ISBA Surface Scheme in the GEM DM Numerical Forecast Model

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The ISBA Land Surface Scheme

This improved version of the Interactions Soil-Biosphere-Atmosphere (ISBA) scheme, originally developed by Noilhan and Planton (1989), has been included in the RPN physics package. Its main purpose is to determine the lower boundary conditions for the vertical diffusion of temperature, moisture, and momentum, as well as evaluating the evolution of ten prognostic variables [i.e., the surface temperature T_{surf}, the mean (or deep-soil) temperature T_p, the near-surface soil moisture w_g, the liquid and frozen bulk soil water contents w_p and w_f, the liquid water W_r retained on the foliage of the vegetation canopy, the equivalent water content W_S of the snow reservoir, the liquid water W_L retained in the snow pack, the snow albedo α_S, and the relative snow density ρ_S] and the hydrological budget of the surface.

a. Entry parameters

The entry parameters have been chosen in a way to characterize the main physical processes occurring at the surface, while attempting to reduce the number of independent variables. As shown in 1, they can be divided in two categories: *primary parameters* that need to be specified at each model grid point, and *secondary parameters* which values can be derived (using association tables) from the primary parameters. The primary parameters describe the nature of the land surface and its vegetation coverage by means of only four numerical indices: the percentage of sand and clay in the soil, the dominant vegetation type, and the land-water mask. The secondary parameters associated with the soil type are evaluated from the sand and clay composition of the soil, according to the continuous formulation discussed in Giordani (1993) and Noilhan and Lacarrère (1995), whereas those related to vegetation can either be derived from the dominant vegetation type or from existing classification or observations.

b. Thermal properties of the surface

The prognostic equations for the superficial and mean surface temperatures (T_{surf} and T_p) are obtained from the force-restore method following:

$$\frac{\partial T_{surf}}{\partial t} = C_{TOT}\left(R_n - H - LE\right) + C_T\, L_f\left(freez_g - melt_g + freez_s - melt_s\right) - \frac{2\pi}{\tau}\left(T_{surf} - T_p\right) \quad (1)$$

$$\frac{\partial T_p}{\partial t} = \frac{1}{\tau}\left(T_{surf} - T_p\right) \quad (2)$$

in which H, LE, and R_n are the sensible heat, latent heat, and net radiational fluxes at the surface, C_{TOT} is a thermal coefficient, L_f is the latent heat of fusion, $freez_s$ and $melt_s$ are fluxes of freezing and melting snow, and τ is a time constant of one day. The first term on the rhs of (1) represents the forcing from radiative fluxes at the surface; the second term is for the release of latent heat due to freezing and melting of soil water and snow; and the last term of (1) [like the only rhs term in (2)], is a "restoring" or relaxation term.

Since only a single energy budget is done for the soil portion of a model grid area, the heat coefficient C_{TOT} includes the effect of bare soil, vegetation, and snow:

$$C_{TOT} = \frac{1}{\dfrac{(1-veg)(1-p_{sng})}{C_g} + \dfrac{veg\,(1-p_{snv})}{C_v} + \dfrac{p_{sn}}{C_s}} \quad (3)$$

where the contributions of ground, vegetation, and snow (i.e., C_g, C_v, and C_s) are weighted by the fraction of the model grid area covered by vegetation (veg), the fraction of the bare soil covered by snow (p_{sng}), the fraction of vegetation covered by snow (p_{snv}), and the fraction of the model grid area covered by snow (p_{sn}). These fractional grid areas are given by:

$$p_{sng} = \min\left(\frac{W_S}{W_{crn}}, 1.\right) \quad ; \quad p_{snv} = \frac{h_s}{h_s + 5000 z_0} \quad ; \quad p_{sn} = (1-veg)\,p_{sng} + veg\;p_{snv} \quad (4)$$

in which W_{crn} = 10 kg m^{-2}(or mm), and $h_s = W_S / \rho_S$ is the thickness of the snow canopy.

