Parametrization of snow accumulation under forest canopy for INM RAS-MSU land surface model

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Abstract. Snow accumulation and melting is one of the most important processes to be simulated by a land surface model. Snow parameterization schemes have been constructed to be applicable to areas without vegetation until the 90s of the 20th century. Even nowadays, only a few of the most advanced Earth system models (for example, CLM, SiB) include sophisticated parameterizations of snow cover including under-canopy snow storage. This paper presents a new snow cover parameterization in the INM RAS-MSU land surface scheme explicitly taking into account snow accumulation under vegetation canopy. The thus updated scheme has been tested using measurement data for the Severnaya Dvina River basin. The new snow scheme takes into account more physical processes that are specific to forested river catchments. This is a necessary step to improve the simulation of water and heat balances in boreal regions.

1. Introduction
Snow cover is an important part of the climate system because of its physical characteristics: high albedo and low thermal conductivity. Snow cover is a reservoir of melt water. It affects all important hydrological cycles such as stream flow, evaporation rate, and precipitation. Thus, correct reproduction of "snow processes" (including snow cover formation, its evolution and melting) is of great practical importance for various fields of science (hydrological forecasts, numerical weather forecast, seasonal forecasts, climate modeling).

1.1 Forest canopy effect
There are several main factors in the LSM models which help us to reproduce the influence of forest on the redistribution of snow mass. This is the canopy structure, snow intercepted by canopy, snow evaporation, snow transport at the forest [1], etc.

The structure of the forest canopy in the models is taken into account using the leaf area index LAI (total plant leaf area per unit surface area), the steam index SAI (steam surface area), or the sum of the two indices [2]. There are also some quantities that describe the forest structure, except the previous one, but the physical meaning is identical based on the following concept: a reservoir is introduced into the calculation scheme of the model, in which precipitation trapped by vegetation can accumulate and be removed from it [3]. The amount of moisture in this reservoir cannot exceed a predetermined value, the maximum intercept capacity $W_{\text{max}}$ which can be set as a function of the leaf index LAI:

$$W_{\text{max}} = x \text{LAI},$$

(1)
where \( x \) is the thickness of the water layer on the leaf surface, \( x \) is usually equal to 0.1 or 0.2 for liquid precipitation [4].

One of the problems in this area is the problem of accurately estimating the reallocation of snow mass between the open area and in the forest. Scientists cannot agree. For example, Hardy in [5] claims that according to measurements at the Burns site in Canada, up to 60% of the mass of snow located in the open space can accumulate in spaces under the forest canopy. On the other hand, [6] demonstrates that a snow density decrease under the canopy leads to an increase in the snow water equivalent (SWE) on the same territories in the Russian temporal forest.

1.2 Radiation processes in the canopy
Vegetation may cause changes in the radiation regime of the surface layer. For instance, snow mass under dense forest canopy changes the albedo of the forest area [7], and this effect increases simultaneously with an increase in the sparseness of the forest cover [8].

The radiation processes in the forest canopy are more complicated. There are multiple reflections, attenuation, and scattering inside. The most computationally efficient radiation scheme is a two-steam approximation [9]. This scheme was developed by Kubelka and Munk in 1931. Today the common variant is summarized in [10].

There are other schemes describing the radiation transfer, for example, the Monte Carlo method. These models are only for describing radiation. They are not suitable for general circulation models (GCMs) due to high computational time costs.

1.3 Turbulence in the canopy
The turbulent exchange between forest and atmosphere consists of two parts: turbulent exchange at the upper border of the forest and exchange within the forest canopy. Note that the day temperature is always several degrees higher under the canopy than over open sites during the cold period. As a result, unstable stratification is established above forest and stable stratification is established under forest canopy [11].

Turbulence in the land-surface models is described by using a “big leaf” concept implemented in [12] and in other models like SiB (Simple Biosphere Model), BATS (Biosphere – Atmosphere Transport Scheme), etc. The forest canopy is presented as a homogeneous large leaf without vertical structure. Heat and moisture fluxes are calculated separately with Monin–Obukhov theory.

The international community has organized some projects to compare the new simulation results. Programs such as the Atmospheric Model Intercomparison Project AMIP [13], a project for intercomparison of land surface parameterization schemes PILPS [14], and the Snow Model Intercomparison Project SnowMIP [15] are widely known. Each project made a significant contribution to improving the parametrization of the Earth system models in the field of snow accumulation.

