposure, the average depth of the active layer is 0.2–0.8 m; on slopes of southern exposure and at the peatlands, the average depth of the active layer reaches 2.0 m. The study area has a sub-arctic continental climate, characterized by a large temperature range and low precipitation. Mean annual temperature is –7.8°C. Mean temperatures of January and July are –26.4°C and +15.4°C, respectively. Mean annual precipitation is 290 mm, of which about 100 mm falls as snow. The maximum depth of snow in the uplands is 10–20 cm, while in the lowlands it can exceed 40 cm and 70 cm in the areas covered by trees and shrubs, respectively. The majority of runoff results from snowmelt in the late spring and summer. Only minimal runoff occurs during the summer because of rainfall. Peak runoff discharges during snowmelt exceeded 300 m^3/s, while peak runoff events during summer were only about 30 m^3/s.

To calculate the characteristics of snow cover, a system of vertically-averaged equations of snow processes has been applied (Kuchment *et al*., 2000). The system includes a description of temporal change of the snow depth, content of ice and liquid water, snow density, snowmelt, sublimation, refreezing melt water, and snow metamorphism. In order to simulate the effect of wind redistribution of snow, we applied the Simplified Blowing Snow Model, SBSM, presented in Essery *et al.* (1999). This simulation calculates downwind blowing snow transport and in-transit sublimation losses.

Vertical water and heat transfer associated with seasonal freezing and thawing of the active soil layer is described with the equations (Gelfan, 2006):

$$\frac{\partial W}{\partial t} = \frac{\partial}{\partial z}\left(D\frac{\partial \theta}{\partial z} + D_I \frac{\partial I}{\partial z} - K \right) \tag{1}$$

$$c_T \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda \frac{\partial T}{\partial z} \right) + \rho_w c_w \left(D\frac{\partial \theta}{\partial z} + D_I \frac{\partial I}{\partial z} - K \right)\frac{\partial T}{\partial z} + \rho_w \chi \frac{\partial W}{\partial t} \tag{2}$$

where W, θ and I are the total water content, liquid water content and ice content of soil, respectively, where $W = \theta + (\rho_i / \rho_w)I$; $K = K(\theta, I)$ is the hydraulic conductivity of soil; T is the soil temperature; λ represents the thermal conductivity of soil; $D = K(\partial\psi / \partial\theta)_I$ and $D_I = K(\partial\psi / \partial I)_\theta$; $\psi = \psi(\theta, I)$ is the capillary potential of soil; $c_T = c_{eff} + \rho_w \chi(\partial\theta / \partial T)$; c_{eff} is the effective heat capacity of soil $c_{eff} = \rho_g c_g (1 - P) + \rho_w c_w \theta + \rho_i c_i I$; ρ and c are the density and the specific heat capacity, respectively, where indexes *w*, *i* and *g* refer to water, ice and soil matrix, respectively; P is the soil porosity; χ is the latent heat of ice fusion.

The capillary potential $\psi = \psi(\theta, I)$ and the hydraulic conductivity $K = K(\theta, I)$ of the frozen soil are determined from the relationships (Gelfan, 2006) obtained by transformation of van Genuchten's formulas.

Melt water detention in the topographic depressions and peat mats is simulated under the assumption that free storage capacity is exponentially distributed over the basin area with the mean value DET_0. Under this assumption, cumulative water detention, DET_Σ, by the basin surface is calculated by:

$$DET_\Sigma = DET_0\left[1-\exp\left(-\frac{R_\Sigma}{DET_0}\right)\right] \tag{3}$$

where R_Σ is the snowmelt outflow from snowpack accumulated from the beginning of melt.

The rate of evaporation, E, from an unfrozen, snow-free soil is calculated as:

$$E = K_E d_a S_1 \tag{4}$$

where S_1 is the relative saturation of the upper soil layer; K_E is the empirical coefficient.

