The Simulation of Surface Heat Fluxes in a Land Surface–Atmosphere Model

HUANG XINMEI AND T. J. LYONS

Environmental Science, School of Biological and Environmental Sciences, Murdoch University, Murdoch, Western Australia, Australia

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ABSTRACT

A modified soil–canopy–boundary layer model based on Ek and Mahrt is presented to simulate surface heat fluxes on a daily basis. The model is validated against independent published bare soil data from Agassiz, Canada, and observations taken over sparse and dense vegetation as part of the HAPEX-MOBILHY (Hydrological Atmospheric Pilot Experiment-Modélisation du Bilan Hydrique) experiment. Simulated surface energy components are shown to be in close agreement with the observations, and the sensitivity of the parameterization is evaluated.

1. Introduction

The interaction between the atmosphere and underlying land surface has been noted by many researchers. Observations and numerical models illustrate that the fluxes of radiation, sensible and latent heat, and momentum are deeply affected by land surface properties (Sato et al. 1989). As this influence extends from the microscale through the mesoscale to the global circulation, much work has been devoted to developing appropriate land surface–atmosphere parameterizations for use in numerical meteorological models (i.e., Dickinson 1984; Sellers et al. 1986; Mahrt and Pan 1984; Noilhan and Planton 1989). These have been reviewed by Avissar and Verstraete (1990).

A one-dimensional, soil–canopy–boundary layer model originally developed by Oregon State University (Ek and Mahrt 1989), OSU1, has been used to simulate the interaction between the land surface and the atmosphere and has been incorporated into other regional or global models. It calculates the flux of water vapor to the atmosphere separately from both soil and vegetation. Soil evaporation is estimated by a threshold formulation, whereas canopy transpiration is computed as a function of soil water content. However, such techniques may not estimate the water vapor flux correctly as transpiration also depends on solar radiation, air water deficit, and air temperature (Deardorff 1978; Dickinson 1984; Noilhan and Planton 1989). Mahfouf and Noilhan (1991) have found that threshold methods strongly underestimate bare soil evaporation.

Also, OSU1 does not account for the ground heat flux difference between bare soil and a vegetated surface. Since ground heat flux is generally smaller for vegetated surfaces in comparison with bare soil surfaces under the same net radiation conditions (Clothier et al. 1986; Choudhury et al. 1987), the computation of ground heat flux for vegetated land could be overestimated if the bare soil heat flux scheme is used.

Thus, the ground and latent heat flux parameterizations of OSU1 (Ek and Mahrt 1989) have been modified to simulate the surface energy balance on a daily basis over a snow-free surface. A comparative study of the parameterization of these fluxes between the original scheme and the modified scheme is presented. Surface heat fluxes calculated from both schemes are compared with independent field observations. Subsequent papers will describe the application of this model to addressing the meteorological impact of the replacement of native perennial vegetation by agriculture in southwestern Australia (Smith et al. 1992; Lyons et al. 1993).

2. Model

OSU1 is a soil–canopy–boundary layer model (Ek and Mahrt 1989) that has been developed and examined by Mahrt and Pan (1984), Mahrt et al. (1984), Troen and Mahrt (1986), Pan and Mahrt (1987), and Mahrt et al. (1987). It has two soil layers to account for soil evaporation and canopy transpiration.

In OSU1, ground heat flux G is computed from the soil temperature gradient through

$$G = K\frac{T_{so} - T_{so1}}{0.5z_{so1}}, \quad (1)$$

where K (W m^{-1} K^{-1}) is the thermal conductivity, T_{so} and T_{so1} the temperatures at the soil surface and midpoint of the first soil layer, respectively, and z_{so1} the

Corresponding author address: Dr. T. J. Lyons, School of Biological and Environmental Sciences, Murdoch University, Murdoch, WA 6150, Australia.

depth of the first soil layer (the full soil model is described in the appendix).

Over a dense crop surface, the ground heat flux has a smaller value than over bare soil under the same net radiation conditions (Clothier et al. 1986; Kustas and Daughtry 1990), suggesting that Eq. (1) could overestimate the ground heat flux over vegetated land. This has been addressed in the modified model, NOSU, whereby the ground heat flux is estimated separately for the full vegetated surface and the bare soil. The total ground heat flux over partially vegetated land is weighted by the fraction of vegetation cover F_c.

In NOSU, ground heat flux over bare soil G_{so}, is computed from Eq. (1), whereas under a fully vegetated area G_c, the heat flux is taken as proportional to the net radiation R_n; that is,

$$G_c = C_c R_n. \tag{2}$$

Observations of ground heat flux reviewed by Choudhury et al. (1987) show that for a mature arable crop, G/R_n ranges from 5% to 10%, whereas for natural grasslands and irrigated lawn it ranges from 10% to 20%. From field observations of spring wheat, they suggested an empirical relationship of the form

$$G = 0.4R_n \exp(-0.5\ \mathrm{LAI}), \tag{3}$$

where LAI is the leaf area index. Thus, for full canopy cover in Eq. (2), C_c would be written as

$$C_c = 0.4 \exp(-0.5\ \mathrm{LAIF}), \tag{4}$$

where LAIF is leaf area index under fully vegetated conditions, which is related to LAI by the fraction of vegetation cover F_c,

$$\mathrm{LAIF} = \frac{\mathrm{LAI}}{F_c}. \tag{5}$$

Under this parameterization, the ground heat flux for a fully vegetated surface is not functionally dependent on the soil temperature gradient. When the land surface is partially vegetated, the ground heat flux is taken as a weighted fraction of the vegetation cover; that is,

TABLE 1. Input parameters for model calculation at the Agassiz site for 30 May 1978.

Soil type	Albedo	$T(0)$ (°C)	$q(0)$ (g kg^{-1})
Slit loam	0.075	10	5.9

Initial soil volumetric water content (m^3 m^{-3}) at the Agassiz site.

z (m)	−0.05	−0.1	−0.20	−0.5	−1.0
	0.37	0.43	0.44	0.44	0.43

TABLE 2. Input parameters for the four case studies simulated from HAPEX-MOBILHY.

	Case 1	Case 2	Case 3	Case 4
Day	16 June	16 June	28 June	16 June
Site	Castelnau	Lubbon 2	Lubbon 1	Estampon
Soil	sandy clay loam	sand	sand	sand
Vegetation	maize	maize	oat	forest
z_0 (m)	0.02	0.10	0.15	1.0
Albedo		0.15		0.10
LAI	0.3	2.0	3.0	3.5
R_{smin} (s m^{-1})	40	40	600	100
F_c	0.40	0.80	0.90	0.99
W_{so1} (m^3 m^{-3})	0.18	0.17	0.10	0.14
W_c (m^3 m^{-3})	0.25	0.17	0.22	0.20

$$G = F_c G_c + (1 - F_c) G_{so}, \tag{6}$$

where G_{so} is defined by Eq. (1) and G_c by Eq. (2).

In OSU1, the latent heat flux LE at the surface is determined by the canopy transpiration and direct soil evaporation as

$$LE = F_c LE_c + (1 - F_c) LE_s, \tag{7}$$

where E_c and E_s represent the transpiration of the canopy and evaporation from the soil, respectively, and L (J kg^{-1}) is the latent heat associated with the change of phase. Following Mahrt and Pan (1984), soil evaporation is evaluated from

$$E_s = \left[D(W_{so}) \frac{\partial W_{so}}{\partial z} \right]_{z=0} + K_w(W_{so})|_{z=0}, \tag{8}$$

where D (m^2 s^{-1}) is the coefficient of diffusivity, K_w (m s^{-1}) the hydraulic conductivity, and W_{so} (z = 0) (m^{-3} m^{-3}) the surface volumetric soil water content.

Assuming that the soil is undergoing potential evaporation E_p, Eq. (8) can be solved for surface soil moisture W_{so} (z = 0). If the surface soil moisture is greater than the air dry value W_d, soil evaporation is equal to the potential rate, whereas if the surface soil moisture is less than W_d, soil evaporation is computed from Eq. (8) by setting $W_{so}(z = 0) = W_d$.