The thermal coefficient for the vegetation (C_v) has fixed values of 1.x10^{-5} or 2.x10^{-5} K m^{-2} J^{-1} for high and low vegetation respectively. For bare ground, the thermal coefficient is given by:

$$C_g = C_{gsat}\left(\frac{w_{sat}}{w_2}\right)^{b/2\ln 10} \qquad C_g \le 2\,\mathrm{x}10^{-5}\ \ \mathrm{Km^2J^{-1}} \quad (5)$$

For snow, the heat coefficient is given by:

$$C_S = 2\left(\frac{\pi}{\lambda_S c_S \tau}\right)^{1/2} \qquad (6)$$

where $\lambda_S = \lambda_i \rho_S^{1.88}$; $c_S = c_i(\rho_S/\rho_i)$; in which ρ_i is relative density of ice [see Douville (1994) and Douville et al. (1995)].

c. Hydraulic properties of the surface

Soil water in both the superficial and deep reservoirs also evolve according to force-restore equations:

$$\frac{\partial w_g}{\partial t} = \frac{C_1}{\rho_w d_1}\left\{(1-veg)(1-p_{sng}) P_r + (1-p_{sn})R_{veg} + R_{snow} - E_g - R_{surf}\right\} - \frac{C_2}{\tau}(w_g - w_{geq}) \quad (7)$$

$$\frac{\partial w_2}{\partial t} = \frac{1}{\rho_w d_2}\left\{(1-veg)(1-p_{sng})P_r + (1-p_{sn})R_{veg} + R_{snow} + melt_g - freez_g - E_g - E_{tr} - R_{surf}\right\} - \frac{C_3}{d_2 \tau}\max\left\{0,(w_2 - w_{fc})\right\} \quad (8)$$

in which w_g and w_2 are the soil volumetric water contents, d_1 and d_2 are the depths of the superficial and deep soil layers, P_r is the rain rate, R_{veg}, R_{snow}, and R_{surf} are the water "runoffs" from vegetation, snow, and ground, w_{geq} is an equilibrium volumetric water content depending on the soil texture, E_g is the evaporation from bare soil, E_{tr} is the evapotranspiration from vegetation, and C_1, C_2, and C_3 are expressions and constants for the infiltration [first terms on the rhs of (7) and (8)], restore [second term on the rhs of (7)], and drainage [second term on the rhs of (8)] terms.

The mathematical expression for C_1 depends on the soil moisture content. For relatively wet soils (i.e., $w_g > w_{wilt}$, w_{wilt} is the soil water content at the wilting point), this coefficient is given by:

$$C_1 = C_{1sat}\left(\frac{w_{sat}}{w_g}\right)^{b/2+1} \qquad (9)$$

in which w_{sat} is the water content at saturation. For dry soils (i.e., $w_g < w_{wilt}$), the vapor-phase transfers need to be considered in order to reproduce the physics of water exchange. These transfers are parameterized as a function of the wilting point w_{wilt}, the soil water content w_g, and the surface temperature T_{surf}, using the Gaussian expression (see Braud et al. 1993, Giordani 1993)

$$C_1 = C_{1\max} \exp\left[-\frac{\left(w_g - w_{\max}\right)^2}{2\,\sigma^2} \right] \tag{10}$$

where w_{max}, C_{1max}, and σ are respectively the maximum abcissa, the mode, and the standard deviation of the Gaussian functions. The other coefficient, C_2, and the equilibrium water content, w_{geq}, are given by

$$C_2 = C_{2ref} \left(\frac{w_p}{\max(w_{sat} - w_p, 0.01)} \right) \tag{11}$$

$$w_{geq} = w_p - a\, w_{sat} \left(\frac{w_p}{w_{sat}} \right)^p \left[1 - \left(\frac{w_p}{w_{sat}} \right)^{8p} \right] \tag{12}$$

in which *a* and *p* are constants.

For the w_2 evolution, (7) represents the water budget over the soil layer of depth d_2. The first term on the rhs includes the effect of rainfall, runoff from snow and vegetation, melting/freezing, evaporation, and surface runoff. The other term, for drainage, is proportional to the water amount exceeding that at field capacity (i.e., w_2-w_{fc}) (see Mahfouf et al. 1994). In this second term, the coefficient C_3 does not depend on w_2 but simply on the soil texture.

