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*Summer Colloquium on the Physics of Weather and Climate*

## Workshop on
## Land-Atmosphere Interactions in Climate Models

*(28 May - 8 June 2001)*

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# Land-surface Modeling of Intermediate Complexity for NWP and Climate: the ECMWF Experience

## Lecture 3

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These are preliminary lecture notes, intended only for distribution to participants

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# Land-surface modelling of intermediate complexity for NWP and climate: the ECMWF experience

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ICTP, May 2001

# Layout

- Introduction
- General remarks
- Model development and validation
- The surface energy budget
- Soil heat transfer
- Soil water transfer
- Surface fluxes
- Initial conditions
- Snow
- Conclusions and a look ahead

Lecture 1

Lecture 2

Lecture 3

# Schematics

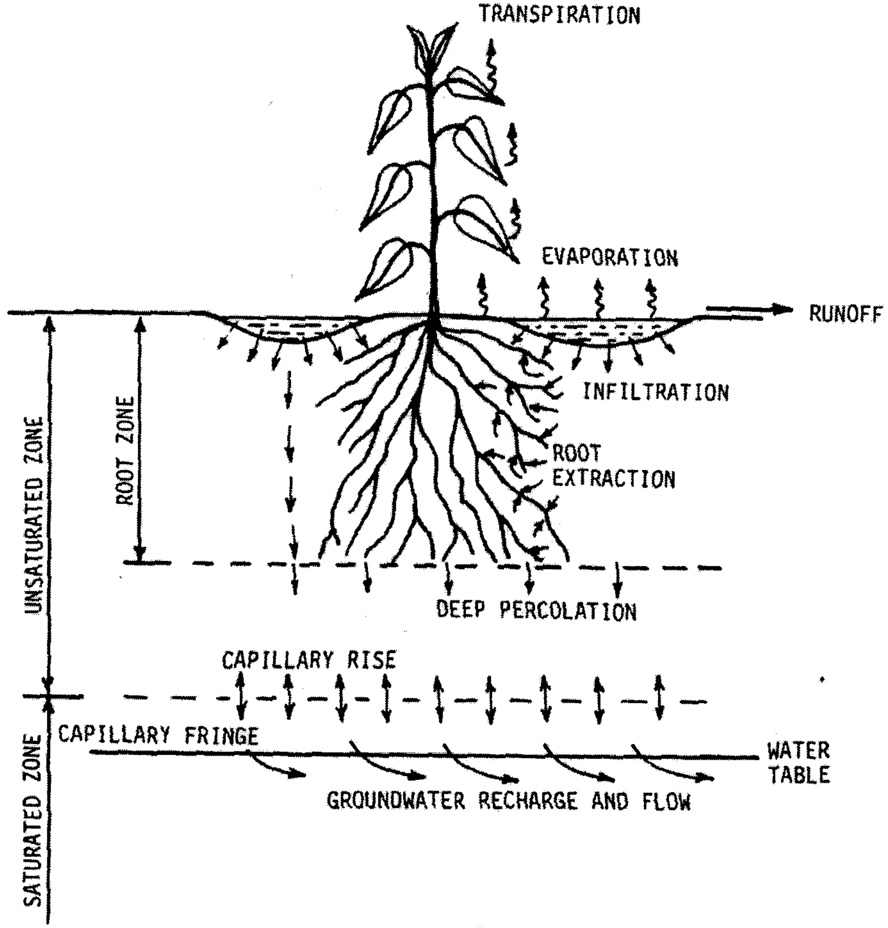



**Fig. 17.1.** The water balance of a root zone (schematic).

**Hillel 1982**

$$\rho_w \frac{f\theta}{ft} = -\frac{fF}{fz} + \rho_w S_\theta$$

$\theta$ soil water $[\;] = m^3 m^{-3}$

$F$ Soil water flux $[\;] = kgm^{-2}s^{-1}$

$S_\theta$ Soil water source/sink, ie root extraction

**Boundary conditions:**

**Top** **See later**

**Bottom** **Free drainage or bed rock**

**Root extraction**

**The amount of water transported from the root system up to the stomata (due to the difference in the osmotic pressure) and then available for transpiration**

# Soil water flux

$$F = -\rho_w \left(\lambda \frac{f\theta}{fz} - \gamma\right)$$

**Darcy's law**

$\lambda$ hydraulic diffusivity $[\lambda] = m^2 s^{-1}$

$\gamma$ hydraulic conductivity $[\gamma] = m s^{-1}$

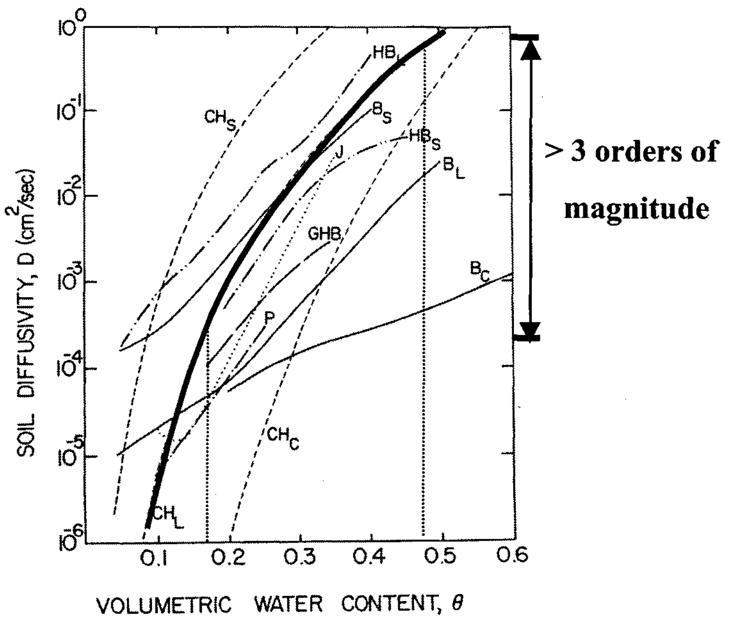



Fig. 2. Examples of the dependence of soil hydraulic diffusivity on volumetric soil water content for loam ($HB_L$, Hanks and Bowers, 1962); (J, Jackson, 1973); (GHB, Gardner *et al.*, 1970); silt loam ($HB_S$, Hanks and Bowers, 1962); clay (P, Passioura and Cowan, 1968); results approximated from Gardner (1960) for sand ($B_S$), loam ($B_L$), and clay ($B_C$); relationship from Clapp and Hornberger (1978) for sand ($CH_S$), loam ($CH_L$), and clay ($CH_C$).

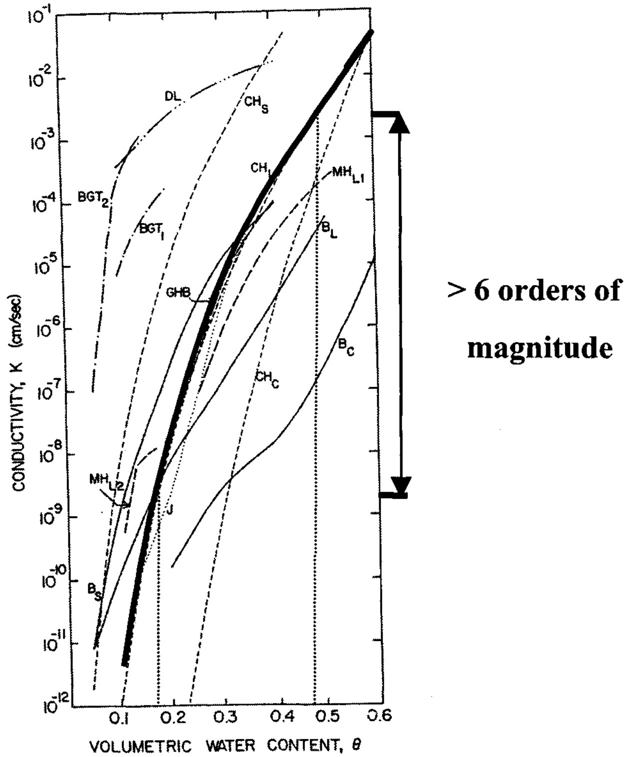



Fig. 3. Examples of the dependence of hydraulic conductivity on volumetric soil water content for sand (DL, Day and Luthin, 1956); (Black *et al.*, 1970, 0–50 cm-$BGT_1$, 50–150 cm-$BGT_2$); loam (J, Jackson, 1973); ($MH_{L1}$ and $MH_{L2}$, Marshall and Holmes, 1979); (GHB, Gardner *et al.*, 1970); results approximated from Gardner (1960) for sand ($B_S$), loam ($B_L$), and clay ($B_C$); relationship from Clapp and Hornberger (1978) for sand ($CH_S$), loam ($CH_L$), and clay ($CH_C$).

**Mahrt and Pan 1984**

# More soil science miscellany

Hillel 1982

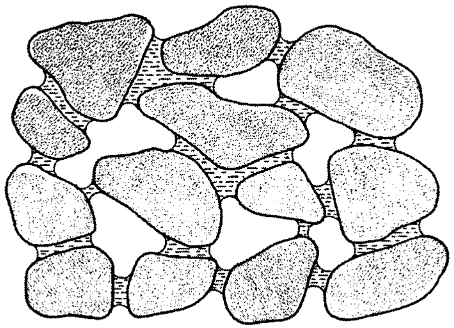


Fig. 7.1. Water in an unsaturated coarse-textured soil.

TABLE I **Jacquemin and Noilhan 1990**

Critical water contents of soils derived from the classification of Clapp and Hornberger (1978): saturated moisture $w_{sat}$, field capacity $w_{fl}$, wilting point $w_{wilt}$. The field capacity is associated with a hydric conductivity of 0.1 mm/day. The wilting point corresponds to a moisture potential of −15 bar

| Soil type | $w_{sat}$ ($m^3/m^3$) | $w_{fc}$ ($m^3/m^3$) | $w_{wilt}$ ($m^3/m^3$) |
|---|---|---|---|
| Sand | 0.395 | 0.135 | 0.068 |
| Loamy sand | 0.410 | 0.150 | 0.075 |
| Sandy loam | 0.435 | 0.195 | 0.114 |
| Silt loam | 0.485 | 0.255 | 0.179 |
| Loam | 0.451 | 0.240 | 0.155 |
| Sandy clay loam | 0.420 | 0.255 | 0.175 |
| Silty clay loam | 0.477 | 0.322 | 0.218 |
| Clay loam | 0.476 | 0.325 | 0.250 |
| Sandy clay | 0.426 | 0.310 | 0.219 |
| Silty clay | 0.482 | 0.370 | 0.283 |
| Clay | 0.482 | 0.367 | 0.286 |

- **3 numbers defining soil water properties**
  - **Saturation (soil porosity) Maximum amount of water that the soil can hold when all pores are filled** **0.472 $m^3m^{-3}$**
  - **Field capacity "Maximum amount of water an entire column of soil can hold against gravity"** **0.323 $m^3m^{-3}$**
  - **Permanent wilting point Limiting value below which the plant system cannot extract any water** **0.171 $m^3m^{-3}$**

# TESSEL: soil water budget

- Solution of Richards equation on the same grid as for energy
- Clapp and Hornberger (1978) diffusivity and conductivity dependent on soil liquid water
- Free drainage bottom boundary condition
- Surface runoff, but no subgrid-scale variability; It is based on infiltration limit at the top
- One soil type for the whole globe: “loam”