Within OSU1, potential evaporation is defined through the surface energy balance,

$$S_g(1 - A) + F_d - \sigma T_p^4 = G + H_p + LE_p, \tag{9}$$

with

$$H_p = \rho C_h C_p(\theta_p - \theta_1) \tag{10}$$

$$E_p = \rho C_h[q_s(T_p) - q_1], \tag{11}$$

where S_g (W m^{-2}) is the downward solar radiation, A the albedo, F_d the downward longwave radiation, σ the Stefan–Boltzmann constant, ρ the air density, C_p the specific heat for air, C_h the heat exchange coefficient, T_p and θ_p the apparent temperature and potential temperature, respectively, if the surface is so wet that potential evaporation occurs, and q_1 and θ_1 the specific humidity and potential temperature, respectively, of the first air level.

The exchange coefficients are defined as (Louis 1979; Louis et al. 1982)

$$C_h = k^2 u\,\mathrm{GH}(z, z_0, \mathrm{Ri}_B)\left[\ln\left(\frac{z}{z_0}\right)\ln\left(\frac{z}{z_0}\right)\right]^{-1} \quad (12)$$

where z_0 is the surface roughness, and k the von Kármán constant.

The function GH is defined as

$$\mathrm{GH} = \begin{cases} 1 - 15\mathrm{Ri}_B\left[1 + 7.5\dfrac{k^2}{\ln(z/z_0)\ln(z/z_0)}10\left(-\mathrm{Ri}_B\dfrac{z}{z_0}\right)^{0.5}\right]^{-1}, & \mathrm{Ri}_B < 0 \\ e^{-\mathrm{Ri}_B}, & \mathrm{Ri}_B \geqslant 0 \end{cases} \quad (13)$$

where Ri_B is the bulk Richardson number for the surface layer, which is given by

$$\mathrm{Ri}_B = \frac{gz[\theta_v(z) - \theta_v(z_0)]}{\theta_v(z)u^2}, \quad (14)$$

where θ_v is the virtual potential temperature and u the wind speed. Surface roughness for heat and water vapor are set to be the same as z_0.

Transpiration depends on the potential evaporation, soil water content, and a plant coefficient,

$$E_c = E_p P_c F_2, \quad (15)$$

where P_c, the plant coefficient, has values between 0 and 1, and F_2 is a function of soil water content.

It is common to express the transpiration in terms of canopy resistance,

$$E_c = \frac{\rho[q_s(T_c) - q_1]}{r_c + 1/C_h}, \quad (16)$$

where r_c is the canopy resistance, $q_s(T_c)$ the saturated specific humidity of the canopy, and T_c the canopy temperature.

By equating the expression for transpiration given by Monteith (1965), OSU1 relates P_c to canopy resistance as

$$P_c = \frac{RR + \Delta}{r_c RRC_h + (\Delta + RR)} \quad (17)$$

with

$$RR = \frac{4\sigma T_1^3}{\rho C_h C_p} + 1$$

$$\Delta = \frac{L}{C_p}\frac{dq_s(T_1)}{dT}, \quad (18)$$

where either P_c or r_c is assumed constant.

From a comparison with observations, Mahfouf and Noilhan (1991) found that aerodynamic methods provided comparable estimates of soil evaporation, while threshold formulations, based on the concept of water supply and demand, underestimated bare soil evaporation. They also suggested that this formulation of Mahrt and Ek (1984) was more sensitive to potential evaporation due to the overestimation of surface temperature calculated from the surface energy balance equation. Hence, in NOSU, bare soil evaporation has been calculated by a bulk aerodynamic method (Lee and Pielke 1992),

$$LE_s = \rho L C_h \beta[q_s(T_{so}) - q_1] \quad (19)$$

with

$$\beta = \begin{cases} 1, & W_{so1} \geqslant W_{fc} \\ 0.25\left[1 - \cos\left(\dfrac{W_{so1}}{W_{fc}}\pi\right)\right]^2, & W_{so1} < W_{fc} \end{cases}, \quad (20)$$

where W_{so1} is the water content of soil in the first soil layer, and W_{fc} the field capacity.

In NOSU, transpiration is modified using Eq. (16). The canopy resistance expression suggested by Noilhan and Planton (1989) is used, which means transpiration is estimated not only as a function of soil water content, but also a function of photosynthesis, leaf water potential, and meteorological condition through

$$r_c = \frac{R_{\min}}{\mathrm{LAI}}\frac{F_1}{F_2 F_3 F_4}, \quad (21)$$

where $R_{\min}$ is the minimum stomatal resistance.

The factor F_1 accounts for the influence of the photosynthetically active radiation, F_2 is a function of soil water content, whereas F_3 accounts for the effect of air vapor deficit on the canopy resistance, and F_4 is the dependence of canopy resistance on air temperature.

The factors F_1 and F_4 are evaluated following Noilhan and Planton (1989), and F_3 is taken from Jacquemin and Noilhan (1990). Factor F_2 is represented as (Pan and Mahrt 1987)

$$F_2 = \frac{1}{z_n}\sum_{i=1}^{i=n} g(z_i)(z_i - z_{i-1}) \quad (22)$$

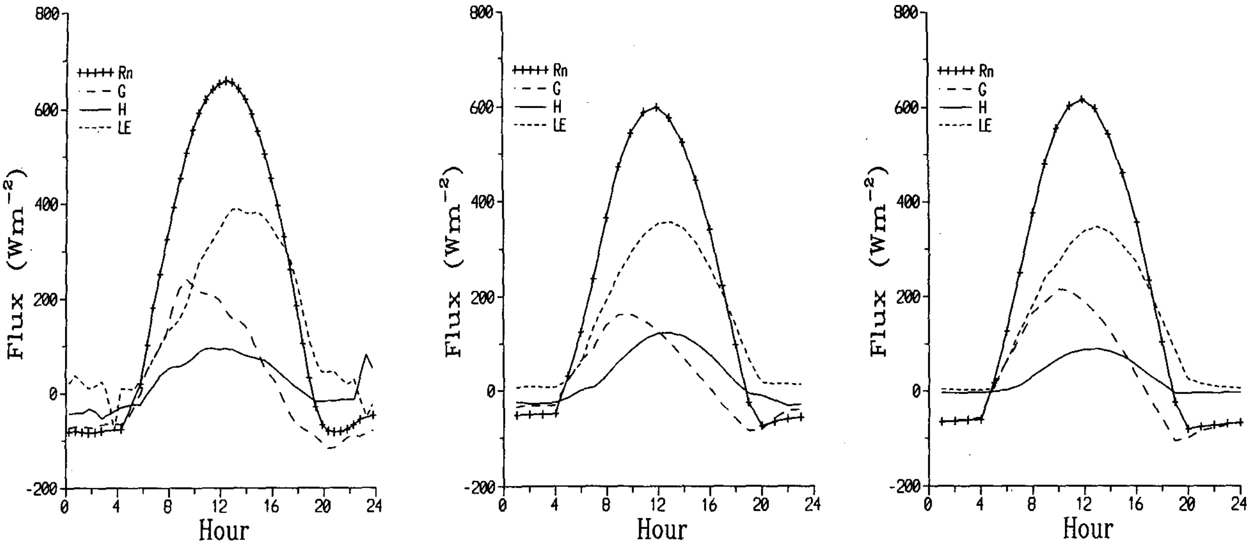

FIG. 1. Surface heat fluxes for 30 May 1978 at Agassiz as (a) observed and simulated by (b) OSU1 and (c) NOSU.

with

$$g(z_i) = \begin{cases} 1, & W_{soi} > W_{cr} \\ \dfrac{W_{soi} - W_{wilt}}{W_{cr} - W_{wilt}}, & W_{wilt} \leqslant W_{soi} \leqslant W_{cr}\,, \\ 0, & W_{soi} < W_{wilt} \end{cases} \quad (23)$$

where W_{soi} is the soil water content in z_i soil level, n the number of soil layers, W_{cr} the critical value of soil water stress (set equal to $0.75W_{sat}$), and W_{sat} and W_{wilt} are the saturated and wilting point of the volumetric moisture content of the soil, respectively.

