EXPERIMENTATION WITH A SURFACE MODEL AT THE GERMAN WEATHER SERVICE

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## 1 INTRODUCTION

During the last few years an increasing amount of contributions to the problem of land surface parameterization appeared, mainly initiated by the Greenbelt-Conference (ICSU/WMO, 1981). Elaborate surface models were developed by Dickinson (1984, abbreviated D84 in the forthcoming) and Sellers et al. (1986). At the German Weather Service, a surface model was developed by Bauer et al. (1983, abbreviated B83; 1986). Here the surface temperature is computed by using the extendend force-restore method (Jacobsen and Heise, 1982), which is an optimized version of a two-layer soil model. For vertically homogeneous soil this method allows the simulation of the earth's surface temperature in accordance with the solution of the heat conduction equation, if the forcing consists only of two harmonic components. The two periods $\tau_1$ and $\tau_2$ of these components may be preselected. The depths of the two layers depend mainly on the two periods and - to a smaller extent - on the soil type. In general circulation models, the selection of the diurnal ($\tau_1$) and the annual ($\tau_2$) period is a natural choice. This results in an upper layer thickness of roughly 15 cm. The lower layer then has a thickness of about $\sqrt{\tau_2/\tau_1} \approx$ 19.1 times the thickness of the upper layer.

In the hydrological part of the B83-model the evaporation E both from bare soils and from plants is based on a Budyko-type model as was first introduced into numerical modelling by Manabe (1969). In this parameterization

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potential evaporation Epot is reduced depending on the water content W (Vol %) of the soil:

$$E = \begin{cases} \text{Epot} & ; W \geq \text{Wcrit} \\ \text{Red}\cdot\text{Epot}; & \text{Wzero} < W < \text{Wcrit} \\ 0 & ; W \leq \text{Wzero} \end{cases} \tag{1}$$

with

$$\text{Red} = (W - \text{Wzero})/(\text{Wcrit} - \text{Wzero})$$

The values of Wzero and Wcrit are different for bare soil and plant foliage:

$$\text{Wzero} = \begin{cases} \text{air dryness point ADP (bare soil)} \\ \\ \text{permanent wilting point PWP (plants)} \end{cases}$$

and

$$\text{Wcrit} = \begin{cases} \text{field capacity FC (bare soil)} \\ \\ \text{turgor loss point TLP (plants)} \end{cases}$$

The parameters ADP, PWP, FC and TLP depend on the soil type (see part 3).

The soil water content W is modelled by using two soil layers and an additional surface (interception) store (Figure 1). The surface store evaporates at the potential rate. In order to facilitate the computation of freezing and melting processes, the depths of the two soil layers are like those in the thermal part, although a thickness of about 1 m for the whole model seems to be better for hydrological purposes. Precipitation partly fills the surface store and partly infiltrates into the ground. The maximum possible infiltration rate is determined by soil type and water content. Fluxes between the two active layers and between the lower layer and the subsoil (with constant water content) are simulated by the well known Darcy equation

$$F = k\ (d\psi/dz - 1), \tag{2}$$

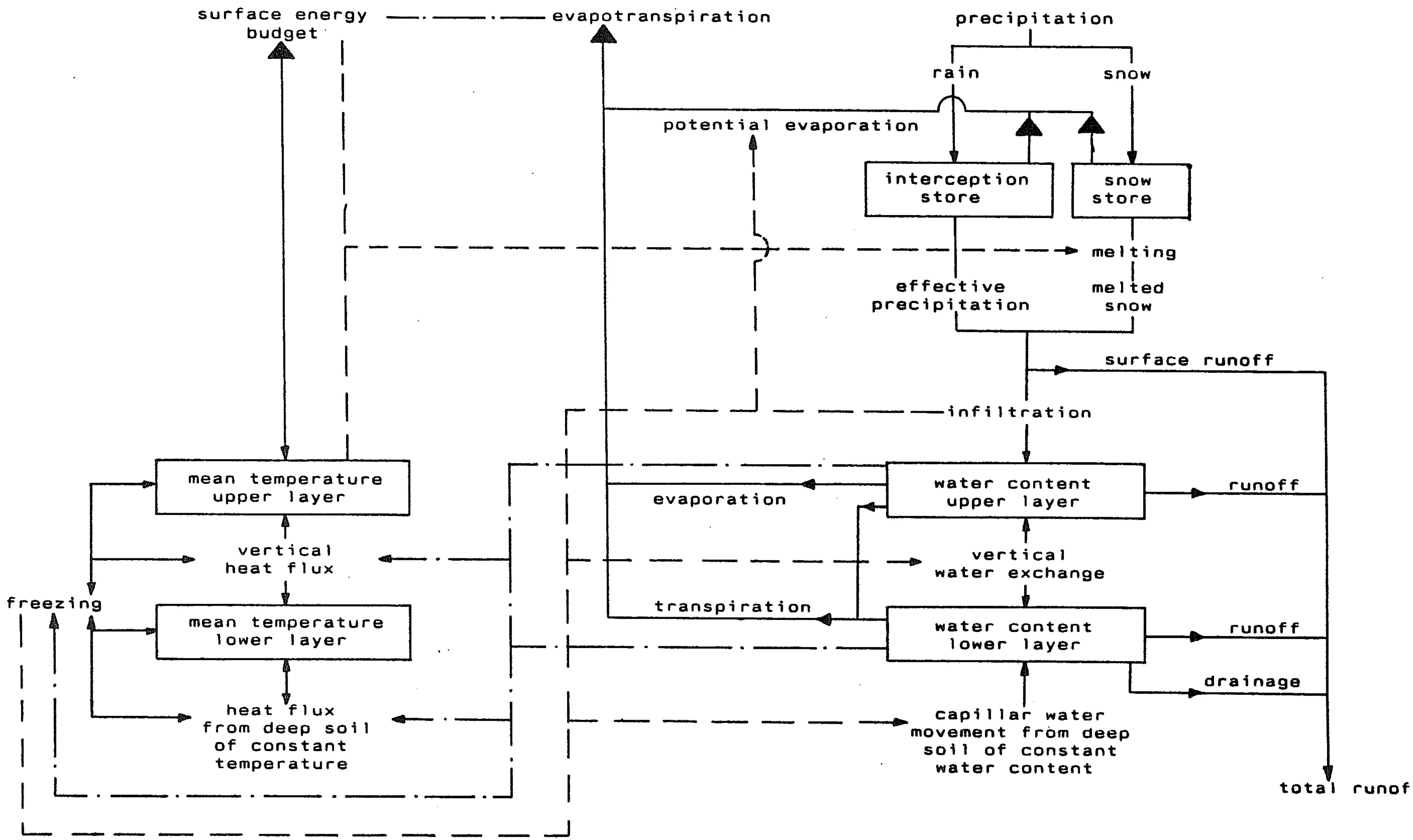



Figure 1: Transport and exchange processes of heat and water in the surface model of Bauer et al., 1983.

i. e. dependent on the vertical suction gradient $d\psi/dz$ and the hydraulic conductivity k. The relation between volumetric water content and suction is taken from Rijtema's (1969) tables. The hydraulic conductivity at each interface between two layers, which strongly depends on the suction, is taken from Rijtema (1969), too. A logarithmic average of the water content values of the adjacent layers is used to compute the suction at the interface.