2. Model description

2.1 Land surface scheme
This research uses an INM RAS-MSU land–surface model [16] with a resolution of 0.5×0.5 degrees in longitude and latitude, with a one-hour time step. The computational domain is the Northern Dvina basin (42°–79° N, 20°–70° E). ERA-Interim reanalysis with a resolution of 1×1 degrees in latitude and longitude and a time step of 6 hours was taken as input data. The time period is 01.01.1979 - 31.12.1994. Reanalysis precipitation was corrected using station data via an algorithm described in [17].

The main physical processes affecting the snow cover evolution in the new version of INM RAS-MSU land surface model are given in Figure 1.
The main result of the land–surface model is the values of turbulent momentum fluxes $\tau$, sensible heat $H$, and latent heat $LE$. Turbulent fluxes are calculated by using the equations

$$\tau = \rho c_P u^2, \quad (2)$$

$$H = -c_p \rho C_u C_\theta u (\theta - \theta_s), \quad (3)$$

$$LE = -L \rho \frac{1}{\tau} (q - q_i), \quad (4)$$

where $L$ is the evaporation heat; $q$ is the specific humidity in the surface layer; $q_i$ is the vapor pressure over the i-type of the surface; $\rho$ is the average density of air in the surface layer; $\theta_s$ is the average cell surface temperature; $u^2 = (u_x^2 + u_y^2)$.

The heat flux spent on melting $L_T M$ is specified as the residual of the heat balance which is spent on evaporation/condensation:

$$L_T M = R(T_{snow}) - H(T_{snow}) - L_v E_{snow}(T_{snow}) - B_T, \quad (5)$$

where the first term to the right is the net radiation, the second term is the sensible heat flux, the third term is the latent heat flux, and the fourth term is the heat flux into the soil. Assume that $B_T = 0$, i.e. the excess heat on the surface is spent on the phase transition when calculating the heat flux for melting snow.

The model assumes that precipitation reaches the ground and falls evenly throughout the cell. It can be liquid or solid, which is determined by the air temperature. However, precipitation can be of only one type at the moment and cannot be transformed.

The calculation algorithms and some important physical quantities for each surface type are discussed below.

2.1.1 *Vegetation.* The leaf area index (LAI) is defined as

$$LAI = \sum_{i=1}^{N_v} LAI_i, \quad (6)$$
where \( N_v \) is the number of vegetation types, it is assumed to be a function of soil temperature and soil moisture and is described by the annual cycle. It is used to calculate the intercepted moisture.

### 2.1.2 Snow cover

The cell area fraction covered by snow is calculated with the equation

\[
f_{\text{snow}} = \max \left[ 0.01, \min \left( \frac{S_{\text{we}}}{S_{\text{we,cr}}}, 0.9 \right) \right],
\]

where \( S_{\text{we,cr}} = 0.4 \) cm is the critical snow cover moisture content (or snow water equivalent). If \( S_{\text{we}} \) is greater than 0.4 cm, the snow cover covers the maximum possible part of the cell, 0.9. On the other side, if \( S_{\text{we}} \) is less than \( S_{\text{we,min}} = 10^{-4} \) cm (the minimum value of the snow cover moisture content), it is assumed that there is no snow cover in the cell.

In the basic version of the model, the presence of different surface types in a cell ceases when the cell is completely covered with snow. However, it is obvious that the snow on the open soil and snow on the vegetation have different physical properties.

In the snow scheme of the model, the snow water equivalent (SWE) is interpreted as the value averaged over the cell, \( S_{\text{we}} \). Its evolution equation can be expressed as follows:

\[
\frac{\partial S_{\text{we}}}{\partial t} = P_{\text{sn}} - (E_{\text{sn}} + M),
\]

where \( P_{\text{sn}}, E_{\text{sn}}, M \) are the precipitation rate, the evaporation rate, and the melting rate of snow, respectively.

### 2.1.3 Intercepted water

Intercepted water is atmospheric precipitation (liquid or solid) trapped by vegetation. In the baseline version of the model, only liquid precipitation is intercepted. When the temperature becomes negative, the intercepted precipitation is assumed zero.

