THE SENSITIVITY OF WINTER EVAPORATION TO THE FORMULATION OF AERODYNAMIC RESISTANCE IN THE ECMWF MODEL

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Abstract. In atmospheric models, the roughness length for momentum, heat and moisture are often taken equal, and tuned to the momentum budget problem. In this paper, it is shown that the roughness lengths have considerable impact on the evaporation in winter. One-column simulations of the land-surface scheme are driven with a long time series of observations for Cabauw in The Netherlands. It is shown that with the operational roughness lengths for this location (as in use at ECMWF in May 1993), evaporation in January, February and March is overestimated by more than a factor 2. More realistic parameters, as documented for this site, virtually eliminate the error. This study shows the importance of the surface roughness lengths in determining evaporation from wet surfaces. It also illustrates the strength of long observational time series in identifying model deficiencies.

1. Introduction

In the parameterization of evaporation over land, evaporation is often represented as a moisture flow between a saturated sub-surface and the reference level (observation level or lowest model level) through a resistance network consisting of a surface resistance r_s and an aerodynamic resistance r_a. The surface resistance reflects the effect of stomatal resistance in the vegetation whereas turbulent diffusion in the surface layer is represented by the aerodynamic resistance. The aerodynamic resistance is often smaller than the surface resistance. The effect of the aerodynamic resistance becomes dominant when the vegetation is wet or when the radiative forcing at the surface is weak. Both conditions predominate in winter evaporation at mid-latitudes.

The aerodynamic resistance in the surface layer is to a large extent determined by wind speed and by the surface roughness lengths for momentum and moisture. The aerodynamic resistance is the key parameter determining the surface drag due to turbulence (e.g., Mason, 1988, 1991) and the roughness length for moisture is its counterpart for the moisture transfer problem (see e.g., Brutsaert, 1982). In operational models, it is customary to select an "effective" aerodynamic roughness length (Fiedler and Panofsky, 1972) to account for sub-grid orographic effects and terrain inhomogeneities. The idea is to adjust the aerodynamic roughness in such a way that the model produces the correct area-averaged surface drag. Methods are available now to determine effective roughness lengths, although this is extremely difficult in practice because of the lack of detailed topographic data (Taylor *et al.*, 1989; Mason, 1991). The situation is much less

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clear for the roughness lengths for heat and moisture. In the ECMWF model, the aerodynamic roughness length and the roughness length for heat and moisture are assumed to be identical and are tuned to the momentum budget. Garratt (1993) reviewed more than 30 GCM's and all have identical roughness lengths for momentum, heat and moisture.

It has been known for a long time that the roughness length for heat and moisture are smaller than the one for momentum (Garratt and Hicks, 1973; Garratt and Francey, 1978; Brutsaert, 1982), but it is only recently that ideas are being developed to account for the effects of sub-grid orography (Mason, 1991) and inhomogeneous terrain (Claussen, 1991, 1994). Mahrt and Ek (1993) found orders of magnitude difference between the roughness length for momentum and heat for the HAPEX experiment, and partially attribute that result to inhomogeneous terrain effects.

In this paper, we shall study the effect of the aerodynamic resistance on evaporation using the ECMWF single-column land-surface model. The model consists of the land-surface scheme and surface-layer part of the vertical diffusion scheme. The model variables at the lowest model level are replaced by an observed time series of wind, temperature and specific humidity. Downward shortwave radiation, downward longwave radiation, and precipitation as needed by the surface scheme are also taken from observations. The purpose of this exercise is to isolate problems in the land-surface parameterization from other model problems. In this paper we shall use data from Cabauw in The Netherlands (Driedonks *et al.*, 1978).

Validation of land-surface schemes in single-column mode will be carried out extensively in the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS, Henderson-Sellers *et al.*, 1993), and the Cabauw data will most likely be among the data sets used for this study (see Section 3 for more details on the data).

2. Model Description

The model used in this study is the ECMWF land-surface parameterization scheme (as operational in May 1993; see Blondin, 1991) combined with an experimental version of the surface-layer part of the boundary-layer scheme. The land-surface scheme has two prognostic layers for soil moisture and temperature and a fixed boundary condition in the so-called climate layer which is specified with help of the monthly climate fields of Mintz and Serafini (1989). The top layer, the deep layer and the climate layer have depths of 7, 42 and 42 cm, respectively. The soil diffusivities and conductivity are constant.

For the purpose of parameterization of evaporation, four different surface fractions are considered in a single grid box: (i) a snow fraction depending on the depth of the snow layer, (ii) a fraction with wet vegetation or wet bare soil, (iii) a dry bare soil part, and (iv) a dry vegetated part. The part with snow cover and the wet fraction have zero surface resistance to evaporation; over the dry bare soil part, relative humidity is assumed dependent on the moisture content of the top soil layer and the dry vegetated part has a bulk surface resistance that is determined by a minimal stomatal resistance, by the soil moisture in the root zone (average of top layer and deep layer), and by the short-wave radiation. Apart from the temperature and moisture in the soil, the model also has variables for water in the skin reservoir and for the skin temperature. For more details on the soil parameterization, see Blondin (1991); the skin temperature parameterization is documented by Beljaars and Betts (1992).

