Part V

Soil processes

1 ISBA soil-vegetation scheme

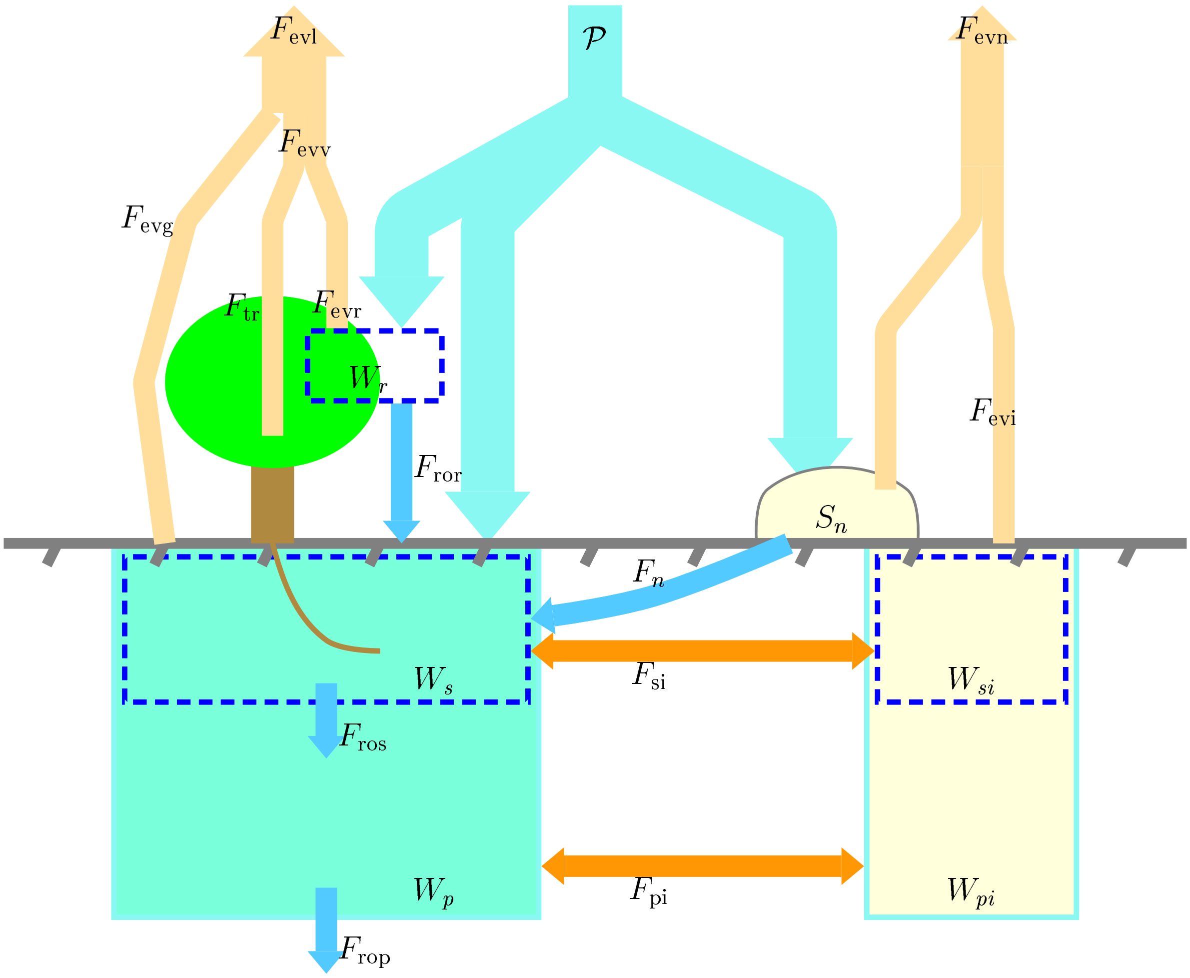

Figure 3: Synthetic view of the different reservoirs and water transfers in the operational ISBA scheme

The ISBA (from the french: *"Interactions Sol Biosphère Atmosphère"*) scheme elaborated by Noilhan and Planton [1989] parameterizes the energetic and hydrous exchanges at the continental surface, using a minimum set of parameters preserving a good representation of the physical phenomena at the interface.
Important effects are the soil thermo-hydrous properties, linked to their composition and water contents; also the feeble thermal inertia of the vegetation and its ability to re-evaporate immediately the dew and precipitation it intercepts, and to modulate the soil surface evaporation.

1.1 "Climatologic" description

1.1.1 Primary parameters: soil and vegetation type

The original implementation used a classification with 18 vegetation types, derived from a 53-types classification from Wilson and Henderson-Sellers [1985], and 23 soil types. The ISBA scheme uses various fields to describe vegetation, as the vegetation fraction, leaf area index, root depth,... [see Giard and Bazile, 2000] so that we may reduce the particular field "vegetation types" to only four classes iveg:

1. sea,
2. ice cap,

constant fields				
VARSF(5)	δland	'SURFIND.TERREMER'	Land-sea mask	$(0,1)$
VARSF(6)	veg	'SURFPROP.VEGETAT'	vegetation fraction	$0 \le$ veg ≤ 1
ISBA "constant" fields				
VCLIV(1)	iveg	'SURFIND.VEG.DOMI'	land-use type	(1, 2, 3, 4)
VCLIV(2)	$R_{s\,\min}$	'SURFRESI.STO.MIN'	minimum surface resistance	SRSMN $\le$ RSMIN $\le$ SRSMX $=$ RSMAX (RSMIN=SRSMX over sea)
VCLIV(3)	arg	'SURFPROP.ARGILE '	percentage of clay	SARGN $\le$ arg $\le$ SARGX (arg=SARGN over sea)
VCLIV(4)	sab	'SURFPROP.SABLE '	percentage of sand	SSABN $\le$ sab $\le$ SSABX (sab=SSABN over sea)
VCLIV(5)	d_2	'SURFEPAIS.SOL '	useful soil depth	SDEPN $\le d_2 \le$ SDEPX ($d_2 =$ SDEPX over sea)
VCLIV(6)	lai	'SURFIND.FOLIAIRE'	leaf-area index	$0 \le$ lai (lai $= 0$ over sea)
VCLIV(7)	hv	'SURFRES.EVAPOTRA'	resistance to evapotranspiration, intermediate field	$0 \le$ hv ≤ 1; up to now hv is set to 1 in 927
VCLIV(8)	GZ0H	'SURFGZ0.THERM '	g*thermal roughness length	STHER $*$ SZZ0N $*$ RG $\le$ GZ0H over land, recomputed over sea ($g *$ SZZ0M at the end of 927). The ratio of thermal to dynamical roughness lengths is equal to 1 over sea, STHER over land
Surface variables with ISBA				
SOILB(1,1)	T_p	'PROFTEMPERATURE '	mean soil temperature	
SOILB(1,2)	W_p	'PROFRESERV.EAU '	total liquid soil water contents	$0 \le W_p \le$ WPMX(i) computed by ACSOLW ($W_p =$ WPMX(i) over sea) WPMX$(i) =$ GCONV $* d_2(i) *$ $W_{\mathrm{sat}}(\mathrm{sab}(i), \delta\mathrm{land}(i), \mathrm{iveg}(i))$
SOILB(1,3)	W_{pi}	'PROFRESERV.GLACE'	total frozen soil water contents	$0 \le W_{pi} \le \min(\mathrm{GWPIMX}, \mathrm{WPMX}(i))$ and $0 \le W_p + W_{pi} \le$ WPMX(i) ($W_{pi} = 0$ over sea)
SNOWG(1)	S_n	'SURFRESERV.NEIGE'	snow reservoir water contents	$0 \le S_n$ ($S_n = 0$ over sea)
RESVR(1)	T_s	'SURFTEMPERATURE '	superficial temperature,	representing bare ground, vegetation and snow
RESVR(2)	W_s	'SURFRESERV.EAU '	superficial liquid soil water contents	$0 \le W_s \le W_{s\,\max}(i)$ computed by ACSOLW ($W_s = W_{s\,\max}(i)$ over sea) $W_{s\,\max}(i) =$ GCONV $* \mathrm{d}_1 *$ $W_{\mathrm{sat}}(\mathrm{sab}(i), \delta_{\mathrm{land}}, \mathrm{iveg}(i))$
RESVR(3)	W_r	'SURFRESERV.INTER'	interception water contents, water on the leaves	$0 \le W_r \le W_{r\,\max}(i)$ computed by ACVEG $W_{r\,\max}(i) =$ GWLMX $*$ lai$(i) *$ veg(i) up to now W_r is set to 0 in 927
RESVR(4)	W_{si}	'SURFRESERV.GLACE'	superficial frozen soil water contents	$0 \le W_{si} \le W_{s\,\max}(i)$ and $0 \le W_s + W_{si} \le W_{s\,\max}(i)$ ($W_{si} = 0$ over sea)

Figure 4: ISBA scheme variables, their names in the code and in the "climatological" files

3. low vegetation or desert,
4. high vegetation.

iveg is only used in surface resistance computations through $F_{\odot\,\mathrm{lim}}$ in function F_1 and parameter γ in function F_3 (see §1.5.1).
The other physical descriptors (see secondary parameters) provide the necessary refinements.

