TerraE Global Land Model

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The TerraE Global Land Model ("TerraE") simulates water and heat transport on a global grid for a set of land tiles within each grid cell. Currently, we have one tile for bare soil and one for vegetation (assuming both exist within a grid cell). For each tile, we have a discretized soil column. The default setup in ModelE2.1 (and previous versions back to at least the 1990s) had a column that extends to a depth of 3.5 m. Each of these soil columns is discretized into six vertical layers; the layer thicknesses are 0.1 m, 0.17 m, 0.30 m, 0.51 m, 0.89 m and 1.53 m (Abramopoulos et al., 1988; Rosenzweig and Abramopoulos, 1997).

ModelE3 features a land column with the following layering (bottom layer depths): 0.05 m, 0.14 m, 0.30 m, 0.587 m, 1.102m, 2.024 m, 3.676 m, 6.635 m, 11.936 m, 21.431 m. The soils are characterized by new texture data from De Lannoy et al. (2014), new data on land-surface slope (add ref), and new depth-to-bedrock data based on Pelletier et al. (2016). The deeper land column together with the depth-to-bedrock information allow us to implement groundwater in the model.

The state of each layer i is described by two prognostic variables: heat content of the layer and the total amount of water in the layer. Soil texture is specified as mixtures of five different components: sand, silt, clay, peat, and bedrock. The vegetation model uses soil texture information as part of the water stress calculation. This soil texture information is also used for soil biogeochemistry.

Interface Between Land Physics and Vegetation

The land physics of the TerraE interacts with vegetation, currently simulated by the Ent Terrestrial Biosphere Model (TBM) (Kim et al., 2015). The interactions occur through the module 'ent_mod' in ent_mod.f. All internal data in Ent TBM is intentionally hidden from an external code that calls it, except for the data that is explicitly passed using the subroutines 'ent_get_exports' and 'ent_set_forcings' from module 'ent_mod'. In particular, the data listed in Tables ##2-1 and 2-2## are exchanged between Ent TBM and TerraE using these subroutines each time step, Δt. For TerraE, Δt ranges from 5 seconds to 30 minutes (ModelE's time-step size), depending on stability constraints for the explicit solutions of the water and energy balance equations. The subroutine 'advnc' in module 'sle001' (GHY.f) is called to advance the water- and energy-related variables.

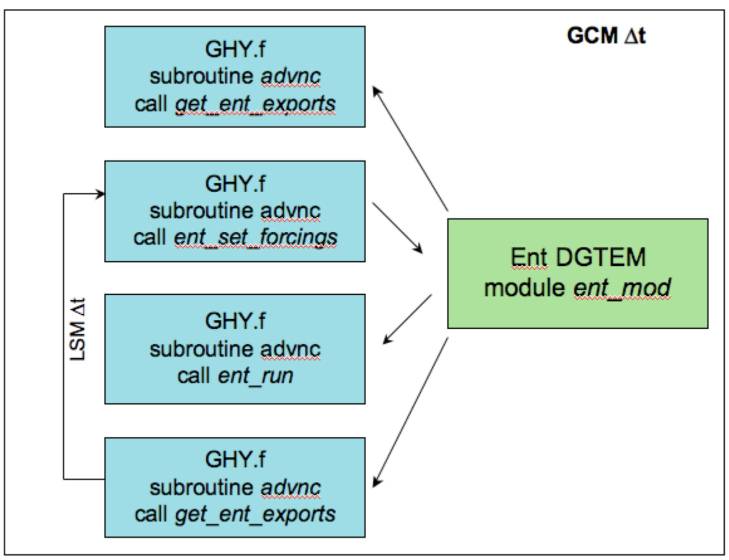

Figure 1: Schematic diagram of the interface between TerraE and Ent TBM. The physics code of TerraE is contained within the file GHY.f. The interactions of TerraE with Ent TBM come through calls to *get_ent_exports*, *ent_set_forcings*, and *ent_run*, which are all part of the Ent module *ent_mod*.