For the surface layer, the experimental scheme, as described by Beljaars and Holtslag (1991), is used. The transfer coefficients or resistances between the surface and the lowest model level are not expressed as a function of the bulk Richardson number (as in Louis, 1979) but rather as a function of the Obukhov length. The advantage is that the empirical functions can be specified as they have been measured and that the surface roughness lengths for momentum, heat and moisture can be chosen independently (we shall take the roughness length for heat equal to that for moisture). The disadvantage is that the Richardson number needs to be converted to the Obukhov length every time step by an iterative method.

Because we concentrate in this paper on the effect of the aerodynamic resistance, we reproduce the model formulation for the aerodynamic resistance Equation (1) and the Penman–Monteith equation for evaporation (Equation (2), see Monteith 1981):

$$r_a^{-1} = \frac{u_* k}{\ln\left(\frac{z}{z_{0h}}\right) - \Psi_H\left(\frac{z}{L}\right)}, \tag{1}$$

$$\lambda E = -\frac{s(Q_N - G_0) + \rho C_p \Delta q / r_a}{s + \gamma(1 + r_s/r_a)}, \tag{2}$$

where

$$H = -(Q_N - G_0) - \lambda E, \quad L = \frac{u_*^3}{k \frac{g}{T_v} u_* \theta_{*v}},$$

$$u_* = \frac{U_h k}{\ln\left(\frac{z}{z_{0m}}\right) - \Psi_M\left(\frac{z}{L}\right)}, \quad \theta_* = \frac{H}{\rho C_p u_*},$$

$$q_* = \frac{E}{\rho u_*}, \quad \theta_{*v} = \theta_* + 0.61 T q_*,$$

$$U_h = \{U^2 + w_*^2\}^{1/2}, \quad w_* = \left\{-z_i \frac{g}{T} u_* \theta_{*v}\right\}^{1/3}, \quad z_i = 1000\,\mathrm{m},$$

$$\Psi_M = 2\ln\{(1+x)/2\} + \ln\{(1+x^2)/2\} - 2ATAN(x) + \pi/2 \quad \text{for } z/L < 0,$$
$$\Psi_H = 2\ln\{(1+x^2)/2\} \quad \text{for } z/L < 0,$$

where $x = (1 - 16z/L)^{1/4}$,

$$-\Psi_M = a\frac{z}{L} + b\left(\frac{z}{L} - \frac{c}{d}\right)\exp\left(-d\frac{z}{L}\right) + \frac{bc}{d} \quad \text{for } z/L > 0,$$

$$-\Psi_H = \left(1 + \frac{2}{3}a\frac{z}{L}\right)^{3/2} + b\left(\frac{z}{L} - \frac{c}{d}\right)\exp\left(-d\frac{z}{L}\right) + \frac{bc}{d} - 1 \quad \text{for } z/L > 0,$$

with $a = 1$, $b = 0.667$, $c = 5$ and $d = 0.35$;

where Q_N is the net radiation; L is the Obukhov length; H is the sensible heat flux; ρ is the air density; λE is the latent heat flux; C_p is the specific heat at constant p; G_0 is the ground heat flux; λ is the latent heat of vaporization; u_* is the friction velocity; k is the Von Karman constant (0.4); T is the temperature at z; Ψ_M is the momentum stability function; q is the specific humidity at z; Ψ_H is the heat/moisture stability function; U is the horizontal wind speed at z (component average); U_h is the average of absolute horizontal wind speed; z is the reference height (observation height or height of lowest model level above displacement height); z_{0m} is the aerodynamic roughness length; z_{0h} is the roughness length for heat/moisture; r_a is the aerodynamic resistance; $s = dq_{\text{sat}}/dT$; r_s is the surface resistance; q_{sat} is the saturation specific humidity; $\gamma = C_p/\lambda$; $\Delta q = q - q_{\text{sat}}$, the moisture deficit at z; w_* is the free convection velocity scale (= 0 for stable situations).

These expressions are consistent with Monin Obukhov similarity in the formulation proposed by Beljaars and Holtslag (1991). The stability functions for the unstable boundary layer are the well known Dyer and Hicks forms (see, Dyer 1974 or Högström, 1988). For stably stratified turbulence, the functions proposed by Beljaars and Holtslag (1991) are used. They have linear dependence of z/L with factors of proportionality of 5 at small z/L, gradually decreasing to 1 for large z/L (see Holtslag and De Bruin, 1988 and Hicks, 1976).

