Supplement

Description of newly modeled processes

SOLVEG is a one-dimensional multi-layer model that consists of four sub-models for the atmosphere near the surface, soil, vegetation, and radiation within the vegetation canopy. Since full descriptions of the model are available in the papers by Nagai (2004), Katata (2009), Ota, Nagai, and Koarashi (2013), and Katata & Ota (2017), we give details about the processes newly modelled in the present study.

Modeling snow accumulation and melting processes

A multi-layer snow module is newly incorporated into the SOLVEG model. Most of the variables in the following equations are based on either the Community Land Model (CLM: Oleson *et al.*, 2010) or SNTHERM (Jordan 1991), while the model is unique in including the gravitational and capillary liquid water flows in unsaturated snow layer based on van Genuchten's concept of water flow in the unsaturated zone (c.f., Hirashima, Yamaguchi, Sati, & Lehning, 2010).

The temporal change in snow temperature T_{sn} (K) is expressed by the heat conduction equation based on Yamazaki (2001) as

$$C_{sn}\rho_{sn}\frac{\partial T_{sn}}{\partial t} = \frac{\partial}{\partial z}\left(\lambda_{sn}\frac{\partial T_{sn}}{\partial z}\right) - \frac{\partial I_n}{\partial z} - l_f E_{smel} - lE_{sb}, \qquad (1)$$

where C_{sn} and ρ_{sn} are the specific heat of snow (J kg^{-1} K^{-1}) and the density of the bulk snow (kg m^{-3}), respectively, λ_{sn} is the thermal conductivity of snow (Wm^{-1}K^{-1}), I_n is the net solar flux in the snow layer (W m^{-2}), l_f and l are the latent heats of fusion and sublimation (J kg^{-1}) , respectively, and E_{smel} is the melting or freezing rate in the snow layer (kg m^{-3} s^{-1}), and E_{sb} is the sublimation rate of water vapor from the snow layer (kg m^{-3} s^{-1}). I_n is calculated as:

$$(1-r)(1-A_b)S_{down}\,exp(-\mu z), \quad (2)$$

where r is the absorptivity of solar radiation at the snow surface, A_b is the albedo of the snow surface as a sum of the direct and the diffuse visible and near-infrared solar and long-wave radiations (Wiscombe & Warren, 1980), and μ is the extinction coefficient of the solar radiation in the snow layer (Jordan, 1991).

The sublimation rate E_{sb} is calculated only at the snow surface by assuming that water vapor is saturated over the snow as:

$$E_{sb0} = \sigma_{sn}\rho c_{E0}|u|[q_{sat}(T_{sn0}) - q_r], \quad (3)$$

where σ_{sn} is the fractional area of snow cover parameterized using physical snow height (Essery, Morin, Lejeune, & Menard, 2013), ρ is the density of air (kg m^{-3}), c_{E0} is the bulk coefficient, $q_{sat}(T_{sn0})$ is the saturated specific humidity (kg kg^{-1}) at the snow surface temperature T_{sn0} (K), and $|u|$ and q_r are the horizontal wind speed (m s^{-1}) and specific humidity (kg kg^{-1}) at the lowest atmospheric layer, respectively.

Melting or freezing rate in the snow layer is calculated from snow temperature as:

$$E_{smel} = \frac{C_{sn}\rho_{sn}}{l_f}\frac{T_{sn}-T_m}{\partial t}, \quad (4)$$

where T_m is the melting point of 273.15 K. Using E_{smel}, the ice content in snow w_i (kg m^{-2}) at each snow layer is determined as:

$$\frac{\partial w_i}{\partial t} = -E_{smel}\Delta z, \tag{5}$$

where Δz is the snow layer thickness (m).

The mass balance equation for liquid water in the snow layer is given as:

$$\rho_w \frac{\partial \eta_{sw}}{\partial t} = \frac{\partial}{\partial z}\left(D_{sw}\frac{\partial \eta_{sw}}{\partial z} + K_{sw}\right) - E_{smel}, \tag{6}$$

where η_{sw} is the volumetric liquid water content (m^3 m^{-3}), D_{sw} is the liquid water diffusivity (m^2 s^{-1}), K_{sw} is the snow unsaturated hydraulic conductivity (m s^{-1}), and ρ_w is the density of liquid water (kg m^{-3}) in the snow layer. The equations for D_{sw} and K_{sw} are similar to those for soil water content in the capillary region (Katata 2009), except for using the empirical parameters for the snow cover that are given by Hirashima *et al.* (2010).

