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The updated version of SPONSOR land surface scheme: PILPS-influenced improvements

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Abstract

The parameterization scheme SPONSOR (Semi-distributed ParameterizatiON Scheme of the ORography-induced hydrology) participating in PILPS (Project for Intercomparison of Land-surface Parameterization Schemes) experiments since 1993 is described in more detail than before, taking into account a range of recent modifications. Improvement of the scheme in several aspects (e.g., soil water movement) resulted in significantly improved results for the Cabauw site (used for PILPS (2a) experiments). Then, parameterization of cold seasons/regions processes (water phase transformations within soil and snow cover) was developed for PILPS (2d) experiments carried out with Valdai data. Testing of the scheme against the data of Kolyma water balance station shows that it is able to reproduce the main features of heat and water exchange at the land surface in the permafrost zone quite satisfactorily. It was found that the scheme results are rather sensitive to the soil heat conductivity, especially in the cold seasons. The original method for the calculation of this parameter was developed using a square root function. The surface temperature and dates of crossing the 0°C temperature threshold for Kolyma station were reproduced with satisfactory accuracy. The temporal variation of the deep soil layers' temperatures was modelled satisfactorily too, but the seasonal amplitude of deep soil temperatures was overestimated by the scheme. This disadvantage can possibly be improved by inclusion of vertical inhomogeneity of soil thermal and hydraulic properties in the model.

Keywords: energy/water balance; land surface; parameterization; permafrost

1. Introduction

During the last decades, great attention has been paid to the problem of parameterization of physical processes for numerical modelling of the atmosphere. This problem becomes especially urgent for experiments with climate change and its consequences. The energy/water exchange at the land surface is an important block of the models as it represents the lower boundary condition for the atmosphere and regulates the whole cycle of energy transformations; moreover, the land cover itself can undergo transformations resulting from climate changes. The processes at the land surface have rather complicated nature, and some their aspects are not well

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studied yet. At the same time, in most cases the processes cannot be described in all details because of the unavailability of the land cover data with fine resolution, and/or limitations of the computer resources. Typically, the developers of such schemes try to find a compromise between the necessity of detalization of physical processes and these limitations. Moreover, the land surface schemes are intended for different tasks with different types of atmospheric or any other (e.g., ecological) models, or as an independent tool, without any model. As a result, one can find in the scientific community a variety of land surface parameterization schemes strongly differing in underlying philosophy and modelling approaches. The international Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) started in 1992 in order to improve the understanding of the existing parameterizations of interactions between the atmosphere and the land surface in climate and weather forecast models (PILPS Workshop Report, 1992). This project gives an opportunity to test the participating parameterization schemes against the observational data obtained in different natural conditions, as well as to compare the modelled results with each other. To date, several PILPS stages have been carried out, and many interesting results obtained (Chen et al., 1997; Shao et al., 1994; and others). As a rule, before the end of each stage the participants do not have access to validation data, so any improvements in the schemes according to their behaviour become possible only after the end of the given stage. Hence, the publications prepared within the framework of PILPS program usually contain the results obtained before such improvements.

The scheme known to PILPS community as SPONSOR (Semi-distributed ParameterizatiON Scheme of the ORography-induced hydrology) took part in many experiments, and it has been changed significantly since the first description in English (Shmakin et al., 1993). Some procedures were modified in the scheme, and several new blocks were developed too. Many of these modifications were made after some intercomparisons within the PILPS community, or according to the requirements of some PILPS stages (for example, the advanced cryospheric block was developed for the intercomparisons at Valdai site with long cold season, though it is also planned to be used for other purposes). This paper presents the latest updated version of SPONSOR scheme and some results of its validation.

2. Description of the model

The Land Surface Scheme (LSS) allows one to calculate the heat and water budget components at the land surface if the atmospheric forcing variables are known. The latter include incoming solar and longwave radiation, precipitation rate, air temperature, air humidity, wind velocity and atmospheric pressure. All these variables are available in PILPS experiments or when the scheme is used within an atmospheric model; otherwise they must be calculated in special blocks. SPONSOR includes a description of the whole cycle of energy and water exchange at the land surface: interception of precipitation by vegetation, formation of runoff by different mechanisms, infiltration of water into the soil, water exchange between soil layers and with underlying ground, evapotranspiration, turbulent sensible heat flux to the atmosphere, conductive heat flux into the soil and snow cover with correspondent changes of their temperatures, soil freezing and thawing, snowmelt. At every time step, we calculate the values of landscape surface temperature, soil temperature at 2 levels, temperature at the boundary between the soil and snow cover, intercepted water storage, storages of liquid and frozen water within both of the 2 soil layers, and the amount of unfrozen water within snow cover. The lower boundary conditions include ground temperature at the depth from 3 to 15 m (according to available data), and free drainage condition if the water table is situated deeper than 4 m (otherwise its depth should be specified).