The temporal evolution of volumetric water content of frozen soil water (i.e., w_f) is represented in ISBA using the following equation:

$$\rho_w\, d_2 \frac{\partial w_f}{\partial t} = freez_g - melt_g \tag{13}$$

in which ρ_w is the liquid water density, d_2 is the soil rooting depth, and $freez_g$ and $melt_g$ are the fluxes of freezing and melting water in the ground, respectively. Based on the work of Giard and Bazile (1999), these fluxes are proportional to differences between the soil-surface and the freezing (melting) temperatures (i.e., T_0=273.16 K):

$$\begin{aligned} freez_g &= \rho_w\, d_2\, K \left(\frac{w_2}{w_{sat}} \right) \max\left[0, -\left(T_n - T_0\right)\right] \\ melt_g &= \rho_w\, d_2\, K \max\left[0, \left(T_n - T_0\right)\right] \end{aligned} \tag{14}$$

where T_n is a representative temperature of the soil under the snow and vegetation canopies:

$$T_n = \left(1 - p_{sn}\right)\left[\left(1 - veg\right) T_{Surf} + veg\, T_p\right] + p_{sn}\, T_p \tag{15}$$

A series of "off-line" experiments has shown that K=1.5x10^{-6} is appropriate for representing in a realistic manner the freezing and thawing of soil water in typical fall and spring conditions. It should be noted that soil water freezing and melting only influence the evolution of w_2 [in (8)]; w_g can thus be interpreted as the sum of liquid and solid water in the superficial layer. It is this last quantity that is used to calculate the surface fluxes of water vapor.

Obviously, the presence of frozen water in the soil should have an important impact on many of the hydraulic processes represented in the above equations (i.e., infiltration, water redistribution, drainage, and runoff). Because it is not clear whether infiltration of liquid water reaching the surface (rain or melting of snow) should increase or decrease due to the presence of frozen water in the soil, the treatment of water infiltration [related to the coefficient C_1 in (7)] in the soil is not modified. One should note however that the evaporation E_g and evapotranspiration E_{tr} appearing in the infiltration terms will be significantly reduced due to the presence of frozen water (see discussion below). Also, as soil water freezes, it loses its fluidity and becomes more static, with less possibility of movement. Ice rings form and become fixed within the soil pores. In these conditions, gravity and suction forces have a lesser impact, and the redistribution of water in the soil is greatly reduced. To represent this effect, the restore term in (7) is forced to zero as the soil water freezes, since w_2 (the *liquid* portion of soil water) tends toward zero in the equation for C_2:

$$C_2 = C_{2ref}\left[\frac{w_2}{w_{sat} - w_2 + 0.01}\right] \tag{16}$$

Using the same argument, it can be asserted that drainage at the bottom boundary of the soil layer greatly decreases as soil water freezes. Finally, the infiltration simulated in ISBA should dramatically decrease (and runoff increase) for frozen soils because evaporation and sublimation of soil water is greatly reduced when the soil is frozen. Because soil water is not redistributed by the restore term and because there is no drainage at the bottom of the soil layer, water should quickly accumulate in the upper layers of the soil (due to precipitation and melting), thus reducing infiltration and increasing surface runoff.

d. Intercepted water

Rainfall and dew intercepted by the canopy foliage feed a reservoir of water content W_r. This water evaporates in the air at a potential rate from the fraction δ of the foliage covered with a film of water, as the remaining part 1-δ of the leaves transpires. Following Deardorff (1978),

$$\frac{\partial W_r}{\partial t} = veg\ P_r - \left(E_v - E_{tr}\right) - R_{veg} \quad ; \quad 0 \le W_r \le W_{r\max} \qquad (17)$$

where P_r is the precipitation rate at the top of the vegetation, E_v is the evaporation from the vegetation including the transpiration E_{tr} and the direct evaporation E_r when positive, and the dew flux when negative (in this case E_{tr} = 0), and R_{veg} is the vegetation throughfall. This throughfall occurs when W_r exceeds a maximum value W_{rmax} that depends on the canopy density, i.e., roughly proportional to *veg LAI*. According to Dickinson (1984), we use the simple equation:

$$W_{r\max} = 0.2\ veg\ LAI \qquad (18)$$

e. Subgrid-scale runoff

The model for subgrid-scale runoff of precipitation reaching the ground is based on the so-called Nanjing model (see Wood et al. 1992, Habets and Noilhan 1996). According to this technique, each model grid area (with soil, not water) is supposed to include a set of subgrid reservoirs with an infinite range of infiltration capacity (continuously varying from 0 to a maximum value i_m). If we suppose that precipitation falls uniformly over each subgrid-scale reservoirs, it is possible to show that the runoff is:

$$R_{surf} = R_g + \frac{i_m}{b_r + 1}\left[\left(1 - \frac{i_0}{i_m} + \frac{R_s}{i_m}\right)^{b_r+1} - \left(1 - \frac{i_0}{i_m}\right)^{b_r+1}\right] \qquad (19)$$

where

$$i_m = \left(1 + b_r\right) w_{sat} d_2 \qquad (20)$$

$$\left(\frac{i_0}{i_m}\right) = 1 - \left(1 - \frac{w_p + w_f}{w_{sat}}\right)^{\frac{1}{b_r+1}} \qquad (21)$$

and b_r is an adjustable parameter that depends on the surface heterogeneity, slopes, etc. This parameter should be different for each grid point. In the current version of ISBA, b_r=1 is used everywhere. One should also note that there is no runoff, of course, when R_g=0 (R_g is the water reaching the soil).

f. Snow model

The evolution of the equivalent water content of the snow reservoir is given by

$$\frac{\partial W_S}{\partial t} = R_S - E_S + freez_S - melt_S - melt_{rain} \tag{22}$$

where R_S is the snowfall rate, E_S is the sublimation from the snow surface, and $freez_s$ and $melt_s$ are freezing and melting terms.

The freezing and melting terms in (22) are given by:

$$freez_s = \min\left[\frac{p_{sn}\left(T_0 - T_n^*\right)}{C_s\ L_f\ \Delta t},\quad \frac{W_L}{\Delta t}\right] \qquad ; \quad freez_s \geq 0 \tag{23}$$

$$melt_s = \min\left[\frac{p_{sn}\left(T_n^* - T_0\right)}{C_s\ L_f\ \Delta t}, \frac{W_S}{\Delta t}\right] \qquad ; \quad melt_s \geq 0 \tag{24}$$

in which $T_n^* = \left(1 - veg\right) T_S + veg\, T_2$ is a temperature representative of the surface under the vegetation canopy, C_s is the thermal coefficient for snow, and W_L is a prognostic variable newly introduced in ISBA for the liquid water retained in the snow pack.

Another aspect that needs to be considered for the evolution of W_S is related to the modification of the internal energy of the snow due to incident liquid precipitation. It often happens, for instance, that warm precipitation falls on snow and accelerates its melting due to energy transfers between the liquid water and the rest of the snow pack.

By using the concept of enthalpy conservation (which reduces here to conservation of internal energy $h = c_p T$), it can be shown that the new surface temperature T_S' after redistribution of energy could be expressed:

$$T_S' = T_S + p_{sn} \Delta T_{snow} \tag{25}$$

where

$$\Delta T_{snow} = \frac{P_r\ \Delta t\left(T_{rain} - T_S\right)}{\left(W_S^* + W_L^*\right) + P_r\ \Delta t} \tag{26}$$

where $W_S^* + W_L^*$ is the total amount of water in the snowpack affected by the energy transfer between the incident liquid precipitation and the snowpack. The quantity $\left(P_r\ \Delta t\right)$ is the liquid water reaching the surface during a single timestep (in kg m^{-2}, or mm), and T_{rain} is the temperature of the rain falling on the snow (taken as the low-level air temperature for the moment).

If we suppose that $W_S^* + W_L^* = P_r \, \Delta t$ then (26) becomes

$$\Delta T_{snow} = \frac{T_{rain} - T_S}{2} \qquad (27)$$

Since an increase of temperature ΔT_{snow} leads to a melting rate $melt_{rain}$ following

$$\Delta T_{snow} = C_S \; L_f \; (melt_{rain}) \; \Delta t \qquad (28)$$

then the melting rate due to incident rain on the snowpack could be calculated using

$$melt_{rain} = \frac{T_{rain} - T_S}{2 \, C_S \; L_f \; \Delta t} \qquad (29)$$

The new variable W_L evolves according to the following equation:

$$\frac{\partial W_L}{\partial t} = p_{sn} \left(P_r + R_{veg} \right) - R_{snow} + melt_s - freez_s \qquad (30)$$

