CLASS -

“Canadian Land Surface Scheme”

Background to Version 3.0
Technical Data on CLASS

- ~10,000 lines of code
- 45 subroutines
- 14 prognostic variables
- almost 700 other internal and background variables
- time step = 30 minutes or less
- written in FORTRAN
- best run at 64 bit precision to avoid round-off problems

Inputs Required to Drive CLASS at each Time Step

- day of year and time of day
- incoming solar radiation: visible, near-IR and diffuse (partitioning can be estimated)
- incoming longwave radiation
- wind speed, specific humidity and adiabatically-extrapolated air temperature from lowest model level, at a specified reference height in the atmosphere
- rainfall and snowfall rate (partitioning can be estimated)
- surface air pressure

Energy Balance Calculations in CLASS

Bare soil:

The surface energy balance equation for bare soils can be written as follows:

\[ K^* g + L^* g + Q_{H,g} + Q_{E,g} = G(0) \]

The net shortwave radiation, \( K^* g \), is calculated as the difference between the incoming and reflected radiation, determined by the ground surface albedo. The albedo is calculated as a function of soil texture and moisture content of the top layer.
The net longwave radiation is determined as the difference between the incoming and outgoing radiation, expressed as $\sigma T(0)^4$.

The ground sensible heat flux is calculated from the relation $Q_{H,g} = \rho_a c_p \frac{(T_a - T(0))}{r_a}$, where $\rho_a$ is the air density, $c_p$ is the specific heat, and $r_a$ is the aerodynamic resistance. The latter is calculated using Monin-Obukhov theory, according to Abdella and McFarlane.

The ground latent heat flux is calculated from the relation $Q_{E,g} = \beta L \rho_a \frac{(q_a - q(0))}{r_a}$, where $\beta$ is an evaporation efficiency parameter depending on the soil texture and soil moisture, $L$ is the latent heat of vaporization, and $q$ is the specific humidity.

The ground heat flux, $G(0)$, is expressed as a linear function of one unknown, $T(0)$, by making use of the following equations and assumptions:

$G(z) = -\lambda(z) \left( \frac{\delta T}{\delta z} \right)$ for heat flow between soil layers, where $\lambda$ is the thermal conductivity;

$\Delta T/\Delta t = -(1/C(z)) \left( \frac{\Delta G}{\Delta z} \right)$ for temperature change of soil layers over a given time step, where $C$ is the volumetric heat capacity;

$T$ is expressed as a quadratic function of depth, as $T(z) = az^2 + bz + c$, and integrating this equation over the thickness of a soil layer leads to an expression for the average temperature of the layer;

$G(z) = 0$ at the bottom of the lowest soil layer;

$G(z)$ at the bottom of a soil layer = $G(z)$ at the top of the layer below;

$T(z)$ at the bottom of a soil layer = $T(z)$ at the top of the layer below.

Combining these equations for each soil layer and making use of these assumptions leads to an expression for $G(0)$ that is a function only of $T(0)$ and a set of known variables including the thermal conductivities of the soil layers and the layer average temperatures.
The thermal conductivity of each soil layer is calculated as the geometric mean of the thermal conductivities of the constituent elements (water, ice, air and soil minerals) for mineral soils, and according to an empirical equation derived by Farouki for organic soils.

The soil volumetric heat capacity is calculated as the arithmetic mean of the heat capacities of the constituent elements for all soils.

Combining all of the equations for the ground energy balance terms leads to a non-linear expression for the overall surface energy balance in one unknown, $T(0)$, which is solved by iteration. $T(0)$ is then back-substituted to determine the values of the individual components of the energy balance, and to determine the rates of heat conduction between soil layers.

**Snow cover:**

If snow is present, it is treated like a fourth “soil” layer, using the same formulation for the surface energy balance as for bare soil, and the same formulation for heat conduction between layers but with values of $\lambda$ and $C$ appropriate for snow.

The snow albedo decreases exponentially with time from a specified value appropriate for new snow, using an empirical relation derived from field data. The density increases exponentially with time from a fresh-snow value.