The cell fraction occupied by the intercepted water is

\[
f_{\text{ir}} = \frac{W_{\text{skin}}}{W_{\text{skin,max}}},
\]

where \( W_{\text{skin}} \) is the intercepted water layer thickness; \( W_{\text{skin,max}} \) is the maximal moisture of the intercepted precipitation calculated by using the following equation:

\[
W_{\text{skin,max}} = W_{\text{skin,max,0}} \left[ (1 - f_{\text{veg}}) + f_{\text{veg}} * \text{LAI} \right],
\]

where \( W_{\text{skin,max,0}} = 0.05 \) cm is the typical value of the moisture layer thickness.

The prognostic equation for water intercepted by vegetation in the baseline model version is

\[
\frac{\partial W_{\text{skin}}}{\partial t} = (M_{\text{soil}} + C_{\text{eff}} P_{\text{rain}}) f_{\text{veg}} (1 - f_{\text{snow}}) - S_{\text{ir}} E_{\text{ir}} \rho_{\text{ir}}^2 - Q_{\text{skin}},
\]

where \( P_{\text{rain}} \) is the liquid precipitation; \( E_{\text{ir}} \) is the evaporation of intercepted precipitation, and \( C_{\text{eff}} \) is the interception efficiency.

### 2.1.4 Bare ground

The cell area occupied by open soil depends on the vegetation wilting:

\[
f_{\text{soil}} = f_{\text{soil}}^* + (f_{\text{veg}} - f_{\text{veg}}),
\]

where \( f_{\text{soil}}^* \) is the smallest value of \( f_{\text{soil}} \).

Radiation transport under the forest canopy is based on two-stream approximations from [18] and [10].

### 2.2. Modification of land–surface scheme

#### 2.2.1 Modifications in surface types

The new model scheme includes a new distribution of the surface types covering the land cell. The definitions of these types have changed. The surface types consist of the following new parts: \( f_{\text{veg}} \) is vegetation; \( f_{\text{veg,ew}} \) is the part of vegetation wetted by precipitation (wet vegetation); \( f_{\text{snow}} \) is snow cover; \( f_{\text{soil}} \) is open soil; \( f_{\text{lake}} \) is the surface of water bodies.

Now the type "surface waterlogged by precipitation" is replaced by the type "wet vegetation"; this is due to the ambiguity of the physical interpretation of the original type. “Wet vegetation” is the part
of vegetation covered by intercepted precipitation. Based on this definition, the cell’s fractions $S_i$ occupied by the surface types mentioned above can be expressed as

$$S_{\text{snow}} = f_{\text{snow}} (1 - f_{\text{lake}}),$$

$$S_{\text{wveg}} = (1 - f_{\text{snow}}) (1 - f_{\text{lake}}) * f_{\text{wveg}} * f_{\text{veg}},$$

$$S_{\text{soil}} = (1 - f_{\text{snow}}) (1 - f_{\text{veg}}) (1 - f_{\text{lake}})$$

Moreover, the total vegetation cell fraction is, obviously, $S_{\text{veg}} = S_{\text{wveg}} + S_{\text{dveg}}$.

Changes in the cell division by surface types are the reason for changes in some equations containing fractions of these types. The maximum of the wet vegetation layer moisture content (or the maximum layer of intercepted precipitation) now depends only on the cell area covered with vegetation:

$$W_{\text{skin,max}} = W_{\text{skin,max,0}} * f_{\text{wveg}} * TL,$$

where $TL$ is the sum of leaf and stem indexes, and the parameter $W_{\text{skin,max,0}}$ acquires an accurate physical meaning: this is the maximum thickness of water or snow in winter (in water equivalent) that can be retained by leaves and stems of plants.

### 2.2.2. Modifications in surface water balance

One of the main model modifications is related to the surface water balance in the winter season, including the melting and evaporation processes, which are determined by the thermal balance net radiation at the surface (5).

The albedo is a key parameter of the net radiation, which is especially important during the period of snow melting. In the new scheme, the cell-average albedo is used for calculating the heat balance; the averaging method follows [19]. The albedo of any of the basic surface types is averaged with areal weights. When vegetation is present in the model cell ($f_{\text{veg}} > 0$), the albedo is averaged over the vegetation types:

$$a_{\text{dveg}} = \frac{1}{2} \sum_{n=1}^{N_{\text{veg}}} f_{\text{veg,n}} * a_n * (a_{\text{dn}} + a_{\text{ln}}),$$

where $a_{\text{dn}}$ is the albedo of direct shortwave radiation of the $n$-th vegetation type; $a_{\text{ln}}$ is the albedo of shortwave scattered radiation of the $n$-th vegetation type.