1.1.2 Secondary parameters: physical characteristics

The soil is described through

- clay arg and sand sab percentages, with $\mathsf{arg}+\mathsf{sab}+\mathsf{silt}=100$ [%]:

$$\mathsf{SARGN} \le \mathsf{arg} \le \mathsf{SARGX}\ (\mathsf{arg}=\mathsf{SARGN}\text{ over sea})$$
$$\mathsf{SSABN} \le \mathsf{sab} \le \mathsf{SSABX}\ (\mathsf{sab}=\mathsf{SSABN}\text{ over sea})$$

- Useful soil depth $d_2[m]$ represents the depth at which the moisture flux in the soil vanishes after about 1 week of integration: it is bigger than the root depth and controls deep soil runoff. We have $\mathsf{SDEPN} \le d_2 \le \mathsf{SDEPX}$, and over sea $d_2 \equiv \mathsf{SDEPX}$. $d_1 \equiv \mathsf{RD1}[m]$ and d_2 represent the respective capacities of the surface and the deep reservoirs: corresponding maximum contents $[kg\,m^{-2}]$ are obtained by multiplying them by w_{sat}:

$$\begin{array}{lcll} w_{\mathrm{sat}} & \equiv & \mathsf{G1WSAT}\cdot\mathsf{sab}+\mathsf{G2WSAT}\ \ [pu] & \\ & & \mathsf{G1WSAT}=-1.08\cdot 10^{-3},\ \ \mathsf{G2WSAT}=494.314\cdot 10^{-3} & \\ W_{s\,\max} & = & d_1 w_{\mathrm{sat}}\mathsf{GCONV}\ \ [kg\,m^{-2}] & \mathsf{GCONV}=10^3\ \ [kg\,m^{-3}] \\ W_{p\,\max} & = & d_2 w_{\mathrm{sat}}\mathsf{GCONV}\ \ [kg\,m^{-2}] & \end{array} \tag{1}$$

In this document we'll note W the reservoirs contents $[kg\,m^{-2}]$ as used in the code, and w volumetric specific contents $[m^3$ water$/m^3$ soil$]$. Remember that water fluxes are expressed in $[kg\,m^{-2}\,s^{-1}]$.

The saturation fraction of the total reservoir is then

$$Fs_p \equiv \frac{W_p}{W_{p\,\max}} \tag{2}$$

Under key LIMW, the wilt point gives a lower boundary for Fs_p:

$$W_{\mathrm{wilt}} \ = \ \mathsf{GWWILT}\cdot\mathsf{arg}^{\mathsf{EWWILT}}\ [kg/m^2] \qquad \mathsf{GWWILT}=37.1342\cdot 10^{-}3,\ \mathsf{EWWILT}=0.5$$

The surface water contents at equilibrium between capillarity forces and gravity is computed as:

$$\begin{array}{lcll} W_{\mathrm{seq}} & = & \left(Fs_p - a(Fs_p)^p(1-(Fs_p)^{8p})\right)W_{s\,\max} & \\ a & \equiv & \mathsf{GA}\cdot\mathsf{arg}^{\mathsf{EA}} & \mathsf{GA}=732.42\cdot 10^{-3},\ \mathsf{EA}=-0.54 \\ b & \equiv & \mathsf{G1B}\cdot\mathsf{arg}+\mathsf{G2B} & \mathsf{G1B}=0.137,\ \mathsf{G2B}=3.5 \\ p & \equiv & \mathsf{G1P}\cdot\mathsf{arg}+\mathsf{G2P} & \mathsf{G1P}=0.134,\ \mathsf{G2P}=3.4 \end{array} \tag{3}$$

Field capacity contents is given by

$$w_{\mathrm{fc}} \ = \ \mathsf{GWFC}\cdot\mathsf{arg}^{\mathsf{EWFC}}\ \ [kg/m^3] \qquad \mathsf{GWFC}=89.0467\cdot 10^{-3},\ \mathsf{EWFC}=0.35 \tag{4}$$

and represents the volumetric contents of the deep reservoir above which a drainage towards deeper ground will start.

- the vegetation fraction $\mathsf{PVEG0} \equiv \mathsf{veg}$ is defined as the part of the canopy preventing radiation from reaching the ground. Related parameters are SURFPROP.VEGETAT and SURFPROP.VEG.MAX.

- Surface resistance R_s is the resistance to water transfers from the roots up to the leaves surface. It modulates vegetation transpiration through a dependency to various factors: air temperature and moisture, solar radiation, soil moisture around the roots. Its minimal value $R_{s\,\min}$ for a given leaf has been estimated from numerical simulations.
$\mathsf{SRSMN} \le R_{s\,\min} \le \mathsf{SRSMX} = \mathsf{RSMAX}$ and over sea $R_{s\,\min} = \mathsf{SRSMX}$.

- Leaf area index lai is defined as the leaf surface per unit soil surface. It was deduced from Mahfouf et al. [1995]'s classification following the dominant vegetation type. It is 0 over sea.
- Albedo (SURFALBEDO) and roughness length (GZ0H $\equiv$ SURFGZ0.THERM) were derived from Wilson and Henderson-Sellers [1985]. The ratio of thermal to dynamical roughness lengths is equal to 1 over sea, STHER $= 10$ over land.
- Emissivity (SURFEMISSIVITE) was obtained by combining various sources.

To this we add climatological values, derived from a re-analysis over 1 year with the pre-operational ISBA (Bazile and Giard), for the parameters

- Surface and deep soil temperature (SURFTEMPERATURE, PROFTEMPERATURE),
- Different reservoirs relative levels

1.1.3 Water transfers physics

Theory [see Clapp and Hornberger, 1978] describes physics of soil water transfers through parameters:

- w_{sat} saturation water volumetric contents $[m^3/m^3]$;
- K_{sat} saturation hydraulic conductivity $[m/\text{day}]$;
- ψ_{sat} saturation matricial potential associated to capillarity forces $[m]$;
- b slope of the retention curve on a logarithmic chart (showing the soil inclination to retain water or let it percolate);

Following equations link the matricial potential ψ and the hydraulic conductivity K to the volumetric water contents W:

$$\begin{aligned} \psi &= \psi_{\text{sat}}(w_{\text{sat}}/w)^b \\ K &= K_{\text{sat}}(w/w_{\text{sat}})^{2b+3} \end{aligned} \tag{5}$$

From them you can derive

- the volumetric water contents at wilt point $w_{\text{wilt}} \equiv w(\psi = -150m)$;
- the volumetric water contents at field capacity $w_{\text{fc}} \equiv w(K = 10^{-4}m/d)$

Ground volumetric heat capacity c_g and thermal conductivity λ depend to w and ψ following laws from Mc Cumber and Pielke [1981]:

$$\begin{aligned} c_g &= (1 - w_{\text{sat}})c_i + w \cdot c_w \qquad [J\,m^{-3}\,K^{-1}] \\ \lambda &= \begin{cases} 418\,e^{-\log\psi+2.7} & \text{if } \log|\psi| \leq 5.1 \\ 0.171 & \text{otherwise} \end{cases} \qquad [Js^{-1}m^{-1}K^{-1}] \end{aligned}$$

c_i and c_w being the volumetric heat capacities for soil and water. λ and c_g appear in the ground thermal coefficient C_T derived below.