Figure 1 presents the interface between TerraE and Ent. At the beginning of the GCM's timestep, the data that TerraE requires from Ent are obtained in subroutine 'advnc' through a call to the subroutine 'ent_get_exports'. The data then needed by Ent from TerraE are then transferred to module 'ent_mod' through a call to subroutine ‘ent_set_forcings’. Once these data are transferred, TerraE calls subroutine ‘ent_run’ to simulate the processes in the Ent TBM for a single grid cell and single Ent TBM time step. After this call, TerraE makes an additional call to the Ent TBM with the subroutine ‘ent_get_exports’ to obtain the variables listed in Table ##2-1##. These updated variables are then used to provide fluxes of water and heat between the canopy and the atmosphere (including evapotranspiration in subroutine ‘evapo_limits’) and among the canopy and soil layers. The values passed from Ent are all Ent-cell averaged values (averaged over the cohorts and then over the patches of Ent). An additional step of accumulating variables from Ent in time (averaged or summed, as appropriate) to the GCM’s time step is necessary, because TerraE had variable Δt due to its explicit solution procedure. For a model that has the same time step size for the land and atmosphere portions of its GCM, this accumulation step can be avoided and the interface presented in Figure 1 is simplified.

Main processes linking land hydrology with vegetation

The processes represented in the Ent TBM directly control fluxes of water and energy for the vegetated section, while bare-soil dynamics are indirectly affected through changes in the extent of vegetation cover and climatic feedbacks. For the vegetated fraction of a grid cell, the representation of vegetation structure differs between Ent and TerraE, as shown in Figure 2. TerraE treats the canopy as a single leaf (a so-called big-leaf model) for certain water and energy processes, while Ent simulates ecosystem processes for cohorts (i.e., ensembles of identical individuals that are distinguished by plant functional type and size). These vegetation-structure differences are relevant for processes that control ecosystem interactions with incoming precipitation and energy. We describe selected processes of TerraE below to highlight important processes for users linking Ent with other land models, which is especially useful given that TerraE has a modeling framework that is common to most GCM land surface schemes. Our focus is on evapotranspiration, which is GISS LSM process directly coupled to Ent’s predictions of water and carbon. In particular, the incorporation of Ent to TerraE has highlighted the need to improve the model’s representation of evapotranspiration components: canopy evaporation, E_c, transpiration, E_t, vegetated soil evaporation, E_{vs}, and snow evaporation, E_{vsn}.

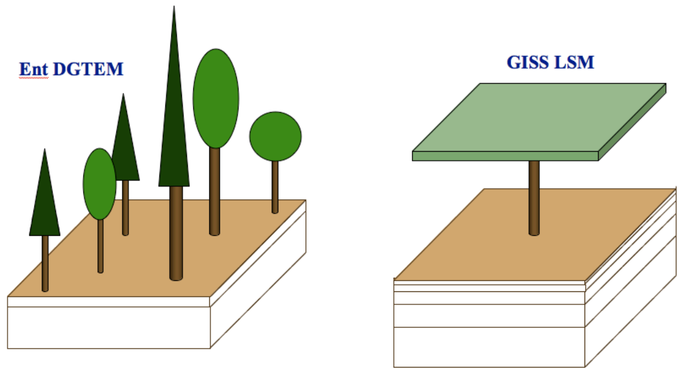

Figure 2: Sketch demonstrating the differences in vegetation structure for Ent and TerraE, which has implications for water and energy dynamics.

Potential evapotranspiration

As described above, potential evaporation, E_p, is expressed as

$$E_p = \frac{\rho_a}{\rho_w} C_q V_s \left(q_{sat} \left(T_{can} \right) - q_s \right)$$

where ρ_a is the density of surface air [kg/m^3], ρ_w is water density [kg/m^3], q_s is the humidity of surface air [-], q_{sat} is the saturated humidity at the temperature of the canopy T_{can}, V_s is wind speed, and C_q is the humidity transfer coefficient [-].

TerraE code for potenial Canopy Evaporation

```
!      pot_evap_can is the evaporation that will occur
!      based on atmospheric demand
!       pot_evap_can = betat*rho3*cna*(qsat(tp(0,2)+tfrz,lhe,pres) - qs)
pot_evap_can = betat*rho3*ch*(vs*(qsat(tp(0,2)+tfrz,lhe,pres) - qs)-(vs-
vs0)*qprime)

c     potential evaporation for bare and vegetated soil and snow
c     epb  = rho3*cna*(qb-qs)
c     epbs = rho3*cna*(qbs-qs)
c     epv  = rho3*cna*(qv-qs)
c     epvs = rho3*cna*(qvs-qs)
v_qprime=(vs-vs0)*qprime
epb  = rho3*ch*( vs*(qb-qs) -v_qprime )
epbs = rho3*ch*( vs*(qbs-qs)-v_qprime )
epv  = rho3*ch*( vs*(qv-qs) -v_qprime )
epvs = rho3*ch*( vs*(qvs-qs)-v_qprime )
```

