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The land surface parameterization scheme SWAP: description and partial validation

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Abstract

A full description of the latest version of the land surface parameterization scheme SWAP (Soil Water–Atmosphere–Plants) describing the interaction between the land surface and the atmosphere and being oriented on the coupling with atmospheric models is presented. The results of scheme validation on the long term at the local scale for the two sites (Cabauw, the Netherlands, and Petrinka, Russia) with different climatic and land-surface conditions are given. Analysis of the results shows that SWAP functions well under non-water-stress conditions, but less well during water stress. The possible reasons for discrepancies between the calculated and observed values are discussed. © 1998 Elsevier Science B.V. All rights reserved.

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1. Introduction

Parameterization of the land surface processes for the incorporation into atmospheric general circulation models (GCMs) is one of the most important problems in studying and modelling the climate system the lower boundary of which is the earth's surface. In solving this problem, the main difficulties are as follows: (i) the land surface is extremely inhomogeneous (from relatively flat and uniform deserts to regions with complex features of relief and vegetation), (ii) many processes of heat and water exchange occur in a complex and multifactor soil-vegetation/snow cover-atmosphere system at spatial scales below the spatial resolution of atmospheric models, i.e., at subgrid scales, (iii) usually it is difficult to provide the land-surface models with parameters for the whole globe and with adequate data for their validation under different natural conditions.

The above mentioned problems resulted in the appearance of a great number of so-called Soil-Vegetation–Atmosphere-Transfer (SVAT) schemes describing the land surface–atmosphere interactions with different degrees of detail and complexity and operating with more or less reliability in those regions for which they have been developed and calibrated. It is not possible to give a review of the existing SVAT schemes within the

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scope of the paper. It can only be pointed out that SVAT schemes range from very simple schemes without considering vegetation, for example, Budyko's 'bucket' model—the first parameterization scheme used in a GCM (Manabe, 1969), to complex models including a large (20–50) number of parameters, for example, SiB with 44 parameters (Sellers et al., 1986), for a detailed description of vegetated surfaces and their distribution within the territory under study. Simple schemes, as a rule, cannot give an adequate description of the land surface–atmosphere interactions. As to the complex models, the main emphasis is usually on the detailed description of the processes occurring during the warm season of a year whereas the cold season processes are treated very schematically. Besides that, such models are usually too large, poorly provided with parameters and require a lot of computer resources. The behavior of different SVAT schemes under the same natural conditions and model parameters, as one can see from the results of comparison of nearly 20 schemes within the frame of the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) initiated in 1992 (Henderson-Sellers et al., 1993), may quite differ from each other and from observations. Therefore, the problem of improvement of the existing land-surface schemes and development of new physically based and sufficiently rational (from the viewpoint of providence with parameters and consuming computer resources) models adequately describing the land surface–atmosphere interactions has not been solved yet. The present work represents an attempt to contribute a solution to this problem.

The aim of this paper is to present a new physically based model SWAP (Soil Water–Atmosphere–Plants), describing the interactions between the land surface and the atmosphere and being oriented on the coupling with atmospheric models. SWAP is a one-dimensional model, based on a system of physical–mathematical equations for the surface energy balance, the water balances of the canopy, the soil root zone and snow cover, and for the heat and water transfer within a soil-vegetation/snow cover-atmosphere system (SVAS). The distinctive features of SWAP are as follows: (i) it is a physically based model especially with respect to the description of the processes of the winter–spring period which are usually described very schematically; (ii) in solving the systems of equations, we tried to use analytical methods contrary to the usual practice of the application of numerical ones, that allowed us to avoid many problems associated with solving the numerical equations (such as instability, great consumption of the computer resources and the calculational time and so on); (iii) a relatively small number (18) of model parameters is needed most of which can be obtained from literature; (iv) when calculating the partition of non-intercepted rainfall into infiltration and surface runoff, the spatial variability of hydraulic conductivity at saturation is taken into account. Relatively simple mathematical formalism and application of the analytical approach make the model compact and sufficiently rational.

SWAP has participated in PILPS since Phase 2(a), (Chen et al., 1997). The first version, developed in 1994 and used in PILPS 2(a), is described in Gusev and Nasonova (1997b). Since then the model has been advanced in its representation of the evaporation from a bare soil, formation of snow cover and drainage, as well as the calculation of some soil and snow parameters. Here we give a detailed description of the latest version and the results of its partial validation on a local scale. For convenience all mathematical symbols are systematized in Appendix A.

2. Model description

As mentioned above, the approach used consists of a physically based modelling of heat and water transfer within the SVAS on a local scale throughout the year, with the main emphasis on application of analytical methods.

Using an analytical mathematical formalism usually requires the distinction of the main aspects of the phenomena and the neglect of insignificant details. The difficulty is that during different periods of a year, different hydrothermal processes in SVAS dominate, as a result of seasonal variation of heat and water exchange at the land surface. For this reason it is impossible to develop a general, simple and convenient analytical model for a whole year. At the same time it is possible to divide a year into several periods with prevalence of some kind of processes to develop a model for each period and then to link the models into one package allowing one to perform simulations continuously throughout a year and year by year. Such approach has been successfully applied when developing an agrometeorological model which describes the formation of soil water regime for various ecosystems of the steppe and forest steppe zones of the former Soviet Union (Gusev and Nasonova, 1997a).

In the present work, a year was divided into warm and cold seasons. For each season, a separate analytical model was developed. These two models were then linked as individual blocks into one general model, named SWAP. The scheme of SWAP is given in Fig. 1. The cold season model is used only in the case of the fulfillment of, at least, one of the following conditions: (1) the mean daily air temperature is below 0°C continuously during several days (here, not less than 7 days); (2) the land surface is covered by snow; (3) the soil freezing depth is greater than zero.

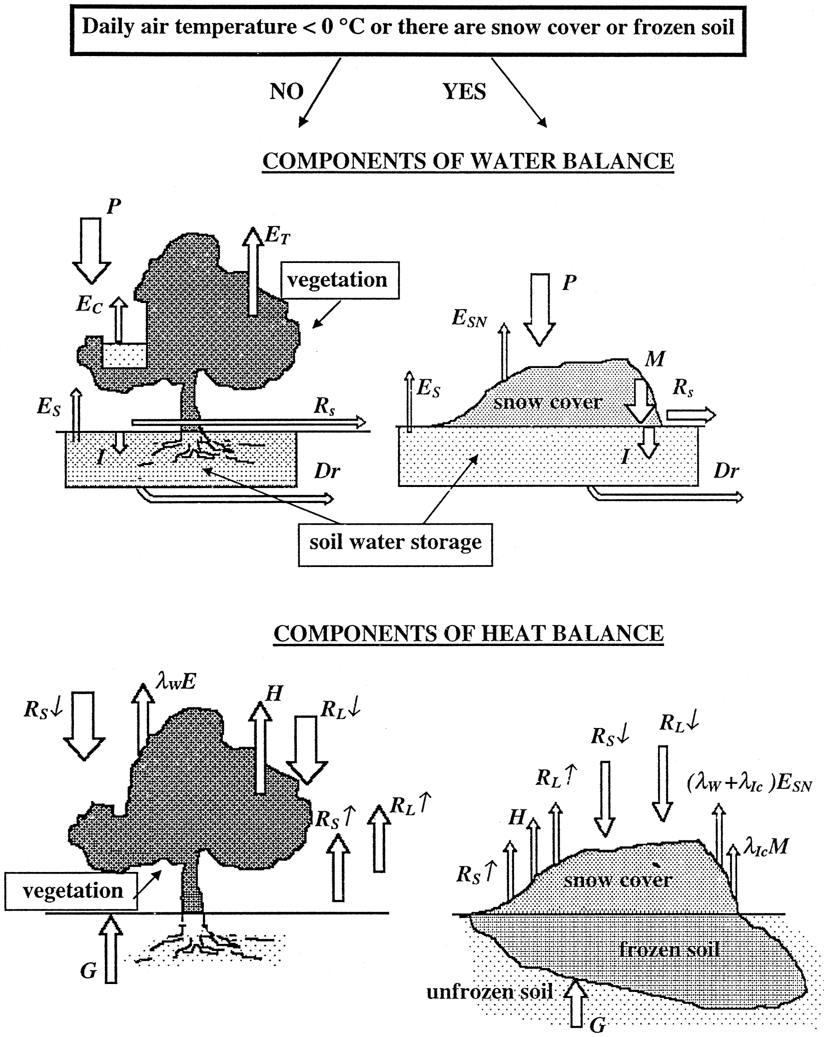

Fig. 1. Schematic representation of SWAP.

It should be noted that in SWAP there is a common technique for the whole year for the calculation of potential evaporation using atmospheric forcings from the lowest atmospheric layer of GCMs. That is why besides the two above mentioned blocks, SWAP includes a block for the calculation of potential evaporation.

2.1. Calculation of potential evaporation using atmospheric forcings from the lowest atmospheric layer of GCMs

On the basis of meteorological information from the lowest calculational layer of GCMs, used as atmospheric forcings, potential evaporation from the land surface (soil or snow) is calculated. Since the lowest layer can be situated at a height of an order of 10^1–10^2 m from the land surface, the turbulent fluxes and the surface temperature should be calculated taking into account the atmospheric stability. Here, for this purpose we use the main outcome of Monin–Obukhov (MO) similarity theory (Zilitinkevich, 1970; Zilitinkevich and Monin, 1971).

The system of equations for the calculation of potential evaporation is listed below.

(a) The equation for the surface energy balance which includes, along with the common components, the component $\lambda_{Ic} M$ representing the heat losses on snowmelt and soil thawing:

$$R_n = \lambda E + H + G + \lambda_{Ic} M \tag{1}$$

(b) Equations for the heat and water turbulent fluxes at the land surface, defined in terms of MO scaling parameters:

$$H = -T_* c_p æ \rho_s U_* \tag{2}$$

$$E = -q_* æ \rho_s U_* \tag{3}$$

(c) Equations for the vertical profiles of wind speed U, potential air temperature θ and air specific humidity q from the land surface to a reference level $(z = z_a)$, which result from the MO theory:

$$U_a æ = U_* \left[f_U((z_a - d_0)/L_*) - f_U(z_0/L_*) \right] \tag{4}$$

$$\theta_a - \theta_s = T_* \left[f_\theta((z_a - d_0)/L_*) - f_\theta(z_0/L_*) + f_2 \right] \tag{5}$$

$$q_a - q_s = q_* \left[f_q((z_a - d_0)/L_*) - f_q(z_0/L_*) + f_2 \right] \tag{6}$$

where:

$$L_* = \frac{U_*^2}{æ^2 g(T_*/T_a + 0.61 q_*)} \tag{7}$$

$$f_2 = 0.13 \left(\frac{z_0 U_*}{\nu} \right)^{0.45} \tag{8}$$

$$\theta = T(1000/p)^{R_a/c_p} = T(1000/p)^{0.288} \tag{9}$$

It should be noted that without the term f_2 Eqs. (5) and (6) would be valid from a height of roughness length z_0 to z_a, whereas incorporating f_2 allows one to expand the application of the equations to the land surface (Zilitinkevich and Monin, 1971).

