SURFACE PARAMETERIZATION

History and fundamental principles

- Why do we need a surface parameterization in a NWP model ?

  - water cycle

  - heat exchange between soil and atmosphere

- Why type of surface ?

  - water (sea, lake, river)

  - ice, snow

  - bare ground, vegetation

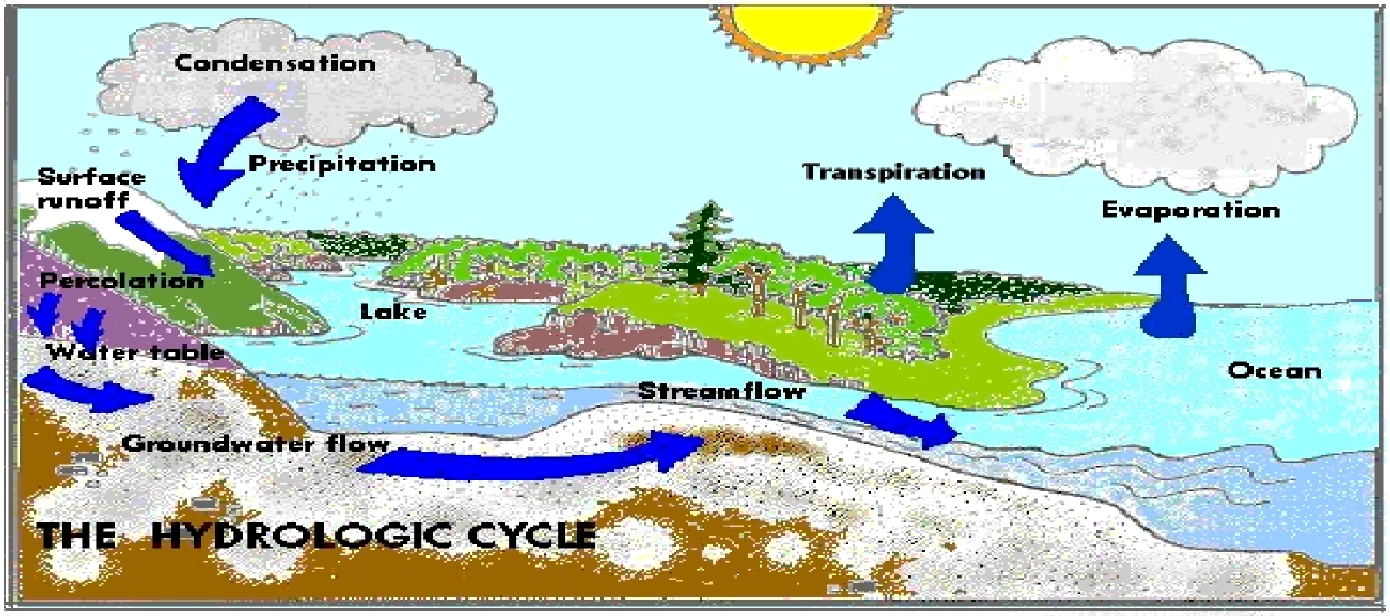
Water distribution:

De Marsily, 1995	Fraction (%)
Sea	97.4
Ice, snow, sea ice	1.99
Underground water	0.59
Lake and inland sea	0.015
Surface water	0.005
Water in atmosphere	-
River	-

Two water cycle:

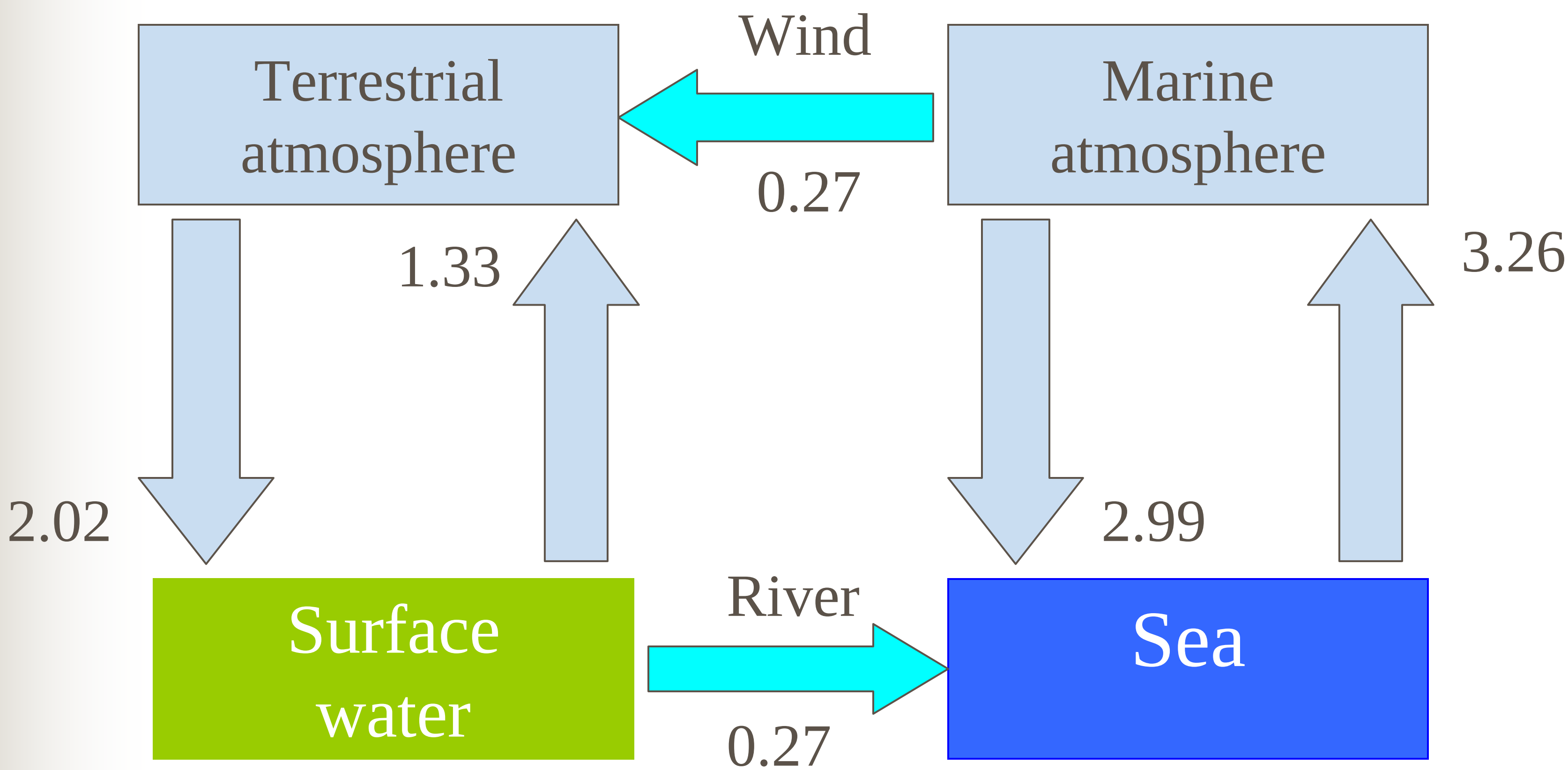
-atmospheric: more visible with clouds, rain.... The mean time of the water residence in the atmosphere is about 8 days.

