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SMR.1313 - 5

Summer Colloquium on the Physics of Weather and Climate

Workshop on Land-Atmosphere Interactions in Climate Models

(28 May - 8 June 2001)

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

Lecture 2

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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

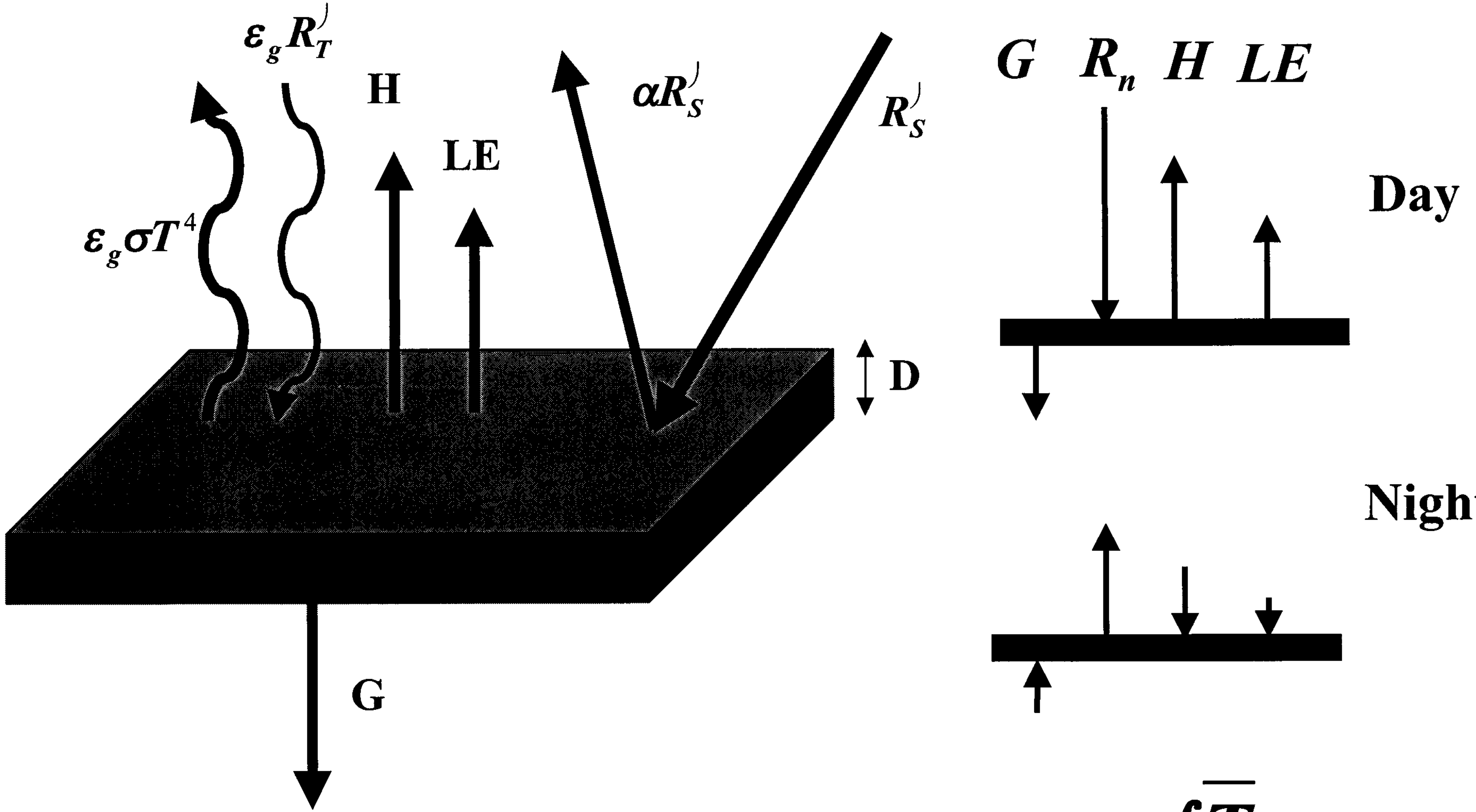
- Introduction
- General remarks
- Model development and validation
- <u>The surface energy budget</u>
- <u>Soil heat transfer</u>
- Soil water transfer
- Surface fluxes
- Initial conditions
- Snow
- Conclusions and a look ahead

Lecture 1 (Introduction – Model development and validation)

Lecture 2 (The surface energy budget – Soil heat transfer)

Lecture 3 (Soil water transfer – Conclusions and a look ahead)

Thermal budget of a ground layer at the surface



$$R_n + H + L?E + G = (\rho C)_g D \frac{f\overline{T}_s}{ft}$$

Energy budget: Summer examples

Dry

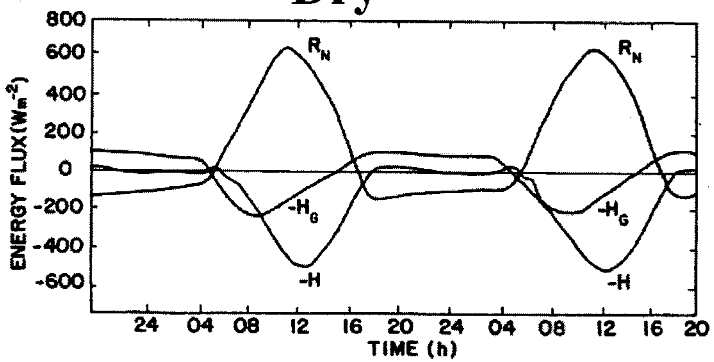


Fig. 2.3 Observed diurnal energy balance over a dry lake bed at El Mirage, California, on June 10 and 11, 1950. [After Vehrencamp (1953).]

Fir canopy

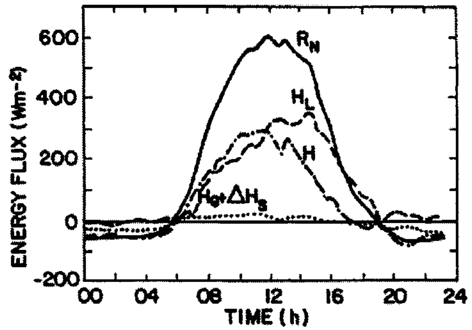


Fig. 2.5 Observed energy budget of a Douglas fir canopy at Haney, British Columbia, on July 23, 1970. [From Oke (1987); after McNaughton and Black (1973).]