(d) The equation for net radiation at the land surface:

$$R_n = (1 - \alpha) R_S \downarrow + R_L \downarrow - \sigma \varepsilon T_s^4. \tag{10}$$

To close the system of equations, the air adjacent to the surface is assumed to be saturated, i.e., the surface specific humidity q_s is equal to the saturated specific humidity $q_{sat}(t_s)$ at the surface temperature t_s. This assumption is reasonable because the system of equations is intended for the calculation of potential evaporation, i.e., evaporation from wetted surface. This allows one to use the Magnus formula for the air humidity at saturation (Khromov and Mamontova, 1974):

$$q_s = q_{sat}(t_s) = \frac{0.623 \cdot 6.1}{p} \exp\left(\frac{17.1 t_s}{235 + t_s}\right) \tag{11}$$

The universal functions, used in Eqs. (4)–(6), can be determined (under æ = 0.43) empirically as follows (Zilitinkevich, 1970):

$$f_U\left(\frac{z}{L_*}\right) = f_\theta\left(\frac{z}{L_*}\right) = f_q\left(\frac{z}{L_*}\right) = \begin{cases} \ln(z/L_*) + 10(z/L_*) & , \quad \text{if } z/L_* > 0 \\ \ln|z/L_*| & , \quad \text{if } -0.07 \le z/L_* \le 0 \\ 0.25 + 1.2(z/L_*)^{-1/3} & , \quad \text{if } z/L_* < -0.07 \end{cases} \tag{12}$$

Albedo is calculated by Ghan et al. (1982):

$$\alpha = \begin{cases} \alpha_s + (\alpha_{sn} - \alpha_s)\sqrt{0.1 \cdot \mathrm{Sn}} & , \quad \text{if Sn} < 10 \\ \alpha_{sn} & , \quad \text{if Sn} \ge 10 \end{cases} \tag{13}$$

here, the snow free surface albedo α_s we determine by:

$$\alpha_s = \alpha_{vg}(1 - \Psi(L)) + \alpha_{soil}\Psi(L) \tag{14}$$

where $(1 - \Psi(L))$ represents the fractional vegetation cover, $\Psi(L)$ can be approximated by the following expression:

$$\Psi(L) = \exp(-0.45L) \tag{15}$$

This expression was derived from experimental data on evaporation from different types of vegetation which were obtained by heat and water balance methods in different regions of the Former Soviet Union (FSU) (Budagovskiyi and Dzhogan, 1980; Budagovskiyi, 1986; Busarova and Shumova, 1987).

The potential evaporation depends, in particular, on the relation between net radiation R_n and ground heat flux G. During warm seasons, for a time step not less than one day, G is nearly an order of magnitude less than R_n (for example, Pavlov, 1979), i.e., G is relatively small in its contribution to potential evaporation (especially for the vegetated surfaces). For this reason we estimate G approximately using the empirical expression for G derived from the appropriate information given in Budyko (1956). For the cold season when the surface temperature is $t_s \le 0$ and in the absence of liquid fraction in snow cover ('dry snow') G can be calculated in the following manner:

$$G = \frac{t_s}{h/\lambda_1 + \xi/\lambda_2} \tag{16}$$

If snow cover has liquid water ('wet snow'), which freezes at $t_s \le 0$, we use the simple parameterization for G:

$$G = \lambda_1 \frac{t_s}{h/2} \tag{17}$$

In the case of soil thawing ($t_s > 0$), G is calculated as follows:

$$G = \frac{t_s}{\xi_{th}/\lambda_3} \tag{18}$$

Solution of the described system of equations gives us the turbulent heat flux from the wetted surface, i.e., potential evaporation $E_{PE} = E$, and the temperature of this surface. In doing so, the values of parameters z_0 and d_0 are specified for snow cover (if any) which allows us to calculate potential snow evaporation and the surface temperature of snow cover. In the absence of snow cover, z_0 and d_0 are taken not for the given type of vegetation, but for a hypothetical rough surface ($d_0 = 0$, $z_0 = 0.05$ m), that allows the calculation of potential evaporation from this surface. The latter is necessary to follow semi-empirical theory by Budagovskiyi (1981) which deals with potential evaporation from the land surface and potential transpiration, i.e., transpiration by full plant cover under unlimited water supply, when calculating the actual values of transpiration and soil evaporation during the warm season (see Section 2.2.3).

2.2. Description of the processes occurring during the warm season

This block includes a description of the following processes (Fig. 1): interception of precipitation by the canopy and evaporation of intercepted precipitation, partitioning non-intercepted precipitation into infiltration and surface runoff, transpiration, soil evaporation, water exchange at the lower boundary of the soil root zone, changes in the soil root zone water storage and formation of the surface energy balance.

2.2.1. Rainfall interception and evaporation of intercepted precipitation

To estimate the intercepted precipitation we use the bucket concept, i.e., the canopy is modeled simply as a bucket, or reservoir, of fixed capacity W_{cmax} that can be filled by precipitation P and emptied by evaporation E_{C}. While the bucket is not full it is assumed that all precipitation can be intercepted by the canopy. In this case, the change in the canopy water storage W_{c} is calculated using the following water balance equation:

$$\frac{\mathrm{d}W_{\mathrm{c}}}{\mathrm{d}\tau} = P - E_{\mathrm{C}}, \quad P_{\mathrm{s}} = 0, \qquad \text{if } W_{\mathrm{c}} < W_{\mathrm{cmax}} \tag{19}$$

When the bucket is full ($W_{\mathrm{c}} = W_{\mathrm{cmax}}$) the excess precipitation is assumed to leave the canopy and reaches the soil surface in the form of drainage P_{s}:

$$P_{\mathrm{s}} = P - E_{\mathrm{C}}, \qquad \text{if } W_{\mathrm{c}} = W_{\mathrm{cmax}} \text{ and } P - E_{\mathrm{C}} > 0 \tag{20}$$

The rate of evaporation of intercepted precipitation E_{C} is calculated under the following assumptions: (i) all intercepted water is concentrated at the surface of the canopy which is fully wetted, (ii) evaporation occurs at the rate which equals the potential evaporation E_{PE}. This allows us to simplify the calculational algorithm for E_{C}:

$$E_{\mathrm{C}} = \begin{cases} E_{\mathrm{PE}}[1 - \Psi(L)] & , \quad \text{if } W_{\mathrm{c}} > 0 \\ 0 & , \quad \text{if } W_{\mathrm{c}} = 0 \end{cases} \tag{21}$$

2.2.2. Calculation of infiltration rate with taking into account the spatial variability of hydraulic conductivity at saturation

The infiltration rate I is calculated on the basis of the relationship between the precipitation rate P_{s} and the infiltration curve $I_{\mathrm{P}} = I_{\mathrm{P}}(\tau)$, which characterizes the rate of infiltration under pressure. The infiltration curve can be derived from a modification of the Green–Ampt equation (Gusev, 1989), according to which the infiltration rate under pressure can be written as:

$$I_{\mathrm{p}} = k_0\left(1 + \frac{H_{\mathrm{k}}}{z_{\mathrm{w}}}\right) \tag{22}$$

Here, the maximum value of the soil effective capillary potential is determined as (Gusev, 1993):

$$H_{\mathrm{k}} = \int_{-\infty}^{\phi_0}\left(\frac{k}{k_0}\right)\mathrm{d}\phi \tag{23}$$

When calculating the integral in Eq. (23) one can use dependencies of k and ϕ on the volumetric soil moisture W. In particular, here, we use the relationship by Clapp and Hornberger (1978):

$$k = k_0 \left(\frac{W}{W_{\mathrm{sat}}} \right)^{2B+3}, \ \phi = \phi_0 \left(\frac{W}{W_{\mathrm{sat}}} \right)^{-B} \tag{24}$$

Combining Eq. (22) together with the water balance equation of wetted zone and supposing that soil water content before wetting front is approximately equal to field capacity, we can find a solution to the problem of the dynamics of $I(\tau)$. Simplification of this solution results in the following expression for the infiltration curve (which is similar in structure to the infiltration equation derived by Philip, 1957):

$$I_{\mathrm{P}} = k_0 + 0.5\sqrt{2 k_0 H_{\mathrm{k}} \rho_{\mathrm{w}} \Delta W \tau_{\mathrm{pr}}^{-1}} \tag{25}$$

where ΔW is the difference between soil porosity and field capacity. The time interval τ_{pr} is calculated from the beginning of current rainfall. If there is an interval, exceeding 24 h, without rainfall, τ_{pr} is set to be zero and for the next rainfall τ_{pr} should be calculated from the very beginning.

It should be noted that spatial variability of the coefficient of hydraulic conductivity at saturation k_0 is high. Thus, in Gusev (1993) it is pointed out that, according to observations, k_0 may differ within order of several meters. Therefore, even for a small area the spatial variability of k_0 should be taken into account when calculating water infiltration into the soil. To a first approximation, the spatial variability of k_0 can be characterized by its standard deviation σ_{k} from its mean value $\bar{k}_0$.