# TESSEL soil water equations (1)

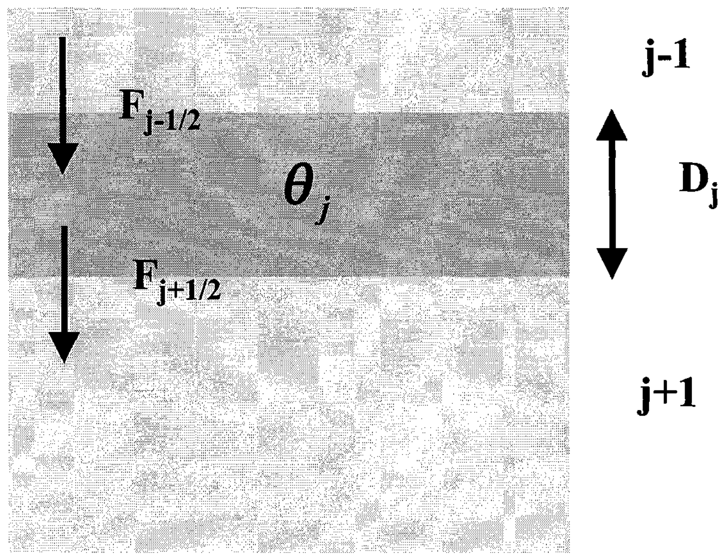
$$\rho_w \frac{\left(\theta_j^{n+1} - \theta_j^{n}\right)}{\Delta t} = -\frac{\left(F_{j+1/2}^{n+1} - F_{j-1/2}^{n+1}\right)}{D_j} + \rho_w S_{\theta,j} \qquad j = 1,\ldots,4$$

$$F_{j+1/2}^{n+1} = -\rho_w \overline{\lambda}_{j+1/2} \frac{\theta_{j+1}^{n+1} - \theta_j^{n+1}}{0.5\left(D_j + D_{j+1}\right)} - \gamma_{j+1/2}$$

Boundary conditions

$$F_{1/2} = T - Y_s + E_{1/2}$$

$$F_{41/2} = \rho_w \gamma_{41/2}$$



# TESSEL soil water equations (2)

Root extraction at layer j, separate for high/low (H/L) vegetation

$$\left[\rho_w S_{\theta,j}\right]_{H/L} = C_{H/L} \frac{R_{j,H/L}\theta_{j,liq}D_j}{\sum_j R_{j,H/L}\theta_{j,liq}D_j} \qquad j=1,\ldots,4$$

Hydraulic coefficients

$$\gamma = \gamma_{sat}\left(\frac{\theta}{\theta_{sat}}\right)^{2b+3}$$

$$\lambda = \frac{b\gamma_{sat}(-\psi_{sat})}{\theta_{sat}}\left(\frac{\theta}{\theta_{sat}}\right)^{b+2}$$

# FIFE: Time evolution of soil moisture

Viterbo and Beljaars 1995

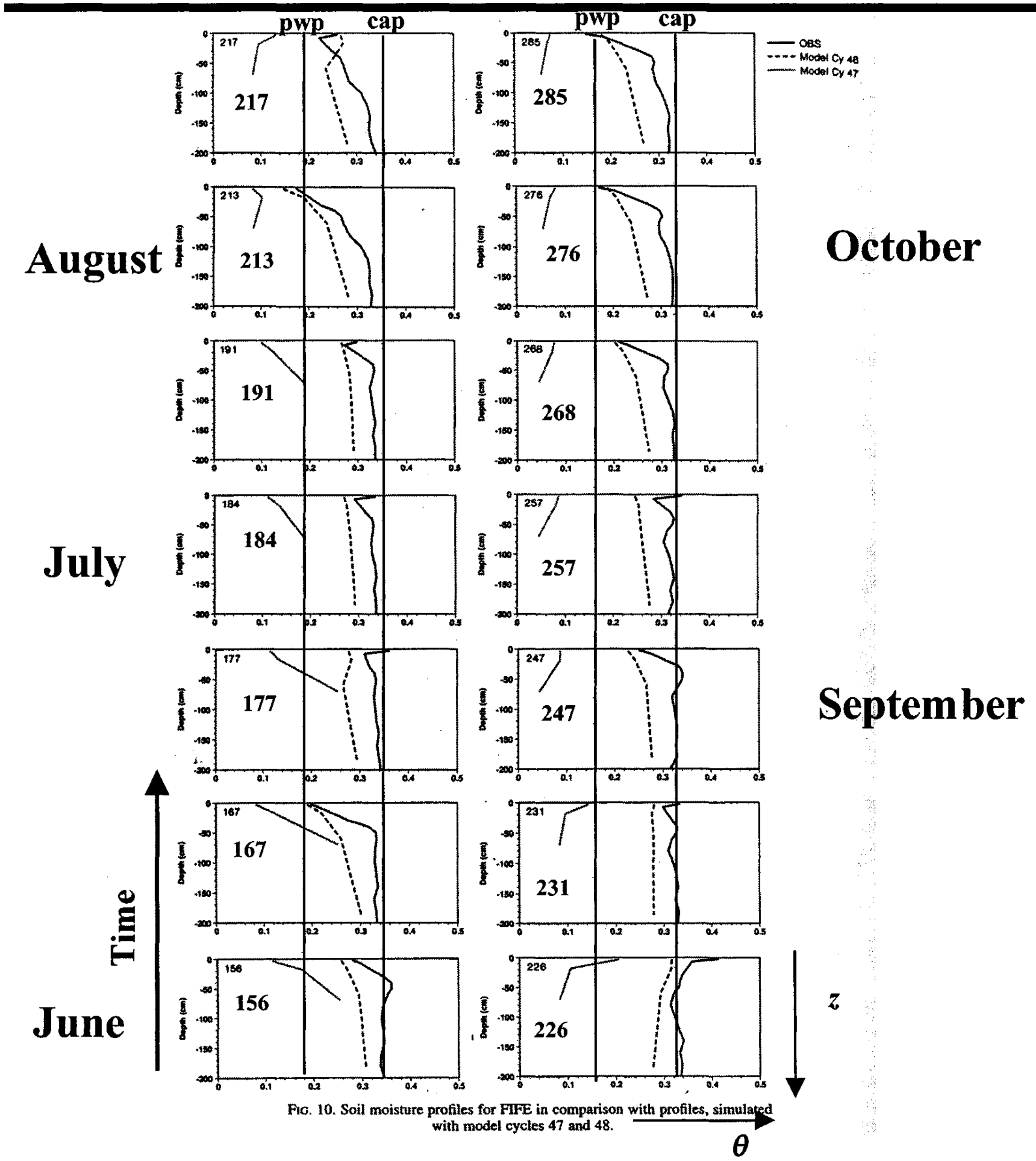



FIG. 10. Soil moisture profiles for FIFE in comparison with profiles, simulated with model cycles 47 and 48.

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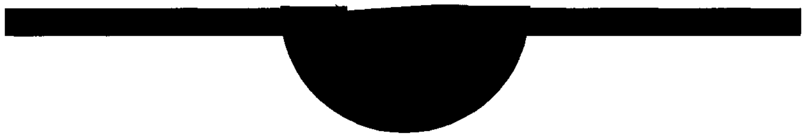
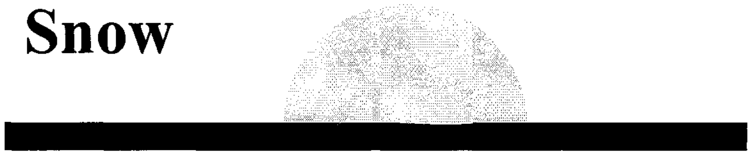
# Evaporation: Idealized surfaces

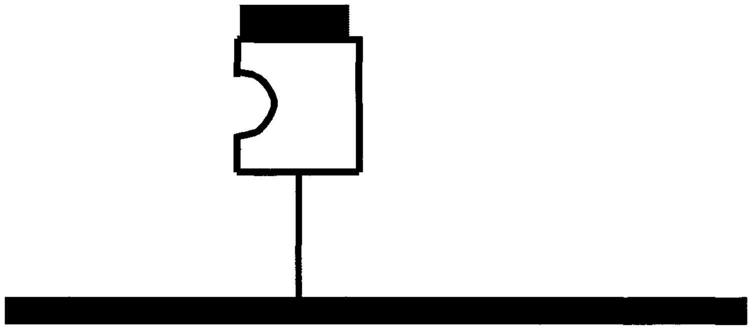
**Lake**



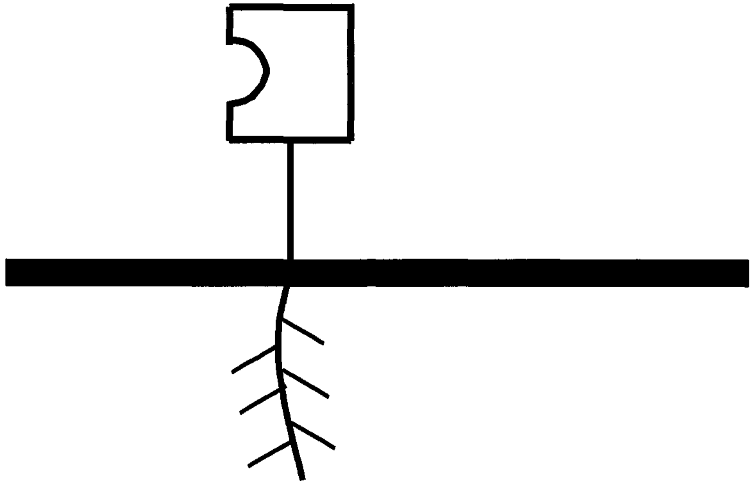
**Snow**



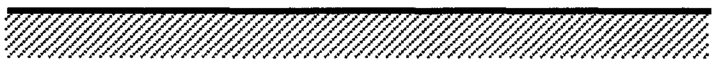
**Wet vegetation**



**Dry vegetation**



**Bare ground**



# Potential evaporation, bucket model

- **Potential evaporation**
  - **The evaporation of a large uniform surface, sufficiently moist or wet (the air in contact to it is fully saturated)**

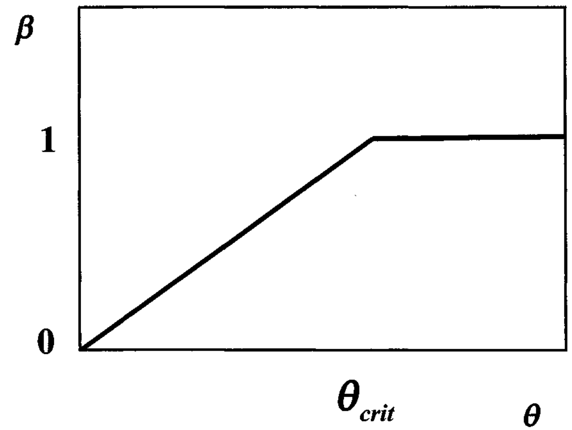
$$E_{pot} = \rho C_h u_L \left[ q_L - q_{sat}(T_{sk}, p_s) \right]$$

- **Evaporation efficiency** $\quad E = \beta E_{pot}$
  - **Ratio of evaporation to potential evaporation** $\quad 0 < \beta < 1$

- **Bucket model Budyko 1963, 1974 Manabe 1969**

$$\beta = \begin{cases} 1 & \theta > \theta_{crit} \\ \dfrac{\theta}{\theta_{crit}} & \theta > \theta_{crit} \end{cases}$$

Ideally, $\beta$ should be a well defined function of soil cover (vegetation type vs. bare ground) and soil properties



# A general, algebraic formulation

$$E = \rho C_h u_L \left[ a_L q_L - a_s q_{sat}(T_{sk}, p_s) \right]$$

$$a_{L,s} = f(q_L, T_s, \text{state and nature of the soil, soil cover})$$

- **Two limit behaviours**
  - **Bare soil: Evaporation dependent on soil water (and trapped water vapour) in a top shallow layer of soil (~ 20 mm).**
  - **Vegetated surfaces: Evaporation controlled by a canopy resistance, dependent on shortwave radiation, water on the root zone (~ 1-5 m deep) and other physical/physiological effects.**