As D84 pointed out, in reality evaporation is mainly controlled by the water transport capability of the soil/plant system rather than by the water content of the soil, if the water content is not too low. Additionally, over dense vegetation transpiration is not bounded by the potential evaporation as is assumed in Budyko-type models. Therefore, as the transpiration by plants seems to be most seriously affected by the B83 parameterization being too simple, this parameterization has been revised in order to include additionally the effects on transpiration of the foliage density, the radiation balance and the type of vegetation (Heise et al., 1988). The other features of the B83-model were left unchanged.

In chapter 2 the revised surface model is described, eked by an examination of the models sensitivity against variations of the input parameters. Chapter 3 contains some remarks on the implementation of the scheme in a general circulation model and on the preparation of global soil and plant datasets. Chapter 4 provides results of tests within the framework of a one-dimensional model and in chapter 5 some results of January simulations with a GCM are shown.

## 2 THE SURFACE MODEL

As mentioned in the introduction, the general features of the B83-model were left unchanged, only the computation of the transpiration by plants was modified. Figure 1 displays all relevant transport and exchange processes of heat and water and the connections between the thermal and the hydraulic parts of the model.

## 2.1 The revised formula for the transpiration by plants

Following D84 we have the heat flux from a foliage of plants to the air within the canopy

$$H_f = \sigma_f \, \mathrm{LAI} \, r_{la}^{-1} \rho_a c_p (T_f - T_{af}) \qquad (3)$$

and the heat flux from the air within the canopy to the atmosphere

$$H_a = \rho_a c_p C_D |v_a| (T_{af} - T_a). \qquad (4)$$

Here $\sigma_f$ is the ground fraction covered by plants, LAI the leaf area index, $r_{la}$ an aerodynamic resistance coefficient for heat and vapour flux from leaves. The other symbols have their usual meaning. The indices denote values at the foliage (f), the air within the canopy (af) and the air above the canopy (a), respectively.

Similarly, the fluxes of water vapour are

$$E_f = r' \sigma_f \, \mathrm{LAI} \, r_{la}^{-1} \rho_a (q_{sat}(T_f) - q_{af}) \qquad (5)$$

and

$$E_a = \rho_a C_D |v_a| (q_{af} - q_a), \qquad (6)$$

where r' is the reduction of the transpiration by dry leaves compared to wet leaves.

Neglecting the fluxes of sensible heat and water vapour from the ground to the air within the canopy and assuming a vanishing storage capacity of the air within the canopy, equations (3) through (6) may be combined to yield

$$T_{af} = (C_F T_f + C_A T_a)/(C_F + C_A) \qquad (7)$$

$$H_a = \rho_a c_p C_F (T_f - (C_F T_f + C_A T_a)/(C_F + C_A)) \qquad (8)$$

$$q_{af} = (C_A q_a + C_V q_{sat}(T_f))/(C_A + C_V) \qquad (9)$$

$$E_a = \rho_a (q_{sat}(T_f) - q_a) C_A C_V / (C_A + C_V), \tag{10}$$

with the abbreviations

$$C_A = C_D |v_a| ,$$

$$C_F = \sigma_f \, \mathrm{LAI} \, r_{la}^{-1}$$

and

$$C_V = r' C_F .$$

Expanding $q_{sat}(T_f)$

$$q_{sat}(T_f) \approx q_{sat}(T_a) + (T_f - T_a)(\partial q_{sat} / \partial T)_{T=T_a} ,$$

$T_f$ may be replaced using $E_a$, and from (8) we have

$$H_a = c_p \frac{C_A + C_V}{C_A + C_F} \frac{C_F}{C_V} \left(\frac{\partial q_{sat}}{\partial T}\right)^{-1} E_a - \rho_a c_p \frac{C_A C_F}{C_A + C_F} \left(\frac{\partial q_{sat}}{\partial T}\right)^{-1} \Delta q \tag{11}$$

with

$$\Delta q = q_{sat}(T_a) - q_a .$$

If additionally we require the net radiation balance $R_n$ to be balanced by the sum of sensible and latent heat flux, the flux of water vapour is

$$E_a = \frac{R_n / L + \rho_a \frac{C_A C_F}{C_A + C_F} Be \, \Delta q}{1 + \frac{C_A + C_V}{C_A + C_F} \frac{C_F}{C_V} Be} \tag{12}$$

with the abbreviation

$$Be = c_p / L \left(\frac{\partial q_{sat}}{\partial T}\right)^{-1} .$$

As soon as the transport coefficients $C_A$, $C_F$ and $C_V$ are determined, (12) may be used as a parameterization for the transpiration by plants.

The accuracy of (12), especially the accuracy of the linear expansion of $q_{sat}(T)$, was recently investigated by Paw U and Gao (1988). Compared to measurements they found a mean relative error of 8.8 %. This seems to be tolerable for parameterization purposes.

## 2.2 The parameterization of the transfer coefficients

The atmospheric values (Index a) in the expressions for the transfer coefficients have to be determined from the lowest model layers by some interpolation formula, depending on the vertical structure of the numerical model. Using the method developed by Kasahara and Washington (1967), estimates of near surface winds and temperatures may be calculated from values given near the top of the Ekman-layer. The specific humidity is computed assuming a constant relative humidity throughout the lowest model level.

a) $C_D$ is the normal drag coefficient, any appropriate value may be used, here $C_D = 0.0012$ is chosen, consequently

$$C_A = 0.0012 \, |v_a|$$

b) The coefficient $r_{la}$ describes the resistance to heat and water vapour transfer from the leaves into the air between the leaves. D84 uses

$$r_{la}^{-1} = C_f \sqrt{U_{af} / D_f}$$

$D_f$ is the characteristic dimension of the leaves in the direction of the wind flow and $C_f$ a constant of proportionality ($C_f = 0.01$ m/s$^{1/2}$). Since both these parameters are hard to assess, they are combined to yield one rather questionable constant

$$C' = C_f / \sqrt{D_f} = 0.05 \, (\mathrm{m/s})^{1/2} .$$

$U_{af}$, the wind velocity effective on the leaves is estimated from

$$U_{af} = u_*,$$

the friction velocity $u_*$ being taken from the boundary layer parameterization.

c) The coefficient r' describes the reduction of the transpiration from dry leaves compared to wet leaves. Since the transpiration from wet leaves is accounted for by using the interception store, D84's formula for r' may be simplified to read

$$r' = f_d \frac{r_{ea}}{r_{ea} + r_s} , \tag{13}$$

where $f_d$ is the transpiring fraction of the canopy. The crucial parameter in r' is $r_s$, the resistance encountered by water transfer from the soil to outside the leaves. $r_s$ has to take into account all - or at least the most important - factors that control transpiration from plants. Following B83, D84 and the concept used in the ECMWF-model (see Research Manual 3, ECMWF Forecast Model, Physical Parameterizations, 2nd Edition, 1/88) the following form is adopted:

$$r_s = r_{min} \left[ \mathrm{RAD} \cdot F(W) \right]^{-1} , \tag{14}$$

with

$$\mathrm{RAD} = \frac{PAR/R_c + r_{min}/r_{max}}{PAR/R_c + 1}$$

and

$$F(W) = \begin{cases} 1 & ; W \geq W_{cr} \\ \dfrac{W - PWP}{W_{cr} - PWP} & ; PWP < W < W_{cr} \\ 0 & ; W \leq PWP \end{cases}$$

In this parameterization for $r_s$, PAR denotes the photosynthetically active radiation (approximated by the net radiation in the first spectral interval 0.215$\mu$ - 0.685$\mu$ of the Hense et al., 1982, radiation scheme) with the normalization factor $R_c$ = 25 W/m$^2$. The maximum and minimum values of the stomatal resistance are quite uncertain. According to Federer (1979), D84 and Sellers and Dorman (1987) $r_{min}$ = 100 s/m and $r_{max}$ = 4000 s/m are used. The rather small value for $r_{min}$ was chosen when one-dimensional tests revealed much too small transpiration rates with $r_{min}$ = 250 s/m, which was the first guess from the values given in the literature. The last parameter to be determined is the critical water content $W_{cr}$. In order to keep the parameterization as simple as possible, $W_{cr}$ is computed from the field capacity $W_{cr} = 0.6 \cdot FC$, instead of using the turgor loss point, which itself would have to be parameterized.