2.1. Water balance components

The interception of precipitation by vegetation, runoff formation and infiltration into the soil are calculated in the same manner as earlier (Shmakin et al., 1993), with only some improvements incorporated. Interception of precipitation is parameterized using a method proposed by many authors (e.g., Noilhan and Planton, 1989). The amount of precipitation intercepted by plants at the n-th time step (w_{in}) is equal to:

$$w_{\text{in}} = \min\left(w_{\text{in}-1} + P\tau, 0.1(\text{LAI} + \text{SAI})\,\sigma_f\right) \tag{1}$$

where P is the precipitation rate, τ is the time step, LAI and SAI are the leaf area index and the similar index for stems, branches and other non-green parts of plants, σ_{f} is the vegetation cover fraction for the given territory. The precipitation rate under vegetation cover (P_{v}) is given by:

$$P_{\text{v}} = \max(P - w_{\text{in}}/\tau, 0) \tag{2}$$

With respect to the soil water storage, it is assumed that the relative soil water content (equal to 0 at wilting point and 1 at field capacity) represents the relative area of saturated soil zones, while the remainder of the territory is occupied by completely dry plots (we define ‘dry’ soil as soil with water content equal to wilting point). Then, it is assumed that the saturated zones are located in the lowest parts of valleys and the adjacent parts of slopes of elementary valleys (the ‘contributing areas’), while the completely dry zones occupy the upper parts of slopes and water divides (which are considered to be generally flat). The morphometric parameters of the elementary valleys (their depth and total length per square unit, fraction occupied by flat water divides) should be specified. The increase of soil humidity can be imagined as ‘ascending’ the slopes by the saturated zone with its spreading, and vice versa.

In the case of rainfall, precipitation falling on the contributing areas goes immediately to surface runoff flow due to Dunne’s mechanism (Bevan et al., 1988; Wood et al., 1988). The runoff rate r_w is equal to:

$$r_w = \min(P_{\text{v}}, k_{\text{sa}})\, w_{1\text{eff}}^2$$

$$w_{1\text{eff}} = w_1 F_{\text{c}}/\theta_w + F_{\text{rs}1}/(d_1\theta_w)$$

$$k_{\text{sa}} = k_{\text{s}}(b\psi_{\text{s}}/d_1 + \cos X) \tag{3}$$

where w_1 and d_1 are the relative water content and the depth of the upper soil layer, F_{c} and θ_w are soil field capacity and porosity respectively, $F_{\text{rs}1}$ is the amount of frozen water within the upper soil layer [mm], k_{s} is the saturated hydraulic conductivity of the soil, ψ_{s} is the matrix potential at saturation, b is the Clapp–Hornberger parameter, X is the steepness of slope of elementary basin (certainly, the term $\cos X$ must be replaced by unity for the flat part of the territory). Parameters k_{s}, b and ψ_{s} are taken as constants for the given soil texture (Clapp and Hornberger, 1978). It is assumed also that the ice fraction within the soil can occupy all pores (i.e., it can reach porosity as a maximum), and the liquid soil water content cannot exceed field capacity.

The rain falling on the ‘dry’ areas can generate Hortonian runoff, if the precipitation intensity is greater than the soil infiltration rate. The non-negative Hortonian runoff intensity r_{h} can be calculated as follows:

$$r_{\text{h}} = (P_{\text{v}} - k_{\text{sa}})(1 - w_1) \tag{4}$$

The subsurface runoff r_{g} is assumed to be formed by the exfiltration from the lower soil layer. It is calculated using the free drainage condition:

$$r_{\text{g}} = k_{\text{s}} w_2^{2b+3} \cos X \tag{5}$$

where w_2 is the relative water content of the lower soil layer.

The increase of the soil water storage due to precipitation falling on the ‘dry’ areas can be imagined as the saturation of some soil layer of depth h, in accordance with the precipitation rate. Then, all this soil water, being distributed over the whole depth of the upper soil layer, will be concentrated in the lower parts of slopes adjacent to the saturated zones, more or less as indicated by Zaslavski and Sinai (1981). So, the extension of the saturated zones in two soil layers due to precipitation (d_{pr1} and d_{pr2}) can be calculated as:

$$d_{pr1} = h(1 - w_1)/d_1 < 1 - w_1$$
$$d_{pr2} = (h - d_1)(1 - w_2)/d_2, \text{ if } h > d_1$$
$$h = \min(P_v, k_{sa})\tau/F_c \tag{6}$$

where d_2 is the depth of lower soil layer which is equal to the depth of the soil root zone of the given landscape, or 1 m if there is no vegetation. It was assumed following Entekhabi and Eagleson (1989) that water percolates to the dryer soil through the thin saturated layer, formed by ponding at the very surface.

The decrease of the saturated zones due to evapotranspiration is calculated by the same approach:

$$d_{e1} = E_1\tau(w_1 + d_{pr1} + d_{m1})/(d_1 F_c)$$
$$d_{e2} = E_2\tau(w_2 + d_{pr2} + d_{m2})/(d_2 F_c) \tag{7}$$

where E_1 and E_2 are the rates of evapotranspiration from the corresponding soil layers, d_{m1} and d_{m2} are the increments of soil humidity due to melting of snow and soil ice, calculated by the same method as in Eq. (6), when these processes occur. The rates of evapotranspiration from the two soil layers (E_1 and E_2), evaporation from snow cover (E_{sn}) and evaporation of intercepted precipitation (E_i) are calculated by

$$E_i = \min\left(\frac{E_p w_{in}}{0.1(\mathrm{LAI} + \mathrm{SAI})\sigma_f}, \frac{w_{in}}{\tau}\right)$$
$$E_{sn} = \min(E_{ps}, S_n/\tau)$$
$$E_1 = w_1(E_p - E_i)(1 - \sigma_f + R_1\sigma_f f_d)\mathrm{LAI}/(\mathrm{LAI} + \mathrm{SAI})$$
$$E_2 = w_2 f_d(E_p - E_i - E_1)(1 - R_1)\sigma_f \mathrm{LAI}/(\mathrm{LAI} + \mathrm{SAI}) \tag{8}$$

