A Study of the Sensitivity of Land Surface Parameterizations to the Inclusion of Different Fractional Covers and Soil Textures

D. T. MIHAILOVIC

University of Novi Sad, Department of Meteorology, Novi Sad, Yugoslavia

H. A. R. DE BRUIN

Agricultural University, Department of Meteorology, Wageningen, The Netherlands

M. JEFTIC

University of Novi Sad, Department of Meteorology, Novi Sad, Yugoslavia

A. VAN DIJKEN

Agricultural University, Department of Meteorology, Wageningen, The Netherlands

(Manuscript received 15 November 1991, in final form 29 May 1992)

ABSTRACT

The inclusion of processes relating to soil type and vegetation is very important in an attempt to improve a land surface parameterization for use in different scale atmospheric models. There is already sample micrometeorological information about the sensitivity of surface energy fluxes and hydrology to plant communities, fractional plant covers, and some soil parameters, but not to soil texture. This study is concerned with the question of how much the land surface parameterization is sensitive to the inclusion of different fractional plant covers and soil-texture classes.

The plant community used at the atmosphere–land interface in the sensitivity tests was a maize canopy planted at the experimental site De Sinderhoeve, the Netherlands, in the growing season of 1988. Four days at different stages of maize growth have been selected. For the time integrations in the sensitivity tests, a one-dimensional biophysical scheme was used. Nine soil textures have been chosen for simulations in the sensitivity tests.

First, the model accuracy was established with observations using the soil texture, which approximately corresponded to the texture of the soil type where the experimental site was located. Then the authors investigated the sensitivity of the other soil textures for the fixed vegetation data and atmospheric conditions. Using the surface fluxes as the output, the sensitivity of the land surface parameterization scheme to the prescribed soil textures and fractional vegetation covers was established.

1. Introduction

In recent years, meteorologists have invested large efforts in developing models of land surface processes for use in climate simulations, numerical weather prediction, and air quality assessments. These efforts are directly determined by the growing demand for a better understanding of climate processes over the land surface, which have to be studied in considerable detail because of their importance for food production, use of water resources, ecological processes, and other human activities. The most comprehensive models of land surface processes, usually called *biophysical schemes*, include those developed by McCumber (1980), Dickinson et al. (1986), and Sellers et al. (1986). They have already been incorporated in different atmospheric models. In the development of a model, the numerical modelers have to concentrate on a crucial question (Wilson et al. 1987): How do the chosen model variables and parameterizations express 1) the impact of climate (weather) changes on land surface processes and 2) the impact of land surface processes on climate (weather) changes?

A considerable number of authors have already devoted their attention to the sensitivity of surface energy fluxes and hydrology to the values of various important parameters: Saxton (1975), Deardorff (1978), Luxmoore et al. (1981), Thom (1975), Beven (1979), Sellers and Lockwood (1981), and Mihailovic (1990). Consequently, we already have extensive micrometeorological information about sensitivity studies to some parameters as reviewed by Wilson et al. (1987). All of these sensitivity studies dealt with different plant com-

Corresponding author address: Dr. Dragutin T. Mihailovic, Associate Professor of Meteorology, University of Novi Sad, Department of Meteorology, Novi Sad Trg Dositeja Obradovica 8, 21000 Novi Sad, Yugoslavia.

TABLE 1. Soil parameters used in the sensitivity tests. All values were taken from Wilson et al. (1987). Parameter 6 was calculated to fit (1) in Clapp and Hornberger (1978).

Number			Soil texture								
			1	2	3	4	5	6	7	8	9
1	p	porosity (volume of voids to volume of soil)	0.39	0.42	0.45	0.48	0.51	0.54	0.57	0.60	0.63
2	Φ_0	soil suction (m)	0.03	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
3	K_{w0}	saturated hydraulic conductivity (m s^{-1})	3.2×10^{-5}	1.3×10^{-5}	0.89×10^{-5}	0.63×10^{-5}	0.45×10^{-5}	0.32×10^{-5}	0.22×10^{-5}	0.16×10^{-5}	0.11×10^{-5}
4	γ_i	ratio of saturated thermal conductivity to that of saturated loam	1.3	1.2	1.1	1.0	0.95	0.90	0.85	0.80	0.75
5	B	exponent b defined in Clapp and Hornberger (1978)	4.5	5.0	5.5	6.0	6.8	7.6	8.4	9.2	10.0
6		moisture content relative to saturation at which transpiration ceases	0.161	0.266	0.300	0.332	0.378	0.419	0.455	0.487	0.516

munities, fractional plant covers, and soil parameters. Most of them, however, did not consider the sensitivity of the land surface parameterization to the choice of soil texture. The only exception was Wilson et al. (1987), who studied the sensitivity of the Biosphere–Atmosphere Transfer Scheme (BATS) to the inclusion of variable soil characteristics. Their primary aim was to examine the effect of the incorporation of some soil parameters and color. For that purpose they used the land-cover type and soil properties of five sample surfaces that represented a wide range of conditions simulated by the land surface parameterization scheme.

This study will be concerned with the question of how much the land surface parameterization is sensitive to the inclusion of the fractional plant covers and soil-texture classes. To do that, we examined the effect of a wide range of soil textures, from sand to clay, on atmospheric fluxes from a single type of plant cover for the four different fractional covers.

For the time integrations performed in the study, we have used the BATS model developed by Dickinson et al. (1981), which now includes the vegetation parameterization of Dickinson (1984) and Dickinson et al. (1986), and has been adapted for one dimension by Mihailovic (1990). The soils data are taken from the archives of Wilson and Henderson-Sellers (1985). We used 9 soil textures—from 3 to 11—out of the 12 ranging from sand (texture 1) to clay (texture 12), as shown in Table 1 (Wilson et al. 1987). The soil textures were chosen arbitrarily to illustrate a wide range of soil conditions used in land surface parameterization.

For the plant community at the atmosphere–land interface in the sensitivity tests, we have employed a maize canopy (*Zea mays L,* cv Brutus) planted at the experimental site De Sinderhoeve, the Netherlands, in the growing season of 1988. The sensitivity tests have consisted of running the BATS model for the four selected days at different stages with differing leaf-area indices (LAI) (Jacobs et al. 1990): 7 June (LAI = 0.44), 18 August (LAI = 4.3), 8 September (LAI = 4.0), 4 October (LAI = 2.0), and the nine soil textures. First, we established the model accuracy by observing soil texture 1, which approximately corresponded to the texture of the soil type where the experimental site was located. Once the set of vegetation data and atmospheric conditions were fixed, we changed only soil textures. Consequently, this design of sensitivity tests, using the surface fluxes as the output, allowed us to establish the sensitivity of the land surface parameterization scheme to the prescribed soil texture and fractional vegetation covers.