## 2.3 Sensitivity of the parameterization with respect to input parameters

The parameterization of plant transpiration contains a lot of more or less well known parameters. In this chapter the sensitivity of (12) with respect to these parameters shall be examined. In order to do this, (12) is differentiated with respect to the different input parameters. The numerical evaluation of the results is based on a standard case that uses the following values:

$|v_a| = 6$ m/s
$C_D = 0.012$
$u_* = 0.6$ m/s
Be = 0.5
$R_n = 300$ W/m$^2$
$R_c = 25$ W/m$^2$
$\rho_a = 1$ kg/m
$\Delta q = 0.004$
$\sigma_f = 1$
LAI = 5
$f_d = 1$
W = 8 Vol %
FC = 19.6 Vol %
PWP = 4.2 Vol %
$W_{cr} = 11.8$ Vol %
$r_{min} = 100$ s/m
$r_{max} = 4000$ s/m
$r_{la} = 25.82$ s/m
$C_A = 0.06$ m/s
$C_F = 0.1936$ m/s
F(W) = 0.5026
$r_s = 263$ s/m

PAR/$R_c$ is replaced by $R_n/(4 \cdot R_c)$ in this evaluation.

In this case the transpiration computed from (12) is $7.82 \cdot 10^{-5}$ kg/(m$^2$s) (or 6.76 mm/d).

The following Table 1 gives the rate of change of $E_a$ with respect to variations of the different parameters and the absolute and relative differences in $E_a$ for a 10 % modification of each parameter:

| Parameter | Rate of change $\partial E_a / \partial$(Parameter) | Change of $E_a$ for a 10% Modification of the Parameter kg/(m$^2$s) | % |
|---|---|---|---|
| W | $1.61 \cdot 10^{-5}$ kg/(m$^2$s Vol%) | $1.29 \cdot 10^{-5}$ | 16.50 |
| $R_n$ | $1.48 \cdot 10^{-7}$ kg/(Ws) | $4.44 \cdot 10^{-6}$ | 5.68 |
| LAI | $6.92 \cdot 10^{-6}$ kg/(m$^2$s) | $3.46 \cdot 10^{-6}$ | 4.42 |
| $\Delta q$ | $8.50 \cdot 10^{-3}$ kg/(m$^2$s) | $3.40 \cdot 10^{-6}$ | 4.35 |
| $\lvert v_a \rvert$ | $- 1.48 \cdot 10^{-7}$ kg/m$^3$ | $- 8.90 \cdot 10^{-8}$ | 0.11 |
| $W_{cr}$ | $- 4.60 \cdot 10^{-6}$ kg/(m$^2$s Vol%) | $- 5.41 \cdot 10^{-6}$ | 6.92 |
| $r_{min}$ | $- 3.46 \cdot 10^{-7}$ kg/(ms$^2$) | $- 3.46 \cdot 10^{-6}$ | 4.42 |
| $R_c$ | $- 3.37 \cdot 10^{-7}$ kg/(Ws) | $- 8.42 \cdot 10^{-7}$ | 1.08 |
| $r_{max}$ | $- 7.20 \cdot 10^{-11}$ kg/(ms$^2$) | $- 2.88 \cdot 10^{-8}$ | 0.04 |
| $r_{ea}$ | $8.95 \cdot 10^{-9}$ kg/(ms$^2$) | $2.31 \cdot 10^{-8}$ | 0.03 |

Table 1: The dependence of the plant transpiration on the different parameters in equation (12).

The last result in the upper half of Table 1 - the dependence of the transpiration rate on the wind velocity - looks a bit strange, normally one would expect increasing transpiration with increasing wind velocity. A closer inspection of this point revealed that the extremely intricate dependence of $E_a$ on $|v_a|$ due to the prescribed balance of different energy fluxes causes this unrealistic result. At the same time this behaviour indicates the values in the table being limited to the standard case. With LAI=6, e.g., the transpiration increases with wind speed. But at least the dependence of the transpiration rate on the wind speed is small.

Despite of this problem, Table 1 indicates

i) the large influence of the soil water content W on $E_a$ and

ii) a worrying large dependence of the transpiration rate on the critical water content $W_{cr}$ and the minimum stomatal resistance $r_{min}$ .

Although the relative error for a 10% change is larger for $W_{cr}$ than for $r_{min}$ , this last parameter is much more critical, as its value is much more uncertain. These two parameters - and to a smaller extent the normalization factor $R_c$ - have to be determined very carefully in the future to avoid large errors in the transpiration computation. Obviously they depend on the genus of the plants forming the canopy. This makes the preparation of complete datasets for global models an even more troublesome task.

## 3 ADAPTATION OF THE SURFACE MODEL TO A GENERAL CIRCULATION MODEL

There are a lot of problems to be solved prior to using the surface model in a general circulation model. Only the more important problems will be mentioned here.

### 3.1 Area-averaged mean values of sensible and latent heat fluxes

The transpiration formula (12) is assumed to be nothing more than a new formula for the parameterization of plant transpiration. It does not influence other parameterizations. That is, the sensible heat flux is computed in the usual manner from the boundary layer parameterization of the numerical model without taking into account (3) and (4). Additionally, a temperature tendency for the soil is computed from the extended force-restore method mentioned in the introduction, using the area-mean values of radiation balance, sensible and latent heat fluxes. This is in contradiction to (12), where the net forcing of the surface is assumed to be zero. But applying this condition strictly would lead to soil temperatures being constant in time for completely plant covered grid elements. The latent heat flux over bare soil is taken from (1), where Epot is computed from the boundary layer formulation of the numerical model. The results of (1) and (12) are averaged, weighted by the ground fraction covered by plants.

### 3.2 Maximum interception store and dry fraction of the canopy

Only an area-mean value of the maximum storage capacity of the interception store is prescribed:

$$SI_{max} = SI_0(1 - \sigma_f + \sigma_f LAI)$$

with $SI_0$ = 0.2 mm. As the maximum value of LAI is about 10, the range of the interception store is 0.2 mm $\leq$ SI $\lesssim$ 2 mm.

The dry fraction of the canopy is computed from

$$f_d = 1 - (SI/SI_{max})^2,$$

where SI is the actual content of the interception store.