- in the soil.



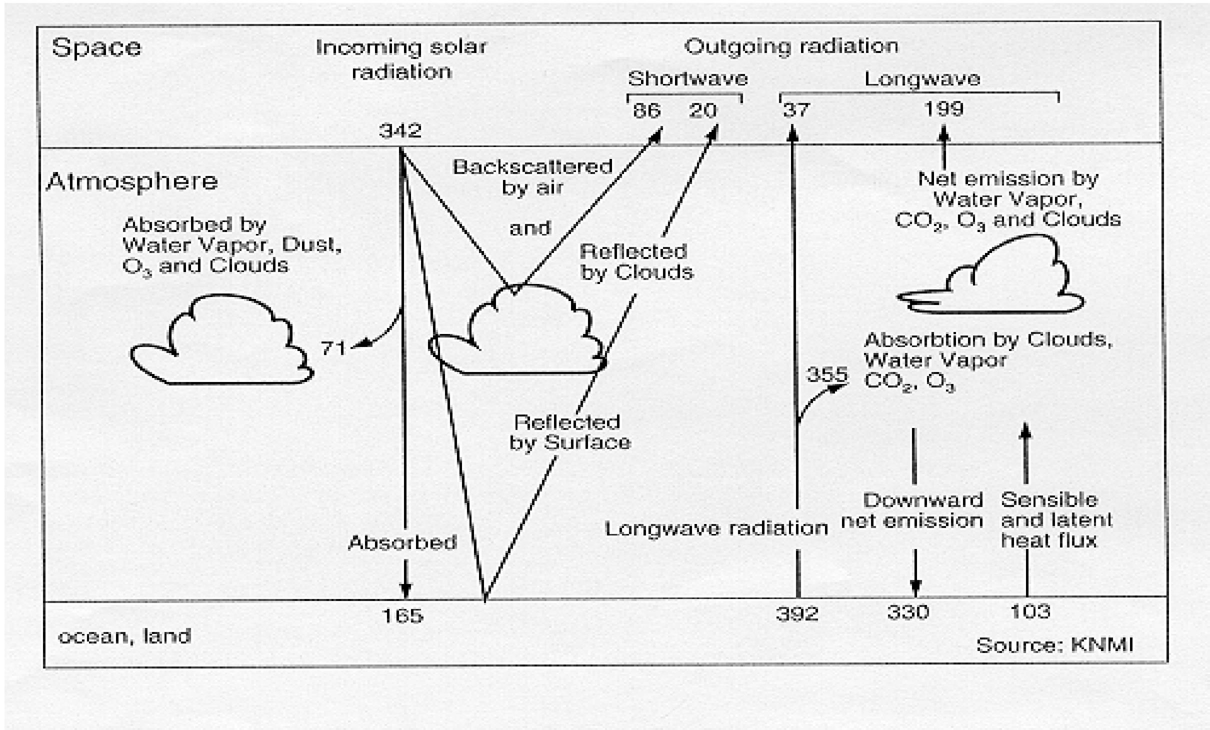
METEO FRANCE

Water Budget



Total evaporation and rain : 2.7 mm/day
from Arkin, 1994

Heat Budget



The heat budget at the earth's surface is apparently out of balance. Against the absorption of 165 W/m2 of solar radiation there is 392 W/m2 of outgoing infrared radiation and again 103 W/m2 through evaporation and sensible heat. The deficiency is compensated for by 330 W/m2 infrared radiation,which is emitted back to earth by gases and clouds.

History for the ground surface temperature

At the end of 1970, the ground surface temperature is usually computed by solving the heat balance equation at the surface with no heat capacity for the soil (Rn – H – LE = 0)

Delsol et al. (1971 QJRMS) include explicitly a soil heat flux (diffusion equation) : Rn – H – LE – G = 0 with G= soil heat flux.

Corby et al. (1972 QJRMS) proposes to compute the ground surface temperature by solving a prediction equation but neglects the conduction from below with a non-zero heat capacity for the soil.

$$c \frac{\partial T_g}{\partial t} = R_n - LE - H$$

Bhumralkar (1975,JAM) and Blackadar (1976,AMS) propose the force restore method = Corby + G

TABLE 1. Formulations of soil heat flux G used in Eq. (3).