TerraE code for Penman-Monteith Potential Evapotranspiration

```
c**** calculation of penman value of potential evaporation, aepp
h0=fb*(snsh_tot(1)+elh*evap_tot(1))+fv*(snsh_tot(2)+elh*evap_tot(2))
el0=elh*1d-3
qsats=qsat(ts,lhe,pres)
dqdt = dqsatdt(ts,lhe)*qsats
! cpfac=sha*rho*ch*vs
! epen=(dqdt*h0+cpfac*(qsats-qs))/(el0*dqdt+sha)
epen=(dqdt*h0+sha*rho*ch*vs*(qsats-qs)-(vs-vs0)*qprime ))/(el0*dqdt+sha)
aepp=epen*dt
```

Interception

The canopy interception algorithm of a land surface model has a significant effect on its partitioning of precipitation between evapotranspiration and runoff (Pitman et al., 1990; Wang et al., 2007) as well as on its partitioning of evapotranspiration between canopy evaporation and transpiration (e.g., Wang and Eltahir, 2000; Wang et al., 2005). Most land surface models represent the reservoir on the vegetation's canopy with a water balance equation for a single point, but difference among models arise due to varying representation of the reservoir's interception, drip, and evaporation rates (Wang et al., 2007). The existing interception scheme of Rosenzweig and Abramopoulos (1997) is expressed as

$$\frac{\partial w_c}{\partial t} = I_{c,net} - f_{c,wet}\left(1 - f_{sn} f_{sn,c}\right) E_{v,w}$$

where w_c is the canopy water storage [m], t is time [s], $I_{c,net}$ is the canopy interception rate not including evaporation losses [m/s], $f_{c,wet}$ is the wet fraction of the canopy [-], f_{sn} is the fraction of the grid cell that is covered by snow [m^2 snow/m^2 ground], $f_{sn,c}$ is the fraction of the canopy that is covered by snow [m^2 snow/m^2 canopy], and $E_{v,w}$ is the evaporation rate from the canopy reservoir (wet canopy) [m/s]. These rates are typically parameterized differently to account for the effects of subgrid precipitation variability. The evaporation from the wet canopy, $E_{v,w}$, is compute based on potential evaporation, E_p (described above) and the water stored in the canopy as

$$E_{v,w} = \min\left[E_p, \frac{w_c}{\Delta t}\right]$$

The importance of the canopy interception to evapotranspiration partitioning is evident in Equation **Error! Reference source not found.**, because the amount of water in the canopy's reservoir limits $E_{v,w}$.

Global simulations with the GISS GCM reveal that the model overestimates interception (leading to greater canopy evaporation at the expense of transpiration). We therefore modify the Rosenzweig and Abramopoulos scheme to improve the models' transpiration predictions (linked to carbon uptake).

The interception overestimation is a consequence of the model's failure to account for temporal correlation in storm position, which is relevant for precipitation events that only cover a fraction of a GCM's grid cell (Koster and Suarez, 1996). For example, rainfall wets a fraction of the canopy within a grid cell for time step *i*. If, on the next time step *i*+1, the same storm has not moved significantly, then rainfall continues to fall over mostly the same area. Even though a typical convective storm can span multiple time steps in TerraE, the Rosenzweig and Abramopoulos scheme does not adjust its drip rate to account for this temporal correlation in storm position. Another consequence is that the scheme's behavior is dependent on the length of the model's time step, which is a property that should be avoided (Koster and Suarez, 1996). It is important to note that the Rosenzweig and Abramopoulos scheme is not the only interception model with this problem. Wang and Wang (2007) also found that the widely used interception model of Shuttleworth (1988) has a time-step depend behavior due to its failure to account for storm position.

The new interception scheme follows the algorithm presented in Koster and Suarez (1996) and is similar to the modified Shuttleworth scheme of Wang et al. (2007). The scheme primarily involves modification of the wet fraction of the canopy ($f_{c,wet}$). Essentially, the drip rate is computed by taking into account the fraction of precipitation that falls onto a previously wetted portion of the canopy with the grid cell. This scheme also distinguished between stratiform and convective precipitation, because the stratiform type covers a greater portion of the grid cell that the convective type (e.g., Koster and Suarez, 1996; Wang et al., 2007). See Appendix 1.