OSU1 assumes that $T_{so} = T_c$, and these are evaluated from the surface energy balance similar to Eq. (9). However, since we compute the ground and latent heat fluxes separately for bare soil and vegetation, T_{so} (the bare soil temperature) and T_c (the canopy surface temperature) are evaluated separately in NOSU by closure of the respective surface energy balance equations. That is, for bare soil,

$$S_g(1 - A) + F_d - \sigma T_{so}^4 = G_{so} + H_{so} + LE_s, \quad (24)$$

and for a vegetated canopy,

$$S_g(1 - A) + F_d - \sigma T_c^4 = G_c + H_c + LE_c. \quad (25)$$

Huang et al. (1993) found that using the aerodynamic potential temperature $\theta(z_0)$ instead of the surface potential temperature improved the estimated sensible heat flux. Thus, the canopy sensible heat flux H_c is expressed as

$$H_c = \rho C_p C_h[\theta_c(z_0) - \theta_1], \quad (26)$$

where ρ is the air density and C_p the specific heat of air. The difference between $\theta(z_0)$ and θ_c is a function of θ_* and $R_0 = u^*z_0/\nu$. When $R_0 \leqslant 1000$, following Zilitinkevich (1970) and Mahrer and Pielke (1977), the expression for $\theta(z_0)$ is

$$\theta(z_0) = \theta_c + 0.0962\,\frac{\theta_*}{k}\left(\frac{u_*z_0}{\nu}\right)^{0.45}, \quad (27)$$

where ν is the dynamic viscosity coefficient of air and θ_* is the temperature scale. When $R_0 > 1000$,

$$\theta(z_0) = \theta_c + \frac{2.15\theta_*}{k}. \quad (28)$$

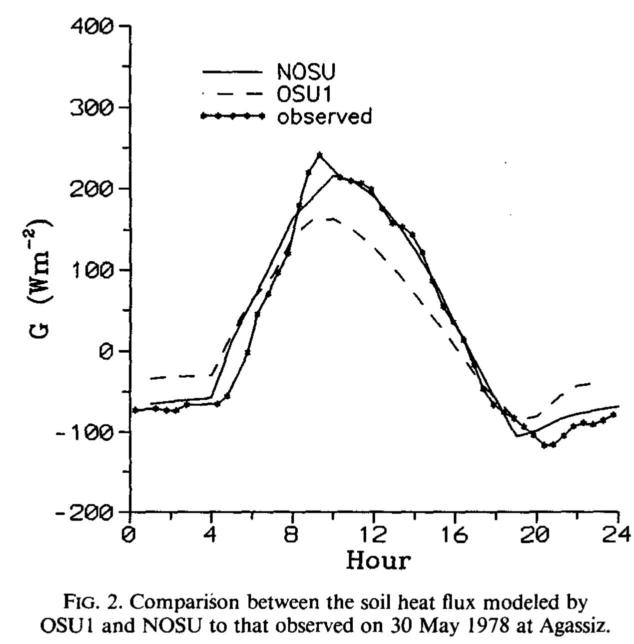

FIG. 2. Comparison between the soil heat flux modeled by OSU1 and NOSU to that observed on 30 May 1978 at Agassiz.

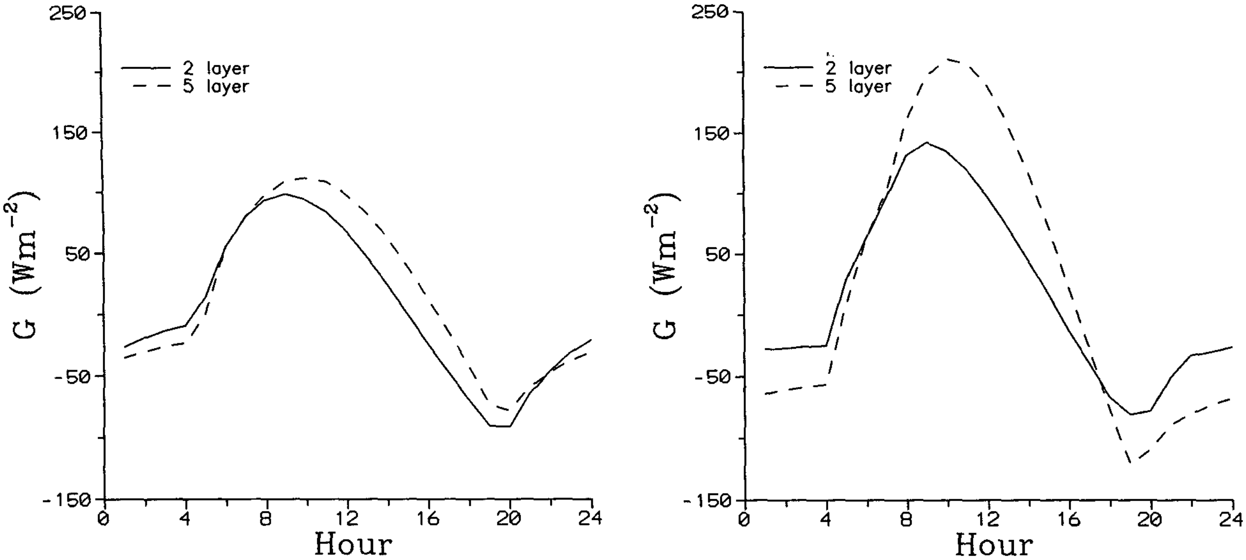

FIG. 3. Influence of the number of soil layers on the bare soil heat flux calculated by OSU1 (a) initial soil water content $W_{so} = 0.10$ m^3 m^{-3} and (b) $W_{so} = 0.35$ m^3 m^{-3}.

The term H_{so} is defined similarly to H_c by substituting θ_{so} for θ_c, and the total sensible heat flux H is expressed as

$$H = F_c H_c + (1 - F_c) H_{so}, \quad (29)$$

where θ_{so} and θ_c are the potential temperatures for the soil and canopy surfaces, respectively.

3. Results and discussion

Both OSU1 and NOSU were evaluated against field data for bare soil and sparsely and densely vegetated cases. The bare soil observations used were the 0.5-h-average surface energy budget measurements obtained at Agassiz, British Colombia, Canada (49.25°N, 121.77°W) on 30 May 1978. During this experiment, solar radiation was measured with a Kipp and Zonen CM5 pyranometer and net radiation was measured with a Swissteco radiometer (Novak and Black 1983). The ground heat flux G was calculated using the null-alignment method (Kimball and Jackson 1975), whereas LE and H were calculated from the measured Bowen ratio and closure of the surface energy balance (Novak and Black 1983).

A z_0 of 0.0004 m, as used by Novak (1991), was adopted and the geostrophic wind speed was estimated as 3.8 m s^{-1}. Other input parameters are listed in Table 1. As atmospheric profiles of temperature and humidity were not available, profiles given by

$$T(z) = \begin{cases} T(0) + 0.0067z, & z \leqslant 30 \text{ m} \\ T(30) - 0.0065(z - 30), & z > 30 \text{ m} \end{cases} \quad (30)$$

$$q(z) = \begin{cases} q(0), & z \leqslant 30 \text{ m} \\ q(0) - \dfrac{q(0)}{5000}(z - 30), & z > 30 \text{ m} \end{cases} \quad (31)$$

were assumed to initialize the model in accord with the analysis of Novak (1991). The initial soil temperature was set equal to $T(0)$ (Table 1) and the volumetric soil water content measured at midday (Table 1) were assumed to be the initial profiles.

The sparse and dense vegetation case studies used data collected during HAPEX-MOBILHY (Hydrological Atmospheric Pilot Experiment-Modélisation du Bilan Hydrique). HAPEX-MOBILHY was designed to study evaporation and transpiration processes over the land surface of southern France and a preliminary analysis of this experiment was reported by Andre et al. (1988).