Experiment	Formula for soil heat flux G	Source
Control	$G=0$	Gates *et al.* (1971)
1	$G=\frac{\lambda}{D}(T_g-T_D)$	Estoque (1963), (L); Pandolfo *et al.* (1965), (L); Myrup (1969), (L); Delsol *et al.* (1971), (GCM); Sasamori (1970), (L)
2	$G=\mu R_N$	Gadd and Keers (1970), (GCM)
3	$G=0.3H$ and $LE=H$	Kasahara and Washington (1971), (GCM)

Note: In source column, L=local atmospheric and soil boundary layer models; GCM=atmospheric general circulation models.

from Bhumralkar (1975)

TABLE 1. Methods of Calculating T_g

Method	Designation of T_g	Description	Reference
Multiple (12) soil layers	T_{gm}	Finite difference solution of diffusion equation for $T(z)$; $G=-\lambda(\partial T/\partial z)_0$	*Carslaw and Jaeger* [1959]; *Benoit* [1976]
Insulated surface	T_{gi}	$G=0$	*Gates et al.* [1971]; *Manabe et al.* [1974]
H_{sg} dependence	T_{gs}	$G=1/3H_{sg}$	*Kasahara and Washington* [1971]
R_{net} dependence	T_{gr}	$G=-0.19R_{net}$ $R_{net}<0$ (down); $G=-0.32R_{net}$ $R_{net}>0$ (up)	*Nickerson and Smiley* [1975]
H_A forcing	T_{gf}	$\partial T_g/\partial t=-\pi^{1/2}H_A/(\rho_s c_s d_1)$	*Arakawa* [1972]; *Corby et al.* [1972]; *Rowntree* [1975]
Force restore rate equation	T_{gfr}	$\partial T_g/\partial t=-2\pi^{1/2}H_A/(\rho_s c_s d_1)-(2\pi/\tau_1)(T_g-T_2)$	*Bhumralkar* [1975], *Blackadar* [1976]

from Deardorff (1978)

Heat conduction equation: $$\frac{\partial T}{\partial t}=-\frac{1}{c}\frac{\partial}{\partial z}\left(-\lambda\frac{\partial T}{\partial z}\right)\quad(1)$$

λ($Wm^{-1}K^{-1}$) = thermal conductivity , c ($Jm^{-3}K^{-1}$) = volumetric heat capacity

We assume that surface temperature T_g can be written as:

$$T_g(0,t)=T_m+\Delta T_0\ \sin(\omega t) \text{ with } \omega=2\pi/\tau$$

The solution of (1) is: $T(z,t)=T_m+\Delta T_0\, e^{-z/d}\sin(\omega t-z/d)$ (2)

with $d=(2\lambda/c\omega)^{0.5}$ is the depth at which the amplitude of ΔT_0 is significant

The soil heat flux reads: $G(z,t) = -\lambda \frac{\partial T}{\partial z}$ (3).
Combining (2) and (3) we obtain:

$$G(z,t) = \sqrt{\frac{\omega c \lambda}{2}} \cdot \left(\frac{1}{\omega} \frac{\partial T(z,t)}{\partial t} + T(z,t) - T_m \right) \qquad (4)$$

with $G(0,t) = R_n - LE - H = \Delta T_0 (\lambda c \omega)^{0.5} e^{-z/d} \sin(\omega t - z/d + \pi/4)$
If we consider a soil layer from the surface (z=0) to a depth z:

$$c \frac{\partial T(z,t)}{\partial t} = -\left(\frac{G(z,t) - G(0,t)}{z} \right) \qquad (6)$$

Applying in (6) to a soil layer of 1cm and assuming that the average T for this layer is the ground surface temperature $T_g = T(0.01,t)$:

$$c^* \frac{\partial T_g}{\partial t} = Rn - LE - H - \sqrt{\frac{\omega c \lambda}{2}} \cdot (T_g - T_m) \qquad c^* = c \cdot z + \sqrt{\frac{\lambda c}{2\omega}} \approx \sqrt{\frac{\lambda c}{2\omega}}$$

$$\Rightarrow \frac{\partial T_g}{\partial t} = 2\sqrt{\frac{\pi}{\tau \lambda c}} \cdot (Rn - LE - H) - \frac{2\pi}{\tau} (T_g - T_m)$$

k_s = soil thermal diffusivity (m^2s^{-1}) = λ/c and $d_1=(k_s\tau)^{0.5}$

$$\Rightarrow \frac{\partial T_g}{\partial t} = \frac{2\sqrt{\pi}}{c \cdot d_1} \cdot \left(Rn - LE - H\right) - \frac{2\pi}{\tau}\left(T_g - T_m\right)$$

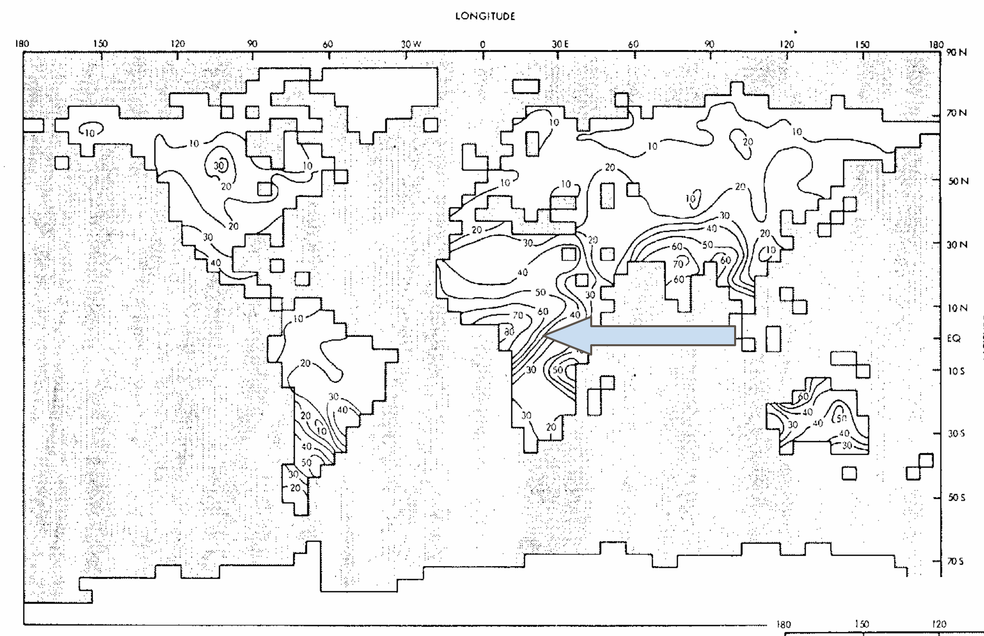


FIG. 4. Diurnal range of ground surface temperature (°C): Control Experiment (

From Bhumralkar (1975): « the diurnal oscillation values for the control experiment (CE) are rather unrealistic… observations (Sinclair, 1922) have shown a diurnal range of 56°C in the extreme, whereas the CE shows a diurnal range in excess of 70°C over wide areas of tropical regions »

From Bhumralkar (1975) with the soil heat flux.