Given the function of distribution of k_0, one can calculate the mean value of the infiltration rate for heterogeneous areas. Here, as a first approximation, k_0 is assumed to be uniformly distributed on some interval from a to b. This assumption allows the derivation of analytical expression for the areally averaged infiltration rate:

$$\bar{I} = \begin{cases} \dfrac{2A_1}{3} k_0^{3/2} \Big|_a^{k_0*} + \dfrac{A_2}{2} k_0^2 \Big|_a^{k_0*} + P_{\mathrm{s}} A_2 k_0 \Big|_{k_0*}^{b} & , \quad \text{if } a < k_0* \le b \\ P_{\mathrm{s}} & , \quad \text{if } k_0* \le a \\ \dfrac{2A_1}{3} k_0^{3/2} \Big|_a^{b} + \dfrac{A_2}{2} k_0^2 \Big|_a^{b} & , \quad \text{if } k_0* > b \end{cases} \tag{26}$$

$$A_1 = \frac{0.5\sqrt{2 H_{\mathrm{k}} \rho_{\mathrm{w}} \Delta W \tau_{\mathrm{pr}}^{-1}}}{2\sqrt{3}\,\sigma_{\mathrm{k}}}, \ A_2 = \frac{1}{2\sqrt{3}\,\sigma_{\mathrm{k}}}, a = \bar{k}_0 - \sqrt{3}\,\sigma_{\mathrm{k}}, \ b = \bar{k}_0 + \sqrt{3}\,\sigma_{\mathrm{k}}$$

Here, k_0* is the hydraulic conductivity at saturation at which the infiltration rate, calculated by Eq. (25), is equal to the precipitation rate:

$$P_{\mathrm{s}} = 0.5\sqrt{2 k_0* H_{\mathrm{k}} \rho_{\mathrm{w}} \Delta W \tau_{\mathrm{pr}}^{-1}} + k_0* \tag{27}$$

The calculation of surface runoff R_{s} is based on the description of the Hortonian surface runoff mechanism, i.e., R_{s} occurs when non-intercepted precipitation P_{s} exceeds the soil infiltration rate. For inhomogeneous areas, R_{s} is absent on the patches where $k_0 \ge k_0*$, while for the other patches:

$$R_{\mathrm{s}} = P_{\mathrm{s}} - I_{\mathrm{P}} \tag{28}$$

so, the areally averaged value of R_{s} is equal to:

$$\bar{R}_{\mathrm{s}} = \begin{cases} P_{\mathrm{s}} - \bar{I} & , \quad \text{if } k_0 < k_0* \\ 0 & , \quad \text{if } k_0 \ge k_0* \end{cases} \tag{29}$$

The standard deviation σ_k can be approximately estimated by (Gusev, 1993):

$$\sigma_k = \begin{cases} 0.9 \cdot \bar{k}_0^{0.6} & , \quad \text{if } \bar{k}_0 > \sqrt{3}\,\sigma_k \\ \bar{k}_0/\sqrt{3} & , \quad \text{if } \bar{k}_0 \le \sqrt{3}\,\sigma_k \end{cases} \tag{30}$$

where $\bar{k}_0$ and σ_k are in mm/min.

2.2.3. Calculation of soil water balance components

The calculation of the soil water balance components is based on the water balance equation for the soil root zone, which can be written in the following form:

$$\rho_w h_r \frac{dW}{d\tau} = \bar{I} - E_T - E_S - Q \tag{31}$$

Simulation of the actual values of transpiration E_T and soil evaporation E_S is based on the semi-empirical theory of Budagovskiyi (1964, 1989), according to which:

$$E_T = \begin{cases} \beta_T E_{PT}(1 - \Psi(L)) & , \quad \text{if } W_c = 0 \\ 0 & , \quad \text{if } W_c > 0 \end{cases} \tag{32}$$

$$E_S = \beta_S E_{PS} \Psi(L) \tag{33}$$

where β_S and β_T are functions dependent on the soil hydrothermal regime and characterize, respectively, the deviation of the actual soil evaporation and transpiration rates from their potential values. The function β_T can be expressed by (Budagovskiyi, 1989):

$$\beta_T = \begin{cases} 0 & , \quad \text{if } W \le W_{wp} \\ \dfrac{W - W_{wp}}{W_{cr} - W_{wp}} & , \quad \text{if } W_{wp} < W < W_{cr} \\ 1 & , \quad \text{if } W \ge W_{cr} \end{cases} \tag{34}$$

$$W_{cr} = W_{wp} + 0.06 + 362.9 \cdot E_{PT} \tag{35}$$

Here, all empirical constants and equations were obtained on the basis of numerous field experiments in various agro- and natural ecosystems mainly in the steppe and forest-steppe zones of the FSU.

To calculate β_S we use a physically based model describing evaporation from the drying bare soil, developed and validated by Gusev (1997), according to which:

$$\beta_S = \frac{1}{1 + \zeta\delta}, \zeta = \left(\frac{\rho_w c_p \lambda \Delta}{\lambda_\delta + \rho_w c_p d_v} + \frac{c_p}{d_v} \right) \cdot \frac{D_S}{c_p + \lambda \Delta} \tag{36}$$

Here, the slope of the saturation air humidity curve at the air temperature at 2 m:

$$\Delta = \frac{d q_{sat}}{dt}\Big|_{t = t_2}$$

is calculated using the Magnus equation (Eq. (11)). Aerodynamic conductance between the soil and 2 m, D_S, is calculated by (Budagovskiyi, 1989):

$$D_S = \frac{0.76 \cdot 10^{-2} U_2}{1 + 0.82\sqrt{U_2}} \tag{37}$$

Coefficient of diffusivity for the soil water vapor d_v can be determined using the empirical relationship between d_v and the wind speed at 2 m (Gusev, 1997):

$$d_v = 0.66 \cdot 10^{-4}(W_{sat} - W_{mh})(0.24 + 0.42U_2) \quad (38)$$

Temporal variation of the thickness of the dry soil layer δ during a rainless period is described by the water balance equation for the dry layer:

$$(W - W_{mh})\rho_w \frac{d\delta}{d\tau} = E_S + Q_\delta \uparrow \quad (39)$$

The rate of water influx from underlying layers to the lower boundary of the dry layer $Q_\delta \uparrow$ is parameterized by (Gusev, 1997):

$$Q_\delta \uparrow = \frac{2B\phi_0 k_0}{h_r(B+3)} \left(\frac{W}{W_{sat}} \right)^{B+3} \quad (40)$$

In the case of rainfall, the upper soil layer becomes wet and δ is equal to zero.

Potential transpiration rate E_{PT} can be derived from the potential soil evaporation rate E_{PS}. The latter is equal to E_{PE}. As it was mentioned by Budagovskiyi (1989), the relationship between E_{PT} and E_{PS} is linear and intimate; that is:

$$E_{PT} = \omega \cdot E_{PS} \quad (41)$$

where the coefficient of proportionality ω ranges within a narrow interval and, in the average, it can be taken to equal 1.15. We have derived an empirical dependence of ω on the average linear size of foliage Lf on the basis of simulations of E_{PS} and E_{PT} for different ecosystems in the Kursk region (Russia) for the years 1956–1988:

$$\omega = 0.041/\sqrt{\mathrm{Lf}} + 0.78 \quad (42)$$

Now let us consider water exchange at the lower boundary of the soil root zone Q during the warm period. The upward flux of water is calculated empirically (Dzhogan, 1990):

$$Q = Q_1\Psi(L) + Q_2(1 - \Psi(L))$$

$$Q_1 = E_{PS}\exp(-n_1 h_g),\ Q_2 = E_{PT}\exp(-n_2(h_g - h_r)) \quad (43)$$

where n_1 and n_2 are the empirical coefficients dependent on the soil texture.

The downward flux of water Q usually occurs when the water content in the soil root zone calculated at the end of a current time step $W_{e,c}$ exceeds field capacity W_{fc}. The excess of water defines Q. In this case the dynamics of water content in the soil root zone W can be parameterized by the following equation which takes into account the soil hydraulic conductivity k:

$$\rho_w h_r \frac{dW}{d\tau} = k(W_{fc}) - k(W) \approx k'(W) \cdot (W_{fc} - W),\ k' = \frac{dk}{dW} \quad (44)$$

As seen from Eq. (44), when W tends to W_{fc}, water transfer in the soil comes to an end. An approximate integration of Eq. (44) under a small time step $\Delta\tau$ (on the assumption that the downward redistribution of water in the soil column occurs during $\Delta\tau$) gives:

$$W_{e,r} = W_{fc} + (W_{e,c} - W_{fc})\exp\left[-\frac{k'(W_{e,c}) \cdot \Delta\tau}{h_r \rho_w} \right] \quad (45)$$

where $W_{e,r}$ is the recalculated value of the water content in the soil root zone at the end of time step. Using the relationship between k and W (Budagovskiyi, 1955):

$$k = k_0 \left(\frac{W - W_{wp}}{W_{sat} - W_{wp}} \right)^4 \tag{46}$$

we can rewrite Eq. (45) in the following manner:

$$W_{e,r} = W_{fc} + (W_{e,c} - W_{fc}) \exp\left[- \frac{4 k_0 \cdot \Delta\tau}{(W_{sat} - W_{wp}) h_r \rho_w} \left(\frac{W_{fc} - W_{wp}}{W_{sat} - W_{wp}} \right)^3 \right] \tag{47}$$

In doing so, the value of Q, averaged over $\Delta\tau$, is calculated as:

$$Q = (W_{e,r} - W_{e,c}) h_r \rho_w / \Delta\tau \tag{48}$$

2.2.4. Recalculation of the surface temperature

When calculating the potential evaporation (Section 2.1) we have estimated the temperature of the wet surface. After the calculation of evapotranspiration, which includes E_C, E_S and E_T, it is possible to determine the actual surface temperature T_s and the appropriate sensible heat flux and longwave emission of the surface. Heat exchange between the land surface and the reference height (here, 2 m) can be expressed by:

$$H = \rho_2 c_p D (T_s - T_2) \tag{49}$$

Aerodynamic conductance between the land surface and the reference height, D, can be calculated taking into account the fraction of vegetation per unit area:

$$D = D_S \Psi(L) + D_{PL}(1 - \Psi(L)) \tag{50}$$

Here, the aerodynamic conductance between the soil and 2 m, D_S, is calculated by Eq. (37). The aerodynamic conductance between the canopy and 2 m, D_{PL}, treating heat exchange between the foliage and the adjacent air, as well as the turbulent heat transfer between the foliage and 2 m, can be represented in the following manner:

$$D_{PL} = \frac{1}{\dfrac{1}{D_0} + \dfrac{1}{L \cdot D_{T0}}} \tag{51}$$

$$D_0 = 5.1 \cdot 10^{-2} U_2^{2/3} \tag{52}$$

$$D_{T0} \approx 0.109 \frac{U_2^{2/3}}{\sqrt{\mathrm{Lf}}} \tag{53}$$

Empirical Eqs. (52) and (53) are taken from Budagovskiyi (1981), and Budagovskiyi and Lozinskaya (1976). The values of meteorological factors at 2 m, which are necessary for Eqs. (49)–(53), are obtained from the vertical profiles (see Section 2.1).

Substitution of Eqs. (10) and (49), and the actual evapotranspiration into the heat balance equation for the land surface (Eq. (1)) gives the actual surface temperature T_s.

2.3. Description of the processes occurring during the cold season

This block includes a description of the processes associated with the formation of the hydrothermal regime of snow cover and soil, namely, snow accumulation and snowmelt, evaporation from snow cover (if any) or soil, soil freezing and soil thawing, dynamics of the frozen and liquid water in the soil column (Fig. 1). The canopy is not considered for the cold season. In doing so, we adhere to the following philosophy. First, in the cold season transpiration is negligible if any. Second, the snow formation processes (accumulation, evaporation and melting) occurs qualitatively similar both on the soil surface and the canopy, but quantitative description of snow processes on the canopy is poorly developed. For this reasons we decided to exclude the canopy from consideration and supposed that all snowpack is located on the land surface.

