

Simple soil-plant water balance model

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1 Model overview

1.1 Design principles

The water balance model calculates temporal variations in soil moisture on a daily step basis for the input forest stand and for the period corresponding to input weather data.

The model considers only the vertical spatial dimension of the stand and not the horizontal distribution of plants within it. In other words, the model is not spatially explicit (i.e., plants do not interact for space explicitly). Still, the stand is divided into groups of plants, here referred to as ‘plant cohorts’ of different species, height and leaf area index (*LAI*). Height and *LAI* values determine competition for light. *LAI* values also drive competition for water. The soil water balance follows the design principles of SIERRA (Mouillot et al., 2001, Ruffault et al., 2014, 2013) and BILJOU (Granier et al., 2007, 1999), although some features are taken from other models (Kergoat 1998). Potential evapotranspiration (*PET*) is given as input and the model determines maximum canopy transpiration (Tr_{\max}) using an empirical relationship between the *LAI* of the stand and the ratio Tr_{\max}/PET (Granier et al. 1999). Actual plant transpiration is calculated using a simple function depending on current soil moisture level and degree of shading.

When lateral water transport is also considered the water inputs of the stand includes not only precipitation but also runoff water from cells of its upper microwatershed, as done in SIERRA (Mouillot et al., 2001).

1.2 State variables

The model has the following state variables:

- Cumulative growth degree days, *GDD*, are tracked by the model and determine leaf phenological status of winter deciduous species.
- Daily soil moisture content dynamics on each layer *s* are tracked using $W = \theta(\Psi_s)/\theta_{fc,s}$, the **proportion of volumetric soil moisture in relation to field capacity**, where field capacity, $\theta_{fc,s}$, is assumed to

correspond to $\Psi_{fc} = -0.033$ MPa. Soils are not allowed to contain more water than dictated by their field capacity.

- If snow pack is considered, the model also tracks S_{snow} , the snow water equivalent (in mm) of the snow pack storage over the surface.
- If irreversible cavitation is activated, the model also tracks $P_{embolized,i}$, the maximum value of daily drought stress so far experienced.

1.3 Water balance

Daily variations in soil water content (mm) can be summarized as:

$$\Delta SWC = Pr + Sm - In - Ru - Dd - Es - Tr \quad (1)$$

where Pr is precipitation as rainfall, Sm is water from melting the snow pack (if considered), In is water evaporated after being intercepted by the canopy, Ru is surface runoff, Dd is deep drainage, Es is evaporation from soil and Tr is plant transpiration. If snow pack is considered, its balance is the following:

$$\Delta S_{snow} = Ps - Sm \quad (2)$$

where Ps is precipitation as snowfall and Sm is snow melt. When landscape hydrological processes are considered, daily variations in soil water content is then summarized as:

$$\Delta SWC = Pr + Sm + Ro - In - Ru - Dd - Es - Tr \quad (3)$$

where Ro is surface runoff water entering the cell from neighbouring cells at higher elevation.

1.4 Process scheduling

Every day water balance is performed as follows. The model first updates leaf area values according to the phenology of species and calculates light extinction. After that, the model updates soil water content of soil layers in two steps:

1. If snow dynamics are included, increase snow pack from snow precipitation (Ps) and decreases it from snow melting (Sm);
2. Increase soil moisture due to rainfall (Pr), runoff (Ro) and (if considered) snow melting (Sm), after accounting for canopy interception loss (In), surface runoff (Ru) and deep drainage (Dd);
3. Decrease water content due to bare soil evaporation (Es), and plant transpiration (Tr).

4. Determines drought stress index for each plant cohort.

When landscape hydrological processes are not considered, daily water balance simulations can be done for cells independently (i.e., the whole simulation period can be done one cell before going to the next one). When lateral water transport is also considered, however, water balance of a given day is conducted for all cells before starting the next day. Moreover, a discharge parameter table is prepared at the beginning of the simulation, cells are processed in an order determined by elevation (i.e. cells at higher elevation are processed before cells at lower elevation) and the water balance of a given target cell is influenced by surface runoff (Ro) coming from those neighboring cells that are at higher elevation.

2 Model inputs and outputs

2.1 Soil description

Soil can be described using between 1 and 5 soil layers. For each soil layer s the following parameters are needed for each soil layer (see function `defaultSoilParams()`):

- Z_s [`widths`]: Soil layer width (d_s , in mm).
- $P_{clay,s}$ [`clay`]: Percent of clay.
- $P_{sand,s}$ [`sand`]: Percent of sand.
- OM [`om`]: Percentage of organic matter per dry weight (can be missing).
- BD_s [`bd`]: Bulk density (g/cm³) of each soil layer.
- $P_{rocks,s}$ [`rfc`]: Percentage of rock fragment content ($P_{rocks,s}$).

Soil texture (i.e. percent of sand, silt and clay), bulk density and rock fragment content can differ between soil layers. Specifying a deep rocky layer is important because Mediterranean plants may extend their roots into cracks existing in the parent rock (Ruffault et al., 2013). For those users lacking soil information, package `medfate` provides a function to get the soil description from SoilGrids.org (see help for `soilgridsParams()`).

2.1.1 Soil initialization

Soil initialization is needed to estimate soil hydrological parameters from the physical description. requires a list of soil parameters that is given as input to function `soil()`:

- $P_{macro,s}$ [**macro**]: Percentage of macroporosity corresponding to each soil layers. Macroporosity values are calculated for each soil layer using the equations given in Stolf et al. (2011).
- γ_{soil} [**Gsoil**]: The maximum daily evaporation from soil($\text{mm}\cdot\text{day}^{-1}$).
- γ_{soil} [**Gsoil**]: The maximum daily evaporation from bare soil($\text{mm}\cdot\text{day}^{-1}$).
- κ_{soil} [**Ksoil**]: Extinction parameter to regulate the amount of water extracted from each soil layer during bare soil evaporation.
- $n_s, \alpha_s, \theta_{sat,s}, \theta_{res,s}$ [**VG_n, VG_alpha, VG_theta_sat, VG_theta_res**]: Parameter of the Van Genuchten-Mualem equations.

Parameters of the van Genuchten-Mualem model are estimated from the physical description of the soil using two kinds of pedotransfer functions (see help for `soil()`):

1. Using the USDA texture classification and the average texture class parameters given by Carsel and Parrish (1988).
2. Directly from the soil texture, organic matter and bulk density, using the pedotransfer functions in Toth et al. (2015).

