A SENSITIVITY STUDY OF A COUPLED SOIL-VEGETATION BOUNDARY-LAYER SCHEME FOR USE IN ATMOSPHERIC MODELING

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Abstract. During the last two decades, several land surface schemes for use in climate, regional and/or mesoscale, hydrological and ecological models have been designed. For all of them, of great importance is a correct parameterization of the energy and momentum transport from the surface. Mainly, the land-surface schemes have been developed in order to parameterize land-surface variations both on daily and annual time scales. In this paper we describe the Land Air Parameterization Surface Scheme (LAPS) designed for use in atmospheric models and mainly for use in simulations of thermal circulations. It has two modules for parameterizing the processes at the boundary-layer interface: one for a bare surface and one for a vegetated one.

Incorrect parameterization of land-surface processes and prescription of the surface parameters in atmospheric modeling, can result in artificial changes of horizontal gradient of the sensible heat flux. Thus, an error in horizontal temperature gradient within the lower atmosphere may be introduced. Of course, the reliability of the model depends on the quality of boundary-layer scheme implemented and its sensitivity to the bare soil and vegetation parameters.

In this study, the sensitivity of the LAPS scheme to parameterization of bare soil processes, ground roughness length, vegetation parameters (stomatal resistance, leaf area index and canopy height) and albedo, has been examined. Tests based on time integrations using real data were performed. For sensitivity tests with the non-plant module the data sets for June 3, 1981; June 4, 1982; June 11, 1982 and June 24, 1982 from the experimental site at Rimski Šančevi (Yugoslavia) were used. The performance of the vegetation module was tested by employing the data set for September 8, 1988 measured at the experimental site of De Sinderhoeve (The Netherlands). The computed latent heat flux and the ground temperature outputs were compared with the observations. Finally, the sensitivity of the LAPS scheme to the prescribed parameters was examined by using a simple statistical analysis.

1. Introduction

As shown in a number of previous publications (Mahrer and Pielke, 1977; Ross and Orlanski, 1982; Rosenthal, 1978; Plantin, 1985; McCumber, 1980; Garrett, 1982; Mahfouf et al., 1987; Segal et al., 1989b) during the last two decades, an accurate description of a large variety of atmospheric phenomena such as sea breezes, frontal perturbations, tropical cyclones and squall lines, could be achieved by using limited area atmospheric models. Such models were also used in order to investigate the role of mesoscale circulations in the dispersion of air pollutants, released from different type of sources (McNider, 1981; Kallos, 1989; Kallos, 1990; Moran, 1992; Kallos et al., 1993; Pielke and Uliasz, 1993). Sensitivity studies by McCumber and Pielke (1981) and Zhang and Anthes (1982) point to the extremely important role of the surface fluxes in determining the diurnal variations within the lower troposphere. However, this fact is often not of great assistance to numerical modelers because of the complexity of the land surface-atmosphere interface in all these processes. Namely, the fluxes of momentum and heat strongly depend on the surface characteristics such as albedo, roughness length and soil moisture content. Sometimes, they vary temporally and spatially within a broad range of values. All this information is incorporated into atmospheric models with the aid of various parameterization schemes which, in general, are crude and sometimes unrealistic.

Pielke et al. (1991), documented that landscape variability (heterogeneous terrain, albedo and vegetation state) on scales smaller than those which can be resolved by general circulation models (GCMs) or operational numerical weather prediction models (NWPMs) (i.e., scales from $\approx$4 to several hundred kilometers) can result in mesoscale heat and moisture fluxes which are larger and have a different vertical structure than those due to turbulent boundary-layer fluxes over the same area. Later, using satellite observations and data provided by the United State Geological Survey (USGS), in conjunction with the Regional Atmospheric Modeling System (RAMS) for real case studies, Pielke et al. (1993) demonstrated the influence of land surface variability on heat and moisture fluxes over GCM grid areas.

A number of sensitivity tests related to the influence of surface boundary-layer conditions on mesoscale processes have been considered in the last decade. Mostly, these numerical experiments should be classified into two groups: In the first are the sensitivity tests of the land-air parameterization scheme by itself. These are performed, usually, by using available field micrometeorological data sets. Thus, Wilson et al. (1987) 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 incorporation of some soil parameters and color. For this 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. In a very detailed study, Sellers and Dorman (1987) investigated the sensitivity of a simple biosphere scheme using micrometeorological field data. The work of Mihailović et al. (1992) has investigated the sensitivity of nine different soil textures with fixed vegetation and atmospheric conditions using the BATS scheme. Recently, Dekić et al. (1995) studied the sensitivity of bare-soil evaporation schemes included in the land surface LAPS scheme to soil surface wetness. The second group of sensitivity studies concern the influence of the land-air interface conditions on mesoscale circulations. There are a considerable number of papers dealing with this topic, starting from the pioneering work by McCumber (1980). For example, in the study of Garrett (1982), it was indicated that the vegetative cover, the soil moisture, and roughness length affect the location of convective cells via the development of the PBL. Avissar and Pielke (1989) examined the influence of the horizontal variability of land cover on the various regional and mesoscale circulations. Mahfouf et al. (1987) studied the influence of soil and vegetation on the development of mesoscale circulations considering the atmospheric response to soil and vegetation inhomogeneities. The influence of vegetation inhomogeneities on mesoscale circulation is considered by Segal et al. (1988), Segal et al. (1989b) and Pinty et al. (1989). Finally, Segal et al. (1989b) evaluated the mesoscale circulations forced by surface gradients of heating arising from irrigated areas adjacent to dry land utilizing a combination of satellite, observational and modeling approaches. As was shown, thermal circulations similar to the sea-breezes should develop between adjacent irrigated-non irrigated areas. They called them Non-Classical Mesoscale Circulations (NCMC).

Incorrect parameterization of land-surface processes and prescription of the surface parameters can result in the wrong partitioning of the surface energy into latent and sensible heat fluxes. It could result in artificial increasing or decreasing of the horizontal gradient of the sensible heat flux and in the Planetary Boundary Layer (PBL) growth (Segal et al., 1989a). Consequently, a change of horizontal temperature gradient within the lower atmosphere may be introduced. Further, it will determine changes in mesoscale circulations simulated by an atmospheric model. In order to avoid these uncertainties in mesoscale and other scale modeling there is a need for more information about the sensitivity of parameterizations, of bare and partly plant-covered surfaces, to the different state of the surface.

This study was made in order to improve our knowledge in this crucial field. For this purpose, the coupled soil-vegetation scheme LAPS was used in order to examine its sensitivity to parameterization of some key processes and variables which must be modeled with more attention than others. In our opinion these are: (i) evaporation from bare soil and vegetated surface and (ii) ground temperature. These are considered as very important in modeling atmospheric processes of various scales and especially the thermally-driven ones. Also, we examined sensitivity to some parameters representing the state of surface: ground roughness length, stomatal resistance, leaf area index, canopy height and canopy albedo (Pielke et al., 1993).

2. Description of Surface Scheme used in Sensitivity Tests

2.1. Bare-soil part

The part of the surface scheme used in the numerical tests with bare soil is based on the model for predicting the soil moisture in three layers, as described by Mihailović (1991). The equation for ground temperature, T_g, is (Deardorff, 1978)

$$C_g \frac{\partial T_g}{\partial t} = R_{ng} - H_g - LE_g - \left(\frac{\lambda C \omega}{2}\right)^{1/2} (T_g - T_d), \tag{1}$$

where R_{ng} is the absorbed net radiation, H_g is the sensible heat flux, LE_g is the latent heat flux, L is the latent heat of vaporization, λ is the thermal conductivity,

C is the volumetric heat capacity, $\omega = 2\pi/\tau$, τ is the day length and T_d is the temperature for deep soil. In Equation (1), C_g represents the bulk heat capacity per unit area which is computed following Zhang and Anthes (1982). The volumetric heat capacity of the soil depending on the volumetric soil moisture content is considered according to de Vries (1963) taking into account the volumetric soil moisture content in the top soil layer, ϑ_1. The thermal diffusivity K_t is parameterized following the approximate formula for a loam soil by de Vries (1963)

$$K_t = \frac{\gamma_i\,(2.9 + 0.04\vartheta_1)K_0}{[(1 - 0.65\vartheta_1)\vartheta_1 + 0.09](0.23 + \vartheta_1)}, \tag{2}$$

where γ_i is the ratio of saturated thermal conductivity for a given soil texture to that of saturated loam and K_0 = 10^{-7} m^2 s^{-1}. The value of γ_i for different soil textures can be found in Table III of Wilson et al. (1987). The thermal conductivity was calculated using the expression $\lambda = K_t C$.