Transpiration

The transpiration component of TerraE remains the same as in previous versions, except for modifications to canopy conductance and water stress (see Section ##4##). In particular, transpiration in the GISS SVAT model is computed with the standard resistance formula, which is used in many land surface models. The potential transpiration, $E_{t,i}$, from soil layer *i* is

$$E_{t,i} = \beta_i \frac{1}{\langle C_c \rangle_{\text{ent}}^{-1} + C_a^{-1}} \frac{\rho_a}{\rho_w} \left(q_{sat}\left(T_{can}\right) - q_s \right)$$

where the symbol $\langle \cdot \rangle_{\text{ent}}$ specifies that the value is an average Ent-cell value, q_{sat} is the saturated humidity at the temperature of the canopy T_{can}, $\langle C_c \rangle_{\text{ent}}$ is the average canopy conductance [m/s], C_a is the conductance of the atmosphere [m/s], and β_i is the water stress for each layer based on the formulation given in Section ##4.1.1.5##. C_a described turbulent transport and is computed as $C_a = C_q v$, where C_q is the humidity transfer coefficient [-] and v is the wind speed at the surface [m/s]. The land surface model receives an average canopy conductance (i.e., Ent-cell level) from the Ent TBM. The actual transpiration is then computed by limiting the potential transpiration based on the amount of water divided by Δt.

Soil evaporation

In previous versions of TerraE, all surface conductance of water vapor for vegetated land was assigned to the vegetation, which meant that there was no soil evaporation under canopies. The rationale for this approach in the Abramopoulos et al. (1988) version was that vegetation was simply a surface with a particular albedo and conductance. The Friend and Kiang (2005) version had biophysics, but transpiration was tuned to total evapotranspiration measured by eddy covariance methods at FLUXNET sites. Numerous studies have demonstrated the importance of soil evaporation as a component of evapotranspiration, especially for semi-arid ecosystems (e.g., Lawrence et al., 2007).

We incorporate a representation of soil evaporation based on the combined work of Lawrence et al. (2007) and Sakaguchi and Zeng (2009). We have two approaches for soil evaporation. Version A involves a simple linear btw the transfer coefficient values for bare soil and for a thick canopy.

TERRAE CODE FOR VERSION A SOIL EVAPORATION

```
!    Transfer coefficient is simply interpolated btw the
!    values for bare soil and for a thick canopy
!    see Zeng et al (2005) and Lawrence et al (2007)
!    in the CLM
ch_dense_veg = 0.01d0*cna
eta = exp( -f_clump*(lai + sai) )
ch_vg = ch*eta + ch_dense_veg*(1.d0-eta)
epvg  = rho3*ch_vg*( vs*(qvg-qs)-v_qprime )
```

The Version B approach for soil evaporation is an attempt to be slightly more mechanistic. We write the expression for soil evaporation as

$$E_g = \frac{\rho_a}{\rho_w} \cdot \frac{q_{sat}\left(T_g\right) - q_a}{r_{aw} + r_{soil} + r_{litter}}$$

where E_g is soil evaporation [m s^{-1}], T_g is the ground temperature [°C], r_{aw} is the aerodynamic resistance [s/m] to water vapor transport between the ground and the air, r_{soil} is the soil resistance [s/m], and r_{litter} is the litter resistance [s/m]. That is, the Equation **Error! Reference source not found.** is based on Fick's law of diffusion applied to soil evaporation (Troeh et al., 1982), where it is assumed that the vapor transport within the dry layer is dictated by molecular diffusion due to the gradient of water vapor and is constant along its path.

We use a formulation of soil resistance that physically represents molecular diffusion of water vapor through the dry fraction of the soil as (Moldrup et al., 1999; Sakaguchi and Zeng, 2009)

$$r_{soil} = \frac{L}{D}$$

where D is the reduced vapor diffusivity within the soil ($m^2\ s^{-1}$) and L is the thickness of the dry part of the soil [m]. The value of D depends on soil texture, which can be estimated using an empirical expression as (Moldrup et al., 1999)

$$D = D_0 \theta_{sat} \left(1 - \frac{\theta_r}{\theta_{sat}}\right)^{2+3b_s}$$

where D_0 is the molecular diffusion coefficient of water vapor in air (assumed to be a constant value of 2.2×10^{-5} m^2 s-1), θ_{sat} is the volumetric soil moisture at saturation, and θ_r is the residual (hygroscopic) volumetric soil moisture, and b_s is a soil-texture-dependent parameter for the soil-moisture characteristic curve. Sakaguchi and Zeng (2009) derived an expression for dry layer thickness, L, which we use here as

$$L = d_1 \frac{\exp\left[\left(1 - \theta_1/\theta_{sat}\right)^w\right] - 1}{e - 1} \quad (0.1)$$

the parameter w is a shape-fitting parameter arbitrarily set to 5 by (Sakaguchi and Zeng, 2009).