Barley field

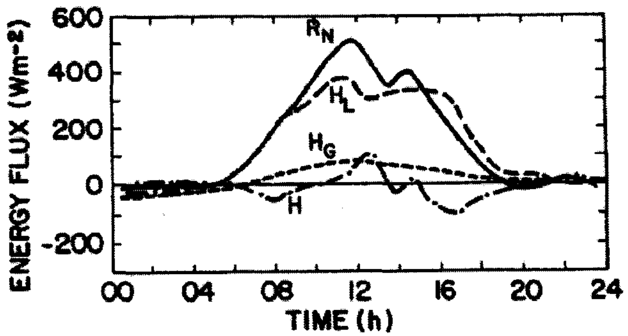


Fig. 2.4 Observed diurnal energy budget of a barley field at Rothamsted, England, on July 23, 1963. [From Oke (1987); after Long *et al.* (1964).]

$$R_N = R_n$$

$$-H_G = G$$

$$-H = H$$

$$-H_L = LE$$

Arya, 1988

The surface radiation

$$R_n = (1-\alpha)R_S^{\downarrow} + \varepsilon_g R_T^{\downarrow} - \varepsilon_g \sigma T_{sk}^4$$

α **Surface albedo**

ε_g **Surface emissivity**

T_{sk} **Skin temperature**

$\overline{T}_s$? T_{sk}

- **In some cases (snow, sea ice, dense canopies) the impinging solar radiations penetrates the "ground" layer and is absorbed at a variable depth. In those cases, an extinction coefficient is needed.**

Table 3.1

Radiative Properties of Natural Surfaces[a]

Surface type	Other specifications	Albedo (a)	Emissivity (ε)
Water	Small zenith angle	0.03–0.10	0.92–0.97
	Large zenith angle	0.10–0.50	0.92–0.97
Snow	Old	0.40–0.70	0.82–0.89
	Fresh	0.45–0.95	0.90–0.99
Ice	Sea	0.30–0.40	0.92–0.97
	Glacier	0.20–0.40	
Bare sand	Dry	0.35–0.45	0.84–0.90
	Wet	0.20–0.30	0.91–0.95
Bare soil	Dry clay	0.20–0.35	0.95
	Moist clay	0.10–0.20	0.97
	Wet fallow field	0.05–0.07	
Paved	Concrete	0.17–0.27	0.71–0.88
	Black gravel road	0.05–0.10	0.88–0.95
Grass	Long (1 m) Short (0.02 m)	0.16–0.26	0.90–0.95
Agricultural	Wheat, rice, etc.	0.10–0.25	0.90–0.99
	Orchards	0.15–0.20	0.90–0.95
Forests	Deciduous	0.10–0.20	0.97–0.98
	Coniferous	0.05–0.15	0.97–0.99

[a] Compiled from Sellers (1965), Kondratyev (1969), and Oke (1978).

Arya, 1988

The other terms

Sensible heat flux

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

$$C_h = f(Ri_B, z_{oh}, z_{om})$$

z_{oh}, z_{om} **specify the surface**

Evaporation

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

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

Ground heat flux

$$(\rho C)_g \frac{fT_s}{ft} = -\frac{fG}{fz} = \frac{f}{fz} \lambda_T \frac{fT}{fz}$$

$$(\rho C)_g, \lambda_T = f(\textbf{soil type, other soil characteristics})$$

Recap: The surface energy equation

$$(1-\alpha)R_S^{J} + \varepsilon_g R_T^{J} - \varepsilon_g \sigma T_{sk}^4 + \rho C_h u_L (C_p T_L + gz - C_p T_{sk}) +$$

$$\rho C_h u_L \left[a_L q_L - a_s q_{sat}(T_{sk}, p_s) \right] + G(T_s, T_{sk}) = (\rho C)_g D \frac{fT_s}{ft}$$

- Equation for T_s, T_{sk}
- For:
  - a thin soil layer at the top $(\rho C)_g D \frac{fT_s}{ft} \cup 0$
  - G (T_s, T_{sk}) is known, or parameterized or G << R_n

we have a non-linear equation defining the skin temperature

TESSEL

- **Skin layer at the interface between soil (snow) and atmosphere; no thermal inertia, instantaneous energy balance**
- **At the interface soil/atmosphere, each grid-box is divided into fractions (tiles), each fraction with a different functional behaviour. The different tiles see the same atmospheric column above and the same soil column below.**

$$G_i = \Lambda_{sk,i}\left(T_s - T_{sk,i}\right)$$

i index for tile

$i = 1,\ldots,N$

- **If there are N tiles, there will be N fluxes, N skin temperatures per grid-box**
- **There are currently up to 6 tiles over land (N=6)**

TESSEL skin temperature equation

$$
\begin{aligned}
&(1-\alpha_i)R_S^{\downarrow} + \varepsilon_g R_T^{\downarrow} - \varepsilon_g \sigma T_{sk,i}^4 + \\
&\rho C_{h,i} u_L (C_p T_L + gz - C_p T_{sk,i}) + \\
&\rho C_{h,i} u_L \left[a_{L,i} q_L - a_{s,i} q_{sat}(T_{sk,i}, p_s) \right] + \\
&\Lambda_{sk,i}(T_s - T_{sk,i}) = 0
\end{aligned}
$$

- **Grid-box quantities**

$$H = \underset{i}{\quad} C_i H_i$$

$$E = \underset{i}{\quad} C_i E_i$$

$$T_{sk} = \underset{i}{\quad} C_i T_{sk,i}$$

C_i Tile fraction

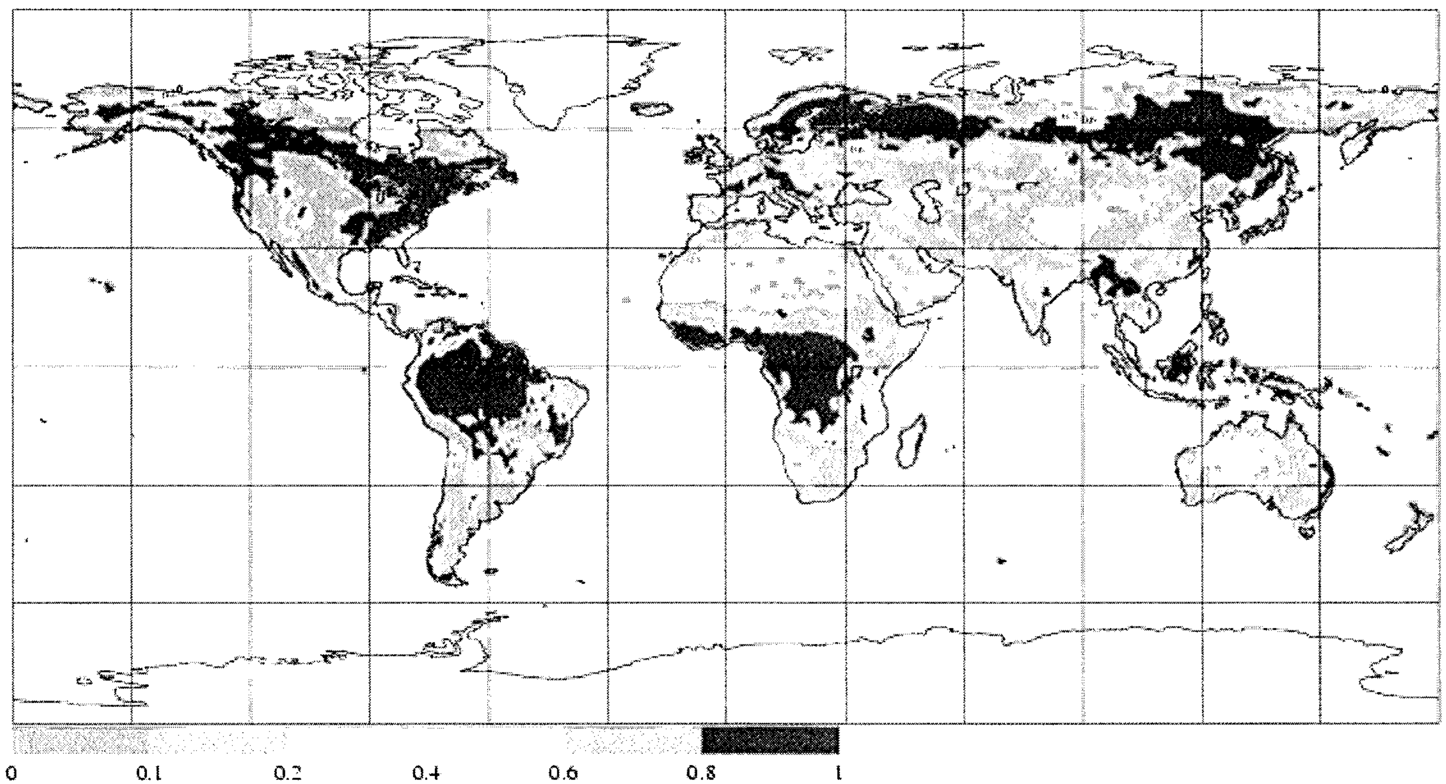
Tiles

Land	Sea and ice
High vegetation	Open sea / unfrozen lakes
Low vegetation	Sea ice / frozen lakes
High vegetation with snow beneath	
Snow on low vegetation	
Bare ground	
Interception layer	