1.2 Prognostic variables

The model had initially 5 prognostic variables:

- T_s: surface temperature, concerning the ground surface and the vegetation $[K]$;
- T_p: deep soil temperature $[K]$;
- W_s surface reservoir water contents $[kg/m^2]$;
- W_p deep (or total) reservoir water contents $[kg/m^2]$;
- W_r interception reservoir volumetric water contents $[kg/m^2]$.

to which were added afterwards:

- S_n the snow reservoir $[kg/m^2]$;

- ρ_n snow density, $[kg/m^3]$ and A_n snow albedo, for the scheme from DOUVILLE et al. [1995] (not in the operational model);
- W_{pi} the total soil ice reservoir $[kg/m^2]$;
- W_{si} the surface ice reservoir $[kg/m^2]$.

1.3 Heat transfers in the ground

Evolution of prognostic variables T_s and T_p is taken from BHUMRALKAR [1975] and BLACKADAR [1976].

1.3.1 Underground temperature

Model implementation allows to define (under key LSOLV) $n+1 = \mathsf{NCSS}+1 \leq 10$ inner ground levels through their depths: $z_0 = 0, z1, ...z_{n+1}$ and thicknesses: $\triangle z_i \equiv z_{i+1} - z_i$.
Between the layers, the sensible heat flux PFCHSP $\equiv F_{\rm sp}$ is $F^0_{\rm sp} = G(t)$ at the surface, $F^1_{\rm sp}$, $F^2_{\rm sp}$, ..., $F^n_{\rm sp}$ below, and $F^{n+1}_{\rm sp} = 0$.

Underground temperature evolution is lead by

$$c_g \frac{\partial T_p}{\partial t} = -\frac{\partial F_{\rm sp}}{\partial z}, \qquad F_{\rm sp}(z) = -\lambda \frac{\partial T_p}{\partial z} \tag{6}$$

$$\Longrightarrow \qquad \frac{\partial T_p}{\partial t} = \frac{\lambda}{c_g}\frac{\partial^2 T_p}{\partial z^2} \tag{7}$$

To evacuate λ and c_g coefficients, we introduce reduced coordinates:

- $\omega_0 = 2\pi/86400$ is the diurnal pulsation
- $H_0 = \sqrt{\frac{2\lambda}{c_g\omega_0}}$ is the depth at which the diurnal wave is attenuated by a factor $1/e$: we call it the *diurnal wave e-folding depth*. It is easily obtained by solving system (6), letting $\hat{T}(z,\omega)$ be the Fourier time transform of $T(z,t)$:

$$\hat{T}(z,\omega) = \hat{T}(0,\omega)e^{-(1+i)z/H(\omega)} \qquad \Longrightarrow H(\omega) = \sqrt{\frac{2\lambda}{c_g\omega}}$$

  hence $H_0 \equiv H(\omega_0)$.

- The reduced coordinate $x \equiv z/H_0$, the reduced thicknesses are $\mathsf{SODELX(0:9)} \equiv \triangle x$

$$\triangle x_0 = 1/\sqrt{2\pi+1} \quad , \quad \triangle x_k = 2\pi \triangle x_{k-1}$$

Using PCT $\equiv C_T = \frac{1}{c_g H_0 \triangle x_0} = \sqrt{\frac{\omega_0(2\pi+1)}{2\lambda c_g}}$ (see actual calculation §1.3.3) and $u_k \equiv \frac{\triangle x_k}{\triangle x_0}$ we have:

$$\frac{\partial T^k_p}{\partial t} = \delta_{\rm land} C_T \frac{1}{u_k}(F^k_{\rm sp} - F^{k+1}_{\rm sp}) = \frac{1}{c_g H_0 \triangle x_k}(F^k_{\rm sp} - F^{k+1}_{\rm sp}) \qquad k = 1 \ldots n-1 \tag{8}$$

$$\frac{\partial T^n_p}{\partial t} = \delta_{\rm land} C_T \left(\frac{1}{u_n} F^n_{\rm sp} + L_{w-i} F_{pi}\right) \tag{9}$$

The deep soil freezing flux F_{pi} is added (under LFGEL) at the lowest underground level. It is also at this place that an exchange with deeper ground may be introduced (see parameter OMTPRO below).

Dependence to C_T (taking into account thermal properties of soil, vegetation and snow) is illusory (excepted at the surface), because there is a simplification between the conservation and the conduction equation in (6). The ground is supposed homogeneous over the vertical between the surface and the depth where fluxes are negligible (seasonal flux extinction). The only parameters are the levels, expressed as multiples of the e-folding depth of the diurnal wave, which takes into account the horizontal inhomogeneities.

The heat flux between two ground levels is computed in ACDIFUS as:

$$F_{\rm sp}^k = \frac{\omega_0}{C_T(\triangle x_0)^2}\frac{1}{u_{k-1}+u_k}(T_p^{k-1}-T_p^k) = \frac{\omega_0}{C_T\triangle x_0(\triangle x_{k-1}-\triangle x_k)}(T_p^{k-1}-T_p^k) \qquad k=2\dots n \quad (10)$$

$$F_{\rm sp}^1 = \frac{\omega_0}{C_T\triangle x_0(\triangle x_0-\triangle x_1)}(T_s^+-T_p^1) \quad (11)$$

T_s^+ appears in surface temperature computation in ACDIFUS.

If LSOLV is not set, we just have to set $\triangle x_0=1,\ \triangle x_k=0,\ k=2\dots n$ in previous formulae.
In this case, only the highest ground layer temperature evolves, the lower ones keeping a constant temperature:

$$\frac{\partial T_p^1}{\partial t} = \delta_{\rm land}\frac{\mathsf{HSOL}}{\mathsf{RTINER}}F_{\rm sp}^1$$

where HSOL is the inverse of the surface soil thermic inertia and RTINER is the ratio of the deep thermal inertia to the surface one.

The tendency is stored by CPTENDS in PTDTP and the physical time step (CPWTS) writes:

$$\begin{aligned} T_{p1}^k &= T_{d1}^k + \triangle t\cdot \mathsf{PTDTP}^k \qquad\qquad k=1,n-1 \\ T_{p1}^n &= T_{d1}^n + \triangle t\cdot(\mathsf{PTDTP}^n + \delta_{\rm land}\mathsf{OMTPRO}\cdot(T_{\rm clim}-T_{p1}^n)) \end{aligned}$$

where OMTPRO $\equiv 2\pi/$(deep temperature time constant) has its default at 0: so we may introduce a relaxation to a climatological value at this stage, by setting OMTPRO to some positive value.

The operational model uses only 2 layers: NCSS $=1$, which doesn't allow to take into account the thermal inertia between winter and summer and would induce an exaggeration of the seasonal cycle - if there was no analysis of the surface temperature (operational version) or a deep soil relaxation (climate version). With 4 judiciously chosen layers, precision and discretization are satisfying over time scales between hours and a year. The operational choice (concerned with short and medium forecast ranges) makes a compromise between the precision and the computational load.