in which P_r is the rainfall and R_{veg} is the runoff from the vegetation canopy. When the amount of liquid water in the snow approaches and exceeds a critical water content W_{Lmax}, there is percolation (snow throughfall) of liquid water towards the ground, following:

$$\begin{aligned} R_{snow} &= \frac{W_{L\max}}{\tau_{hour}} \exp\left\{ W_L - W_{L\max} \right\} && \text{if} \quad W_L \le W_{L\max} \\ R_{snow} &= \frac{W_{L\max}}{\tau_{hour}} + \frac{W_L - W_{L\max}}{\Delta t} && \text{if} \quad W_L > W_{L\max} \end{aligned} \qquad (31\text{-}32)$$

where $$W_{L\max} = c^R \; W_s \qquad (33)$$

in which τ_{hour} is a time constant of one hour and c^R is a retention factor depending on the density of snow:

$$c^R = \begin{cases} c^R_{\min} & if \;\; \rho_S \ge \rho_e \\ c^R_{\min} + \left(c^R_{\max} - c^R_{\min} \right) \dfrac{\rho_e - \rho_S}{\rho_e} & if \;\; \rho_S < \rho_e \end{cases} \qquad (34)$$

with $c^R_{\min} = 0.03$, $c^R_{\max} = 0.10$, and $\rho_e = 0.2$.

The presence of snow covering the ground and vegetation can greatly influence the energy and mass transfers between the land surface and the atmosphere. Notably, a snow layer modifies the radiative balance at the surface by increasing the albedo. To consider this effect, the albedo of snow α_S is treated as a prognostic variable. Depending if snow is melting or not, α_S decreases linearly or exponentially with time:

$$\alpha_S(t) = \alpha_S(t-\Delta t) - \tau_a \frac{\Delta t}{\tau} + \frac{(Snowrate)\,\Delta t}{W_{crn}}\left(\alpha_{S\max} - \alpha_{S\min}\right) \qquad (35)$$

for cold snow cases (i.e., without melting), and

$$\alpha_S(t) = \alpha_{S\min} + \left[\alpha_S(t-\Delta t) - \alpha_{S\min}\right]\exp\left\{-\tau_f \frac{\Delta t}{\tau}\right\} \qquad (36)$$

for warm snow (with melting). Here, $\tau_a = 0.008$, $\alpha_{smax} = 0.80$, and $\tau_g = 0.24$.

The snow density ρ_S currently evolves in ISBA according to the following mechanisms. First, the density increases due to gravitational settling following the exponential function:

$$\begin{aligned} \rho_S^* &= \rho_{S\max} - \left(\rho_{S\max} - \rho_S(t-\Delta t)\right)\exp\left\{-\tau_f \frac{\Delta t}{\tau}\right\} \quad && if\ \ \rho_S(t-\Delta t) < \rho_{S\max} \\ \rho_S^* &= \rho_S(t-\Delta t) && if\ \ \rho_S(t-\Delta t) \geq \rho_{S\max} \end{aligned} \qquad (37\text{-}38)$$

in which ρ_{Smax} is the maximum value for the density of snow (note that this density is relative to that of liquid water), $\tau_f = 0.24$ and $\tau = 3600$ s are constants.

Second, the snow density decreases when new snow falls on top of the snow pack:

$$\rho_S^{**} = \frac{\left(W_S^* - R_S\,\Delta t\right)\rho_S^* + R_S\,\Delta t\,\rho_{S\min}}{W_S^*} \qquad (39)$$

where $W_S^* = \max\left(W_S\,,\,R_S\,\Delta t\right)$, ρ_S^* is an intermediate value of snow density after gravitational settling, and R_S is the snowfall rate. From this second equation, it is clear that $\rho_S \geq \rho_{S\min}$ (= 0.1).

Because liquid water in the snowpack is now evaluated prognostically in the new version of ISBA, and that exchanges of water between the W_L and W_S reservoirs are known through the $freez_S$ and $melt_S$ tendencies, it is possible to include the impact of freezing water in the snow pack, which effect is to increase the density of snow. This is represented following:

$$\rho_S(t) = \left(\frac{W_S}{W_S + freez_S\,\Delta t}\right)\rho_S^{**} + \left(\frac{\Delta W_S}{W_S + freez_S\,\Delta t}\right)\rho_{ice} \qquad (40)$$

in which ρ_{ice} = 0.9 is the density of ice relative to that of water. With this new equation, the snow density can have values much higher than ρ_{Smax}, which is normal near the end of the cold season.