2. Model description

As already indicated, the sensitivity test performances used the BATS model, which was developed by Dickinson et al. (1981) but which now includes the vegetation parameterization of Dickinson (1984). This biophysical scheme for some physical processes is still highly simplified with respect to reality, but it incorporates most of the surface parameters, including a vegetation canopy, surface and rooting-zone soil layers, variable albedo, and hydraulic parameters. In this paper, we used the BATS scheme as adapted for one-point physics by Mihailovic (1990).

a. Ground parameterization

In the sensitivity tests presented in the next section, we parameterized the soil temperature following Deardorff (1978). The soil surface temperature T_g and subsurface temperature T_2 are obtained from

$$\frac{\partial T_g}{\partial t} = -\frac{c_1}{\rho_s c_s d_1} H_s - c_2 \frac{(T_g - T_2)}{\tau_1} \qquad (1a)$$

$$\frac{\partial T_2}{\partial t} = -\frac{H_s}{\rho_s c_s d_2}, \qquad (1b)$$

where the symbols have the following meaning: ρ_s is the density of the subsurface soil layer, c_s is the volumetric specific heat of the subsurface layer, H_s is the sum of fluxes to the atmosphere (positive when directed upward), $c_2 = 2\pi$, $c_1 = 2\pi^{1/2}$, $d_1 = (k_s\tau_1)^{1/2}$, τ_1 is the daily period of heating (8.64×10^4 s), k_s is the thermal diffusivity, and $d_2 = (365k_s\tau_1)^{1/2}$.

As it is well known, the thermal diffusivity and volumetric specific heat depend strongly on their moisture and organic content. According to de Vries (1975), we can get the following approximate formulas for the volumetric specific heat c_v and thermal diffusivity k_t for a loam soil:

$$c_v = (0.23 + x_w)c_w \tag{2}$$

$$k_t = \gamma_i \frac{(2.9 + 0.04x_w) \times 10^{-7}}{[(1 - 0.65x_w)x_w + 0.09](0.23 + x_w)}, \tag{3}$$

where

$x_w = p(x_u + x_t)/2$ is the weighted average volumetric soil water, x_u and x_t are the degrees of saturation of the upper and total soil layers, respectively, and p is the soil porosity, which is listed in Table 1 for different soil textures;

$c_w = 4.186 \times 10^6$ J m^{-3} K^{-1} is the volumetric specific heat of water; and

γ_i = ratio of the thermal conductivity of texture i to that of loam.

Shown in Fig. 1 are the calculated values of thermal diffusivity and thermal conductivity $\lambda_t = k_t c_v$, as functions of the weighted average soil water per unit soil volume x_w for the nine different soil textures.

The soil temperatures T_g and T_2 in (1a) and (1b) are calculated by an explicit backward-differencing scheme over a time step Δt:

$$T_g^{n+1} = T_g^n + \Delta t\left(-\frac{c_1}{\rho_s c_s d_1} - c_2\frac{T_g - T_2}{\tau_1}\right)^n \tag{4a}$$

$$T_2^{n+1} = T_2^n + \Delta t\left(-\frac{H_s}{\rho_s c_s d_2}\right)^n. \tag{4b}$$

b. Soil moisture and surface flux parameterization

Incoming moisture on the ground either infiltrates into the soil or is lost to surface runoff. These processes were parameterized in accordance with Dickinson et al. (1986) considering total water W_2 in the rooting-zone depth h_0, and surface soil water W_g in the upper layer of soil depth h_u is restricted to a thickness between 1 and 20 cm. The rooting zone h_0 and the upper soil depth h_u are both functions of land-cover type. In the presence of vegetation cover, the equations for W_g and W_2 can be written in the form

$$\frac{\partial W_g}{\partial t} = P_r(1 - \sigma_f) - R_s + F_g - \beta E_t - E \tag{5a}$$

$$\frac{\partial W_2}{\partial t} = P_r(1 - \sigma_f) - R_s - R_g - E_t - E, \tag{5b}$$

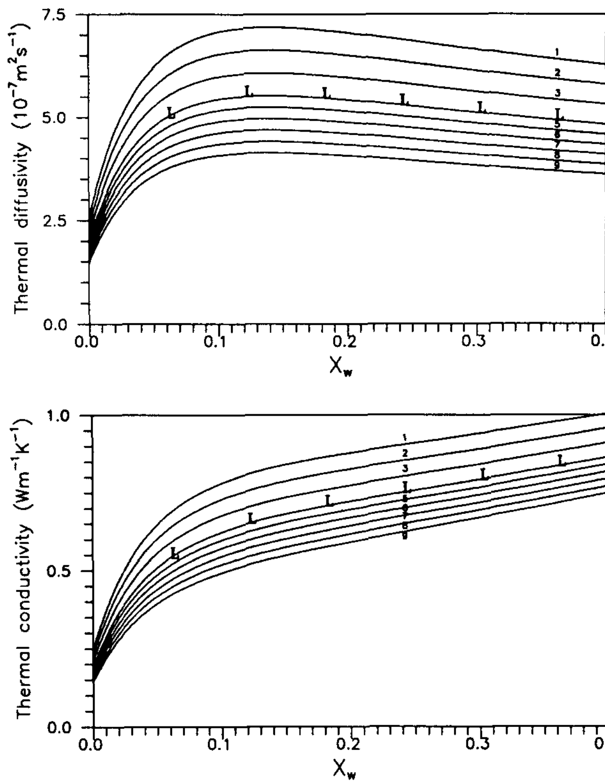

FIG. 1. Dependence of thermal diffusivity k_t (above) and thermal conductivity λ_t (below) on the weighted average soil water per unit soil volume x_w for nine different soil texture classes (see Table 1). The letter L denotes loam soil.

where P_r is the rainfall (m s^{-1}) and σ_f is the fractional vegetation cover. The other terms included in (5a) and (5b) are described hereafter.

The surface runoff R_s (m s^{-1}) is parameterized as

$$R_s = \left[\frac{1}{2}(x_u + x_t)\right]^4 G, \tag{6}$$

where G is the net input of water to soil surface.

The subsurface runoff R_g (m s^{-1}) is given by the expression (Clapp and Hornberger 1978)

$$R_g = K_{w0}x_t^{2B+3}, \tag{7}$$

where K_{w0} (m s^{-1}) is the saturated hydraulic conductivity and B is the b exponent in Clapp and Hornberger (1978). The values of K_{w0} (m s^{-1}) and B are listed in Table 1.

The gravitational drainage F_g (m s^{-1}) is included using the expression

$$F_{g_t} = F_{g0} - \left(\frac{x_u}{x_t}\right)^{B+1.5} R_g, \tag{8}$$

where

$$F_{g0} = \alpha D\left(x_t - x_u\frac{h_u}{h_t}\right)\frac{1}{(h_u h_t)^{1/2}} \tag{9}$$

and h_t is the depth of the soil active layer (between 0.5 and 2 m).

The α term is parameterized as in Dickinson et al. (1986), and the average soil diffusivity D is calculated from x_u and x_t as

$$D = 1.02\left(\frac{B\phi_0 K_{w0}}{x_{wm}}\right)^B (x_u)^2 \left(\frac{x_u}{x_t}\right)^{B_f}, \qquad (10)$$

where B_f is a function of B given by Dickinson (1984), x_{wm} is a fraction of saturated soil filled by water, and ϕ_0 is the maximum soil suction, listed in Table 1.