1.3.2 Surface temperature

For the surface temperature, we have:

$$\frac{\partial T_s}{\partial t} = \delta_{\rm land}C_T(Q_R+Q_{\rm sens}+Q_{\rm lat}-F_{\rm sp}-L_{w-i}(F_n-F_{si})) \quad (12)$$

- Q_R is the surface net radiative energy flux,
- $Q_{\rm sens}$ is the surface sensible heat flux,
- $Q_{\rm lat}=Q_{w\rm lat}+Q_{i\rm lat}$ is the surface latent heat flux (associated to liquid and solid water),
- $F_{\rm sp}$ is the heat flux between surface and deep ground.
- F_n is the snow melting flux, $L_{w-i}\equiv$ RLMLT is the melting heat;
- F_{si} is the surface freezing flux (under LFGELS, otherwise use F_{pi} instead);

The tendency is stored by CPTENDS in PTDTS and the physical time step writes (CPWTS):

$$T_{s1} = T_{s1} + \triangle t\cdot \mathsf{PTDTS}$$

1.3.3 Ground thermic coefficient

The thermic coefficient C_T $[K\,m^{-2}\,J^{-1}]$ is estimated as follows:

- for bare ground, we use the saturation fraction of the total reservoir Fs_p, with a possible lower boundary value at the wilt point under key LIMW:

$$C_G = C_{G\mathrm{sat}}(Fs_p)^{-\frac{b\ln 10}{2}} \qquad C_{G\mathrm{sat}} = \mathsf{G1CGSAT}\cdot\mathsf{sab} + \mathsf{G2CGSAT}\cdot\mathsf{arg} + \mathsf{G3CGSAT}$$
$$\mathsf{G1CGSAT} = -1.5571\cdot 10^{-8},\ \mathsf{G2CGSAT} = -1.441\cdot 10^{-8},\ \mathsf{G3CGSAT} = 4.70217\cdot 10^{-6}$$

Under key LFGEL, the resulting value is modified by

$$\frac{1}{C'_G} = \frac{W_p}{(W_p+W_{pi})}\frac{1}{C_G} + \frac{W_{pi}}{(W_p+W_{pi})}\frac{1}{\mathsf{RCTGLA}}, \qquad \mathsf{RCTGLA} = 5.5\,10^{-6}$$

When the soil is very dry, we limit C_G to the value $\mathsf{RCGMAX} = 0.8\cdot 10^{-5} = \mathsf{RCTVEG}$, to cure some problems of too cold night temperatures associated to a very big C_T. Model key LIMC controls this limitation.

- for vegetation alone, we take:

$$C_V = 0.8\cdot 10^{-5}[K\ m^2 J^{-1}] \equiv \mathsf{RCTVEG}$$

- for the snow, under key LSNV (not set in the operational run) we suppose

$$\lambda_n = 2.22\,\rho_n^{1.885} \qquad c_n \propto \rho_n \qquad \rho_i = 0.910^3[kg\,m^{-3}] \qquad c_i = 1.89510^6[J\,m^{-3}\,K^{-1}]$$

Final expression is then

$$\mathsf{PCN} \equiv C_N = \sqrt{\frac{2\omega_0}{\lambda_n c_n}} = 2\sqrt{\frac{\mathsf{GSNC1}}{(\rho_n)^{\mathsf{GSNC2}}}}$$
$$\mathsf{GSNC1} = \frac{\pi}{2.22\cdot 2106\cdot 86400\cdot 1000} \qquad \mathsf{GSNC2} = 2.885$$

- coefficient for the whole system ground-vegetation-snow is obtained by weighting previous contributions by snow and vegetation fractions:

$$\frac{1}{C_T} = (1-\mathsf{veg})(1-P_{nc})\frac{1}{C_G} + \mathsf{veg}_a\frac{1}{C_V} + ((1-\mathsf{veg})P_{nc} + \mathsf{veg} - \mathsf{veg}_a)\frac{1}{C_N}$$

with

  - S_n is the snow reservoir, P_{nc} is the fraction of the mesh covered by snow, for use in thermal calculations (see §1.4.3).
  - veg is the vegetation fraction, so $(1-\mathsf{veg})(1-P_{nc})$ is the fraction of bare ground not covered by snow;
  - veg_a is the "apparent vegetation", i.e. the vegetation fraction not covered by snow (§1.4.3).

If LSNV is not active, just set $\mathsf{veg}_a = \mathsf{veg}$ and $P_{nc} = 0$ in previous expression.

1.4 Water transfers in the ground

The surface reservoir W_s represents a thin layer ($d_1 \sim 1cm$) in direct interaction with the atmosphere, while the deep reservoir W_p, corresponds to a soil column of depth d_2.
The model includes now also the interception reservoir W_r, the surface and deep ice reservoirs W_{si} and W_{pi}.

1.4.1 Interception Reservoir

Precipitation and dew intercepted by vegetation feed the interception reservoir W_r. If a part δ of the leaves is covered by a thin layer of water, this part is going to evaporate while the rest $1-\delta$ is going to transpire. The evolution equation writes:

$$\frac{\partial W_r}{\partial t} = \mathsf{veg}_e\cdot\mathcal{P} + F_{\mathrm{evv}} - F_{\mathrm{tr}} - F_{\mathrm{ror}} \qquad (13)$$

with

- $\mathcal{P}$ is the precipitation at the top of the vegetation. The fraction veg of the precipitation is used to fill the reservoir; as it has a very small capacity, it could be saturated within a time step in case of intense precipitation.
  As convective precipitation concerns by implementation only a fraction of the mesh, we consider only an effective vegetation fraction veg_e for their interception:

$$\mathsf{veg}_e = \frac{\mathsf{veg}}{1 + \frac{\mathcal{P}_{cl}}{\mathcal{P}_{\text{crit}}}}$$

  $\mathcal{P}_{cl}$ represents the *liquid* convective precipitation at the surface and $\mathcal{P}_{\text{crit}} \equiv 1/\mathsf{USUPRC}$ is a critical value limiting the amount of convective rain intercepted by vegetation.

- F_{evv} it the evapotranspiration, sum of the transpiration F_{tr} and the evaporation from the interception reservoir F_{evr}:

$$F_{\text{evv}} = \begin{cases} F_{\text{tr}} + F_{\text{evr}} & \text{if } F_{\text{evr}} \geq 0 \\ F_{\text{dew}} & \text{if } F_{\text{evr}} < 0 \end{cases}$$

  negative evaporation from the interception reservoir implies there is no more transpiration.

- F_{ror} is the runoff from the interception reservoir, starting when W_r reaches a value $W_{r\,\text{max}}$ depending on the density of the vegetation cover:

$$\mathsf{WLMX} \equiv W_{r\,\text{max}} = 0.2 \cdot \mathsf{veg} \cdot \mathsf{lai} \;\; [kg\,m^{-2}] \tag{14}$$

1.4.2 surface and deep reservoirs

With 2 reservoirs, evolution was lead by

$$\frac{\partial w_s}{\partial t} = \underbrace{\frac{C_1}{\rho_w d_1}(\mathcal{P}_g + F_{\text{evg}})}_{\text{surface atmospheric fluxes}} - \underbrace{\frac{C_2}{\tau}(w_s - w_{\text{seq}})}_{F_{wsp}\text{ diffusion in ground}} \tag{15}$$

$$\frac{\partial w_p}{\partial t} = \frac{1}{\rho_w d_2}(\mathcal{P}_g + F_{\text{evg}} + F_{\text{tr}}) - \underbrace{\frac{C_3}{\tau}\max(0, W_p - W_{\text{fc}})}_{F_{\text{rop}}} \tag{16}$$

Present system is slightly more sophisticated:

$$\frac{\partial W_s}{\partial t} = \delta_{\text{land}}\left[C_1\left\{(1-\mathsf{veg}_e)\mathcal{P} + F_{\text{evl}} - F_{\text{evv}} + F_{\text{ror}} + F_n\right\} - F_{wsp} - F_{\text{ros}} - F_{si}\right] \tag{17}$$

$$\frac{\partial W_{si}}{\partial t} = \delta_{\text{land}}(F_{si} + F_{\text{evi}}) \tag{18}$$

$$\frac{\partial W_p}{\partial t} = \delta_{\text{land}}\left[\left\{(1-\mathsf{veg}_e)\mathcal{P} + F_{\text{evl}} - F_{\text{evv}} + F_{\text{ror}} + F_n\right\} + F_{\text{tr}} - F_{\text{rop}} - F_{pi} - F_{si}\right] \tag{19}$$

$$\frac{\partial W_{pi}}{\partial t} = \delta_{\text{land}}(F_{pi} + F_{si} + F_{\text{evi}}) \tag{20}$$

where

- $\mathcal{P}_g = (1-\mathsf{veg}_e)\mathcal{P}$ is the precipitation flux at the ground (i.e. precipitation that was not intercepted by vegetation),
- F_{evi} is evaporation flux from ice reservoir, F_{evl} is evaporation flux from liquid water reservoir,
- $F_{\text{evg}} = F_{\text{evl}} - F_{\text{evv}} = F_{\text{evl}} - F_{\text{evr}} - F_{\text{tr}}$ is the bare ground evaporation flux,
- F_n is snow melting, F_{si} the surface freezing/melting flux, F_{pi} the deep freezing/melting flux,
- F_{ror}, F_{ros} and F_{rop} are the runoffs of interception, surface and deep reservoirs,
- $\tau \equiv \mathsf{RDAY}$ is the period of 1 day (86400 s)

All transpiration originates from reservoir W_p, where the roots are situated.
Runoff begins when W_s or W_p overtakes W_{sat}.
When W_p reaches the field capacity W_{fc}, a gravitational drain appears at the bottom.