TESSEL geographic characteristics

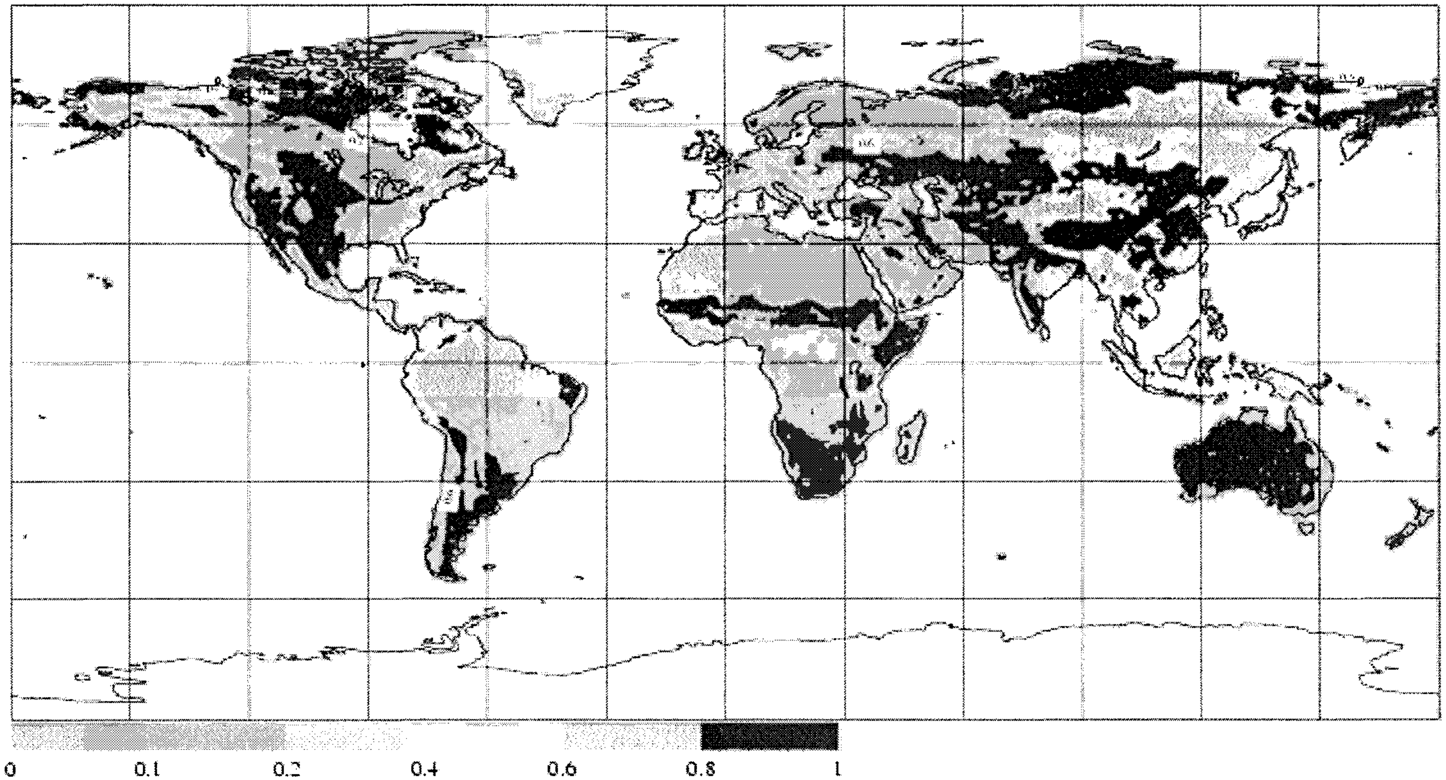
Fields	ERA15	TESSEL
Vegetation	Fraction	Fraction of low Fraction of high
Vegetation type	Global constant (grass)	Dominant low type Dominant high type
Albedo	Annual	Monthly
LAI r_{smin}	Global constants	Dependent on vegetation type
Root depth Root profile	1 m Global constant	Dependent on vegetation type