The net radiation R_{ng} at the soil surface accounts for the contributions of solar radiation, R^S, and thermal radiation from the atmosphere, R_L, absorbed by the ground. It also accounts for the component of solar radiation $\alpha_g R^S$ reflected from the ground where α_g is its albedo. The radiation outgoing from the ground is calculated following the Stephan-Boltzmann law, also taking into account the emissivity of the ground. The albedo is considered by taking into account its variability with volumetric soil moisture content according to Idso et al. (1975)

$$\alpha_g = \begin{cases} 0.31 - 0.34\vartheta_1/\vartheta_s & \vartheta_1/\vartheta_s \le 0.5 \\ 0.14 & \vartheta_1/\vartheta_s > 0.5 \end{cases}, \tag{3}$$

where ϑ_s is the volumetric soil moisture content at saturation for the top soil layer. For thermal radiation from the atmosphere we used an expression proposed by Staley and Jurica (1972)

$$R_L = [n + (1 - n)0.67(1670q_r)^{0.08}]T_r^4, \tag{4}$$

where n is the cloud fraction and q_r and T_r are the specific humidity and temperature of the air at the reference level z_r.

In Equation (1), H_g and LE_g are defined by

$$H_g = \rho c_p \frac{T_g - T_r}{r_A}, \tag{5}$$

and

$$LE_g = \frac{\rho c_p}{\gamma}\frac{1}{r_A}[\alpha e_*(T_g) - e_r], \tag{6}$$

where ρ is the density of air, c_p the specific heat of air at constant pressure, r_A the aerodynamic resistance between ground surface and the reference level, $e_*(T_g)$

is the is the saturation vapor pressure at ground temperature T_g, e_r is the vapor pressure of the air at the reference level, γ is the psychrometric constant, and α is considered as a function of the volumetric soil moisture content of the top soil layer, ϑ_1, and field capacity, ϑ_{fc} (Mihailović et al., 1993)

$$\alpha = \begin{cases} 1 - [(\vartheta_{fc} - \vartheta_1)/\vartheta_{fc}]^2 & \vartheta_1 \le \vartheta_{fc} \\ 1 & \vartheta_1 > \vartheta_{fc} \end{cases} . \tag{7}$$

The aerodynamic resistance r_A in Equations (5) and (6) under neutral condition was calculated according to

$$r_A = \frac{1}{k^2 u_r} \ln^2 \left[\frac{z_r}{z_g} \right] , \tag{8}$$

where k is von Karman's constant taken to be 0.41, z_g is the ground roughness length and u_r is the wind speed at the reference level. The effect of atmospheric stability on the aerodynamic resistance is determined following Mihailović (1991).

The equation for deep soil temperature T_d is

$$C_g \frac{\partial T_d}{\partial t} = 2(R_{ng} - LE_g - H_g)/\sqrt{365\pi}. \tag{9}$$

The soil moisture content was parameterized using three equations for three soil moisture storages

$$\frac{\partial \vartheta_1}{\partial t} = \frac{1}{D_1} \left[P_1 - Q_{12} - \frac{1}{\rho_w} E_g \right] , \tag{10}$$

$$\frac{\partial \vartheta_1}{\partial t} = \frac{1}{D_2} [Q_{12} - Q_{23}], \tag{11}$$

$$\frac{\partial \vartheta_3}{\partial t} = \frac{1}{D_3} [Q_{23} - Q_3], \tag{12}$$

where P_1 is the infiltration of precipitation into the top soil layer, ϑ_1 is the volumetric soil moisture content in ith layer, ρ_w is the density of water, E_g is the evaporation rate, D_i is the thickness of the ith soil layer, $Q_{i,i+1}$ is the water flux between i and $i+1$ soil layers, and Q_3 is the gravitational drainage from a recharge soil moisture store.

The governing equations for the three soil moisture contents for bare soil are given by Equations (10)–(12) while in the presence of vegetation cover they have the form described by Equations (35)–(37). The terms E_g and E_{tf} are defined by Equations (6) and (38) while other terms in the governing equations will be described below.

The precipitation P_1 that infiltrates into the top soil layer is given by

$$P_1 = \begin{cases} \min(P_0, K_s) & \vartheta_1 < \vartheta_s \\ 0 & \vartheta_1 = \vartheta_s \end{cases}, \tag{13}$$

where K_s is the saturated hydraulic conductivity, ϑ_s is the volumetric soil moisture content at saturation ϑ_1 and P_0, the effective precipitation rate on the soil surface, is given by

$$P_0 = P - (P_f - D_f). \tag{14}$$

where P is the precipitation rate above the canopy. The rate of interception (inflow) for the canopy, P_f, is given by

$$P_f = P(1 - e^{-\eta})\sigma_f, \tag{15}$$

where η is a constant depending on the leaf area index and σ_f the fractional cover of the ground by the vegetation. It is assumed that the interception of the rainfall can be considered via this expression describing the exponential attenuation (Sellers et al., 1986). The rate of drainage of water stored on the vegetation (outflow) for canopy, D_f, is given by

$$D_f = \begin{cases} 0 & w_f < w_{\text{max}} \\ P_f & w_f = w_{\text{max}} \end{cases}, \tag{16}$$

where w_f is the canopy interception store described by Equation (33) and w_{max} is the maximum amount of water held by the canopy which is parameterized according to Dickinson (1984).

The transfer of water between adjacent layers $Q_{i,i+1}$ is given by

$$Q_{i,i+1} = K_{ef}[2(\Psi_i - \Psi_{i+1})/(D_i + D_{i+1}) + 1], \tag{17}$$

where Ψ_i is the soil moisture potential of the ith layer that is parameterized, as it is usually done, after Clapp and Hornberger (1978),

$$\Psi_i = \Psi_s(\vartheta_i/\vartheta_s)^{-B}, \tag{18}$$

where Ψ_s is the soil water potential at saturation and B is a soil type constant. In Equation (17) K_{ef} is the effective hydraulic conductivity between soil layers given by

$$K_{ef} = (D_i K_i + D_{i+1} K_{i+1})/(D_i + D_{i+1}), \tag{19}$$

where K_i is the hydraulic conductivity of the ith soil layer determined by the empirical formula

$$K_i = K_{si}(\vartheta_i/\vartheta_s)^{2B+3}, \tag{20}$$

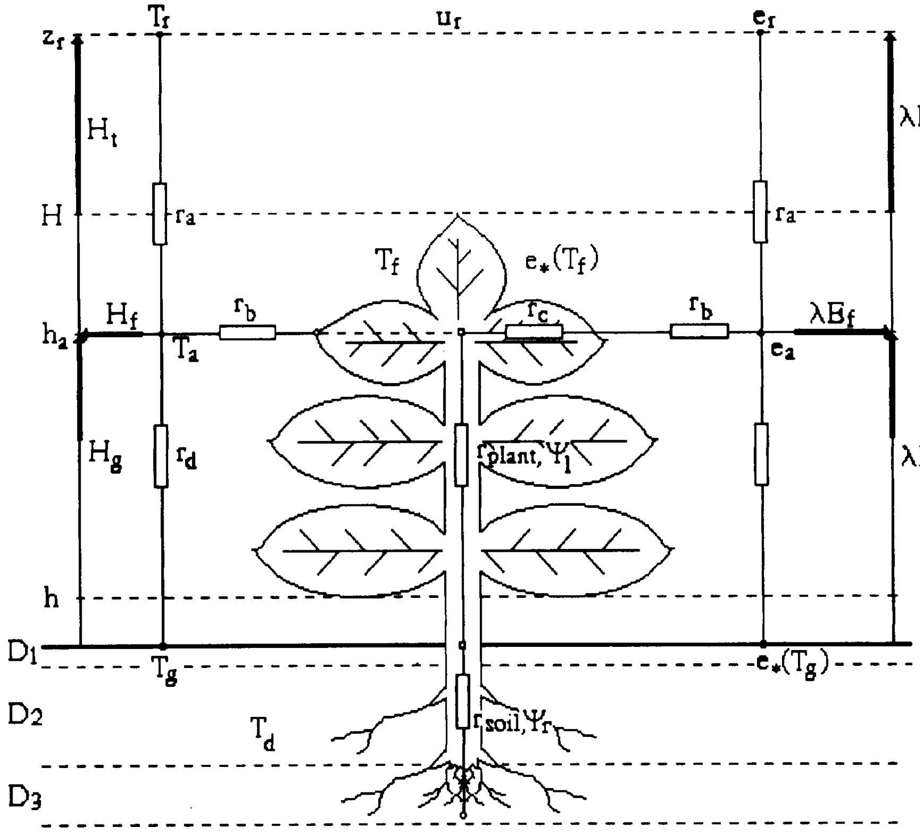

Figure 1. Schematic diagram of transfer pathways for latent and sensible heat fluxes.

while K_{si} is the hydraulic conductivity at saturation of the ith soil layer.