The maximum snow density is a diagnostic variable determined this way:

$$\rho_{S\max} = 0.60 - \frac{20470}{h_s}\left\{1 - \exp\left(-\frac{h_S}{67.3}\right)\right\} \quad if \quad melt_S > 0 \qquad (41)$$

$$\rho_{S\max} = 0.45 - \frac{20470}{h_s}\left\{1 - \exp\left(-\frac{h_S}{67.3}\right)\right\} \quad if \quad melt_S = 0 \qquad (42)$$

in which h_s is the depth of the snowpack (in cm).

The hydrological budget simulated in ISBA is summarized in Fig. 1.

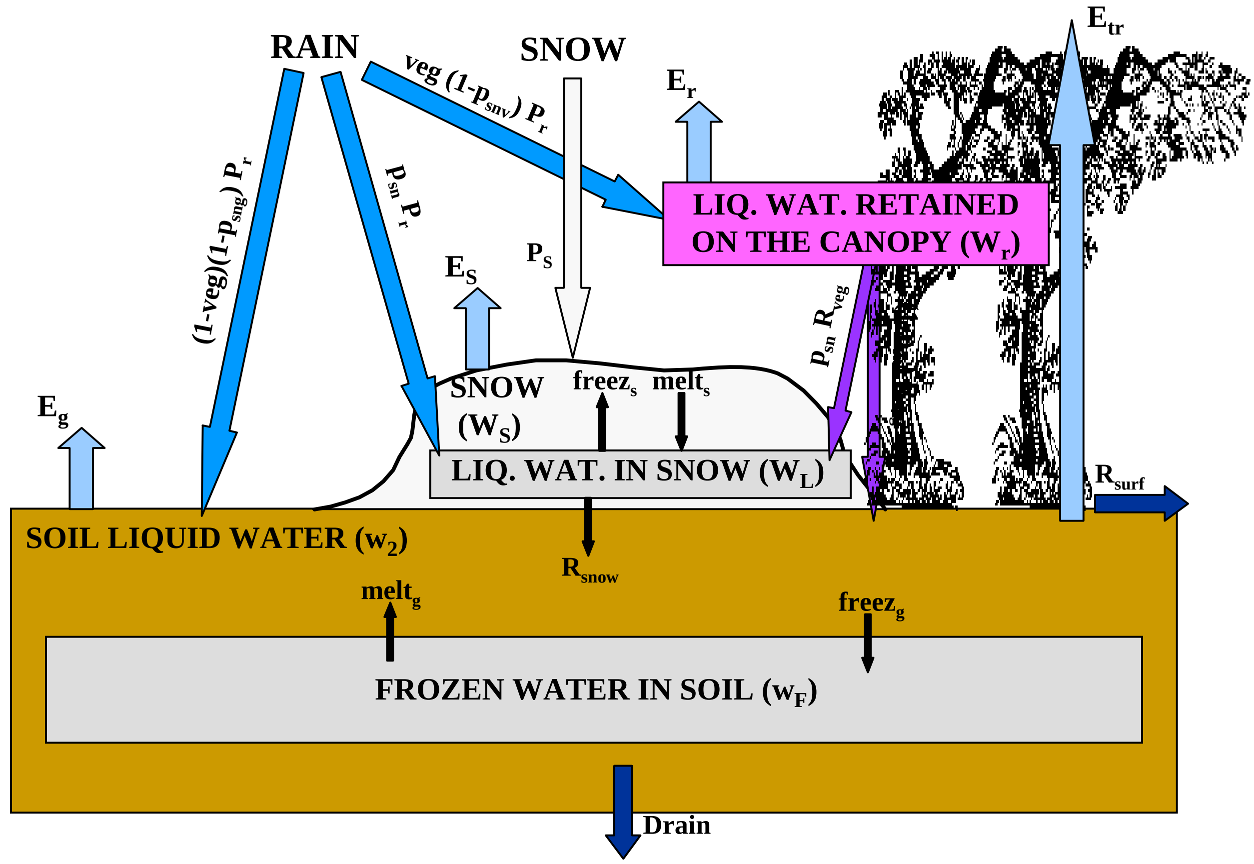

Fig. 1. Hydrological budget in ISBA.