The hydrous coefficients C_1, C_2, C_3 are calibrated below.

The potential of capillarity forces follows the diffusion equation:

$$\begin{aligned} \frac{\partial \psi}{\partial t} &= \frac{K}{C_w}\frac{\partial^2 \psi}{\partial z^2} \\ \Longrightarrow \psi(z,t) &= \psi_m + (\triangle\psi)e^{-z/d}\sin(\omega t - \frac{z}{d}) \end{aligned}$$

where

- d is the depth reached by the diurnal wave, of pulsation ω ($d^2 = \frac{K}{C_w \omega}$);
- ψ_m is a linear function of z;
- $C_w = \frac{\partial w}{\partial \psi}$ is the hydraulic capacity $[m]$.

From this we deduce the equations giving the potential at initial time, and the superficial reservoir. By identification with (15), we get C_1:

$$\begin{aligned} C_1 &= C_{1\text{sat}}\left(\frac{w_{\text{sat}}}{w_s}\right)^{\frac{b}{2}+1} \\ C_{1\text{sat}} &= (\mathsf{G1C1SAT}\cdot\mathsf{arg} + \mathsf{G2C1SAT})\cdot\mathsf{GC2} \\ & \mathsf{G1C1SAT} = 5.58\cdot 10^{-3},\ \mathsf{G2C1SAT} = 84.88\cdot 10^{-3},\ \mathsf{GC2} = 10 \end{aligned}$$

C_1 increases when the soil dries, as hydraulic conductivity is then reduced.

C_2 characterizes the speed by which the water profile returns to the equilibrium profile. It increases with hydraulic conductivity K.
Let be $F_w(z)$ the water flux at depth z. Setting boundary conditions:

$$\begin{aligned} F_w(0) &= \mathcal{P} + F_{\text{evg}} = 0 \\ F_w(d_2) &= \rho_w K(\frac{\partial\psi}{\partial z} - 1)|_{z=d_2} = 0 \end{aligned}$$

(expressing that the total water contents in the soil column are conserved as long as saturation is not reached) to integrate (15) yields

$$(w_s - w_{\text{seq}})(t) = (w_s - w_{\text{seq}})(t_0)e^{-C_2\frac{t-t_0}{\tau}}$$

If $t_1 - t_0$ is the time constant for this equation, we can estimate C_2 as

$$C_2 = \frac{\tau}{t_1 - t_0}$$

but this approach ignored any runoff. A multilayer model using water profiles differing on their shape as well as on the initial water contents lead to the expression:

$$\begin{aligned} C_2 &= C_{2\text{ref}}\frac{Fs_p}{1 - Fs_p + \varepsilon} \\ & C_{2\text{ref}} = \mathsf{GC2REF}\cdot\mathsf{arg}^{\mathsf{EC2REF}} \qquad \mathsf{GC2REF} = 13.82,\ \mathsf{EC2REF} = 0.95 \end{aligned}$$

ε is a small value limiting C_2 at saturation.

C_3 characterizes the speed by which the water profile W_p is relaxed to the field capacity W_{fc}. Supposing that all water lost is produced by a gravitational drain at the base of the soil layer, we have:

$$\frac{\partial w_p}{\partial t} = -\frac{K}{\tau d_2}$$

K being expressed in m/day and $\tau = 86400s$/day. Using (5) and integrating from saturation ($w_p = w_{\text{sat}}, t = 0$) to (w_p, t):

$$w_p(t) = w_{\text{sat}} \left(1 + (2b+2)\frac{K_{\text{sat}}}{\tau d_2 W_{\text{sat}}} t\right)^{-\frac{1}{2b+2}} \tag{21}$$

Writing the evolution equation as a Newtonian relaxation:

$$\frac{\partial w_p}{\partial t} = C_3 \frac{w_p - w_{\text{fc}}}{\tau} \quad \Longrightarrow \quad w_p = w_{\text{fc}} + (w_{\text{sat}} - w_{\text{fc}}) e^{-\frac{C_3 t}{\tau}}$$

the contents at time $t = \tau / C_3$ are

$$w_p^* = w_{\text{fc}} + \frac{w_{\text{sat}} - w_{\text{fc}}}{e}$$

which we can report in (21) yielding C_3:

$$C_3 = \frac{(2b+2) K_{\text{sat}}}{d_2 w_{\text{sat}} \left[\left(\frac{w_{\text{sat}}}{w_p^*}\right)^{2b+2} - 1\right]}$$

(with K in m/day). Actual expression uses the approached formula:

$$C_3 = \mathsf{GC31} \cdot \mathsf{arg}^{\mathsf{GC32}} \qquad \mathsf{GC31} = 5.3275,\ \mathsf{GC32} = -1.043$$

On sea, we set $C_1 = C_2 = C_3 = 1$.

1.4.3 Snow related variables

Under key LNEIGE, snow reservoir evolution follows:

$$\frac{\partial S_n}{\partial t} = \delta_{\text{land}} (\mathcal{P}_i - F_n + F_{\text{evn}} - F_{\text{evi}}) \tag{22}$$

If LSNV is not set, snow albedo A_n takes constant value ALCRIN, and snow emissivity ϵ_n the value EMCRIN. Net values are then

$$\begin{aligned} \alpha &= (1 - P_n) A_g + P_n A_n \\ \epsilon &= (1 - P_n) \epsilon_g + P_n \epsilon_n \end{aligned}$$

where P_n is the snow fraction, A_g and ϵ_g the bare ground values.
Under key LSNV, snow albedo A_n and density ρ_n are also treated:

$$\frac{\partial A_n}{\partial t} = \delta_{\text{land}} \frac{\mathcal{P}_i}{\mathsf{WNEW}} - \begin{cases} \mathsf{TOEXP}(A_n - A_{n\,\min}) & \text{if } F_n > 0 \\ \mathsf{TOLIN} & \text{otherwise} \end{cases} \tag{23}$$

$$\frac{\partial \rho_n}{\partial t} = -\mathsf{TOEXP}(\rho_n - \rho_{n\,\max}) - \min\left(\frac{\mathcal{P}_i}{S_n}, \frac{1}{2\triangle t}\right)(\rho_n - \rho_{n\,\min}) \tag{24}$$

where

- $\mathcal{P}_i$ is the solid precipitation,
- TOEXP is the rate of exponential decrease in snow albedo and density during snow melting,
- TOLIN the rate of linear decrease in snow albedo during snow accumulation,
- WNEW the minimum snowfall (mm) to renew snow albedo.
- RHOMIN $\equiv \rho_{n\,\min}$ and RHOMAX $\equiv \rho_{n\,\max}$ are minimum and maximum snow densities.

The net albedo α combines bare soil albedo A_g, vegetation albedo A_v and snow albedo A_n, while for the emissivity we only distinguish the value for the ground with no snow ϵ_f from the value for the snow ϵ_n:

$$\begin{aligned} \alpha &= \underbrace{(1 - \mathsf{veg})(1 - P_{ng})}_{\text{bare ground}} A_g + \underbrace{\mathsf{veg}_a}_{\text{apparent vegetation}} A_v + (\underbrace{(1 - \mathsf{veg}) P_{ng}}_{\text{snowy ground}} + \underbrace{\mathsf{veg} - \mathsf{veg}_a}_{\text{snowy vegetation}}) A_n \\ \epsilon &= (1 - \mathsf{veg})(1 - P_{ng}) \epsilon_f + \mathsf{veg}_a \epsilon_f + ((1 - \mathsf{veg}) P_{ng} + \mathsf{veg} - \mathsf{veg}_a) \epsilon_n \end{aligned} \tag{25}$$

P_{ng} is the fraction of the ground covered by snow.
Snow fraction, apparent vegetation fraction and roughness length are affected by the snow as follows.