Snow accumulation and compaction at each snow layer are modelled as:

$$\frac{1}{\Delta z}\frac{\partial \Delta z}{\partial t} = C_{snf} - C_{met} - C_{over} - C_{mel}, \tag{7}$$

$$C_{met} = c_1\, exp[-c_2(T_m - T_s) - c_3\, max(0, \rho_s - \rho_0)], \tag{8}$$

$$C_{over} = \frac{-P_s}{\eta_{sn}}, \tag{9}$$

$$C_{mel} = -\frac{1}{\Delta t} max\left(0, \frac{f_{ice} - f_{ice}^{+}}{f_{ice}}\right), \tag{10}$$

where C_{snf}, C_{met}, C_{over}, and C_{mel} are the change rates in Δz (s^{-1}) due to snowfall, metamorphism, overburden, and melting, respectively, and f_{ice} and f_{ice}^{+} the fractions of ice before and after the melting, respectively. C_{snf} is calculated as $S_f\, \rho_{fs}/\rho_w$, where S_f is the snowfall rate (mm s^{-1}) given by either the input data or the empirical equation using total rainfall rate and wet bulb temperature (Yamazaki 2001), and ρ_{fs} the fresh snow density (kg m^{-3}) obtained by Boone (2002).

Values for the parameters in the above equations are given by Oleson *et al.* (2010).

Snow grain growth (i.e., change in grain size in the snow layer) is calculated based on Jordan (1991) as:

$$\frac{\partial d_{sn}}{\partial t} = \begin{cases} \frac{g_1|U_v|}{d_{sn}} & \eta_{sw} = \eta_{swilt} \\ \frac{g_2}{d_{sn}}(\eta_{sw} + 0.05) & \eta_{swilt} < \eta_{sw} < 0.09, \\ 0.14\frac{g_2}{d_{sn}} & 0.09 < \eta_{sw} \end{cases} \quad (11)$$

where d_{sn} is the snow grain diameter (m), U_v the mass vapor flux in the snow layer (kg m^{-2} s^{-1}), and g_1 and g_2 the parameters. The formulation of U_v and the values of g_1 and g_2 are given by Jordan (1991).

After the above calculations for temperature, liquid and ice water contents, and accumulation and compaction in snow, the number of snow layers is adjusted by either combining or subdividing layers (Jordan, 1991) to obtain the physical snow height.

Modeling freeze-thaw process in soil

In the soil module, freeze–thaw processes in soil are considered based on heat conduction and liquid water flow equations as follows:

$$C_s\rho_s\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z}\left(\lambda_s\frac{\partial T_s}{\partial z}\right) - lE_b - l_f E_{mel}, \quad (12)$$

$$\rho_w\frac{\partial \eta_w}{\partial t} = \frac{\partial}{\partial z}\left(D_w\frac{\partial \eta_w}{\partial z} + K\right) - E_b - E_{mel}, \quad (13)$$

where C_s and ρ_s are the specific heat of soil (J kg^{-1} K^{-1}) and the density of the bulk soil (kg m^{-3}), respectively, λ_s is the thermal conductivity of soil (W m^{-1} K^{-1}), l_f and l are the latent heat of fusion and sublimation (J kg^{-1}) , respectively, η_w is the volumetric soil water content (m^3 m^{-3}), D_w is the soil water diffusivity (m^2 s^{-1}), K is the unsaturated hydraulic conductivity (m s^{-1}), E_b is the evaporation or condensation or sublimation of soil water (kg m^{-2} s^{-1}), and E_{mel} is the melting or freezing rate in soil (kg m^{-3} s^{-1}). The soil water diffusivity D_w (m^2 s^{-1}) is expressed by:

$$D_w = K\frac{\partial\psi}{\partial\eta_w}, \tag{14}$$

where ψ is the water potential in the soil layer (m). ψ and K (m s^{-1}) in frozen soil are modeled based on the concept of freezing point depression (Zhang, Sun, & Xue, 2007):

$$\psi = \psi_{unfrozen}(1 + C_k\eta_i)^2, \tag{15}$$

$$K = K_{unfrozen}10^{-E_i\eta_i}, \tag{16}$$

where C_k and E_i are the empirical parameters, and $\psi_{unfrozen}$ and $K_{unfrozen}$ are the ψ and K in unfrozen soil described by Katata (2009), respectively.

Ice content at each soil layer η_i (m^3 m^{-3}) is determined similar to snow ice content in Eq. (5) as:

$$\frac{\partial\eta_i}{\partial t} = -\frac{E_{mel}}{\rho_i}, \tag{17}$$

$$E_{mel} = \frac{C_s\rho_s}{l_f}\frac{T_s-T_m}{\partial t}, \tag{18}$$

where ρ_i is the density of ice (kg m^{-3}).