The gravitational drainage from the bottom soil layer is defined as

$$Q_3 = K_{si}(\vartheta_3/\vartheta_s)^{2B+3} \sin x, \tag{21}$$

where x is the mean slope angle (Sellers et al., 1986).

2.2. Vegetation Part

The vegetation part of the scheme is based on a single layer approach commonly used by numerical modelers (Henderson-Sellers, 1993). The vegetation is represented as a block of constant density porous material sandwiched between two constant stress layers, the height of the canopy top is H, and the height of the canopy bottom is h (Figure 1).

The prognostic equation for canopy temperature T_f, is,

$$C_f \frac{\partial T_f}{\partial t} = R_{nf} - H_f - LE_f - G, \tag{22}$$

where C_f is the heat capacity of the canopy, R_{nf} the net radiation at the vegetated surface, H_f the sensible heat flux, E_f the evaporation rate from the vegetated surface and G the soil heat flux which is parameterized using the "force-restore" method.

The net radiation absorbed by canopy, R_{nf}, is calculated as a sum of short and long wave radiative flux. The short wave radiation absorbed by canopy, R_f^s, is:

$$R_f^s = R^s(\sigma_f - \alpha_f)[1 + (1 - \sigma_f)\alpha_g], \tag{23}$$

where R^s is the incident downward directed short wave flux, assumed to be known, as the forcing variable; α_f is the foliage albedo. There is no distinction between direct and diffuse radiation and it is assumed that albedo does not vary with zenith angle. Both short and long wave radiation are reflected once between the soil surface and canopy.

The radiative flux absorbed by the canopy, R_f^1 is

$$\begin{aligned} R_f^1 &= R^s\sigma_f\varepsilon_f - 2\sigma_f\varepsilon_f\sigma_B T_f^4 + \sigma_f\varepsilon_f[R_L(1 - \sigma_f)(1 - \varepsilon_g) \\ &\quad + \sigma_f\varepsilon_f(1 - \varepsilon_g)\sigma_B T_f^4 + \varepsilon_g\sigma_B T_g^4], \end{aligned} \tag{24}$$

where ε_g and ε_f are the emissivities of the ground and the canopy, respectively, and R_L is the incident downward radiation parameterized by Equation (4).

The fluxes H_f and LE_f are parameterized as

$$H_f = \frac{2(T_f - T_a)}{r_b}\rho c_p, \tag{25}$$

and

$$LE_f = [e_*(T_f) - e_a]\frac{\rho c_p}{\gamma}\left[\frac{W_f}{r_b} + \frac{1 - W_f}{r_b + r_c}\right], \tag{26}$$

where T_a and e_a are temperature and vapor pressure in the canopy air space, $e_*(T_f)$ is saturation vapor pressure at temperature T_f, $W_f = w_f/w_{\max}$, the wetness fraction of canopy, r_b is the bulk boundary-layer resistance and r_c is the bulk stomatal resistance. The bulk boundary-layer resistance, r_b, is calculated as

$$r_b = P_S C_t \beta(\sin\beta)^{1/4}/(L_d H u_H^{1/2})\int_{\alpha_w\beta}^{\beta}(\sin y)^{1/4}\,\mathrm{d}y, \tag{27}$$

where P_S is the leaf shelter factor, C_t the transfer coefficient, β the extinction factor, L_d the stem and leaf area density related to leaf area index, LAI, as LAI $= L_d(H - h)$, u_H the wind speed at the canopy top, $\alpha_w = h/H$ and $y = \beta z/H$. The extinction factor, β, depends on the plant morphology and is defined as

$$\beta = \left(\frac{C_d \mathrm{LAI} H}{2\sigma_s}\right)^{1/2}, \tag{28}$$

where C_d is the leaf drag coefficient. The constant σ_s is defined following Goudriaan (1977)

$$\sigma_s = i_w \left(\frac{4w_d}{\pi L_d}\right)^{1/2}, \tag{29}$$

where i_w is the relative turbulence intensity and w_d is the width of the square leaves. The typical values for parameters used in Equations (27)–(29) can be found in Goudriaan (1977).

The prognostic equation for the ground temperature T_g has the form of Equation (1). However, the corresponding sensible and latent heat fluxes H_g and LE_g now take the following forms:

$$H_g = \frac{(T_g - T_a)}{r_d}\rho c_p, \tag{30}$$

and

$$LE_g = \frac{\rho c_p}{\gamma}\frac{1}{r_A}[\alpha e_*(T_g) - e_r], \tag{31}$$

where $e_*(T_g)$ is the saturation vapor pressure at temperature T_g and r_d is the resistance to water vapor and heat flow from the soil surface to air space within the canopy which is parameterized as

$$r_d = \frac{1}{k^2 u_H}\left[\frac{\sinh(\beta)}{\sinh(\alpha_w \beta)}\right]^{1/2} \ln^2\left(\frac{h}{z_g}\right). \tag{32}$$

It is worth noting that the Equations (30) and (31) will become (5) and (6), respectively in the case of bare soil.

The governing equation for the canopy interception water store is

$$\frac{\partial w_f}{\partial t} = P_f - D_f - \frac{E_{wf}}{\rho_w}, \tag{33}$$

where the rate of evaporation from the wetted part of the vegetation, E_{wf}, is

$$LE_{wf} = \frac{[e_*(T_f) - e_a]}{r_b}\frac{\rho c_p}{\gamma}. \tag{34}$$

The governing equations for the soil wetness in the three soil layers are

$$\frac{\partial \vartheta_1}{\partial t} = \frac{1}{D_1}\left[P_1 - Q_{12} - \frac{1}{\rho_w}(E_g + E_{tf,1})\right], \tag{35}$$

$$\frac{\partial \vartheta_2}{\partial t} = \frac{1}{D_2}\left[Q_{12} - Q_{23} - \frac{1}{\rho_w} E_{tf,2}\right], \tag{36}$$

$$\frac{\partial \vartheta_3}{\partial t} = \frac{1}{D_3}[Q_{23} - Q_3], \tag{37}$$

where P_1 is the infiltration of precipitation into the upper soil moisture; $E_{tf,1}$ and $E_{tf,2}$ are the canopy extraction of moisture by transpiration from the first and second layer, respectively. The transpiration part E_{tf}, is calculated by using the equation

$$LE_{tf} = \frac{[e_*(T_f) - e_a]}{[r_b + r_c]}\frac{\rho c_p}{\gamma}(1 - W_f). \tag{38}$$

The diagnostic variables, the temperature T_a and vapor pressure e_a within the canopy air space are determined from the energy balance equations:

$$H_t = H_f + H_g = \frac{(T_a - T_r)}{r_a}\rho c_p, \tag{39}$$

$$LE_t = LE_f + LE_g = \frac{(e_a - e_r)}{r_a}\frac{\rho c_p}{\gamma}, \tag{40}$$

where LE_t and H_t are the latent and sensible heat flux respectively, directed to the atmosphere, and r_a the aerodynamic resistance representing the transfer of heat and moisture from the canopy to the reference level, z_r, which is calculated as

$$r_a = \frac{1}{ku_*}\ln\frac{z_r - d}{H - d}, \tag{41}$$

where u_* is the friction velocity, z_0 the roughness length and d the zero plane displacement.