### 3.3 Soil types

From the soil map of the world (FAO/UNESCO, 1971-78) the dominating soil type was derived, basically using a $1^\circ \times 1^\circ$ grid. Eight soil types were distinguished:

1) Ice
2) Rock formation
3) Sand
4) Sand/Loam
5) Loam
6) Loam/Clay
7) Clay
8) Peat

The basic characteristics of the soils have to be prescribed for these types. Firstly, the heat capacity is computed as the sum of the contributions from the dry soil and the water and the ice content of the soil:

$$\rho c = \rho c_d (1 - PV/100) + \rho c_{WAT} W/100 + \rho c_{ICE} W_I/100$$

Here $\rho c_d$ is the heat capacity of dry soils (cf. Table 3), $\rho c_{WAT} = 4.18 \cdot 10^6$ J/(m$^3$K) and $\rho c_{ICE} = 1.92 \cdot 10^6$ J/(m$^3$K) are the heat capacities of water and ice, respectively, (Linke and Baur, 1970). $W_I$ is the ice content in Vol %.

| a W/(m K) | b W/(m K) | | |
|---|---|---|---|
| 0.30 | 4.80 | $W \leq 0.25$ FC | anorganic soils |
| 1.33 | 0.67 | 0.25 FC < W < FC | anorganic soils |
| 2.00 | 0.00 | $W \geq$ FC | anorganic soils |
| 0.05 | 0.45 | | organic soils |

Table 2: Constants for the computation of the thermal conductivity.

| Soil type | $\rho c_d$ $10^6$ J/(m$^3$K) | PV Vol % | FC Vol % | PWP Vol % | ADP Vol % |
|---|---|---|---|---|---|
| Ice | 1.92 | - | - | - | - |
| Rock formation | 2.10 | - | - | - | - |
| Sand | 1.93 | 35.6 | 19.6 | 4.2 | 1.2 |
| Sand/Loam | 2.10 | 40.2 | 26.8 | 7.6 | 2.3 |
| Loam | 2.25 | 44.8 | 34.0 | 11.0 | 3.5 |
| Loam/Clay | 2.36 | 47.4 | 40.2 | 18.4 | 5.0 |
| Clay | 2.48 | 50.0 | 46.3 | 25.7 | 6.5 |
| Peat | 3.45 | 85.5 | 50.0 | 26.5 | 9.8 |

Table 3: Basic thermal and hydraulic properties of the eight soil types (heat capacity of the dry soil $\rho c_d$, porosity PV, field capacity FC, permanent wilting point PWP and air dryness point ADP).

The thermal conductivity is parameterized following Benoit (1976) and Van Wijk and De Vries (1966):

$$\lambda_T = a + b \cdot W/FC$$

with a and b as given in Table 2.

According to Linke and Baur (1970) we use $\lambda_T$ = 2.26 W/m K for ice and $\lambda_T$ = 2.45 W/m K for rock formation.

The basic hydraulic properties of the soil are characterized by certain values of the water suction (e. g. Scheffer and Schachtschabel, 1979). Using the conversion between suction and volumetric water content given by Rijtema (1969), the values in Table 3 were derived.

### 3.4 Plants

For evaluating (12) a number of global datasets are required. In co-operation with the University of Osnabrück (Lieth and Esser, personal communication) global datasets of the fractional area covered by plants, the leaf area index and and the root depths were constructed. (For details see Heise et al., 1988.) No attempt has been made to prescribe horizontally varying values of the minimum stomatal resistance and the critical water content, which proved to be very influential parameters of the transpiration formula (cf. Table 1). As an example Figure 2 shows the global distribution of the annual mean values of the ground fraction covered by plants and of the leaf area index.

The B83 parameterization additionally requires the value of the turgor loss point TLP. This parameter is computed according to Denmead and Shaw (1962):

$$TLP = PWP + (FC - PWP)\,(0.81 + 0.121\,\mathrm{arctg}(4.75(Epot/PE_N - 1))),$$

with

$$PE_N = 4.75 \text{ mm/day}.$$

## 4 ONE-DIMENSIONAL TESTS

The one-dimensional model used for the first tests of the transpiration parameterization includes all the parameterizations of the general circulation

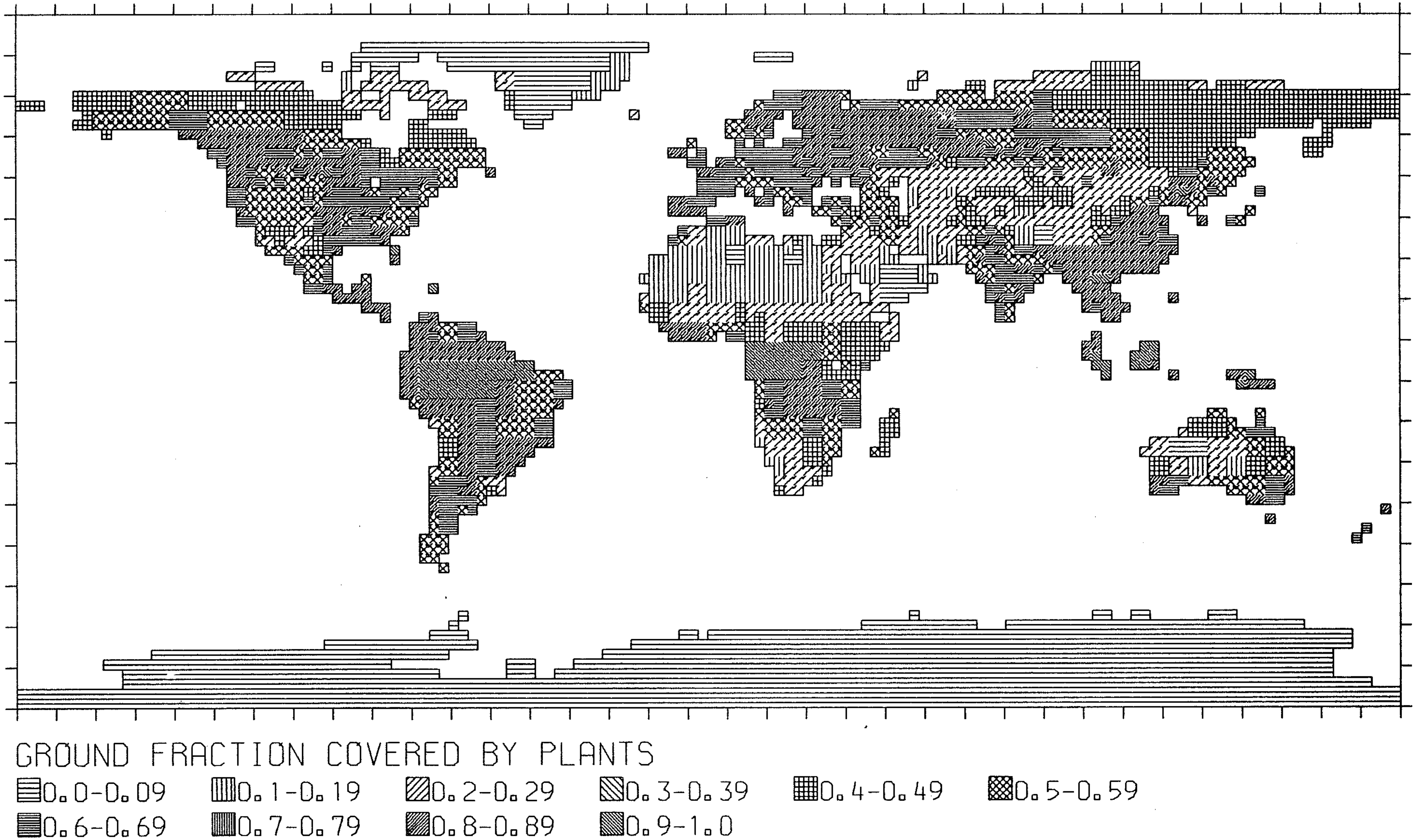



Figure 2a: Global distribution of annual mean values of basic plant data. Ground fraction covered by plants.