where E_{ps} is the rate of potential evaporation of snow, S_n is snow water equivalent (mm), R_1 is the fraction of vegetation roots in the upper soil layer, f_d is the fraction of green leaves which are not covered by intercepted water at the given time step. Potential evapotranspiration rate E_p follows (Thom, 1975):

$$E_p = K_t(q_s - q_a)$$
$$K_t = \frac{\rho u f \kappa^2}{\log^2\left(\frac{z_u - d_0}{z_0}\right)} \tag{9}$$

where q_s is the surface specific humidity at saturation according to Magnus formula; q_a is the air specific humidity; ρ is the air density; u is the wind velocity measured at the height z_u; f is the factor of correction for atmospheric lapse rate according to Brutsaert (1982); κ is the von Karman constant; d_0 and z_0 are, respectively, the zero plane displacement and the roughness length of vegetation calculated from its height, LAI and SAI according to Thom (1975). Potential evaporation rate of snow E_{ps} is calculated with $d_0 = 0$ and $z_0 = 0.0024$ m (this assumption is maybe too rough for the case of snow underlying tall vegetation, but it is better to use these aerodynamic parameters for snow evaporation than those calculated for vegetation cover).

The water loss due to subsurface runoff is calculated in the same manner as that due to evapotranspiration:

$$d_{g2} = \frac{r_g\tau}{d_2 F_c} \tag{10}$$

Water flux between soil layers q_z (positive if directed upward) is calculated by Darcy equation in generalized form, if there are available (not occupied by water and/or ice) pores in the given soil layer:

$$q_z = -k_s\left(\frac{\partial\psi}{\partial z} + 1\right) \tag{11}$$

where ψ is the soil matrix potential, z is the vertical coordinate. This equation is transformed into finite-difference form considering approximations given in (Clapp and Hornberger, 1978), according to the depths of the soil layers. The water inflow to the lower soil layer from the ground water table (e.g., like in Cabauw where the ground deeper than 1 m is permanently saturated) is calculated by the same manner. The soil water storage at every time step is calculated as the algebraic sum of all water inflows and outflows in the given soil layer.

2.2. Heat balance components

The effective surface temperature is calculated by iterative solution of the heat budget equation. All heat budget components must be parameterized for this procedure. The shortwave and longwave incoming radiation are given as inputs for the surface block and land cover albedo must be prescribed. In the presence of snow, the surface albedo α is calculated according to Ghan et al. (1982), with our own modification for the vegetation height and bare soil wetness:

$$\alpha = \begin{cases} \alpha = \alpha_{sn} & , Z_{sn} > Z_{sncr} \\ \alpha = \sigma_f \alpha_v + \left(\alpha_s + (\alpha_{sn} - \alpha_s)\sqrt{\dfrac{Z_{sn}}{Z_{sncr}}}\right)(1-\sigma_f) & , Z_{sn} < Z_{sncr} \end{cases}$$
$$\alpha_s = \alpha_{sd}(1 - w_1/2)$$
$$Z_{sncr} = \max(0.1, h_v) \tag{12}$$

where α_{sn} is the snow albedo assumed to be equal to 0.75 (0.5 after the beginning of snowmelt), Z_{sn} is the snow cover depth in meters, α_v is the snow-free albedo of the vegetation, α_{sd} is the dry soil albedo and h_v is the vegetation height.

The latent heat flux is calculated by multiplying the total evapotranspiration rate calculated by Eq. (8) including evaporation of intercepted water and by latent heat of vaporization. The sensible heat flux to the atmosphere H is given by:

$$H = K_t(T_s - T_a) \tag{13}$$

where K_t is the turbulent exchange coefficient according to Eq. (9), T_s is the effective surface temperature, T_a is the air temperature.

The ground heat flux at the surface G is calculated as:

$$G = \lambda_1 \frac{T_s - T_{g1}}{d_1/2}$$
$$\lambda_1 = \lambda_d + \sqrt{W_{eff}}(\lambda_w - \lambda_d) \tag{14}$$

where λ_1 is the soil heat conductivity; T_{g1} is the temperature in the middle of the upper soil layer; W_{eff} is the total soil water content including liquid and frozen phases, and water unavailable for evaporation which is equal to the wilting point; λ_w and λ_d are the heat conductivities for wet and dry soil (i.e., with water content equal to field capacity and wilting point, respectively), they are taken as 2.5 and 0.25 W/(mK). Parameter λ_1 is multiplied by 1.3 if all soil water is frozen (Kalyuzhny and Pavlova, 1981); the coefficient is interpolated between 1 and 1.3 in dependence of the fraction of frozen water in the total soil water storage.

For rather rough surfaces, effective surface temperature T_s corresponds to the level of zero plane displacement plus roughness length, but it is suggested to be equal to the temperature of soil or snow surface. The conductive heat flux between the two soil layers and between the lower soil layer and the deep level of prescribed ground temperature are calculated similar to Eq. (14), with temperatures and effective soil wetness of corresponding layers. For the lower boundary temperature condition, the value equal to the average annual air temperature is taken, and is assumed to be constant during a year at the depth of 15 m. Special tests showed that this condition works well, and the same results were obtained if the measured soil temperatures at depths of 3–4 m were taken as the lower boundary condition. For evaluation of soil temperature changes, the total conductive heat flux to each layer is divided by corresponding heat capacity.