The zero plane displacement height, d, and roughness length, z_0, we calculated according to Goudriaan (1977)

$$d = H - \frac{1}{k}\left[\frac{\sigma_s H}{\beta}\right]^{1/2}, \tag{42}$$

and

$$z_0 = (H - d)\exp\left\{-\frac{H}{\beta(H - d)}\right\}. \tag{43}$$

In the LAPS scheme, stomatal resistance, r_s, depends upon both atmospheric factors and water stress. This dependence is given in the form of

$$r_s = r_{s\min}\phi_1\phi_2^{-1}\phi_3^{-1}\phi_4^{-1}. \tag{44}$$

The minimum stomatal resistance $r_{s\min}$ is defined as the value of stomatal resistance observed at high solar flux, with well irrigated soil, saturated air and optimal air temperature. The factor ϕ_1 gives the dependence on the solar radiation. It is parameterized following Dickinson (1984) who suggested the form

$$\phi_1 = \frac{1+f}{f + \frac{r_{s\min}}{r_{s\max}}}, \tag{45}$$

where f is calculated according to Dickinson et al. (1986)

$$f = \frac{1.1R^s}{R_{gl}\mathrm{LAI}}, \tag{46}$$

and R_{gl} is the limiting value of 30 W m^{-2} for a forest and 100 W m^{-2} for crop. For $r_{s\max}$ the value of 5000 s m^{-1} was used.

The factor ϕ_2 takes into account the effect of water stress on the stomatal resistance and is parameterized in the following way

$$\phi_2 = \begin{cases} 1 & \vartheta_a > \vartheta_{fc} \\ 1 - \left[\dfrac{\vartheta_{\mathrm{wilt}}}{\vartheta_a}\right]^{1.5} & \vartheta_{\mathrm{wilt}} \le \vartheta_a \le \vartheta_{fc} \\ 0 & \vartheta_a < \vartheta_{\mathrm{wilt}} \end{cases}, \tag{47}$$

where ϑ_a is the mean volumetric soil moisture content of the first and the second soil layer and $\vartheta_{\mathrm{wilt}}$ is the wilting point.

The factor ϕ_3 describes the dependence of the stomatal resistance on the air temperature. According to Dickinson et al. (1986) this factor can be written in the form

$$\phi_3 = 1 - 0.0016(298 - T_r)^2, \tag{48}$$

where T_r is the air temperature at the reference level. Following the same authors, the factor ϕ_4, which represents the effect of atmospheric vapor pressure deficit, could be parameterized as

$$\phi_4 = 1 - 0.025[e_*(T_f) - e_r], \tag{49}$$

where $e_*(T_f)$ and e_r are the saturation vapor pressures for the canopy temperature T_f, and for the temperature at the reference level. The bulk stomatal resistance, r_c, represents the effective stomatal resistance per unit ground surface area and it is given by

$$r_c = r_s/\mathrm{LAI}. \tag{50}$$

More details about the parameterization for the whole vegetation module are provided in Mihailović and Jeftić (1994).

3. Sensitivity of the Bare Soil Module

Regarding bare soil, the main processes parameterized in land surface schemes are evaporation and heat and momentum exchange between the atmosphere and the surface. The exchange of energy, mass and momentum between the atmosphere and the surface is controlled by atmospheric conditions (including wind speed, temperature and moisture), and the surfaces conditions (including temperature and soil moisture in the top soil layer and aerodynamic ground roughness length). In order to examine the sensitivity of the bare soil module of the LAPS scheme to the selection of the evaporation scheme and some parameters characterizing the state of bare soil surfaces, some tests were performed. For this purpose, the latent heat flux outputs produced by the bare soil module were compared with one point micrometeorological measurements obtained over a bare soil experimental site Rimski Šančevi (Yugoslavia).

3.1. Description of data sets, boundary and initial conditions used in the tests with the bare soil module

The experimental site was located in the northeastern part of Yugoslavia at the meteorological station of Rimski Šančevi (45.33°N, 19.5°E), altitude 84 m, on the chernozem soil of the loess terrace of southern Bačka. Description of its structure and its distribution was given by Živković et al., (1972). Some of its hydraulic properties: soil pore space, and hydraulic conductivity at saturation were determined by Vučić (1964). They are listed in Table I. In this study, we used the four data sets which are part of a larger measurement campaign which examined the exchange processes of heat, mass, and momentum above bare soil, winter wheat, and soya bean planted surfaces during the growing season in the period 1981–85. They refer to June 3, 1982 (BS03), June 4, 1982 (BS04), June 11, 1982 (BS11) and June 24, 1982 (BS24). The sensible and latent heat fluxes for these situations were calculated by the Bowen ratio method using gradient measurements within the surface boundary layer.

Except for the set of ground thermometers, the rain gauge, and the measuring instruments in the thermometer screen, all sensors were fixed to a minitower. The wind speed was measured by cup anemometers. The wet and dry bulb temperature gradients, for determining the Bowen ratio, were measured using platinium resistance thermometers. The thermometers and anemometers were set at 0.2, 0.5, 0.8, 1.1, 1.4, 1.7, and 2 m above the ground. Two solarimeters were used in order to measure the incoming and reflected solar radiation (Kipp and Zonen CM5). The soil temperature was measured at 0.02-, 0.05-, 0.1-, 0.2-, 0.3-, 0.5-, and 1-m depths. Volumetric soil moisture contents were measured at 10-cm intervals up

Table I
Hydraulic constants of the chernozem soil of the loess terrace of Southern Bačka (density, $\rho_s = 1290$ kg m^{-3}, roughness length, z_g = 0.01 m)

Hydraulic properties		
Saturated moisture potential	Ψ_s	-0.036 m
Saturated hydraulic conductivity	K_s	3.2×10^{-5} m s^{-1}
Clapp-Hornberger's constant	B	6.5
Field capacity	ϑ_{fc}	0.36 m^3 m^{-3}
Volumetric soil moisture content at its saturation	ϑ_{si}	0.52 m^3 m^{-3}
Photometric properties		
Emissivity	ε_g	0.97

to 1-m depth by gypsum blocks manufactured in the Department of Meteorology (Mihailović, 1983). The soil heat flux was estimated from gradient measurements of soil temperature using Ceytin's method (Ceytin, 1953; Vereshnin et al., 1959) which is given in more detail in Mihailović et al. (1995).

The Bowen ratio was derived from the air temperature and vapor pressure measured at the seven heights mentioned above. Their average values were derived from the measured profiles following the methodology of Monteith (1973).

In some cases, mainly near sunset, we have obtained incorrect results by using the Bowen ratio method. A reason for the poor results during this period could be that as sunset approaches and the surface layer makes the transition from the unstable to the stable regime, the gradients should be very small. This can introduce large errors in calculating the fluxes when using the Bowen ratio method. In this case, the calculated fluxes were not considered. The 66 hourly values of the latent and sensible heat fluxes from the four chosen cases were available for comparison with the simulated ones. In the numerical tests, we used the chernozem soil properties listed in Table I.

In all datasets, the atmospheric boundary conditions at the reference height $z_r =$ 2 m were derived from measurements of global and reflected radiation, cloudiness, precipitation, wet-bulb and dry-bulb temperatures, and average wind speed over 1-h intervals. Then, the measured values were interpolated to the beginning of each time step, which was 600 s in this study. Distribution of the soil layers was D_1 = 0–0.1 m, D_2 = 0.1–0.5 m and D_3 = 0.5–1.5 m. The initial conditions for the volumetric soil moisture contents ϑ_1, ϑ_2, ϑ_3 and the ground temperature T_g for all datasets are given in Table II. The initial condition for deep soil temperature T_d was assumed to be equal to 291.45 K for all cases. The initial condition for the atmospheric pressure was always the same -101.6 kPa.

Equation (1) was solved using an implicit backward-differencing scheme, while an explicit scheme was applied for solving Equations (10)–(12) and Equation (9).

Table II

A list of initial conditions for the 4 cases used in numerical tests. The variables ϑ_1, ϑ_2, and ϑ_3 are the volumetric soil moisture contents in the three model layers from the top to the bottom, respectively, and T_g is the ground temperature. The stars refer to next day

Dates	ϑ_1	ϑ_2	ϑ_3	T_g	Time interval	Number of observations
June, 1981						
3	0.15	0.22	0.23	295.25	0500–0400*	18
June, 1982						
4	0.16	0.23	0.26	291.75	0500–0400*	17
11	0.12	0.18	0.20	292.25	0500–0400*	11
24	0.18	0.20	0.21	292.85	0500–0400*	20

3.2. Sensitivity to Partitioning the Surface Energy into Sensible and Latent Heat Portions

As is well known, the bare soil evaporation E_g (or the bare soil latent heat flux, LE_g), is driven by the humidity difference between the atmosphere and the ground. Also, the amount of this flux is additionally modified by the turbulence in the atmosphere as well as the rate at which water can be diffused towards the soil-atmosphere interface. The actual evaporation from unsaturated soils differs from the potential one. Its correct simulation in land air parameterization is impressive, starting from the well known Philip's (1957) approach to the recently designed schemes which have been reviewed by Mahfouf and Noilhan (1991) and Mihailović et al. (1993). Basically, all available methods can be classified into the three groups: the α, the β and the threshold method.