g. Surface fluxes

Only a single energy balance is considered for the whole system ground-vegetation-snow. As a result, heat and mass transfers between the land surface and the atmosphere are related to averaged values of the surface temperature and humidity. Before calculating the surface energy budget, it is necessary to determine land-averaged values for the albedo, emissivity, and roughness length:

$$\alpha_t = (1 - p_{sn})\alpha + p_{sn}\alpha_S \tag{43}$$

$$\varepsilon_t = (1 - p_{sn})\varepsilon + p_{sn}\varepsilon_S \tag{44}$$

where ε_S = 1.0 is the emissivity of the snow. Thus, the overall albedo and emissivity of the ground for infrared radiation is enhanced by snow. For the roughness length z_{0tot} over land:

$$z_{0tot} = (1 - p_{snz0})z_{0M} + p_{snz0}\, z_{0S} \tag{45}$$

with

$$p_{snz0} = \frac{W_S}{W_S + W_{crn} + \beta_S g\, z_0} \tag{46}$$

Here, β_S = 0.408 s^2 m^{-1} and g = 9.80665 m s^{-2} are physical constants, whereas z_{0S} is the roughness length of the snow.

The net radiation at the surface is

$$R_n = F_{SS}^-(1-\alpha_t) + \varepsilon_t\left(F_{SI}^- - \sigma_{SB}T_{surf}^4\right) \tag{47}$$

where F_{SS}^-, F_{SI}^- are the incoming solar and infrared radiation at the surface, and σ_{SB} is the Stefan-Boltzmann constant. The turbulent fluxes are calculated by means of the classical aerodynamic equations (see section 2). For the sensible heat flux:

$$H = \rho_a c_p C_T u_*\left(T_{surf} - T_a\right) \tag{48}$$

where c_p is the specific heat; ρ_a and T_a are for the air density and temperature at the lowest atmospheric level; and C_T is the thermal drag coefficient which depends on the stability of the atmosphere.

The water vapor flux E is the sum of the evaporation from bare ground (i.e., E_g), from the vegetation (i.e., E_v), and from the snow (i.e, E_S):

$$\begin{aligned}
LE &= L_{eff}E_g + L_v E_v + L_i E_S \\
E_g &= (1-veg)(1-p_{sng})\rho_a C_T u_*\left(h_u q_{sat}(T_{surf}) - q_a\right) \\
E_v &= veg(1-p_{snv})\rho_a C_T u_* h_v\left(q_{sat}(T_{surf}) - q_a\right) \\
E_S &= p_{sn}\,\rho_a C_T u_*\left(q_{sat}(T_{surf}) - q_a\right)
\end{aligned} \tag{49}$$

where L_v and L_i are the specific heat of evaporation and sublimation, $q_{sat}(T_{surf})$ is the saturated specific humidity at the temperature T_{surf}, and q_a is the atmospheric specific humidity at the lowest model level.

For the bare-soil portion of the surface, the latent heat constant is modified so as to account for sublimation of soil ice. It is proposed to use an effective latent heat constant, following:

$$L_{eff} = f_{ice}\, L_i + \left(1 - f_{ice}\right) L_v \qquad (50)$$

in which

$$f_{ice} = \frac{w_f}{w_f + w_p} \qquad (51)$$

is the fraction of ice in the soil.

The relative humidity h_u of the ground surface is related to the superficial soil moisture w_g following

$$\begin{aligned} h_u &= \frac{1}{2}\left[1 - \cos\left(\frac{w_g}{w_{fc}}\pi\right)\right] && \text{if} \quad w_g < w_{fc} \\ h_u &= 1 && \text{if} \quad w_g \geq w_{fc} \end{aligned} \qquad (52)$$