TerraE code for Version B soil evaporation

```
!    1) Soil resistance computed according to the formulation of
!       Sakaguchi and Zeng (2009)
D_vapor = D0*(thets(1,2)**2.d0) (1.d0-thetm(1,2)/thets(1,2))**(2.d0+3.d0*5.d0)
L_dry = dz(1)* (exp((1.d0-theta(1,2)/thets(1,2))**5.d0)-1.d0) / (exp(1.d0)-1.d0)
r_soil = L_dry / D_vapor

!    2) Under-canopy transfer coefficient is weighted average of
!       bare soil and thick canopy values
!       (e.g. Zeng et al 2005 and Lawrence et al 2007)
ch_dense_veg = 0.01d0*cna
eta = exp( -f_clump*(lai + sai) )
ch_vg = ch*eta + ch_dense_veg*(1.d0-eta)

!    3) Litter layer resistance estimated in a similar manner to
!       Sakaguchi and Zeng (2009)
llai_eff = llai * (1.d0 - min(1.d0,fr_snow(2)))
r_litter = 1.d0 - exp(-llai_eff)

!    Potential evaporation from vegetated soil (additional limits follow)
epvg  = rho3*(qvg-qs) / (r_soil + r_litter + 1.d0/(ch_vg*vs) )
```

The atmospheric resistance term under the canopy, r_{aw}, is computed in an manner analogous to atmospheric resistance in the GISS GCM (Hansen et al., 1983) as

$$r_{aw} = \frac{1}{C_s U}$$

where U is wind speed and C_s is the turbulent transfer coefficient under the canopy. This transfer coefficient must take into account a canopy's sheltering effect on turbulent transfer. We use a simple expression that is a function of leaf and stem area index as (Lawrence et al., 2007)

$$C_s = C_{s,bare} e^{-(LAI+SAI)} + C_{s,dense}\left(1 - e^{-(LAI+SAI)}\right)$$

where LAI is leaf area index [m^2/m^2] and SAI is stem area index [m^2/m^2]. We recognize that in reality this transfer coefficient is also dependent on vegetation density and structure. FLUXNET-based measurements at Morgan Monroe State Forest indicate that soil evaporation is overestimated in the non-growing season. We include a litter resistance that is current of the form $1/(kU)$, where k is a constant depending on the amount of litter on the ground.

Energy Balance

TerraE solves explicitly for soil and canopy temperature by calculating change in heat content. Belowground, root mass does not affect the thermodynamic properties of the soil. Aboveground, because of the instabilities inherent in the explicit numerical solution, the heat capacity of the canopy plus canopy air is artificially large, and TerraE may go to shorter time steps. Therefore, currently, canopy heat capacity, while varying with plant biomass, is not directly equal to the heat capacity of the canopy and its air space. The shorter time steps mean that TerraE calls Ent's biophysics more often to calculate canopy conductance, but this does not affect the rate of calls to other Ent modules, such as growth/allocation, phenology, and disturbance, because these are updated at longer time scales than the LSM's primary half-hour time step.

In TerraE, the canopy energy balance is computed as (Rosenzweig and Abramopoulos, 1997)

$$\frac{dH_c}{dt} = R_{n,c} - F_{h,v} - \lambda F_{q,v} + F_{h,P} - F_{h,d}$$

where H_c is the canopy heat, $R_{n,c}$ is the net canopy radiation [W/m^2], λ is the latent heat of vaporization, $F_{h,v}$ is the sensible heat flux from the canopy, $F_{h,v}$ is the moisture flux from the canopy, $F_{h,P}$ is the heat flux transported by precipitation, and $F_{h,d}$ is the heat flux transported by throughfall. The canopy temperature is then calculated by dividing Hc by the specific heat of the canopy

$$T_c = \frac{H_c}{c_{p,can} + w_c c_p}$$

where $c_{p,\,can}$ is the specific heat capacity of the canopy and cp is the specific heat capacity of water. The above equation is modified, if frozen water is on the canopy. The canopy heat capacity is based on an empirical expression expressed as

$$c_{p,can} = \left(0.01 + 0.002LAI + 0.001LAI^2\right)c_p\rho_w$$

The land-surface model computes the average ground temperature for each time step and passes it to the atmospheric surface routine of the GCM, which then returns the corresponding drag coefficient and air surface temperature to the LSM (Rosenzweig and Abramopoulos, 1997).