The latent heat flux LE_g from a bare soil can be generally written in the form

$$LE_g = [e_*(T_g) - e_r]\frac{\rho c_p}{\gamma}\frac{h_s}{r}, \tag{51}$$

where r is equal r_A or $r_s + r_A$ depending upon whether the α or β method is applied, r_s is the surface resistance, and h_s is a factor that adjusts for the relative humidity of the air at the soil surface.

In numerical models for the adjustment factor h_s, two methods have been commonly used in order to estimate the soil surface humidity of water vapor pressure which is

$$h_s = \begin{cases} [\alpha e_*(T_g) - e_r][e_*(T_g) - e_r]^{-1}, & \alpha \text{ formulation} \\ \beta, & \beta \text{ formulation} \end{cases}, \tag{52}$$

where α and β are the functions of soil wetness.

In designing the schemes, the commonly used approach is the expression proposed by Philip (1957). From thermodynamic considerations, he derived an expression for α in the form

$$\alpha = \exp\left(\frac{\psi_g g}{R_w T_g}\right), \tag{53}$$

where Ψ_g is the soil water potential at the surface, g is the acceleration of gravity, and R_w the gas constant for water vapor. This scheme produces somewhat unrealistic results, especially when the upper soil layer is dry. This fact is considered in more detail by Kondo et al. (1990). They showed that the α factor depends strongly on the ratio of the specific humidity of air and the specific humidity of the saturated air at the ground temperature for small values of the volumetric soil moisture content, otherwise α changes rapidly from 0 to 1. This rapid change of α was also noted by Mihailović et al. (1993). Moreover, they found that for different soil texture α is still close to 1, even when the volumetric soil water content has dropped below the permanent wilting point (0.17 m^3 m^{-3} for chernozem soil). For example, for T = 293 K and Ψ_g = -160 m, α takes values of 0.989.

Some researchers tried to overcome this deficiency of Philip's formula, resulting in an entire class of new formulae in either the α or β form. Their detailed description can be found in the recent comprehensive overviews (Mahfouf and Noilhan, 1991; Lee and Pielke, 1992; Mihailović et al., 1993). All of these formulae have been designed either from experimental data (Kondo et al., 1990; Barton, 1979) or in a rather *ad hoc* manner (Deardorff, 1978; Mihailović et al., 1993); assuming some relationship between β and ϑ/ϑ_{fc} for the volumetric soil moisture content ϑ below the field capacity ϑ_{fc}. Figure 2 shows the α and β factors for a chernozem soil computed for the different formulations listed in Table III.

All the above mentioned methods have the common problem of correctly partitioning the surface energy into sensible and latent heat fluxes. According to Avissar and Pielke (1989) and Segal et al. (1989a), the parameterization of the Bowen ratio is of prime importance in achieving accurate simulations of thermal circulations with atmospheric models. Thus, an inaccurate parameterization of the latent heat flux can seriously disturb this ratio, i.e., the energy partitioning between the sensible and the latent heat at the surface and, consequently, the accuracy of the model simulation. In that sense, Figure 3 should be used as a clear illustration of differences in the simulation of bare soil evaporation due to its different formulations. This figure shows results of time integrations with the bare soil (BS24 data set) which have been performed using different "α" and "β" schemes listed in Table III. The BS24 data set was chosen over other bare soil sets since it included the largest number (19) of observed hourly values of the latent heat flux and the ground temperature. The experiment consisted of running the LAPS bare soil module with the different schemes included in Equation (51). The left panels show the diurnal variations of the computed latent heat flux. The panels on the right side present the

Table III

Collection of various formulae for α and β. Variable ϑ is the volumetric soil water content, ϑ_{fc} is the volumetric soil water content at field capacity, g is the acceleration due to gravity, Ψ_g is the soil water potential at the surface, R_w is the gas constant for water vapor and T_g is the ground temperature

Investigator	Abbrev.	
Barton (1979)	B	$\beta = \begin{cases} 1.8\vartheta/(\vartheta + 0.3), & \vartheta \le 0.375 \\ 1, & \vartheta > 0.375 \end{cases}$
Deardorff (1978)	D	$\beta = \min(1, \vartheta/\vartheta_{fc})$
Lee and Pielke (1992)	L	$\beta = \begin{cases} \frac{1}{4}\left[1 - \cos\left(\frac{\vartheta\pi}{\vartheta_{fc}}\right)\right]^2, & \vartheta \le \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$
Philip (1957)	P	$\alpha = \exp\left(\dfrac{g\psi_g}{R_w T_g}\right)$
Jacquemin and Noilhan (1990)	J	$\alpha = \begin{cases} \frac{1}{2}\left[1 - \cos\left(\frac{\vartheta\pi}{\vartheta_{fc}}\right)\right], & \vartheta \le \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$
Mihailović et al. (1993)	M	$\alpha = \begin{cases} 1 - \left(1 - \frac{\vartheta}{\vartheta_{fc}}\right)^2, & \vartheta \le \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$

observed latent heat values plotted against the computed ones using BS03, BS04, BS11 and BS24 data sets which included 66 hourly values.

The left panels of Figure 3 show very large differences in the simulation of diurnal variations of latent heat flux regardless of the method used. Generally, neither the "α" nor the "β" schemes reproduce the diurnal variations of the latent heat flux in a proper way. Moreover, both methods simulate the daily course in a similar manner. They either considerably overestimate or underestimate the observed diurnal course of the latent heat flux over a bare soil. For mesoscale models, it is very important that a land surface scheme correctly parameterizes the energy and momentum transport from the surface on daily and hourly time scales. Beyond the fact that all selected schemes do not exactly follow this condition, most of them are already included in many current atmospheric models (Henderson-Sellers et al., 1993). Among them there is one exception: the M (defined in Equation (7)) and D schemes simulate the daily cycle better than the other approaches. The improvement in the calculated latent heat flux from these two schemes, compared to other ones, is emphasized by the right panels of the Figure 3 where the computed 66 hourly values, are plotted against the observations. In this figure, all the above data

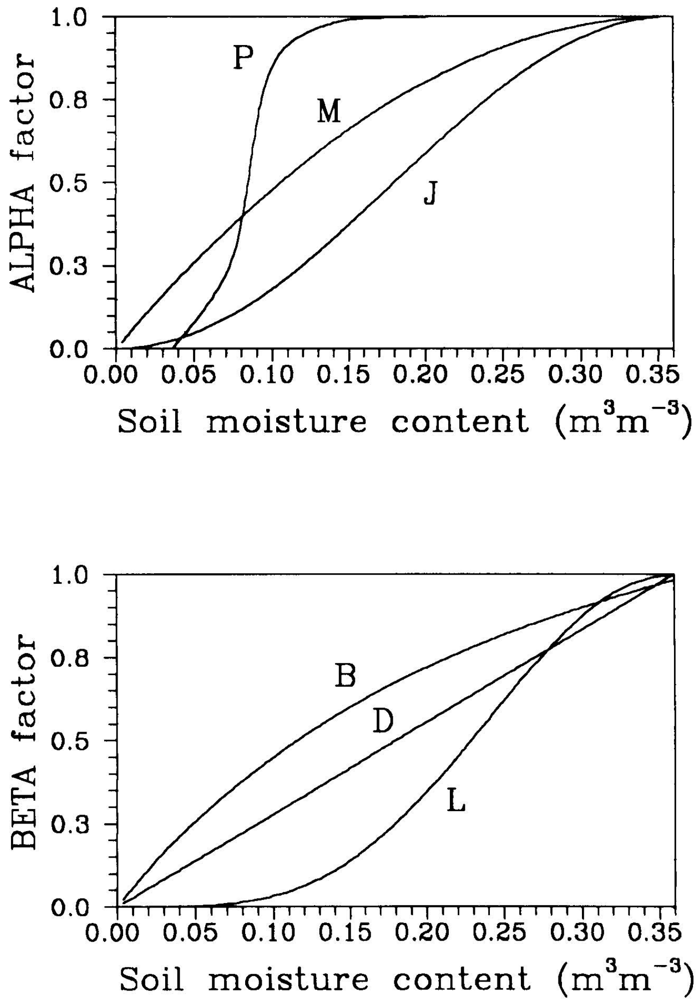

Figure 2. The "α" and "β" factors for a chernozem soil for different α (above) and β (below) formulations.

sets were used. The concentrations of D and M letters is more pronounced around the diagonal while the other letters are scattered towards the larger and smaller computed values of the latent heat flux. This example illustrates that the various bare soil evaporation schemes yield quite varying results. Certainly, it will seriously affect the quality of the model results. The evaporation scheme incorporated in the land surface scheme suggested here, could be considered as a satisfactory choice but it is still far from the desirable quality.