Now consider the changes in representation of the water-equivalent precipitation layer intercepted by vegetation $P$ (cm/s) and the total flux of liquid water into the soil $P_{\text{soil}}$. If the air temperature is higher than the melting temperature ($T_a \geq T_f$),

$$P = \beta \alpha P_{\text{liq}} f_{\text{veg}},$$

$$P_{\text{soil}} = P_{\text{liq}} (1 - \beta \alpha f_{\text{veg}}) + M_{\text{soil}},$$

where $\alpha = 0.85$ is the efficiency of interception of liquid precipitation by vegetation; $\beta$ is the cell fraction occupied by atmospheric precipitation (taken equal to 1). This takes into account the fact that the part of the snow cover has not yet had time to melt when a positive surface air temperature occurs. Note that the precipitation fraction intercepted by vegetation was 0.5 in the previous model scheme.

When the negative air temperature is established ($T_a < T_f$), atmospheric precipitation comes in solid form, and so the liquid moisture enters the soil only as a result of snow melting:

$$P_{\text{soil}} = M_{\text{soil}}$$

Snow evaporation rates can be expressed as

$$E_{\text{snow}} = -L S_{\text{snow}} \left( \frac{1-f_{\text{veg}}}{r_{\text{snow}}} + \frac{f_{\text{veg}}}{\gamma_{\text{veg, snow}}} \right) (q - q_{\text{snow}}),$$

where $\gamma_{\text{veg}}$ is the “reflecting factor” which expresses an increase in the resistance to evaporation from the snow surface under tall vegetation. It is assumed 10 in the new snow scheme. The intercepted snow is added to the trapped water reservoir:

$$\frac{dW_{\text{skin}}}{dt} = P - E_{\text{veg}}.$$
If the SWE intercepted by vegetation exceeds the maximum possible value \( W_{\text{skin}} > W_{\text{skin,max}} \), the excess is added to the average water-equivalent thickness of the snow cover \( S_{\text{we}} \).

2.2.3. SWE underneath the canopy and on top the bare soil. Two new variables, i.e. water-equivalent snow on bare surfaces averaged over the cell \( S_{\text{we,soil}} \) and water-equivalent snow under vegetation averaged over the cell \( S_{\text{we,veg}} \), were introduced into the new scheme.

The SWE under vegetation is an excess of snow which cannot be retained by canopy. It can evaporate and melt:

\[
\frac{\partial S_{\text{we,veg}}}{\partial \tau} = P_{\text{unload}} - (E_{\text{sn}} + M_{\text{soil}}) \frac{S_{\text{we,veg}}}{S_{\text{we}}}. \tag{24}
\]

Here a closing hypothesis is used, stating that the evaporation and melting rate of the snow under canopy is proportional to the rate of these processes on average over the cell. It is also proportional to the ratio of the SWE in the forest to the SWE averaged over the cell.

The open surface SWE is then computed diagnostically:

\[
S_{\text{we,soil}} = S_{\text{we}} - S_{\text{we,veg}} \tag{25}
\]

Comparison with observation data requires averaging the SWE over the areas of vegetation and open sites, respectively (shown by asterisks):

\[
S_{\text{we,soil}}^* = S_{\text{we,soil}}/(1 - f_{\text{veg}}), \tag{26}
\]

\[
S_{\text{we,veg}}^* = S_{\text{we,veg}}/f_{\text{veg}}. \tag{27}
\]

2.3. Numerical experiment setup

The measurement data used to validate the model is an array of snow surveys available from the Russian Scientific Research Institute of Hydrometeorological Information - World Data Center [http://meteo.ru/]. The time period is 1979 - 2019.

The data array "Snow survey" consists of 19 parameters. Three of them were selected for this research: mean snow cover height, mean snow density, snow–water equivalent.

Seventeen meteorological stations located within the Northern Dvina basin were selected for comparison with the model results. Snow survey data were available for 17 out of 34 meteorological stations located in the specified area. The data were checked for completeness and reliability for each type of location (field, forest, and ravines).

3. Results

This section compares the simulation results with the new snow scheme versus observational data.