Melting occurs if the surface iteration process projects a snow surface temperature greater than $0^\circ$C, or the forward stepping of the snow layer temperature projects a layer temperature greater than $0^\circ$C. In these cases the excess energy is used to melt part or all of the snow pack, and the temperature of the affected layer is set back to $0^\circ$C. Surface snow meltwater percolates into the pack and refreezes, releasing latent heat, until the temperature of the entire pack reaches $0^\circ$C; thereafter it reaches the soil surface, enters the surface ponded water store, and becomes available for infiltration.

Snow cover is assumed to reach 100% of the grid surface when the snow depth is $\geq 0.1$ m. If the projected snow depth falls below 0.1 m, it is reset to zero and a fractional ground coverage is calculated based on conservation of mass.
Vegetation:

The surface energy balance equation for vegetation can be written as follows:

\[ K^*_c + L^*_c + Q_{H,c} + Q_{E,c} = C_c \Delta T_c / \Delta t \]

The net shortwave radiation \( K^*_c \) is determined by the bulk albedo and transmissivity of the vegetation canopy, and the albedo of the ground surface below. The canopy albedo and transmissivity are calculated separately for the four main vegetation types (coniferous trees, deciduous trees, crops and grass) and separately for visible and near-IR radiation as functions of the solar zenith angle and the leaf area index, and then lumped.

The net longwave radiation \( L^*_c \) is determined by the downwelling and upwelling longwave radiation from the sky and the ground surface respectively, the canopy temperature, and the canopy gap fraction (calculated separately as a function of leaf area index for the four main vegetation types and then lumped).

The sensible heat flux from the canopy is calculated using the relation \( Q_{H,c} = \rho_a c_p \left[ T_a - T_c \right] / r_a \), where \( r_a \) is the aerodynamic resistance. The latter is calculated after Abdella and McFarlane, as for soils. (Modifications have been made to this formulation in version 3.0.)

The latent heat flux from the canopy is calculated as \( Q_{E,c} = L \rho_a \left[ q_a - q_c \right] / [r_a + r_c] \), where \( q_c \) is the saturation specific humidity at the canopy leaf surfaces, and \( r_c \) is the stomatal resistance to vapour transfer. The stomatal resistance is a function of the limiting minimum value of \( r_c \) (depending on vegetation type); the leaf area index; and four empirical functions in the incoming shortwave radiation, the vapour pressure deficit of the air, the air temperature, and the soil moisture. (If the canopy is covered with intercepted rain or snow, \( r_c \) is assumed to be zero until the intercepted moisture has disappeared.)

For sensible and latent heat fluxes from the ground under the canopy, the wind speed is assumed to be small, so fluxes are considered to be negligible under stable conditions (canopy warmer than ground). For unstable conditions, the transfer coefficient is calculated after Deardorff as a function of the surface-air temperature difference.
As in the case of the ground surface temperature, substituting these expressions into the canopy energy balance equation yields an equation in one unknown, the canopy temperature $T_c$. This equation is solved by iteration, correcting for phase changes of intercepted water if the temperature crosses 0°C, and $T_c$ is then back-substituted to determine the individual components of the canopy energy balance.

**Moisture Balance Calculations in CLASS**

**Bare soil:**

Under conditions of no precipitation, the water flux at the soil surface, $F(0)$, is given by the surface evaporation rate. The flux between soil layers is calculated from the Darcy flow equation, as

$$F(z) = k(z) \left[ \frac{\delta \psi}{\delta z} \right]_{z + 1}$$

where $k(z)$ is the thermal conductivity and $\psi(z)$ is the soil moisture suction at the layer interface. Both are calculated according to empirical functions of soil texture and moisture, derived by Cosby et al. (Parameters appropriate for organic soils in these equations have been derived separately by Letts et al.)