In atmospheric modeling, an accurate prediction of the ground temperature for bare soil surfaces is of great importance. The ground temperature might differ depending on type and state of the surface (Segal et al., 1989a; Dalu et al., 1991).

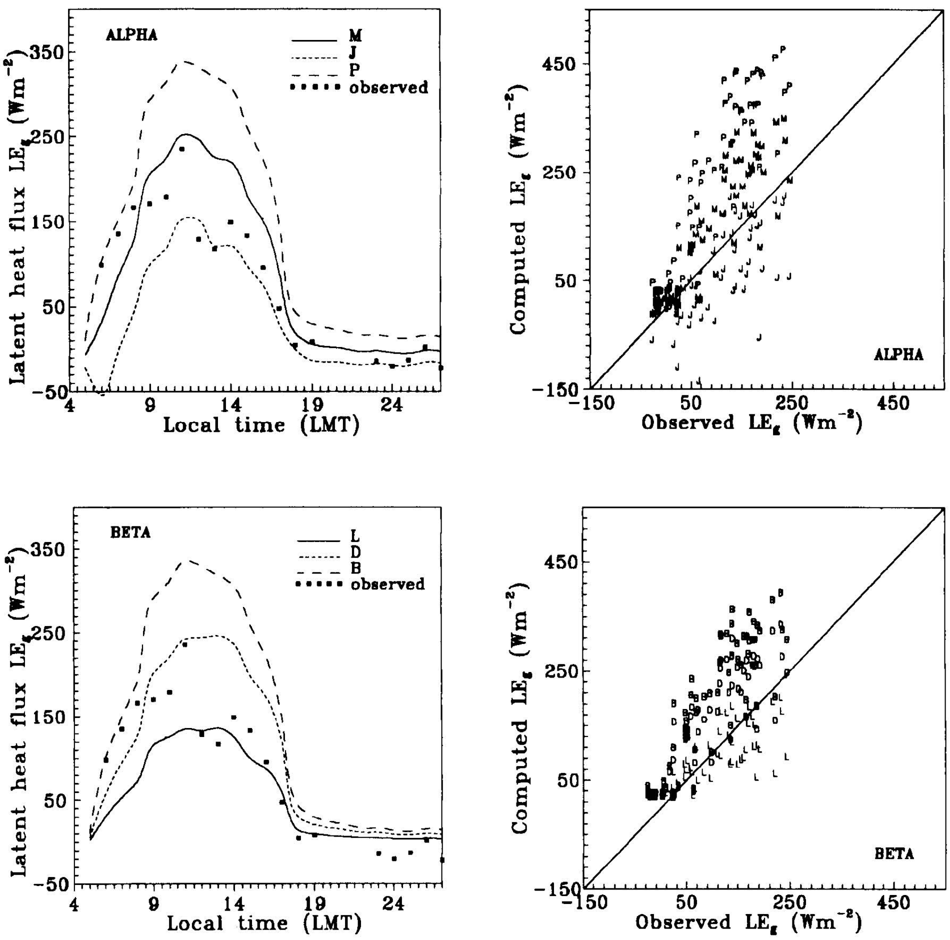

Figure 3. Diurnal variations of latent heat fluxes observed and computed by the LAPS scheme using different the "α" and "β" approaches over bare soil at Rimski Šančevi (Yugoslavia) for June 24, 1982 (left panels). Computed values plotted against observations using BS03, BS04, BS11 and BS24 data sets (right panels). The capitals indicating the designer of the scheme are listed in Table III.

Its calculation is strongly governed by the surface fluxes. Thus, an inaccurate parameterization of the latent heat flux can seriously disturb the Bowen ratio, i.e., the partitioning of energy between the latent and the sensible heat at the surface, and, consequently, accurate calculation of the ground temperature. A good illustration of such effects is provided in Mihailović and Rajković (1994). These differences are evident in the left and the right panels of Figure 4. On the left, the diurnal variations of the ground temperatures, observed and computed by the LAPS scheme are displayed using both the "α" and "β" schemes applied to the BS24 data set. On the right, the computed values plotted against observations, using the BS03, BS04, BS11 and BS24 data sets, are also shown. The diurnal course of the ground temperatures obtained by the P and B schemes considerably underestimate the observations. This is something expected because they give higher amounts of evaporation compared to reality. Thus, an intensive cooling of the ground is evident. On the right panels these letters are mostly concentrated below the diagonal. However, the J and L schemes give results which are in a good agreement with the observations around the noon although during the rest of the day they overestimate them. The right panels show that these schemes tend to overestimate the measured values of ground temperature. There is a physical reason for that: because they predict lower evaporation, more heat is retained by the ground, resulting in its higher temperatures. The D and M schemes used in the LAPS predict the ground temperature more accurately than the other ones.

To quantify the differences between the considered schemes in partitioning the surface energy, we have computed some statistical quantities. More specifically, we have computed the root-mean-square (RMSE), the mean absolute (MAE) and mean error (AVER), and the standard deviation (SDEV). The results of this analysis are shown in Table IV. The statistics in this table support the above conclusions, as it was expected. The largest deviations from the observed latent heat fluxes are exhibited by the P and B schemes. These two schemes exhibit the tendency to overestimate the observations, although the B scheme has a tendency towards values that are lower than in the case when the P scheme is applied. Generally, the L scheme has the lowest deviation from the observations. All the other schemes with their statistical quantities are placed in between the P and L ones. Looking at the latent heat flux statistics of the D and M schemes, we can see that they are very close to each other. The J scheme underestimates the observed values like the L one. However, a distinction between them can be made. Namely, the values of the latent heat fluxes computed by the L schemes are more evenly distributed around the diagonal than in the case of application of the scheme (right panels of Figure 3). Moreover, the J scheme has a more emphasized tendency toward the lower values.

A further inspection of Table IV shows that the largest error in ground temperature is made when the B scheme is employed. It seems that this scheme, among the other considered here, is partitioning the energy by introducing the largest error in computing the ground temperature. The P and L schemes introduce smaller errors which are similar in magnitude but have a quite different partitioning of energy since the L overestimates and the P underestimates the observed values of the ground temperature. The J scheme has a smaller error than the foregoing ones. The best results in predicting the ground temperature are achieved by the D and M schemes. These schemes have similar statistical performance and therefore minimize the error in partitioning the surface energy between the latent and sensible heat. Consequently, they result in a more accurate estimation of the ground temperature.

Table IV

Error analysis of the predicted latent heat flux and ground temperature obtained by the bare soil module of the LAPS scheme, using different "α" and "β" approaches. RMSE is the root mean square error, MAE is the mean absolute error, AVER is the mean error and SDEV is the standard deviation. r and α_r refer to correlation coefficient and slope angle of the regression line, respectively. Analysis was performed using BS03, BS04, BS11 and BS24 data sets with 66 hourly values.