The w_* effect in the horizontal wind has been included to ensure a proper free convection formulation as proposed by Miller *et al.* (1991). It prevents the surface wind from dropping to zero when heating is forced from the surface (i.e., it permits a convection-induced surface wind).

The Penman-Monteith Equation (2) is not used in the model as such, but is used implicitly since the surface scheme adjusts the surface temperature to satisfy the surface energy balance. Equation (2) is included here because it gives us quantitative insight into the parameterization of evaporation in winter. The evaporation in the Penman-Monteith equation consists of 2 terms: the net radiation term and the moisture deficit term. The net radiation term is more important in summer than it is in winter, because the radiation is larger and the slope of the saturation specific humidity curve increases with temperature. Thus the regimes of temperature and radiation in winter oblige us to put more emphasis on the moisture deficit term. In winter the vegetation is often wet and therefore the aerodynamic resistance can become the key parameter determining evaporation.

3. The Cabauw Data Set

The data used in this study have been collected on the 200 m meteorological mast in Cabauw in the Netherlands (51°58′N and 4°56′E). This site is located in flat terrain consisting mainly of grassland interrupted by narrow ditches. Up to a distance of 200 m from the mast, there are no obstacles or perturbations of any importance; farther on, some scattered trees and houses are found for most wind directions (see Driedonks *et al.*, 1978, for a more detailed description).

In this paper, we make use of the observations of wind, temperature and specific humidity at a height of 20 m as a boundary condition to the one-column soil model (the lowest model level in the operational ECMWF model is about 30 m above the surface). Additionally, surface observations of precipitation, solar downward and longwave downward radiation are used. All quantities are averages for 30 min intervals. For verification, fluxes of sensible and latent heat have been derived from net radiation, ground heat flux and profiles of wind temperature and moisture (see Beljaars, 1982). Different methods can be used: (i) Sensible heat flux from profiles of temperature and wind and latent heat as a residual from the surface energy balance, (ii) Sensible and latent heat flux from the Bowen method, and (iii) Sensible and latent heat fluxes from temperature, moisture and wind profiles. The first method is used most of the time, but the second or third method is selected occasionally when data are missing. When data are still missing because of instrument failure or data transmission problems, the gaps are filled in with observations from the SYNOP station De Bilt 30 km away. In the latter case, the FLUXLIB software package is used to simulate the surface fluxes, to interpolate to the 20 m level and to correct for differences in terrain roughness (see Beljaars and Holtslag, 1990). Although the methods used in the package have been extensively verified against data (Holtslag and Van Ulden, 1983), the simulated data can not be considered to be real verification material. However, the amount of synthetic data is small (about 10–20%, the percentage dependent on the specific parameter) and the procedure enables one to look at integrated budgets.

The data set has been prepared for the entire year of 1987, although we concentrate on January, February and March in this study. Figure 1 illustrates a few components of the surface energy budget as a function of day number. The basic data set consists of half-hour averages, but diurnal averages have been computed and plotted in the figures. The figure contains net short-wave radiation (albedo is estimated as 0.23), net long-wave radiation, the sum of sensible and latent heat fluxes, and the ground heat flux. The sum of the atmospheric components minus

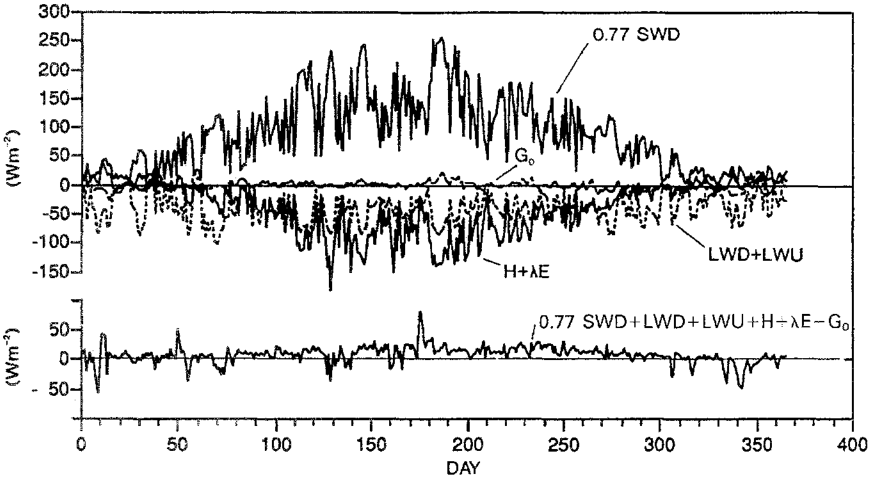

Fig. 1. Observed diurnal averages of surface energy components as a function of day number for Cabauw in 1987. Net shortwave (0.77SWD; albedo 0.23 is assumed), net longwave (LWD+LWU), sensible plus latent heat flux ($H + \lambda E$) and the ground heat flux (G_0) are distinguished. Downward fluxes are positive. Fig 1b shows the residual in this budget.

the ground heat fluxes is a residual, which should be zero for the observations. It gives information about the consistency of the data. The residual is quite large in summer, when the mismatch can reach about 20% of net radiation in the diurnal averages. This inconsistency is already present when the sum of the long- and short-wave observations is compared with the total net radiation. It appears most likely that the net radiation instrument underestimates the net radiative forcing in summer. Because this observation is used as a basis to determine the sensible and latent heat fluxes, it is possible that also the sensible and latent heat fluxes are underestimated in summer. In winter, the different observed terms in the surface energy budget are in better balance.