- If LSNV is not set:

$$\begin{aligned} \mathsf{PNEIJ} &\equiv P_n \equiv \frac{S_n}{S_n + W_{nc}} \\ \mathsf{PVEG} &\equiv \mathsf{veg}_a = \mathsf{veg} \\ z_0 &= z_{0f} + \delta_{\text{land}} \frac{S_n(z_{0n} - z_{0f})}{S_n + W_{nc}(1 + \frac{z_{0f}}{z_{0cn}})} \end{aligned}$$

  with $W_{nc} \equiv$ WCRIN, $z_{0n} \equiv$ ALRCN1 represents the roughness length of the snow, and $z_{0cn} \equiv$ ALRCN2 is a critical roughness length taking into account that more snow is required to cover the whole surface of broken ground.

- under key LSNV, we compute in ACSOL:

  – P_{ng} is the snow fraction covering the ground:

  $$P_{ng} = \frac{S_n}{S_n + W_{nc}(1 + g z_{0g} x_{cg})}$$

  where the factor $x_{cg} \equiv \mathsf{XCRINR} = \frac{1}{g}$ is the inverse of a critical roughness length allowing to increase the critical snow reservoir due to accumulation against the slopes of the orography.

  – P_{nv} is the snow fraction covering the vegetation: it must take into account the height of the vegetation, represented by its bare roughness length z_{0v}:

  $$\begin{aligned} z_{0v} &= \sqrt{z_{0f}^2 - z_{0g}^2} \\ W_{ncz} &= \max(W_{ncg}, \rho_n z_{0v} x_{cv}) \\ P_{nv} &= P_{ng} \frac{S_n + W_{ncg}}{S_n + W_{ncz}} \end{aligned}$$

  z_{0g} is the roughness length of bare ground, while z_{0f} is the net roughness length with no snow. $W_{ncg} \equiv$ WCRING and $x_{cv} \equiv \mathsf{XCRINV} = 1.$ is an adimensional factor allowing to increase the critical snow reservoir to take into account the accumulation of snow supported by high vegetation.

  – $P_{nc} = \min(1, S_n/W_{ncc})$ with parameter $W_{ncc} \equiv$ WCRINC is used locally in computation of thermal coefficient C_T in ACSOL. It has the property to converge faster to 1.

  Net values for the mesh are computed in ACHMT:

  $$\begin{aligned} \mathsf{PNEIJ} \equiv P_n &= \delta_{\text{land}} \left\{ (1 - \mathsf{veg}) P_{ng} + \mathsf{veg} P_{nv} \right\} \\ \mathsf{PVEG} \equiv \mathsf{veg}_a &= \mathsf{veg}(1 - P_{nv}) \\ z_0 &= \sqrt{(1 - P_{nv}) z_{0f}^2 + P_{nv}(z_{0n}^2 + z_{0g}^2)} \end{aligned}$$

  $z_{0n} \equiv$ ALRCN1 represents the roughness length of the snow.

- Model key LDZ0H activates the use of a different roughness length for thermic phenomena: in this case, z_{0f} is replaced by z_{0Hf} the net thermic roughness length in absence of snow, and the net value z_{0H} is used in the computation of the surface exchange coefficients in ACHMT (see part III).

1.5 Surface fluxes

A single energetic equilibrium is considered for the whole soil-vegetation scheme. Heat and mass transfers between surface and the atmosphere depend directly of the average values of T_s and W_s.

1.5.1 Moisture fluxes

Various experiences have shown that the surface evaporates at the potential rate (maximal rate) as soon as its water contents reach the field capacity W_{fc}.
We estimate two weighting coefficients, HU for the saturating moisture and HQ for the air moisture contributions to the surface moisture:

$$q_s = \mathsf{HU}\, q_{s\text{sat}} + \mathsf{HQ}\, q^L \qquad q_{s\text{sat}} \equiv q_{\text{sat}}(T_s, p_s) \tag{26}$$

With the ISBA scheme:

$$\mathsf{HU} = \mathsf{veg}_a \cdot \mathsf{Hv} + (\mathsf{veg} - \mathsf{veg}_a) + (1 - \mathsf{veg})\mathsf{Hub} \qquad \mathsf{HQ} = \mathsf{veg}_a(1 - \mathsf{Hv}) \tag{27}$$

Bare ground coefficient:

$$\mathsf{Hub} = \begin{cases} (1 - P_{nr})f(\mathcal{W}) + P_{nr} & \text{if} \quad \mathcal{W} \leq 1 \\ 1 & \text{if} \quad \mathcal{W} > 1 \end{cases} \tag{28}$$

$$\text{if } \mathsf{Hub} \cdot q_{s\mathrm{sat}}(T_s, p_s) < q^L \Rightarrow \mathsf{Hub} = \min(1, q^L / q_{s\mathrm{sat}}(T_s, p_s))$$

$$\mathcal{W} \equiv W_s / W_{\mathrm{fc}} \qquad f(\mathcal{W}) = \frac{1 - \cos(\pi \mathcal{W})}{2}$$

If LSNV is not set, we just have to set $\mathsf{veg} = \mathsf{veg}_a$ and $P_{nr} = P_n$, simplifying previous expressions.

Halstead's evapotranspiration coefficient Hv is expressed by:

$$\mathsf{Hv} = \begin{cases} (1 - \delta)\frac{\mathsf{R}^L}{\mathsf{R}^L + R_s} + \delta \equiv \mathsf{Htr} + \delta & \qquad \mathsf{R}^L \equiv \frac{1}{C_H|\mathbf{V}^L|} \quad \text{if } F_{\mathrm{evv}} > 0 \\ 1 & \qquad \text{otherwise} \end{cases} \tag{29}$$

When there is no evapotranspiration, the dew flux is supposed to occur at the potential rate.
R^L represents the resistance to evaporation between the interception reservoir and the lowest model level, while R_s is the surface resistance to evapotranspiration.
δ is the fraction of the leaves covered by the intercepted water, and the following expression is proposed by Deardorff [1978]:

$$\delta = \left(\frac{W_r}{W_{r\max}} \right)^{\frac{2}{3}}$$

and we have

$$F_{\mathrm{evv}} = \rho^L \mathsf{veg} \frac{\delta}{\mathsf{R}^L}(q_{s\mathrm{sat}} - q^L) \tag{30}$$

$$F_{\mathrm{tr}} = \rho^L \mathsf{veg} \frac{1 - \delta}{\mathsf{R}^L + R_s}(q_{s\mathrm{sat}} - q^L) \tag{31}$$

Surface resistance R_s is parameterized by

$$R_s = \frac{R_{s\min}}{\mathsf{lai}} \frac{F_1}{F_2 F_3 F_4} \tag{32}$$

- $R_{s\min}$ is the minimum stomatal resistance, depending on the vegetation type and its maturation.
- F_1 measures the influence of solar radiation on photo-synthesis, it expression is given by Dickinson [1984]:

$$F_1 = \frac{1 + f}{f + R_{s\min}/R_{s\max}} \qquad f = 0.55 \frac{F_\odot}{F_{\odot\lim}} \frac{2}{\mathsf{lai}}$$

  where the limit value of the net solar flux $F_{\odot\lim}$ is $30[W\ m^{-2}]$ for forest and $100[W\ m^{-2}]$ for low vegetation.

- F_2 introduces the effect of the hydrous stress on R_s:

$$F_2 = \begin{cases} 1 & \text{if } W_p > W_{\mathrm{fc}} \\ \frac{W_p - W_{\mathrm{wilt}}}{W_{\mathrm{fc}} - W_{\mathrm{wilt}}} & \text{if } W_{\mathrm{wilt}} \leq W_p \leq W_{\mathrm{fc}} \\ 0 & \text{if } W_p < W_{\mathrm{wilt}} \end{cases}$$

- F_3 represents the effects of the atmospheric saturation deficit on R_s. Following Sellers et al. [1986]:

$$F_3 = 1 - \gamma(e_{\mathrm{sat}}(T_s) - e^L)$$

  where γ is an empirical parameter between 0 and 40 depending on plant species.