High vegetation fraction at T511



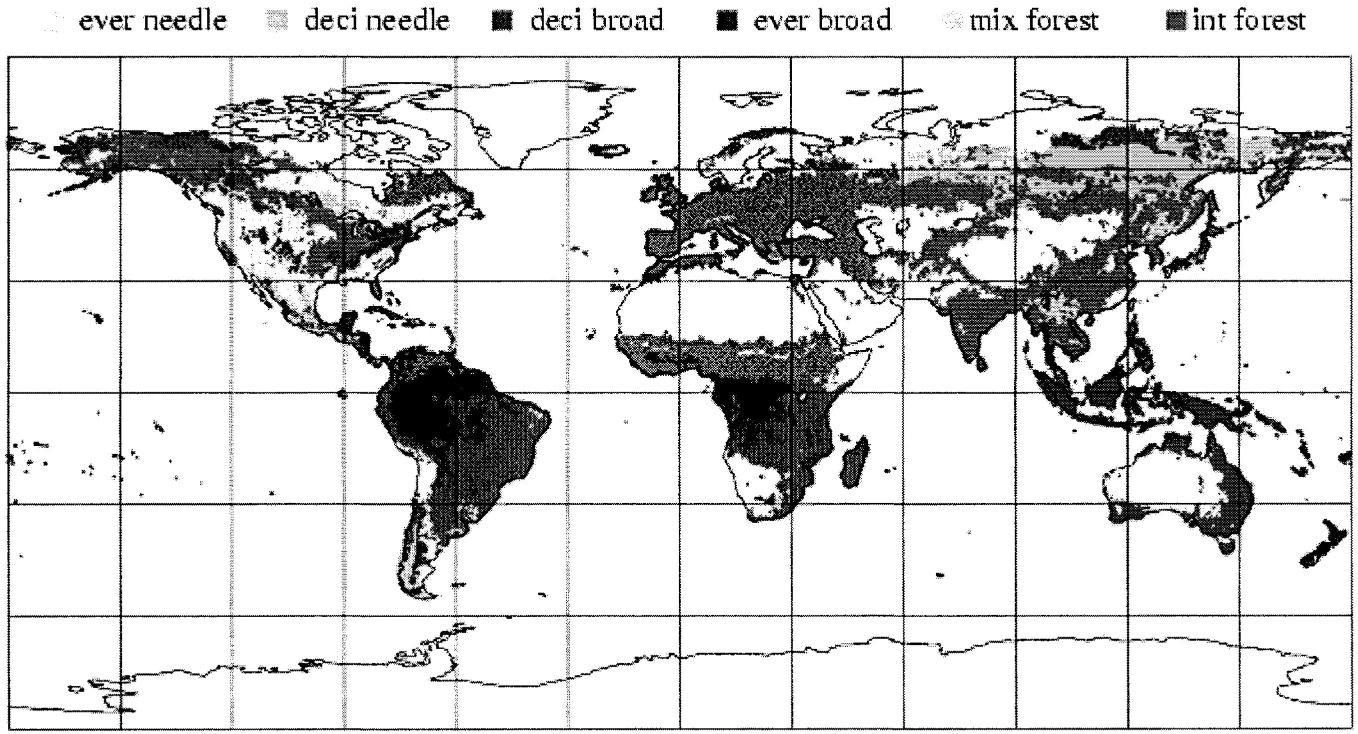
Aggregated from GLCC 1km

Low vegetation fraction at T511



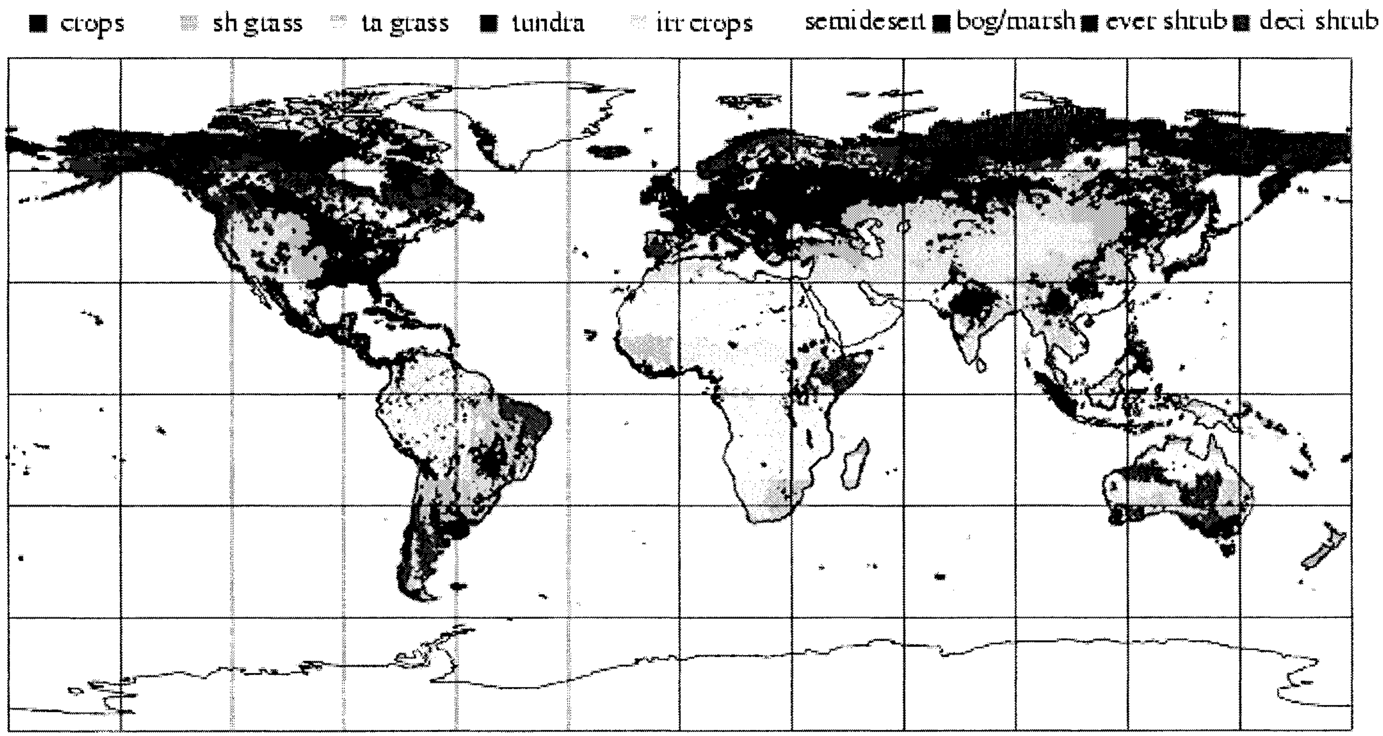
Aggregated from GLCC 1km

High vegetation type at T511



Aggregated from GLCC 1km

Low vegetation type at T511



Aggregated from GLCC 1km

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Ground heat flux

In the absence of phase changes, heat conduction in the soil obeys a Fourier law

$$(\rho C)_g \frac{\partial T_s}{\partial t} = -\frac{\partial G}{\partial z} = \frac{\partial}{\partial z} \lambda_T \frac{\partial T}{\partial z}$$

$(\rho C)_g$ **Soil volumetric heat capacity**

λ_T **Thermal conductivity**

$k = \frac{\lambda_T}{(\rho C)_g}$ **Thermal diffusivity**

For an homogeneous soil,

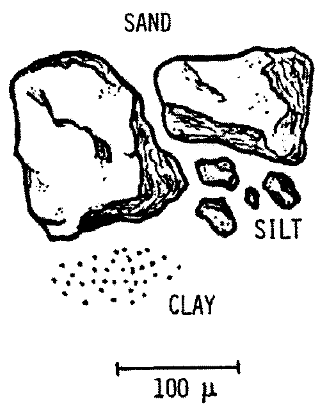
$$\frac{\partial T_s}{\partial t} = k \frac{\partial^2 T}{\partial z^2}$$

Boundary conditions:

- **Top** **Net surface heat flux**
- **Bottom** **No heat flux OR prescribed climate**

Soil science miscellany (1)

- **The soil is a 3-phase system, consisting of**
  - **minerals and organic matter** **soil matrix**
  - **water** **condensate (liquid/solid) phase**
  - **moist air trapped** **gaseous phase**
- **Texture - the size distribution of soil particles**



Hillel 1982

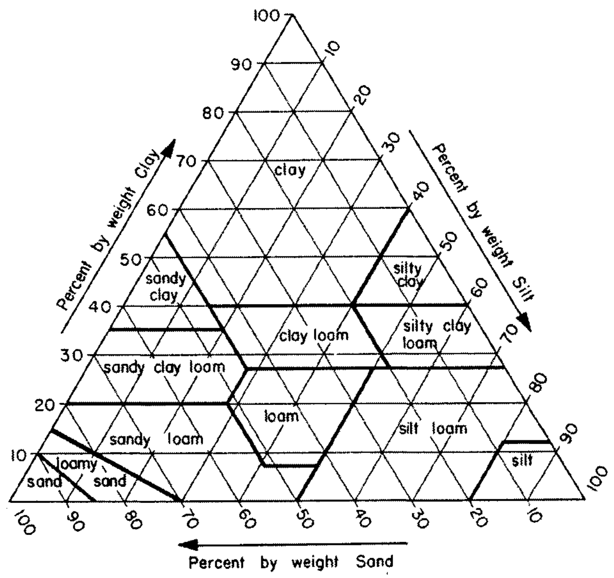


Fig. 3.5. Textural triangle, showing the percentages of clay (below 0.002 mm), silt (0.002–0.05 mm), and sand (0.05–2.0 mm) in the basic soil textural classes.