4. Model Simulations

The aim of the one-column simulations is to identify model problems related to the aerodynamic resistance, so the first simulations were done with the roughness-length parameters as they are used in the global model for the Cabauw grid box. The roughness lengths z_{0m} and z_{oh} are both taken equal to 0.4 whereas the albedo is set to 0.23, which is thought to be appropriate for the experimental site but higher than the value of 0.18 as in the operational model for this location. The results for sensible and latent heat flux are shown in Figure 2 in comparison with observations. The integrated evaporation is shown in Figure 3. The evaporation is substantialy overestimated; after 100 days from 1 January, the evaporation is about 120 mm in the simulation whereas only 50 mm has been observed. The

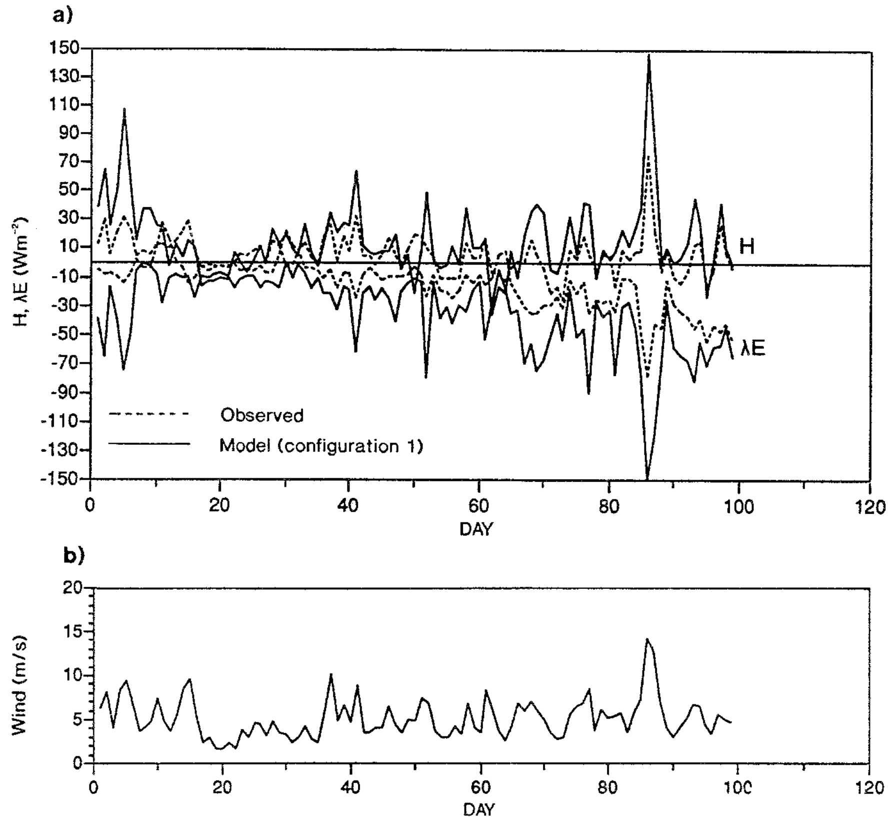

Fig. 2. Observed (dashed) and simulated (solid) diurnal averages of sensible and latent heat (Figure 2a). The wind speed is shown in Fig 2b.

energy for this evaporation comes from a downward sensible heat flux. This is realistic, but most of the time the sensible and latent heat fluxes (with opposite sign) are larger in magnitude than observed. Comparison with the wind speed in Figure 2b shows that the errors become particularly large when the wind speed is high, which suggests that the problem is related to the formulation of the aerodynamic resistance. Figure 4 shows the daily sums of precipitation together with the diurnal average of the skin reservoir content of the model (maximum value is 0.8 mm on vegetation). It is clear that the vegetation is wet during a considerable fraction of the time, implying that the surface resistance to evaporation is small.