To simulate overland and subsurface flow over the Pravaya Hetta River basin, the catchment area was schematized as a series of rectangular reaches located along the main channel. Overland flow along each of the schematized reaches was described by the length-integrated kinematic wave equation written as:

$$L\frac{dh}{dt} = RL - q_l \tag{5}$$

$$h = \frac{m}{m+1}\left(\frac{q_l n_l}{i_l^{0.5}}\right)^{\frac{1}{m}} \tag{6}$$

where h is the average flow depth; m equals 5/3, L is the length of the reach; R is the rate of water inflow per unit length of the reach; q_l is the lateral overland inflow rate per unit length of the channel; i_l is the slope of the overland flow; n_l is the Manning's roughness coefficient.

It is also assumed that the horizontal movement of water in the active layer occurs only if the soil moisture content exceeds the soil field capacity θ_f. As a result, the subsurface flow is generated with the rate $R_g = \begin{cases}\frac{\partial}{\partial t}\left[\left(\theta-\theta_f\right)h_g\right], & \theta > \theta_f \\ 0, & \text{otherwise}\end{cases}$ where h_g is the depth of the subsurface flow.

Subsurface flow was described by the equations

$$(P-\theta_f)L\frac{dh_g}{dt} = R_g L - q_g \tag{12}$$

$$h_g = \frac{q_g}{2K_g i_l} \tag{13}$$

where q_g is the discharge; K_g is the coefficient of horizontal conductivity of soil.

To simulate channel flow, the one-dimensional kinematic wave equation is applied.

Calibration and validation of the model was carried out using the available daily hydro-meteorological measurement data for the period from 1 January 1981 to 31 December 1986. Initial snow depth and density values were assigned from snow survey measurements. Initial liquid water content in snowpack was assumed to equal zero. Because of the absence of the necessary measure-ments, initial soil heat and water content were assigned as constant on soil depth, equalled to the corresponding climatic values. Using these values, the initial ice content of soil was calculated by the model.

Soil porosity and density, as well as field capacity and horizontal hydraulic conductivity of the organic soil, were assigned *a priori* on the basis of the available measurements for the similar types of soil (Pavlov *et al.*, 1997; Sławiński *et al.*, 2004). The snow sub-model was calibrated using snow survey measurements as well as data of special observations on snow evaporation (Kuz'min, 1976). A comparison of the measured and calculated snow depth is shown in Fig. 1.

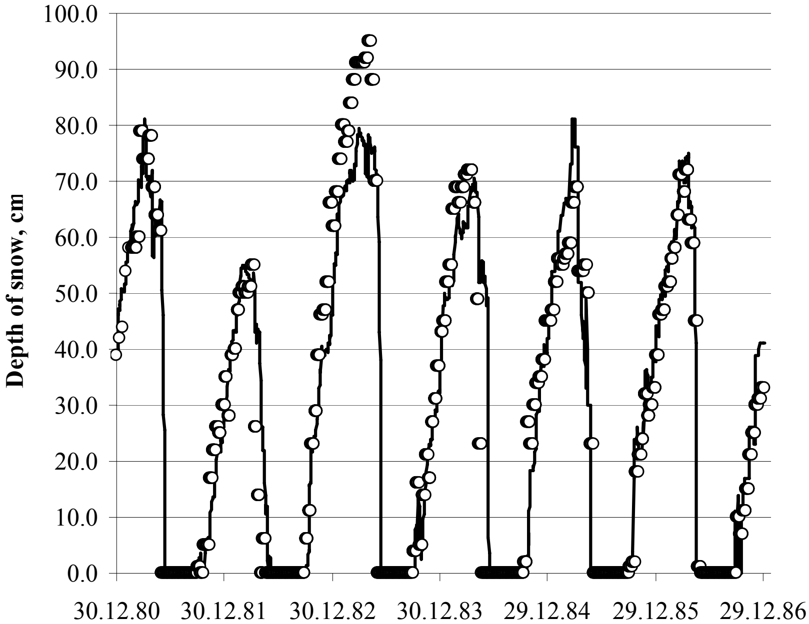

Fig. 1 Comparison of the observed (points) and calculated (line) depth of snow at the meteorological station Pangody.

The model showed that seasonal snow losses due to sublimation during blowing snow transport vary within 17–31 mm, depending on the meteorological conditions of the specific season, and reach 29% of the maximum snow accumulation; this result is close to that reported by Essery *et al.* (1999) for the shrub tundra.