Surface and soil parameters for the four case studies from HAPEX-MOBILHY are listed in Table 2. Soil types were adopted from Andre et al. (1988), whereas the other parameters for cases 1, 2, and 4 follow Noilhan and Planton (1989), except that the LAI of 3.5 for case 4 is based on Pinty et al. (1989) (2.3 for pine canopy and 1.2 for the bracken understorey). For case 3, the first level of soil water content W_{so1} is from Mehrez et al. (1992), whereas the other parameters follow the analysis of Mahfouf (1990).

To initialize the five-layer soil model, we use the soil water content at root zone W_r as the water content at layers 3, 4, and 5. For layer 2, the soil water content was interpolated from W_{soi1} and W_r. For the OSU1 simulation, $R_c = R_{s\min}$.

The initial atmospheric temperature and humidity profile were from the 0600 UTC radiosonde. Wind speed measured at 3000 m was used as the geostropic wind speed. Initial soil temperature was set to the lowest-model-level air temperature.

Observed surface heat fluxes were measured by the SAMER (Système Automatique de Mesure de l'Evapotranspiration Régionale) station and forest tower. Observed heat fluxes for case 2 were adopted from Noilhan and Planton (1989), whereas cases 1 and 3 used the more recent recalibrated data (M. Ek 1994, personal communication). For cases 1 and 3, measured net radiation was used to force the models rather than compute the net radiation.

Figure 1 shows the observed surface energy components and associated model results for the bare soil case of 30 May 1978, where the model has been initialized from the morning sounding. A value of 0.07 $m^3 m^{-3}$ for the air dry value W_d was assumed in OSU1. The estimated latent heat flux in the afternoon is smaller than that observed, which may be directly attributable to the lower simulated net radiation. Both modeled latent and sensible heat fluxes give similar results, although NOSU provides the closest simulation of the observations. Ground heat flux is underestimated by OSU1, and NOSU provides a better simulation of ground heat flux for both daytime and nighttime (Fig. 2).

The impact of the number of soil layers on bare soil heat flux for wet and dry conditions is illustrated in Figs. 3 and 4, where the model configurations are listed in Table 3. Soil heat fluxes calculated by the two-layer versions (OSU-2L) of both OSU1 and NOSU are lower than the five-layer versions (OSU-5L), and this difference increases when the soil is wet. Since the difference between five and nine layers in NOSU (Fig. 4) is not significant, we have adopted five layers for computational efficiency.

Figure 5 illustrates the sensitivity of NOSU to different soil evaporation schemes, namely those of Lee and Pielke (1992) (henceforth LP92) and Mahrt and Pan (1984) (henceforth MP84), where the initial atmospheric conditions for 30 May at Agassiz have been used and we have assumed an initial constant soil water content profile. Under MP84, evaporation is very sensitive to soil water content when the soil is dry, especially in the transfer from potential evaporation to soil water supply limited evaporation. Such sensitivity is not as apparent in LP92.

TABLE 3. Soil levels used in each configuration of the model.

Number of soil layer	Grid of level (m)
2	z: −0.05, −1.0
5	z: −0.05, −0.1, −0.2, −0.5, −1.0
9	z: −0.025, −0.06, −0.10, −0.15, −0.22, −0.32, −0.47, −0.7, −1.0

The difference in latent heat flux evaluated by the two schemes is shown in Fig. 6. When the soil water content is greater than or near field capacity (a value of 0.255 $m^3 m^{-3}$ has been used in this case), both MP84 and LP92 simulated potential evaporation and give similar soil evaporation. However, when the soil is dry or at least less than field capacity, the latent heat fluxes calculated from the two schemes are significantly different. In particular, MP84 yields a much lower evaporation rate leading to a considerable reduction in estimated daily evaporation. This is consistent with the results of Mahfouf and Noilhan (1991), who showed that threshold methods underestimate bare soil evaporation.

Figure 7 shows the observed heat fluxes and model results over sparse vegetation under dry conditions

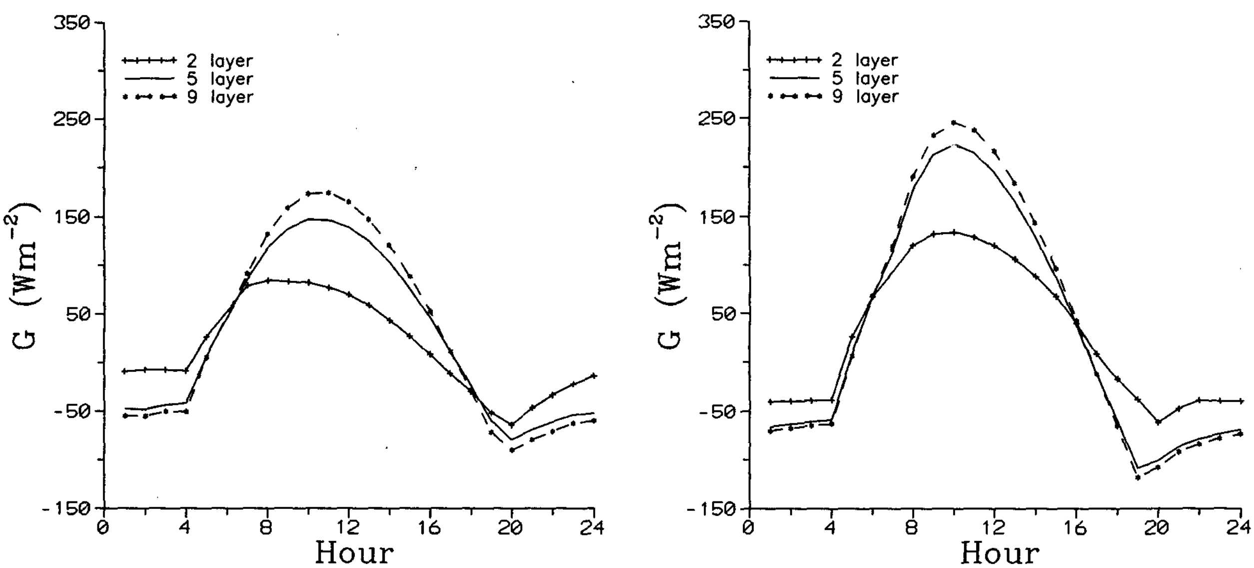

FIG. 4. Influence of the number of soil layers on the bare soil heat flux calculated by NOSU (a) initial soil water content W_{so} = 0.10 $m^3 m^{-3}$ and (b) W_{so} = 0.35 $m^3 m^{-3}$.

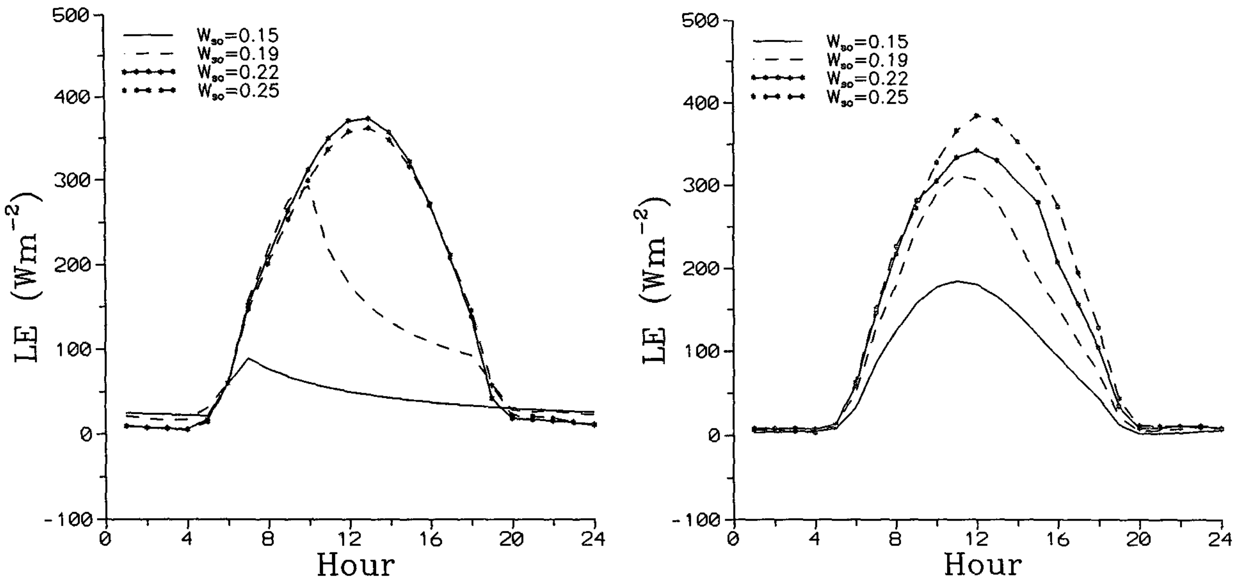

FIG. 5. Sensitivity of latent heat flux over a bare surface to the soil water content as estimated by (a) MP84 and (b) LP92, where W_{so} is the initial soil water content.