We shall now modify the surface roughness lengths to increase the aerodynamic resistance. Table I summarizes the list of parameters that has been used; configurations 1 to 5 correspond to decreasing aerodynamic resistance. Configuration 1 represents the operational ECMWF model for the Cabauw site; configuration 2 has new climate fields for z_{0m} and z_{0h} where Mason's (1991)

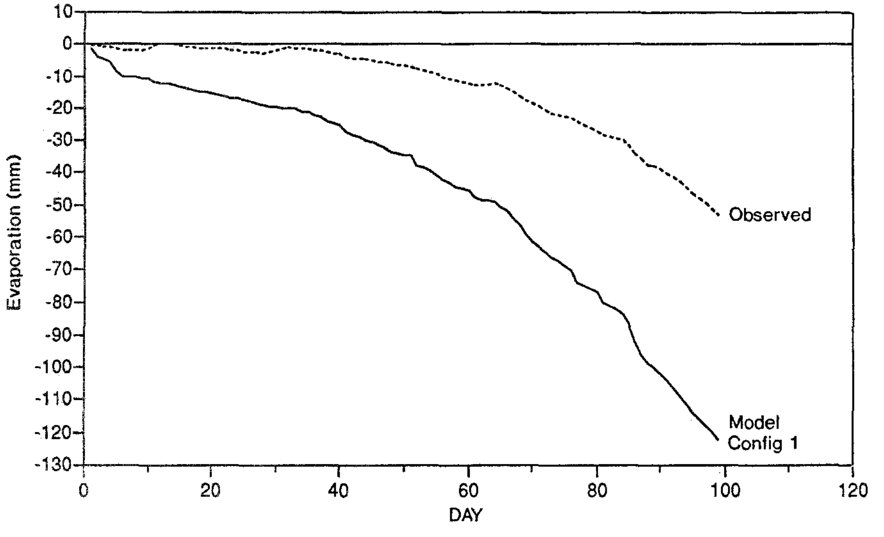

Fig. 3. Time-integrated evaporation (in mm water; upward flux is negative), observed (dashed) and simulated (configuration 1, solid) for the first 100 days of 1987 of the Cabauw data set.

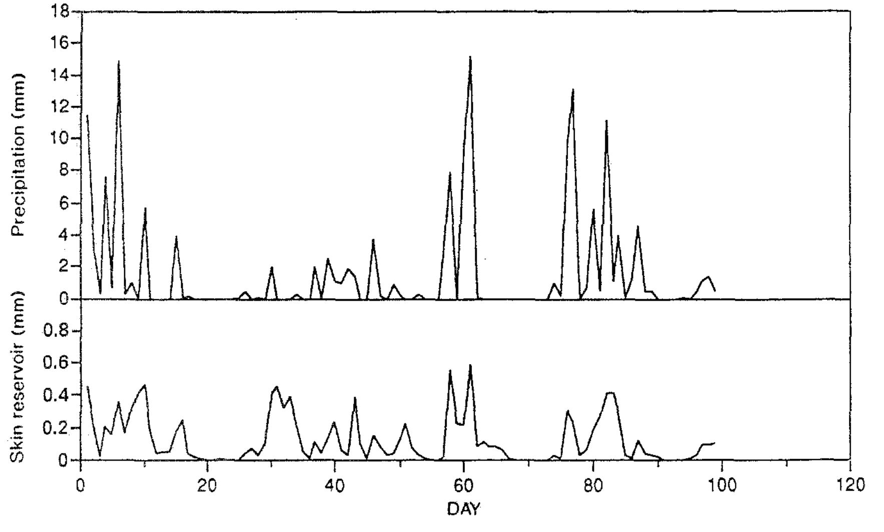

Fig. 4. Day sums of precipitation (observed, a) and diurnal average of skin reservoir content as simulated by the model in configuration 1 (b).

suggestions have been used to compute the orographic contribution in z_{0m}. The value of z_{0h} is taken as 10% of z_{0m} for the vegetation contribution, but z_{0h} is reduced even further in mountainous terrain with the help of the blending height concept (Mason 1991; Claussen, 1991).

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

Configuration 3 has an aerodynamic roughness length which is typical for the observational site (Beljaars and Holtslag, 1991) and $z_{0h} = z_{0m}$. Configuration 4 and 5 have further decreasing values for z_{0h}. It is argued by Beljaars and Holtslag (1991) that configuration 5 is the most realistic one on the basis of an analysis of summer observations of radiative surface temperature, wind/temperature profiles and the surface energy balance at the Cabauw site. The large ratio of z_{0m}/z_{0h} is believed to be due to the effects of inhomogeneous terrain.

The model-simulated accumulated evaporation is displayed in Figure 5 for the different model configurations. Figure 6 shows diurnal averages of sensible and latent heat fluxes for the different model configurations in comparison with observations. It is clear that the increase of the aerodynamic resistance by reducing the surface roughness lengths for momentum and heat has a considerable impact on the simulated evaporation. The model evaporation over the first 100 days of 1987 could be reduced from about 120 to 60 mm when going from the operational parameter setting to the parameter values that are considered to be appropriate for the Cabauw site. Also the annual evaporation is much closer to observations, but it is more difficult to draw firm conclusions for the summer data because the aerodynamic resistance is less dominant in the parameterization of evaporation. In summer, there may be some influence from the ground hydrology on the surface resistance, which is difficult to separate from the aerodynamic resistance.

