Surface Modelling in Northern Europe

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Abstract

This is a short description and discussion about surface models. An overview about the development of surface models, for applications in NWP and climate models is described. Different formulation of the uppermost soil layer, and the use uf tiled schemes is also discussed. A particular formulation of a snow scheme, is also included.

1 Introduction to surface modelling

1.1 What is a surface model supposed to do?

Most of the energy, unevenly distributed by the solar radiation is reaching the ground, which is then transferred to the atmosphere via the planetary boundary layer. For a detailed description of the atmospheric flow, the fluxes of heat, moisture and momentum from the boundary layer, is crucial. These processes strongly depend on the surface conditions, and include both physical and biological processes. Moreover, the local conditions in the boundary layer, strongly dependent on the surface conditions, is important information itself. Also secondary information about the soil conditions, runoff, frost, etc. is an important outcome of the surface model.

1.2 Surface energy balance

The surface scheme must solve the *surface energy balance*:

$$R_n = H + \lambda E + G(+F)$$

here R_n is the net radiation balance for the surface:

$$R_n = S(1-\alpha) + \epsilon(L_w - \sigma {T_s}^4)$$

and S is the incoming solar radiation, $(1-\alpha)$ of which is absorbed at the ground, and L_w is the incoming longwave radiation. H and λE are the sensible and latent heat fluxes between the surface and the atmosphere and G is the heat flux that goes to the ground via heat conduction. The term F is the chemical energy stored during the photosynthesis and is usually omitted (<1 % of Rn, but might be important in climate models). The sensible and latent heat fluxes are often described as:

$$H = \frac{T_s - T_a}{r_a}\rho c_p \quad \text{and} \quad \lambda E = \frac{q_{sat}(T_s) - q_a}{r_a + r_s}\rho\lambda$$

here r_a and r_s are the atmospheric resistance and surface resistance respectively. The partioning of the energy into heat and moisture fluxes, as well as the heat conduction into the ground, can be crucial, and there is a risk that compensating errors evolve, between the surface scheme, the vertical diffusion formulation, and the radion scheme.

1.3 Surface water balance

One of the most important things in surface modelling is the water cycle, i.e., from the atmosperic point of view, to model the surface resistance r_s in the energy equation, which is a function of how the evapotranspiration is treated in the surface scheme. A lot of water processes are connected with the surface. The precipitation of rain and snow, are intercepted on the vegetation, and, at that stage, subject to direct evaporation to the atmosphere, on a short time scale. The rest of the water, percolates into the soil, sometimes stored as snow on the surface for a long time. In the soil, the water is entering the biological cycle of photosynthesis, by supplying water to the plants, and trees, in the root zone. The fluxes of water from the root zone to the leaves is regulated by the stomata in the leaves, and the photosynthesis is dependent on external conditions, like PAR (Photosynthetically Active Radiation), temperature, carbon dioxide concentration etc.

The water in the snow can melt and refreeze, several times, and thus delaying the excess runoff during the spring. Besides that, the snow is important also due to the good insulation to the ground, which implies a low heat capacity, and thus rapid surface temperature changes.

2 Overview of evolution of surface models, focused mainly on evaporation

2.1 First generation

The first surface model, used in a climate model was made by Manabe (1969). It was not designed to describe the seasonal and diurnal cycles, and did not contain any heat conduction. This means that the surface temperature was calculated diagnostically at each time by using the energy balance with zero heat capacity for the surface.

Only one single layer of soil moisture, and excessive water became runoff after the "bucket" was filled. The evapotranspiration process was not explicitely adressed, and instead the formula for the latent heat flux was written:

$$\lambda E = \beta \frac{(q_{sat}(T_s) - q_a)}{r_a} \rho \lambda \ , \quad 0 \leq \beta \leq 1$$

i.e. the reduction from potential evaporation was modelled with a single parameter β which was a simple function of the soil moisture w.
A simple relation could be linearily varying between 0 and 1 for w varying from 0 (or w_{wilt}) to w_{max}.

2.2 Second generation

A big step was taken when the effect of vegetation was introduced. The modelling of evaporation on two different timescales, became possible. The very rapid timescale where the intercepted water evaporates, and the slower evapotranspiration, where the photosynthesis is parameterized.