Snow

Snow cover is needed to calculate active leaf area and canopy albedo. The snow computations are based on the model of Lynch-Stieglitz (1994) and were recently outlined in Aleinov and Schmidt (2006). The model includes three snow layers unless the thickness of total snow goes below a threshold value (15 cm), in which case the snow is modeled as a single layer. Three prognostic variables are used to describe each snow layer: water equivalent W_i [m], heat content H_i [J/m^2], and layer thickness Z_i [m]. The total snow fraction covering a grid cell is computed according to Roesch et al. (2001) as

$$f_{sn} = 1 - e^{-\frac{S}{\langle V_h \rangle_{\text{ent}} \rho_{sn} g}}$$

where $\langle V_h \rangle_{\text{ent}}$ is the average vegetation height [m], S is the water equivalent snow depth [m], and ρ_{sn} is the density of snow [kg/m^3]. The snow masking depth is currently computed as $0.1\langle V_h \rangle_{\text{ent}}$, which affects both the canopy albedo calculation and the calculation of active leaf area.

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Appendix 1

Scheme for Interception of Precipitation by Vegetation Canopies

Canopy Interception Reservoir

The scheme for precipitation interception by a canopy is based on a water-balance equation applied to a canopy reservoir

$$\frac{\partial w_c}{\partial t} = I_{c,net} - f_{c,wet}\left(1 - f_{sn}f_{sn,c}\right)E_{v,w} \tag{2}$$

where w_c is the canopy water storage [m], t is time [s], $I_{c,net}$ is the canopy interception rate not including evaporation losses [m/s], $f_{c,wet}$ is the wet fraction of the canopy [-], f_{sn} is the fraction of the grid cell that is covered by snow [m^2 snow/m^2 ground], $f_{sn,c}$ is the fraction of the canopy that is covered by snow [m^2 snow/m^2 canopy], and $E_{v,w}$ is the evaporation rate from the canopy reservoir (wet canopy) [m^3/m^2 ground/s]. $I_{c,net}$ is expressed as

$$I_{c,net} = P_{tot,tot} - D_{r,tot} \tag{3}$$

where $P_{tot,tot}$ is the total (convection and large-scale) precipitation in the grid cell [m/s] and $D_{r,tot}$ is the total drip (rain and snow) from the canopy [m/s]. The equation is solved using the Euler's method (explicit in time) as

$$w_c^{t+\Delta t} = w_c^t + \left(P_{tot}^t - D_{r,tot}^t - f_{c,wet}^t\left(1 - f_{sn}^t f_{sn,c}^t\right)E_{v,w}^t\right)\Delta t \tag{4}$$

```
w(0,2) = w(0,2) + ( fc(1) - fc(0) )*dts
ht(0,2)=ht(0,2) + ( fch(1) - fch(0) )*dts
fc(0) = -pr+evapvw*fw*(1.d0-fm*fr_snow(2))
          ! snow masking of pr is ignored since
! it is not included into drip
fc(1) = -dripw(2) - drips(2)
```

where Δt is the time-step size [s]. The constitutive equations for the canopy water balance are based on empirical relationships for $D_{r,tot}$, $f_{c,wet}$, f_{sn}, f_m, and $E_{v,w}$..

Interception Reservoir Capacity and Wet Canopy Fraction

The expression for wet fraction of the canopy is written as

$$f_{c,wet} = \left(\frac{w_c}{w_{c,\max}}\right)^{2/3} = \left(\frac{w_c}{0.0001L_t}\right)^{2/3} \tag{5}$$

where $w_{c,max}$ is the maximum storage capacity of the canopy reservoir [m]. This variable is currently computed in the Ent DGTEM as $0.0001L_t$ (or 0.1 L_t for units of kg/m^2).

Fraction of Grid Cell Covered by Snow

The fraction of the grid cell that is cover by snow, f_{sn}, is computed from the three-layer snow model, while the fraction of vegetation covered by snow is given as

$$f_{sn,v} = 1 - \exp\left(\frac{S}{V_h G_s}\right) \tag{6}$$

where S is the water-equivalent snow depth [m] assuming the snow is distributed uniformly in space over the snow cover fraction of the cell, V_h is the height of the vegetation [m], and G_s is the specific gravity of snow [-] assumed to be equal to 0.1.

Precipitation Types

The precipitation currently passed to GHY.f is the total (convection and large-scale) precipitation in the grid cell, $P_{tot,tot}$. The large-scale precipitation, P_{LS}, passed to GHY.f is currently set to zero.