2.1.2 Water retention curves

Water retention curve of a soil layer s is the relationship between the water content, θ_s (in $\text{m}^3\cdot\text{m}^{-3}$ of soil), and the soil water potential, Ψ_s (in MPa). This curve is characteristic for different types of soil, and is also called the soil moisture characteristic. Two types of water retention curves are offered for the water balance model:

- **Saxton model:** In this model, volumetric soil moisture $\theta(\Psi_s)$ corresponding to a given water potential Ψ (in MPa) is calculated using:

$$\theta(\Psi) = (\Psi/A)^{(1/B)} \quad (4)$$

where A and B depend on the texture and, if available, organic matter in the soil layer. If organic matter is available, A and B are calculated P_{clay} , P_{sand} and OM following Saxton and Rawls (2006). Otherwise, they are calculated from P_{clay} and P_{sand} as indicated in Saxton et al. (1986).

- **Van Genuchten model:** The well known van Genuchten (1980) model is:

$$\theta(\Psi) = \theta_{res} + \frac{\theta_{sat} + \theta_{res}}{[1 + (\alpha \cdot \Psi)^n]^{1-1/n}} \quad (5)$$

where $\theta(\psi)$ is the water retention, θ_{sat} is the saturated water content, θ_{res} is the residual water content, α is related to the inverse of the air entry pressure (and here has to be expressed in MPa^{-1}), and n is a measure of the pore-size distribution. When initializing soil using function `soil()`, the parameters of the Van Genuchten model are estimated from texture, using the USDA texture classification and the parameters in Carsel and Parrish (1988), but they can be modified manually afterwards.

Soil water holding capacity (V_s , in mm) in soil layer s is defined as the volumetric water content at field capacity:

$$V_s = d_s \cdot ((100 - P_{rocks,s})/100) \cdot \theta_{fc,s} \quad (6)$$

where d_s is the depth of the soil layer (in mm) and $P_{rocks,s}$ is the percentage of rock fragments. Thus, soil water holding capacity depends on the water retention curve used to determine $\theta_{fc} = \theta(-0.033)$.

2.2 Vegetation description

Vegetation in the stand is described using a set of plant cohorts, in an object of class `spwbInput`. Each plant cohort i is primarily defined by its species identity (SP_i ; with R name `[SP]`).

2.2.1 Aboveground parameters

The aboveground structure of each cohort is defined using the following attributes:

- H_i [`H`]: Total tree or shrub height (in cm).
- CR_i [`CR`]: Crown ratio (i.e. the ratio between crown length and total height).
- LAI_i^{live} [`LAI_live`]: (Maximum) leaf area index (one-side leaf area of plants in the cohort per surface area of the stand).
- LAI_i^{dead} [`LAI_dead`]: Dead leaf area index (one-side dead leaf area of plants in the cohort per surface area of the stand).
- LAI_i^ϕ [`LAI_expanded`]: Current expanded leaf area index (one-side expanded leaf area of plants in the cohort per surface area of the stand).

All vegetation characteristics stay constant during water balance simulations, although the actual expanded leaf area and dead leaf area may vary if the species is winter deciduous.

2.2.2 Belowground parameters

The root system of each plant cohorts is described using the proportion of fine roots in each soil layer:

- $v_{i,s}$ [V]: The proportion of fine roots in each soil layer s .

The rooting system of each cohort i (i.e. the proportions $v_{i,s}$) can be defined assuming conic distribution of fine roots (see `root.conicDistribution()`). In this case, only rooting depth parameter is needed to determine fine root proportions. Alternatively, one can adopt the linear dose response model (Collins and Bras, 2007; Schenk and Jackson, 2002):

$$Y_i(z) = \frac{1}{1 + (z/Z_{50,i})^{c_i}} \quad (7)$$

where $Y_i(z)$ is the cumulative fraction of fine root mass located between surface and depth z ; $Z_{50,i}$ is the depth above which 50% of the root mass is located; and c_i is a shape parameter related to $Z_{50,i}$ and $Z_{95,i}$ as $c_i = 2.94/\ln(Z_{50,i}/Z_{95,i})$ (see `root.ldrDistribution()`).

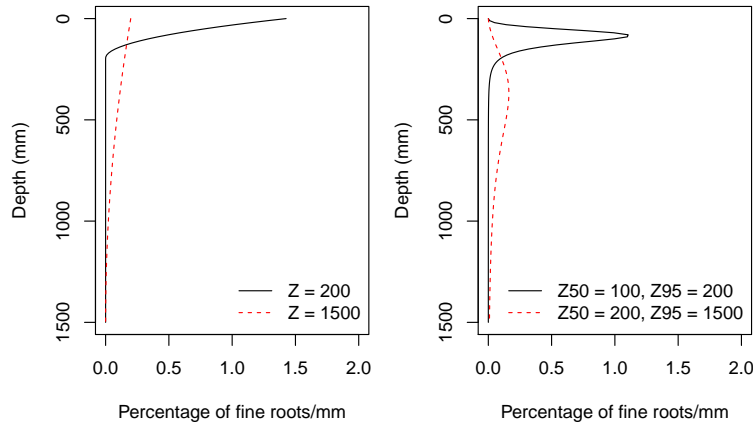


Fig. 1: Examples of root density profile according to a conic distribution (left) and the linear dose response model (right).

2.3 Vegetation functional parameters

Vegetation functional attributes are normally filled for each cohort by function `spwbInput()` from species identity. However, different parameters can be specified for different cohort of the same species if desired. Basic functional parameters relate to light extinction, water interception and leaf phenology:

- $k_{PAR,i}$ [k]: PAR extinction coefficient.

- $s_{water,i}$ [g]: Crown water storage capacity (i.e. depth of water that can be retained by leaves and branches) per LAI unit (in $\text{mm H}_2\text{O} \cdot \text{LAI}^{-1}$).
- $S_{GDD,i}$ [Sgdd]: Growth degree days corresponding to leave budburst (in degrees Celsius).

A second set of parameters are related to plant transpiration and photosynthesis.

- $\Psi_{extract,i}$ [Psi_extract]: Soil water potential (in MPa) corresponding to 50% of water extractive capacity.
- $WUE_{max,i}$ [WUE]: Maximum water use efficiency (in $\text{g C} \cdot \text{mm H}_2\text{O}^{-1}$).
- $K_{rootdisc,i}$ [pRootDisc]: Relative conductance of roots that leads to hydraulic disconnection from soil.

2.4 Metereological input

Weather input data must include variables calculated at the **daily** scale.

- T_{mean} [MeanTemperature]: Mean daily temperature (in $^{\circ}\text{C}$).
- Rad [Radiation]: Solar radiation, after accounting for clouds (in $\text{MJ} \cdot \text{m}^{-2}$), only necessary if snow pack dynamics are simulated (i.e., if `snowpack = TRUE`).
- P [Precipitation]: Daily precipitation (in $\text{L} \cdot \text{m}^{-2} = \text{mm of water}$).
- PET [PET]: Daily potential evapotranspiration (in $\text{L} \cdot \text{m}^{-2} = \text{mm of water}$), preferably calculated using Penman's equation.