2.3. Soil water and snow phase transformations

If the temperature of any soil layer or its surface must be lower than 273.16 K according to the heat budget at the given level, and the soil water storage exceeds its minimal unfreezing value w_{unfr}, the temperature is kept at 0°C, and all energy needed for this is assumed to result from the freezing of soil water; the increase in soil frozen water content is calculated from the amount of available energy. The value w_{unfr} is determined according to Kalyuzhny and Pavlova (1981):

$$W_{\mathrm{unfr}} = 1.04\, F_{\mathrm{c}} - 0.06\, \rho_{\mathrm{s}} \tag{15}$$

where ρ_s is the soil density (specified as 2.67 kg/m^3).

The effective surface temperature of the landscape is assumed to be representative for the uppermost quarter of the upper soil layer, and freezing/melting of water within this sublayer is described by the changes in this temperature. It is suggested that the frozen part of soil water can occupy all pores, while the liquid phase cannot exceed the field capacity. So, if the freezing occurs in a soil layer, the ice particles are suggested to be placed firstly in the pores representing the excess of porosity above the field capacity. This leads to partial ‘drying’ of the soil, and water flow to the freezing layer from that with higher liquid water content can appear according to Eq. (11). This allows one to parameterize soil water flux from unfrozen to the frozen zone, the phenomenon known from experimental studies (e.g., Gusev, 1993).

The soil thawing takes place if there is frozen water within the given layer and its temperature exceeds 0°C. Again, the temperature is kept at this value, and the increase of liquid soil water content (terms d_{m1} or d_{m2} in Eq. (7)) is calculated from the quantity of energy excess. If the liquid soil water content exceeds field capacity, all water excess goes to runoff.

Non-intercepted precipitation is accumulated as snow if the air temperature is below 0°C. The density of fresh snow is suggested to be equal to 100 kg/m^3. Then it is increased 1.5 times for a period of 4 months due to an increase of snow mass (after reappearance of disappeared snow its density is equal to 100 kg/m^3 again). Moreover, the snow density increases due to refreezing of melted snow within snow cover. If the snow depth calculated from its water equivalent and density is larger than the vegetation height, parameters w_{in}, LAI, SAI and d_0 are set to zero (the intercepted precipitation is added to the snow mass), and z_0 is taken as for snow cover (0.0024 m). The conductive heat flux through snow cover is calculated in the same manner as in Eq. (14), while the heat conductivity is calculated by Yanson formula:

$$\lambda_{\mathrm{sn}} = \lambda_{\mathrm{sn0}}\left(6\rho_{\mathrm{sn}}^4 + 1.9\rho_{\mathrm{sn}} + 0.005\right) \tag{16}$$

where $\lambda_{\mathrm{sn0}} = 0.67$ W/(mK), ρ_{sn} is the snow density (kg/m^3) divided by 1000. Snow cover can melt if the temperature at one of its surfaces exceeds 273.16 K due to the balance of energy fluxes, or in case of warm rainfall and/or condensation of warm (i.e., warmer than 0°C) water vapour. In the latter cases, the energy flux to the snow surface is equal to the liquid water inflow multiplied by its heat capacity with the given air temperature. If the snow surface temperature before the warm rain or condensation was lower than 0°C, it is suggested that some amount of snow (according to the heat capacity of water inflow) is heated by the incoming water up to 0°C and melts, while the remainder of snow preserves its original surface temperature. Snow cover is assumed to be able to retain as much liquid water as 12% of its mass and all excess is directed to runoff. The liquid water within snow cover can appear due to snowmelt and/or rainfall. It can also refreeze if the temperature at the bottom of snow cover tends to be below the freezing point (snow surface temperature is not considered as water concentrates in the lower part of snow cover); this process increases the snow density.

3. Realization of the scheme

The scheme was involved in PILPS since 1993 and took part in the following PILPS experiments: (1a), (1b), (1c) (off-line runs with GCM-generated weather forcing), (2a) (with forcing data and validation fluxes measured in Cabauw, the Netherlands), (2c) (calculations for the large-scale basin of Arkansas and Red River), and (2d) (with forcing and validation data measured in Valdai, Russia, with the main emphasis on cold season processes). The scheme was also used for several additional stand-alone experiments (see, for example, Shmakin et al., 1993). The hydrological block of SPONSOR, along with the scheme of lateral redistribution of river water was coupled with a GCM of the Hydrometeorological Centre of Russia (Shmakin et al., 1996).