- F_4 measures the influence of the air temperature on R_s. Following DICKINSON [1984]:

$$F_4 = 1.0 - 1.6\ 10^{-3}(298.0 - T^L)^2$$

From this, the surface vapour flux is computed in ACDIFUS (see part III) using

$$J_q = -C_H|\mathbf{V}^L|(q^L - q_s) = F_{\text{evg}} + F_{\text{evv}}$$

Repartition of J_q between F_{evl}, F_{evn} and F_{evi} uses the fraction of the surface evaporating in ice phase: P_{ie} and the ice fraction in the surface reservoir P_{wsi}:

$$P_{wsi} = \begin{cases} \frac{W_{si}}{W_{si}+W_s} & \text{under LFGELS} \\ 0 & \text{otherwise} \end{cases}$$

$$P_{ie} \equiv \delta_{\text{land}}(P_{nc} + (1 - P_{nc})P_{wsi}) + \begin{cases} (1 - \delta_{\text{land}}) & \text{if TMERGL} > T_s \\ 0 & \text{otherwise} \end{cases}$$

where TMERGL is the melting temperature of floating ice.
The different water vapour fluxes are:

$$\begin{array}{lcll} F_{\text{evl}} & = & J_q(1 - P_{ie}) & \text{above water or moist ground} \\ F_{\text{evn}} & = & J_q P_{ie} & \text{above snow or ice} \\ F_{\text{evi}} & = & J_q(1 - P_{nc})P_{wsi} = J_q(P_{ie} - P_{nc}) & \text{above frozen ground (not covered by snow)} \end{array}$$

1.5.2 Heat fluxes

We must take into account:

- Q_R the net radiative energy flux at the surface, sum of the net solar flux $\overline{F^L_{\odot}}$ (taking into account the surface albedo α) and the net thermal flux $\overline{F^L_{\dagger}}$ (taking into account the surface emissivity ϵ).

- PFCS $\equiv Q_{\text{sens}}$ is the surface sensible heat flux (cf part III):

$$Q_{\text{sens}} = c_p C_H|\mathbf{V}^L|(T_s - T^L)$$

  where $\mathbf{V}^L$, T^L are air density, speed and temperature at the lowest model level, and C_H is an exchange coefficient depending on the thermal stability of the atmosphere.

- Latent heat fluxes PFCLL $\equiv Q_{w\text{lat}}$ over water or moist ground, and PFCLN $\equiv Q_{i\text{lat}}$ over ice or snow are computed from the moisture fluxes as subdivided by P_{ie}, using the vaporisation and sublimation latent heats respectively for the liquid and solid fractions.

- PFCHSP $\equiv F_{sp}$ computation was explained in §(1.3).

2 Practical implementation

2.1 ACSOL, ACSOLW: surface characteristics

Those two modules compute various coefficients required for the soil-vegetation scheme, namely

- Thermo-hydrous coefficients, from the reservoir contents and the soil composition
  - PWSEQ $\equiv W_{\text{seq}}$ the surface reservoir contents at the equilibrium between capillarity and gravity;
  - PWSMX $\equiv W_{s\max}$ and PWPMX $\equiv W_{p\max}$ the surface and total reservoir maximum contents;
  - PWFC $= w_{\text{fc}}$ the volumetric water contents at field capacity;
  - PWWILT $= w_{\text{wilt}}$ the volumetric water contents at wilt point.

- Thermal coefficients:
  - PCG $\equiv C_G$ for bare soil;
  - PCN $\equiv C_N$ for snow;
  - PCT $\equiv C_T$ for complete soil-snow-vegetation system.

- hydrous coefficients:
  - PC1 $\equiv C_1$ measuring the contribution of the surface fluxes to the surface water contents;
  - PC2 $\equiv C_2$ and PC2REF measuring the intensity of the transfers between the two reservoirs;
  - PC3 $\equiv C_3$ measuring the deep reservoir draining.

  On the sea, the three coefficients are set to 1.

- auxiliary coefficients: PNEIJG $\equiv P_{ng}$ the snow fraction on bare soil, PNEIJV $\equiv P_{nv}$ the snow fraction on vegetation.

2.2 ACVEG: vegetation-related computations

This routine computes the surface characteristics linked to vegetation:

- Surface resistance (equation (32))
- Maximum water contents of the interception reservoir (equation (14))
- Relative weights of air and saturation moisture in surface humidity, HU and HQ, with equations (27), (29)

2.3 ACDROV: water transfers in the soil-vegetation scheme

This routine computes water fluxes within the soil-vegetation system: runoffs, transfers between the three water reservoirs, and water flux linked to snow melting. It is called under key LSOLV, else it is replaced by ACDRO.
Reservoir contents evolute in regard to the associated fluxes; runoff is then computed by comparing the final contents W^+ to the maximal contents $W_{\max}$:

$$F_{\text{ro}} = \max(0, W + \left(\frac{\partial W}{\partial t}\right)^+ \triangle t - W_{\max})$$

but if $W^+ < 0$ we take a negative runoff to ensure non negative water contents.

Actual computation follows:

- Interception reservoir evolution.
  Interception reservoir runoff $F_{\text{ror}} \equiv$ PRUISL.

- Snow melting.
  If key LSNV is not set, snow melting occurs when a temporary temperature T_s^{+*} is bigger than the triple point temperature and the snow reservoir is not empty:

$$\begin{aligned} \mathsf{ZTSP} = T_z^{+*} &= T_s + C_T \triangle t_{\text{phys}} \delta_{\text{land}} \left\{ Q_R + Q_{\text{sens}} + Q_{w\text{lat}} + Q_{i\text{lat}} - F_{\text{sp}} \right\} \\ F_n \equiv \mathsf{PFONTE} &\equiv \triangle S_n = \frac{T_s^{+*} - T_t}{C_T L_{w-i}} \end{aligned}$$

  where F_{sp} is the heat transfer between surface and deep soil, PCT $\equiv C_T$ is the thermal coefficient of soil-vegetation and L_{w-i} the melting heat.
  If LSNV is set, the temperature of the vegetation may be positive (tree tops) while there is snow on the ground. Furthermore, bare soil not covered by snow may have a positive temperature.
  Snow temperature is taken equal to the deep soil temperature T_p, which is close to the diurnal mean mesh temperature:

$$T_n = (1 - \mathsf{veg}_a) T_s^{+*} + \mathsf{veg}_a T_p$$

  where veg_a is the "apparent" vegetation fraction, hence if all vegetation is covered by snow, $T_n = T_s^{+*}$.
  Melting concerns the snowy fraction of the mesh P_n and only the snow cover (thermal coefficient C_N):

$$F_n \equiv \mathsf{PFONTE} \equiv \triangle S_n = P_n \frac{T_n - T_t}{C_N L_{w-i}}$$

- Freezing computations
  Under key LFGEL freezing flux for deep reservoir PFGEL is computed; and under LFGELS a freezing flux PFGELS for the surface reservoir.
  PWPI $\equiv W_{pi}$ is the total frozen water contents, PWSI $\equiv W_{si}$ is the frozen water contents of the surface reservoir.
  If both keys are set, surface freezing writes:

$$
\begin{aligned}
T_s^+ &= T_s^{+*} + C_T \triangle t_{\text{phys}} \delta_{\text{land}} L_{w-i} \triangle S_n \\
\mathsf{ZZK} &= \delta_{\text{land}} \mathsf{GCGELS} \max(0, 1 - \frac{\mathsf{veg}}{\mathsf{veg}_{s\max}}) \max(0, 1 - \frac{\mathsf{lai}}{\mathsf{lai}_{s\max}}) \max(0, 1 - \frac{P_n}{P_{ns\max}}) \\
& \mathsf{veg}_{s\max} \equiv \mathsf{GVEGMXS},\ \mathsf{lai}_{s\max} \equiv \mathsf{GLAIMXS},\ P_{ns\max} \equiv \mathsf{GNEIMXS} \\
\mathsf{ZZ} &= \mathsf{ZZK} \frac{T_t - T_s^+}{C_{Ti} \cdot L_{w-i}} \\
\mathsf{ZFGELS} &= \begin{cases} \min(\frac{W_s}{W_{s\max}} \cdot \mathsf{ZZ}, W_s/\triangle t_{\text{phys}}) & \text{if } \mathsf{ZZ} > 0 \\ 0 & \text{otherwise} \end{cases} \\
\mathsf{ZFONTES} &= \begin{cases} -\min(|\mathsf{ZZ}|, \frac{W_{si}}{\triangle t_{\text{phys}}}) & \text{if } \mathsf{ZZ} < 0 \\ 0 & \text{otherwise} \end{cases} \\
F_{si} &\equiv \triangle W_{si} \equiv \mathsf{PFGELS} = \mathsf{ZFGELS} + \mathsf{ZFONTES}
\end{aligned}
$$