# Transpiration: The big leaf approximation

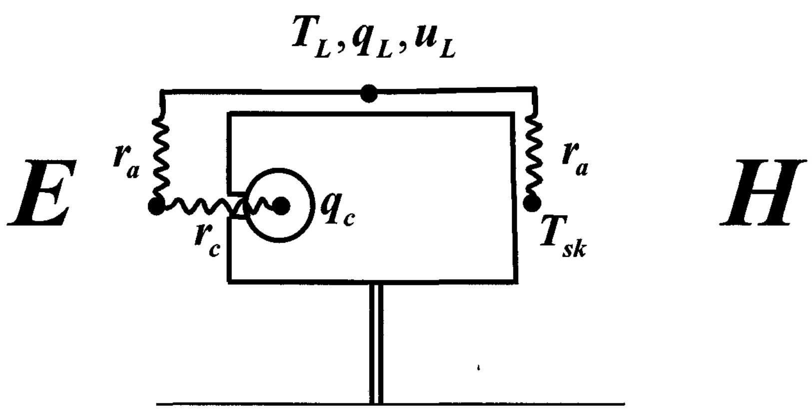
- **Sensible heat (*H*), the resistance formulation**

$$H = \rho C_p C_h u_L (T_L - T_{sk}) = \rho C_p \frac{T_L - T_{sk}}{r_a}$$

$r_a$ aerodynamic resistance, $[r_a] = sm^{-1}$

$$r_a = \frac{1}{C_h u_L}$$

- **Evaporation (*E*), the resistance formulation (the big leaf approximation, Deardorff 1978, Monteith 1965)**




$$E = \rho \frac{q_L - q_c}{r_a + r_c} = \rho \frac{q_L - q_{sat}(T_{sk})}{r_a + r_c}$$

$q_c = q_{sat}(T_{sk})$ Specific humidity for the interior of the stomata, ie, for saturated conditions

$r_c$ canopy resistance

# Some plant science

$$r_c = \frac{\langle r_s \rangle}{L}$$

$$\langle ( \ ) \rangle = \frac{L}{dL/( \ )}$$

$L$ Leaf area index (see later for a definition)

- **$r_s$, the stomatal resistance of a single leaf. Physiological control of water loss by the vegetation. Stomata (valve-like openings) regulate the outflow of water vapour (assumed to be saturated in the stomata cells) and the intake of $CO_2$ from photosynthesis. The energy required for the opening is provided by radiation (Photosynthetically Active Radiation, PAR). In many environments the system appears to be operate in such a way to maximize the $CO_2$ intake for a minimum water vapour loss. When soil moisture is scarce the stomatal apertures close to prevent wilt and dessication of the plant.**

# Jarvis approach(1)

$$r_c = r_{c\min} f_1(PAR) f_2(T_s) f_3(D_a) f_4(\theta)$$

$T_s$ soil temperature

$D_a = e_{sat}(T_a) - e$ near - surface saturation deficit

$f_i > 1$

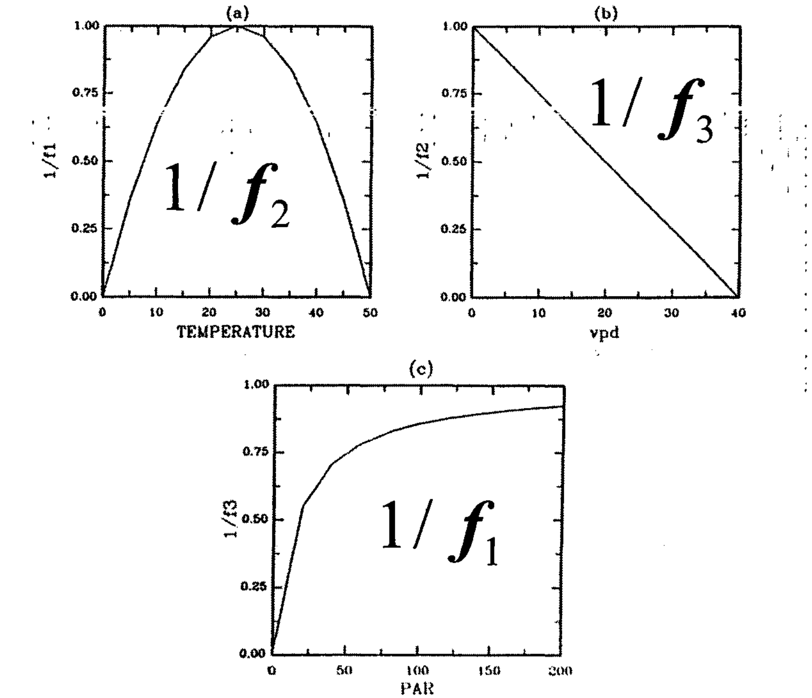



Fig. 4. Environmental dependencies of the inverse of stomatal resistances (i.e. conductance) in BATS model: (a) dependence of conductance on temperature; (b) dependence of conductance on vapor pressure deficit; (c) dependence of conductance on PAR.

**Dickinson et al 1991**

# Jarvis approach(2)

**Shuttleworth 1993**

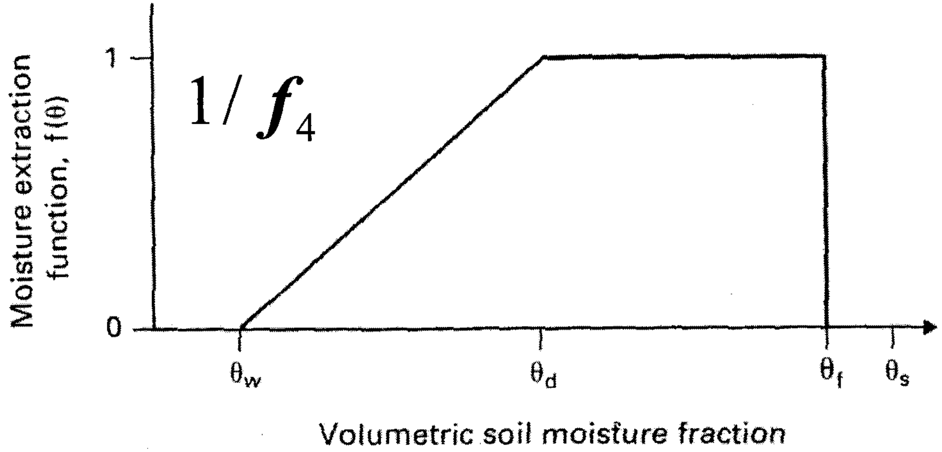



**FIGURE 4.4.3** Typical variation of the moisture extraction function $f(\theta)$ which modifies the potential crop coefficient in response to changes in the volumetric soil moisture content $\theta$ in (portions of) the plants' rooting zone. $\theta_s$, $\theta_f$, and $\theta_w$ are the values of $\theta$ at saturation, field capacity, and wilting point, respectively, and $(\theta_d/\theta_f)$ is typically 0.5 to 0.8. These values are determined by soil type.

| | |
|---|---|
| $\theta$ | soil water in the root zone |
| $\theta_d = \theta_{cap}$ | in many models (ECMWF) |
| $\theta_{ava} = \theta_d - \theta_{pwp}$ | availability |
| $\theta_{ava}$ ?$d$ | available soil water (water holding capacity) |

# Bare ground evaporation

- **Soil (bare ground) evaporation is due to:**
  - **Molecular diffusion from the water in the pores of the soil matrix up to the interface soil atmosphere ($z_{0q}$)**
  - **Laminar and turbulent diffusion in the air between $z_{0q}$ and screen level height**
- **All methods are sensitive to the water in the first few cm of the soil (where the water vapour gradient is large). In very dry conditions, water vapour inside the soil becomes dominant**

$\alpha$ method

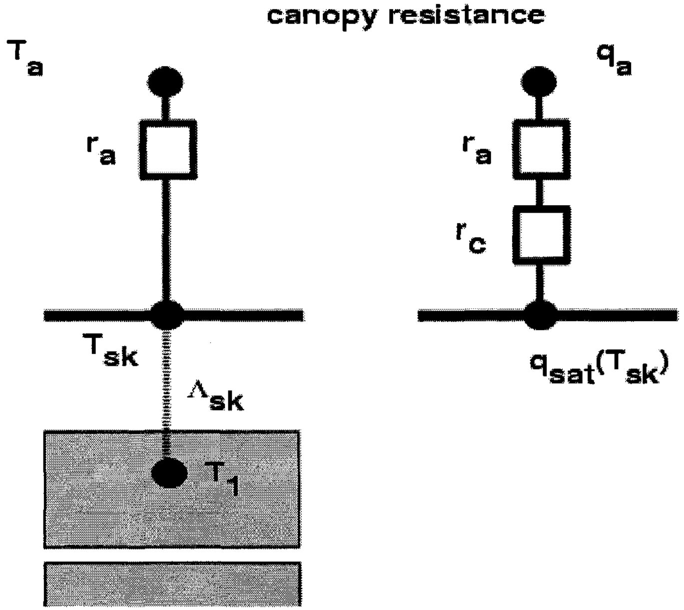
$$E = \rho \frac{q_L - \alpha q_{sat}(T_{sk})}{r_a}$$

$\alpha = f(\theta_1)$ "Relative humidity of the soil"

$\theta_1$ Top soil layer (a few cm) water

# TESSEL transpiration

$i$ is the high/low vegetation tile




$$E_i = \frac{\rho_a}{r_{c,i} + r_{a,i}} \left[ q_a - q_{sat}(T_{sk,i}) \right]$$

$$r_c = \frac{r_{s,\min}}{LAI} f_1(R_S^{\downarrow}) f_3(\bar{\theta}) f_4(D_a)$$

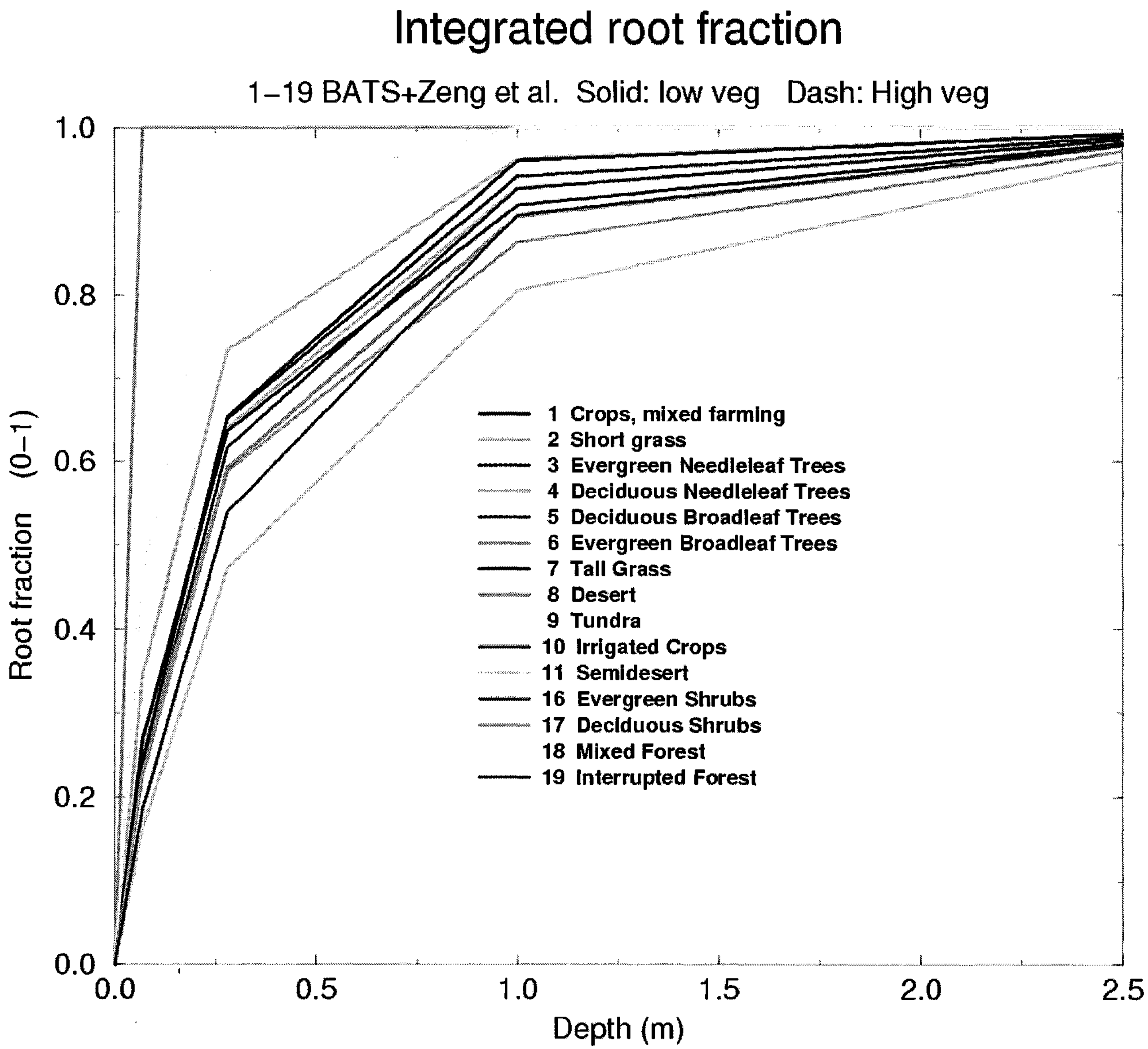
$$f_1(R_S^{\downarrow})^{-1} = \min\left[1, \frac{a(1 + bR_S^{\downarrow})}{bR_S^{\downarrow} + c}\right]$$