Optionally, the user may specify values of wind speed (u , in $\text{m} \cdot \text{s}^{-1}$, `WindSpeed`) that are used to control leaf fall in autumn.

2.5 Control parameters

Control parameters are a list of parameter values, initialized using function `defaultControl()`, that the user can modify to change the general behavior of the model. The control values relevant for the simple water balance model are:

- `verbose` [=TRUE]: Whether extra console output is desired during simulations.
- `soilFunctions` [= "SX"]: Water retention curve model, either "SX" (for Saxton) or "VG" (for Van Genuchten).
- `snowpack` [= TRUE]: Whether dynamics of snow pack are included.

- `transpirationMode` [= "Simple"]: Transpiration model, in this case "Simple".
- `defaultWindSpeed` [= 5]: Default value for wind speed (in $\text{m}\cdot\text{s}^{-1}$) when this is missing (only used for leaf fall).
- `cavitationRefill` [= TRUE]: If FALSE the model operates in a irreversible cavitation mode.

3 Details of processes

3.1 Leaf phenology and leaf fall

Given a base temperature ($T_{base} = 5^\circ\text{C}$), the growth degree days (GDD) are zero for all those days where mean temperature T_{mean} is below T_{base} and start increasing when temperatures become warmer than this threshold. In other words, the GDD function accumulates $\max(0.0, T_{mean} - T_{base})$ for all days previous to the current one. At the end of a year the cumulative value is set again to zero. Plant species can have either evergreen or winter deciduous phenology. Evergreen plants maintain constant leaf area over the year, whereas in deciduous plants leaf-phenological status is updated daily, represented by ϕ_i , the fraction of maximum leaf area. Leaf area index (LAI) values of deciduous plants are adjusted for leaf phenology following (Prentice et al., 1993; Sitch et al., 2003):

$$LAI_i^\phi = LAI_i^{live} \cdot \phi_i \quad (8)$$

Budburst occurs when daily temperature exceeds T_{base} and ϕ_i increases linearly from 0 to 1 as function of the degree days above T_{base} , until a the value $S_{GDD,i}$ is reached (i.e. until $GDD > S_{GDD,i}$). In autumn, ϕ_i drops to 0 when average daily temperature falls again below T_{base} (Sitch et al., 2003). The drop of ϕ_i causes live expanded leaves to become dead leaves. To avoid a sudden decrease of leaf area, dead leaves are kept in the canopy and they are reduced daily using a negative exponential function of wind speed:

$$LAI_i^{dead} = LAI_i^{dead} \cdot e^{-u/10} \quad (9)$$

where u is wind speed in ($\text{m}\cdot\text{s}^{-1}$).

After updating the leaf area of each cohort, the model updates the total leaf area of the stand. To simplify the notation, let us call LAI_i^{all} the sum of dead and live expanded leaves of a cohort i :

$$LAI_i^{all} = LAI_i^\phi + LAI_i^{dead} \quad (10)$$

If there are c plant cohorts, the leaf area index of the whole stand, LAI_{stand} is then:

$$LAI_{stand} = \sum_{i=1}^c LAI_i^{all} = \sum_{i=1}^c LAI_i^\phi + LAI_i^{dead} \quad (11)$$

Although simulations will normally start in winter and GDD will be zero at the beginning, the user can specify a starting non-zero value for GDD in the object created by function `spwbInput()`.

3.2 Light extinction

The proportion of photosynthetic active radiation (PAR) decreases with leaf area following the Beer-Lambert's light extinction equation. To calculate the proportion of PAR available for a given plant cohort one must accumulate the light extinction caused by cohorts whose crown is above that of the target cohort:

$$L_i^{PAR} = e^{-\sum_{h=1}^c k_{PAR,h} \cdot LA_i^{all} \cdot p_{ih}} \quad (12)$$

where $k_{PAR,h}$ is the PAR extinction coefficient of cohort h . Because plant cohorts may differ in height only slightly, leaf area is multiplied by p_{ih} , the proportion of the crown of cohort h that overtops that of cohort i :

$$p_{ih} = \max(0, \min(1, (H_h - H_i \cdot (1 - CR_i)) / (H_h - H_h \cdot (1 - CR_h)))) \quad (13)$$

where CR_i and CR_h are the crown ratio of cohorts i and h . In other terms, cohorts whose crown is completely above that of i reduce the amount of light available more strongly than cohorts that are only slightly taller. L_{ground}^{PAR} , the proportion of PAR that reaches the ground, is calculated as:

$$L_{ground}^{PAR} = e^{-\sum_{i=1}^c k_{PAR,i} \cdot LA_i^{all}} \quad (14)$$

The shortwave radiation (SWR; 400-3000 nm) energy absorbed by each plant cohort needs to be calculated to determine plant transpiration, and the radiation absorbed by the soil is needed to calculate soil evaporation. Foliage absorbs a higher proportion of PAR than SWR; thus, the extinction coefficient is higher for PAR than for SWR. However, values for the ratio of extinction coefficients are rather constant. Following Friend et al. (1997) here it is assumed that the extinction coefficient for PAR is 1.35 times larger than that for SWR (i.e. $k_{SWR,i} = k_{PAR,i} / 1.35$).

To calculate radiation absorption, where the vertical dimension of the plot is divided into 1 m deep layers, and the SWR absorbed is calculated for each plant cohort in each layer. Let n be the number of canopy layers. The fraction of radiation incident on layer j that is absorbed in the same layer is:

$$f_j = 1 - e^{-\sum_{i=1}^c k_{SWR,i} \cdot LA_{i,j}^{all}} \quad (15)$$

where $LA_{i,j}^{all} = LA_{i,j}^{\phi} + LA_{i,j}^{dead}$ is the leaf area index of cohort i in layer j . Hence, the fraction transmitted is $(1 - f_j)$. The fraction of radiation incident on layer j that is absorbed by expanded leaves of plant cohort i in

that layer (f_{ij}) is calculated from the relative contribution of these leaves to the total absorption in the layer:

$$f_{ij} = f_j \cdot \frac{k_{SWR,i} \cdot LAI_{i,j}^\phi}{\sum_{h=1}^c k_{SWR,h} \cdot LAI_{h,j}^{all}} \quad (16)$$