The evaporative term E is very difficult to parameterize. Based on multilayer soil model integrations and theoretical arguments, Dickinson et al. (1986) suggested the following parameterization:

$$E = \min\left[E_0, \frac{\alpha D x_t}{(x_u x_t)^{1/2}}\right], \qquad (11)$$

which, in our opinion, is a suitable choice. In (11), α is parameterized as in Dickinson et al. (1986), and E_0 (m s^{-1}) is the potential evaporation, calculated in accordance with Deardorff (1978) as the evaporation from a wet surface with the same aerodynamic characteristics as the soil surface; for example,

$$E_0 = \rho_a C_D u_r f_g (q_g - q_r), \qquad (12)$$

where ρ_a is the surface air density, C_D is the drag coefficient depending on atmospheric stability of the atmosphere, u_r is the wind speed at the reference level, q_g is the saturated specific humidity at the temperature of the ground surface, q_r is the specific humidity at the reference level, and f_g is the wetness factor, which has the value of 1.0, except for diffusion-limited soil surfaces where it is defined by the ratio of actual-to-potential evaporation.

The transpiration E_t (m s^{-1}) is parameterized following Dickinson et al. (1986):

$$E_t = \sigma_f \rho_a \frac{1-\delta}{r_a + r_s}(q_f^{\mathrm{sat}} - q_{af}), \qquad (13)$$

where r_s is the stomatal resistance, r_a is the aerodynamic resistance, q_f^{sat} is the saturated specific humidity at the foliage temperature, q_{af} is the specific humidity within the canopy, and δ is a power function of the moisture content of the interception reservoir (Deardorff 1978) defined as

$$\delta = \left(\frac{w_d}{w_{\max}}\right)^{2/3}, \qquad (14)$$

where w_d is the total water intercepted by the canopy and $w_{\max}$ is the maximum water the canopy can hold (Deardorff 1978). The aerodynamic resistance r_a is parameterized as

$$r_a = \frac{1}{C_D u_r}. \qquad (15)$$

The stomatal resistance r_s is taken to be

$$r_s = \frac{r_{s\min}}{\mathrm{LAI}} I_1 I_2 I_3 I_4, \qquad (16)$$

where $r_{s\min}$ is the minimum stomatal resistance and LAI is the leaf-area index. Terms I_1, I_2, I_3, and I_4 were parameterized in accordance with Dickinson et al. (1986) and Noilhan and Planton (1989). The factor I_1 varies between about 1 for overhead sun and $r_{s\min}/r_{s\max}$ for nighttime, where $r_{s\max}$ is the cuticular resistance of leaves. It is assumed to have a form (Dickinson 1984)

$$I_1 = \frac{1+f}{f + r_{s\min}/r_{s\max}}, \qquad (17)$$

with $r_{s\max} = 5000$ s m^{-1} and $f = 1.1(S/S_g)(\mathrm{LAI})^{-1}$, where S is the incident shortwave radiation and S_g is a limit value of 30 W m^{-2} for a forest and 100 W m^{-2} for a crop. The moisture factor I_2 depends on the soil moisture and ability of plant roots to take up water from the soil. For the initial value, we set $I_2 = 1$. If the plant transpiration exceeds a maximum value, depending on soil moisture, I_2 is increased so that the transpiration is maintained at the maximum value (Dickinson et al. 1986). The factor I_3 represents the effect of vapor-pressure deficit of the atmosphere. In accordance to Sellers et al. (1986),

$$I_3 = 1 - 0.0025\ \mathrm{hPa}^{-1}\,[e_{\mathrm{sat}}(T_f) - e_r], \qquad (18)$$

where e_{sat} is the saturated vapor pressure at the temperature of the leaf and e_r is the vapor pressure at the reference level. For the seasonal temperature factor I_4, we used (Dickinson 1984)

$$I_4 = 1 - 0.0016(298.0 - T_r), \qquad (19)$$

where T_r is the air temperature at the reference level.

Finally, the fraction of transpiration from the topsoil-layer parameter β in (5a) is parameterized as in Dickinson et al. (1986).

3. Data used in sensitivity tests

The sensitivity tests consisted of running the above model for different fractional covers and nine soil textures. The set of experiments was performed for a single type of plant cover. Once the set of vegetation data and atmospheric conditions were externally prescribed, then we changed only the soil textures. This set of tests was repeated for each fractional vegetation cover at four different growth stages. In our opinion, the experiments designed that way—that is, using the surface fluxes as the output—allowed us to establish the sensitivity of the parameterization scheme for the prescribed soil texture and fractional vegetation cover.

Once we established the model's accuracy with the observations of a single soil texture, we could proceed to investigate the sensitivity of the others. For that purpose, we used real datasets that are the part of a larger measurement program to examine the exchange pro-

TABLE 2. Parameters of maize for the four different growth stages measured at De Sinderhoeve (Jacobs et al. 1990) used in the sensitivity tests.

Parameter	Units	MJ	MA	MS	MO
Maize height	(m)	0.25	2.20	2.30	2.20
Roughness length	(m)	0.03	0.114	0.114	0.114
Displacement height	(m)	0.10	1.76	1.86	1.76
Leaf-area index	(m^2 m^{-2})	0.44	4.3	4.0	2.0
Fractional vegetation cover	—	0.40	0.85	0.80	0.50
Minimum stomatal resistance	(s m^{-1})	200	200	200	200

cesses of heat, mass, and momentum just above and within a maize canopy during its growing season in the De Sinderhoeve. The experimental site was in the center of the Netherlands (51°59′N, 5°45′E). The site was 250 m × 300 m, surrounded by other agricultural fields in which mainly maize was grown. The maize was planted in north-northeast–south-southwest rows, with a row spacing of 0.75 m and with 0.11-m spacing in the row (12 plants per square meter). We selected a few situations that represented the maize in different growth stages: 7 June, 18 August, 8 September, and 4 October, which will be denoted as MJ, MA, MS, and MO, respectively. These datasets were selected because they covered a wide range of fractional covers and the leaf-area index (LAI; one-sided leaf area per unit ground surface). These parameters, together with the maize height, the roughness length, and the displacement height, were measured (Jacobs et al. 1990). The minimum stomatal resistance was not measured and was assumed to be equal to 200 s m^{-1} (Table 2).

The MJ case was an example with a significant fraction of bare ground ($\sigma_f = 0.44$). The meteorological conditions for that day, however, were affected by the amount of cloudiness (0.9 on average), with a visible effect on the net radiation and the diurnal course of the air temperature. The MA and MS cases correspond to the periods of the growing season when the maize plants were tall (2.20 and 2.30 m). As a result, the values of LAI and σ_f increased significantly. During these days, u_r varied between 0.41 m s^{-1} (1.23 m s^{-1}) at night, and 3.28 m s^{-1} (3.83 m s^{-1}) in the afternoon for the MA and MS cases, respectively. The maximum incoming shortwave radiation was 636 W m^{-2} for the MA and 604 W m^{-2} for the MS case. The MO case was already the end of the growing season, and the fraction of the bare soil was visible ($\sigma_f = 0.50$). The incoming shortwave radiation was considerably reduced so that the maximum reached just 409 W m^{-2}.