$$f_4(D_a)^{-1} = \exp(-g_D D_a) \qquad g_D = 0 \quad \text{for low vegetation}$$

$$f_3(\bar{\theta})^{-1} = \begin{cases} 0 & \bar{\theta} < \theta_{pwp} \\ \dfrac{\bar{\theta} - \theta_{pwp}}{\theta_{cap} - \theta_{pwp}} & \theta_{pwp} < \bar{\theta} < \theta_{cap} \\ 1 & \bar{\theta} > \theta_{cap} \end{cases}$$

$\bar{\theta}$ is a root weighted average of the liquid soil water

# TESSEL root profile

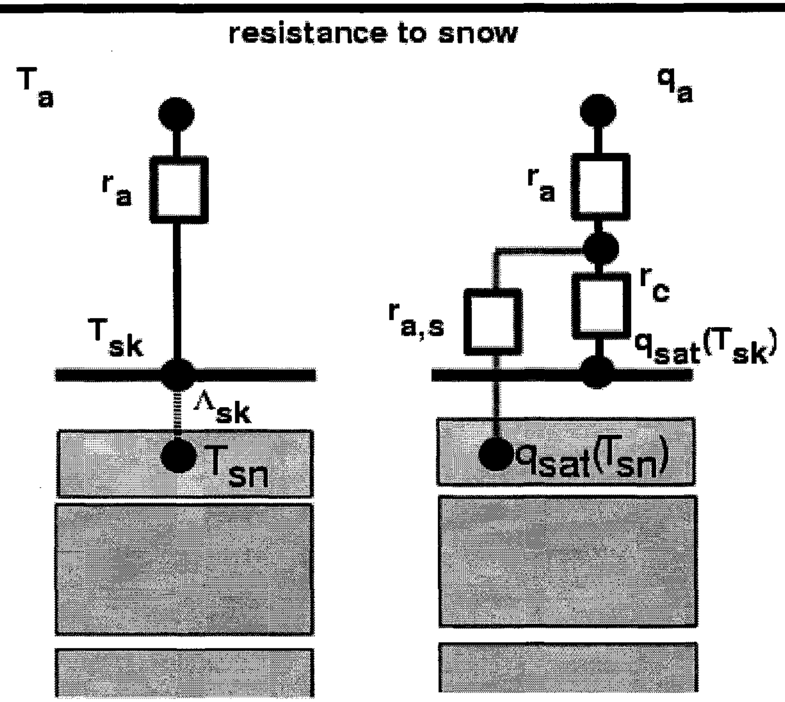


# TESSEL evaporation: shaded snow

$i$ is the shaded snow tile

$$E = E_{veg} + E_{snow}$$

$$E_{snow} = \frac{\rho_a}{r_{a,s} + r_a}\left[q_a - q_{sat}(T_{sn})\right]$$




- **Evaporation is the sum of the contribution of the snow underneath (with an additional resistance, $r_{a,s}$ ,to simulate the lower within canopy wind speed) and the exposed canopy. The former is dominant in early spring (frozen soils) and the latter is dominant in late spring.**

# TESSEL bare ground evaporation

$i$ is the bare ground tile

$$E_i = \frac{\rho_a}{r_{soil} + r_{a,i}} \left[ q_a - q_{sat}(T_{sk,i}) \right]$$

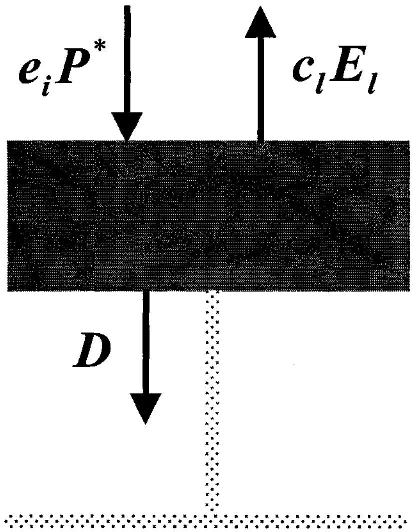
$$r_{soil} = r_{soil,min} f_3(\theta_1)$$

$$f_3(\theta_1)^{-1} = \begin{cases} 0 & \theta_{1,liq} < \theta_{pwp} \\ \dfrac{\theta_{1,liq} - \theta_{pwp}}{\theta_{cap} - \theta_{pwp}} & \theta_{pwp} < \theta_{1,liq} < \theta_{cap} \\ 1 & \theta_{1,liq} > \theta_{cap} \end{cases}$$

# Interception (1)

- Interception layer represents the water collected by interception of precipitation and dew deposition on the canopy leaves (and stems)
- Interception (I) is the amount of precipitation (P) collected by the interception layer and available for "direct" (potential) evaporation. I/P ranges over 0.15-0.30 in the tropics and 0.25-0.50 in mid-latitudes.
- Leaf Area Index (LAI) is (total area of leaf surface)/(projected area) $0.1 < LAI < 6$
- Two issues
  - Size of the reservoir
  - $C_I$, fraction of a gridbox covered by the interception layer
- T=P-I; Throughfall (T) is precipitation minus interception

# Interception: Canopy water budget




$$\frac{fw_l}{ft} = e_i P^* + c_l E_l + D = I + c_l E_l$$

$w_l$ Intercepted water

$e_i$ efficiency of interception

$P^*$ modified precipitation

$c_l E_l$ evaporation of intercepted water

$D$ rate of drainage at the bottom of the canopy

# TESSEL: interception

- **Interception layer for rainfall and dew deposition**

$$\frac{fw_l}{ft} = I + c_l E_l$$

$$I = max\ \overline{\ e}_i c_l P^*, \frac{w_{lmx} - w_l^n}{\Delta t}$$

$P^* = P / k$ **modified precipitation**

$k$ **fraction of grid - box covered by precipitation**

$T = P - I$ **Throughfall (input to top soil)**

# Example: Deep tropics interception

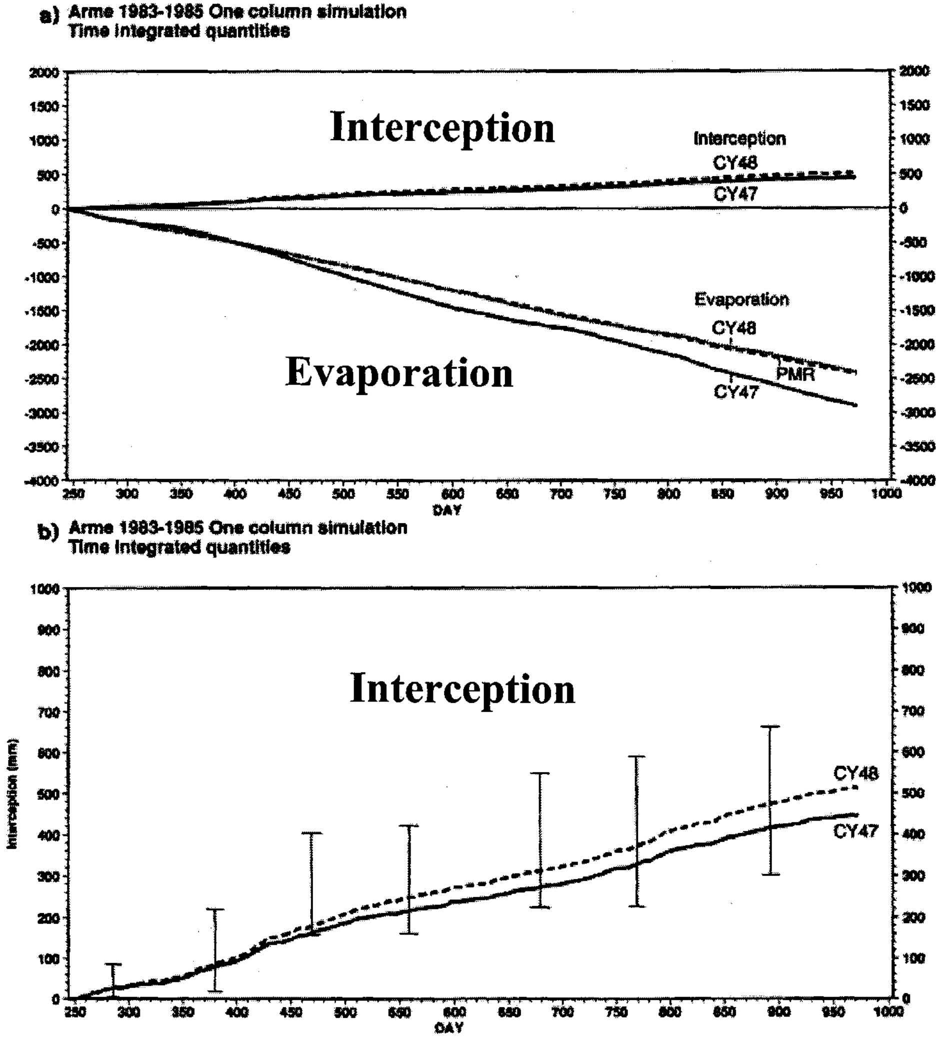



FIG. 17. Accumulated interception and evaporation for ARME with model cycles 47 and 48. Curve PMR is an estimation of evaporation from the Penman–Monteith–Rutter model as described by Shuttleworth (1988) with some improvements by Dolman et al. (1991). The accumulated interception is repeated in panel (b), notice the different scale, in comparison with the range of observations (error bars) as published by Sellers et al. (1989).