2.3.1. Calculation of the water balance components of snow cover

Snowpack Sn consists of the hard Snh and liquid Snl fractions. Their dynamics can be determined from the water and heat balances of snow cover. If $t_s < 0°C$, then:

$$\frac{\mathrm{dSnh}}{\mathrm{d}\tau} = \begin{cases} P_h - E_{SN} - \dfrac{\mathrm{dSnl}}{\mathrm{d}\tau}, & \text{if Snl} > 0 \\ P_h - E_{SN}, & \text{if Snl} = 0 \end{cases} \tag{54}$$

$$\lambda_{Ic} \frac{\mathrm{dSnl}}{\mathrm{d}\tau} \approx 2\lambda_1 T_s / h \tag{55}$$

Here, P_h is the rate of precipitation in a hard form (precipitation is assumed to fall as snow, if air temperature is below 0°C); the snow evaporation rate E_{SN} is taken to be equal to potential rate, i.e., $E_{SN} = E_{PE}$; $h = \mathrm{Sn}/\rho_{sn}$ is the snow depth; the snow density ρ_{sn} can be calculated following Yosida (1955):

$$\rho_{sn}(\tau_i) = \rho_{sn}(\tau_{i-1})\left[1 + 0.1 \cdot \mathrm{Sn}(\tau_{i-1})\exp\left(0.08 \cdot \bar{t}_{sn} - 21\rho_{sn}(\tau_{i-1})\right)\right] \tag{56}$$

where $\bar{t}_{sn}$ is the temperature of snow cover averaged over the snow depth (°C), i is the number of the current time step equaled to 1 day. The dimensions of Sn and ρ_{sn} in Eq. (56) are cm and g cm^{-3}, respectively.

If $t_s \geq 0°C$, Snh and Snl are calculated from the following equations:

$$\frac{\mathrm{dSnh}}{\mathrm{d}\tau} = -M - E_{SN}, \tag{57}$$

$$\frac{\mathrm{dSnl}}{\mathrm{d}\tau} = \begin{cases} M + P, & \text{if Snl} < \mathrm{Snl}_{max} \\ -\dfrac{\varepsilon_{sn}}{1-\varepsilon_{sn}} \dfrac{\mathrm{dSnh}}{\mathrm{d}\tau}, & \text{if Snl} = \mathrm{Snl}_{max} \end{cases} \tag{58}$$

$$\mathrm{Snl}_{max} = \mathrm{Snh} \cdot \frac{\varepsilon_{sn}}{1-\varepsilon_{sn}} \tag{59}$$

In this case, the water, formed as a result of snowmelt, reaches the soil surface with the rate of water yield of snow cover V:

$$V = M + P - \frac{\varepsilon_{sn}}{1-\varepsilon_{sn}} \frac{\mathrm{dSnh}}{\mathrm{d}\tau} \tag{60}$$

This water is then partitioned into infiltration and surface runoff.

Maximum liquid water holding capacity of snow ε_{sn} is parameterized following Kuchment et al. (1983)

$$\varepsilon_{sn} = 0.11 \frac{\rho_w}{\rho_{sn}} - 0.11 \tag{61}$$

2.3.2. Calculation of soil freezing and soil thawing depths

The calculation of potential evaporation for the cold season (see Section 2.1) requires simultaneous solution of the problem of soil freezing or soil thawing. The soil freezing depth ξ is calculated when the surface temperature $t_s < 0°C$. The equation for the ξ was obtained by solving the equation for the rate of movement of the soil freezing front which, in its turn, is based on the heat balance equation of the frozen soil (Gusev, 1985, 1988):

$$\lambda * \frac{d\xi}{d\tau} = Q_f - Q_{un} \tag{62}$$

$$Q_f = -G, \quad Q_{un} = \frac{2\lambda_3 \tilde{t}}{\sqrt{2.25 \cdot \xi^2 + 12 a_3(\tau + \tau_0)} - 1.5 \cdot \xi}, \lambda * = \lambda_{Ic}\, \rho_w (W - u_{min}) + \frac{c_2 |t_g|}{2}, \quad t_g = \frac{t_s \xi}{\xi + h_\lambda},$$

$$h_\lambda = h \frac{\lambda_2}{\lambda_1}$$

The final calculational equation for the soil freezing depth ξ at the time step τ_i can be written as (Gusev, 1988, 1993):

$$\xi(\tau_i) = -h_\lambda(\tau_i) - \frac{Q_{un}(\tau_{i-1})\Delta\tau}{\lambda * (\tau_{i-1})} + \sqrt{\left[\xi(\tau_i - 1) + h_\lambda(\tau_i)\right]^2 - \frac{2\lambda_2 t_s \Delta\tau}{\lambda * (\tau_{i-1})} + \left(\frac{Q_{un}(\tau_{i-1})\Delta\tau}{\lambda * (\tau_{i-1})}\right)^2} \tag{63}$$

The soil thawing depth ξ_{th} can be obtained from the heat balance equation for thawing the frozen soil when $t_s > 0$ (Gusev et al., 1993):

$$\lambda * \frac{d\xi_{th}}{d\tau} = Q_{ts} \tag{64}$$

$$Q_{ts} = G, \quad \lambda * = \lambda_{Ic}\, \rho_w (W - u_{min})$$

When integrating Eq. (64) we come to the following expression for ξ_{th}:

$$\xi_{th} = \begin{cases} 0 & , \quad \text{if } h > 0 \\ \sqrt{\dfrac{2\lambda_3 \Delta\tau \sum t_s}{\lambda * (\tau_{i-1})}} & , \quad \text{if } h \le 0 \end{cases} \tag{65}$$

when $\sum t_s$ is the accumulated positive surface temperature from the beginning of the soil thawing period.

2.3.3. Calculation of the soil water balance components

The water balance equation of the soil root zone h_r for the cold season can be written in the following form:

$$h_r \rho_w \frac{dW}{d\tau} = I - E_S - Q. \tag{66}$$

Soil evaporation E_S is absent if the soil is covered by snow, in the absence of snow E_S is calculated in the same manner as for the warm season. The only distinction is that during the cold season the leaf area index is set to be zero.

The soil root-zone water content W is connected with the liquid water u and soil ice Ic:

$$W = u + \mathrm{Ic} \frac{\rho_w}{\rho_{Ic}} \tag{67}$$

The downward water flux at the lower boundary of the soil root layer Q is forming when the amount of unfrozen water at the end of time step $u_{e,c}$ exceeds field capacity W_{fc}. In this case the recalculated value of liquid water $u_{e,r}$ is calculated in the same manner as for the warm season, but accounts for the ice content in the frozen soil:

$$u_{e,r} = W_{fc} + (u_{e,c} - W_{fc})\exp\left[- \frac{4k_0 \cdot \Delta\tau}{(W_{sat} - W_{wp})h_r \rho_w} \left(\frac{u_{fc} - W_{wp}}{W_{sat} - W_{wp}} \right)^3 \frac{1}{(1 + 8\mathrm{Ic})^2} \right] \tag{68}$$

The rate of Q is parameterized by:

$$Q = h_r \rho_w (u_{e,r} - u_{e,c}) / \Delta\tau \tag{69}$$

The upward water flux, Q, resulted from the water migration from the underlying unfrozen zone to the frozen one, was neglected to simplify the calculations.

The infiltration rate I during snowmelt depends on u and Ic. At the beginning of the time step τ_i the ice content Ic can be calculated as:

$$\mathrm{Ic}_b = \left[W(\tau_{i-1}) - u_b \right] \frac{\rho_w}{\rho_{Ic}} \tag{70}$$

where the liquid water content at the beginning of the time step u_b is obtained from (Gusev, 1989):

$$u_b = \begin{cases} u_{min} & , \quad \text{if } t_g \le t* \\ u_{min} + \left(W(\tau_{i-1}) - u_{min} \right) \dfrac{Q_T}{Q_{T0}} & , \quad \text{if } t_g > t* \text{ and } Q_T \le Q_{T0} \\ W(\tau_{i-1}) & , \quad \text{if } t_g > t* \text{ and } Q_T > Q_{T0} \end{cases} \tag{71}$$

$$Q_{T0} = \lambda_{Ic} \rho_w \left(W(\tau_{i-1}) - u_{min} \right) \cdot (\xi - \xi_{th}) \tag{72}$$

$$Q_T = 2\lambda_2 \frac{|t*|}{0.25 \cdot (\xi - \xi_{th})} \Delta\tilde{\tau} \tag{73}$$

Here, Q_T is the amount of heat obtained by the frozen zone from its upper and lower boundaries and expended for ice melt; Q_{T0} is the amount of enthalpy which is necessary for melting all ice in the frozen zone; $\Delta\tilde{\tau}$ is the duration of the latest time interval with the soil temperature above $t* = -0.1$°C.

To a first approximation, the value of u at the time step τ_i can be estimated with no regard for capillary forces near the front of wetting (that is possible when the water infiltrates into sufficiently wet soil (Gusev, 1989):

$$u = \min\left[u_s + (W_{sat} - u_s) \cdot \left(\frac{V \cdot (1 + 8 \cdot \mathrm{Ic})^2}{k_0} \right)^{1/4}, W_{sat} - \mathrm{Ic} \right] \tag{74}$$

In doing so, the rate of infiltration is determined by:

$$I = \begin{cases} V & , \quad \text{if } u < W_{sat} - \mathrm{Ic} \\ k_0 \left(\dfrac{W_{sat} - \mathrm{Ic} - u_s}{W_{sat} - u_s} \right)^4 \dfrac{1}{(1 + 8 \cdot \mathrm{Ic})^2} & , \quad \text{if } u = W_{sat} - \mathrm{Ic} \end{cases} \tag{75}$$

The rate of surface runoff R_s is equal to the difference between the rate of water yield of snow cover V and the infiltration rate I:

$$R_s = V - I \tag{76}$$

3. Data and parameters

The model was validated on a local scale for the two sites with different climatic and hydrological conditions, namely, Cabauw (51°58′N, 4°56′E; the Netherlands) and Petrinka (36°11′E, 51°40′N; Russia). Cabauw is characterized by moderate maritime climate and permanently saturated deep soil (Chen et al., 1997). The latter means that the depth to water table is small (1 m in the calculations). Climate in Petrinka, situated in the forest-steppe zone, is characterized as moderate continental and the depth to water table exceeds 10 m.