3.1. Improvement of soil water fluxes for Cabauw

Carrying out the mentioned experiments for different natural conditions allowed the improvement of the SPONSOR in many aspects. Several improvements were made after the analysis of results obtained during several PILPS stages, when it became obvious that the scheme produced results with some systematic bias. An example of such improvements is shown in Fig. 1 (freezing/melting processes parameterization was taken in its

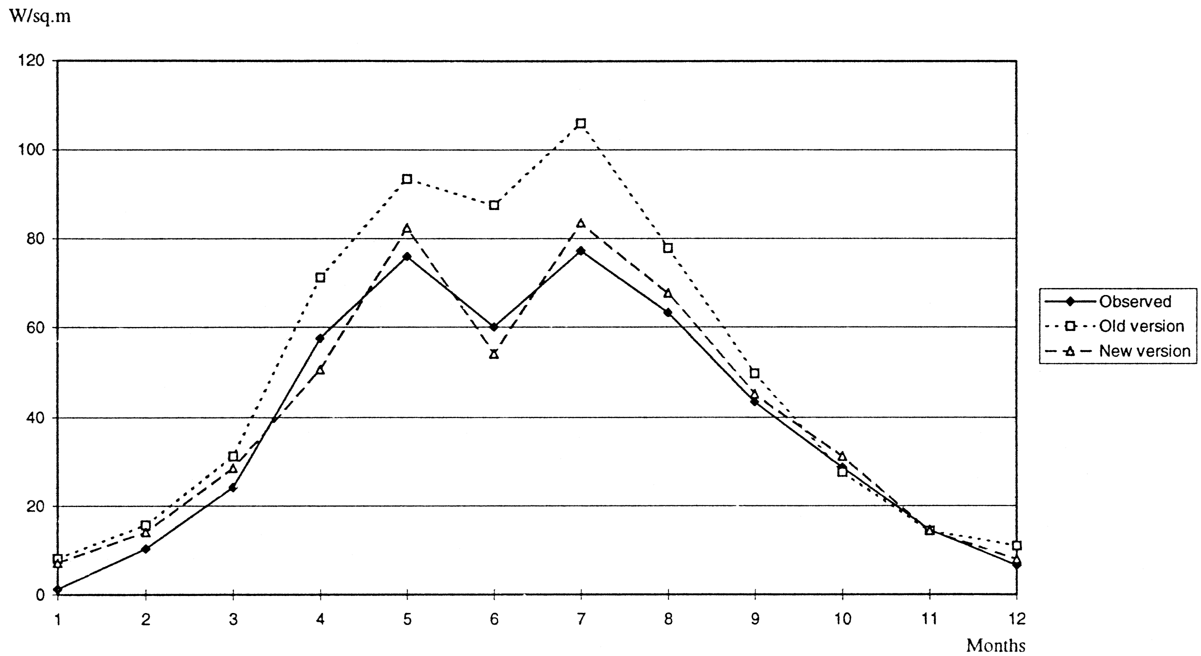

Fig. 1. Seasonal course of latent heat flux (W/m^2) in Cabauw according to measurements (Chen et al., 1997) and 2 versions of the SPONSOR scheme.

simplified version for those experiments, without a detailed description of the heat transfer within soil and snow; soil/snow heat flux was taken as a fraction of the deviation of net radiation from its annual average value). It is seen that the evapotranspiration rates were initially overestimated by SPONSOR scheme in PILPS Phase-(2a) experiments. So, after these runs, the new version of parameterization of water exchange between soil layers (based on Darcy equation) was incorporated into the scheme instead of simple diffusion terms. Moreover, the transpiration rate was multiplied by LAI/(LAI + SAI) in order to include the effect of green leaf fraction. After these changes, the results were found to be much more realistic for the Cabauw site. The description of soil water vertical movement by Darcy equation was especially important for improvement of the results, maybe due to continuous water inflow from saturated ground layers deeper than 1 m, and a more detailed description of this process was necessary. Also, the algorithm was transformed in order to make it possible to use any time step, depending on temporal resolution of the forcing parameters.

After PILPS Phase-(2c) (Arkansas–Red River) experiment, the soil field capacity was used as maximum volume of soil pores which can be occupied by soil water, instead of porosity. Incorporation of the term F_c/θ_w into Eq. (3) allowed us to diminish the runoff rate which was obviously overestimated by SPONSOR in PILPS Phase-(2c) runs. Moreover, the parameterization of soil water behaviour became more physically based due to this improvement. Also, the non-linear dependence of the surface runoff rate on the relative soil water content in this equation gave better results too.

3.2. Development of parameterization of cold seasons / regions processes: testing at permafrost site

For PILPS Phase-(2d) (Valdai) experiment, a new special block describing the soil water phase transformations was developed, along with a more detailed snow block. Before this experiment, snowmelt was parameterized according to the snow surface energy balance only, and the soil freezing/thawing was treated as simultaneous change for the whole upper soil layer. Now all main processes influencing the regime of soil heat/water transfer in the cold seasons/regions are parameterized in SPONSOR.

During preparation of the paper, the results of validation against Valdai data were not available to the author. To test the scheme under different natural conditions, the Kolyma water balance station (KWBS) (61°53′N and 147°43′E, Russia) was chosen. It is situated in the closed watershed (its area is nearly 27 km^2) of several streams belonging to the Kolyma river basin, one of the biggest in the north-eastern Siberia. Absolute heights within the watershed vary from 850 to 1700 m a.s.l. The station provides representative measurements of all the main components of water and heat budgets at several locations within the watershed (there is a network of such stations working by standard program in different geographical zones of the former USSR). KWBS is situated on the permafrost which depth varies from 50 to 350 m, melting for 0.3–3.0 m from the upper boundary in summer. Consequently, there is no water exchange within deeper ground layers. Soils are rather inhomogeneous; they include such different matters as clay and loam in the valleys, and gravel at the ranges. Vegetation represents sparse forest (larch, willow, alder, creeping cedar, polar birch, some species of bushes) and tundra patches (mosses, lichens and bare stones). The meteorological observations and the whole set of heat/water balance measurements are carried out near the lowest point of the station watershed (850 m a.s.l.), at horizontal terrace, surrounded from the northwest, north, east and southeast by low mountains (up to 1400 m a.s.l.) at the distance of 200–500 m (average view factor is 11°). The site is covered by sparse grass and moss, soil represents peat and loam, at deeper levels with gravel.

