Land Surface Schemes

1) CLASS (The Canadian Land Surface Scheme)

Developed at Environment Canada. Originally described in:

![Diagram of land surface schemes](image)

Lateral heat flow is neglected; the finite-difference form of the one-dimensional heat conservation equation is applied to each layer to obtain the change in average layer temperature $T_i$ over a time step $\Delta t$:

$$T_i(t + 1) = T_i(t) + \left[ G(z_{i-1}, t) - G(z_i, t) \right] \frac{\Delta t}{C_i \Delta z_i} + S_i$$  (1)

where $G(z_{i-1}, t)$ and $G(z_i, t)$ are the downward heat fluxes at the top and bottom of the layer, respectively, $C_i$ is the volumetric heat capacity of the soil, $\Delta z_i$ is the layer depth, and $S_i$ is a correction term applied in case of freezing or thawing, or the percolation of ground water (see section 2.2). $G$ and $z$ are both taken to be positive downward.

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To evaluate the surface temperature, the surface energy balance equation is expressed as a non-linear function of $T(0)$ and solved iteratively. The energy balance equation is given by

$$K_* + L_* + Q_H + Q_E = G(0)$$

(7)

where $K_*$ and $L_*$ are, respectively, the net shortwave and net longwave radiation absorbed at the surface, $Q_H$ and $Q_E$ are the sensible and latent heat fluxes, and $G(0)$ is the surface heat flux into the ground.

The net shortwave radiation $K_*$ depends on the incoming shortwave radiation $K^I$ and the ground surface albedo $\alpha_*$:

$$K_* = (1 - \alpha_*)K^I$$

(8)

The net longwave radiation absorbed at the surface, $L_*$, is given by the difference between the incoming atmospheric radiation $L^I$ and the radiation emitted by the surface:

$$L_* = L^I - \sigma T(0)^4$$

(10)

where $\sigma$ is the Stefan–Boltzmann constant. The surface is assumed to radiate as a black body; further refinement is a useless complication at this stage, since the effective emissivity depends not only on the measured surface value but on the effects of local microtopography.

The sensible and latent heat fluxes $Q_H$ and $Q_E$ are given by the bulk transfer formulae

$$Q_H = \rho_a c_p V_a c_D [T_* - T(0)]$$

(11)

and

$$Q_E = L_* \rho_a c_v V_a c_D [q_* - q(0)]$$

(12)

where $\rho_a$, $c_p$, $T_*$, and $q_*$ represent the density, specific heat, temperature, and specific humidity, respectively, of air in the constant flux layer, $V_a$ is the wind speed, $L_*$ is the latent heat of vaporization (or sublimation, if a snow pack is present), and $c_D$ is a drag coefficient that depends on surface roughness length, wind speed and atmospheric stability (McFarlane and Laprise, 1985).

2) ISBA (Interactions of the Soil-Biosphere-Atmosphere)

Available at [http://www.cnrm.meteo.fr/isbadoc/model.html](http://www.cnrm.meteo.fr/isbadoc/model.html)

Developed at CNRM, France. Originally described in:


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

$$\frac{\partial T_{surf}}{\partial t} = C_{TOT} (R_n - H - LE) + C_T L_f \left( \text{freez}_g - \text{melt}_g + \text{freez}_s - \text{melt}_s \right) - \frac{2\pi}{\tau} (T_{surf} - T_p)$$

$$\frac{\partial T_p}{\partial t} = \frac{1}{\tau} (T_{surf} - T_p)$$
in which $H$, $LE$, and $Rn$ are the sensible heat, latent heat, and net radiational fluxes at the surface, CTOT is a thermal coefficient, $Lf$ is the latent heat of fusion, freezes and melts are fluxes of freezing and melting snow, and $t$ 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.

The net radiation at the surface is

$$R_n = F_{SS}^{-}(1 - \alpha) + \varepsilon_F(F_{SI}^{-} - \sigma_{SB}T_{surf}^4)$$

where $F_{SS}^{-}$ and $F_{SI}^{-}$ are the incoming solar and infrared radiation at the surface, and $\sigma_{SB}$ 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:

$$H = \rho_a c_p C_T u_T (T_{surf} - T_a)$$

where $c_p$ is the specific heat; $\rho_a$ 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., $E_s$):

$$E = L_e E_g + L_v E_v + L_s E_s$$

$$E_g = (1 - \text{veg})(1 - p_{avg})\rho_a C_T u_T (h_s q_{sat}(T_{surf}) - q_a)$$

$$E_v = \text{veg}(1 - p_{avg})\rho_a C_T u_T h_v (q_{sat}(T_{surf}) - q_a)$$

$$E_s = p_{sn} \rho_a C_T u_T (q_{sat}(T_{surf}) - q_a)$$

where $L_v$ and $L_s$ 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.

Fig. 1. Hydrological budget in ISBA.