The texture of soil at the experimental site was very similar to soil texture no. 1 in Table 1 (Jacobs et al. 1990). In the model used, we took 0.01 m for the ground roughness length of the soil underneath vegetation, $h_u = 0.01$ m for the upper, and $h_t = 1.0$ m for the total soil layer. The other textures used in the tests were between loamy sand and sandy clay in accordance with the U.S. Department of Agriculture (USDA) textural classification.

The atmospheric boundary conditions at the reference level $h_r = 4.5$ m were derived from measurements of global radiation, cloudiness, precipitation, specific humidity, and average wind speed for 24 h from 0000 LST at 15-min intervals. These values were interpolated at the beginning of each time step $t = 10$ min. For the downward directed longwave radiation R_L we used the expression proposed by Staley and Jurica (1972),

$$R_L = [c + (1 - c)0.67(1670q_s)^{0.08}]\sigma T_g^4, \quad (20)$$

where c is the cloud fraction, q_s is the specific humidity, and σ is the Stefan–Boltzman constant. All time integrations were started at 0000 LST with the initial values of air pressure of 101 500 (MJ), 101 600 (MA), 102 400 (MS), and 101 600 Pa (MO). The initial values of volumetric soil-moisture content and soil temperature for both layers are listed in Table 3.

4. Discussion and comments

There is already considerable information, as well as numerous studies, on the sensitivity of surface energy fluxes and hydrology to various important vegetation parameters and their changes, but much less information is available about the sensitivity of the land surface parameterization to the choice of soil texture (Mihailovic 1990). Here we will discuss the sensitivity of the land surface parameterization to the soil texture and fractional vegetation cover through the changes in the surface fluxes as the output obtained by time integrations.

TABLE 3. Initial volumetric soil-moisture contents and temperatures used in the sensitivity tests.

Parameter	Units	MJ	MA	MS	MO
Volumetric soil-moisture content for total layer	(m^3 m^{-3})	0.15	0.15	0.15	0.15
Volumetric soil-moisture content for upper layer	(m^3 m^{-3})	0.15	0.14	0.18	0.21
Soil temperature for total layer	(K)	287.16	288.16	287.16	285.16
Soil temperature for upper layer	(K)	283.16	287.16	287.16	283.16

First, we shall establish the model's correspondence with the observations. Figures 2a–d show the diurnal variations of the computed surface energy-balance components for the four chosen situations that represent the maize during the growing season at De Sinderhoeve. It was already mentioned that the soil texture 1 (Table 1) approximately corresponds to the texture of the soil type at the experimental site. The observed values of the latent and sensible heat fluxes for the MA, MS, and MO cases are indicated by circles and stars, respectively. These values were available from the eddy flux measurements by van Pul (1992) as a part of the larger measurement program described above. We had no available observations for the MJ case. The computed latent heat fluxes for the MA case agreed well with the observations. The computed sensible heat fluxes showed a somewhat better agreement with their observed values (Fig. 2b). In the MS case (Fig. 2c), the latent heat fluxes show that the model underestimated the observed values most of the time. The largest differences occurred shortly after noon. The computed sensible heat fluxes, however, agree quite well with the observed one. For the MO case, the model completely underestimated the observed sensible heat fluxes and overestimated the observed latent heat fluxes (Fig. 2d).

As it is well known, the partitioning of energy into sensible and latent heat flux is the process most sensitive to changes in vegetation, as well as in soil parameters. It can be clearly seen through the surface fluxes. We made the time integrations for the MJ, MA, MS, and MO cases with the nine different soil textures. The integration results are shown in Figs. 3–6 in the form of diurnal variations of the surface fluxes and the Bowen ratio. These figures present only the results for the first six soil textures, since the differences in the surface fluxes were negligible when the soil textures 7–9 were used.

The MJ case is an example with a significant fraction of bare ground ($\sigma_f = 0.40$) and a low value of leaf-area index (LAI = 0.44) when the evapotranspiration and the latent heat flux from the ground surface has the same order of magnitude. The latent and sensible heat flux curves (Figs. 3a,b) can be readily classified into two groups. The first one includes soil textures 1–3, and the second group includes the rest of the soil textures used; for example, 4–6. In the first group, there are practically no differences in either the latent heat fluxes or the sensible heat fluxes. The same conclusion can be established for the second group of curves. This splitting of the curves is more evident in Fig. 3d where the variations of the Bowen ratio are shown. The second

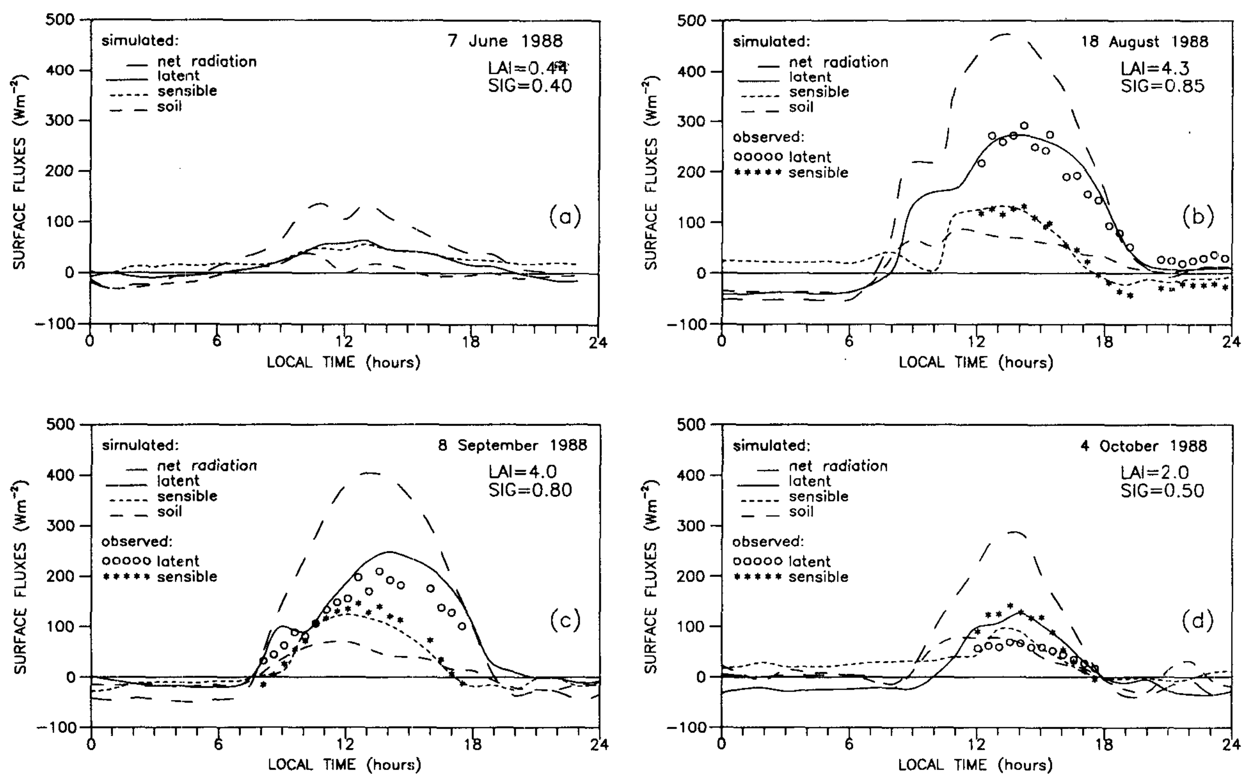

FIG. 2. Diurnal variations of simulated and observed surface fluxes above a maize canopy during growing season at De Sinderhoeve, the Netherlands.