Soil science miscellany (2)

- Structure - The spatial organization of the soil particles
- Porosity - (volume of maximum air trapped)/(total volume)

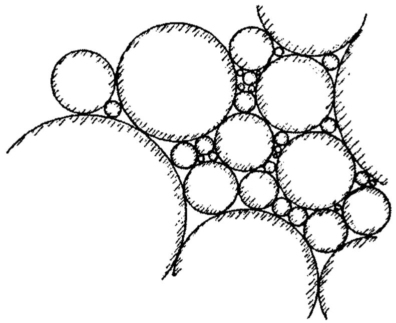
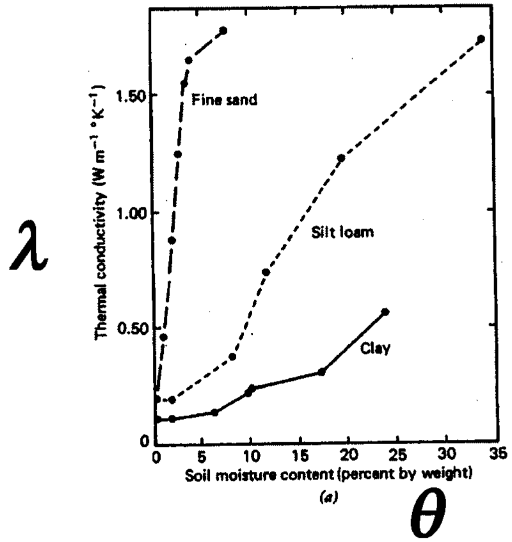


Fig. 4.1. Packing of polydisperse particles (hypothetical).

Hillel 1982

- Composition
- Water content

Soil properties



Rosenberg et al 1983

Table 4.1

Molecular Thermal Properties of Natural Materials[a]

Material	Condition	Mass density ρ (kg m^{-3} × 10^3)	Specific heat c (J kg^{-1} K^{-1} × 10^3)	Heat capacity C (J m^{-3} K^{-1} × 10^6)	Thermal conductivity k (W m^{-1} K^{-1})	Thermal diffusivity α_h (m^2 sec^{-1} × 10^{-6})
Air	20°C, Still	0.0012	1.00	0.0012	0.026	21.5
Water	20°C, Still	1.00	4.19	4.19	0.58	0.14
Ice	0°C, Pure	0.92	2.10	1.93	2.24	1.16
Snow	Fresh	0.10	2.09	0.21	0.08	0.38
Sandy soil	Dry	1.60	0.80	1.28	0.30	0.24
(40% pore space)	Saturated	2.00	1.48	2.98	2.20	0.74
Clay soil	Dry	1.60	0.89	1.42	0.25	0.18
(40% pore space)	Saturated	2.00	1.55	3.10	1.58	0.51
Peat soil	Dry	0.30	1.92	0.58	0.06	0.10
(80% pore space)	Saturated	1.10	3.65	4.02	0.50	0.12

[a] After Oke (1987).

Arya 1988

Diurnal cycle of soil temperature

winter

summer

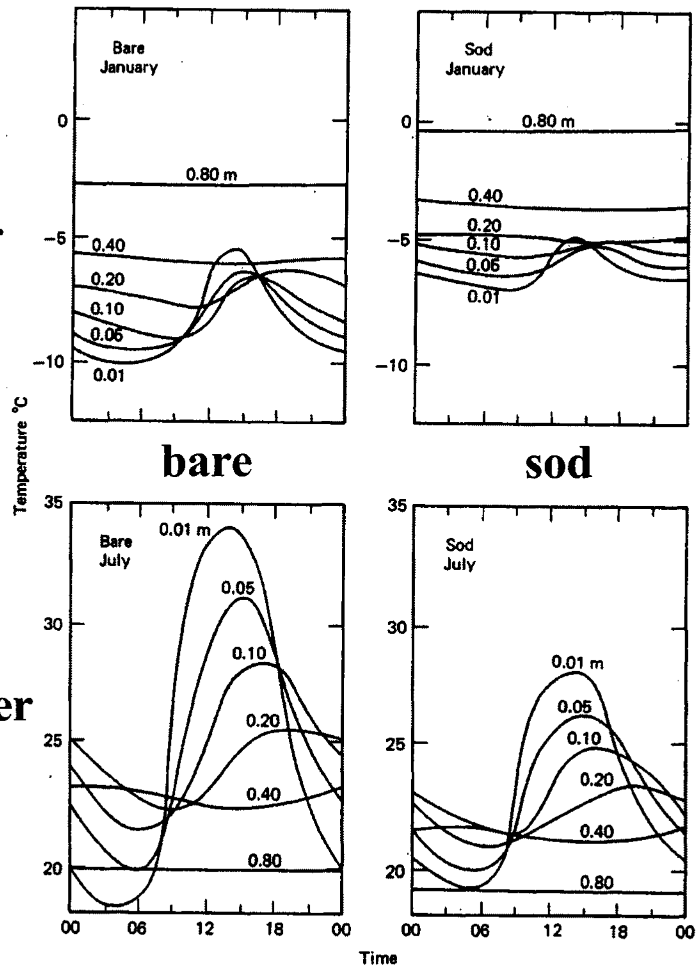
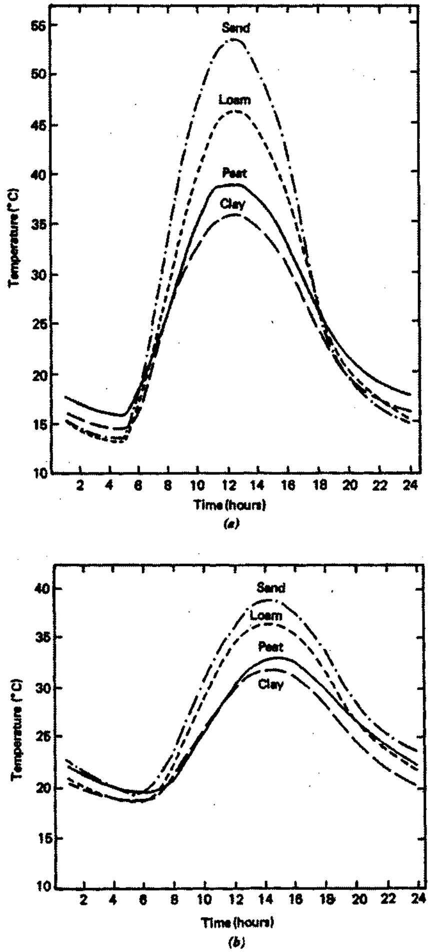


Fig. 2.2 Average hourly soil temperature under bare and sod-covered soil at St. Paul, Minnesota in January (top) and July (bottom). Soil depth is shown in m (after Baker, 1965).

summer



surface

50 cm depth

Fig. 2.6 Daily course of temperature (*a*) at the surface and (*b*) at a depth of 50 mm on clear summer days at Sapporo, Japan (after Yakuwa, 1946).