5. Discussion

The Cabauw near-surface observations have been used as a boundary condition for the single-column version of the ECMWF land-surface and surface-layer turbulence model. The magnitude of the aerodynamic resistance turns out to be quite important for winter evaporation. With small aerodynamic resistance, the evaporation is too large and compensated for by downward sensible heat flux. The correlation of these errors with wind speed is also a clear indication that the aerodynamic resistance is involved. This is consistent with documented

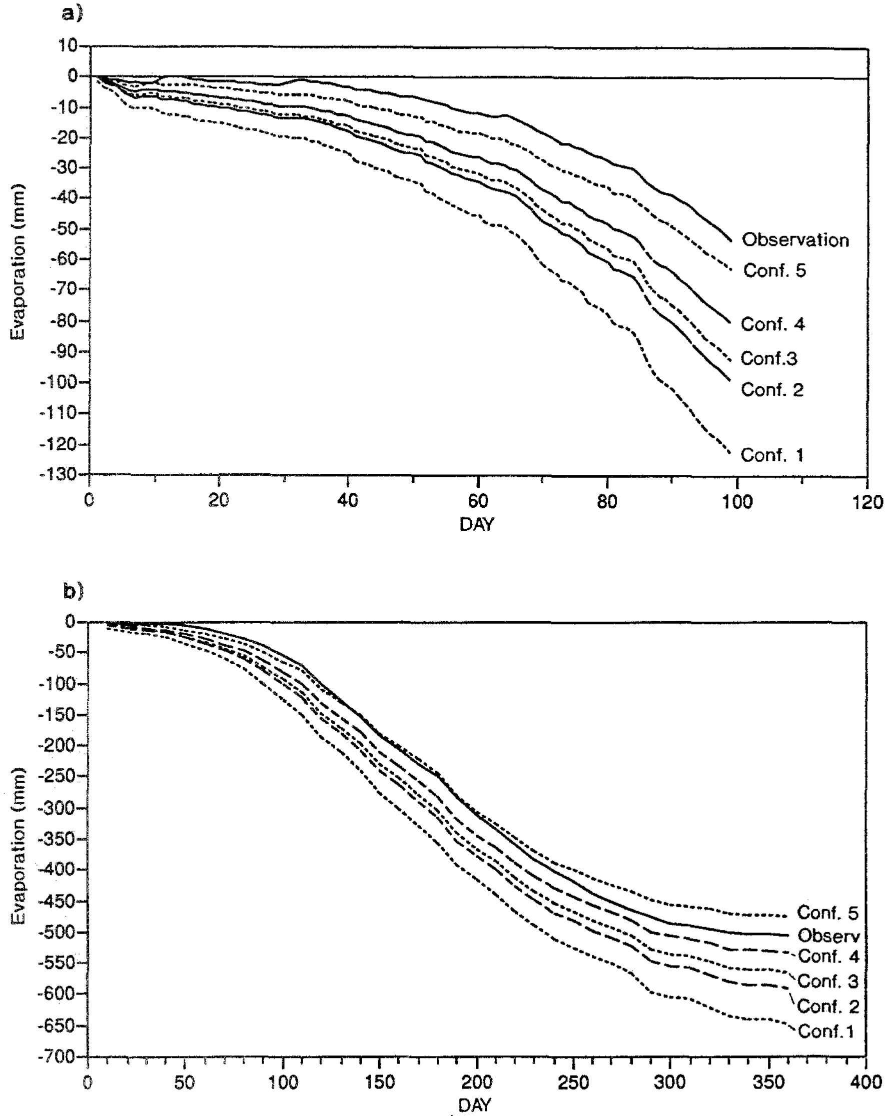

Fig. 5. Accumulated evaporation as observed and simulated with different model configurations (see Table 1) for the first 100 days of 1987 (a) and for the entire year (b).

biases in the boundary-layer budgets of the ECMWF operational model over Europe. In winter the boundary layer tends towards a too moist and too cold state in the forecasts (Beljaars and Betts, 1992). The simulations could easily be improved for the Cabauw site in a single-column mode by specifying more appropriate roughness lengths for momentum and heat (which are reasonably well documented for this site). Improving the global model is quite a different matter, because it involves the specification of global fields for the roughness lengths