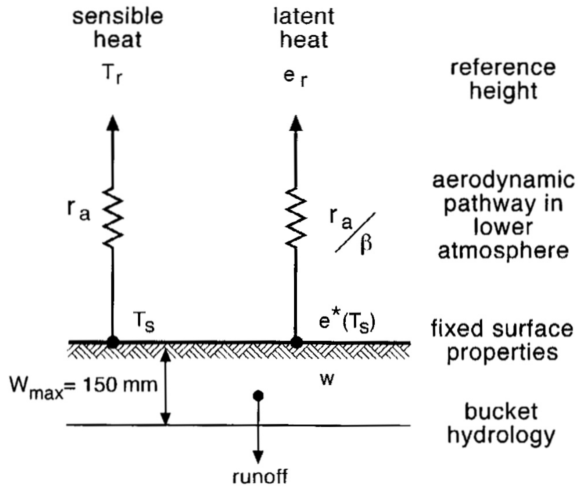


Figure 1: Figure after Pitman(2003)

Also the introduction of the so called *force-restore* formulation where the upper layer temperature forced by the surface energy balance is modified by a "deep" temperature (e.g.Deardorff, 1978, Noilhan and Planton, 1989):

$$\frac{\partial T_s}{\partial t} = C_T G - \frac{2\pi}{\tau}(T_s - T_2)$$

$$\frac{\partial T_2}{\partial t} = \frac{1}{\tau}(T_s - T_2)$$

Here T_s could be interpreted as the temperature of both the canopy and the uppermost soil layer, dependent on how much of the ground that is covered by vegetation (parameter $veg, 0 \leq veg \leq 1$). τ is one day.

In ISBA (Noilhan and Planton,1989) the water in the soil for a surface layer (w_g), and a total layer (including the root zone) (w_2) is given by:

$$\frac{\partial w_g}{\partial t} = \frac{C_1}{\rho_w d_1}(P_g - E_g) - \frac{C_2}{\tau}(w_g - w_{geq}) \ , \quad 0 \leq w_g \leq w_{sat}$$

$$\frac{\partial w_2}{\partial t} = \frac{1}{\rho_w d_2}(P_g - E_g - E_{tr}) \ , \quad 0 \leq w_2 \leq w_{sat}$$

Here C_1 and C_2 are functions of the soil type and have been calibrated using a multilayer model. w_{geq} is the moisture where gravity is balancing the capillary forces (field capacity).

The modelling of the evapotranspiration E_{tr} is important, and the stomata of the leaves regulate the evaporation, and could be modelled by the stomatal resistance r_{st} (Jarvis, 1976):

$$r_{st}^{-1} = g_{st} = g_{st}(PAR)[f(\delta e) f(T) f(w)]$$

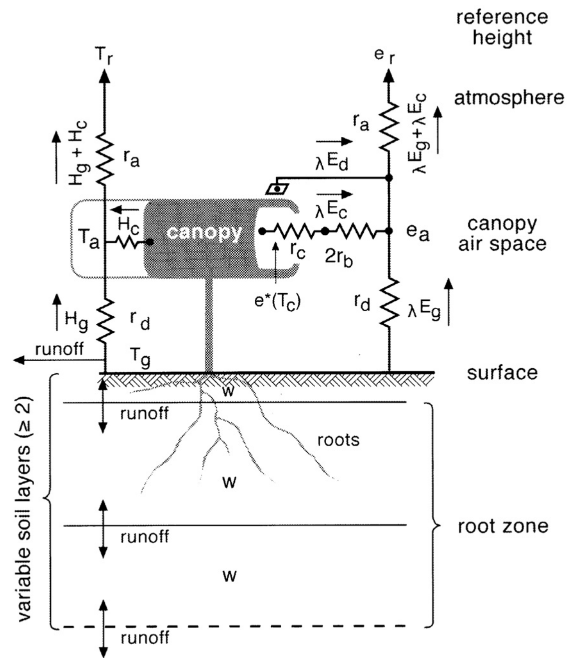


Figure 2: Figure after Pitman(2003)

Here PAR is the photosynthetically active radiation, δe is the vapour difference between the plant and the surrounding air, T the air temperature and f is a function of the water amount in the root zone (water stress).