  (T_s^+ takes into account the snow melting computed above).
  Deep soil freezing writes:

$$
\begin{aligned}
\mathsf{ZZK} &= \delta_{\text{land}} \mathsf{GCGEL} \max(0, 1 - \frac{\mathsf{veg}}{\mathsf{veg}_{\max}}) \max(0, 1 - \frac{\mathsf{lai}}{\mathsf{lai}_{\max}}) \max(0, 1 - \frac{P_n}{P_{n\max}}) \\
& \mathsf{veg}_{\max} \equiv \mathsf{GVEGMX},\ \mathsf{lai}_{\max} \equiv \mathsf{GLAIMX},\ P_{n\max} \equiv \mathsf{GNEIMX} \\
T_p^+ &= T_p + C_T \triangle t_{\text{phys}} \delta_{\text{land}} F_{\text{sp}} \frac{\mathsf{SODELX(0)}}{\mathsf{SODELX(1)}} \\
\mathsf{ZZ} &= \mathsf{ZZK} \frac{T_t - T_p^+}{C_{Ti} L_{w-i}} \\
W_p^+ &= \max(W_p - W_s - W_{si}, 0) \\
\mathsf{ZFGEL} &= \begin{cases} \min(\frac{W_p^+}{W_{p\max}} \cdot \mathsf{ZZ}, \frac{W_p^+}{\triangle t_{\text{phys}}}, \frac{W_{pi\max} - W_{si} + W_{pi}}{\triangle t_{\text{phys}}}) & \text{if positive} \\ 0 & \text{otherwise} \end{cases} \\
\mathsf{ZFONTE} &= \begin{cases} -\min(|\mathsf{ZZ}|, \frac{W_{pi} - W_{si}}{\triangle t_{\text{phys}}}) & \text{if } \mathsf{ZZ} < 0 \text{ and } W_{pi} > W_{si} \\ 0 & \text{otherwise} \end{cases} \\
F_{pi} &\equiv \triangle W_{pi} \equiv \mathsf{PFGEL} = \mathsf{ZFGEL} + \mathsf{ZFONTE}
\end{aligned}
$$

  RCTGLA $\equiv C_{Ti}$ is the thermal coefficient for ice.

- Flux between surface and total (deep) reservoir:
  $F_{\text{sp}} \equiv \mathsf{PFLWSP} = \frac{C_2}{\tau}(W_s - W'_{\text{seq}})$, with $W'_{\text{seq}} = W_{\text{seq}} - W_{si}$ under key LFGELS.

- Surface reservoir evolution.
  The surface reservoir receives the part of the precipitation that was not intercepted by vegetation. See equation (15).
  Surface reservoir runoff is $F_{\text{ros}} \equiv$ PRUISS. Under key LFGELS, $W_{s\max}$ is replaced by $W'_{s\max} = W_{s\max} - W_{si}$ and W_{seq} by $W'_{\text{seq}} = W_{\text{seq}} - W_{si}$.

$$
\begin{aligned}
\triangle F_s &= \triangle t \frac{C_1 \{(1 - \mathsf{veg}_e)\mathcal{P}_w + F_{\text{ror}} + F_{\text{evw}} + F_{\text{evv}} + F_n\} - \frac{C_2}{\tau}(W_s - W'_{\text{seq}})}{1 + C_2 \frac{\triangle t \delta_{\text{land}}}{\tau}} - F_{si} \\
F_{\text{ros}} &= \frac{\delta_{\text{land}}}{\triangle t} \begin{cases} W_s + \triangle F_s - W'_{s\max} & \text{if } W_s + \triangle F_s > W'_{s\max} \\ 0 & \text{if } 0 < W_s + \triangle F_s < W'_{s\max} \\ W_s + \triangle F_s & \text{if } W_s + \triangle F_s < 0 \end{cases}
\end{aligned}
$$

- Total (deep) reservoir evolution (equation (16)).
  Deep reservoir runoff is $F_{\text{rop}} \equiv$ PRUISP. Under key LFGEL, $W_{p\max}$ is replaced by $W'_{p\max} = W_{p\max} - W_{pi}$

and field capacity $w_{\rm fc}d_2$ is replaced by $W'_{\rm fc} = w_{\rm fc}d_2 - W_{pi}$

$$\begin{aligned}
\triangle F_p &= \triangle t\left\{(1-\mathsf{veg}_e)\mathcal{P}_w + F_{\rm ror} + F_{\rm evw} + F_{\rm evv} + F_n + F_{tr} - F_{pi}\right\} \\
F_{\rm rop} &= C_3\frac{\delta_{\rm land}}{\tau}\max(0, W_p - W'_{\rm fc}) + \frac{\delta_{\rm land}}{\triangle t}\begin{cases} W_p + \triangle F_p - W'_{p\,\max} & \text{if } W_p + \triangle F_p > W'_{p\,\max} \\ 0 & \text{if } 0 < W_p + \triangle F_p < W'_{p\,\max} \\ W_p + \triangle F_p & \text{if } W_p + \triangle F_p < 0 \end{cases}
\end{aligned}$$

2.4 ACHMT, ACDIFUS: fluxes computations

ACHMT uses T_s, W_s, W_{si},$W_{\rm fc}$, Hv, P_{ng}, P_{nv}, veg_a, z_{0g}, z_{0f}, z_{0Hf}.
It computes:

- Surface moisture q_s and saturation moisture $q_{s\rm sat}$;
- the weight factors HU and HQ for saturation and air moistures terms in surface moisture;
- veg_a, P_n, z_0;
- other parameters and variables related to the exchanges with the surface (see part III).

ACDIFUS uses T_p, T_s, W_s, W_{si}, as well as the vegetation type iveg and fraction veg, C_T, HQ, HU, Htr, Hv, P_n, and the surface moisture q_s and saturation moisture $q_{s\rm sat}$.
Its ISBA part computes

- the different moisture fluxes: $F_{\rm evi}$, $F_{\rm evl}$, $F_{\rm evn}$, $F_{\rm evv}$, F_{tr};
- the surface sensible heat flux PFCS $\equiv Q_{\rm sens}$;
- the surface latent heat fluxes PFCLL $\equiv Q_{w\rm lat}$ and PFCLN $\equiv Q_{i\rm lat}$;
- the heat flux from the surface to the deep ground PFCHSP $\equiv F_{\rm sp}$;
- it adapts the roughness length z_{0f} over the sea, using Charnock's formula.

2.5 APLPAR: coordination of parameterization calls

Adaptation of the radiative parameters: albedo and emissivity (equations (25)) is performed at the level of APLPAR.

2.6 CPTENDS, CPWTS: tendencies and time step

CPWTS computes the evolution of ground and snow variables: T_s, T_p, W_s, W_{si}, W_p, W_{pi}, W_r, S_n, A_n, ρ_n.

CPTENDS computes tendencies for the former, using C_1, C_2, C_T, $Q_{\rm sens}$, $Q_{w\rm lat}$, $Q_{i\rm lat}$, $F_{\rm sp}$, $F_{\rm evi}$, $F_{\rm evl}$, $F_{\rm evn}$, $F_{\rm evv}$, $F_{\rm tr}$, $F_{\rm ror}$, $F_{\rm ros}$, $F_{\rm rop}$, veg_a, and also Q_R, $\mathcal{P}$, A_n, ρ_n, S_n.