The fraction of canopy radiation absorbed by a plant cohort i across all layers is found by adding the fraction absorbed in each layer:

$$f_i = \sum_{j=1}^n f_{ij} \cdot \prod_{h>j}^n (1 - f_h) \quad (17)$$

where for each layer the fraction of the radiation incident in the canopy that reaches the layer is found by multiplying the transmitted fractions across the layers above it. The proportion of (shortwave) net radiation absorbed by the ground is simply:

$$L_{ground}^{SWR} = 1 - \sum_j^n f_j \quad (18)$$

3.3 Snow pack dynamics

Precipitation is considered be snow when $T_{mean} < 0$. Interception of snow by the canopy is neglected, and all snow is assumed to accumulate in a single storage compartment S_{snow} (i.e. there is no canopy storage). A very simple snow submodel is used for snow pack dynamics (accumulation and melt), taken from Kergoat (1998). When mean air temperature is above 0 Celsius $T_{mean} > 0$, a simple energy budget relates snow melt, Sm (mm), to air temperature and soil-level radiation:

$$Sm = \frac{Rad \cdot L_{ground}^{SWR} \cdot (1 - \alpha_{ice}) + \tau_{day} \cdot T_{mean} \cdot \rho \cdot C_p / r_s}{\lambda_{ice}} \quad (19)$$

where Rad is solar radiation ($\text{MJ} \cdot \text{m}^{-2}$), L_{ground}^{SWR} is the fraction of (short-wave) radiation reaching the ground, $\alpha_{ice} = 0.9$ is the albedo of snow, $\tau_{day} = 86400$ is the day duration in s, ρ is the air density (depending on temperature and elevation), $C_p = 1013.86 \cdot 10^{-6} \text{ MJ} \cdot \text{kg}^{-1} \cdot \text{C}^{-1}$ is the specific heat capacity of the air and $r_s = 100 \text{ s} \cdot \text{m}^{-1}$ is the snow aerodynamic resistance and $\lambda_{ice} = 0.33355 \text{ MJ} \cdot \text{kg}^{-1}$ is the (latent) heat of fusion of snow.

3.4 Rainfall interception loss

When snow dynamics are considered, precipitation occurs in form of rainfall if $T_{mean} > 0$. Otherwise, all precipitation is considered rainfall. Rainfall interception loss, In , is modelled following the Gash et al. (1995) analytical

interception model for sparse canopies, where rain is assumed to fall in a single event during the day. First, the amount of rainfall needed to saturate the canopy is calculated:

$$P_G = -\frac{S/C}{(E/R)} \cdot \ln(1 - (E/R)) \quad (20)$$

where S is the canopy water storage capacity (in mm) – i.e. the minimum amount of water needed to saturate the canopy –, C is the canopy cover and (E/R) is the ratio of evaporation rate to rainfall rate during the rainfall event. Simplifying assumptions are made to determine (E/R) . In De Cáceres et al. (2015) a value of 0.2 is used for all days between December and June, and a value of 0.05 is used for the remaining months (Miralles et al. 2010).

The amount of water evaporated from interception, I (mm), is calculated as:

$$In = C \cdot P_G + C \cdot (E/R) \cdot (P - P_G) \text{ if } P > P_G \quad (21)$$

$$In = C \cdot P \text{ if } P \leq P_G \quad (22)$$

where P is the daily gross precipitation (in mm). Net rainfall, Pr_{net} , is calculated as the difference between gross rainfall and interception loss. Although interception models are normally applied to single-canopy stands, we apply the sparse Gash model to the whole stand (including shrubs). Moreover, in our implementation stem interception is lumped with canopy interception, so that S represents both. Following Watanabe & Mizutani (1996) we estimate S , the canopy water storage capacity, from adjusted LAI values:

$$S = \sum_i s_{water,i} \cdot LAI_i^\phi \quad (23)$$

where $s_{water,i}$ is the depth of water that can be retained by leaves and trunks of a species i per unit of leaf area index ($\text{mm} \cdot \text{LAI}^{-1}$). To estimate the stand cover, C , we use the complement of the percentage of PAR that reaches the ground, i.e. $C = 1 - L_{ground}^{PAR}$ (Deguchi et al., 2006). Fig. 1 below shows examples of relative throughfall, calculated according to the interception model, under different situations (see function `spwb.RainInterception`).

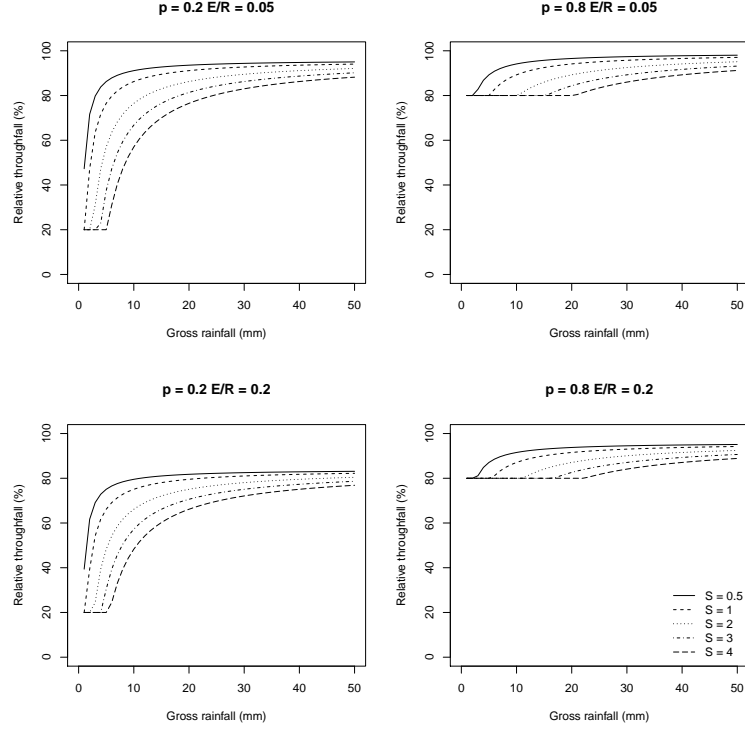


Fig. 1: Examples of canopy interception with different S (canopy water storage capacity), E/R (ratio between evaporation and rainfall rates) and p (throughfall coefficient; $p = 1 - C$).

3.5 Runoff, infiltration and percolation

The water that reaches the soil is the sum of net rainfall (Pr_{net}), runoff (Ro) and melted snow (Sm). The amount of water infiltrating into the soil is $Pr_{net} + Sm + Ro - Ru$, where Ru is the water lost by runoff (see function `spwb.SoilInfiltration`). Runoff (in mm), is calculated using the USDA SCS curve number method, as in Boughton (1989):

$$Ru = \frac{(Pr_{net} + Ro + Sm - 0.2 \cdot V_{soil})^2}{(Pr_{net} + Ro + Sm - 0.8 \cdot V_{soil})} \quad (24)$$

where V_{soil} (in mm) is the overall soil water retention capacity (i.e. the sum of V_s values for topsoil and subsoil).