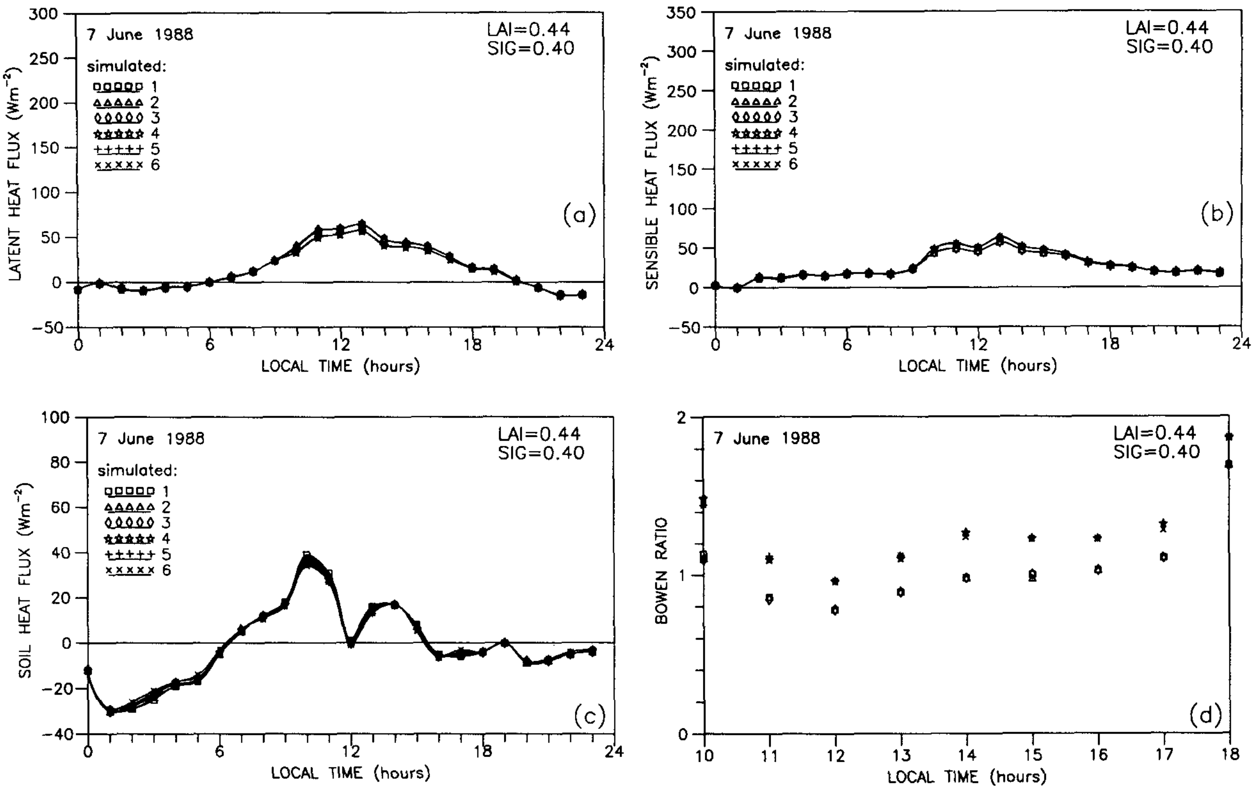

FIG. 3. Diurnal variations of simulated surface fluxes and the Bowen ratio above a maize canopy as a function of different soil textures for 7 June 1988. LAI and SIG denote leaf-area index and fractional vegetation cover, respectively.

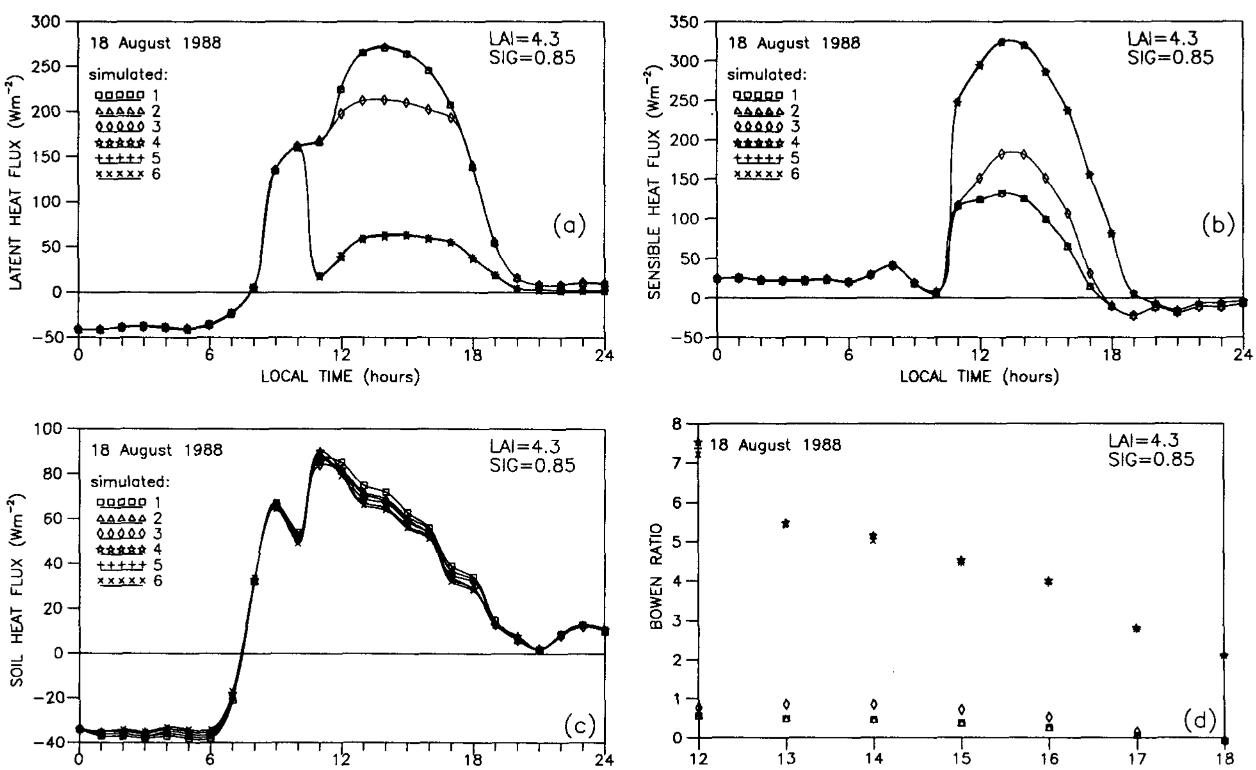

FIG. 4. As in Fig. 3 but for 18 August 1988.

group of curves systematically have higher values of the Bowen ratio than the first one. The differences in the soil heat fluxes are negligible for all soil textures (Fig. 3d).