Only daily average meteorological parameters were available as forcing, so the time step was taken as 24 h. The data for 1976–1978 were used for the scheme validation, representing different conditions of precipitation and air temperature. Forcing for 1976 was used firstly for running the scheme until equilibrium, and then for evaluations for 1976 and two subsequent years. There were no available data on solar and infrared incoming radiation for each day, so these were evaluated using daily values of cloudiness, air temperature and air humidity by the radiation model developed earlier for mountainous areas (Krenke et al., 1991). A description is given in Appendix A. The effect of increasing incoming radiation due to reflection from surrounding slopes (this is especially important in springtime due to the high albedo of snow surface and intensive radiation fluxes) was taken into consideration by a special parameter determined according to Shmakin and Ananicheva (1991).

The observed heat and water balance components were available for validation with poor temporal resolution (10 days), for this reason the validation was made mainly against the soil temperature data. The soil heat conductivity was found to be an important parameter for modelling soil freezing/thawing regime. During the calibration, several versions of its parameterization (in particular, those developed by Ioffe and Revut (1959); McCumber and Pielke (1981)) were tested. Their equations for determination of the soil heat conductivity include rather complicated combinations of different types of functions and hence give unpredictable values in some particular cases (for example, in dry conditions). At the same time, in the mentioned papers the authors use experimental data obtained for a limited range of soil texture and wetness, so their equations hardly can be treated as universal. So, an original parameterization of the soil heat conductivity was developed using the data given by Oke (1978) and other authors who published a lot of data on soil conductivity. The square root function used for this parameterization (see Eq. (14)) was chosen as rather simple one; at the same time it reflects the main peculiarities of the conductivity dependence on soil water content.

Fig. 2 shows the comparison of soil surface temperature for the 1st and 15th day of each month for 1976–1978, calculated by the SPONSOR scheme and measured at KWBS (i.e., averaged for those days using measurements carried out every 6 h). In general, the curve of modelled temperature follows the observed one rather closely, except for some days. Calculated extrema (especially negative) differ from the measured values somewhat more (usually the model represents a smoother course of the surface temperature). It can be mentioned also that the dates of crossing of zero temperature values were simulated with high accuracy. Taking into account the strong annual amplitude of the surface temperature and large amount of hardly determinable

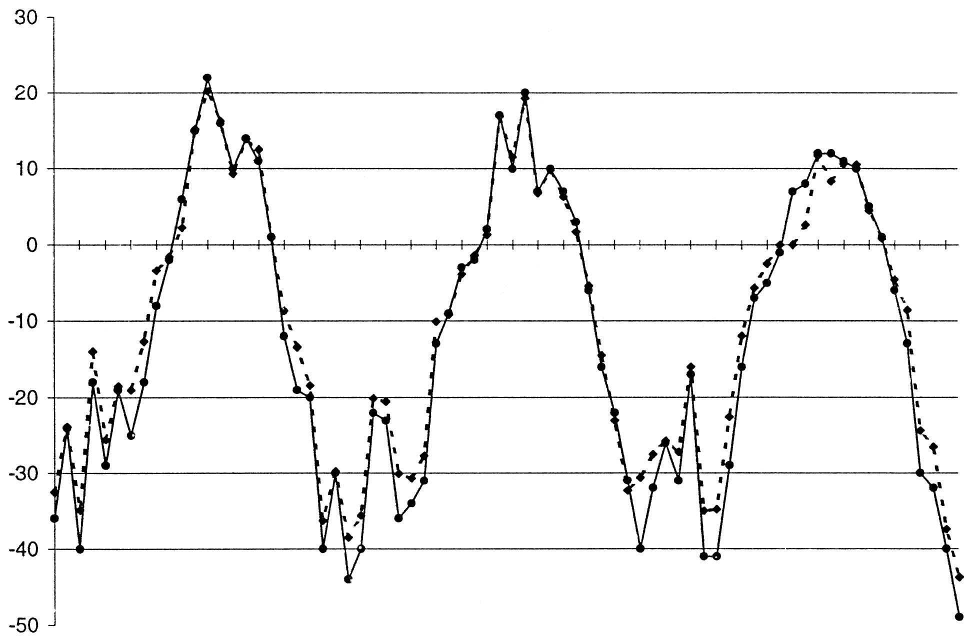

Fig. 2. Measured (solid line) and modeled (dashed line) soil (or snow) surface temperature (°C) at the Kolyma water balance station for 1st and 15th day of each month for 1976–1978. Each tick at the horizontal axis corresponds to a month.

factors (such as vertical inhomogeneity of soil properties, approximate values of main land cover parameters, etc.), one can consider the results as satisfactory.