One interesting concluding remark is: « This method (force restore) may not be applicable to surfaces which are covered with ice/or snow » !

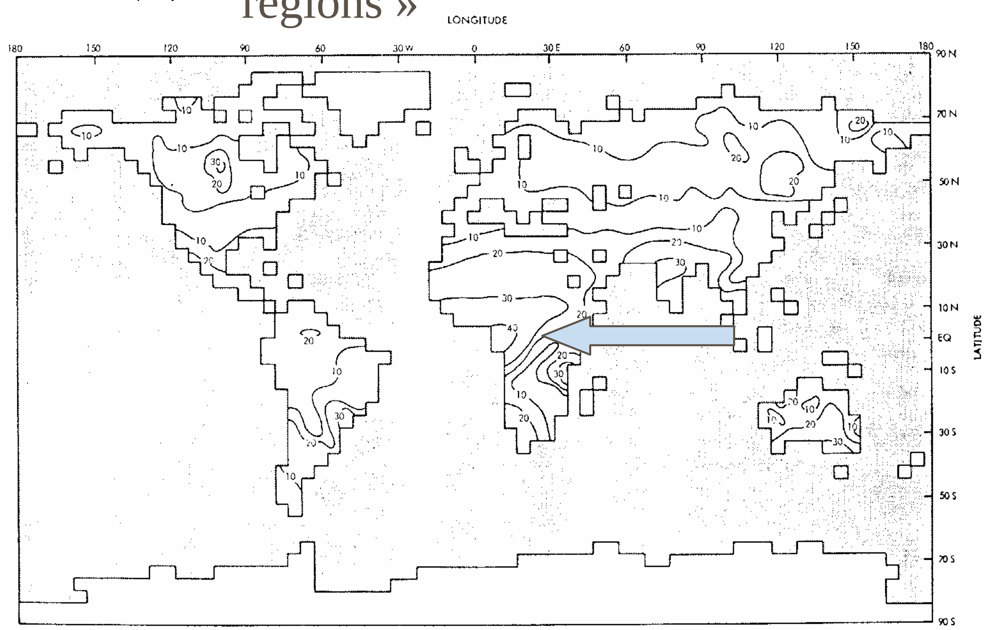


FIG. 13. Diurnal range of ground surface temperature (°C): Experiment 4 (January).

Ground surface moisture

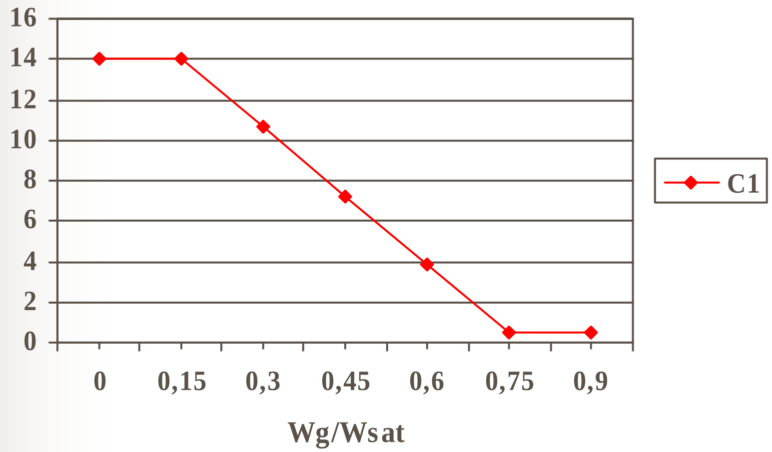
- NWP treats only the bulk soil moisture (Manabe 1969). The evaporation rate is then taken to be a fraction of the potential evaporation rate according to Wg/Wsat.
- Jackson (1973) shows how large the diurnal variation of surface soil moisture for bare soil is.
- Naturally the surface evaporation is closely related to Wg and not directly to W_2.
- Deardorff (1977 JAM) propose a soil-surface moisture parameterization starting from the same ideas as described by Bhumralkar for T, the so-called « force-restore » model.

$$\frac{\partial W_g}{\partial t} = -C_1 \cdot \frac{(E_g - P)}{\rho_w \cdot d_1} - C_2 \cdot \frac{(w_g - w_2)}{\tau}$$

$$\frac{\partial W_2}{\partial t} = -\frac{(E_g - P)}{\rho_w \cdot d_2}$$

With Eg= evaporation rate, P=rain

d1=10cm, d2=50cm, C1 and C2 are dimensionless constants



Eg= Wg/Wfc* Epot
Wfc=0.75*Wsat
Epot = Evaporation rate obtained
if the soil were coated with water

Deardorff 1978 (Journal of Geophysical Research)

- « This method (force-restore) appears even more promising and is still much more efficient than the use of multiple layers » (for two-day forecast with only one phase for water)
- A simple parameterization for a vegetation layer (with no heat capacity) has been developed: Tg, T2, Wg, W2, Wleaf, Tleaf, Qleaf
- He defines σ as a factor associated with the degree to which the foliage prevents SW radiation from reaching the ground = combination of vegetation cover and LAI.
- σ could be a function of season, latitude and soil moisture!