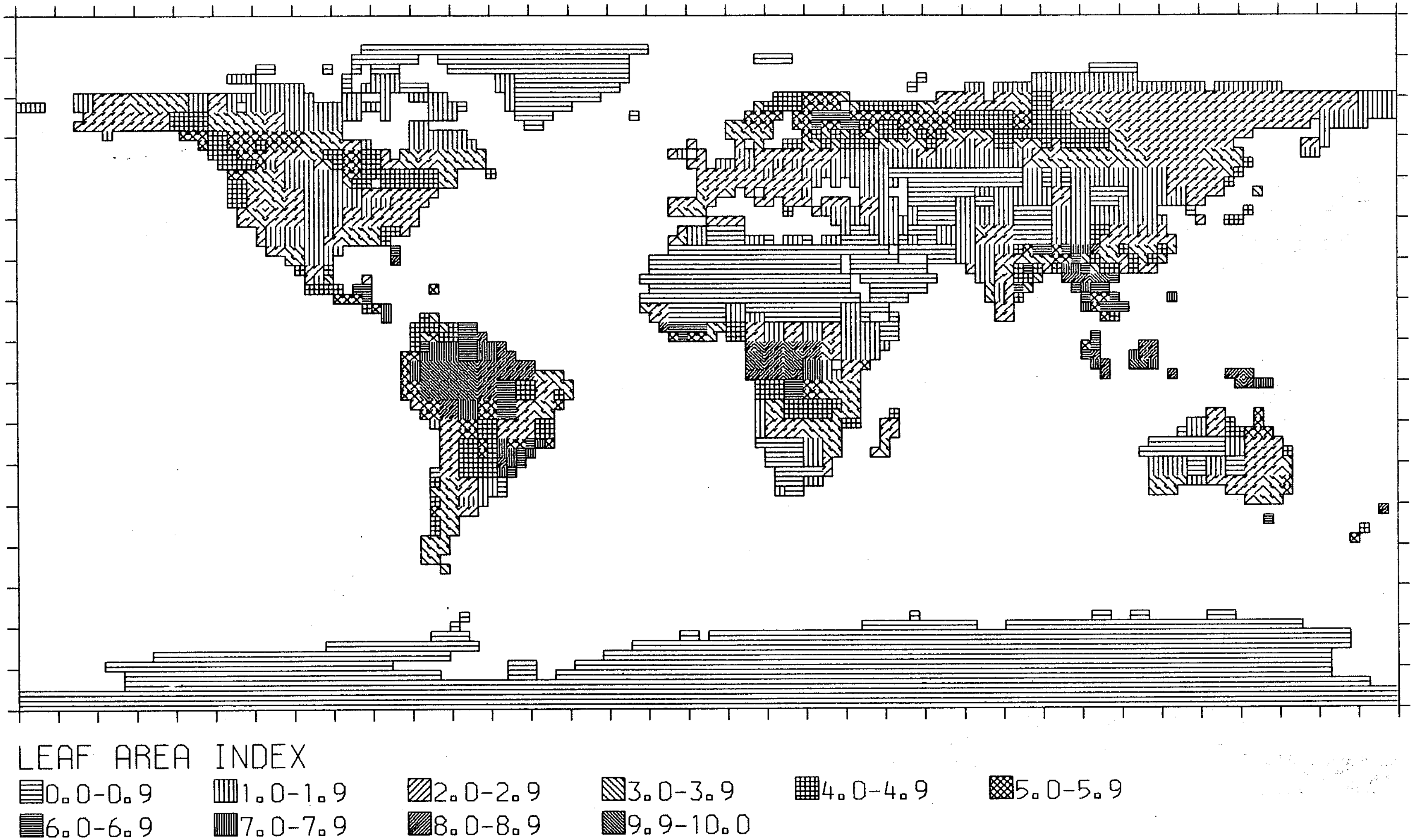



**Figure 2b: Global distribution of annual mean values of basic plant data. Leaf area index.**

model of the German Weather Service. Especially important concerning the surface model are the following components: i) The boundary layer treatment is a somewhat simplified version of the Deardorff (1972) proposal (with a constant boundary layer height of 1 km and a vanishing veering of the wind in the vertical). ii) The convection parameterization basically follows Kuo (1965) with the reduction of the moistening given by Carr and Bosart (1978). iii) The radiation calculation is taken from Hense et al. (1982).

As the one-dimensional model cannot take into account horizontal advection, the computed tendencies in the atmosphere might be too large. To overcome this problem, a simple artificial horizontal advection is prescribed for all prognostic variables:

$$\partial \eta / \partial t = (\eta_{t=0} - \eta)/\tau .$$

A time-constant $\tau$ = 6 hours is used.

In the experiments described below the declination of the sun is 20° and a latitude of 50° N is assumed. Soil type sand is prescribed.

Four different versions of surface models were used:
1) SOIL : The ground fraction covered by plants is zero (bare soil case).
2) B83 : The B83 surface model is used.
3) H88L8: The surface model as described in this text is used with LAI=8.
4) H88L1: As H88L8, but LAI=1.

The bare soil case is independent of using the B83-model or the model of Heise et al. (1988). Exept for this case, the ground fraction covered by plants is 1.

The initial profiles of temperature and humidity in the ten layers of the atmospheric part of the model are chosen as given in Table 4:

| p (hPa) | 50 | 150 | 250 | 350 | 450 | 550 | 650 | 750 | 850 | 950 |
|---|---|---|---|---|---|---|---|---|---|---|
| T (K) | 215 | 230 | 242 | 255 | 267 | 273 | 278 | 283 | 288 | 283 |
| RH (%) | 5 | 15 | 25 | 35 | 45 | 55 | 65 | 57 | 85 | 95 |

Table 4: Initial profiles of temperature and relative humidity used in the one-dimensional tests.

A ground temperature of 280 K is initially prescibed for all layers. No clouds are assumed in the radiation calculation. The experiments start at midnight and are conducted for two days each.

The most interesting part of a comparison of the different versions of the surface model is the differences of evaporation and precipitation. Table 5 shows integrated results of a series of experiments with different values of the initial water content of the soil layers. (The initial value of the interception store was set to zero in all experiments.)

| W (t=0) | Model | $\Delta T$ (K) | $\Delta W$ (mm) | Evap. (mm) | Prec. (mm) |
|---|---|---|---|---|---|
| 5 Vol % | SOIL | 2.003 | - 3.354 | 3.375 | 0.022 |
| | B83 | 2.125 | - 2.220 | 2.246 | 0.026 |
| | H88L8 | 1.963 | - 5.328 | 5.423 | 0.096 |
| | H88L1 | 2.354 | - 1.912 | 1.912 | 0.000 |
| 10 Vol % | SOIL | 1.687 | - 5.675 | 5.793 | 0.131 |
| | B83 | 1.522 | - 6.682 | 7.231 | 0.563 |
| | H88L8 | 1.583 | - 8.891 | 10.844 | 1.966 |
| | H88L1 | 1.873 | - 5.931 | 6.030 | 0.111 |
| 15 Vol % | SOIL | 1.503 | - 6.586 | 7.201 | 0.747 |
| | B83 | 1.366 | - 6.715 | 8.667 | 2.081 |
| | H88L8 | 1.512 | - 9.179 | 11.281 | 2.235 |
| | H88L1 | 1.775 | - 6.470 | 6.691 | 0.353 |
| 20 Vol % | SOIL | 1.390 | -13.965 | 8.077 | 1.749 |
| | B83 | 1.336 | -14.063 | 8.649 | 2.177 |
| | H88L8 | 1.470 | -16.574 | 11.313 | 2.338 |
| | H88L1 | 1.729 | -13.884 | 6.711 | 0.458 |

Table 5: Time integrated values of the changes of temperature and water content of the soil layers and evaporation and precipitation for the four different versions of the soil model with four different values of the initial water content.