Validation of the model requires atmospheric forcings, validation data and model parameters. The data set for the Cabauw site covered the entire year of 1987 and was provided by PILPS 2a organizers and is described in Chen et al. (1997). The Petrinka data set has been prepared by ourselves on the basis of standard records of the agrometeorological station Petrinka and the radiation station Kursk (situated in the vicinity of Petrinka, nearly 10 km away). In addition, we used the results of field observations in the Kursk region and corresponding data from literature.

3.1. Forcing data

The atmospheric forcing data to drive the model includes downward longwave and shortwave radiation, precipitation, air temperature, air specific humidity, wind speed and atmospheric pressure. For the Cabauw site, we used the forcing data provided at an hourly time step. For Petrinka, we have prepared the forcings with a time step of 1 day.

To prepare the forcings for Petrinka we used the daily values of air temperature at 2 m, air humidity at 2 m, total cloudiness, wind speed at 10 m and precipitation measured at the agrometeorological station Petrinka from 1955 to 1988. Downward shortwave $R_S\downarrow$ and longwave $R_L\downarrow$ radiation were computed by a radiation model developed and provided by A.B. Shmakin (Krenke et al., 1991). Its application requires air temperature and air humidity at 2 m, cloudiness, transparency of the atmosphere and solar declination. Transparency was taken from climatic maps presented in Pivovarova (1977). The main problem was to provide the radiation model with data on cloudiness which should include the information both on the amount and the type of clouds. Since the data on the clouds' types were unavailable, we had to modify the cloud attenuation formulae by means of calibration against daily values of $R_S\downarrow$ and R_n (unfortunately, $R_L\downarrow$ is not measured at the radiation stations in Russia) calculated from the standard records of the Kursk radiation station for the years of 1987 and 1988. As a result, for the calculation of $R_S\downarrow$ for cloudy conditions, we have incorporated a correction into the cloud attenuation formulae with the help of data on precipitation for the cases with total cloudiness greater than 0.9 to make some distinction between clouds with different thickness. For calculating $R_L\downarrow$ under cloudy conditions we used a cloud cover adjustment formulae (Krenke et al., 1991):

$$R_L\downarrow = R_{L0}\downarrow(1 + aN^b) \tag{77}$$

where N is the cloudiness (in a fraction of unit), $R_{L0}\downarrow$ is the clear sky downward longwave radiation, a and b are the empirical parameters equaled to 0.17 and 2, respectively (Krenke et al., 1991). In doing so, the calculated values of $R_L\downarrow$ for the cold period of a year (November–April) were much lower than that from literature. For this reason we had to calibrate a and b comparing the calculated by SWAP daily values of the net radiation with observations (for the years 1987 and 1988). As a result, for the cold period, b was found to be equal to 0.5 and a can be presented as a linear function of the air relative humidity (RH) with a coefficient of proportionality 0.4, i.e., $a = 0.4\mathrm{RH}$.

3.2. Validation data

Daily averaged surface turbulent heat and water fluxes, net radiation and surface radiative temperature from Cabauw for the entire year of 1987 were used for the Cabauw validation. The Petrinka observations for validation represent soil water storage in 1-m soil layer measured in three locations under grass (from 1969 to 1981), crops (from 1970 to 1979) and fallow (1955–1959) at the agrometeorological station Petrinka. These measurements were conducted once every 10 days basically during the snow free period of a year. The location with crops represents an agricultural field with a rotation of different crops from year to year (spring wheat in 1970, winter rye in 1971 and 1972, potato in 1973, winter wheat in 1974 and 1975, bare soil in 1976, spring wheat in 1977, clover from the autumn of 1977 to 1978, spring wheat in 1979, potato in 1980). The accuracy of soil water measurements in a 1 m soil layer on a scale under consideration is nearly 20 mm (Verigo and Razumova, 1973).

Table 1
The values of model parameters for Cabauw and Petrinka sites

Parameter	Cabauw	Petrinka
Soil properties		
h_g (m)	1	>10
h_r (m)	0.25 (with more than 95% of the total roots)	1
W_{sat} ($m^3\ m^{-3}$)	0.468	0.62
W_{fc} ($m^3\ m^{-3}$)	0.368	0.34
W_{wp} ($m^3\ m^{-3}$)	0.214	0.115
W_{mh} ($m^3\ m^{-3}$)	by Eq. (82)	$W_{wp}/1.34$ (Rode, 1955)
u_s ($m^3\ m^{-3}$)	W_{mh}	W_{mh}
u_{min} ($m^3\ m^{-3}$)	0.23	0.2
k_0 (kg $m^{-2}\ s^{-1}$)	3.43×10^{-3}	5.0×10^{-3}
B	10.4	7.9
ϕ_0 (m)	-0.045	-0.03
$\tilde{t}$ (°C)	6	7
λ_3 (W $m^{-1}\ K^{-1}$)	0.5	by Eq. (77)
λ_2 (W $m^{-1}\ K^{-1}$)	1.3 λ_3	1.3 λ_3
c_2 (J $m^{-3}\ K^{-1}$)	7.5×10^6	by Eq. (80)
c_{ds} (J $kg^{-1}\ K^{-1}$)	920	920
a_3 ($m^2\ s^{-1}$)	λ_3/c_3	λ_3/c_3
n_1 (in Eq. (43))	1.2	1.4
n_2 (in Eq. (43))	0.7	0.8
Vegetation and snow cover parameters		
Lf (m)	0.04	0.02
ρ_{sn} (kg m^{-3})	250	by Eq. (56)
λ_1 (W $m^{-1}\ K^{-1}$)	by Eq. (83)	by Eq. (84)
W_{cmax} (kg m^{-2})	0.1 mm l	0.1 mm l
Radiative transfer characteristics		
ε	1	1
α_{sn}	0.75	0.75
α_{veg}	0.25	0.23
α_{soil}	–	0.13
Aerodynamic transfer characteristics		
z_0 (m)	snow: 0.0024 snow free: 0.05	snow: 0.0024 snow free: 0.05
d_0 (m)	0	0
æ	0.43	0.43
Others		
z_a (m)	20	2

3.3. Model parameters

The values of parameters are given in Table 1. As it was mentioned above, for Cabauw, most of the parameters were provided by PILPS 2a organizers and described in Chen et al. (1997). Some of the missing parameters, required for the model SWAP, were found in the literature (the references are the same as for Petrinka, see below).

For Petrinka, W_{fc}, W_{wp} and W_{sat} were taken from the observations conducted at the agrometeorological station. Most other soil parameters for Petrinka were determined on the basis of literature while taking into account the type of soil and its texture. Thus, k_0 was obtained with the help of Clapp and Hornberger (1978) and Gusev (1981), B and ϕ_0 are presented in Gusev (1997); u_{min} was determined in accordance with Ivanov and Gavril'ev (1965), c_{ds} is from Palagin (1981), n_1 and n_2 are from Dzhogan (1990). Albedo of vegetation α_{veg} and soil α_{soil} are given in Budyko (1956). The monthly values of the leaf area index taken over many years for different crops in Petrinka were taken from Dzhogan and Lozinskaya (1993). The remaining parameters are calculated by equations presented in Appendix B.

4. Validation of the model SWAP at the local scale

4.1. Results from the Cabauw validation

The stand-alone land surface simulations using meteorological data for the year 1987 from Cabauw (grassland) as the atmospheric forcing were carried out with a 1-day time step (without any calibration). The scheme was run using the same forcings (for 1987) until reaching a dynamic equilibrium (Chen et al., 1997).

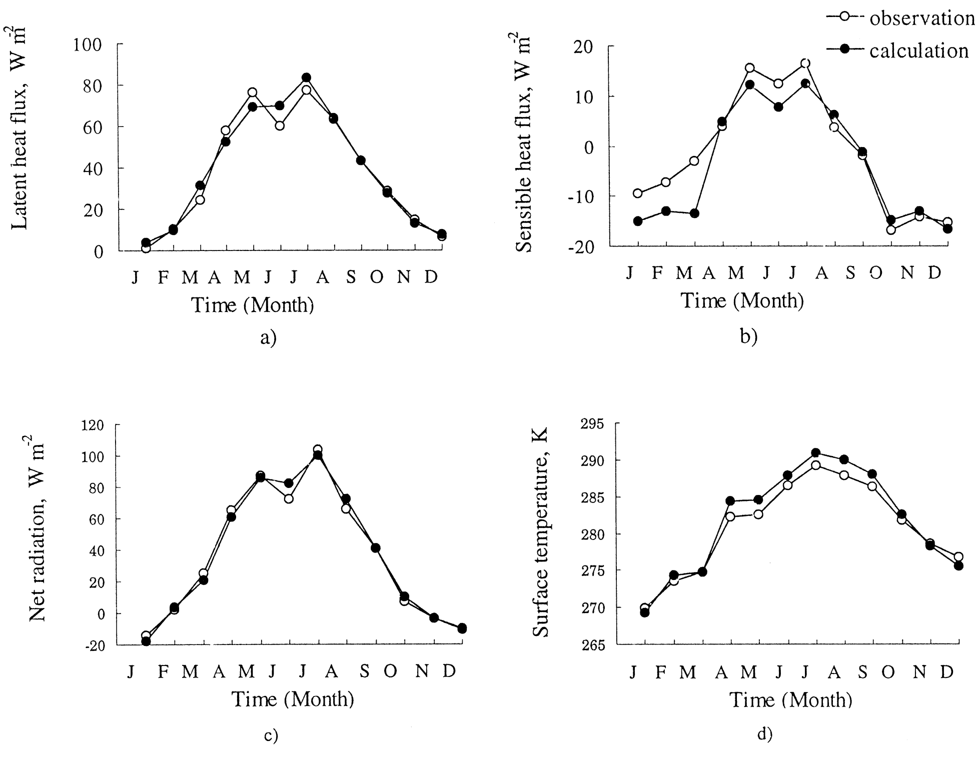

Fig. 2. Simulated monthly mean latent and sensible heat fluxes, net radiation and surface temperature compared with observation (Cabauw, the Netherlands, 1987).

Spinup time is equal to 2 years. As it was mentioned in Chen et al. (1997) the assumption that 1987 was an equilibrium year can induce errors, but it seems to be limited in very wet Cabauw conditions. The simulated annual, month and daily averaged surface turbulent fluxes, net radiation and surface temperature were compared with observations.

The simulated annual means are equal to -3.6 W m^{-2} for sensible heat flux (H), 39.8 W m^{-2} for latent heat flux (λE), 37.1 W m^{-2} for net radiation (R_n) and 281.7 K for surface temperature (T_s). The corresponding observed values are -1.3 W m^{-2} for H, 38.8 W m^{-2} for λE, 36.8 W m^{-2} for R_n and 280.8 K for T_s.