Also the introduction of more layers in the soil, both for temperature and moisture can be regarded as belonging to this generation of surface models. The introduction of a separate temperature of the canopy is a further refinement.

2.3 Third generation

The photosynthesis is a biological process, where the plants import carbon dioxide, and while this is done, the stomata are open and the evapotranspiration takes place. In the third generation these chemical/biological processes are modelled more accurately. This is important for climate simulations, where the amount of CO_2 is increasing, and also the feedback on the vegetation growth can be taken into account. In the second generation models, the photosynthesis is tuned for the present climate, by a simpler paramemerization.

3 Skin temperature and heat conduction

3.1 Zero heat capacity

To describe the diurnal cycle in temperature one must be able to treat the short time scales of the soil and/or the vegetation. One method is to use the so called *skin temperature* which can be derived diagnostically by solving the energy balance and assuming zero heat capacity, i.e.:

$S(1-\alpha) + \epsilon(L_w - \sigma {T_{sk}}^4) - H - \lambda E = G$

G is describing the flux into the soil from the zero heat capacity skin layer and can be described as:

$G = \Lambda_{sk}(T_{sk} - T_1) \qquad \Lambda_{sk}\ [Wm^{-2}K^{-1}]$ is the skin conductivity.

This formulation is used e.g. at ECMWF (except for a small portion of shortwave radiation that goes directly to the uppermost soil layer). For the tile "high vegetation", the different fluxes in stable and unstable conditions, between the trees and the ground is parameterized by using different values of Λ_{sk}. The use of skin temperature makes it possible to tune the temperature evolution.

3.2 Non-zero heat capacity

Another way to describe the ground temperature, is to choose the uppermost layer of soil thin enough (1 cm), and write the time evolution in the same way as the force restore formulation, where now we are dealing with soil with a " given" heat capacity:

$$\rho c \Delta z_1 \frac{\partial T_1}{\partial t} = S(1-\alpha) + \epsilon(L_w - \sigma {T_1}^4) - H - \lambda E - \frac{\lambda}{\Delta z}(T_1 - T_2)$$

This in turn requires, in cases of high vegetation, that the flow between the canopy and the ground below is described in a realistic way, e.g. as a function of the canopy temperature, which now should be modelled as a prognostic variable. The above equation, can cause numerical problems if not treated implicitely.

4 The formulation of tiled schemes

The gridsquare is generally heterogeneous, and this must be taken into account. From the atmospheric point of view, we are interested of grid average values of sensible and latent heat fluxes and the momentum flux. There are in principle two ways to treat inhomogenities, either to estimate mean values over the gridsquare of $z0$, albedo etc. and compute only one energy balance (parameter aggregation), or to use separate energy balances for different fractions of the gridsquare (tiling).

There is also a possibility to use a mix of these methods, which is often the case. Parameter aggregation is simple, but less physical than tiling, since the fluxes are very nonlinear. Most modern schemes use tiling, but parameter aggregation is robust and safe, and are used in earlier schemes.

4.1 Tiling in the soil

Some schemes keep the same profile in the soil for all tiles, e.g. ECMWF, where the tiling is only confined to the skin layer (the net energy flux into the soil is given by a weighted average over the tiles), while a pure tiled scheme, like Hirlam-ISBA is storing the soil variables separately for each tile. The latter is more physical and should be important for e.g. soil under or outide the snow cover.

5 Northern Europe problems, lakes and snow

5.1 Lakes

Generally there is a large inhomogenity in Northern Europe, due to the big number of lakes (in Sweden about 95000 lakes > 100m x 100m, the total volume is about 588 km^3). Both Finland and Sweden has roughly 10 % of the area coverd by lakes. Probably the most important thing is to keep track of wether the lakes are frozen or not. In the beginning of the winter and in spring, there are big differences in this respect between different lakes.

Lake models exist, like PROBE (Ljungemyr et.al., 1996), where the lakes are treated as one-dimensional thermodynamic boxes, covering a part of the gridsquare. A simple parameterization, of wether the lake is frozen or not, is to relate this to a deep temperature in the soil, which has a timescale near that of lakes.