Modeling grassland vegetation growth and development

To simulate the winter-related processes for grassland phenology such as leaf development and senescence due to cold stresses, the relevant scheme in the grass growth model named BASic GRAssland model (BASGRA; Höglind, Van Oijen, Cameron, & Persson, 2016) is coupled with the vegetation sub-model of SOLVEG to simulate the vegetation growth. BASGRA consists of the LINGRA grassland model (Van Oijen, Höglind, Hanslin, & Caldwell, 2005) with models for cold hardening and soil physical winter processes. The three main features that characterize plant growth in BASGRA are: (1) simulation of source-sink relations where the source consists both current photosynthesis and remobilization of reserves; (2) simulation of leaf area dynamics and tillering for vegetative and generative tillers; and (3) cold hardening and the effect of physical winter stress factors on tiller survival and plant growth. BASGRA is evaluated by using experimental datasets of harvestable dry matter of perennial rye grass collected in Europe (Schapendonk, Stol, van Kraalingen, & Bouman, 1998) and from five locations in Norway, covering a wide range of agroclimatic regions, day lengths, and soil conditions (Höglind *et al.*, 2016). Since the full description of BASGRA is available in Höglind *et al.* (2016), only processes related to coupling SOLVEG with BASGRA are summarized below.

Instead of the original scheme of photosynthetic processes in BASGRA, diurnal CO_2 assimilation is calculated as the accumulation of the net assimilation for each time step within the vegetation sub-model (Nagai, 2004). The net CO_2 assimilation rate A_n, which is calculated by subtracting the leaf respiration rate R_d (μmol m^{-2} s^{-1}) from the assimilation rate, is expressed as

$$A_n = \min(f_{cold} w_c,\ w_e,\ f_{cold} w_s) - R_d, \tag{19}$$

where the CO_2 assimilation rate is determined as the minimum of three limiting rates, that is the limitation by efficiency of the photosynthetic enzyme system (Rubisco) w_c (μmol m^{-2} s^{-1}), limitation by the absorbed photosynthetically active radiation (PAR) w_e (μmol m^{-2} s^{-1}), and limitation by the capacity of leaf to export the products of photosynthesis w_s (μmol m^{-2} s^{-1}). When snow covers grasses, no photosynthesis is assumed due to low light availability and only soil respiration is considered. Furthermore, the reduction of photosynthesis under a low air temperature is tested as the empirical factor for cold stress of grasslands (f_{cold}) as:

$$f_{cold} = \min[1, \max\{0, (T_a + 4)/(T_{ph} + 4)\}], \quad (20)$$

where f_{cold} is the empirical factor for cold stress of grasslands, T_a (°C) is the daily and vertical mean air temperature for all canopy layers, and T_{ph} (°C) is the threshold air temperature above which grasslands are photosynthetically active. The determination of the value of this threshold temperature is important to avoid the overestimation (mainly from fall to winter) of photosynthesis at a low temperature (Höglind *et al.*, 2011). In the original BASGRA, T_{ph} is set to 1 °C; that is, V_{cmax} starts decreasing linearly when T_a drops below 1 °C until it becomes zero at −4 °C. However, in the SOLVEG simulation, since the values of T_{ph} may change depending on environmental conditions, this value is calibrated for each site so that the model reproduces the observed CO_2 flux during the extremely warm winter period.

Carbon from photosynthesis and remobilized reserves is allocated between sinks according to the method of BASGRA based on changing sink priorities and changing sink strengths. Sink strengths are determined by the dynamics of leaves and stems and the acclimation to a low temperature. The major occasional disturbance during the growing season is the removal of tillers and leaves by cutting, with subsequent regrowth of the sward. The regrowth rate after cutting depends on the phenological stage at which cutting takes place and on the strengths of the sources and sinks.

BASGRA uses a so-called "big-leaf" approach (Monteith, 1981), thus predicting total LAI of the whole grass vegetation canopy. Since SOLVEG uses a multi-layer structure of canopies, the profile of leaf area density is obtained by simply dividing total LAI by canopy height (h) by assuming vertically uniformity for all canopy layers. Canopy height, which is not simulated in BASGRA, is calculated by the following function of LAI (c.f., Ammann, Flechard, Leifeld, Neftel, & Fuhrer, 2007; Thornley & France, 2007) with fitting parameters:

$$h = 0.025LAI^{1.2}. \quad (21)$$

The natural turnover of leaves and roots is modeled using typical life spans in years (Arora & Boer, 2005), while BASGRA does not simulate the senescence of elongating tillers or roots. The fraction of roots in soil layers and rooting depth are modeled as a function of root biomass (Arora & Boer, 2003). Daily amounts of the dead root biomass (root litter) are used as inputs to SOC in the soil sub-model of SOLVEG.

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