Deardorff 1978 (Journal of Geophysical Research)

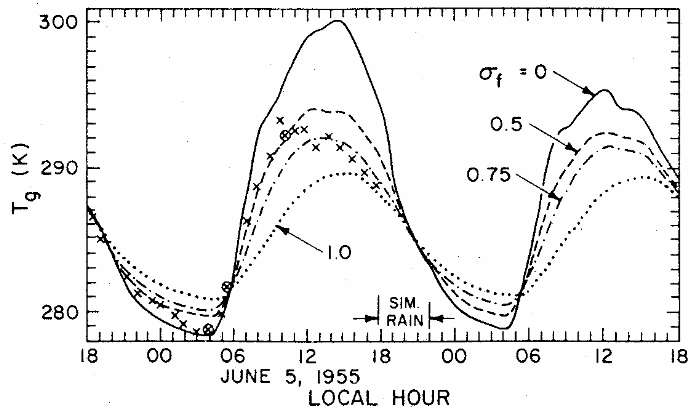


Fig. 3. Variation of the ground surface temperature calculated by the force restore method and vegetation parameterization over a 2-day period with atmospheric forcing as on June 4–5, 1955, from *Penman and Long* [1960]. Results for four different shielding factors, σ_f, are shown, along with extrapolated observed values denoted by crosses.

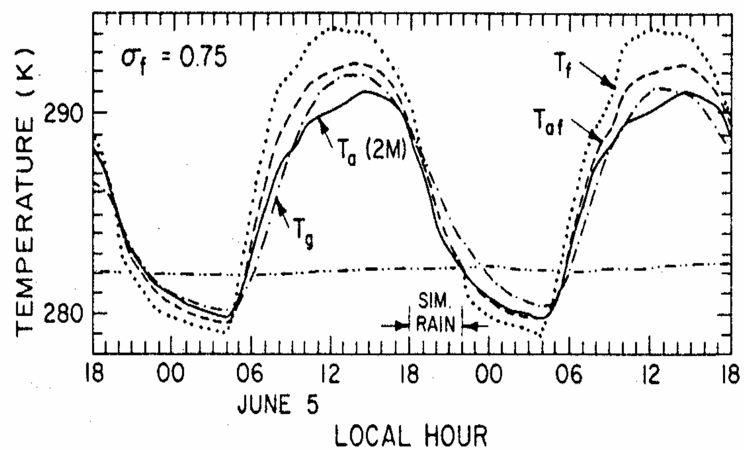


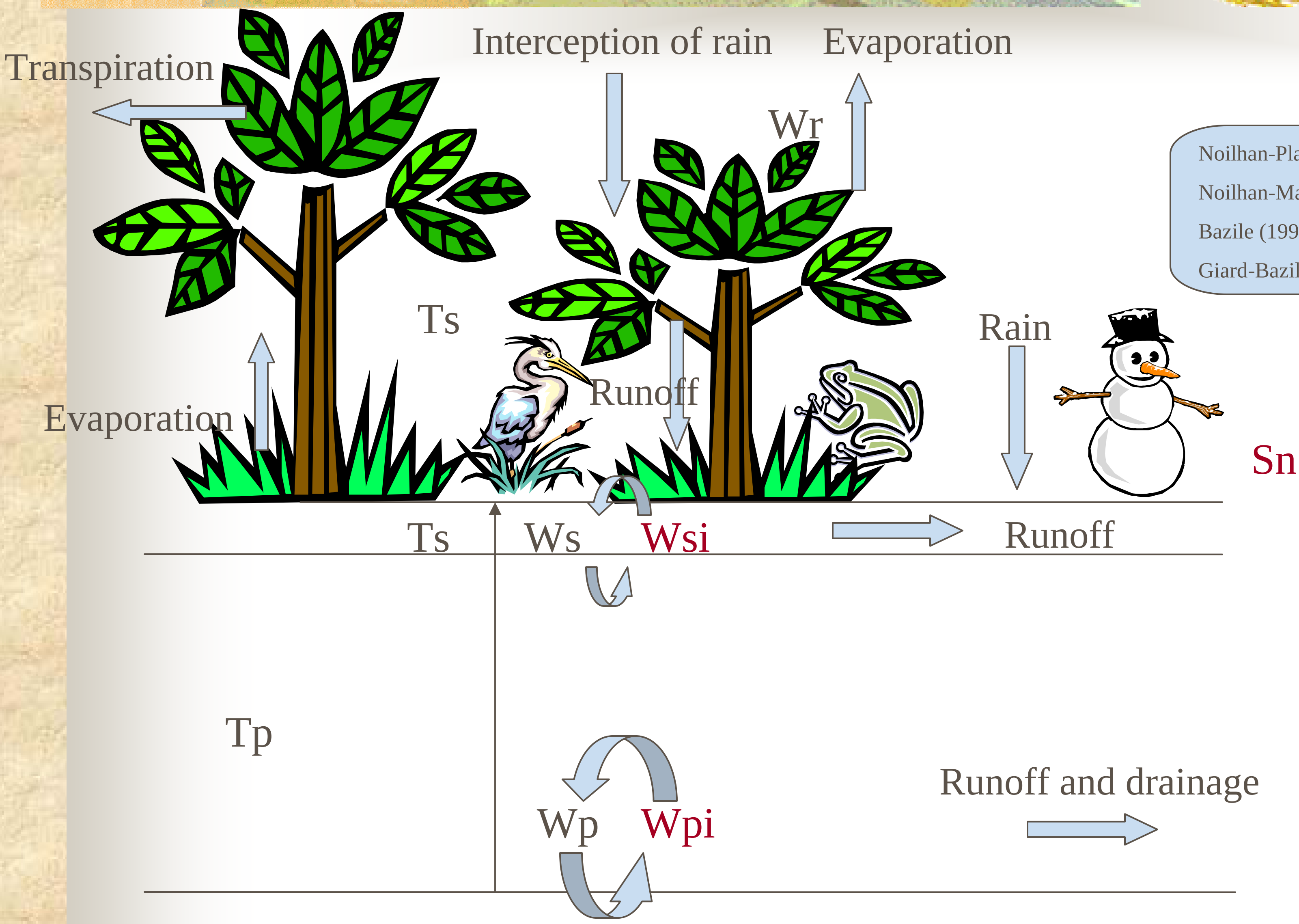
Fig. 4. Variation of the predicted temperatures T_f, T_{af}, T_g, and of T_a over the 2-day period for a shielding factor of 0.75.