5.2 Snow

When modelling snow one of the biggest problems is the estimation of the snowcover. In surface schemes, which are non-tiled, a parameter aggregation of snow/no snow makes the schemes robust. In a tiled scheme, whith a separate temperature for the snow tile, things become more sensitive, since the snow coverage is directly related to the snow depth, and thus the temperature evolution.

As an example of a snow scheme, we take the snow scheme of RCA3 (from the Swedish climate centre, Rossby Centre), developed by P. Samuelsson and S. Gollvik. A common way to estimate the fraction of snow is by:

$frsn = sn/sncrit$ where $sncrit$ is a parameter (≈ 0.015 m watereq.)

There are indications that this relation is not very good. It has been shown by Lindström et.al. 2000, that there is an hysteresis effect in the snow coverage, such that there is less coverage during the melting phase, for the same snow amount. The following algorithm is adopded for the coverage in RCA3:

We use a simple formula for the growth of snow cover (for numerical reasons)

$frsn = frsnlim * tanh(100 * sn)$, $(frsnlim{=}0.95)$

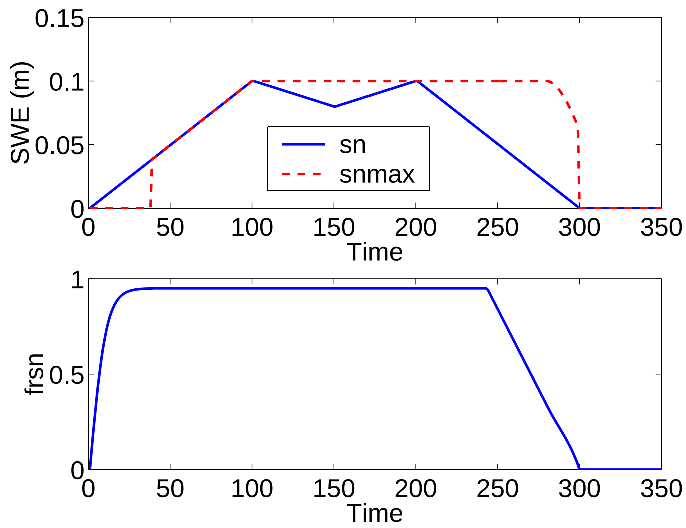
We also define an extra variable $snmax$ which initially is zero, and is put to sn when we reach full coverage. Then for the melting phase, we don't decrease the snow cover until we reach a factor $sndist$ (≈ 0.6) of $snmax$:

$frsn = sn/(snmax * sfdist)$, $frsn \leq frsnlim$

We then let $snmax$ gradually decrease, during the melting period (B. Bringfelt):

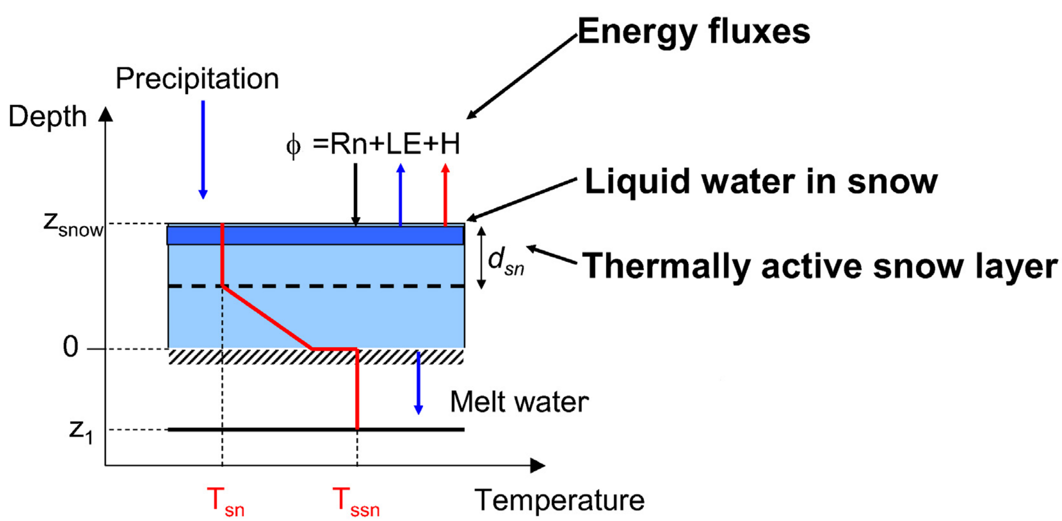
$snmax(\tau + 1) = snmax(\tau) - (zk1 * snmax - sn(\tau + 1) * (1 - zk)/zk1$