Following Granier (1999), part of the water reaching one soil layer percolates quickly through the macropores. The remaining water is retained by the micropores refilling the current soil layer. When this soil layer reaches its field capacity the excess of water percolates to the soil layer below. The water percolating from the lowest layer is considered deep drainage, Dd .

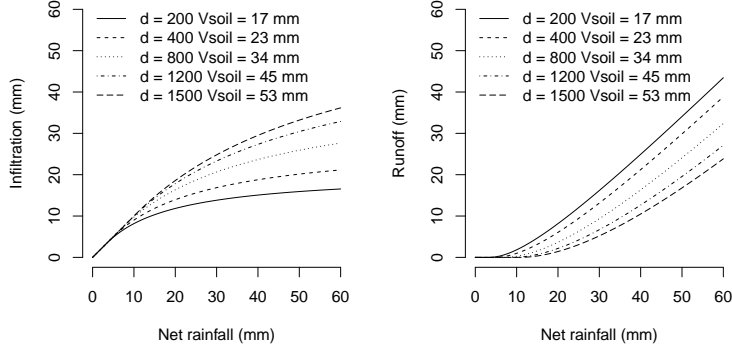


Fig. 2: Examples of infiltration/runoff calculation for different values of net rainfall and overall retention capacity (see function `spwb.SoilInfiltration`), V_{soil} , calculated from different soil depths (topsoil+subsoil) and assuming that soil texture is 15% clay and 25% sand. Rock fragment content was 25% and 40% for the topsoil and subsoil, respectively.

3.6 Plant transpiration and photosynthesis

3.6.1 PET and maximum canopy transpiration

In this model daily potential evapotranspiration (PET ; in $\text{mm}\cdot\text{day}^{-1}$; the amount of evaporation that would occur if a sufficient water source was available) is part of the meteorological input and has to be calculated externally. PET is assumed to represent open water evaporation potential (like in Penman's formula). Maximum canopy transpiration Tr_{\max} not only depends on PET but also on the amount of transpiring surface. To estimate Tr_{\max} we take the approach of Granier et al. (1999), where Tr_{\max}/PET is a function of LAI_{stand} – the cumulative leaf area of the forest stand –, according to the empirical equation:

$$\frac{Tr_{\max}}{PET} = -0.006 LAI_{stand}^2 + 0.134 LAI_{stand} + 0.036 \quad (25)$$

This equation has already been adopted for Mediterranean biomes (Fyllas and Troumbis, 2009; Ruffault et al., 2013).

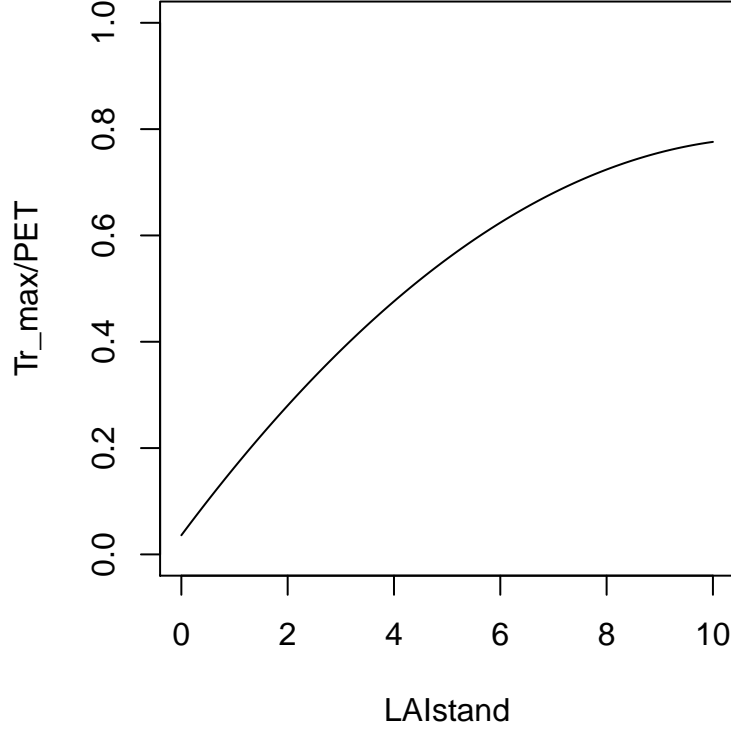


Fig. 3: Experimental relationship between Tr_{\max}/PET and LAI_{stand} .

The maximum transpiration for a given plant cohort i is calculated as the portion of Tr_{\max} defined by the fraction of total absorbed SWR that is due to cohort i :

$$Tr_{\max,i} = Tr_{\max} \cdot \frac{f_i}{\sum_j f_j} \quad (26)$$

3.6.2 Actual plant transpiration

Actual plant transpiration depends on soil moisture and is calculated for each plant cohort and each soil layer separately. $Tr_{i,s}$ (in $\text{mm} \cdot \text{day}^{-1}$) represents the transpiration made by cohort i from layer s . Actual plant transpiration from a given layer is regulated by soil moisture and the resistance to water flow through the plant. For each plant cohort i and soil layer s , the model first estimates the a whole-plant relative water conductance, $K_{i,s}$, which varies between 0 and 1 depending on $\Psi_{extract,i}$, the potential at which conductance is 50% of maximum, and Ψ_s , the water potential in layer s (see function `hydraulics.psi2K()`):

$$K_{i,s} = K_i(\Psi_s) = \exp \left\{ \ln(0.5) \cdot \left[\frac{\Psi_s}{\Psi_{extract,i}} \right]^r \right\} \quad (27)$$

where r is an exponent that modulates the steepness of the decrease in relative conductance when soil potential becomes negative (by default, $r = 3$) and $\ln(0.5)$ is used to ensure that $K_i(\Psi_{extract,i}) = 0.5$ (Fig. 4).