Rosenberg et al 1983

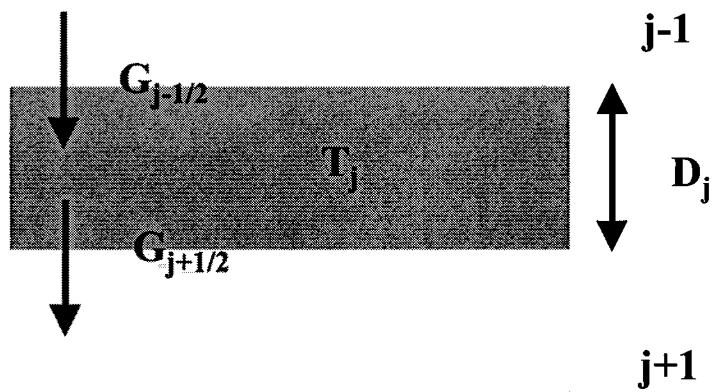
TESSEL

- **Solution of heat transfer equation with the soil discretized in 4 layers, depths 7, 21, 72, and 189 cm.**
- **No-flux bottom boundary condition**
- **Heat conductivity dependent on soil water**
- **Thermal effects of soil water phase change**

TESSEL soil energy equations

$$\frac{(\rho C)_j}{\Delta t}\left(T_j^{n+1} - T_j^{n}\right) = -\frac{\left(G_{j+1/2}^{n+1} - G_{j-1/2}^{n+1}\right)}{D_j} \qquad j = 1,\ldots,4$$

$$G_{j+1/2}^{n+1} = -\lambda_{T,j+1/2}\frac{T_{j+1}^{n+1} - T_j^{n+1}}{0.5\left(D_j + D_{j+1}\right)}$$



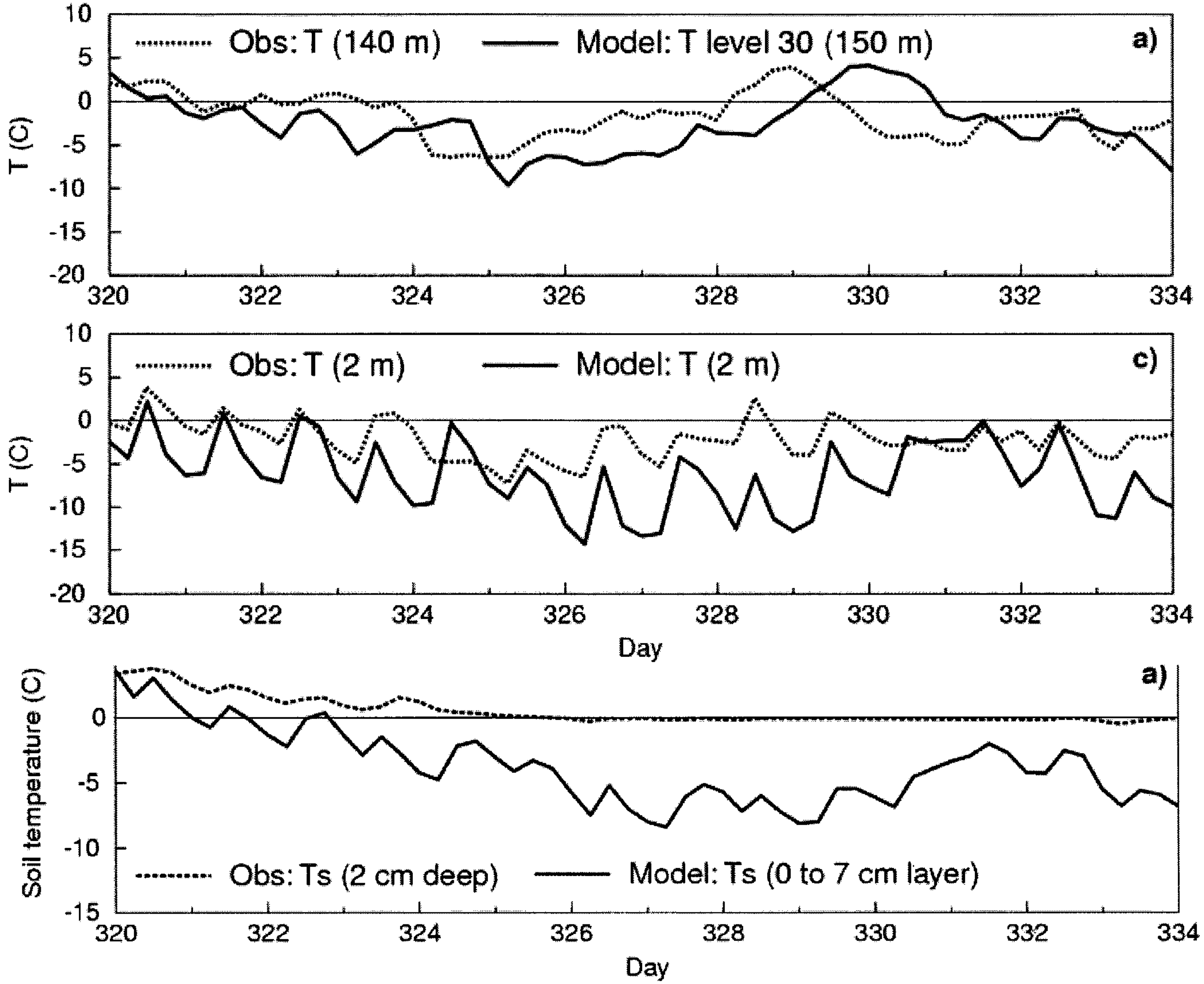
Boundary conditions

$$G_{1/2} = \sum_{i} \Lambda_{sk,i}\left(T_{sk,i} - T_1\right)$$

$$G_{41/2} = 0$$

Case study: winter (1)