The convective rainfall, $P_{rain,c}$, in GHY.f is computed as

$$P_{rain,c} = P_{tot,tot} - P_{LS,tot} - \left(P_{tot,sn} - P_{LS,sn}\right) \tag{7}$$

```
ptmps=prs-snowfs
ptmps=ptmps-evapvw*fw
ptmp=pr-prs-(snowf-snowfs)
```

where $P_{LS,tot}$ is the large-scale precipitation rate [m/s], $P_{tot,sn}$ is the total (convective and large-scale) snowfall rate [m/s], and $P_{LS,sn}$ is the large-scale snowfall rate [m/s].

Precipitation Throughfall from Canopy

The total throughfall is the sum of drip due to rain and snow. It is assumed that the canopy does not intercept snow, such that the drip due to snow, $D_{r,sn}$, is

$$D_{r,sn} = P_{tot,sn} \tag{8}$$

where P_{sn} is the total (convective and large-scale) snowfall rate using equivalent water depth [m s^{-1}].

The drip from rainfall, $D_{r,rain}$, is then computed based on whether $P_{rain,c}$ is greater than zero and based on the value of the factor P_{fac}. P_{fac} is computed as

$$P_{fac} = \frac{\mu}{P_c}\left[P_{\max} - \left(P_{LS,tot} - P_{LS,sn} - f_{c,wet}E_{v,w}\right)\right] \tag{9}$$

```
pfac=(pmax-ptmps)*prfr/ptmp
```

where μ is the fraction of the grid cell over which rainfall is occurring. $P_{\max}$ is the maximum precipitation rate for which there is no throughfall. $P_{\max}$ is taken to be a fixed precipitation rate times the dry canopy fraction, which is given as

$$P_{\max} = P_m\left(1 - f_{c,wet}\right) \tag{10}$$

```
pm=1d-6
pmax=fd0*pm
```

where P_m is a constant equal to 1×10^{-6} m/s. The expression for $D_{r,rain}$ is then for positive P_c as

$$D_{r,rain} = \begin{cases} P_{tot,tot} - P_{LS,tot} - f_{c,wet} E_{v,w} - P_{\max} & P_{fac} < 0 \\ P_c e^{-P_{fac}} & 0 \le P_{fac} < 30 \\ \max\left[P_{LS,tot} - P_{LS,sn} - f_{c,wet} E_{v,w} - P_{\max}, 0\right] & P_{fac} \ge 30 \end{cases} \qquad (11)$$

```
if(ptmp.gt.0.d0)then
     pfac=(pmax-ptmps)*prfr/ptmp
     if(pfac.ge.0.d0)then
      if(pfac.lt.30.d0) dr=ptmp*exp(-pfac)
     else
      dr=ptmp+ptmps-pmax
     endif
endif
```

and P_c equal to 0 as

$$D_{r,rain} = \max\left[P_{LS,tot} - P_{LS,sn} - f_{c,wet} E_{v,w} - P_{\max}, 0\right] \qquad (12)$$

```
drs=max(ptmps-pmax,zero)
dr=drs
```

The following limits are also imposed sequentially:

$$D_{r,rain} = \min\left[D_{r,rain}, P_{tot,tot} - P_{tot,sn} - f_{c,wet} E_{v,w}\right] \qquad (13)$$

$$D_{r,rain} = \max\left[D_{r,rain}, P_{tot,tot} - P_{tot,sn} - f_{c,wet} E_{v,w} - \frac{w_{c,msx} - w_c}{\Delta t}\right] \qquad (14)$$

$$D_{r,rain} = \max\left[D_{r,rain}, 0\right] \qquad (15)$$

```
dr = min( dr, pr-snowf-evapvw*fw )
dr = max( dr, pr-snowf-evapvw*fw - (ws(0,2)-w(0,2))/dts )
dr = max( dr, 0.d0 )   ! just in case (probably don''t need it)
dripw(2) = dr
```

Revisions to Precipitation Throughfall & Wet Canopy Fraction

The determination of precipitation interception is challenging in a global-scale land surface model, because precipitation is rarely uniform over the model's grid cell (Koster and Suarez, 1996). The amount of precipitation intercepted (precipitation loading) is affected by the fraction of the grid cell, μ, covered by precipitation. The mass of precipitation that needs to reach the land surface is the same, such that the effective precipitation rate increases as μ decreases. As a result, less precipitation is intercepted as μ decreases. Some models (SiB) use an exponential function to describe the subgrid distribution of precipitation, but a steeper exponential function is analogous to smaller μ values and will result in less intercepted water. For ecosystem-scale simulations, the fraction of the grid cell, μ, covered by precipitation is set to 1. That is, the size of a storm is much greater than the ecosystem size.