The processing of the SWE time series was carried out according to the following scheme: the data of each station and the model cell closest to the station were averaged for each month of each year, and then the monthly mean values were averaged over 15 years. The maximum monthly values were defined in the same way. Thus, 17 series (by the number of stations) of 15-year mean monthly values and maximum values were obtained. The data were also averaged over the territory of the Northern Dvina basin.

The accuracy of modeling the interannual variability of the maximum moisture content in the snow cover was assessed using monthly mean values for March as the month with the climatologically highest snow-water equivalent. The processing described above was carried out for each of the following values: the SWE average over the cell \( S_{\text{we}} \), the SWE under vegetation \( S_{\text{we,veg}}^* \) and the SWE in open sites \( S_{\text{we,soil}}^* \).
Figure 2. Mean monthly SWE for 1979-1994 averaged over 17 stations. Comparison of simulation data (cell average, $S_{we}$) and snow survey data (blue - “Field data” is snow in open areas, green - “Forest data” is snow under canopy, red – model data).

Figure 2 shows the seasonal course of the SWE averaged over a long-term period (25 years) and over the Northern Dvina basin. The seasonal variation of the SWE is reproduced well, in particular, the characteristic rates of snow accumulation from November to March in the model are close to those observed in this territory. The underestimation of the model data relative to observations by ~ 2 cm can be traced during the entire cold period starting from October, when a stable snow cover is formed on the territory of the Northern Dvina River basin. A similar problem was noticed in other models, for example, TERRA [20].

The simulation of the SWE shows a slight overestimation during the melting period (March, April). The reason may be lack of the longwave radiation parameterization inside canopy. The downward thermal radiation of the canopy increases in March and April when the canopy loses snow and, accordingly, a sharp drop in the albedo of stems, branches, and leaves takes place and, hence, the canopy heating by shortwave radiation increases. This increases downward thermal radiation from the canopy and thereby enhances snow melting.

The spatial distribution of the March SWE correlation coefficients (based on the model data and observation data in the forest and in the field, which is not shown) over the Northern Dvina basin demonstrates that the smallest correlation is characteristic of the stations located in the Southern part of the basin.

Consider the seasonal dynamics of the SWE in the open areas and under the canopy. Figures 3 (a,b) show that the simulated long-term-averaged seasonal course is reproduced satisfactorily. The difference between the "model - observation" averages about ~ 1 cm for the SWE in the field and ~ 2 cm for the SWE under the forest during the cold period due to the late subgrid snow-on dates, excluding the period of snow melting. During the snowmelt period, the difference "model - observations" is minimal.

Figure 3. Mean monthly SWE in open areas (a) and under canopy (b) for 1979-1994 averaged over 17 stations (snow surveys) and the nearest model cells ($S_{we,soil}$). $\alpha = 0.85$ (blue - “Field data” is snow in open areas, red – model data).
Note that the new snow scheme reproduces a feature revealed from the observation data: the SWE under the forest canopy is higher versus open areas. This ratio is controlled in the model using the efficiency coefficient of solid precipitation interception by the canopy $a$. This coefficient is calculated in the equation for water interception by the canopy (19, 23). Numerical experiments were carried out with the $a$ coefficient = 0.5, 0.7, 0.85, 1. It turns out that the higher the efficiency of interception of solid precipitation by vegetation, the more vegetation holds snow, and the less snow falls under the forest canopy.

4. Conclusions
The snow scheme of the INM RAS-MSU land surface model has been improved. The new scheme gives a more thorough understanding of the processes taking place under forest canopy. The new algorithm includes melting, evaporation, and redistribution of snow masses not only in open areas but also under forest canopy. Thus, this physical model more correctly simulates the processes of precipitation interception by vegetation.

We compared modeling data and measurements of the snow water equivalent (SWE) for three key sites: average model cell, open areas, and under forest canopy in the Severnaya Dvina River basin.

The seasonal variation of the SWE is satisfactorily reproduced by the model. The model underestimates the SWE on average by 2 cm per cell throughout the entire cold period (excluding the period of snow melting) for 1979-1994.

There are several reasons why snow storage in the model is underestimated. One is the high sensitivity of the model to the atmospheric data during the snow cover formation period (October). Another is the incorrect reproduction of the albedo of snow during the initial stage of snow cover development. In future studies, it is planned to correct the rate of spring snow melting by using a parameterization of the downward thermal radiation of canopy, as well as to include a physically based parameterization of the snow-covered landscapes’ albedo.

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