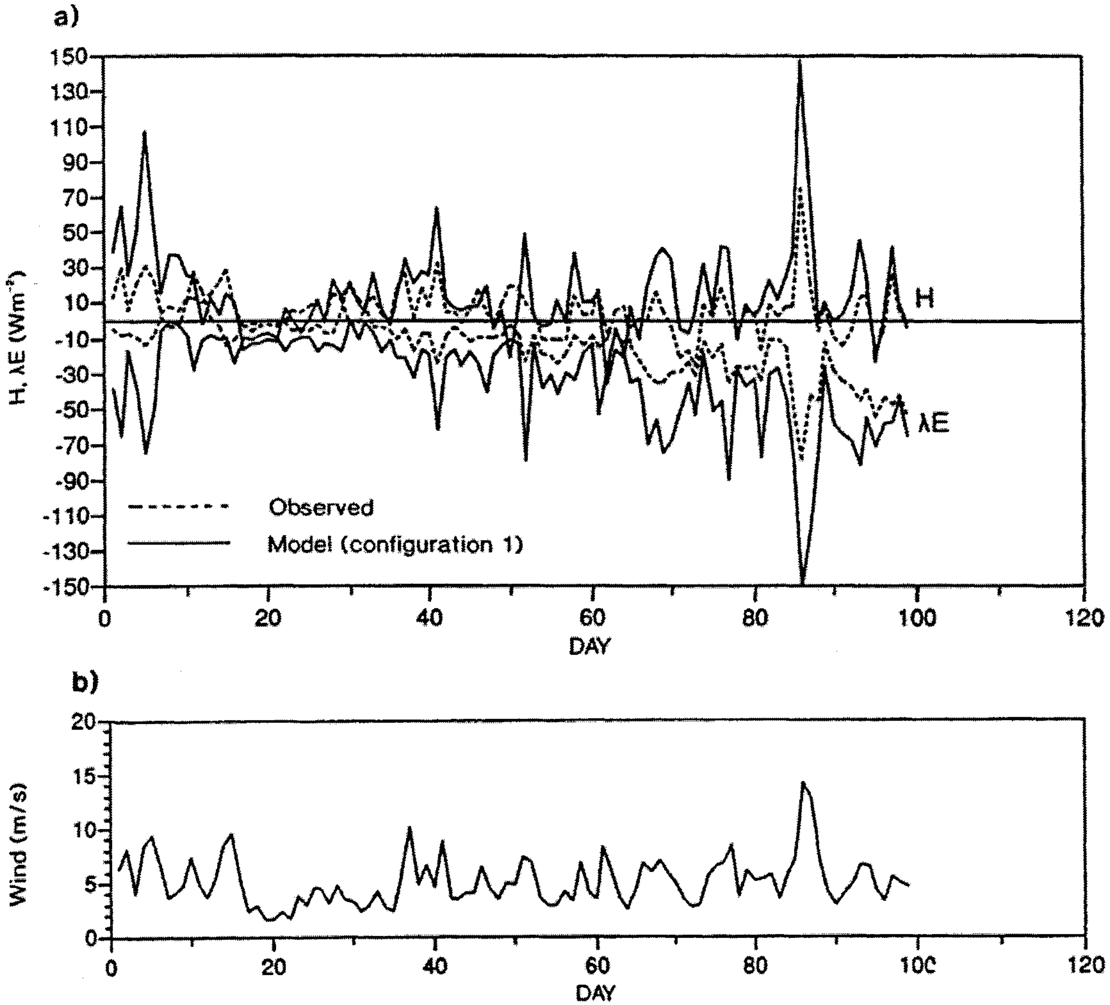
Viterbo and Beljaars 1995

**ARME, 1983-1985, Amazon forest**
**Accumulated water fluxes**

# Case study: Aerodynamic resistance (1)

- Cabauw (Netherlands), is a grass covered area, where multi-year detailed boundary layer measurements have been taken
- Observations were used to force a stand-alone version of the surface model for 1987
- The first model configuration tried had $z_{0h}=z_{0m}$

# Case study: Aerodynamic resistance (2)




**Beljaars and Viterbo 1994**

Fig. 2. Observed (dashed) and simulated (solid) diurnal averages of sensible and latent heat (Figure 2a). The wind speed is shown in Fig 2b.

# Case study: Aerodynamic resistance (3)

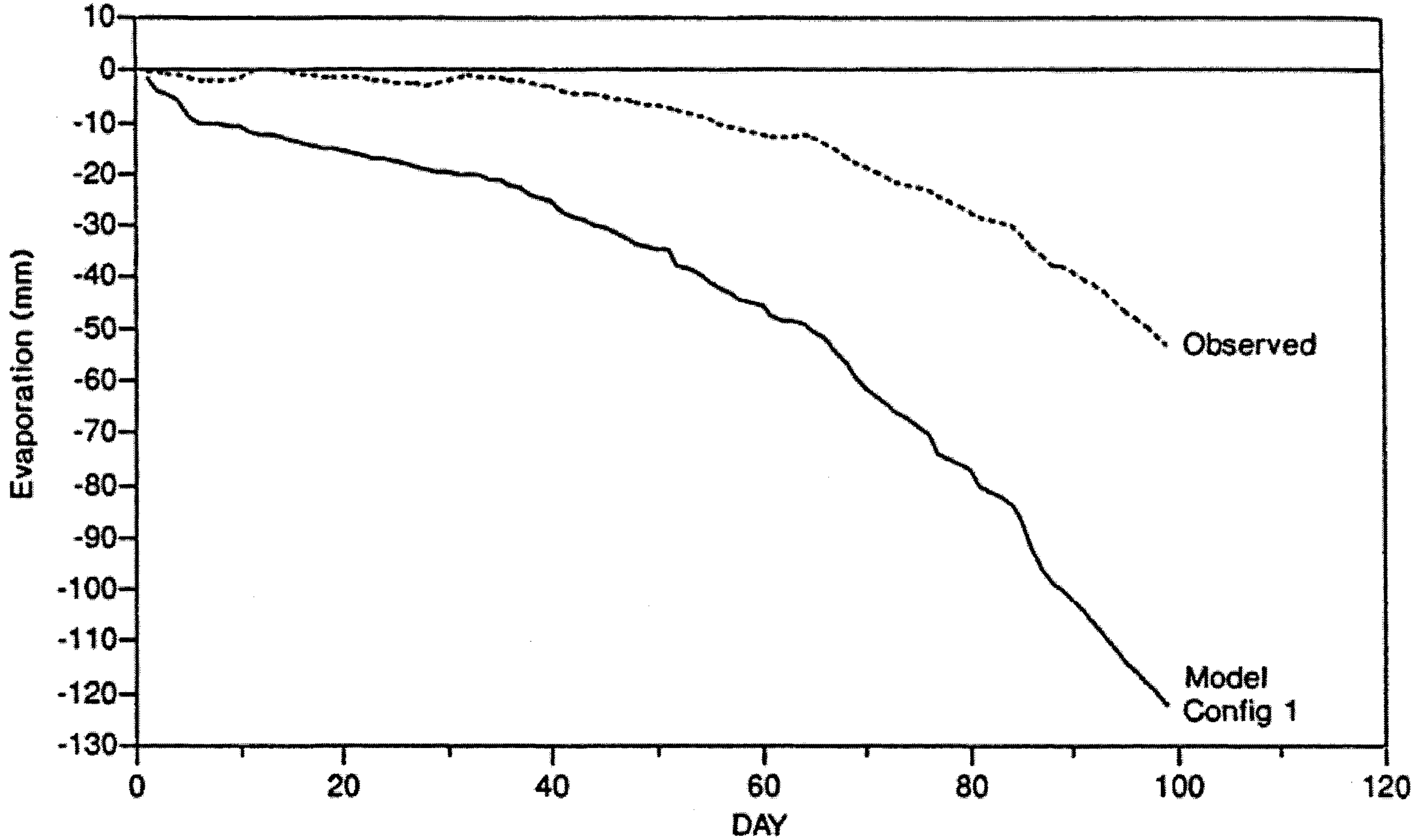



Fig. 3. Time-integrated evaporation (in mm water; upward flux is negative), observed (dashed) and simulated (configuration 1, solid) for the first 100 days of 1987 of the Cabauw data set.

# Case study: Aerodynamic resistance (4)

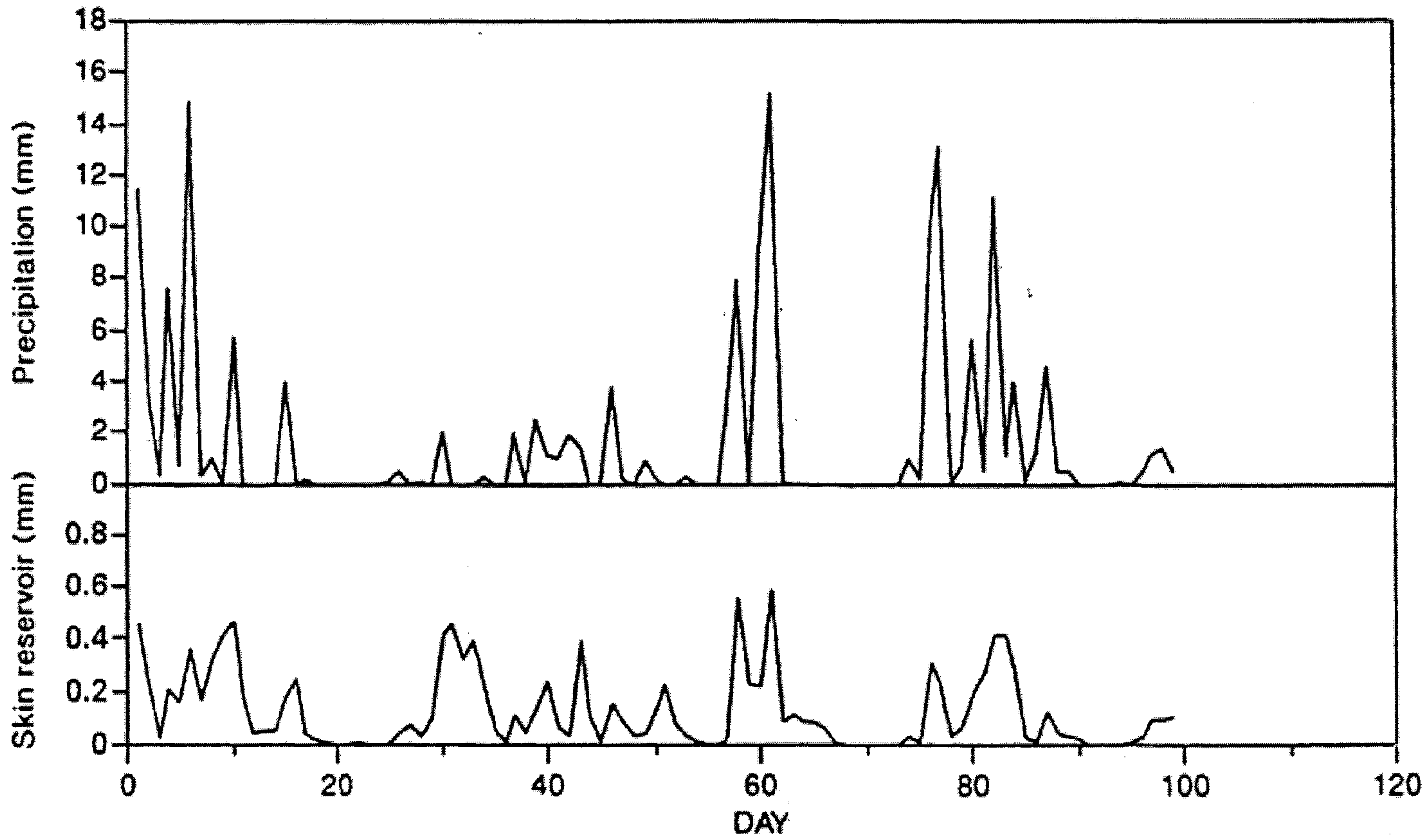



Fig. 4. Day sums of precipitation (observed, a) and diurnal average of skin reservoir content as simulated by the model in configuration 1 (b).