The MA situation is the case of a dense vegetation ($\sigma_f = 0.85$ and LAI = 4.3) when evapotranspiration significantly prevails over the latent heat flux from ground underneath the vegetation. The computed values of the latent, sensible, and soil heat fluxes and the Bowen ratio are shown in Figs. 4a–d. As in the MJ case, we can clearly see their grouping into two groups. The only exception was the soil texture 3. In that case, lower values of the latent heat flux and higher values of the sensible heat flux occurred in comparison with the first group of curves (1 and 2). The computed values of the Bowen ratio show variation, between 0 and 1, for the first group of curves (Fig. 4d). The considerably higher values for the second group indicate that the sensible heat flux is more dominant in the energy-exchange processes when the soil textures 4–6 are used. The soil heat flux curves show the tendency of splitting depending on soil texture. In Fig. 4a, the curves are sorted from the highest value (texture 1) to the lowest (texture 6).

In the MS case, the vegetation is still very dense (σ_f = 0.80, LAI = 4.0). The time integrations with the soil textures 1–3 give the same amounts for the latent heat fluxes (Fig. 5a). For the textures 4–9, however, the computed values of the latent heat fluxes show a tendency of splitting with a dispersion in the range from 5 to 10 W m^{-2} for 1000–1800 LST. The values of the computed sensible heat fluxes (Fig. 5c) and the Bowen ratio (Fig. 5d) follow this dispersion. Figure 5d shows that the soil heat flux curves are stilted and arranged from higher values (texture 9) to lower (texture 1).

In the MO case, the density of the maize canopy is significantly reduced ($\sigma_f = 0.40$, LAI = 2.0), so that the evapotranspiration has approximately the same order of magnitude as the latent heat flux from ground underneath the vegetation. Figure 6a shows that there are no differences in the computed values of the sensible heat fluxes for soil textures 1–3. A simple scrutiny of the second group of curves clearly shows a tendency of splitting, with a dispersion ranging from 8 to 10 W m^{-2}. This splitting is also characteristic for the computed values of the sensible heat fluxes and the Bowen ratio (Figs. 6b and 6d). The difference in the amounts of the computed soil heat fluxes occur in the time interval 1000–1700 LST (Fig. 6c).

This brief discussion of each situation undoubtedly points out that the computed latent and sensible heat fluxes can be classified into two groups. The first one

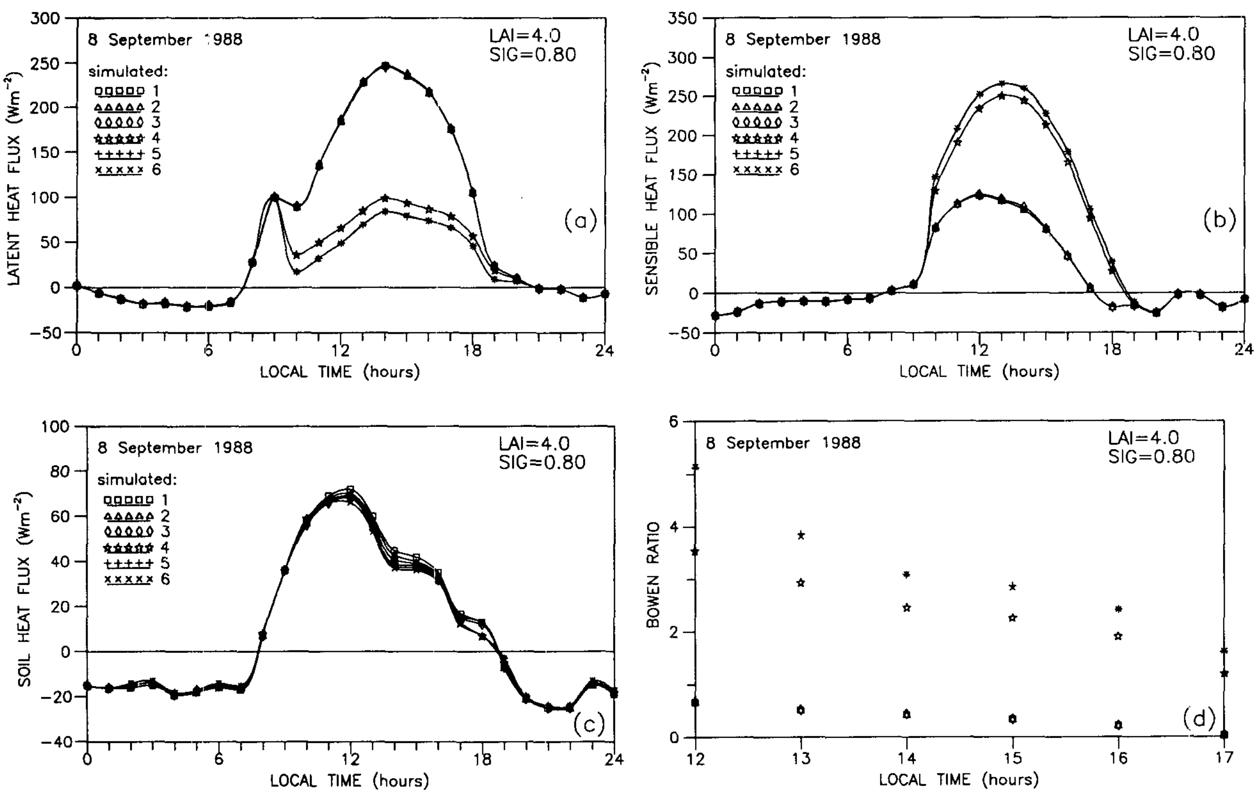

FIG. 5. As in Fig. 3 but for 8 September 1988.