Five parameters (saturated hydraulic conductivity of soil, the Manning's roughness coefficient of river channel, and the coefficients DET_0 and K_E in equations (3) and (4), respectively) were adjusted through calibration against observed hydrographs from 1 January 1981 to 31 December 1983. The model validation was carried out by comparison of the observed and simulated hydrographs for the period from 1 January 1984 to 31 December 1986. The simulated and the observed hydrographs for the six years of the calibration and validation periods are presented in Fig. 2. The Nash and Sutcliffe efficiency criterion for the discharge simulations is 0.74 for the calibration period and 0.73 for the validation period.

According to the simulation results, the largest floods in the basin are caused by overland snowmelt inflow into the channels. Most of the ice melt water formed within the thawing active layer of ground, as well as infiltrated rainwater, does not reach a channel and evaporates during a warm period. After precipitation, melt of ground ice is the second largest component of the basin water balance, and accounts for about 70% of annual precipitation.

UNCERTAINTY OF RUNOFF RESPONSE TO CLIMATE CHANGE

There are many estimates of hydrological consequences of the current and the projected climate change in the permafrost regions, particularly, for the Ob River basin (Yang *et al.*, 2004). These estimates contain significant uncertainty caused, most of all, by the diversity of the future climate scenario. An attempt is made below to account for this source of the uncertainty.

According to the recent IPCC report (*Climate Change*, 2007), annual temperature and precipitation in the Siberian north may increase by 1.3–4.7°C and 5–20%, respectively, depending on the chosen greenhouse gases emission scenario. In order to assess hydrological consequences of the projected climate change and their uncertainty, the following multi-scenario approach was applied (Fig. 3). At first, 1000 random combinations of annual changes of air temperature ΔT and

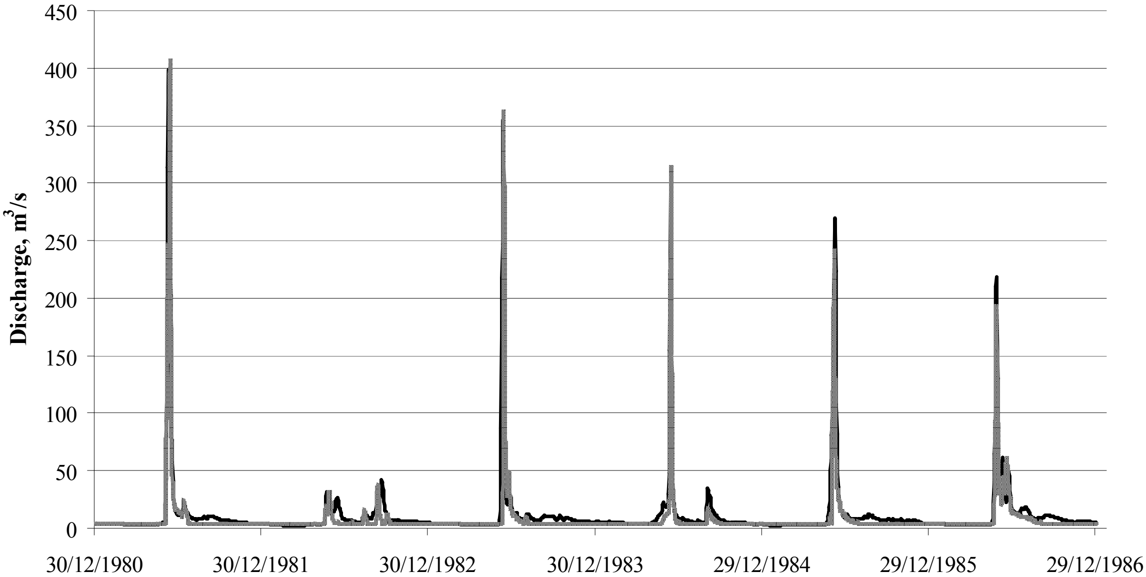

Fig. 2 Comparison of the observed (black line) and calculated (grey line) hydrographs at the gauge Pangody of the Pravaya Hetta River.