Model vs observations, Cabauw, The Netherlands, 2nd half of November 1994



140 m

2 m

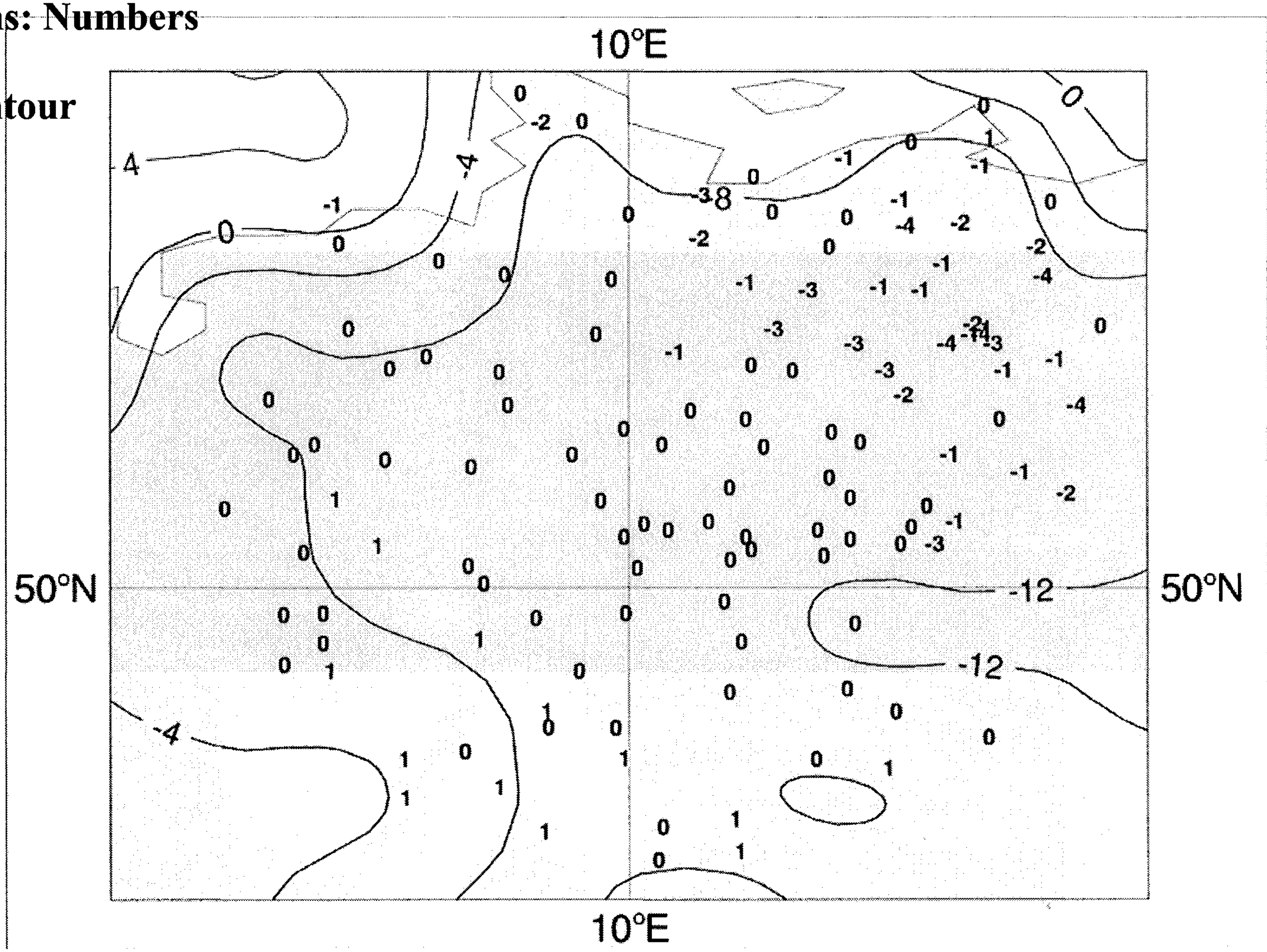
Soil

Case study: winter (2)

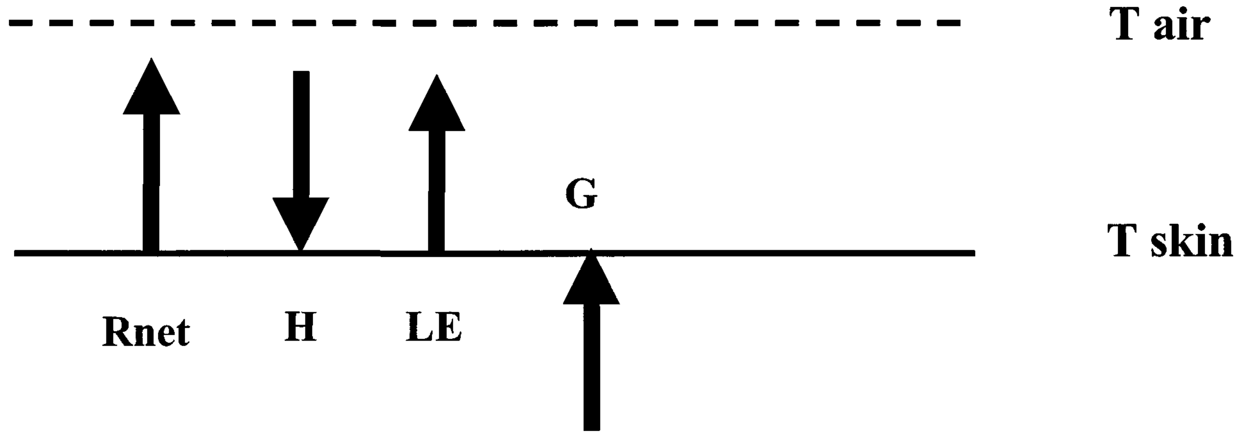
Soil Temperature, North Germany, Feb 1996: Model (28-100 cm) vs OBS 50 cm

Observations: Numbers

Model: Contour



Case study: winter (3)



$$H = \rho C_p |U_{air}| C_{Hn} f(Ri)(T_{air} - T_{sk})$$

Stability functions

Soil water freezing

- **Model bias:**
  - **Net radiation (Rnet) too large**
  - **Sensible heat (H) too small**
    - **But (Tair-Tsk) too large (too large diurnal cycle)**
    - **Therefore f(Ri) problem**
  - **Soil does not freeze (soil temperature drops too quickly seasonally)**

Viterbo, Beljaars, Mahfouf, and Teixeira, 1999: Q.J. Roy. Met. Soc., 125,2401-2426.

Winter: Soil water freezing

Soil heat transfer equation

$$(\rho C)_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \lambda_T \frac{\partial T}{\partial z}$$

Winter: Soil water freezing

Soil heat transfer equation

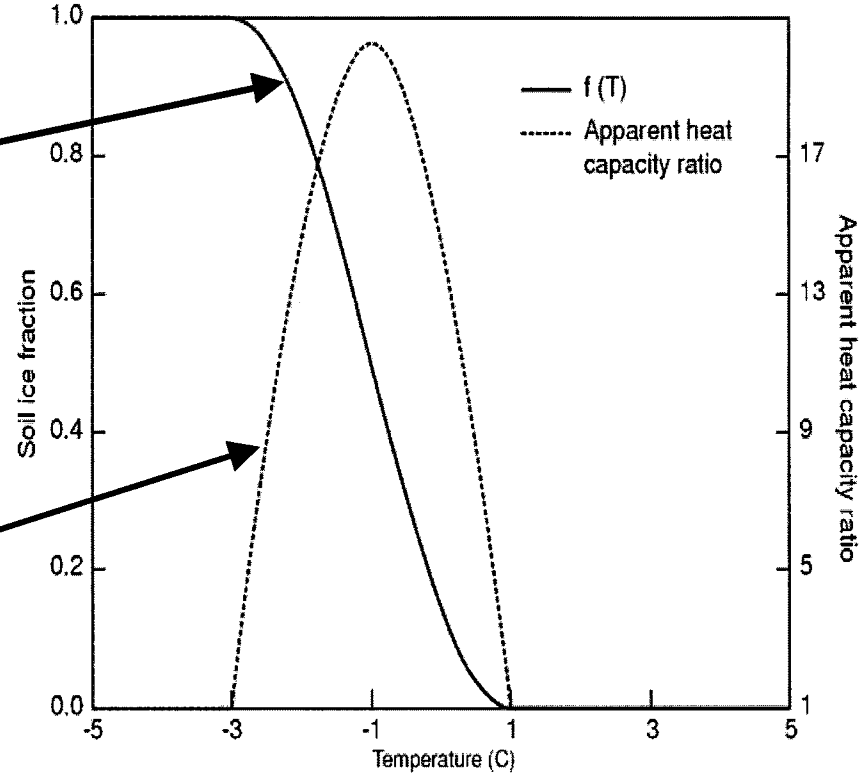
$$(\rho C)_s \frac{fT}{ft} = \frac{f}{fz} \lambda_T \frac{fT}{fz} + L_f \rho_w \frac{f\theta_I}{ft}$$