The precipitation rate for the dry fraction of the canopy is

$$P_{\mathrm{dry}} = \left(1 - f_{c,\mathrm{wet}}\right) P \qquad (16)$$

and the rate for the wet fraction of the canopy, which fall through directly to the ground, is

$$P_{wet} = f_{c,wet} P \tag{17}$$

where P_{dry} and P_{wet} are the precipitation rates occurring on the dry and wet leaves, respectively. The maximum amount of precipitation that can be added in the region with a dry canopy and covered by convective rainfall is

$$w_{c,\text{add}} = \mu \cdot \left(w_{c,\max} - w_c \right) \tag{18}$$

The total intercepted precipitation is updated with

$$I_c^{t+\Delta t} = I_c^t + \min\left(P_{\text{dry}} \Delta t, w_{c,\text{add}} \right) \tag{19}$$

where I_c is the canopy interception [m]. The precipitation throughfall or drip is computed as

$$D_{r,\text{rain}} = P - \frac{I_c^{t+\Delta t} - I_c^t}{\Delta t} \tag{20}$$

The temporal correlation in storm position is a problem that most LSMs do not address, which stems from the fact that a simulated storm can span several timesteps (Koster and Suarez, 1996). On one timestep, a storm will fill a fraction of the interception reservoir. In the next timestep, the model produces precipitation over the same fraction μ (Koster and Suarez, 1996). The amount of intercepted water on the subsequent timestep depends on how and if the model redistributes water from the previous timestep. If a model does redistribute water and has no memory (that is, does not account for temporal correlation in storm position), then it will intercept more water than if the model had memory. The time step size of the GISS LSM ranges from 5 sec to 30 minutes, and, because storms typically last much longer than this, accounting for temporal correlation in storm position is arguably more realistic (Koster and Suarez, 1996).

The fraction of precipitation that falls on leaves previously wetted is

$$f_{previous,wet} = \left(1 - \frac{\Delta t}{\tau_{\text{storm}}} \right) \tag{21}$$

Water added to the dry canopy, covered by convective rainfall, and not previously wetted by a storm is

$$P_{\text{dry}} = \begin{cases} \left(1 - f_{prev\_wet}\right)\left(1 - f_{c,\text{wet}}\right) P & f_{c,\text{wet}} \geq \mu \\ \dfrac{f_{c,\text{wet}}}{\mu}\left(1 - f_{prev\_wet}\right)\left(1 - f_{c,\text{wet}}\right) P & f_{c,\text{wet}} < \mu \end{cases} \tag{22}$$

where τ_{storm} is an arbitrary time scale for storm length that is greater than or equal to Δt (Koster and Suarez, 1996).

Evaporation from the Wet Canopy

The evaporation from the wet canopy, $E_{v,w}$, is compute based on potential evaporation, E_p. E_p is expressed as

$$E_p = \frac{\rho_a}{\rho_w} C_q V_s \left(q_{sat}\left(T_{can}\right) - q_s \right) \tag{23}$$

epv = rho3*ch*(vs*(qv-qs) -v_qprime)

where ρ_a is the density of surface air [kg/m^3], q_s is the humidity of surface air [-], q_{sat} is the saturated humidity at the temperature of the canopy T_{can}, V_s is wind speed, and C_q is the humidity transfer coefficient [-]. The following limits are then used to constrain $E_{v,w}$:

$$E_{v,w} = \min\left[E_p, \frac{w_c}{\Delta t} \right] \tag{24}$$

$$E_{v,w} = \min\left[E_p, \frac{q_{m,1}/\rho_w}{\Delta t} \right] \tag{25}$$

```
ibv = 2; evap_max_wet(ibv) = w(0,2)/dt !+ pr ! pr doesn"t work
for snow
evapvw = min( epv, evap_max_wet(2) )
c**** qm1 has mass of water vapor in first atmosphere layer, kg
m-2
qm1dt=.001d0*qm1/dt
evapvw = max( evapvw,-qm1dt )
```

where $q_{m,1}$ is the mass of water vapor in the first atmospheric layer [kg/m^3].

References

Koster, R.D. and Suarez, M.J., 1996. Energy and water balance calculations in the Mosaic LSM, NASA Goddard Space Flight Center, Greenbelt, Maryland.