Scheme	RMSE	MAE	AVER	SDEV	r	α_r
Latent heat flux (W m^{-2})						
B	96.07	77.90	−74.95	61.57	0.899	53.4
D	64.95	53.95	−44.95	46.87	0.893	48.0
L	53.82	37.64	21.82	49.57	0.814	29.0
J	82.48	53.38	54.21	62.61	0.713	35.2
M	67.06	51.57	−37.14	56.26	0.893	51.3
P	148.10	120.25	−119.09	88.72	0.906	60.0
Ground temperature (°C)						
B	4.69	2.49	−2.85	3.47	0.905	38.2
D	2.74	1.83	−0.24	2.75	0.941	41.7
L	4.16	3.14	−2.78	3.13	0.940	46.4
J	3.55	3.81	3.90	2.92	0.941	44.9
M	2.77	1.89	−0.50	2.74	0.943	40.0
P	4.18	3.29	2.68	3.23	0.937	35.0

3.3. Sensitivity to Ground Roughness Length and Drag Coefficient

The ground roughness length and the corresponding drag coefficient play a very important role in controlling the exchange of momentum between the atmosphere and the ground. Rougher surfaces cause more intense turbulence, which increases the drag and transfer rates across the surface. The proper parameterization in different scale atmospheric numerical models is of great importance but it is not always obvious how to do it correctly. For example, in several models, especially in large-scale numerical weather forecast ones, the lowest level (at some height above the surface, z_1) is so high that the surface layer is not resolved. In order to avoid such problems in 3-D models, Andre and Blondin (1986) suggested the use of effective roughness length $z_{0\text{eff}}$ which dependents on z_1. However, Taylor (1987) suggested that $z_{o\text{eff}}$ is independent of z_1. The sensitivity of mesoscale models to roughness length was studied by Garratt and Pielke (1989). They used the Colorado State University Mesoscale Model (CSU-MM) in order to evaluate the sensitivity of 1 and 2-D models to differences in the aerodynamic roughness length and surface temperature roughness length. The ratio of these parameters quantifies the excess resistance to the transfer of heat and water relative to momentum. In their model, allowance is made for this excess resistance. In the LAPS-like schemes it is not

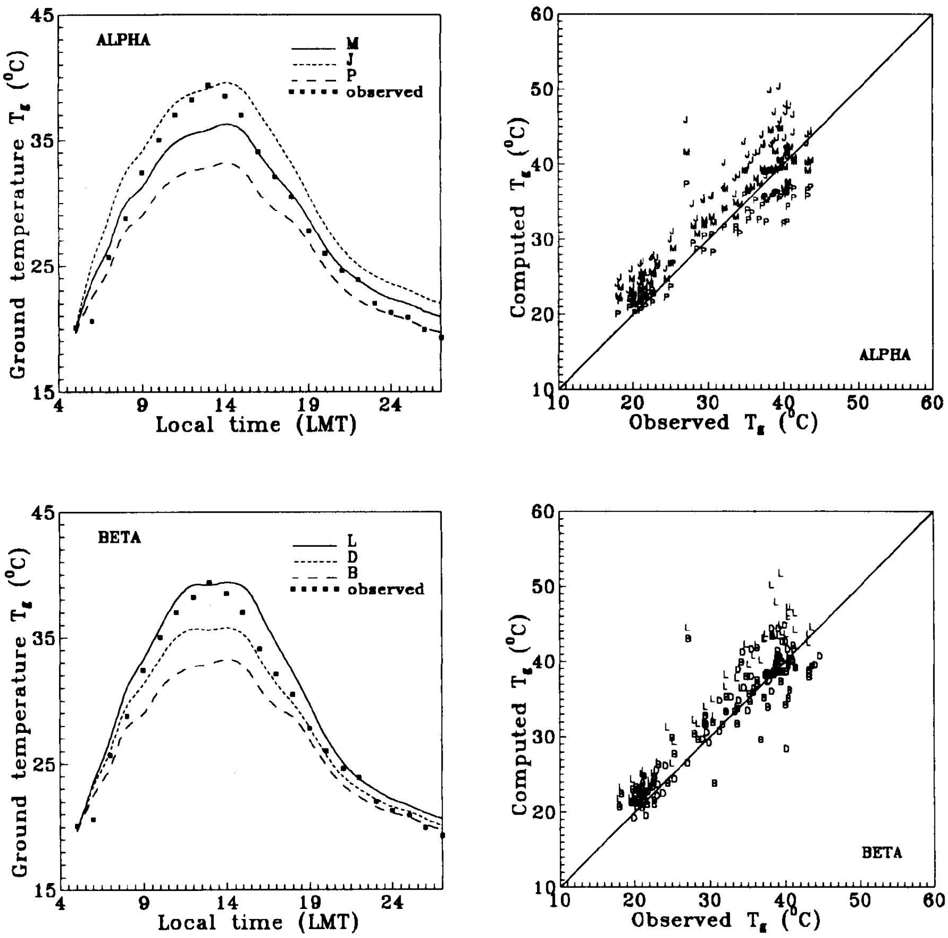

Figure 4. Diurnal variations of ground temperatures observed and computed by LAPS scheme using different the "α" and "β" approaches over bare soil at Rimski Šančevi, (Yugoslavia) for June 24, 1982 (left panels). Computed values plotted against observations using BS03, BS04, BS11 and BS24 data sets (right panels). The capitals indicating the designer of the scheme are listed in Table III.

(e.g., Bush et al., 1976; Zhang and Anthes, 1982). The same is true for several other models using the scheme of Louis (1979).

In order to establish the sensitivity of the bare soil module of the LAPS scheme to the ground roughness length, a set of experiments was performed, using the BS24 data set. The sensitivity tests consisted of running the above scheme for different z_g (0.001 m, 0.003 m, 0.005 m, 0.01 m, 0.02 m, 0.03 m, 0.04 m, 0.05 m, 0.06 m, and 0.1 m). Once the set of bare soil data and atmospheric conditions were externally prescribed, we then changed only the ground roughness length. In order to quantify the differences caused by variations of z_g, we have computed the RMSE. The results from these experiments are presented in Figure 5. In this figure

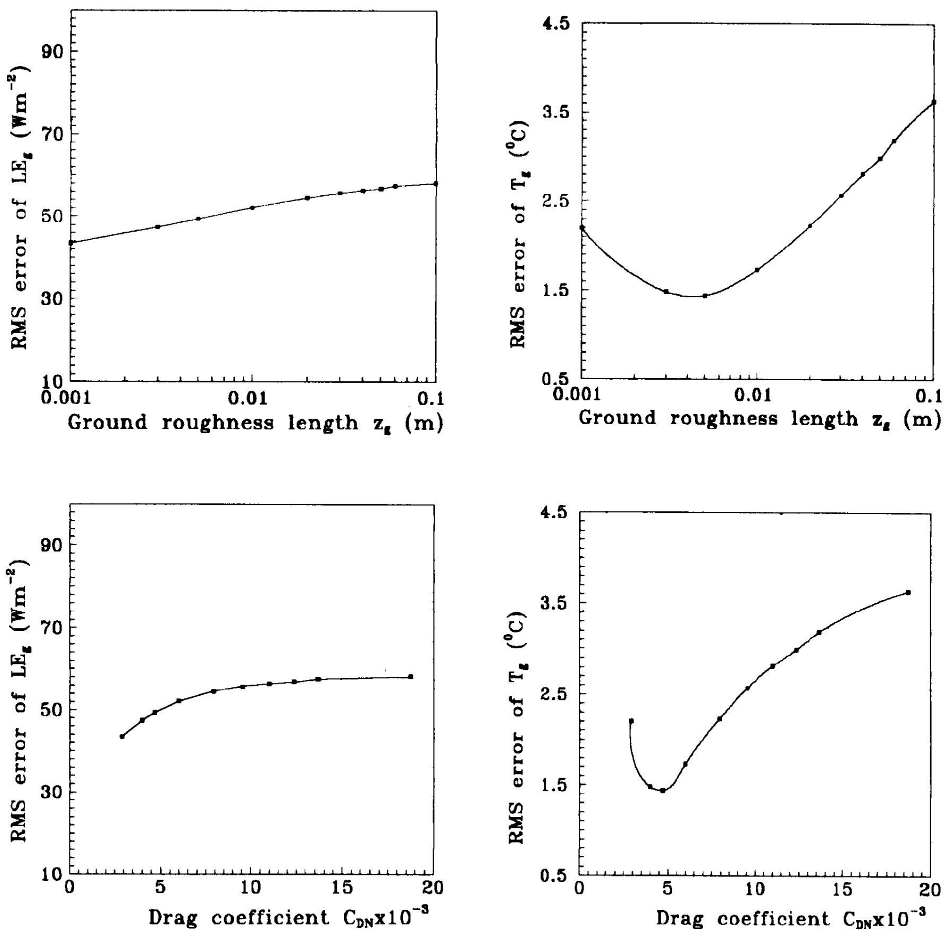

Figure 5. Root-mean-square errors of latent heat flux and temperature as functions of: ground roughness length, z_g, (upper panels) and drag coefficient under neutral conditions, C_{DN}, (lower panels).

the RMSE of the latent heat flux and temperature are shown as functions of the ground roughness length and drag coefficient under neutral conditions. Looking at the left panels it can be seen that increasing z_g (0.001–0.1 m) and C_{DN} ($3–19 \times 10^{-3}$) introduces errors in the range (43–58 W m^{-2}) or 35% of the daily sum of the latent heat flux. The error is also introduced in the computations of the ground temperature T_g. These differences vary from 2.2 °C to 3.6 °C but they are passing through a minimum of 1.4 °C when z_g is around 0.005 m. It is worth noting that this value of the ground roughness length is commonly used for the parameterization of processes over bare soil.