With all values of the initial water content the evaporation and the decrease of the soil water content are largest in H88L8. A reduction of the leaf area index causes a significant decrease of the transpiration. A comparison of H88L1 with the bare soil case suggested the reduction of the minimum stomatal resistance mentioned in chapter 2.2. Nevertheless, evaporation in the bare soil case is sometimes larger than in H88L1, leading to the conjecture that the bare soil evaporation is too large. A next step in the development of the soil model will therefore be a reformulation of the evaporation from bare soils.

In general, the relative differences between the soil model versions decrease with increasing water content of the soil. Thus, the correct formulation of the model seems to be most important in the transition zones between completely dry and completely wet regions, where the water budget of the soil may seriously be distorted by errors in the evaporation calculation.

Of course, the Bowen ratio (not shown) decreases with increasing soil water content. But with all the four different values of initial water content, the version H88L8 has a much smaller value than the others. In the dry soil case (5 %) there is a reduction from 1.6 (B83) to 0.5 (H88L8). For the wet soil case (20 %) there is still a reduction from 0.12 (B83) to 0.08 (H88L8). The relative differences of the sums of the heat fluxes are much smaller than those of the Bowen ratios. Accordingly, the differences in the temperature reactions are rather small, whereas the differences in the evaporation and precipitation rates are exceptionally high.

A characteristic difference in the behaviour of the original (1) and the revised (12) transpiration formulas is depicted in Figure 3. Using the original version, the latent heat flux becomes positive by and large in coincidence with the stratification becoming unstable. This is in contrast to the revised version, where the latent heat flux normally has the same sign as the radiation balance. Accordingly, the difference-plot shows largest values during the first hours after sunrise. Lateron the smaller sensible heat flux partly compensates for the larger latent heat flux. After sunrise the net forcing of the ground (figure 4) is much smaller with the revised version, causing temperature differences of some 3.5 K in the morning. During night there is a larger latent heat flux from the atmosphere to the ground in H88L8, therefore temperatures are warmer by about 2 K.

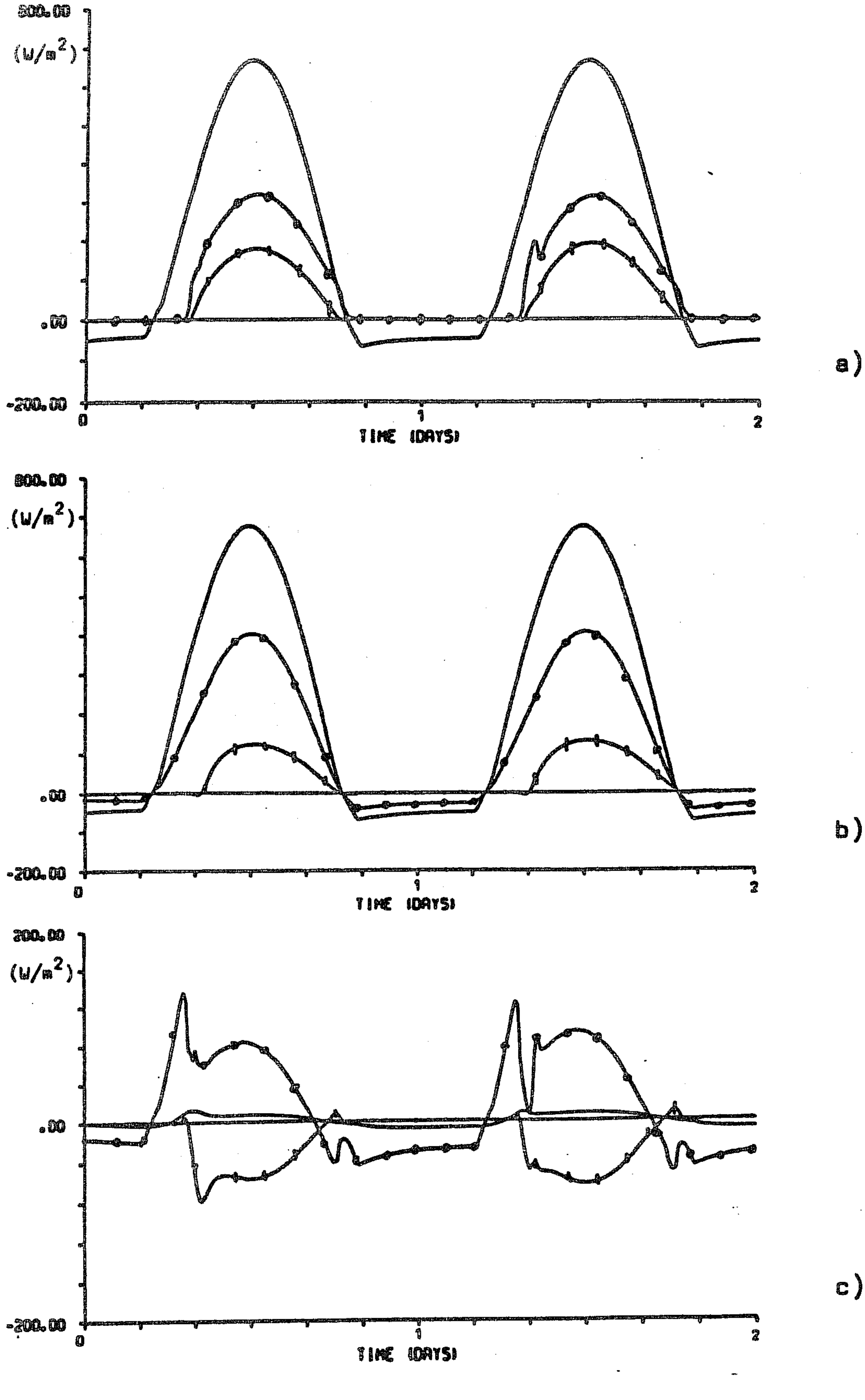
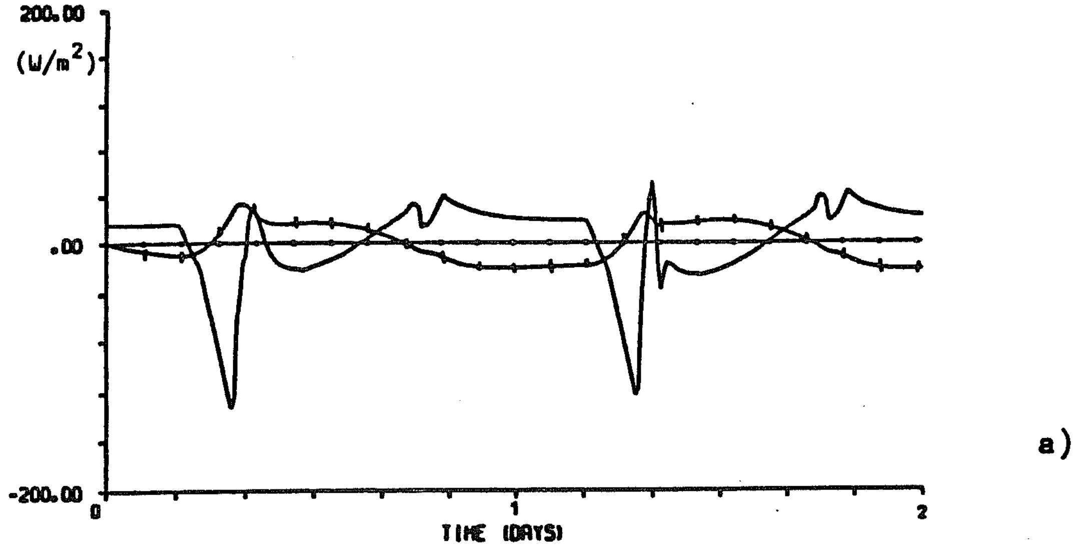



Figure 3: Energy fluxes at the surface calculated in a one-dimensional model:
—— Radiation balance, ⟷ sensible heat flux, ⊙—⊙ latent heat flux.
a) B83-version, b) H88L8-version, c) difference H88L8-B83.
See text for details.