Evaluation of deeper soil temperatures was not so successful by absolute values, though their temporal course was reproduced satisfactorily compared to the observed one (Fig. 3). The soil temperature was calculated at the levels of 5 and 55 cm, but taking into account the heat capacity of soil layers from the surface to 10 cm, and from 10 to 100 cm, respectively. So, the modelled temperature for these layers can be kept at 0°C for several days, until full freezing or thawing soil within the given layer. As seen from Fig. 3, the scheme overestimates the seasonal amplitude of the deep soil temperatures greatly, while the dates of crossing of 0°C are simulated accurately again. The latter also means that main features of water balance in the soil were modelled satisfactorily. In particular, total water storage within the soil was calculated quite accurately; otherwise the time necessary for soil freezing/thawing should noticeably differ from the observed one. More ‘extremal’ behaviour of the modelling results compared to observations can be explained by the fact that the heat conductivity of deeper soil layers is much less than the conductivity near the surface (there are many pores between particles of gravel which underlies the soil, and the air in pores decreases the heat conductivity). At the moment, the vertical inhomogeneity of soil thermal and hydraulic properties is not accounted for in the scheme. This will be one of the main directions of the future development of SPONSOR. Moreover, it is not clear, how to define the heat conductivity of so inhomogeneous matter like gravel. At the same time, the dates of soil freezing/thawing in deep layers were reproduced with quite good accuracy (Table 1). Again, it should be taken into account that the parameterization scheme provides soil temperature characterizing not only an actual level, but the whole soil layer also, while the measured data were obtained at some levels only. Good agreement between modelled and

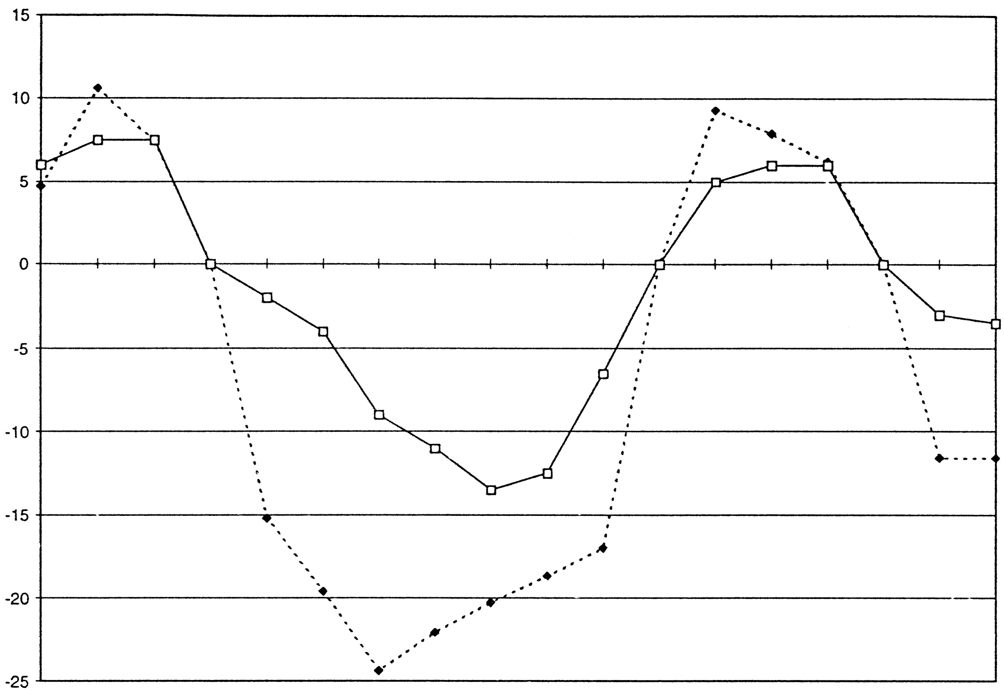

Fig. 3. Measured (averaged for 40 and 80 cm depth; solid line) and modeled (for the layer 10–100 cm depth; dashed line) soil temperature (°C) at the Kolyma water balance station for 1st day of each month for July, 1976–December, 1977. Each tick at the horizontal axis corresponds to a month.

Table 1
Dates with soil temperature equal to 0°C at the Kolyma water balance station according to model simulations and observations

	SPONSOR			Observed			
	Surface	0–10 cm	10–100 cm	Surface	20 cm	40 cm	80 cm
1976	May 23;	May 22–June 12;	May 30–June 29;	May 22;	May 20–23;	No obs.;	June 19;
	Sept. 21	Sept. 30	Oct. 2	Sept. 21	No obs.	Sept. 29	Oct. 3
1977	May 17;	May 16–June 3;	May 26–June 13;	May 17;	May 11;	May 11;	June 13;
	Sept. 17, 28–29	Sept. 17–Oct. 10	Sept. 9–Oct. 10	Sept. 18–20, 28–29	Oct. 9–12	Oct. 13	Sept. 30–Oct. 3
1978	May 23, 27;	May 23–24, 27–28, 30;	June 2–10;	May 26, 30	May 25–30;	No obs.;	June 28–29;
	Sept. 26	Sept. 26	Sept. 26	Sept. 26	Sept. 30	No obs.	Oct. 2

observed dates of soil freezing/thawing allows one to conclude that the scheme reproduces main changes of heat and water regime within the soil in permafrost zone quite satisfactorily.