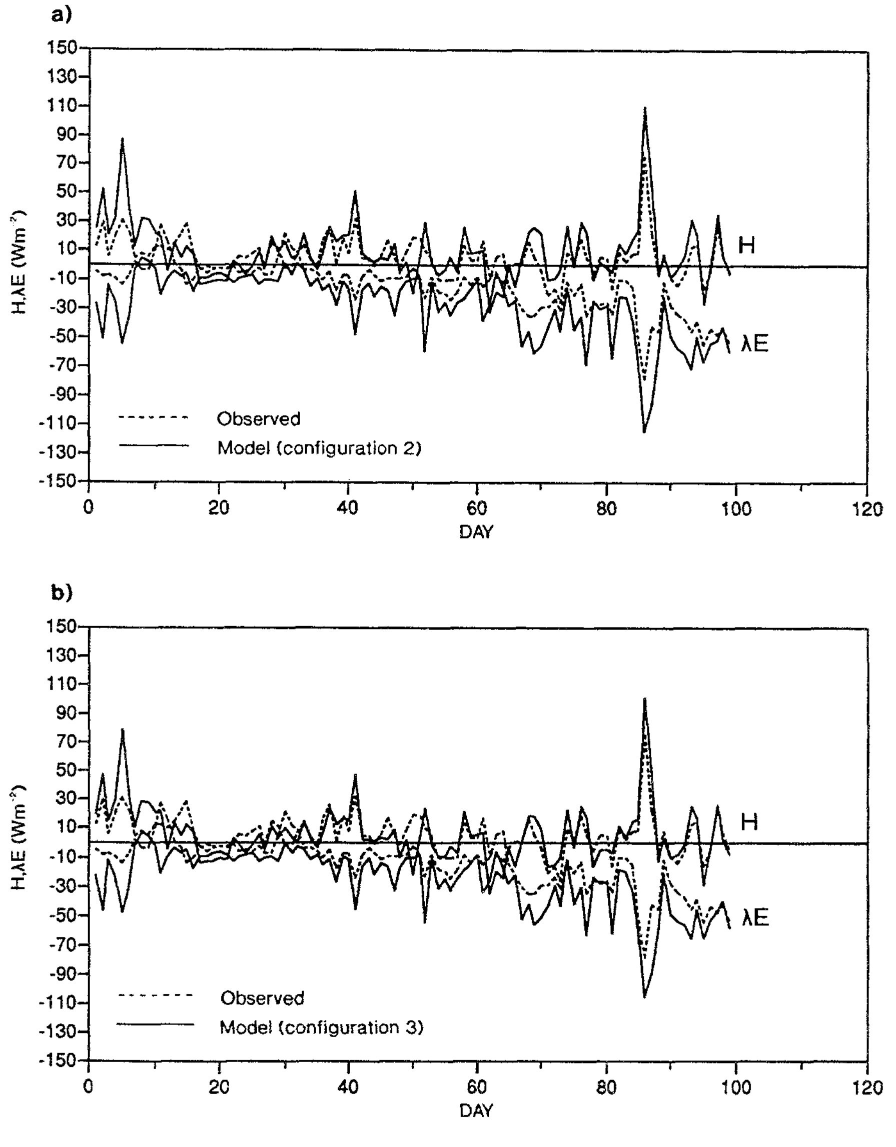

Fig. 6a,b.

for momentum and heat. Until now, little attention has been paid to this problem in large-scale models. The main focus has been on the momentum budget and its implications for the roughness length for momentum. Many models assume $z_{0h} = z_{0m}$ (Garratt, 1993). This aspect is easy to change in large-scale models, and a consensus seems to exist that the ratio between the two parameters is at least 10 for homogeneous vegetation (Garratt and Hicks, 1973). However, the variability in observations is large. A ratio of about 40 has been reported by Kohsiek *et*