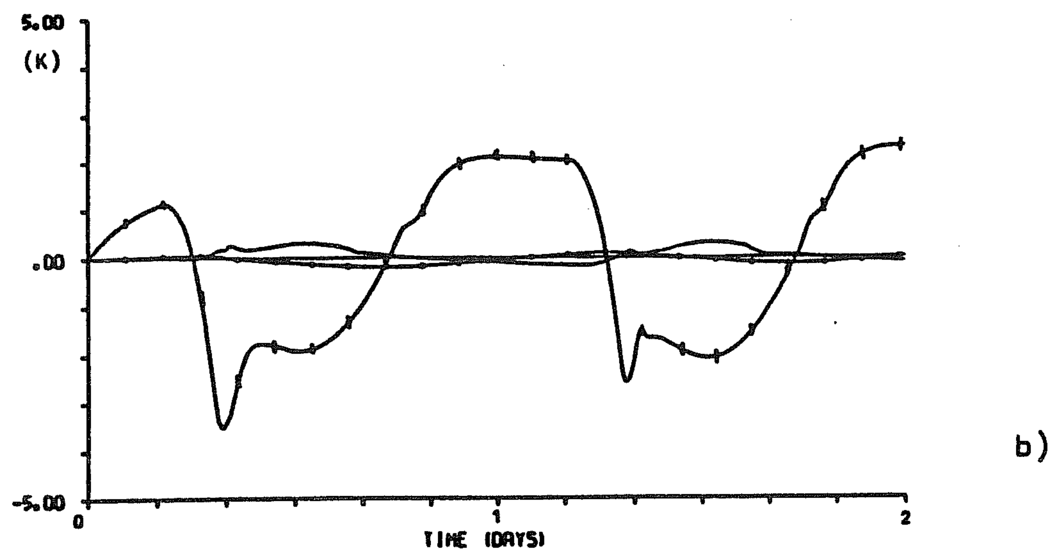




Figure 4: Evaluation of differences obtained from calculations with two model versions (H88L8-B83).
a) ——— net surface energy fluxes, +——+ soil heat flux at the interface between the two soil layers, ●——● soil heat flux at the lower boundary of the soil model.
b) +——+ Surface temperature, ——— temperature of the lowest atmospheric layer (950 hPa), ●——● temperature at the interface between the two soil layers. See text for details.

## 5 GENERAL CIRCULATION MODEL RESULTS

Experiments were carried out with the DWD general circulation model. This model uses a regular lat/lon-grid with a grid spacing of $5^\circ$. In the vertical there are ten layers, equally spaced in $\sigma = p/p_s$, where $p_s$ is the unreduced surface pressure. The difference approximations are based on the potential enstrophy conserving system of Burridge and Haseler (1977) in the C-grid. However, the vertical differencing of humidity follows Arakawa and Lamb (1977). The leap-frog time differencing scheme is used. Near the poles a Fourier filter is applied to the tendencies of the prognostic variables to allow longer time steps. A linear fourth order horizontal diffusion is used, which for temperature and specific humidity is applied on pressure levels. The most important parameterizations were mentioned in chapter 4 already. Additionally, the statistical cloud cover parameterization as described by Heise (1985) is used in the version of Heise et al. (1988).

All experiments start from observed atmospheric fields (December 16, 1978; 12.00 z). The initial fields of temperature and water content of the soil layers are interpolated from climatological fields used at the ECMWF (Research Manual 3, ECMWF Forecast Model, Physical Parameterizations, 2nd Edition, 1/88). January sea surface temperatures are prescribed according to Alexander and Mobley (1976). The model is integrated in time for three 16 day periods, the last two of them are used for evaluating the January statistics in the following part.

Three experiments were conducted, they are comparable to those in the preceeding paragraph:
SOIL: The ground fraction covered by plants is assumed to be zero.
B83 : The B83-model is used.
H88 : The surface model as described in this text is used.

In all three experiments the appropriate soil and plant data as described in 3.3 and 3.4 are used. Owing to the large difficulties in the preparation of global plant datasets, only annual mean values of the ground fraction covered by plants and the leaf area index are prescribed (Figure 2), although the model was run in the perpetual January mode.

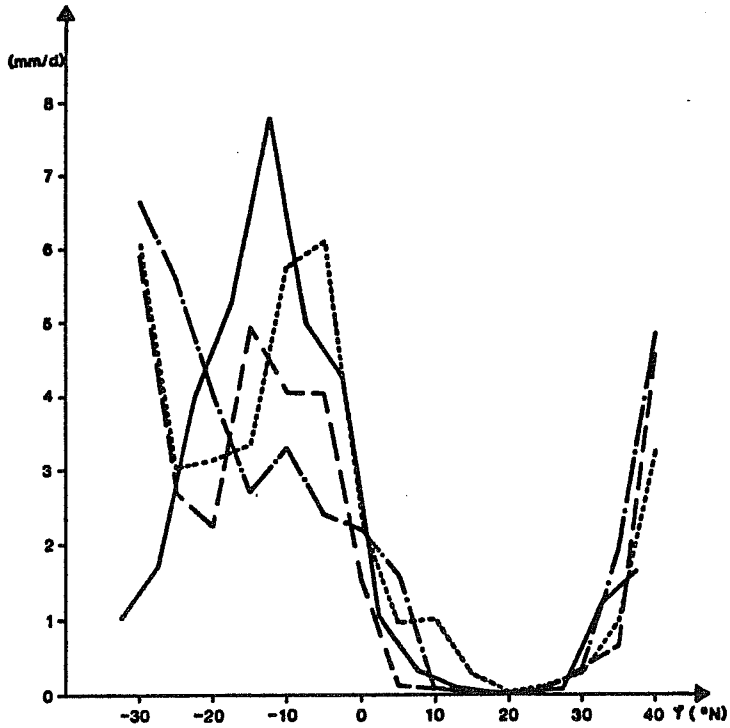



Figure 5: Zonally averaged January precipitation rate over the continent of Africa.
——— Observations (Jaeger, 1976),— — experiment SOIL, —·—· experiment B83, ----- experiment H88.