In case of dew flux when $q_{sat}(T_{surf}) < q_a$, h_u is also set to 1 (see Mahfouf and Noilhan 1991 for details). When the flux E_v is positive, the Halstead coefficient h_v takes into account the direct evaporation E_r from the fraction of the foliage covered by intercepted water, as well as the transpiration E_{tr} of the remaining part of the leaves:

$$\begin{aligned} h_v &= (1-\delta)\frac{R_a}{R_a + R_S} + \delta \\ E_r &= veg\left(1 - p_{snv}\right)\frac{\delta}{R_a}\left(q_{sat}(T_{surf}) - q_a\right) \\ E_{tr} &= veg\left(1 - p_{snv}\right)\frac{1-\delta}{R_a + R_S}\left(q_{sat}(T_{surf}) - q_a\right) \end{aligned} \qquad (53)$$

When E_v is negative, the dew flux occurs at the potential rate, and $h_v = 1$.

Following Deardorff (1978), δ is a power function of the moisture content of the interception reservoir:

$$\delta = \left(\frac{W_r}{W_{r\max}}\right)^{2/3} \qquad (54)$$

The aerodynamic resistance is $R_a = (C_T\, V_a)^{-1}$. The surface resistance, R_S, depends upon both atmospheric factors and available water in the soil; it is given by:

$$R_S = \frac{R_{S\min}}{F_1 F_2 F_3 F_4 LAI} \qquad (55)$$

with the limiting factors F_1, F_2, F_3, and F_4:

$$F_1 = \frac{f + R_{S\min}/R_{S\max}}{1+f}$$

$$F_2 = \frac{w_p - w_{wilt}}{w_{fc} - w_{wilt}} \quad \text{with} \quad 0 \le F_2 \le 1 \tag{56}$$

$$F_3 = 1 - \gamma\left(q_{sat}(T_{surf}) - q_a\right)$$

$$F_4 = 1 - 1.6\text{x}10^{-3}\left(T_a - 298.15\right)^2$$

where the dimensionless term f represents the incoming photosynthetically active radiation on the canopy foliage, normalized by a species-dependent threshold value:

$$f = 0.55 \frac{F_{SS}^{-}}{R_{Gl}} \frac{2}{LAI} \tag{57}$$

Moreover, γ is a species-dependent parameter (see Jacquemin and Noilhan 1990) and R_{smax} is arbitrarily set to 5000 s m^{-1}.

It should be noted that as soil water freezes (i.e., w_2 decreases and w_f increases), the F_2 factor decreases towards zero and the surface resistance R_S increases to infinity. Thus, as would be expected, evapotranspiration is dramatically smaller for frozen soils, due to the unavailability of the soil water.

The surface fluxes of heat, moisture, and momentum that serve as boundary conditions for the vertical diffusion are written in the following way:

$$\left(\overline{w'\,\theta'}\right)_{surf} = \frac{H}{\rho_a c_p \, T_a/\theta_a} \tag{58}$$

$$\left(\overline{w'\,q'}\right)_{surf} = \frac{E}{\rho_a} \tag{59}$$

$$\left(\overline{w'\,V'}\right)_{surf} = C_M^2 \, |V_a|^2 = u_*^2 \tag{60}$$

where w is the vertical motion, θ_a is the potential temperature at the lowest atmospheric level. The primes and overbars denote perturbation and average quantities.

Table 1 Primary and secondary parameters

Primary parameters

SAND	Sand percentage of soil
CLAY	Clay percentage of soil
	Vegetation type
M	Land-water mask

Secondary parameters

w_{sat}	Volumetric water content at saturation
w_{wilt}	Volumetric water content at the wilting point
w_{fc}	Volumetric water content at field capacity
b	Slope of the soil water retention curve
C_{Gsat}	Thermal coefficient at saturation
C_{1sat}	C_1 coefficient at saturation
C_{2ref}	C_2 coefficient for $w_2 = w_{sat} / 2$
C_3	Drainage coefficient
a, p	Parameters for the w_{geq} formulation
w_{geq}	Equilibrium volumetric water content
veg	Fraction of vegetation
d_2	Soil depth
R_{Smin}	Minimum stomatal (surface) resistance
LAI	Leaf Area Index
C_v	Thermal coefficient for the vegetation canopy
R_{Gl}, γ	Coefficients for the surface resistance
z_{0M}, z_{0T}	Roughness length for momentum and heat transfers
α	Surface albedo (vegetation)
ε	Emissivity

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