Fig. 2 illustrates the annual cycle of simulated monthly averaged turbulent fluxes, net radiation and radiative temperature compared with observations. The root-mean-square error of the calculation of monthly values is equal to 4.5 W m^{-2} for H, 4.7 W m^{-2} for λE, 4.2 W m^{-2} for R_n and 1.4 K for T_s, whereas the range of observation errors for monthly averages was estimated to be ± 5 W m^{-2} for sensible heat flux, ± 10 W m^{-2} for both latent heat flux and net radiation (Chen et al., 1997).

The root-mean-square error of the calculation of daily values is greater than that of monthly values and equal to 9.8 W m^{-2} for H, 14.9 W m^{-2} for λE, 13.9 W m^{-2} for R_n and 1.9 K for T_s. Fig. 3 shows differences between the simulated and observed daily values in terms of mean values and root mean square deviations.

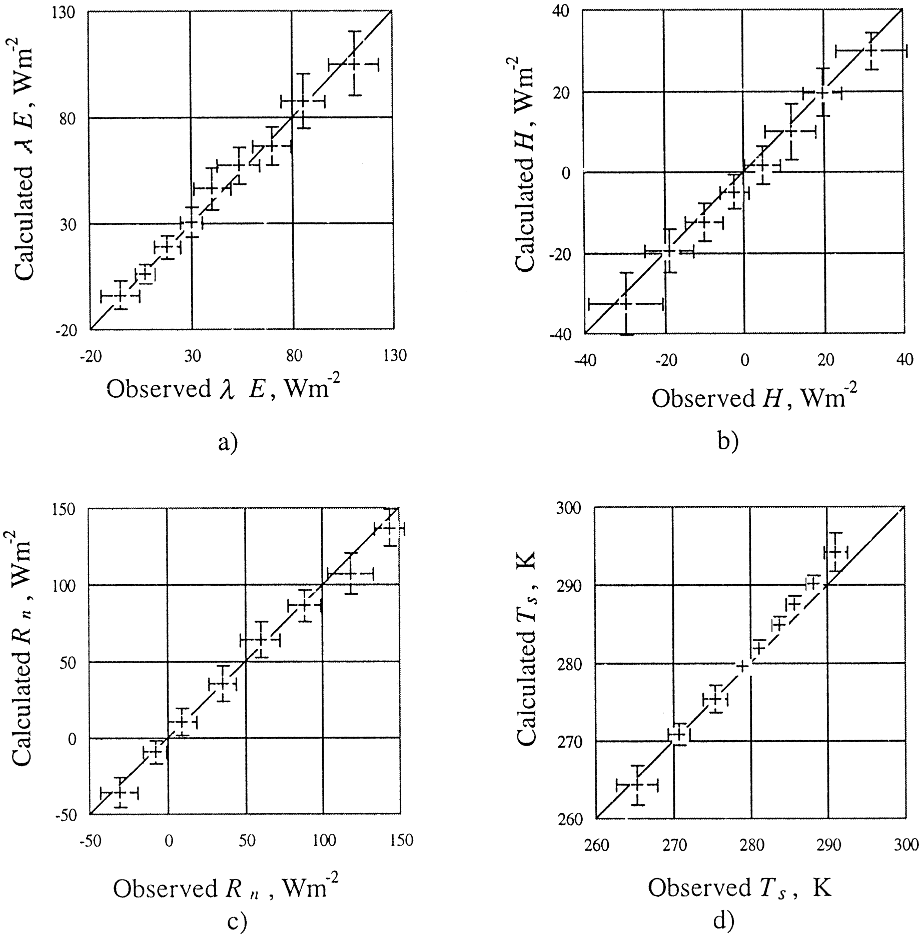

Fig. 3. Simulated daily mean latent λE and sensible H heat fluxes, net radiation R_n and surface temperature T_s vs. observed values in terms of mean values and standard deviations (Cabauw, the Netherlands, 1987).

Analysis of the obtained results shows that SWAP is quite suitable for parameterization of mid-latitude homogeneous grassland with a deep soil permanently saturated throughout a year.

4.2. Results from the Petrinka validation

The stand-alone simulations were carried out for fallow (for 5 years), grassland (for 13 years) and agricultural field (11 years) with a 1-day time step. Unlike the Cabauw, here we used the atmospheric forcings for several additional years (before the first year of the calculations) to run the model until equilibrium and then exclude these years from further analysis to reduce the error resulting from initialization. The values of simulated water storage in the 1-m soil layer were compared with the corresponding values measured at the Petrinka. The results of comparison of annual and interannual variation of simulated and measured values are depicted in Fig. 4. The root mean square error of the calculations is equal to 32, 33 and 41 kg m^{-2} for fallow, grassland and crop area, respectively. The coefficient of correlation between the calculated and measured data is equal to 0.72 for fallow and crops and 0.89 for grass. The latter shows that SWAP reproduce annual and interannual variation of soil water storage fairly well. The obtained errors can be explained by the following reasons: (a) errors in the calculation of downward shortwave and longwave radiation used as inputs; (b) the forcing data may be not fully consistent with the validation data because they are measured in different locations where some micrometeorological differences can take place, especially with respect to precipitation; (c) errors in determination of soil hydrophysical parameters, such as field capacity and plant wilting point, hydraulic conductivity at saturation, minimum amount of unfrozen water in the frozen soil; (d) the absence of real data on the leaf area index and utilization of monthly mean values of LAI generalized over many years and locations; (e) disadvantages of the parameterization scheme.

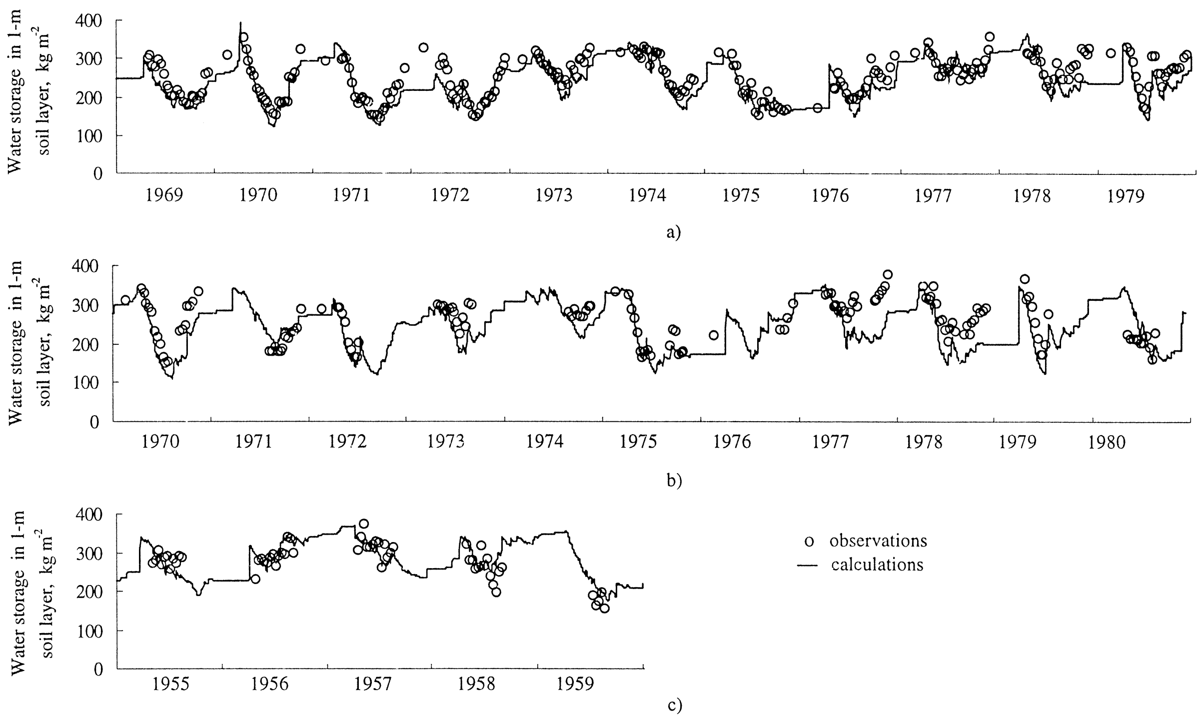

Fig. 4. Interannual variation of measured and simulated water storage in 1-m soil layer below (a) grass, (b) crops and (c) fallow (Petrinka, the Kursk region, Russia).

The major shortcoming of the scheme is connected with the treatment of evapotranspiration. Thus, SWAP does not take into account the impact of dead parts of plants and remains of crops after harvesting on soil evaporation and transpiration. The situation with dead vegetation is not so severe under wet conditions (for example, in Cabauw). This is not the case when water supply is limited that can result in sufficient amount of dead vegetation especially during the second half of the growing season. Such a situation often occurs with grass ecosystems in Petrinka and is not taken into account in SWAP. This can lead to overestimation of evapotranspiration and, respectively, underestimation of soil water storage for the mentioned period.

5. Concluding remarks

In this paper, we have presented a full description of the latest version of the land surface parameterization scheme SWAP and the results of its validation at the local scale for the two sites (Cabauw and Petrinka) with different climatic and land-surface conditions. Choice of these sites allows us to test the scheme for the same type of vegetation (grass) growing under different water-stress condition, namely, absence of water-stress factor at Cabauw and limited water supply at Petrinka. Besides that, using the Petrinka data set we had an opportunity to test SWAP for different types of the land surface: bare soil, grass and crops and on the long term, from 5 to 13 years.

Analyses of the results of validation shows that SWAP functions well under non-water-stress condition, but less well during water stress. However, the extent to which the discrepancies between the calculated and observed values depend on quality and completeness of input variables, validation data and parameters or on unknown scheme-specific errors. For this reason it is necessary to continue the validation of the scheme under a wide spectrum of natural conditions. This will require great efforts directed towards collecting appropriate data of high quality.