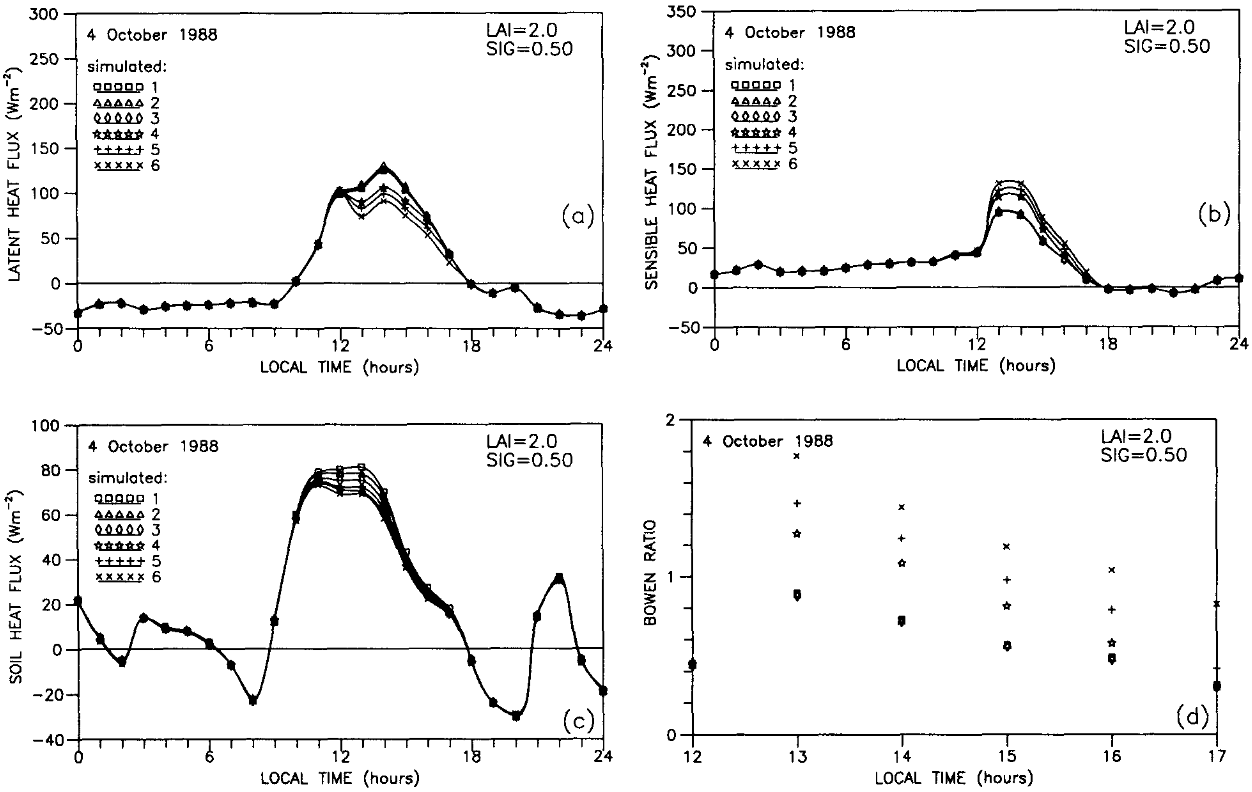

FIG. 6. As in Fig. 3 but for 4 October 1988.

includes the experiments performed with soil textures 1–3, and the second group includes the other soil textures 4–9. In the first group, there are practically no differences in the amount of latent heat flux, regardless of the fractional cover used. This correspondence between the curves seems to be a result of the presence of a very dense vegetation. Then the evapotranspiration is a more dominant process and consequently prevails over the latent heat flux from the ground underneath the vegetation (MA and MS). This correspondence could, however, result from the fact that the eventual differences in the latent heat fluxes are due to the different soil textures used and cannot be separated by the model or the differences are negligible. We support the second possibility, since the MO and MJ experiments support this. In these situations, the evapotranspiration and the latent heat flux from the ground surface have the same order of magnitude. The values of latent heat flux, however, stay practically the same regardless of the soil texture used.

These quantitative descriptions are more understandable if we use Fig. 7, where the normalized diurnal latent heat and sensible heat sums are plotted against the saturated hydraulic conductivity. This basic soil parameter decreases when the soil texture becomes finer, from the coarsest texture (sand) to the finest (clay). To make the normalization of the latent and sensible heat flux sums, we introduced a factor of normalization α_{ij} (FON hereafter), which is defined as

$$\alpha_{ij} = \frac{S_{ij}}{\max\limits_{j=1,M}(S_{ij})} \quad i = 1, 2, 3 \cdots N, \tag{21}$$

where the subscript i indicates the type of the daily sum of surface flux and the subscript j denotes type of soil texture. The introduced factor FON varies from 0 to 1.

If we plot FON of a surface flux S_{ij} against the soil parameters ξ on which it depends, then the domain of its dependence ($\xi_1 < \xi < \xi_2$) can be considered as

$$\mathrm{FON}[S_{ij}(\xi)] = \begin{cases} a\xi + b & (22) \\ c, & (23) \end{cases}$$

where a, b, and c are some arbitrary constants. Equation (22) represents a case of linear dependence of S_{ij} on ξ. It is evident that the degree of linear dependence is governed by the slope of the line. It varies from a strong dependence (the slope goes to infinity) to a weak dependence (the slope tends to zero). Otherwise, we have a case of a strong independence of S_{ij} on ξ as indicated by (23).

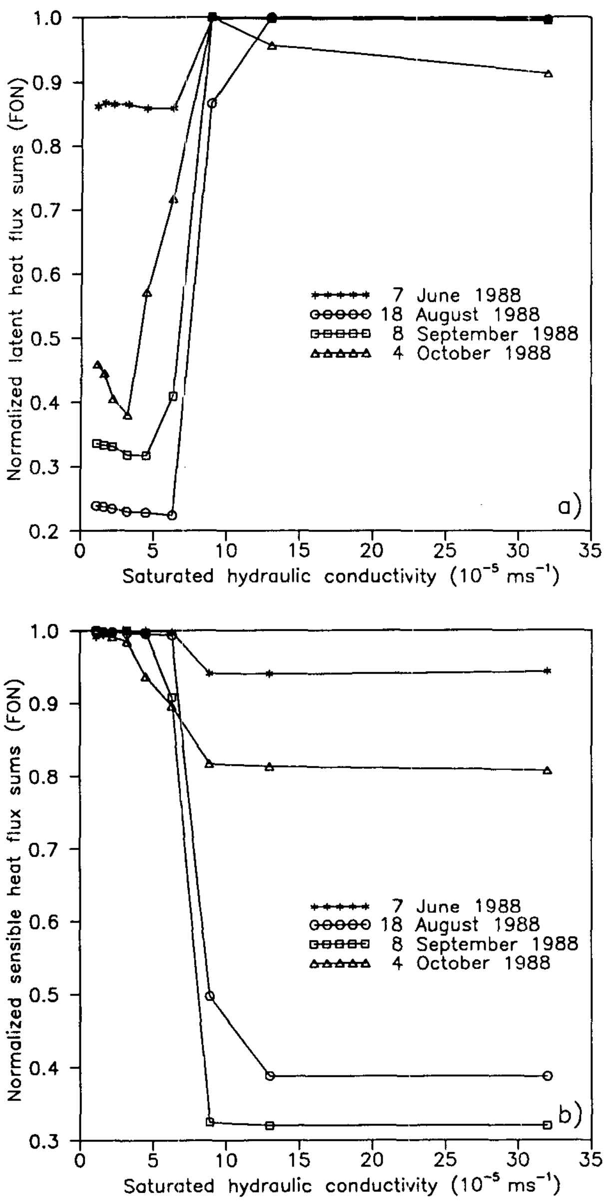

FIG. 7. Normalized values of the latent (a) and sensible (b) daily flux sums in Figs. 3–6. FON is the factor of normalization.