($W_{so1} < W_{fc}$). It shows that OSU1 strongly underestimates the latent heat flux, whereas the LE produced by NOSU is in agreement with the observation. In this case, as soil evaporation is more important ($F_c = 0.4$), the difference in latent heat fluxes is caused mainly by the two different soil evaporation schemes.

Compared with the observations, it illustrates that under both wet, $W_{so1} > W_{fc}$ (Fig. 1), and dry conditions (Fig. 7), LP92 simulates realistic soil evaporation, while MP84 produces realistic soil evaporation under wet soil conditions but underestimates the soil evaporation under the dry soil case. This is in agreement with the sensitivity test of Fig. 6.

Under dense vegetation, modeled and observed surface heat fluxes are shown in Fig. 8. NOSU simulated sensible and latent heat fluxes are in good agreement with the observations. The calculated soil heat flux is small under the dense vegetation cover and latent heat flux is dominant. The heat fluxes simulated by OSU-2L provide a reasonable approximation to the obser-

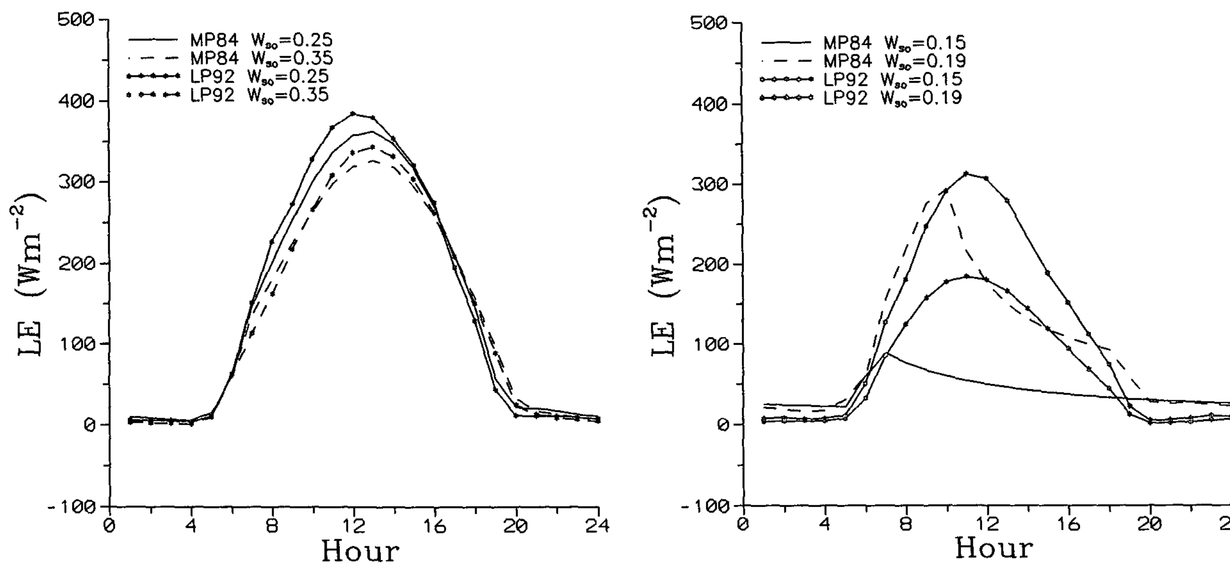

FIG. 6. Latent heat flux over bare soil as estimated by MP84 and LP92 for (a) wet soil conditions and (b) dry soil conditions, where W_{so} is the initial soil water content.

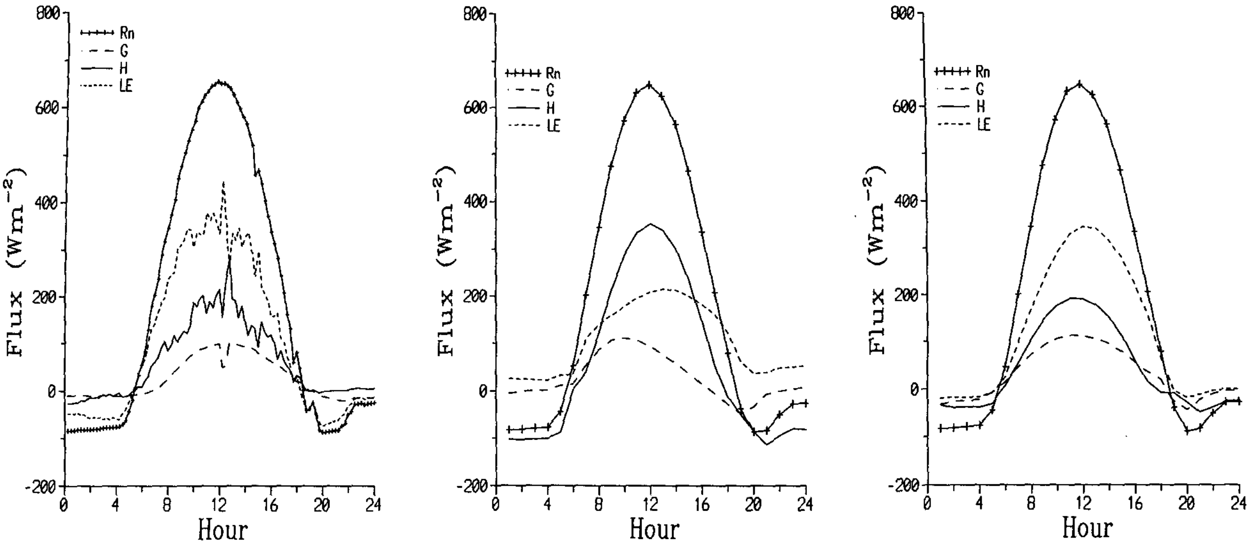

FIG. 7. Surface heat fluxes on 16 June at Castelnau during HAPEX-MOBILHY as (a) observed and simulated by (b) OSU1 and (c) NOSU.

vations. If the number of soil layers is increased from two to five (OSU-5L), it shows a strong overestimate of the ground heat flux (Fig. 9), whereas the five-soil-layer NOSU gives a better estimate of the ground heat flux and closer representation of the observations.

Figure 10 illustrates observed and modeled fluxes over mature oat under conditions of strong sensible heat fluxes (case 3). NOSU simulates a reasonable pattern of surface heat fluxes to those observed. It shows a dominate sensible heat flux and low ground heat flux. The lower latent heat flux simulated by NOSU could result from an $R_{s\min}$ of 600 s m^{-1} (used in this case) being too high as Noilhan and Planton (1989) used 450 s m^{-1} for mature oat and Pinty et al. (1989) used 200 s m^{-1}. OSU-5L strongly overestimates the ground heat flux and consequently simulates a lower sensible and latent heat flux than those observed.

These simulations suggest that relating ground heat flux for fully vegetated surfaces to net radiation should improve the surface heat flux simulation over dense vegetation.