Acknowledgements

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Appendix A. List of mathematical symbols

a_3	thermal diffusivity of unfrozen soil (m^2 s^{-1})
B	‘B’-parameter of Clapp and Hornberger (1978)
c_p	specific heat capacity of air (J kg^{-1} K^{-1})
c_2	volumetric heat capacity of frozen soil (J m^{-3} K^{-1})
c_3	volumetric heat capacity of unfrozen soil (J m^{-3} K^{-1})
c_{ds}	specific heat capacity of dry soil (J kg^{-1} K^{-1})
c_{Ic}	specific heat capacity of ice (J kg^{-1} K^{-1})
c_w	specific heat capacity of water (J kg^{-1} K^{-1})
d_v	coefficient of diffusivity of soil water vapor (m^2 c^{-1})
d_0	zero plane displacement height (m)
D	aerodynamic conductance between the land surface and a reference height (here, 2 m) (m s^{-1})
D_{PL}	aerodynamic conductance between the canopy and 2 m (m s^{-1})
D_S	aerodynamic conductance between the soil and 2 m (m s^{-1})

D_{T0}	aerodynamic conductance between the foliage and adjacent air (m s^{-1})
D_0	aerodynamic conductance between the height of foliage and 2 m (m s^{-1})
Dr	drainage (kg m^{-2} s^{-1})
E	evapotranspiration rate ($E = E_T + E_S + E_C + E_{SN}$) (kg m^{-2} s^{-1})
E_C	evaporation rate of precipitation intercepted by a canopy (kg m^{-2} s^{-1})
E_{PE}	rate of potential evaporation from wetted surface (kg m^{-2} s^{-1})
E_{PS}	rate of potential evaporation from wetted soil (kg m^{-2} s^{-1})
E_{PT}	potential transpiration rate (kg m^{-2} s^{-1})
E_S	soil evaporation rate (kg m^{-2} s^{-1})
E_{SN}	evaporation rate from snow (kg m^{-2} s^{-1})
E_T	transpiration rate (kg m^{-2} s^{-1})
g	acceleration due to gravity (m s^{-2})
G	ground heat flux (W m^{-2})
H	sensible heat flux (W m^{-2})
H_k	maximum effective capillary potential of the soil (m)
h	snow depth (m)
h_g	depth to water table (m)
h_r	depth of the soil root zone (m)
I	infiltration rate (kg m^{-2} s^{-1})
I_P	infiltration rate under pressure (kg m^{-2} s^{-1})
Ic	soil ice content (m^3 m^{-3})
k	hydraulic conductivity of the moist soil (kg m^{-2} s^{-1})
k_0	hydraulic conductivity of the moist soil at saturation (kg m^{-2} s^{-1})
k_0*	hydraulic conductivity at saturation at which the infiltration rate is equal to the precipitation rate (kg m^{-2} s^{-1})
L	leaf area index (one-sided leaf area plus an area of longitudinal cross-sections of steams, branches and ears) (m^2 m^{-2})
L_*	Monin–Obukhov length (m)
Lf	effective linear leaf size (m)
M	the rate of snowmelt and soil thawing (kg m^{-2} s^{-1})
p	air pressure (mbar)
P	precipitation rate (kg m^{-2} s^{-1})
P_h	snowfall rate (kg m^{-2} s^{-1})
P_s	rainfall rate at the soil surface (kg m^{-2} s^{-1})
q_a	air specific humidity at a reference height (kg kg^{-1})
q_s	air specific humidity at the soil surface (kg kg^{-1})
q_{sat}	saturated air specific humidity (kg kg^{-1})
q_*	scaling parameter for air specific humidity (kg kg^{-1})
Q	water exchange at the lower boundary of the soil root zone (kg m^{-2} s^{-1})
Q_T	the amount of heat received by frozen zone from its upper and lower boundaries (J m^{-2})
Q_{T0}	the amount of enthalpy required for melting all ice in frozen zone (J m^{-2})
Q_{un}	conductive heat flux from underlying unfrozen zone to the front of freezing (W m^{-2})
$Q_\delta \uparrow$	upward flux of water at the lower boundary of dry soil layer (kg m^{-2} s^{-1})
R_a	specific gas constant of dry air (J kg^{-1} K^{-1})
$R_L \uparrow$	upward longwave radiation (W m^{-2})
$R_L \downarrow$	downward longwave radiation (W m^{-2})
R_n	net radiation (W m^{-2})
$R_S \uparrow$	upward shortwave radiation (W m^{-2})

$R_S\downarrow$ downward shortwave radiation (W m^{-2})
R_s surface runoff (kg m^{-2} s^{-1})
Sn snowpack (kg m^{-2})
Snh hard fraction of snow (kg m^{-2})
Snl liquid fraction of snow (kg m^{-2})
Snl_{max} maximum liquid water storage in snow cover (kg m^{-2})
t_g soil surface temperature (°C)
t_s air temperature at the land surface (°C)
t_2 air temperature at 2 m (°C)
$t*$ temperature parameter ($= -0.1$) in relationship between the amount of unfrozen water in frozen soil and its temperature (°C)
$\tilde{t}$ constant soil temperature at the depth where seasonal temperature variations are damped out (°C)
T_a air temperature at a reference height (K)
T_s air temperature at the land surface (K)
T_* scaling parameter for air temperature (K)
T_2 air temperature at 2 m (K)
u the amount of unfrozen water in wetted zone of frozen soil (m^3 m^{-3})
u_b liquid water content in frozen soil at the beginning of time step (m^3 m^{-3})
u_e liquid water content in frozen soil at the end of time step (m^3 m^{-3})
u_{min} minimum amount of unfrozen water in frozen soil (m^3 m^{-3})
u_s static soil water (m^3 m^{-3})
U_a wind speed at a reference height (m s^{-1})
U_2 wind speed at 2 m (m s^{-1})
U_* scaling parameter for wind speed (friction velocity) (m s^{-1})
V the rate of water yield of snow cover (kg m^{-2} s^{-1})
W total soil water content in a rooting zone (m^3 m^{-3})
W_c the amount of moisture in the canopy interception reservoir (kg m^{-2})
W_{cmax} the maximum amount of moisture on the canopy (interception capacity) (kg m^{-2})
W_{cr} critical total soil moisture (m^3 m^{-3})
$W_{e,c}$ soil water content calculated at the end of a time step (m^3 m^{-3})
$W_{e,r}$ soil water content recalculated at the end of a time step, if $W_{e,c} > W_{fc}$ (m^3 m^{-3})
W_{fc} soil water content at the field capacity (m^3 m^{-3})
W_{mh} maximum hygroscopicity (m^3 m^{-3})
W_{sat} soil porosity (m^3 m^{-3})
W_{wp} soil water content at the plant wilting point (m^3 m^{-3})
z_a reference height (m)
z_w depth to which water penetrates into soil from the surface (m)
z_0 roughness length (m)
α albedo (dimensionless)
α_s albedo of snow free surface (dimensionless)
α_{sn} snow albedo (dimensionless)
α_{soil} bare soil albedo (dimensionless)
α_{vg} vegetation albedo (dimensionless)
β_T availability of water in the root zone for transpiration (dimensionless)
β_S availability of water in the root zone for evaporation (dimensionless)
δ the thickness of drying soil layer (m)
Δ the slope of the saturation air humidity curve at air temperature at 2 m (kg kg^{-1}°C^{-1})
$\Delta\tau$ time step of calculations (s)

$\Delta\tilde{\tau}$ duration of the last time period with soil temperature above -0.1°C (s)
ε thermal emissivity (dimensionless)
ε_{sn} maximum liquid water holding capacity of snow (dimensionless)
θ potential air temperature (K)
θ_a potential air temperature at a reference height (K)
θ_s potential air temperature at the land surface (K)
æ the von Karman constant (dimensionless)
λ $\lambda_w + \lambda_{Ic}$ (J kg^{-1})
λ_w latent heat of vaporisation of water (J kg^{-1})
λ_{Ic} heat of ice fusion (J kg^{-1})
λ_1 thermal conductivity of snow (W m^{-1} K^{-1})
λ_2 thermal conductivity of frozen soil (W m^{-1} K^{-1})
λ_3 thermal conductivity of unfrozen soil (W m^{-1} K^{-1})
λ_δ thermal conductivity of dry soil layer (W m^{-1} K^{-1})
ν kinematic coefficient of air viscosity (m^2 s^{-1})
ξ soil freezing depth (m)
ξ_{th} soil thawing depth (m)
ρ_s air density at the land surface (kg m^{-3})
ρ_{ds} density of dry soil (kg m^{-3})
ρ_{hs} density of hard fraction of a soil (kg m^{-3})
ρ_{sn} snow density (kg m^{-3})
ρ_w water density (kg m^{-3})
ρ_2 air density at 2 m (kg m^{-3})
σ Stefan–Boltzmann constant (W m^{-2} K^{-4})
σ_k standard deviation of the hydraulic conductivity of the moist soil (kg m^{-2} s^{-1})
τ time (s)
τ_0 parameter characterising the influence of a period prior to freezing on freezing depth formation freezing, here, it is taken to be 10 days (Gusev, 1993)
τ_{pr} duration of the period which is used in infiltration curve and calculated from the beginning of precipitation (s)
ϕ matric potential of soil water (m)
ϕ_0 matric potential of soil water at saturation (m)

Appendix B. Some important soil and snow characteristics used in SWAP

Thermal conductivity of the unfrozen soil is calculated by (Vershinin et al., 1959):

$$\lambda_3 = \left[2.1\rho_{ds}^{(1.2-0.02W)}\exp\left(0.007(W-20)^2\right) + \rho_{ds}^{0.8} + 0.02W\right]\cdot(0.2+0.01W)\cdot\rho_{ds}\cdot 0.001 \quad (77')$$

$$\rho_{ds} = (1 - W_{sat})\,\rho_{hs},\ \rho_{hs} = 2.65 \quad (78)$$

Here, the soil moisture W is in g g^{-1}, λ_3 is in cal cm^{-1} s^{-1} (°C)$^{-1}$, ρ_{ds} and ρ_{hs} are in g cm^{-3}.

Volumetric heat capacity of the unfrozen soil is given by:

$$c_3 = c_{ds}\,\rho_{ds} + c_w\,\rho_w W \quad (79)$$

volumetric heat capacity of the frozen soil is calculated as:

$$c_2 = c_w\,\rho_w\,u + c_{Ic}\,\rho_{Ic}\mathrm{Ic} + c_{ds}\,\rho_{ds} + (u_{min} - u_{min5})\,\lambda_{Ic}\,\rho_w/\Delta t_5,\ \Delta t_5 = 5°\mathrm{C} \quad (80)$$

where the minimum amount of unfrozen water in the frozen soil $u_{\min 5}$ is calculated empirically (Kalyuzhnyi and Pavlova, 1981):

$$u_{\min 5} = 0.94 W_{\mathrm{wp}} + 0.017 \rho_{\mathrm{ds}} \tag{81}$$

where W_{wp} and $u_{\min 5}$ are in g g^{-1}, ρ_{ds} is in g cm^{-3}.