Figure 7 shows the variations of the FON factor for latent and sensible heat fluxes as a function of saturated hydraulic conductivity. Figure 7a shows clearly that the FON factor of latent heat flux can be considered as a case of linear dependence. Practically, for the first three textures, FON is kept in the domain of strong independence, with a value of 1, except for the MO situation, when it goes into the regime of a weak dependence. Regardless of this small discrepancy, we still can conclude that changes in soil texture from loamy sand (texture 1) to silt loam (texture 3) do not induce large changes in the latent heat flux. Moreover, different fractional vegetation cover and LAI values produce latent heat fluxes that are practically independent of the mentioned soil textures. Further inspection of Fig. 7a indicates a leap of the FON factor between the textures of silty loam (texture 3) and loam (texture 4), for all cases, with an earlier increase for the MA situation. In accordance with the introduced definitions, the FON factor can be considered highly dependent in this domain. For soil textures from loam (texture 4) toward the silty clay (texture 9), however, the FON factor shows a tendency of high independence (the MJ and MA cases), or near high independence (the MS case). The only exception in the considered cases is the FON factor for the MO case, which represents a case with the LAI between the MJ and MA and MS cases. The differences between the average values of FON factors for the latent heat flux (0.25 for MA, 0.32 for MS, 0.40 for MO, and 0.88 for MJ) are a result of the presence of different fractional vegetative covers, and of LAI from the lowest to the highest value. It means that in this domain the latent heat fluxes depend on LAI and fractional cover, but not on soil texture.

Similar consideration of the FON factors for the sensible heat fluxes, shown in Fig. 7b, can also be made. Two groups of the textural classes are separated by a leap of FON factors for each situation. For the first group (1–3), sensible heat fluxes depend on LAI and fractional cover but not on soil texture. For the second group (4–9), sensible heat fluxes depend neither on LAI and fractional cover nor on soil texture. As in Fig. 7a, some small discrepancies are also evident in Fig. 7b.

This short analysis supports the well-known fact that surface fluxes are strongly dependent on soil texture. The contribution can be summarized as follows. Finer classifications of soil textures from sand to clay indicate that there exist two regimes of latent heat flux variability, where the latent heat flux is 1) independent regardless of the fractional cover, the LAI of the surface, and the soil texture (from loamy sand to silt loam), and 2) dependent on the fractional cover and the LAI, but independent of the soil texture (from loam to silty clay).

Acknowledgments. The authors thank Addo van Pul for supplying the experimental data for maize. Several people contributed to the collection of this dataset: Adri Jacobs, and the technical staff of the Department of Meteorology of the Wageningen Agricultural University. Part of the work was carried out while the first author spent four months at the Wageningen Agricultural University as a visiting scientist. The government of the Netherlands supported this study financially.

REFERENCES

Beven, K., 1979: A sensitivity analysis of the Penman-Monteith actual evapotranspiration estimates. *J. Hydrol.*, **44**, 169–190.

Clapp, R. B., and G. M. Hornberger, 1978: Empirical equations for some soil hydraulic properties. *Water Resour. Res.*, **14**, 601–604.

Deardorff, J. W., 1978: Efficient prediction of ground surface temperature and moisture with inclusion of a layer vegetation. *J. Geophys. Res.*, **83**, 1889–1903.

Dickinson, R. E., 1984: Modeling evapotranspiration for three-dimensional global climate models. *Climate Processes and Climate Sensitivity*, J. E. Hansen and T. Takahashi, Eds., Amer. Geophys. Union, 58–72.

——, J. Jager, W. M. Washington, and R. Wolski, 1981: Boundary subroutine for the NCAR global climate model. NCAR Tech. Note, 173+1A, 75 pp.

——, A. Henderson-Sellers, P. J. Kennedy, and M. F. Wilson, 1986: Biosphere–Atmosphere Transfer Scheme (BATS) for the NCAR community climate model. NCAR Tech. Note, NCAR/TN-275+STR, 69 pp.

Jacobs, A. F. G., W. A. J. van Pul, and A. van Dijken, 1990: Similarity moisture dew profiles within a corn canopy. *J. Appl. Meteor.*, **29**, 1300–1306.

Luxmoore, R. J., J. L. Staley, and J. T. Holdeman, 1981: Sensitivity of a soil–plant–atmosphere model to changes in air temperature, dew point temperature and solar radiation. *Agric. Meteor.*, **23**, 115–129.

McCumber, M. C., 1980: A numerical simulation of the influence of heat and moisture fluxes upon mesoscale circulation. Ph.D. dissertation, University of Virginia, Charlottesville, 255 pp.

Mihailovic, D. T., 1990: Testing the Biosphere–Atmosphere Transfer Scheme (BATS) using Penman and Long data (1960)—preliminary results. Wageningen Agricultural University, Tech. Note 1990, 40 pp.

Noilhan, J., and S. Planton, 1989: A simple parameterization of land surface processes for meteorological models. *Mon. Wea. Rev.*, **117**, 536–549.

Saxton, K. E., 1975: Sensitivity analyses of the combination evapotranspiration equation. *Agric. Meteor.*, **15**, 343–353.

Sellers, P. J., and J. G. Lockwood, 1981: A numerical simulation of the effects of changing vegetation type on surface hydroclimatology. *Climatic Change.*, **3**, 121–136.

——, Y. Mintz, Y. C. Sud, and A. Dalcher, 1986: A simple biosphere model (SiB) for use within general circulation models. *J. Atmos. Sci.*, **43**, 505–531.

Staley, D. O., and G. M. Jurica, 1972: Effective atmospheric emissivity under clear skies. *J. Appl. Meteor.*, **11**, 349–356.

Thom, A. S., 1975: Momentum mass, and heat exchange of plant communities. *Vegetation and the Atmosphere*, Vol. 1, J. L. Monteith, Ed., Academic Press, 57–109.

van Pul, W. A. J., 1992: The flux of ozone to a maize crop and the underlying soil during a growing season. Ph.D. thesis, Wageningen Agricultural University, 147pp.

Vries, D. A. de, 1963: Thermal properties of soils. *Climatology and Microclimat.* North-Holland Publishing Co., 210–235.

Wilson, M. F., and A. Henderson-Sellers, 1985: Land cover and soils data sets for use in general circulation climate models. *J. Climatol.*, **5**, 119–143.

——, ——, R. E. Dickinson, and P. J. Kennedy, 1987: Sensitivity of the Biosphere–Atmosphere Transfer Scheme (BATS) to the inclusion of variable soil characteristics. *J. Appl. Meteor.*, **26**, 341–362.