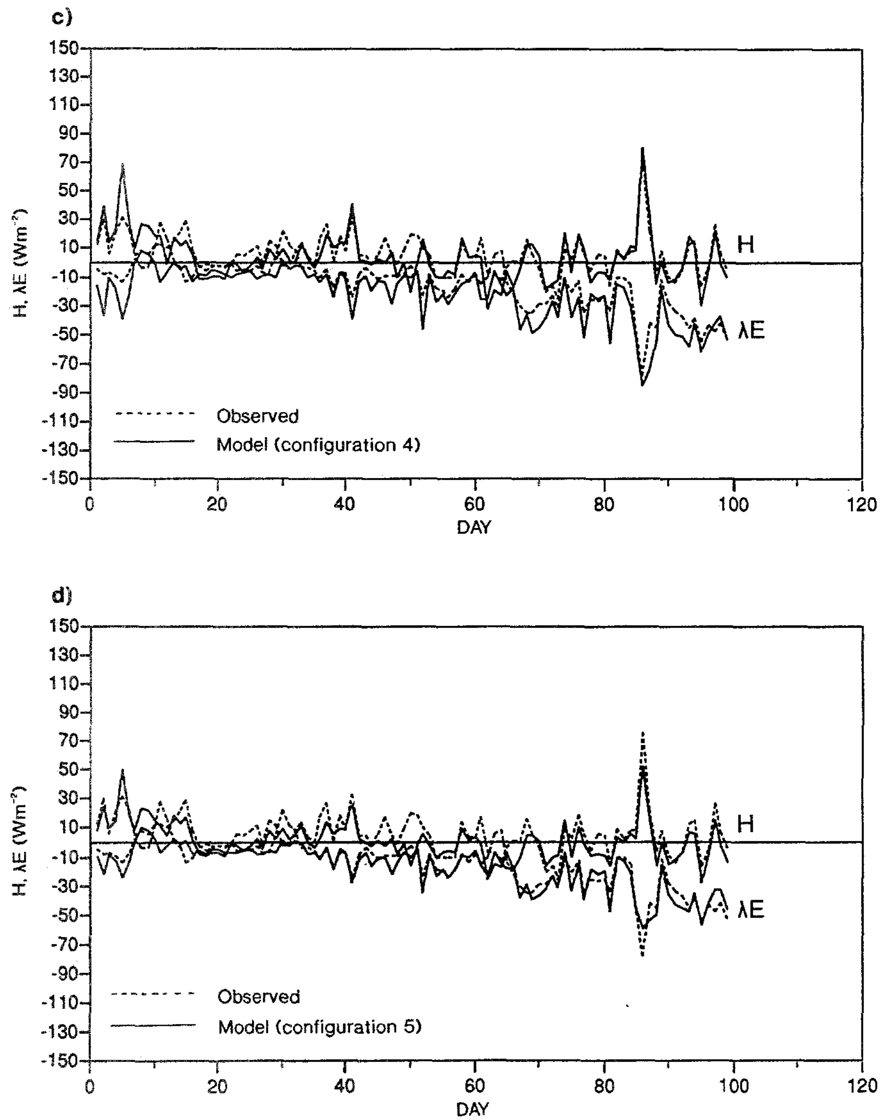

Fig. 6. Diurnally averaged simulated sensible and latent heat fluxes in comparison with observations for the different model configurations.

al.. (1993) for bare soil. Betts and Beljaars (1993) find $z_{0m}/z_{0h} \approx 20$ for the FIFE area with grassland in rolling terrain in the centre of the USA. Sugita and Brutsaert (1990, 1992) and Brutsaert and Sugita (1992) find much larger ratios for the same FIFE area mainly because their aerodynamic roughness length is larger. For the forest area of Les Landes in the South of France, a ratio of order 10 is found by Brutsaert *et al.* (1993). Given the spread in the data and the possible dependence on surface characteristics (e.g., leaf area index and vegetation cover) the parameterization of air-surface interaction with the help of a single parameter z_{0h} (different from z_{0m}) may be an oversimplification (Garratt *et al.*, 1993) but should be considered as an improvement over a parameterization with $z_{0h} = z_{0m}$.

The effects of orography and inhomogeneous terrain are even more difficult to handle. It is clear that enhanced momentum exchange between the atmosphere and the surface due to orographic or inhomogeneous terrain effects does not apply in the same way to the heat and moisture exchange problem. Beljaars and Holtslag (1991) give the example of the weakly inhomogeneous Cabauw site, for which it is suggested to compensate the roughness length of heat for the enhanced aerodynamic roughness length. How this should be done quantitatively and how boundary-layer stability affects this aspect is far from clear yet. Preliminary formulations have been suggested by Mason (1991) and Claussen (1991) and turn out to be beneficial. HAPEX data seem to support the idea of large ratios of z_{0m}/z_{0h} for inhomogeneous terrain (Mahrt and Ek, 1993). However, the lack of detailed topographic information makes it difficult to produce accurate global fields of the surface roughness lengths.

Other recent studies also emphasize the water recycling through the wet skin of the surface (Dolman and Gregory, 1993; Warrilow and Buckley, 1989). In the latter study, sensitivity is found to the surface roughness length. The present study supports this idea by direct comparison with observations. Jacobs and De Bruin (1992) argue that the aerodynamic resistance is of secondary importance, also because of negative feedback from the boundary-layer coupling. They concentrate, however, on situations where the surface resistance dominates, i.e., in summer with dry vegetation.

The present study clearly shows the value of long time series of observational data combined with single-column simulations. It enables the isolation of deficiencies in the surface scheme from other model problems. Deficiencies, however, may look worse in such a test than they actually are, because of feedback from the boundary layer. With a coupled atmosphere land-surface system, the atmospheric feedback is probably negative, e.g., too much evaporation results in too moist boundary layers and therefore reduces the evaporation (e.g., Jacobs and De Bruin, 1992). Experience with atmospheric models with a full physics package indicates that model feedbacks are often difficult to understand. It is therefore felt that stand-alone validation as proposed in the PILPS exercise is very important, particularly if data sets become available for different climatological regimes.

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