Thermal conductivity of the dry soil layer λ_δ is calculated by Eq. (77) in which maximum hygroscopicity W_{mh} is substituted instead of W. The value of W_{mh}, in its turn, can be determined by:

$$W_{\mathrm{mh}} = W_{\mathrm{sat}} \left(\frac{\phi_0}{\phi_1} \right)^{1/B}, \phi_1 = -1000 \ldots -3000\,\mathrm{m} \tag{82}$$

Snow thermal conductivity λ_1 can be calculated, for example, using an empirical relation by Abels (1892) rewritten in SI units as:

$$\lambda_1 = 2.9 \cdot 10^{-6} \rho_{\mathrm{sn}}^2 \tag{83}$$

or by Yanson (Gusev, 1993):

$$\lambda_1 = 10^{-3} \left[(0.05 + 1.9 \rho_{\mathrm{sn}}(\tau_{i-1}) + 6.0 \rho_{\mathrm{sn}}(\tau_{i-1})^4 \right] \tag{84}$$

where λ_1 is in cal cm^{-1} s^{-1} (°C)$^{-1}$, ρ_{sn} is in g cm^{-3}.

References

Abels, H., 1892. Beobachtungen der taglichen periode der temperature im schnee und bestimmum des warmeleitung svermogens des schnees als function seiner dichtigkeit. Rep. Meteorolog, Bd. XVI 1, 1–53.

Budagovskiyi, A.I., 1955. Vpityvanie Vody v Pochvu (Infiltration of water into a soil). Izdatel'stvo Akademii Hauk, Moscow, USSR (in Russian).

Budagovskiyi, A.I., 1964. Isparenie Pochvennoi Vlagi (Evaporation of soil water). Nauka, Moscow, USSR (in Russian).

Budagovskiyi, A.I., 1981. Isparenie Pochvennykh Vod (Evaporation of soil water). In: Nerpin, S.V. (Ed.), Phizika Pochvennykh Vod (Physics of Soil Water). Nauka, Moscow, USSR, pp. 13–95 (in Russian).

Budagovskiyi, A.I., 1986. Utochnenie modelei ispareniya pochvennykh vod. [Refinement of models for evaporation of soil water] Vodnye Resursy 5, 58–69 (in Russian).

Budagovskiyi, A.I., 1989. Principles of the method of calculating the duty of water and irrigation regimes. Water Resour. 16 (1), 27–35.

Budagovskiyi, A.I., Dzhogan, L.Y., 1980. Ways of successive refinement of methods of calculating the duty of water. Water Resour. 7 (6), 483–500.

Budagovskiyi, A.I., Lozinskaya, E.A., 1976. Sistema yravneniyi teplo-i vlagoobmena v rastitel'nom pokhrove. [A system of equations for heat and water exchange within a canopy] Vodnye Resursy 1, 78–94 (in Russian).

Budyko, M.I., 1956. Heat Balance of the Earth's Surface. United States Weather Bureau, Office of Climatology, Washington, DC.

Busarova, O.E., Shumova, N.A., 1987. Biometric characteristics of field crops and their use for evaporation calculations. Water Resour. 14 (2), 194–198.

Chen, T.H., Henderson-Sellers, A., Milly, P.C.D., Pitman, A.J., Beljaars, A.C.M., Polcher, J., Abramopoulos, F., Boone, A., Chang, S., Chen, F., Dai, Y., Desborough, C.E., Dickinson, R.E., Dumenil, L., Ek, M., Garratt, J.R., Gedney, N., Gusev, Y.M., Kim, J., Koster, R., Kowalczyk, E.A., Laval, K., Lean, J., Lettenmaier, D., Liang, X., Mahfouf, J.-F., Mengelkamp, H.-T., Mitchell, K., Nasonova, O.N., Noilhan, J., Robock, A., Rosenzweig, C., Schaake, J., Schlosser, C.A., Schulz, J.-P., Shao, Y., Shmakin, A.B., Verseghy, D.L., Wetzel, P., Wood, E.F., Xue, Y., Yang, Z.-L., Zeng, Q., 1997. Cabauw experimental results from the project for intercomparison of land-surface parameterization schemes (PILPS). J. Climate 10 (6), 1194–1215.

Clapp, R.B., Hornberger, G.M., 1978. Empirical equations for some soil hydraulic properties. Water Resour. Res. 14, 601–604.

Dzhogan, L.Y., 1990. Isparenie c Oroshaemykh Poleyi Sredneyi Azii (Evaporation from Irrigated Fields of the Middle Asia). Nauka, Moscow, USSR (in Russian).

Dzhogan, L.Y., Lozinskaya, E.A., 1993. A method for averaging leaf relative area in mesoscale assessment of heat and moisture exchange between the underlying surface and the atmosphere. Water Resour. 20 (6), 612–619.

Ghan, S.J., Lingaas, J.V., Schlesinger, M.E., Mobley, R.L., Gates, W.L., 1982. A documentation of the OSU 2-level atmospheric general circulation model. Report 35. Climatic Research Institute, Oregon State University, Corvallis.

Gusev, Y.M., 1981. Eksperimental'noe issledovanie napornogo vpityvaniya (Experimental investigation of infiltration under pressure). In: Nerpin, S.V. (Ed.), Phizika Pochvennykh Vod. (Physics of Soil Water). Nauka, Moscow, USSR, pp. 195–205 (in Russian).
Gusev, Y.M., 1985. Priblizhonnyi chislennyi raschet glubiny promerzaniya pochvy. [Approximate numerical calculation of soil freezing depth] Meteorologiya i gidrologiya 6, 94–102 (in Russian).
Gusev, E.M., 1988. Simplified calculation of soil freezing depth with consideration of water migration to the freezing boundary. Water Resour. 15 (1), 9–17.
Gusev, E.M., 1989. Infiltration of water into soil during melting of snow. Water Resour. 16 (2), 108–122.
Gusev, E.M., 1993. Formirovanie Rezhima i Resursov Pochvennykh Vod v Zimne-Vesenniyi Period (Formation of Regime and Resources of Soil Water during Winter–Spring Period). Fizmatlit Publishing, Nauka, Moscow, Russia (in Russian).
Gusev, Y.M., 1997. Isparenie vody prosykhayuscheyi pochvoyi (Evaporation of water from drying soil). Vodnye Resursy (in press) (in Russian).
Gusev, Y.M., Nasonova, O.N., 1997a. Modelling annual dynamics of soil water storage for agro- and natural ecosystems of the steppe and forest-steppe zones on a local scale. Agric. Forest Meteorol. 85, 171–191.
Gusev, E.M., Nasonova, O.N., 1997b. Parametrizatsiya teplo- i vlagoobmena na poverkhnosti sushi pri sopryadzenii gidrologicheskhikh i klimaticheskhikh modelei (Parameterization of heat- and water exchange at the land surface for coupling hydrological and climatic models). Vodnye Resursy (in press) (in Russian).
Gusev, E.M., Busarova, O.E., Yasinskij, S.V., 1993. Impact of mulch layer composed by organic remains on the soil thermal conditions following snowmelt. J. Hydrol. Hydromech. (Slovak Republic) 41 (1), 15–28.
Henderson-Sellers, A., Yang, Z.-L., Dickinson, R.E., 1993. The project for intercomparison of land-surface parameterization schemes. Bull. Am. Meteorol. Soc. 74, 1335–1349.
Ivanov, N.S., Gavril'ev, R.I., 1965. Teplophizicheskie Svoyistva Gornykh Porod (Thermophysical Properties of Rocks). Nauka, Moscow, USSR (in Russian).
Kalyuzhnyi, I.L., Pavlova, K.K., 1981. Formirovanie Poter' Talogo Stoka (Formation of Melted Snow Losses). Gidrometeoizdat, Leningrad, USSR (in Russian).
Khromov, S.P., Mamontova, L.I., 1974. Meteorologicheskii slovar' (Meteorological Glossary). Gidrometeoizdat, Leningrad, USSR (in Russian).
Krenke, A.N., Nikolaeva, G.M., Shmakin, A.B., 1991. The effects of natural and anthropogenic changes on heat and water budgets in the Central Caucasus USSR. Mountain Res. Dev. 11 (3), 173–182.
Kuchment, L.S., Demidov, V.N., Motovilov, Y.G., 1983. Formirovanie Rechnogo Stoka. Phiziko–Matematicheskie Modeli (Formation of River Runoff. Physical–Mathematical Models). Nauka, Moscow, USSR (in Russian).
Manabe, S., 1969. Climate and the ocean circulation: 1. The atmospheric circulation and the hydrology of the earth's surface. Mon. Weath. Rev. 97, 739–805.
Palagin, E.G., 1981. Matematicheskoe Modelirovanie Agrometeorologicheskikh Usloviyi Perezimovki Ozimykh Kul'tur (Mathematical Modelling Winter Agrometeorological Conditions for Winter Crops). Gidrometeoizdat, Leningrad, USSR (in Russian).
Pavlov, A.V., 1979. Teplophizika landshaftov (Thermophysics of landscapes). Nauka, Sibirskoe otdelenie, USSR (in Russian).
Philip, J.R., 1957. The theory of infiltration: 4. Sorptivity and algebraic infiltration equations. Soil Sci. 84, 257–264.
Pivovarova, Z.I., 1977. Radiatsionnye Kharakteristikhi Klimata SSSR (Radiation Characteristics of Climate of the USSR). Gidrometeoizdat, Leningrad, Russia (in Russian).
Rode, A.A., 1955. Pochvovedenie (Soil Science). Goslesbumizdat, Moscow–Leningrad, USSR (in Russian).
Sellers, P.J., Mintz, Y., Sud, Y.C., Dalcher, A., 1986. A simple biosphere model (SiB) for use within general circulation models. J. Atmos. Sci. 43, 505–531.
Verigo, S.A., Razumova, L.A., 1973. Pochvennaya Vlaga (Soil Moisture). Gidrometeoizdat, Leningrad, USSR (in Russian).
Vershinin, P.V., Mel'nikova, M.K., Michurin, B.N., Moshkov, B.S., Poyasov, N.P., Chudnovskiyi, A.F., 1959. Osnovy Agrophiziki (The Basis of Agrophysics). Fizmatgiz, Moscow, USSR (in Russian).
Yosida, Z., 1955. Physical studies on deposited snow. Centrib. Inst. Low Temp. Sci. Sapporo 7, 19–74.
Zilitinkevich, S.S., 1970. Dinamika Pogranichnogo Sloya Atmosfery (Dynamics of the Boundary Layer of the Atmosphere). Gidrometeoizdat, Leningrad, USSR (in Russian).
Zilitinkevich, S.S., Monin, A.S., 1971. Turbulentnost' v Dinamicheskhikh Modelyakh Atmosphery (Turbulence in Dynamical Atmospheric Models). Nauka, Leningrad, USSR (in Russian).