Evaporation from the vegetation takes into account the direct evaporation from Wleaf and the transpiration.
Charney et al (1975) point out the need for including a model of the biosphere within the atmospheric model.

ISBA=Interaction Soil Biosphere Atmosphere

Noilhan & Planton 1989 MWR

- For Tg and T_2 it is the force restore approach proposed by Bhumralkar (75)
- For Wg and W2 it is Deardorff (77)
- The limits of this approach are specified by NP: « …is limited to the case of short range simulations…excluding the case of frozen soils »
- The improvement of ISBA compared to Deardorff is the use of the hydraulic properties from Clapp and Hornberger (78) and the reduction of the number of parameters.
- Mahfouf and Noilhan (1995, JAM) include the gravitational drainage
- The most up-to-date reference paper for ISBA is Noilhan and Mahfouf (1996) but the operational version of ISBA differs on several points: thermal inertia coeff., hydric coefficient for very dry soils and soil water freezing (Giard and Bazile (2000, MWR) , Bazile (1999, Hirlam NL)).



Noilhan-Planton (89,MWR)
Noilhan-Mahfouf (96, GPC13)
Bazile (1999, Hirlam NL 33)
Giard-Bazile (2000,MWR)

ECMWF scheme Viterbo & Beljaars 1995

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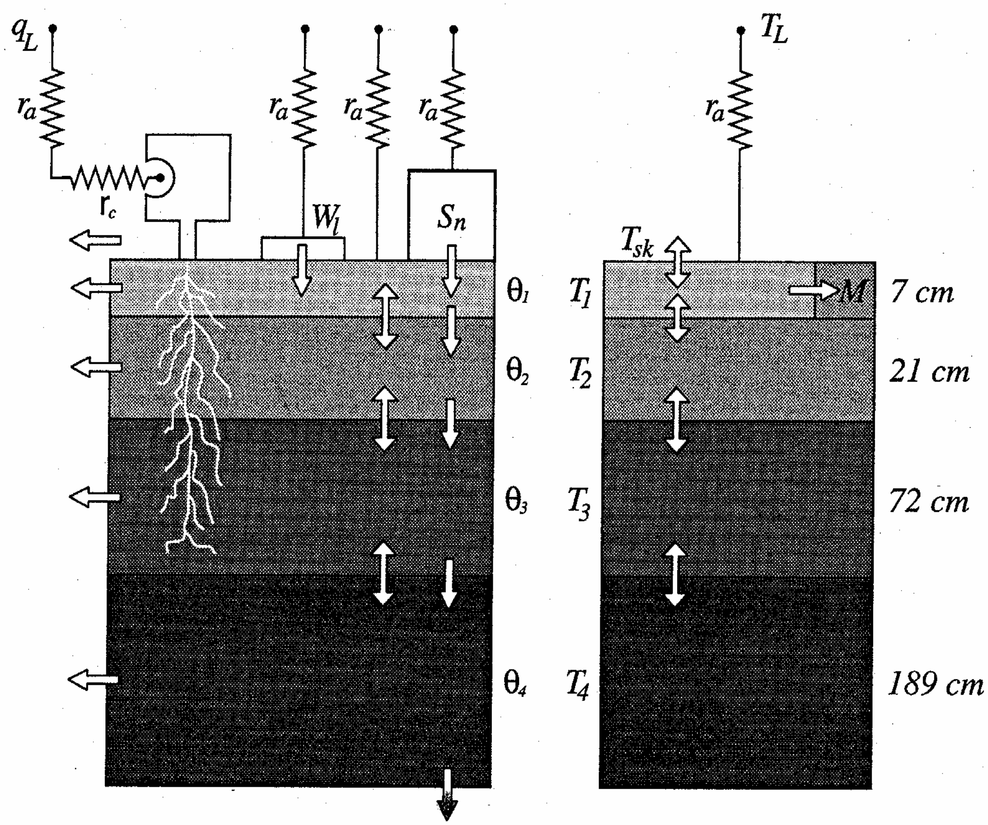


FIG. 1. Schematic description of the structure of the land surface model. Double arrows mean diffusivity processes, single arrows represent "drainagelike" terms (soil drainage, snow melting, and throughfall/top infiltration for the skin reservoir), horizontal arrows represent surface and subsurface runoff (bottom drainage is lost to the model and is therefore a runoff term). The bottom value of the resistance network for evaporation is $q_{sat}(T_{sk})$, except for the bare ground, where a relative humidity α is assumed [see Eq. (19)]. In the heat transfer panel, the snow mass replaces a portion of the first model layer (M), and the horizontal arrow represents heat exchanges due to melting.

TABLE 1. Parameters in the land surface scheme.

D_1	Depth of soil layer 1	0.07 m
D_2	Depth of soil layer 2	0.21 m
D_3	Depth of soil layer 3	0.72 m
D_4	Depth of soil layer 4	1.89 m
R_1	Fraction of roots in layer 1	0.33
R_2	Fraction of roots in layer 2	0.33
R_3	Fraction of roots in layer 3	0.33
R_4	Fraction of roots in layer 4	0
$(\rho C)_s$	Volumetric soil heat capacity	$2.19\ 10^6$ J m^{-3} K^{-1}
θ_{sat}	Soil moisture at saturation	0.472 m^3 m^{-3}
θ_{cap}	Soil moisture at field capacity	0.323 m^3 m^{-3}
θ_{pwp}	Soil moisture at permanent wilting point	0.171 m^3 m^{-3}
Ψ_{sat}	Matric potential at saturation	−0.338 m
γ_{sat}	Hydraulic conductivity at saturation	$4.57\ 10^{-4}$ m s^{-1}
a	Clapp and Hornberger soil parameter	3.8
b	Clapp and Hornberger soil parameter	6.04
L_f	Leaf area index	4
ϵ	Surface emissivity	0.996
W_{max}	Maximum water amount on single leaf	0.0002 m
r_{smin}	Minimum stomatal resistance of single leaf	240 s m^{-1}
Λ_{sk}	Skin layer "conductivity"	7 W m^{-2} K^{-1}
k	Heterogeneity factor for convective precipitation	0.5
c_i	Interception efficiency	0.25