4. Conclusions

The SPONSOR scheme was improved in many aspects since the beginning of its participation in PILPS in 1993. Moreover, several new blocks were developed in the scheme for participation in new PILPS stages and other experiments. The improvements allowed us to obtain better results in several cases, after obtaining the validation data. For the Cabauw site located in the extremely wet area in the Netherlands, it was important to describe accurately continuous water inflow to the deep soil layer from permanently saturated ground. After inclusion of Darcy equation in the scheme instead of simple diffusion terms, and limitation of transpiration by green leaf fraction in total leaf/stem area index, results were found to be much closer to the measured data (before that, SPONSOR overestimated the evapotranspiration rate). After participation in PILPS Phase-(2c) (Arkansas–Red River basin) experiment, some additional improvements were made in the soil water block too. For PILPS Phase-(2d) (Valdai) experiment, a new large block describing the processes, which are typical for cold seasons/regions, was incorporated into the parameterization scheme. Results of PILPS Phase-(2d) experiment were not available at the time of this paper preparation, but the scheme was tested against observations of the Kolyma water balance station located in the permafrost zone of the northeastern Siberia. This testing showed that SPONSOR in its new version is able to reproduce the main features of heat/water regime at the land surface in such exotic natural conditions like mountain valley in permafrost zone with high seasonal amplitude of soil temperatures. The calculated soil surface temperature is in a good agreement with observations. The dates of crossing of 0°C values by the soil temperature both at the surface and at deeper levels were in good agreement with the observations. Consequently, total soil water storages within the model layers were calculated with reasonable accuracy: the time for their freezing/melting corresponds well to the amount of water. Overestimation of the seasonal amplitude of deep soil temperatures as compared to the measurements possibly results from the too high value of heat conductivity taken for the gravel matter underlying the soil at the Kolyma station. It is rather difficult to find a method for estimation of this parameter for such porous media. Moreover, vertical inhomogeneity of thermal and hydraulic properties in the soil is not included in the parameterization scheme at the stage. Such improvement promises future progress in a more correct description of the processes of heat/water exchange within the soil.

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Appendix A. Radiation model used for SPONSOR validation against KWBS data

Direct and diffuse solar radiation are calculated as daily averages from corresponding hourly fluxes. Direct solar radiation at every hour S is calculated according to Bouguer's law in the form given by Pivovarova (1977):

$$S = \frac{(1-N)C_z(S_0/R^2)}{1+m(1-t_2^2)/2t_2^2}$$

$$C_z = \sin\varphi\sin\sigma + \cos\varphi\cos\delta\cos(90.2618(I-12))$$

$$m = \frac{2.0016P_a}{1013\left(Cz+\sqrt{Cz^2+0.003147}\right)} \tag{A1}$$

where N is the cloudiness (in a fraction of unity) averaged between the values for total and low-level clouds, Cz is cosine of the solar zenith angle, S_0 is the solar constant, R is the radius-vector of the Earth's orbit according to season, t_2 is the air transmittance for $m = 2$ at the sea level (m is optical mass of the atmosphere), φ is latitude, δ is solar declination (both in radians), I is hour number during a day, P_a is the atmospheric pressure. Parameter t_2 is determined for all USSR territory (for mountain regions with its vertical gradient) by Pivovarova (1977).

Diffuse solar radiation D is determined by Berlague formula:

$$D_c = 0.38Cz(S_0 - S_c)$$

$$D = (1-N)D_c + 0.34AN(S_c + D_c)$$

$$A = \begin{cases} 1, & \text{if } N < 0.95 \\ \exp(-0.03P\tau), & \text{if } N > 0.95 \end{cases} \tag{A2}$$

where S_c is the direct solar radiation for clear sky (i.e., $S/(1-N)$). Parameter A was included in Eq. (A2) for consideration of the effect of clouds thickness which well corresponds to the precipitation rate P. Total solar radiation is the sum of direct and diffuse solar radiation fluxes ($S+D$).

Infrared downward flux from the atmosphere to the surface is calculated according to Marks and Dozier (1979), with correction for cloudiness (Geiger, 1965, with modification for frosty conditions). The flux intensity I_d is determined by:

$$I_d = B\sigma\delta_a T_a^4$$

$$\delta_a = 1.24(P_a/1013)(E_0/T_0)^{0.143}$$

$$B = \begin{cases} 1+0.17N^2, & T_a > 273.16 \\ 1+0.4Nr, & T_a < 273.16 \end{cases} \tag{A3}$$

where σ is the Stephan–Boltzmann constant, δ_a is the atmospheric emissivity, E_0 and T_0 are the air water vapour pressure and the temperature recalculated for the sea level, B is the correction factor for cloudiness, r is the air relative humidity. For air temperatures higher then 0°C, B is taken according to Geiger (1965), and for frosty conditions according to our modification tested at several locations in Russia.The additional radiation income due to reflection of solar radiation from slopes and their emission to the bottom of valley, where KWBS is situated, is considered by inclusion of special coefficient A_r into the equations for absorbed radiation Q_{abs} and net radiation R_n:

$$\begin{aligned} Q_{abs} &= Q(1-\alpha)(1+A_r\alpha), \\ R_n &= Q_{abs} + I_d - (1-A_r)(I_s) \end{aligned} \tag{A4}$$

where α is the surface albedo, Q is the total solar radiation income, I_s is the surface infrared emission. Coefficient A_r is determined from the relief dissection parameters according to Shmakin and Ananicheva (1991) (for KWBS, A_r is equal to 0.17).

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