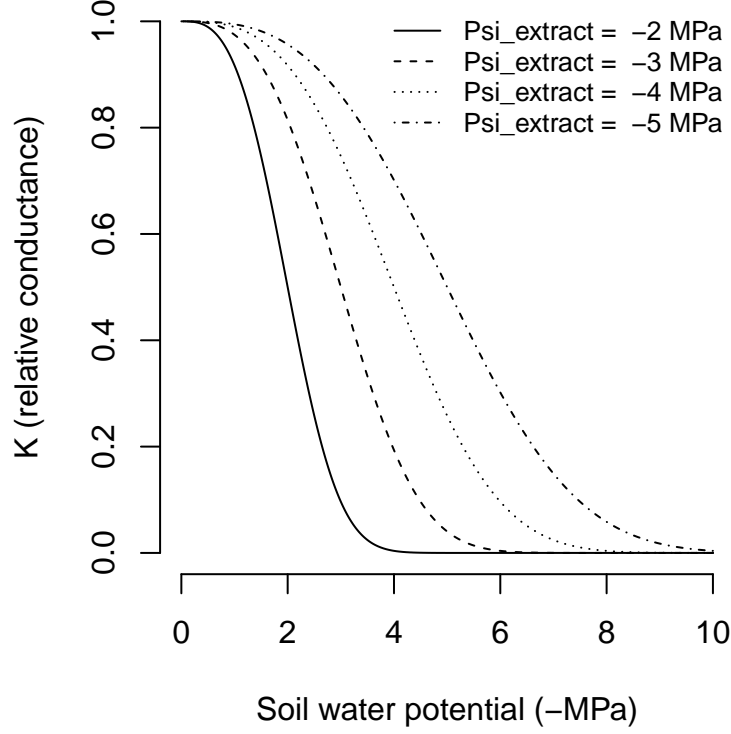


Fig. 4: Whole-plant relative water conductance functions for different $\Psi_{extract,i}$ values ($r = 3$ in all cases).

Actual transpiration of plant cohort i from a given soil layer s , $Tr_{i,s}$, is defined as the product of (Mouillot et al., 2001): (i) the maximum transpiration of the plant cohort; (ii) the relative whole-plant conductance, $K_{i,s}$, corresponding to the species and water potential in layer s ; (iii) the proportion of plant fine roots in layer s , $v_{i,s}$:

$$Tr_{i,s} = Tr_{\max,i} \cdot K_{i,s} \cdot v_{i,s} \quad (28)$$

The total amount of water transpired by plants, Tr (in $\text{mm} \cdot \text{day}^{-1}$), is the sum of $Tr_{i,s}$ values over all plant cohorts and soil layers:

$$Tr = \sum_s \sum_i Tr_{i,s} \quad (29)$$

Assuming no water limitations (i.e. $K_{i,s} = 1$), we have that $Tr = Tr_{\max}$. Total stand transpiration will be lower than Tr_{\max} if soil water potential in

any layer is negative enough to cause a significant reduction in whole-plant conductance. At the plant level, the transpiration of a given plant cohort will be lower than that of others if: (1) the cohort is under the shade (it reduces f_i and hence $Tr_{\max,i}$); (2) the cohort has a lower amount of leaf area (it reduces f_i and hence $Tr_{\max,i}$); (3) the soil layers exploited by the cohort have more negative water potentials (it reduces $K_{i,s}$).

3.6.3 Plant photosynthesis

Because it is useful for growth and forest dynamics, and for compatibility with the 'Complex' transpiration mode, the 'Simple' transpiration mode also calculates net assimilated carbon. Assuming a constant water use efficiency (WUE), photosynthesis for a given plant cohort i (in $\text{g C} \cdot \text{m}^{-2} \cdot \text{day}^{-1}$) is estimated as (Mouillot et al. 2001):

$$A_n = \alpha \cdot WUE_{\max} \cdot Tr_i \quad (30)$$

where Tr_i is the transpiration of plant cohort i , WUE_{\max} is the maximum water use efficiency of the corresponding species (in $\text{g C} \cdot \text{mm}^{-1}$) and $\alpha = T_{\text{mean}}/20$ is bounded between 0 and 1.

3.6.4 Plant drought stress and plant water potential

Similarly to Mouillot et al. (2002), daily drought stress of a given plant cohort i , DDS_i , is defined as the complement of relative whole-plant conductance and is aggregated across soil layers using the proportion of fine roots in each layer as weights:

$$DDS_i = \phi_i \cdot \sum_s (1 - K_{i,s}) \cdot v_{i,s} \quad (31)$$

Leaf-phenological status is included to prevent winter deciduous plants from suffering drought stress during winter. Daily drought stress values can be later used to define drought stress indices for larger temporal scales, as presented in the main text.

The simple transpiration model does not allow estimating a water potential drop from soil to the leaf. Moreover, in a multilayered soil it is difficult to know what would be the water potential of the plant. Despite these limitations, a gross surrogate of 'leaf water potential' ($\Psi_{\text{leaf},i}$; in MPa) may be obtained averaging whole-plant relative conductance values:

$$\Psi_{\text{leaf},i} = K^{-1}(K_i) = K^{-1} \left(\sum_s K_{i,s} \cdot v_{i,s} \right) \quad (32)$$

where K_i is the average whole-plant relative conductance obtained from the scalar product of conductances and fine root proportions and K^{-1} is the inverse of the whole-plant conductance function (see function `hydraulics.K2Psi()`).

3.6.5 Irreversible cavitation and hydraulic disconnection

The water balance model is normally run assuming that although soil drought may reduce transpiration, embolized xylem conduits are automatically re-filled when soil moisture recovers (in other words, cavitation is reversible). It is possible to simulate irreversible cavitation by setting `cavitationRefill = FALSE` in the control parameters. This option causes the model to keep track of the maximum value of drought stress so far experienced:

$$P_{embolized,i} = \max\{P_{embolized,i}, DDS_i\} \quad (33)$$

and then $K_{i,s}$ cannot be larger than the complement of this maximum drought stress:

$$K_{i,s} = \min\{K_i(\Psi_s), 1.0 - P_{embolized,i}\} \quad (34)$$

Another optional behavior consists in allowing the plant to disconnect from the soil when its potential becomes too negative. This may be advantageous for a cavitation-sensitive plant that is competing for water with another plant with higher extraction capacity. Parameter $K_{rootdisc,i}$ can be used to specify the minimum relative conductance value that the plant will tolerate without disconnecting hydraulically from the soil (by default $K_{rootdisc,i} = 0$). If, after possibly accounting for irreversible cavitation, $K_{i,s} < K_{rootdisc,i}$ for a given soil layer, then the model assumes that transpiration from this soil layer is absent. Moreover, $K_{i,s}$ is assumed equal to $K_{rootdisc,i}$ for the sake of determining plant water potential.