The soil water freezing

- The soil moisture freezing and thawing plays an important role in the thermal budget of large continental areas.
- Viterbo et al (1998 Tech. Mem. N°255) introduce this effect on ECMWF scheme with a diagnostic approach:

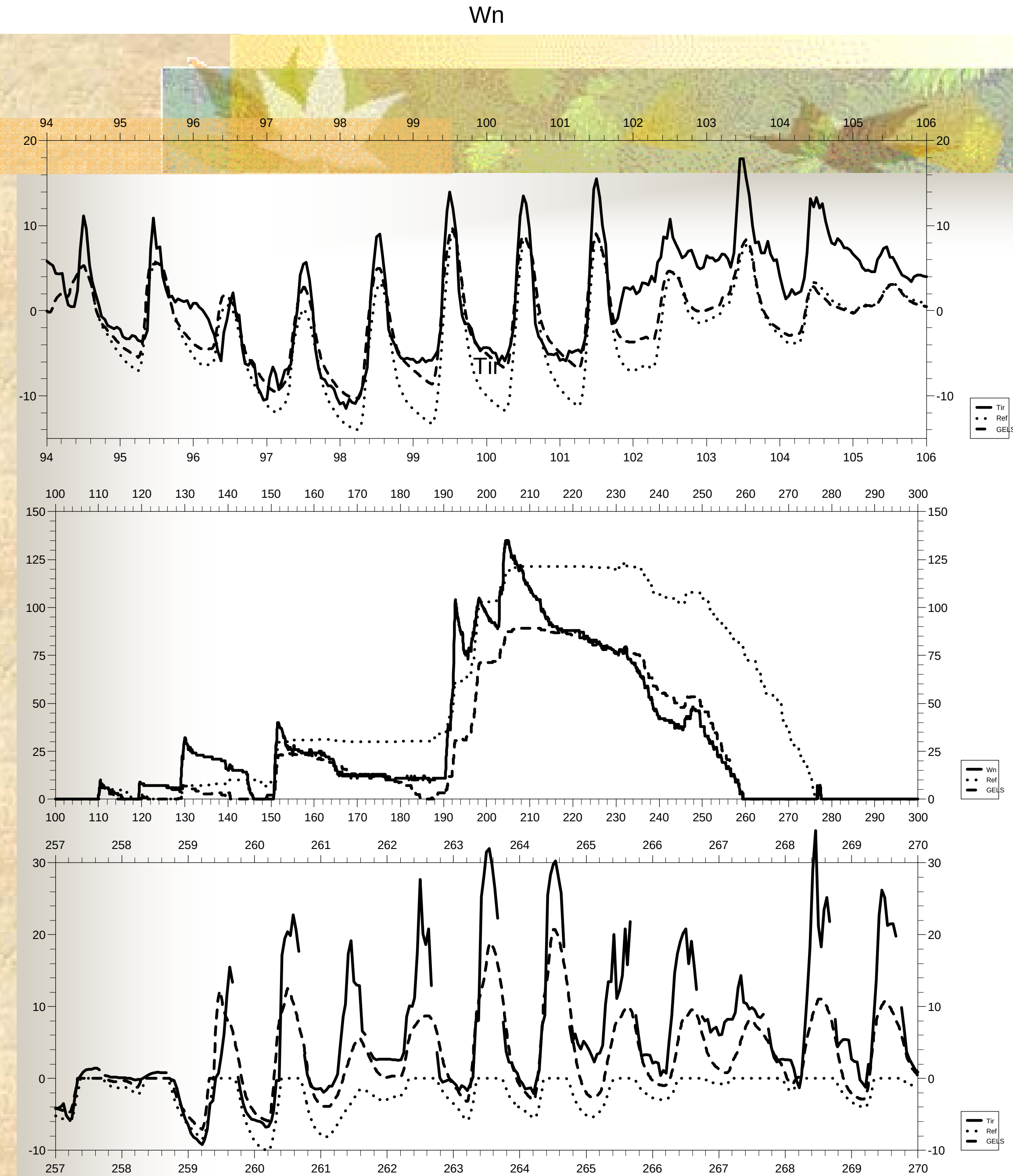
$$C\,\frac{\partial T}{\partial t} = \lambda\,\frac{\partial}{\partial t}\left(\frac{\partial T}{\partial z}\right) + L_f\,\rho_w\,\frac{\partial \theta_i}{\partial t} \quad \text{With } \theta_i = f(T)\,\theta_f$$

$$f(T) = 0.5[1 - \sin(\pi(T - 0.5T_1 - 0.5T_2)/(T_2 - T_1))]$$

f(T)=0 if T > 1°C, f(T)=1 if T<-3°C

The soil water freezing

- It is not realistic to use this approach in a force restore scheme with two layers because we need to keep the memory of the freezing period during a long time.
- The ISBA scheme has been modified with the introduction of two new prognostic variables: superficial frozen water Wsi and the total frozen water Wpi (Bazile 1999, Hirlam NL n33).
- This scheme improves Ts but also has a strong impact on the date of the disappearance of snow in spring (the deep soil temperature is warmer with this scheme).



Experiments with the 1d model on the Col de Porte site (Alps).

During the first days of November 95 we have a cold period without snow cover and Tg has a cold bias during night. The impact of the soil moisture freezing on the snow height is very important and the date of the snow disappearance is really improved due to a warmer Ts during the winter. This change has clearly a strong impact on Ts because the snow melting prevents the increase of Ts in the daytime.