# Case study: Aerodynamic resistance (5)

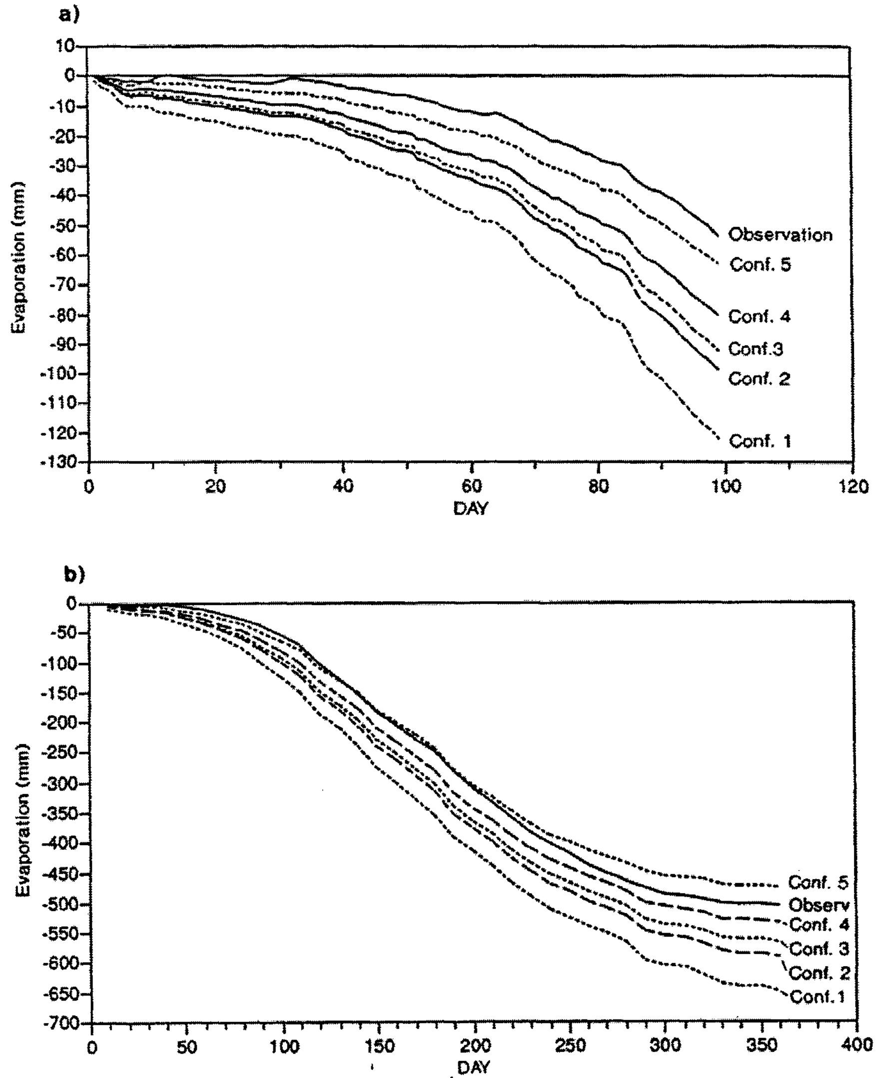



Fig. 5. Accumulated evaporation as observed and simulated with different model configurations (see Table 1) for the first 100 days of 1987 (a) and for the entire year (b).

TABLE I

Parameters in the different model simulations

| Model configuration | $z_{0m}$ (m) | $z_{0h}$ (m) |
|---|---|---|
| 1 | 0.4 | 0.4 |
| 2 | 0.4 | 0.033 |
| 3 | 0.1 | 0.1 |
| 4 | 0.1 | 0.01 |
| 5 | 0.1 | 0.0001 |

# Runoff and infiltration

$$I_f = T - Y$$

$I_f$ infiltration

$T$ throughfall

$Y$ runoff

- **Infiltration is that part of the precipitation flux that contributes to wet the soil**
- **Runoff occurs in**
  - **Parts of the watershed where hydraulic conductivities are lowest (Horton mechanism, in upslope areas)**
  - **Parts of the watershed where the water table is shallowest (Dunne mechanism, in near channel wetlands)**
- **Runoff depends on orography, nature and moisture state of the soil, precipitation intensity, and sub-grid scale variations of all these parameters**
- **Runoff in NWP/climate models should be called runoff generation (upon application of a routeing algorithm, it becomes "runoff")**

# TESSEL: runoff generation

- Surface runoff is based on a maximum infiltration limit concept, but with no sub-grid scale variability of either the precipitation flux or the top soil water content. Physically, it is the Hortonian concept, but applied at the wrong spatial scales (the model grid-box)
- Deep runoff is free drainage at the bottom
- All computations are performed with soil liquid water only. The frozen fraction of the surface is impervious to vertical water fluxes

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# Snow

- **Snow insulates the ground (30% to 90% of the snow mantle is air)**
- **A snow covered surface has a higher albedo than any other natural surface (0.2-0.3 in the presence of forests, 0.5-0.8 for bare ground/low vegetation)**
- **Snow melting keeps the surface temperature at 0 C for a long period in spring**

# Snow energy budget

$$(\rho C)_{sn} D \frac{fT_{sn}}{ft} = R_n + L_s E + H - G - Q_m$$

$L_s$ **Latent heat of sublimation**

$Q_m$ **Heat of melting**

$G$ **Basal heat flux**

$$Q_m = L_f M = L_f \frac{\rho_w}{C_{sn}} \left.\frac{fS}{ft}\right|_m$$

$L_f$ Latent heat of fusion

$M$ Meltwater

$C_{sn}$ Snow fraction

# Snow mass budget

$$\rho_w \frac{fS}{ft} = F + C_{sn}(E - M)$$

$F$ **Snowfall**

$S$ **Snow water equivalent (snow mass)**

- **Snow mass (*S*) and snow depth (*D*)**

$$D = \frac{\rho_w}{\rho_{sn}} \frac{S}{C_{sn}}$$

$\rho_{sn}$ Snow density

$D$ Snow depth for the snow - covered area, $m$

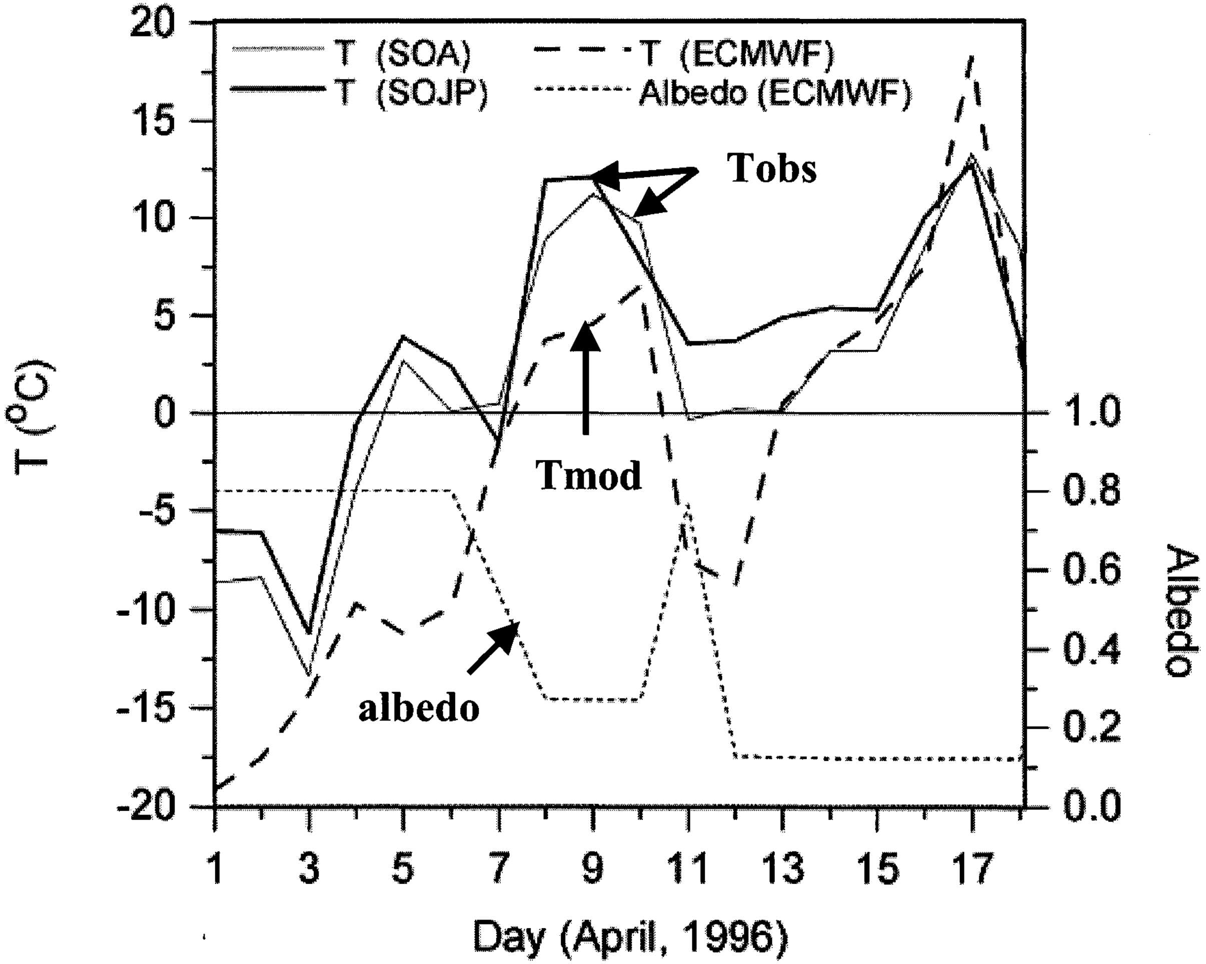
# Metamorphism, density and albedo

- **Density**
  - Weighted average between current density and the density of fresh snow, in case of snowfall
  - Exponential relaxation
- **Albedo**
  - Exponential relaxation with different time scales for melting and non-melting snow