3.7 Bare soil evaporation

Evaporation from the soil surface is the last component of the soil water balance to be calculated. There is a difference in the way that soil evaporative demand is calculated depending on the transpiration mode. Potential evaporation from the soil (PE_{soil} ; in $mm \cdot day^{-1}$) is defined as the product between PET and L_{ground}^{SWR} , the proportion of SWR absorbed by the ground:

$$PE_{soil} = PET \cdot L_{ground}^{SWR} \quad (35)$$

Actual evaporation from the soil surface is modeled as in Mouillot et al. (2001), who followed Ritchie (1972). First, the model determines the time needed to evaporate the current water deficit (difference between field capacity and current moisture) in the surface soil layer:

$$t = \left\{ \frac{V_1 \cdot (1 - W_1)}{\gamma_{soil}} \right\} \quad (36)$$

where γ_{soil} is the maximum daily evaporation ($mm \cdot day^{-1}$). The calculated time is used to determine the ‘supplied’ evaporation, SE_{soil} :

$$SE_{soil} = \gamma_{soil} \cdot (\sqrt{t+1} - \sqrt{1}) \quad (37)$$

The amount of water evaporated from the soil, E_{soil} , is then calculated as the minimum between supply and demand (Federer, 1982), the latter being the product of PET and the proportion of light that reaches the ground (see function `spwb.SoilEvaporation`):

$$E_{soil} = \min(PE_{soil}, SE_{soil}) \quad (38)$$

Finally, E_{soil} is distributed along the soil profile according to an exponential decay function with an extinction coefficient κ_{soil} (Mouillot et al., 2001).

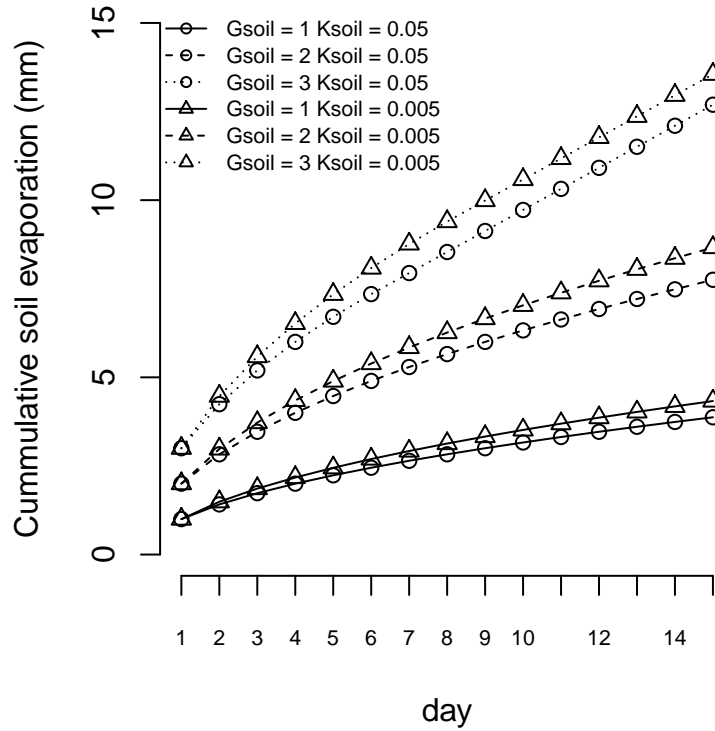


Fig. 4: Cumulative bare soil evaporation for different values of maximum evaporation rate γ_{soil} and extinction coefficient κ_{soil} . Three soil layers (0 – 30 cm; 30 – 150 cm; 150 – 400 cm) are initialized at field capacity ($V_1 = 50mm$; $V_2 = 201mm$; $V_3 = 35mm$). PE_{soil} was assumed not to be limiting. When the extinction coefficient is smaller a higher proportion of the evaporated water is removed from the subsoil and less from the topsoil. This causes more water being available to calculate t in the next step.

3.8 Landscape hydrological processes

To simulate runoff from one cell to the other, the approach of Ostendorf & Reynolds (1993) is used, as in SIERRA (Mouillot et al. 2001). Water lateral transport depends on topography only. The model determines cell

neighbours following the queen rule (up to eight neighbours per cell). The proportion of water runoff of cell i will be an input to a neighbouring cell j is:

$$q_{ij} = \frac{\Delta z_{ij}/L_{ij}}{\sum_j \Delta z_{ij}/L_{ij}} \quad (39)$$

if $\Delta z_{ij} = z_i - z_j > 0$, that is, if the difference in elevation between the two cells is positive (i.e. if $z_j < z_i$). Otherwise there is no discharge from i to j , i.e. $q_{ij} = 0$. L_{ij} indicates the distance between cell i and j (which depends on cell size and on whether the neighbouring cell j is diagonal to cell i). The sumatory of the denominator is done only for neighbours at lower elevation, so that $\sum_i q_{ij} = 1$.

The table of q_{ij} values is calculated at the beginning of simulations only. Every day, cells are processed in order from higher to lower elevation. After the daily water balance of a given cell i , water runoff R_i is divided among the neighbouring cells situated at lower elevation. The runoff of a neighbour j , O_j is updated as:

$$O_j = O_j + R_i \cdot q_{ij} \quad (40)$$

Note that a given cell j can receive water discharge from more than one neighbour. O_j values are set to zero at the beginning of each day.

4 Model output

Function `spwb` returns a list object whose elements are data tables. Each table has dates as rows and has different output variables:

- "DailyBalance": Components of the climatic and soil daily water balance (i.e. net precipitation, infiltration, runoff, plant transpiration...).
- "SoilWaterBalance": Daily variation of soil variables (volume, moisture relative to field capacity, water potential) for soil layers.
- "PlantLAI": Daily leaf area index for each plant cohort.
- "PlantTranspiration": Daily transpiration (in mm) for each plant cohort.
- "PlantPhotosynthesis": Daily net photosynthesis (in g C·m⁻²) for each plant cohort.
- "PlantPsi": Daily water potential of each plant (in MPa).
- "PlantStress": Daily stress level suffered by each plant cohort (relative whole-plant conductance).