Figure 11 shows the ground heat flux difference calculated between the OSU1 scheme represented by Eq. (1) (OG) and the modified scheme of Eq. (6) (NG)

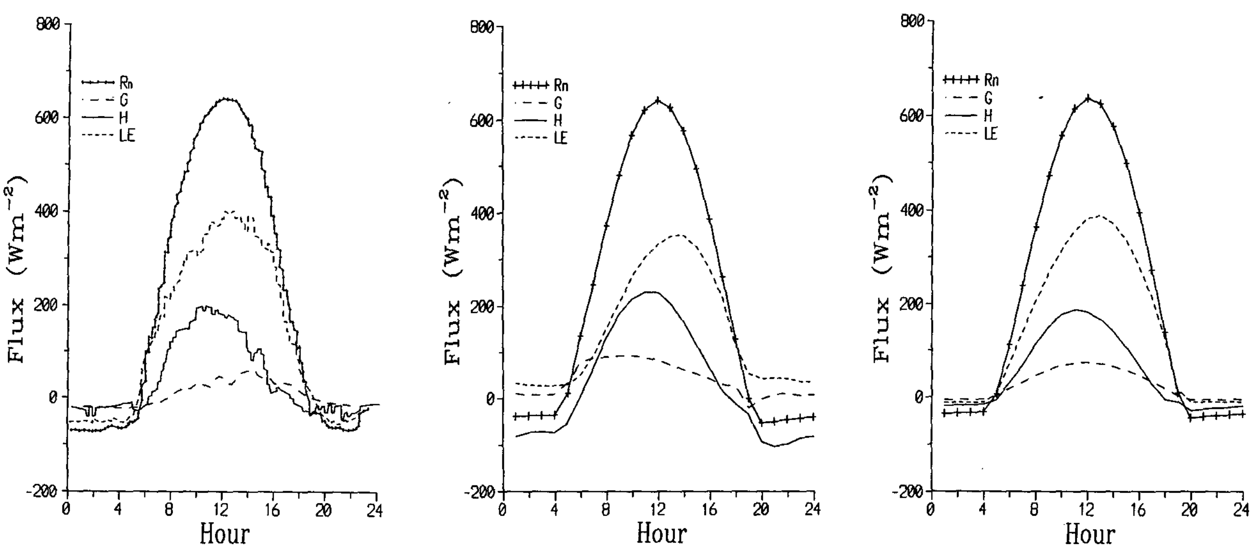

FIG. 8. Surface heat fluxes over maize at Lubbon 2 during the HAPEX-MOBILHY experiment on 16 June as (a) observed and simulated by (b) OSU1 and (c) NOSU.

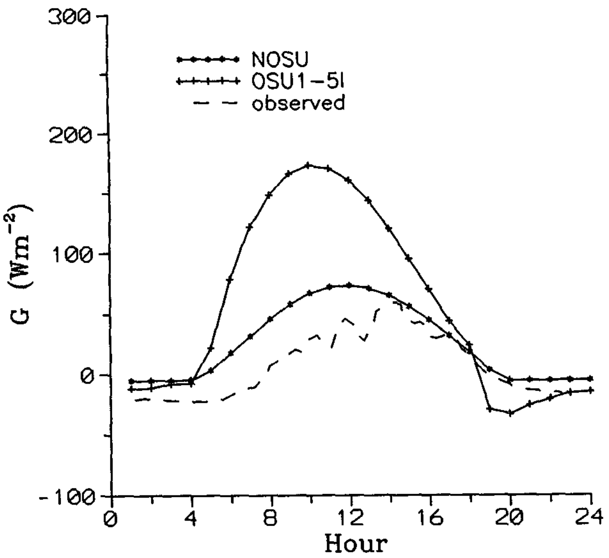

FIG. 9. Comparison between the ground heat fluxes modeled and observed during HAPEX-MOBILHY on 16 June over maize at Lubbon 2.

with five soil layers for dense vegetation. When $F_c = 0.2$, the two schemes estimate similar ground heat fluxes. However, as the vegetation becomes more dense (i.e., $F_c = 0.8$), there is a significant difference between the schemes.

As the ground heat flux for a fully vegetated surface is related to the net radiation, the surface energy balance for a fully vegetated surface can be written as

$$R_n(1 - C_c) = \rho C_p C_h(\theta_c - \theta_1) + \rho C_h L \frac{q_s(T_c) - q_1}{1 + C_h r_c}, \quad (32)$$

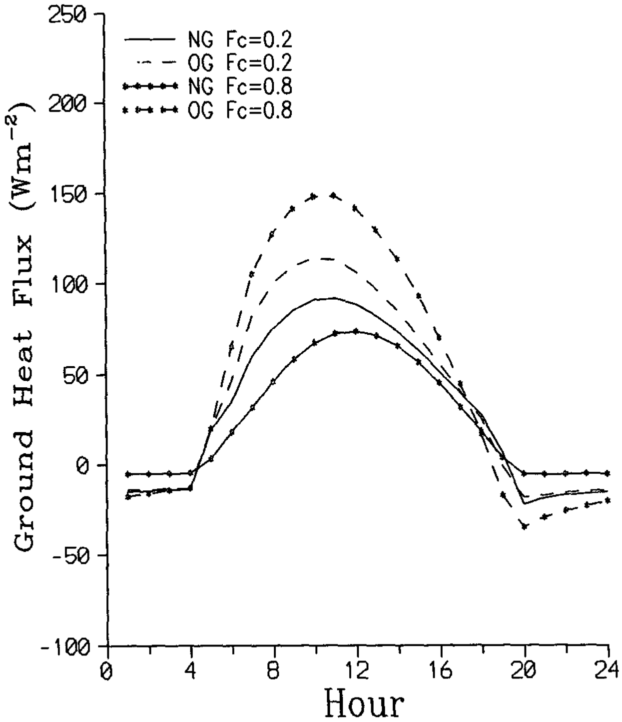

FIG. 11. Comparison between ground heat flux simulated by the soil temperature gradient (OG) and by the new method (NG) over partially vegetated surface.

with an approximation of

$$dq_s(T_c) \approx dq_s(T_1) + \frac{dq_s(T_1)}{dT}(T_c - T_1)$$

$$\theta_c - \theta_1 \approx T_c - T_1 + z_1\gamma_d, \quad (33)$$

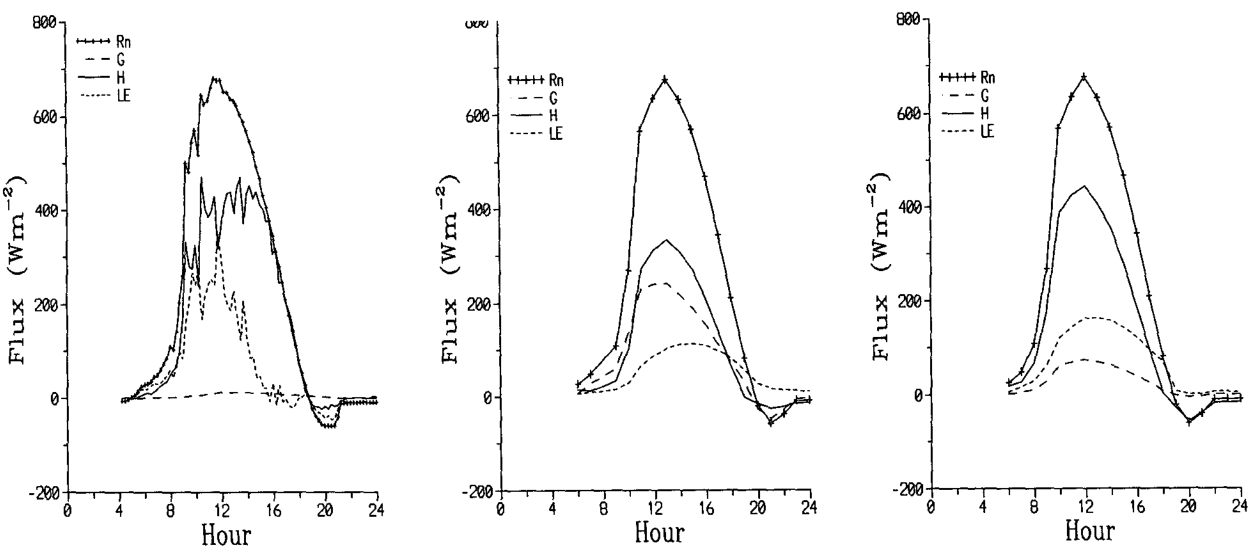

FIG. 10. Surface heat fluxes over oat at Lubbon 1 during HAPEX-MOBILHY on 28 June as (a) observed and simulated by (b) OSU1 and (c) NOSU.