# Case study: Boreal forest albedo (1)

**2m temperature at BOREAS, Canada, 18 LT**

**Observations vs 48 h forecast**



# Case study: Boreal forest albedo (2)

**Observed albedo BOREAS 1994**

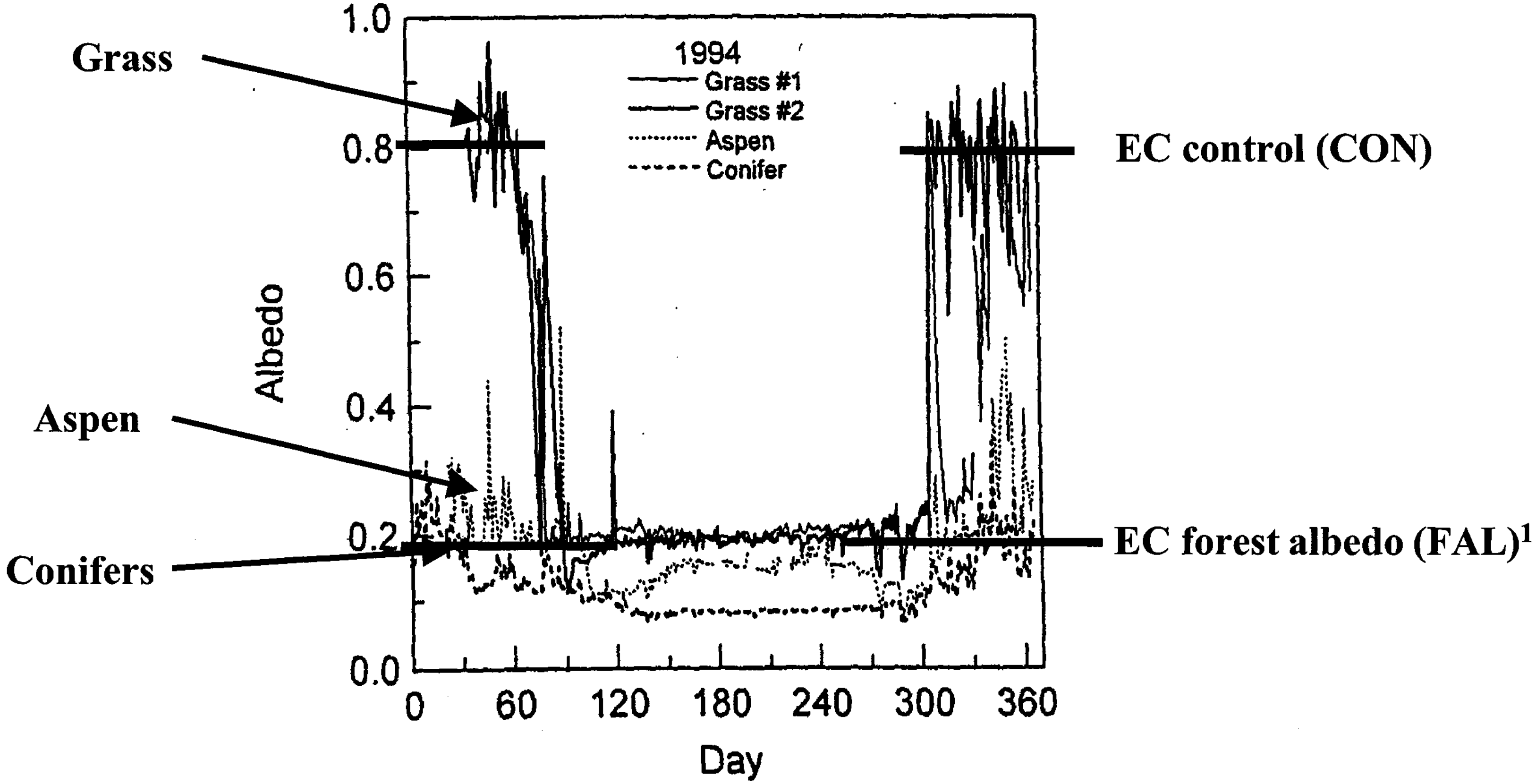
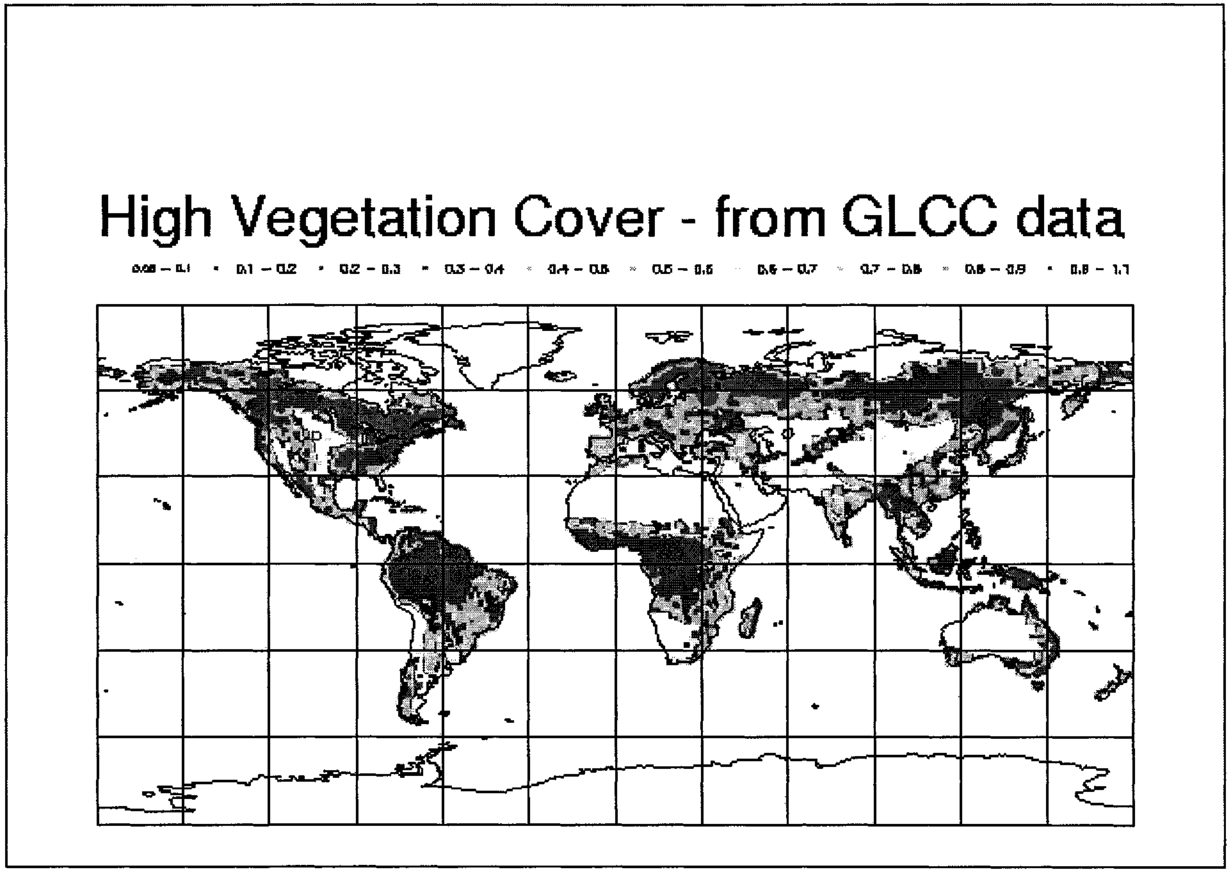



**Figure 4.** Daily average albedo for 10 BOREAS mesonet sites for 1994; showing two grass sites, the aspen site, and an average of the seven conifer sites.

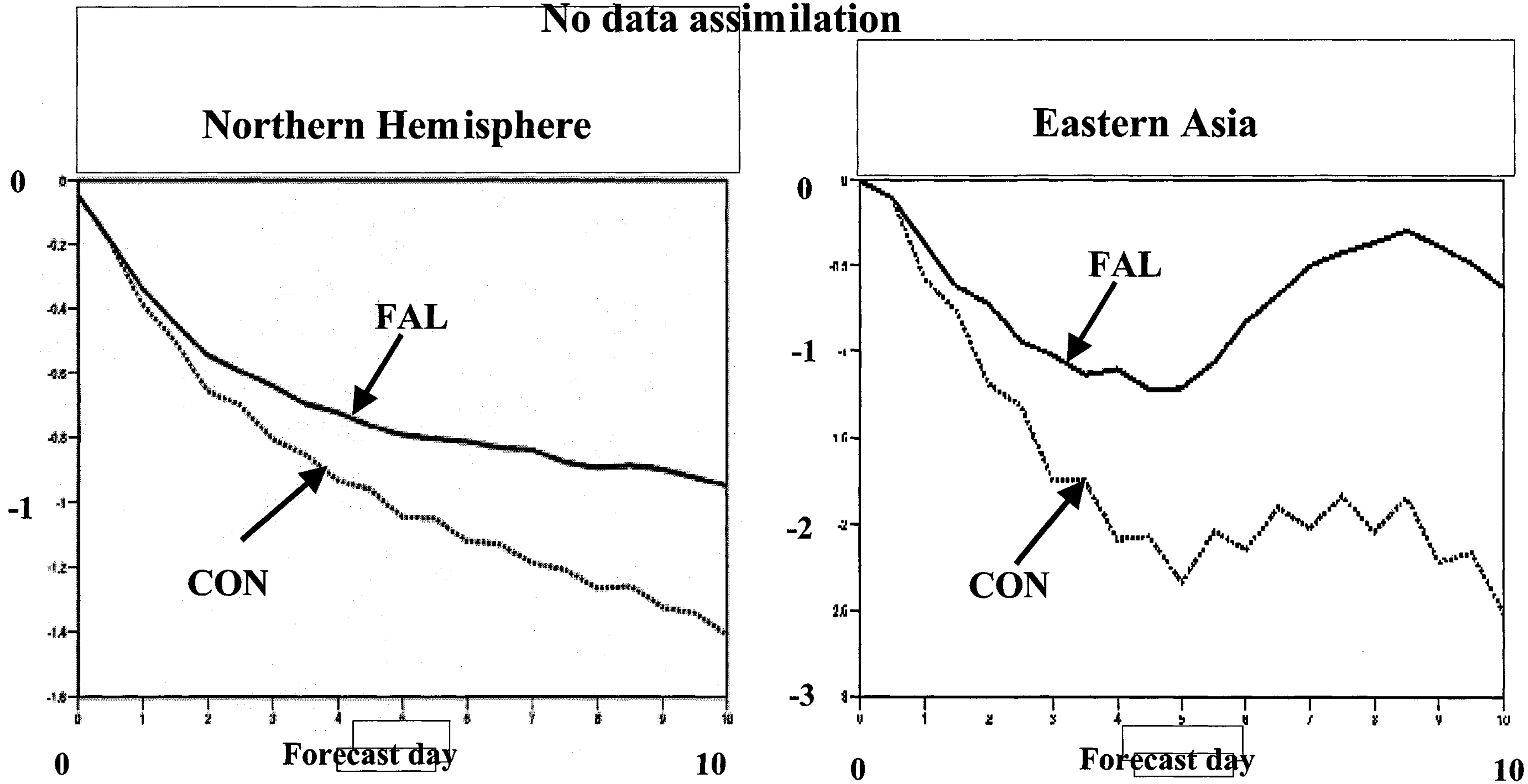
**Viterbo and Betts, 1999: J. Geophys. Res., 104D, 27,803-27,810.**

# Case study: Boreal forest albedo (3)



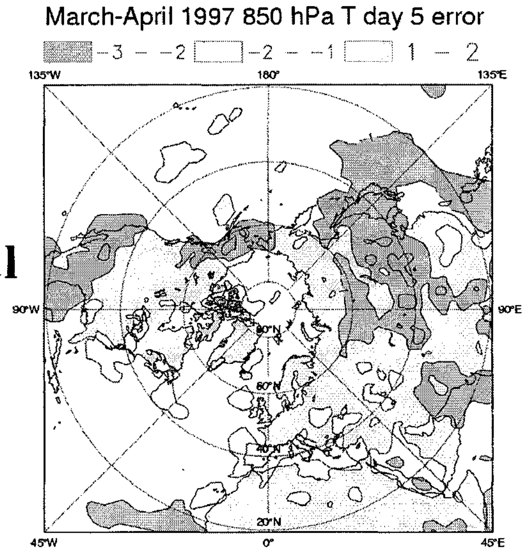
# Case study: Boreal forest albedo (4)