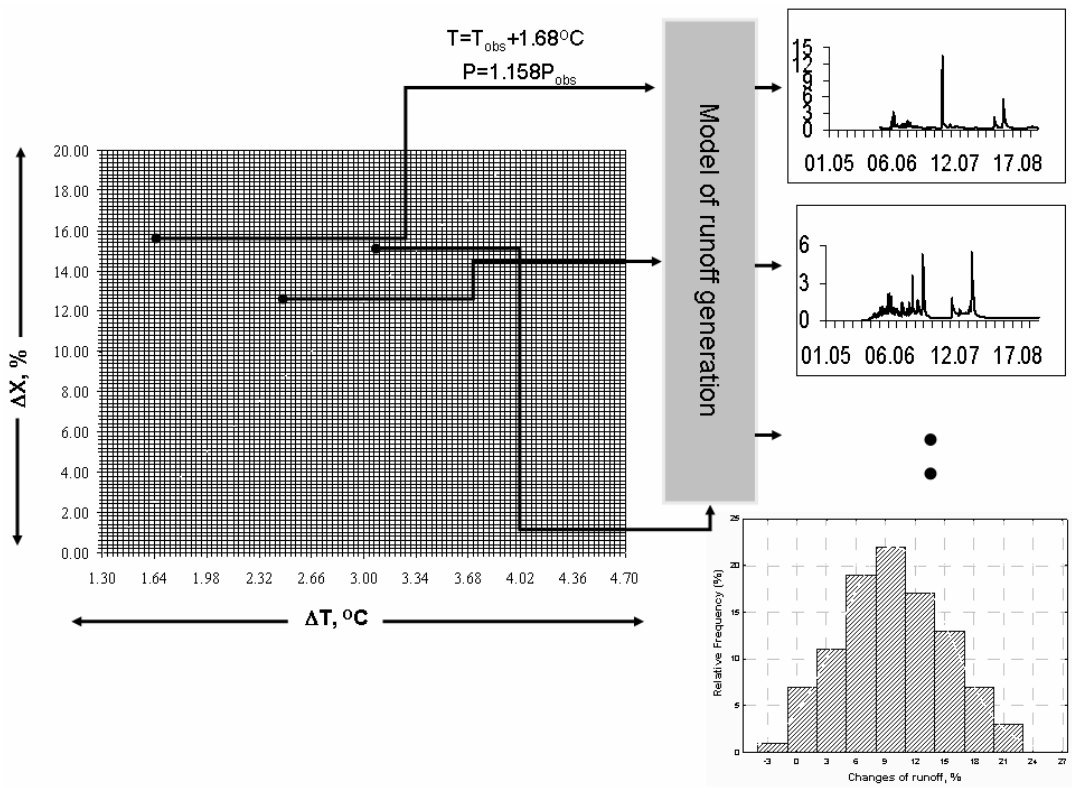

Fig. 3 Multi-scenario approach to assess uncertainty of hydrological consequences of the projected climate change.

precipitation ΔP were sampled from the intervals $\Delta T \in [1.3°, 4.7°]$ and $\Delta P \in [5\%, 20\%]$ by the Latin hypercube method. It was assumed that ΔT and ΔP are independent random variables, uniformly distributed within the respective interval, i.e. any combination of ΔT and ΔP was equally guessed. Then, the daily series of temperature and precipitation measured in the basin under consideration were changed according to the assigned values of ΔT and ΔP. As a result, 1000 synthetic series of these meteorological variables were obtained. Finally, the model was forced by these series and 1000 hydrographs were simulated reflecting uncertainty of the hydrological consequences of climate change.

Mean response of runoff to 1000 constructed scenario of climate change turned out to be insignificant: annual runoff and peak discharge increased by 5.6% and 6.7%, respectively. At the same time, variation of the runoff response is rather high. Changes of the annual runoff can vary within the 95%-confidence interval of [–1%, +16%]. The corresponding interval for the annual peak discharge is [–2%, +26%]. Decrease of runoff can occur under a large increase in air temperature along with a small change in precipitation. In this case, the small increase of precipitation cannot compensate for the increase of the basin evaporation.

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