If infiltration is occurring at the current time step, $F(0) = I$, where $I$ is the surface infiltration rate. How much water infiltrates during the time step is governed by the amount of water available from rainfall, snowmelt or ponded water, and the soil hydraulic properties and presence of impermeable (e.g. saturated frozen) layers. If the time-varying infiltration rate falls below the rainfall rate during the current time step, infiltration continues but water begins to pond on the soil surface, up to a maximum surface detention capacity. This ponded water is retained on the surface for subsequent time steps until it has infiltrated or evaporated.

**Vegetation:**

Precipitation arriving at the vegetation top is either intercepted by the canopy, or falls through gaps in it. Interception proceeds until the interception capacity has been filled, after which rain or snow drips off onto the ground
beneath. The interception capacity is calculated as a linear function of leaf area index for rain. (For snow in version 3.0, a different formulation is used.)

Evaporation from the canopy is assumed to deplete the intercepted moisture store first, since the surface resistance associated with this flux is zero. From dry canopies, moisture is taken, via transpiration, from the soil. The amount of water extracted from a soil layer depends on a weighted function of the fraction of roots contained in that layer, and the soil moisture suction.

Vegetation parameters such as height, rooting depth and mass are assumed to be constant for forests and grass. For crops, an annual cycle is imposed, depending on latitude and longitude of the grid cell, reflecting typical dates of planting and harvest. Thus, “crop” areas can be bare for parts of the year. Leaf area index undergoes an annual variation in temperate and boreal forests, with “leaf-out” and senescence dates triggered by changes in the air and soil temperatures. Crop leaf area indices track the annual variations in growth stage. Lumped, bulk-canopy values of these variables are calculated by areally averaging over the vegetation types present.

**Grid Cell Subdivision in CLASS (version 2.x)**

Each grid cell is divided into four possible subareas: bare soil, vegetation, bare soil with a snow pack, and vegetation with a snow pack. The surface energy balance equations are solved separately for each subarea. The resultant fluxes, and the prognostic soil, snow and vegetation canopy variables, are aggregated after each time step to produce grid cell average values.

Vegetation characteristics are averaged in the pre-processing stage over vegetation types present on the grid cell, to determine parameters for the four major vegetation sub-groups. Instantaneous values of canopy parameters are determined at each time step for the four major vegetation sub-groups, and aggregated to produce bulk canopy parameters.

Fractional snow cover is re-calculated at every time step on the basis of snow mass present. Short vegetation (crops and grass) can be buried by snow, which affects the surface properties.
Vegetation Types Listed in Standard CLASS Lookup Table

*Lumped vegetation canopy:*

- Coniferous trees
  - Evergreen needleleaf trees
  - Deciduous needleleaf trees
- Broadleaf trees
  - Evergreen broadleaf trees
  - Deciduous broadleaf trees
  - Tropical broadleaf trees
  - Drought deciduous trees
  - Deciduous shrub
  - Thorn shrub
- Crops
  - Arable
  - Rice
  - Sugar
  - Maize
  - Cotton
  - Irrigated crop
- Grass
  - Evergreen broadleaf shrub
  - Short grass and forbs
  - Long grass
  - Tundra
  - Swamp
- Urban
Source dataset (Wilson and Henderson-Sellers, 1985) used in GCM runs lists primary and secondary vegetation types at 1° x 1° resolution globally.

In a pre-processing stage, values of the following variables are assigned to each of the W&H-S vegetations classes, based on values reported in the literature:

- visible and near-IR albedo
- annual maximum aerodynamic roughness length
- annual maximum and minimum LAI
- annual maximum canopy mass
- rooting depth

Average values for these variables for each of the four grouped vegetation classes are then determined over each of the model grid cells at the desired resolution, by areally weighting the values for the contributing W&H-S vegetation types present over the grid cell.

At each model time step, values of instantaneous vegetation characteristics are calculated for each of the four grouped vegetation classes, including visible and near-IR albedo, canopy transmissivity, leaf area index, roughness length, minimum stomatal resistance, mass, rooting depth, and sky view factor. These are then averaged to produce aggregated values to characterize the lumped canopy.