**850 hPa temperature bias**
**20 forecasts every 3 days, March-April 1996**
**No data assimilation**

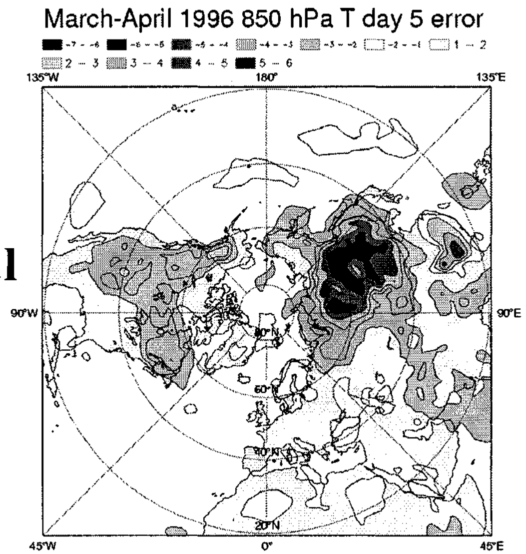


# Case study: Boreal forest albedo (5)

1997
Operational
Bias
(FAL)

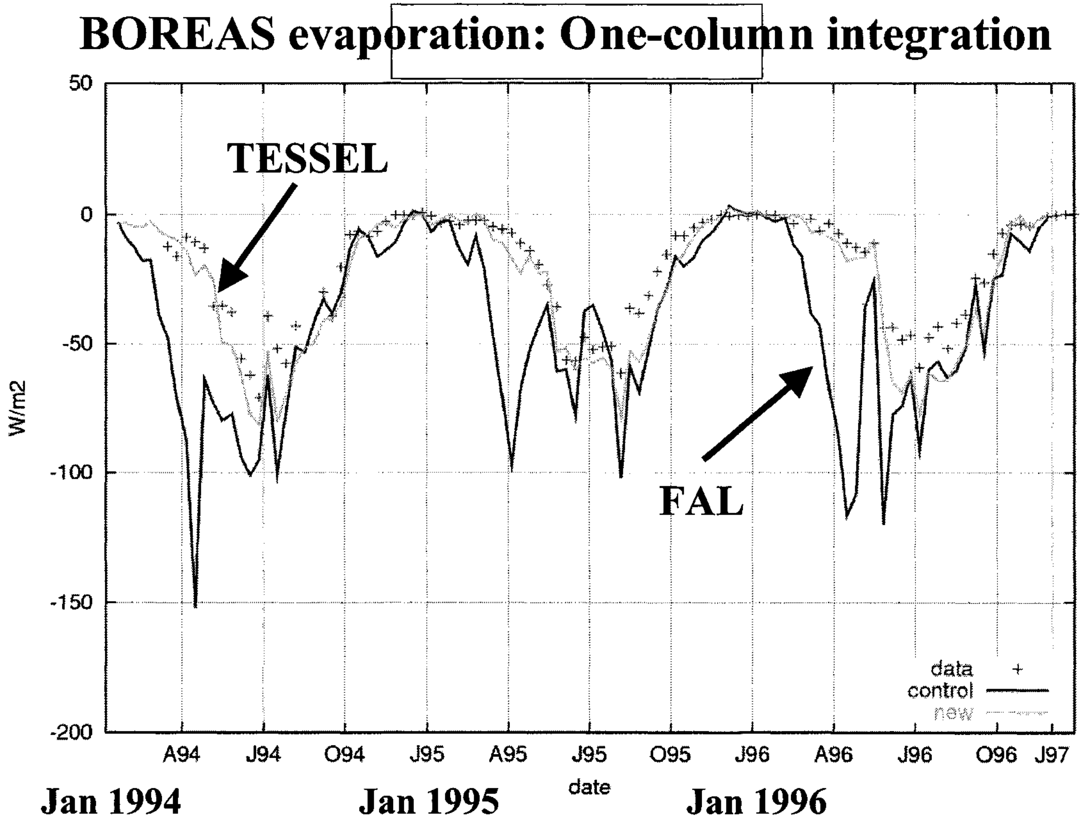



1996
Operational
Bias
(CON)




- **The surface albedo is the direct regulator of the energy available to the surface. The albedo of natural surfaces has a limited range (0.1-0.3), but in non-forested snow covered areas can reach values up to 0.8.**
- **Snow covered boreal forests have a much lower albedo than grassland to their south and tundra to their north; the presence of boreal forests has a direct control on the climate of high-latitudes.**

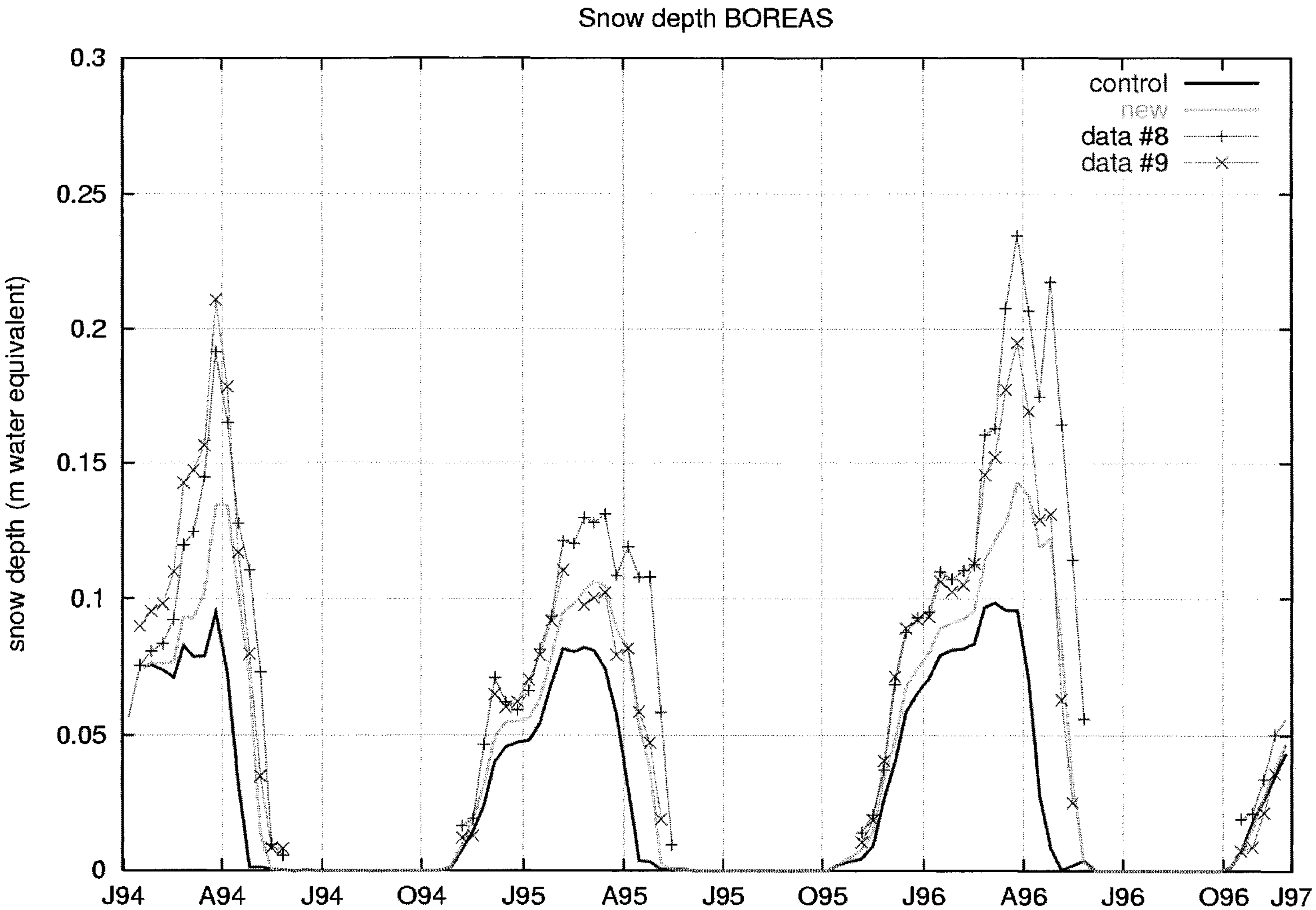
# Evaporation: “The sting in the tail” (1)




van den Hurk et al 2000

- **The model FAL erroneously transform the available energy into evaporation. However, plants have limited transpiration in winter/spring, when the roots are frozen.**
- **The TESSEL model (operational at ECMWF July 2000) simulates that behaviour.**

# Evaporation: "The sting in the tail" (2)



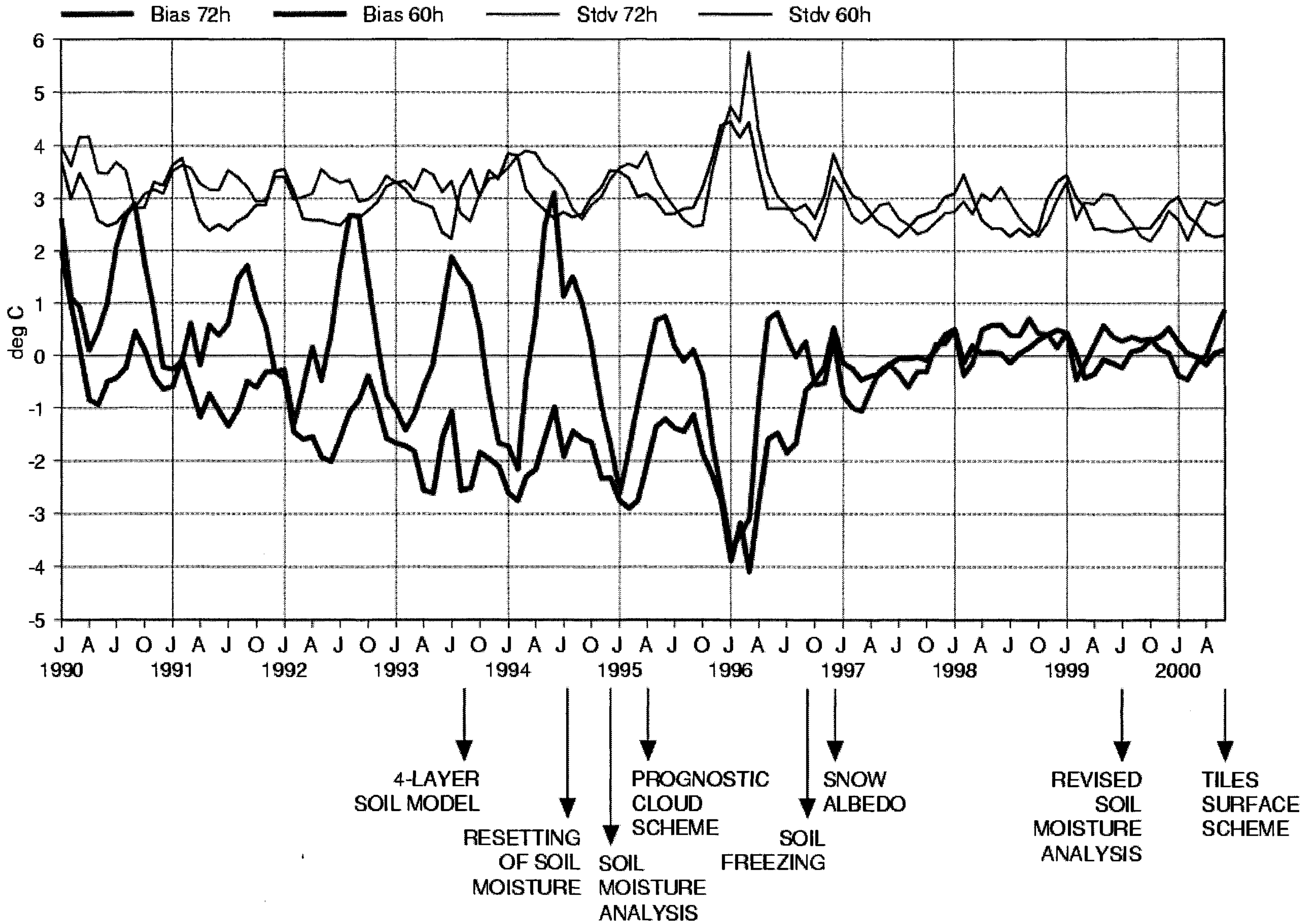
# Layout

- Introduction
- General remarks
- Model development and validation
- The surface energy budget
- Soil heat transfer
- Soil water transfer
- Surface fluxes
- Initial conditions
- Snow
- Conclusions and a look ahead

# Conclusions

- The land surface can have a significant impact on the atmosphere at the synoptic/continental time scale when it affects the partitioning of the net radiation into sensible/latent heat, via the soil water content. This effect can be local or non-local.
- The land surface has a significant impact on the atmosphere at a synoptic/continental scale when it affects the net radiation at the surface, such as a change of the albedo in spring.
- In winter, stable, situations the land surface is decoupled from the atmosphere: Large variations in surface temperature affect only the lowest hundred metres of the atmosphere with no impact on the circulation.
- Forecast systems are sensitive to mis-representation of longer time scales in the land-surface/atmosphere interaction.
- Data assimilation systems are ideal tools to validate parameterisations because of their constant confrontation with observations and the customers' needs for accurate diurnal cycles of weather parameters.

# History of ECMWF 2m T errors



# 3-5 years ahead

- Model
  - Surface runoff taking into account sub-grid scale heterogeneity
  - Distribution of soil types
  - Alternative formulation for hydraulic conductivity/diffusivity
  - Handling of fractional lake/sea cover
  - Urban tile
  - Snow cover and orography
  - Monthly climatology of LAI and biome vegetation fractions
  - Snow over ice
  - Daily production of basin averaged runoff and precipitation
- Data assimilation
  - Use tile skin temperature, assisting on satellite retrievals over land
  - Cloudy and clear sky tiles for satellite retrievals
  - Use of satellite data for snow data assimilation
  - Prototype scheme for surface data assimilation

# 5-10 years ahead

- **Model**
  - Near-real time (weekly) input of satellite derived LAI and vegetation fractions
  - $CO_2$ model and NPP production
  - Crop yield forecasts
  - Lake model
  - Runoff routeing
- **Data assimilation**
  - Studies on assimilation of microwave L-band data to initialise soil water