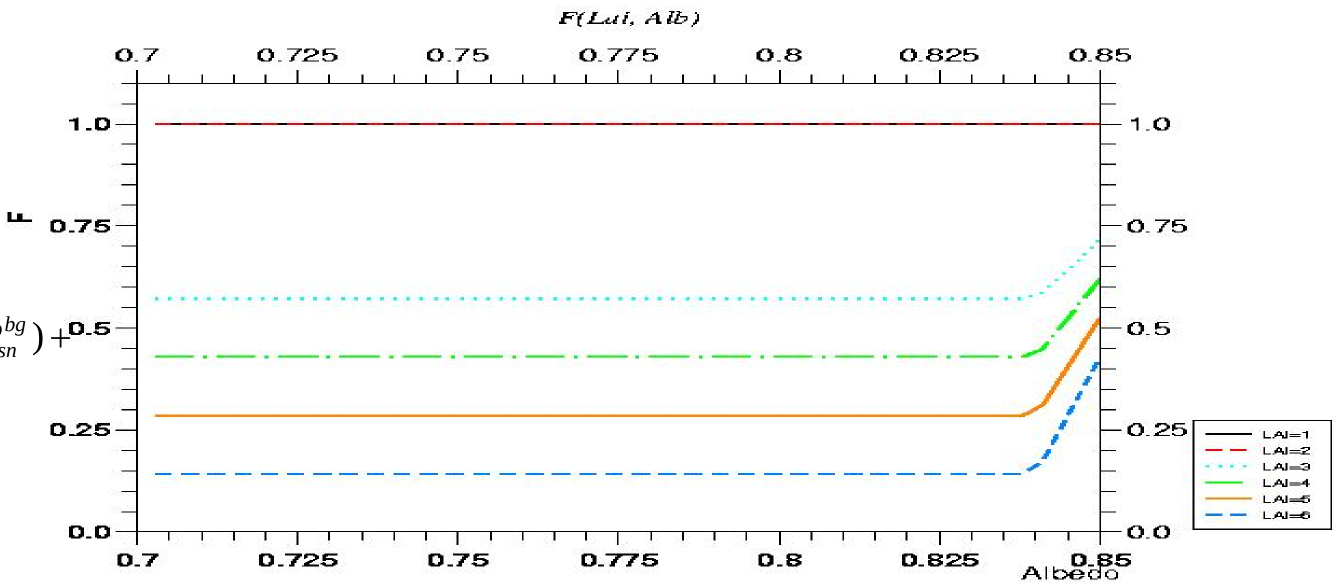
Full line: Observed, dotted line: without, dashed line with soil freezing.

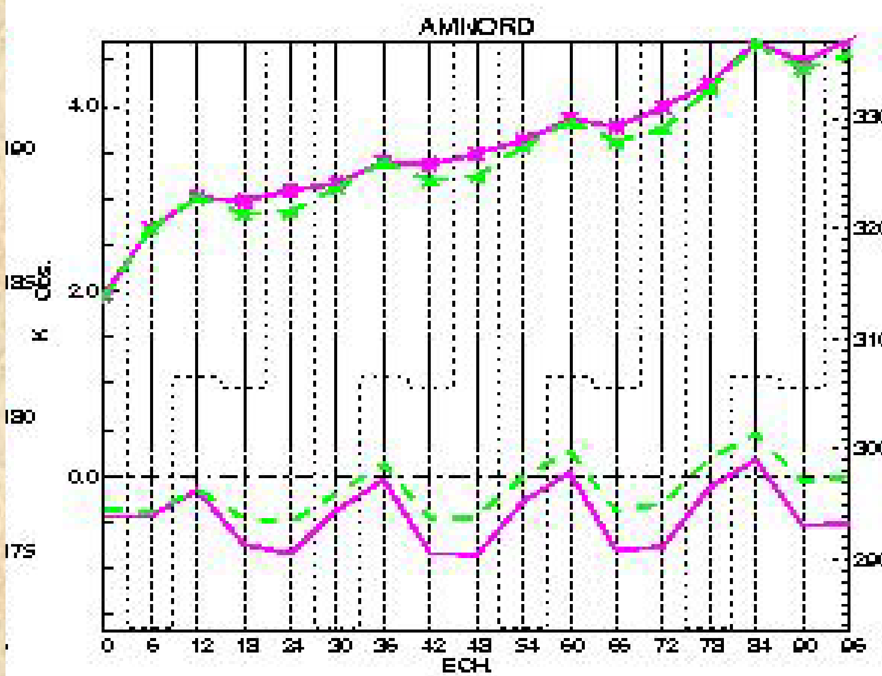
Snow parameterization

- The high reflectivity of snow can increase the surface albedo by as much as 60% . The snow albedo can vary between 0.3 and 0.9
- Fresh snow acts as a thermal insulator.
- Turbulent fluxes are reduced: static stability is increased and z0 decreases.
- Crocus (Brun et al (89, 92)): snow model for operational avalanche forecasting, considered as a reference to compare simple parameterizations !
- Douville et al (1995): density and albedo of the snow are prognostic. Two fractions of snow: on the vegetation part and on bare ground. Already coded in ARPEGE, used in the climate version. Changes required for NWP.
- Fernandez scheme (1998): a combination of Kondo and Yamazaki (1990, JAM) and Tarboton and Luce(1996,Int. Note); evolution of the freezing depth and Ts + snow's liquid water content.

$$P_{sn}^{veg} = F(LAI, \alpha_{sn}) \cdot P_{sn}^{bg}$$

$$\alpha = (1-veg)(\alpha_{bg}(1-P_{sn}^{bg}) + \alpha_{sn}P_{sn}^{bg}) + veg \cdot (\alpha_{veg}(1-P_{sn}^{veg}) + \alpha_{sn}P_{sn}^{veg})$$





SNOW ALBEDO (Douville et al 95)

Day=1 no melting, 2-5 melting, 6 0.5mm/day, 7 no melting, 8-10 melting