5 References

- Carsel, R.F., and Parrish, R.S. 1988. Developing joint probability distributions of soil water retention characteristics. *Water Resources Research* 24: 755–769.
- Collins, D.B.G., Bras, R.L., 2007. Plant rooting strategies in water-limited ecosystems. *Water Resour. Res.* 43, W06407. doi:10.1029/2006WR005541
- De Cáceres, M., Martínez-Vilalta, J., Coll, L., Llorens, P., Casals, P., Poyatos, R., Pausas, J.G., Brotons, L., 2015. Coupling a water balance model with forest inventory data to predict drought stress: the role of forest structural changes vs. climate changes. *Agric. For. Meteorol.* 213, 77–90. doi:10.1016/j.agrformet.2015.06.012
- Deguchi, A., Hattori, S., Park, H.-T., 2006. The influence of seasonal changes in canopy structure on interception loss: Application of the revised Gash model. *J. Hydrol.* 318, 80–102. doi:10.1016/j.jhydrol.2005.06.005
- Federer, C., 1982. Transpirational supply and demand: plant, soil, and atmospheric effects evaluated by simulation. *Water Resour. Res.* 18, 355–362.
- Fyllas, N.M., Troumbis, A.Y., 2009. Simulating vegetation shifts in north-eastern Mediterranean mountain forests under climatic change scenarios. *Glob. Ecol. Biogeogr.* 18, 64–77. doi:10.1111/j.1466-8238.2008.00419.x
- Gash, J., Lloyd, C., Lachaud, G., 1995. Estimating sparse forest rainfall interception with an analytical model. *J. Hydrol.* 170.
- Granier, A., Bréda, N., Biron, P., Villette, S., 1999. A lumped water balance model to evaluate duration and intensity of drought constraints in forest stands. *Ecol. Modell.* 116, 269–283.
- Granier, A., Reichstein, M., Bréda, N., Janssens, I.A., Falge, E., Ciais, P., Grünwald, T., Aubinet, M., Berbigier, P., Bernhofer, C., Buchmann, N., Facini, O., Grassi, G., Heinesch, B., Ilvesniemi, H., Keronen, P., Knohl, A., Köstner, B., Lagergren, F., Lindroth, A., Longdoz, B., Loustau, D., Mateus, J., Montagnani, L., Nys, C., Moors, E., Papale, D., Peiffer, M., Pilegaard, K., Pita, G., Pumpanen, J., Rambal, S., Rebmann, C., Rodrigues, A., Seufert, G., Tenhunen, J., Vesala, T., Wang, Q., 2007. Evidence for soil water control on carbon and water dynamics in European forests during the extremely dry year: 2003. *Agric. For. Meteorol.* 143, 123–145. doi:10.1016/j.agrformet.2006.12.004
- Jarvis, P., McNaughton, K., 1986. Stomatal control of transpiration: Scaling Up from leaf to region. *Adv. Ecol. Res.* 15, 1–49.

- Kergoat, L. 1998. A model for hydrological equilibrium of leaf area index on a global scale. *Journal of Hydrology* 212–213: 268–286.
- Linacre, E.T., 1968. Estimating the net-radiation flux. *Agric. Meteorol.* 93, 49–63.
- Liu, B. Y. H. and Jordan, R. C. “The interrelationship and characteristic distribution of direct, diffuse and total solar radiation,” *Solar Energy*, vol. 4, no. 3, pp. 1–19, 1960.
- Miralles, D.G., Gash, J.H., Holmes, T.R.H., de Jeu, R.A.M., Dolman, A.J., 2010. Global canopy interception from satellite observations. *J. Geophys. Res.* 115, D16122. doi:10.1029/2009JD013530
- Mouillot, F., Rambal, S., Joffre, R., 2002. Simulating climate change impacts on fire frequency and vegetation dynamics in a Mediterranean-type ecosystem. *Glob. Chang. Biol.* 8, 423–437.
- Mouillot, F., Rambal, S., Lavorel, S., 2001. A generic process-based SIMulator for mediterranean landscApes (SIERRA): design and validation exercises. *For. Ecol. Manage.* 147, 75–97. doi:10.1016/S0378-1127(00)00432-1
- Ostendorf, B., Reynolds, J.F., 1993. Relationships between a terrain-based hydrologic model and patch-scale vegetation patterns in an arctic tundra landscape. *Landsc. Ecol.* 8, 229–237. doi:10.1007/BF00125130
- Prentice, I.C., Sykes, M.T., Cramer, W., 1993. A simulation model for the transient effects of climate change on forest landscapes. *Ecol. Modell.* 65, 51–70. doi:10.1016/0304-3800(93)90126-D
- Reynolds, C.A., Jackson, T.J., Rawls, W.J., 2000. Estimating soil water-holding capacities by linking the Food and Agriculture Organization Soil map of the world with global pedon databases and continuous pedotransfer functions. *Water Resour. Res.* 36, 3653–3662. doi:10.1029/2000WR900130
- Ritchie, J., 1972. Model for predicting evaporation from a row crop with incomplete cover. *Water Resour. Res.* 8, 1204–1213.
- Ruffault, J., Martin-StPaul, N.K., Duffet, C., Goge, F., Mouillot, F., 2014. Projecting future drought in Mediterranean forests: bias correction of climate models matters! *Theor. Appl. Climatol.* 117, 113–122. doi:10.1007/s00704-013-0992-z
- Saxton, K.E., Rawls, W.J., Romberger, J.S., Papendick, R.I., 1986. Estimating generalized soil-water characteristics from texture. *Soil Sci. Soc. Am. J.* 50, 1031–1036.

- Saxton, K. E., Rawls, W. J. (2006). Soil water characteristic estimates by texture and organic matter for hydrologic solutions. *Soil Science Society of America Journal*, 70(5), 1569. <https://doi.org/10.2136/sssaj2005.0117>
- Schenk, H., Jackson, R., 2002. The global biogeography of roots. *Ecol. Monogr.* 72, 311–328.
- Sitch, S., Smith, B., Prentice, I.C., Arneth, a., Bondeau, a., Cramer, W., Kaplan, J.O., Levis, S., Lucht, W., Sykes, M.T., Thonicke, K., Venevsky, S., 2003. Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ dynamic global vegetation model. *Glob. Chang. Biol.* 9, 161–185. doi:10.1046/j.1365-2486.2003.00569.x
- Sperry, J.S., Love, D.M., 2015. What plant hydraulics can tell us about responses to climate-change droughts. *New Phytol.* 207, 14–27. doi:10.1111/nph.13354
- Spitters, C.J.T., Toussaint, H.A.J.M., Goudriaan, J. 1986. Separating the diffuse and direct components of global radiation and its implications for modeling canopy photosynthesis. I. Components of incoming radiation. *Agricultural and Forest Meteorology*, 38, 231–242.
- Stolf, R., Thurler, Á., Oliveira, O., Bacchi, S., Reichardt, K., 2011. Method to estimate soil macroporosity and microporosity based on sand content and bulk density. *Rev. Bras. Ciencias do Solo* 35, 447–459.
- Watanabe, T., Mizutani, K., 1996. Model study on micrometeorological aspects of rainfall interception over an evergreen broad-leaved. *Agric. For. Meteorol.* 80, 195–214.
- van Genuchten, M. 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society of America Journal*, 44, 892–898.
- Toth, B., Weynants, M., Nemes, A., Mako, A., Bilas, G., Toth, G. 2015. New generation of hydraulic pedotransfer functions for Europe. *European Journal of Soil Science* 66: 226–238.