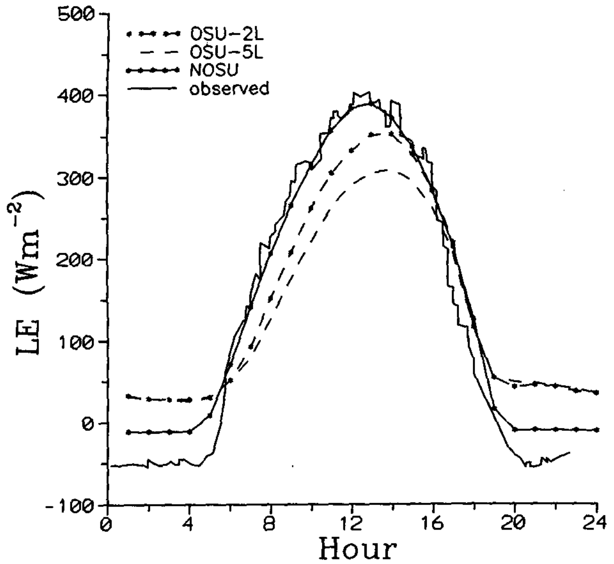

FIG. 12. Latent heat flux observed and simulated during HAPEX-MOBILHY over maize at Lubbon 2 on 16 June.

where z_1 is the first air level and γ_d the dry-adiabatic lapse rate.

Thus, the sensitivity of H and LE to C_c, the ratio of ground heat flux to net radiation, can be expressed as

$$\frac{dH}{dC_c} = \frac{dH}{dT_c}\frac{dT_c}{dC_c} = -HH\frac{R_n}{HH + EE}$$

$$\frac{dLE}{dC_c} = \frac{dLE}{dT_c}\frac{dT_c}{dC_c} = -EE\frac{R_n}{HH + EE}, \quad (34)$$

with

$$HH = \rho C_p C_h$$

$$EE = \frac{\rho C_h L}{1 + C_h r_c}\frac{dq_s(T_1)}{dT}. \quad (35)$$

Assuming typical values for the various parameters for the crop as r_c of 100 s m^{-1}, C_h of 0.025 m s^{-1}, T_1 of 300 K, R_n of 600 W m^{-2}, and changing C_c from 0.0 to 0.1 leads to a change in the sensible and latent heat fluxes of approximately 30 W m^{-2}. Thus, during the daytime, changes in the fluxes over the crop are not overly sensitive to the value of C_c.

The ratio of sensitivity of H and LE to C_c can be expressed as

$$\frac{dH}{dC_c}\left(\frac{dLE}{dC_c}\right)^{-1} = C_p(1 + r_c C_h)\left[L\frac{dq_s(T_1)}{dT}\right]^{-1}, \quad (36)$$

and this shows that the latent heat flux is more sensitive to changes in C_c than the sensible heat flux for short vegetation with small canopy resistance; whereas for higher canopy with large canopy resistance, the sensible heat flux is more sensitive to changes in C_c.

The latent heat flux calculated by OSU-2L, OSU-5L, and NOSU, as well as that observed over maize at Lubbon 2 (case 2) are shown in Fig. 12. OSU1 underestimated the latent heat flux during daytime but overestimated it during nighttime. Since the plant coefficient P_c was assumed constant, it does not account for the radiation effect on plant physiology and leads to an overestimate of transpiration during the night.

Using the relationship of P_c and r_c given by Eq. (17) means that transpiration is calculated in OSU1 as

$$E_{t1} = \frac{\rho C_h F_2}{1 + R_c C_h}[q_s(T_c) - q_1]. \quad (37)$$

If canopy resistance only accounts for the soil water content, Eq. (16) becomes

$$E_{t2} = \frac{\rho C_h}{1 + R_c F_2^{-1} C_h}[q_s(T_c) - q_1]. \quad (38)$$

Figure 13 illustrates the transpiration of the canopy calculated by Eqs. (37) and (38), with $F_c = 1.0$ and $R_c = 20$ s m^{-1}, using the initial conditions for case 2. The term E_{t1} strongly underestimates LE and the difference between these two schemes is given by

$$\frac{E_{t2}}{E_{t1}} = \frac{1 + C_h R_c}{F_2 + C_h R_c} \geqslant 1. \quad (39)$$

Given $R_c = 20$ s m^{-1}, $C_h = 0.045$ m s^{-1}, and $F_2 = 0.45$, the ratio E_{t2}/E_{t1} is about 1.4; that is, E_{t1} underestimates transpiration given by E_{t2} by about 40%. Thus, as the soil becomes drier, the ratio of E_{t2} to E_{t1} increases.

Figure 14 shows the observed and modeled heat fluxes over a forest (case 4). Although both model re-

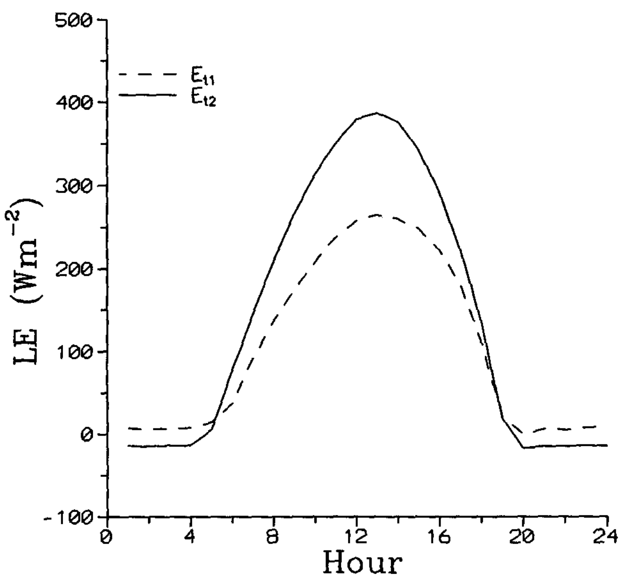

FIG. 13. Comparison between the latent heat flux computed as E_{t1} and E_{t2} for fully vegetated surface.

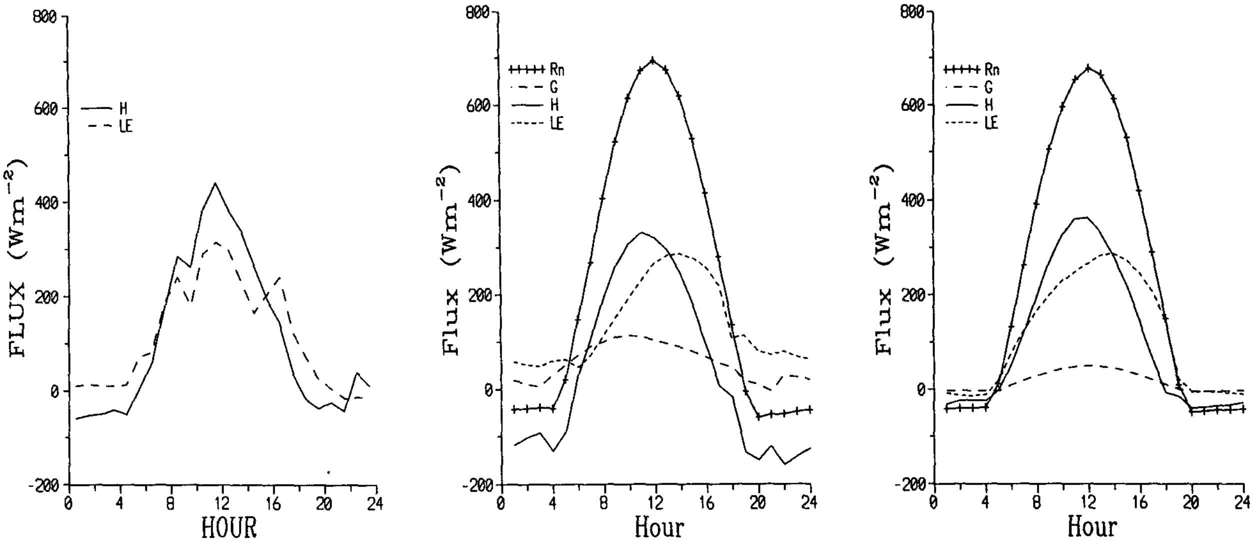

FIG. 14. Surface heat fluxes over forest at Estampon during the HAPEX-MOBILHY experiment on 16 June as (a) observed and simulated by (b) OSU1 and (c) NOSU.

sults are reasonable during the daytime, when compared with the observations, NOSU produces better sensible and latent heat fluxes than OSU1, especially in the morning and during the night.