θ_I **Soil frozen water**

$$\theta_I = \theta_I(T) = f(T)\theta$$

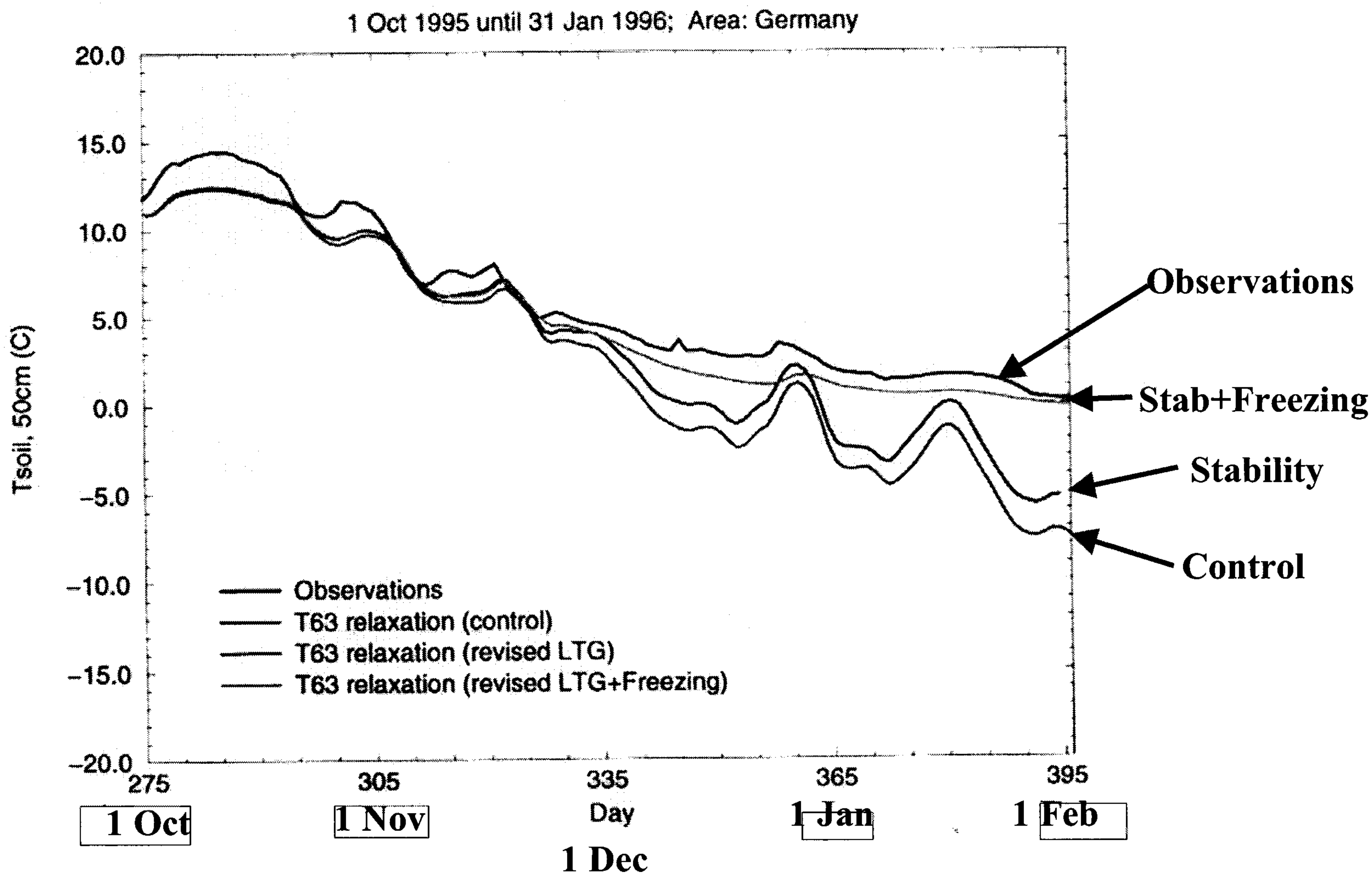
$$(\rho C)_s - L_f \rho_w \theta \frac{ff}{ft} \quad \frac{fT}{ft} = \frac{f}{fz} \lambda_T \frac{fT}{fz}$$

Apparent heat capacity



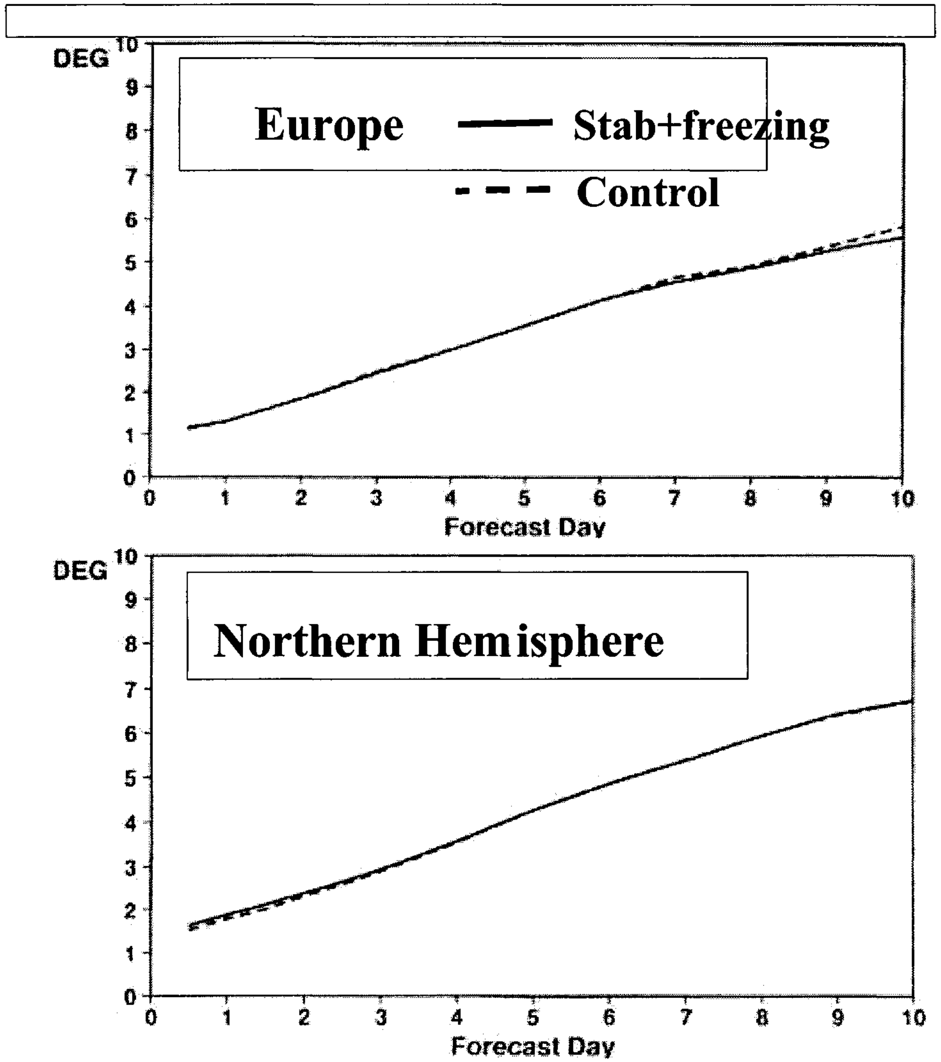
Case study: winter (4)

Germany soil temperature: Observations vs Long model relaxation integrations



Case study: winter (5)

850 hPa T RMS forecast errors



- Soil water freezing acts as a thermal regulator in winter, creating a large thermal inertia around 0 C.
- Simulations with soil water freezing have a near-surface air temperature 5 to 8 K larger than control.
- In winter, stable, situations the atmosphere is decoupled from the surface: large variations in surface temperature affect only the lowest hundred metres and do NOT have a significant impact on the atmosphere.