$zk1 = 0.2$ and $zk = \exp(10^6 * \Delta t)$, used if $sn(\tau + 1) < zk1 * snmax$



5.2.1 Heat conduction

Here we use only one layer of snow, the depth of which is Z_{snow} [m snow]. Only the upper part is thermally active in cases of deep snow:



The time evolution of the snow temperature is given by:

$$\frac{dT_{sn}}{dt} = \frac{1}{c_{snow} * MIN(Z_{snow}, d_{sn})}[\Phi - \alpha_{snow}(T_{sn} - T_{ssn})]$$

$c_{snow} = vhice * \rho_{sn}/\rho_{ice}$

Here the coefficient α_{snow} (formulation from ERA 40) is parameterizing a "fictive" profile through the snow, since the isolation is a function of the snowdepth:

$$\alpha_{snow}^{-1} = 0.5\frac{Z_{snow}}{\lambda_{sn}} + 0.5\frac{Z_1}{\lambda_{soil}} \ ; \quad \lambda_{sn} = \lambda_{ice}\left(\frac{\rho_{sn}}{\rho_{ice}}\right)^{1.88}$$

5.2.2 Melting/freezing

The melting is straight forward and is invoked if the total flux into the snow pack is large enough to reach the melting point. The melted water is kept in the snow, until it reaches 10 % of the total snow water equivalent, the excess is going to the soil. The temperature T_{sn} is kept at 0° C. Freezing of the water in the snow, is less straight forward, since the negative energy flux must be partitioned between freezing and coling of the snow. We parameterize this fraction, *freezefrac*, as a function of snow depth, and water in the snow. Technically the melting/frezing is done by partitioning the timestep into temperature changes and phase shift.

5.2.3 Density and albedo changes

Here we use a simple method from Douville et.al. 1995, where the snow density is increasing with time, (e-folding time ≈ 4 days), and modified with new snow with less density. Then a weighteing between "dry" snow and the water in the snow is done. Also the albedo is following Douville et.al., and varies between 0.5 for old and wet snow, and 0.85 for fresh new fallen snow

6 Something about Hirlam-ISBA surface analysis

The analysis of soil temperatures is difficult, due to lack of observations. Instead the screen level temperature, T_{2m}, is analysed by optimal interpolation. The analysis increments is added to each tile (only tiles 3-5):

$\Delta T_{si} = \Delta T_{2m}$, i = 3,4,5

Then for the restore temperatures T_{2i} :

$\Delta T_{2i} = \Delta T_{si}/2\pi$

This is of course only one assumption of many! In the version of the analysis coupled to the new snow scheme, where we solve a heat conduction problem, we calculate the values in the soil by using analysed values of the uppermost layers

7 Complexity versus tiling problems

When developing surface models, it is possible to do more and more refined physical formulation, of the relevant processes. However, one must always bare in mind, the limitation that comes from the uncertainties in the horizontal representations. In the case of tiles, usually the fractions of different properties are given. In reality also the statistical distribution of these fractions are important, and also subgrid scale flow patterns, can significantly change the grid average fluxes. In other words, if half the gridsquare is forest and the other half is sea, the real fluxes differ if the air is advected from the sea or from the forest. Moreover if the water is more patchy (archipellago), even more complex fluxes can appear. A pure tiled scheme implies that there exists a local turbulent equilibrium for each tile, and that parameters like screen level temperatures are meaningful to estimate. This means that the only communication between the tiles goes via the lowest model layer, which then in turn should be high enough over the ground, to be described by one single value, i.e. the lowest model layer should be above blending height (the height at which the individual inhomogenities of the ground are not felt).

It is often stated that these problems dissapear when the horizontal resolution is increased, so that non-tiled schemes can be used. If there are large horizontal differences in the surface characteristics, there might be numerical problems, caused by these inhomogenities.

8 References

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