The sensitivity of transpiration evaluated by Eq. (37) to the parameterization of canopy resistance (Table 4) is showed in Fig. 15. Without accounting for solar radiation, air vapor deficit, and temperature, r_{c2} leads to a strong overestimate of the latent heat flux and, consequently, a large underestimate of the sensible heat flux. Using the daily mean value of $F_1F_3^{-1}F_4^{-1}$ for canopy resistance, the latent heat flux computed by r_{c3} is lower than r_{c1} in the morning and higher in the afternoon.

4. Conclusions

The soil-canopy-atmosphere model (Ek and Mahrt 1989) has been modified to simulate surface energy components and compared with field observations on a daily basis. Ground heat flux over a bare soil surface is underestimated by the original two-layer soil model, while that simulated by the five-layer soil model provides a more realistic reproduction of the observations. Ground heat flux is significantly overestimated for a dense vegetation surface unless it is taken as a proportion of the net radiation. For partial vegetated land, the ground heat flux is based on a weighted fraction of the vegetation cover.

Using the formulation of canopy resistance proposed by Noilhan and Planton (1989) yields better results for the latent heat flux than the original formulation employed in OSU1 (Ek and Mahrt 1989). Failure to account for solar radiation, air vapor deficit, and air temperature in the canopy resistance leads to an overestimate of transpiration. Soil evaporation calculated by the Mahrt and Pan (1984) scheme is lower than

TABLE 4. Summary of parameterization of canopy resistance tests.

Test	Canopy resistance
r_c—test 1	$R_c = (R_{smin}/\mathrm{LAI})(F_1F_3^{-1}F_4^{-1})$
r_c—test 2	$R_c = R_{smin}/\mathrm{LAI}$
r_c—test 3	R_c = day time average of (0700–1700) $\times (R_{smin}/\mathrm{LAI})(F_1F_3^{-1}F_4^{-1})$

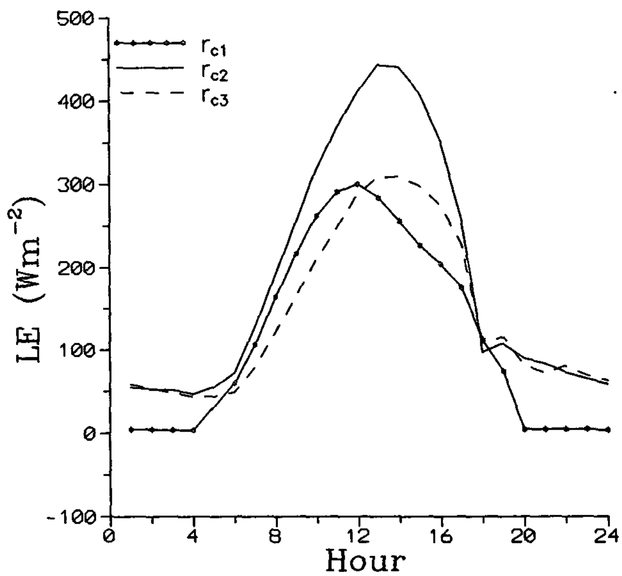

FIG. 15. Comparison of the diurnal variation in latent heat fluxes for the different parameterizations of canopy resistance.

that evaluated by the scheme of Lee and Pielke (1992) when the soil is dry. The soil evaporation evaluated by OSU1 is also sensitive to soil water content when it transfers from potential evaporation to soil water demand limited evaporation and when the soil is dry. The modified model addresses these shortcomings and is shown to reproduce field observations.

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APPENDIX

Soil Model

The soil model does not account for horizontal variations in the soil. Since it uses a short time period (diurnal change), water transfer is neglected at the lower soil boundary and soil temperature is assumed constant at this boundary. The soil model is based on Mahrt and Pan (1984) and Pan and Mahrt (1987).

Prognostic equations for soil temperature and soil water content are described as

$$C_{so}\frac{\partial T_{soi}}{\partial t} = \frac{\partial}{\partial_z}\left(K_T\frac{\partial T_{soi}}{\partial z}\right) \tag{A1}$$

$$\frac{\partial W_{so}(z)}{\partial t} = \frac{\partial}{\partial z}\left[D\frac{\partial W_{so}(z)}{\partial z}\right] + \frac{\partial K_w(z)}{\partial z} - \mathrm{RT}(z), \tag{A2}$$

where T_{soi} is the soil temperature, K_T the soil heat conductivity, C_{so} the volumetric heat capacity, D the coefficient of hydraulic diffusivity, and K_w the hydraulic conductivity. The term $\mathrm{RT}(z)$ is the volumetric soil water loss rate by transpiration at depth z and is a function of root distribution within the soil. Since it is difficult to determine this distribution over large areas with various canopy species, it is simply assumed that root distribution is uniform, so that RT is expressed as

$$\mathrm{RT}(z_i) = \frac{E_c}{\rho_w}\frac{g(z_i)}{\sum_{i=1}^{i=n} g(z_i)(z_i - z_{i-1})}, \tag{A3}$$

where ρ_w is the density of water.

The soil thermal conductivity K_T varies with soil water content and texture. Following McCumber and Pielke (1981), it is estimated as

$$K_T = \begin{cases} 418\exp[-(\log|\psi_w| + 2.7)], & \log|\psi_w| \leq 5.1 \\ 0.171, & \log|\psi_w| > 5.1 \end{cases}, \tag{A4}$$

where $\psi_w(m)$ is the soil moisture potential.

The simple relationships of K_w, D, and ψ_w to saturated volumetric soil moisture W_{sat}, soil water content, saturated hydraulic conductivity K_{wsat}, and saturated soil moisture potential ψ_{sat} are (Clapp and Hornberger 1978)

$$K_w = K_{wsat}\left(\frac{W_{so}}{W_{sat}}\right)^{2b+3} \tag{A5}$$

$$D = -\frac{bK_{wsat}\psi_{sat}}{W_{so}}\left(\frac{W_{so}}{W_{sat}}\right)^{b+3} \tag{A6}$$

$$\psi_w = \psi_{sat}\left(\frac{W_{sat}}{W_{so}}\right)^{b}, \tag{A7}$$

where values for b, W_{sat}, K_{wsat}, and ψ_{sat}, defined for the U.S. Department of Agriculture's 11 soil textural

TABLE A1. Soil parameters for the 11 USDA textural classes (following Clapp and Hornberger 1978).

Soil type	W_{sat} (m^3 m^{-3})	K_{wsat} (m s^{-1})	ψ_{sat} (m^3 m^{-3})	b	W_{fc} (m^3 m^{-3})	W_{wilt} (m^3 m^{-3})
Sand	0.395	1.760	−0.121	4.05	0.135	0.068
Loamy sand	0.410	1.563	−0.090	4.38	0.150	0.075
Sand loam	0.435	0.341	−0.218	4.90	0.195	0.114
Silt loam	0.485	0.072	−0.786	5.30	0.255	0.179
Loam	0.451	0.070	−0.478	5.39	0.240	0.155
Sandy clay loam	0.420	0.063	−0.299	7.12	0.255	0.175
Silt clay loam	0.477	0.017	−0.356	7.75	0.322	0.218
Clay loam	0.476	0.025	−0.630	8.52	0.325	0.250
Sand clay	0.426	0.022	−0.153	10.40	0.310	0.219
Slit clay	0.482	0.010	−0.490	10.40	0.370	0.283
Clay	0.482	0.013	−0.405	11.40	0.367	0.286

classes, were adopted following Clapp and Hornberger (1978) and are listed in Table A1.

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