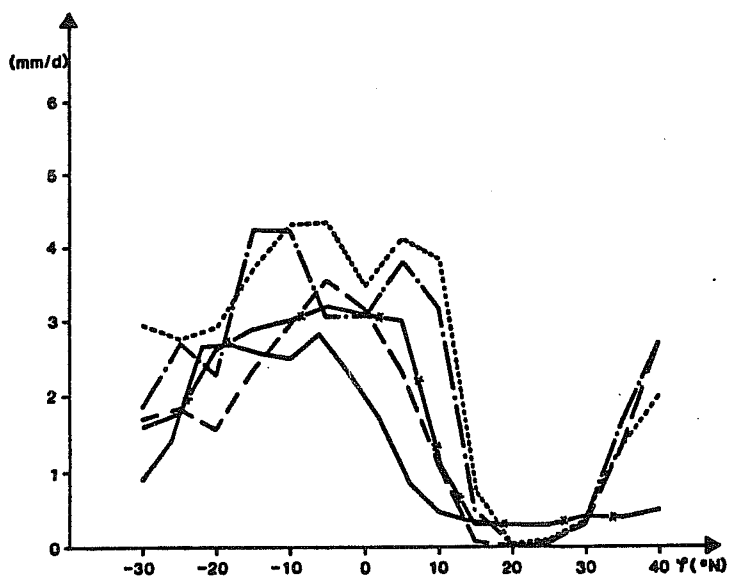



Figure 6: Zonally averaged January evaporation rate over the continent of Africa.
x—x Observations (Mintz and Serafini, 1984),——— observations (Schutz and Gates, 1971), experiments as in Figure 5.

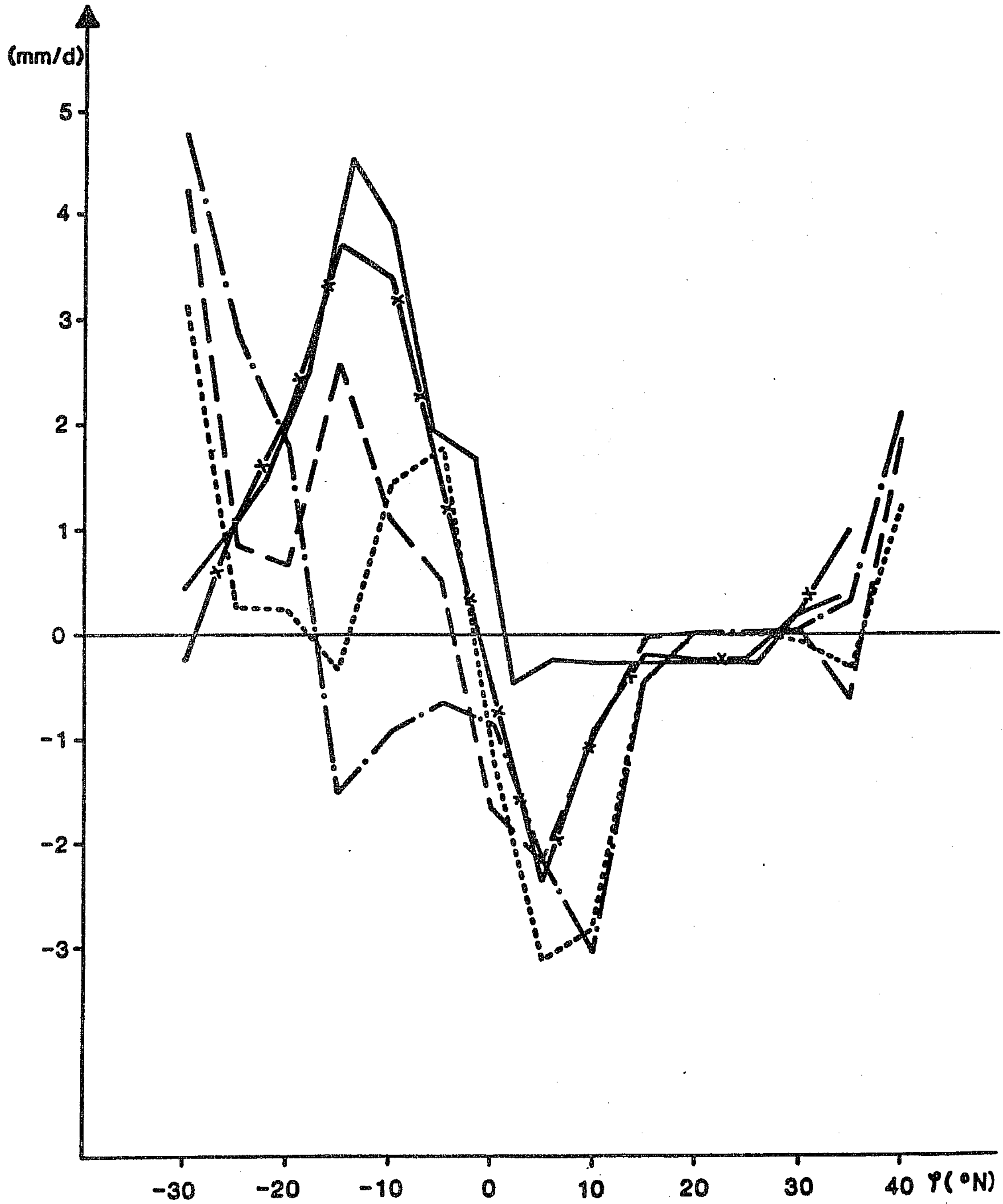



Figure 7: Zonally averaged hydrological budget (precipitation minus evaporation) over the continent of Africa for January.
x——x Observations (precipitation rate Jaeger, 1976, minus evaporation rate Minz and Serafini, 1984), ——— observations (precipitation rate Jaeger minus evaporation rate Schutz and Gates, 1971), experiments as in Figure 5.

The impact of changes in the transpiration formula is unlikely to be large in global statistics of a general circulation model. Therefore the following considerations are confined to zonal mean values of water balance components over the continent of Africa. As Figure 5 shows for the precipitation rate, the different climatic zones are rather well captured by the three model versions. South of the equator a northward shift of these zones is to be observed, most obvious in the appearence of high precipitation rates south of 20° S. In the region of the ITCZ, the H88 model simulates the highest values, close to the observations. Also, in this model the most pronounced decrease to the subtropics of the southern hemisphere occurs. The performance of the SOIL version is comparable to the H88 version, the maximum in the ITCZ-region is not as well developed, the simulation near the southern boundary of the Sahara seems to be better, however.

The comparison of evaporation rates is very problematic due to the fact that observations are parameterizations too. Additionally, Monteith (1985) pointed out that evaporation rates in dry climatic regions normally are severely underestimated. Bearing these difficulties in mind, the model results of evaporation (Figure 6) reproduce sufficiently well the observed distributions.

The difference precipitation minus evaporation is contaminated by the errors of both sets of observations. Nevertheless, some distinct dry and moist regions show up in Figure 7. These are simulated fairly well in SOIL and H88, but only indicated in B83.

## 6 SUMMARY AND CONCLUSIONS

The parameterization of the transpiration by plants in the soil model of Bauer et al. (1983) was reformulated, mainly based on proposals by Dickinson (1984). A sensitivity analysis of the revised parameterization revealed that the soil water content, which was the sole controlling factor in the old version, retains its dominating influence in the revised formula. On the other hand, large differences in the transpiration rate are obtained, if the minimum stomatal resistance or the critical water content are changed. Especially the stomatal resistance appeared to be very critical, as its value - which should depend on the genus of the plant - is rather questionable. Much work therefore remains to be done for the preparation of global datasets of all relevant parameters describing the vegetation cover.

One dimensional tests with the soil model coupled to an atmospheric model showed the largest differences between the original and the revised version during the first hours after sunrise. Temperatures are now rising slower because the latent heat flux is larger. Due to the colder surface temperatures the sensible heat flux is reduced, thus the Bowen ratio is much smaller. This is true also for daily mean values, since at least for a dense vegetation cover the transpiration and hence the precipitation are larger than with the original version.

In January simulations with the general circulation model of the DWD the new version produced a better representation of dry and moist regions over the continent of Africa.

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