Table V

Soil and plant properties of the experimental site in De Sinderhoeve (The Netherlands) and parameters of maize for September 8, 1988

	Symbol	Value
Soil		
Density	ρ_s	1410 kg m^{-3}
Ground roughness length	z_g	0.01 m
Saturated soil moisture potential	ψ_s	-0.0175 m
Saturated hydraulic conductivity	K_s	0.95×10^{-3}
Clapp-Hornberger's constant	B	4.38
Field capacity	ϑ_{fc}	0.15 m^3 m^{-3}
Volumetric soil moisture content at saturation	ϑ_s	0.41 m^3 m^{-3}
Wilting point	$\vartheta_{\rm wilt}$	0.075 m^3 m^{-3}
Emissivity	ε_g	0.97
Plant		
Maize height	H	2.30 m
Roughness length	z_0	0.114 m
Displacement height	d	1.86 m
Leaf area index	LAI	4.0
Fractional vegetation cover	σ_f	0.80
Albedo	α_f	0.2

4. Sensitivity to Some Vegetation Parameters and Albedo

There are several studies concerning the sensitivity of land surface schemes to various vegetation parameters (e.g., Saxton, 1975; Sellers and Lockwood, 1981; Dorman and Sellers, 1989). The crucial point in the vegetation part of each land surface scheme is mainly related to how the stomatal resistance is parameterized; this is because it strongly governs the process of evapotranspiration. This is the reason why we considered stomatal resistance, leaf area index, canopy height and albedo to test the sensitivity of the LAPS scheme to variations in each one.

4.1. Description of the Data Set – Boundary and Initial Conditions Used in the Tests with the Vegetation Model

For these tests we used a data set which is a part of a larger measurement program which examined the exchange processes of heat, mass, and momentum just above and within a maize canopy during its growing season in De Sinderhoeve (The Netherlands). The observed values of the surface fluxes were available from the eddy flux measurements reported by van Pul (1992).

The experimental site was in the center of the Netherlands (51.59°N, 5.45°E). It was an area of 250 m × 250 m, surrounded by other agricultural fields in which maize was grown. Their soil properties are shown in Table V (Jacobs et al., 1990). The maize was planted in the 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 m^2). The selected case is September 8, 1988 which will be denoted as PL89. This data set was chosen because it is considered as a representative case corresponding to the period of the growing season when the maize plants are tall. As a result, the values of the leaf area index, LAI, and fractional cover, σ_f, increased significantly. For this single day, the LAI was very high (4.0). This parameter, together with the maize height, roughness length, and zero plane displacement height were measured (van Pul, 1992; Jacobs et al., 1990). The minimum stomatal resistance was not measured and was assumed to be $r_{s\min}$ = 200 s m^{-1}. Some of the plant parameters used in the sensitivity tests are also listed in Table V while other calculations related to resistance can be found in Mihailović and Ruml (1996). During this day, wind at the reference level u_r varied between 1.2 m s^{-1} at night and 3.8 m s^{-1} in the afternoon. The maximum incoming short-wave radiation was 604 W m^{-2}.

The atmospheric boundary conditions at the reference level z_r = 4.5 m were derived from measurements of global radiation, cloudiness, precipitation, specific humidity, and average wind speed for 24 h from 0000 LMT at 15-min intervals. These values were interpolated at the beginning of each time step (Δt = 600 s). The thickness of soil layers were D_1 = 0–0.1 m, D_2 = 0.1–0.5 m, D_3 = 0.5–1.5 m. The initial conditions for the volumetric soil moisture contents which corresponded to these layers were: ϑ_1 = 0.14 m^3 m^{-3}, ϑ_2 = 0.15 m^3 m^{-3}, and ϑ_3 = 0.15 m^3 m^{-3}. At the initial time, the ground temperature T_g, was 287.15 K while the temperature of the deep soil layer, T_d was 286.15 K. The initial atmospheric pressure was 102.6 kPa.

4.2. Sensitivity to Stomatal Resistance, Leaf Area Index, Canopy Height and Albedo

Numerical modelers must be careful in parameterizing the transpiration since this natural process is very complex. At the same time, it is very important and must be simulated in an accurate way. The comparison between the land-surface schemes and observations, and inter-scheme comparisons by themselves, have shown that the disparity among land-surface schemes is very large in evapotranspiration formulations. This disparity, reflected in the partitioning of surface energy fluxes, seems to be linked to the treatment of canopy processes (Shao et al., 1994). Consequently, we have investigated the sensitivity of the LAPS scheme to the stomatal resistance and the other plant parameters such as the leaf area index, canopy height and albedo. We believe that these parameters significantly determine the quality of the scheme.

Initially, we compared simulations to the observations. Figure 6 shows the diurnal variations of the computed surface energy balance components over a maize field at De Sinderhoeve. The observed values of the latent and sensible heat

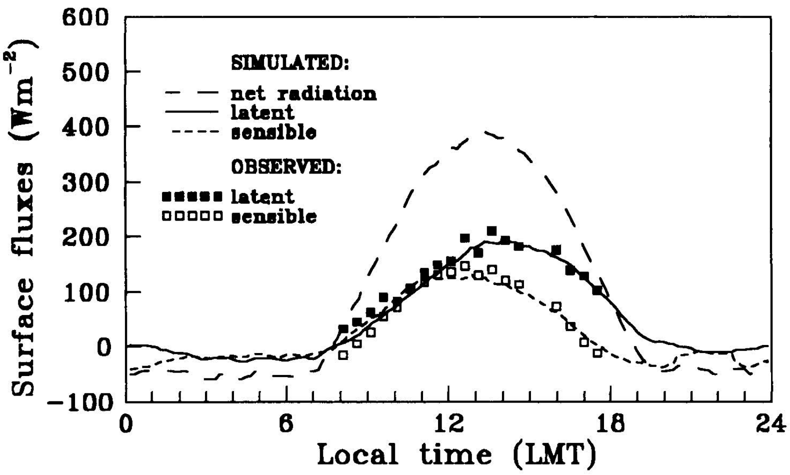

Figure 6. Diurnal variations of surface fluxes simulated by LAPS and observed above a maize canopy at De Sinderhoeve (The Netherlands) for September 8, 1988.

Table VI

Error analysis of the predicted latent and sensible heat fluxes obtained by the LAPS scheme. Statistical quantities have the same meaning as in Table IV. Analysis was performed using the PL89 data set with 18 hourly values

RMSE	MAE	AVER	SDEV	r	α_r
Latent heat flux (W m^{-2})					
19.02	16.72	−0.41	19.02	0.964	49
Sensible heat flux (W m^{-2})					
18.29	15.62	5.84	17.33	0.973	36

fluxes are indicated by black and white squares, respectively. The computed latent heat fluxes agree well with the observations. The computed sensible heat fluxes also agree quite well with the observations. However, the results of the comparison between the simulated values and the observed ones are clarified in Figure 7. Both panels show the computed values of the surface fluxes plotted against the observations. Higher values of the latent heat as well as the sensible heat fluxes are simulated better than the lower ones. In order to present quantitatively the surface fluxes predictions, an error analysis was performed on the PL89 data set, based on statistical parameters mentioned previously. The statistics for the latent and sensible heat fluxes are listed in Table VI. A careful examination of Table VI indicates that the LAPS scheme simulated the surface fluxes quite well for the applied data set, the prescribed vegetation and the soil parameters.

As is well known, the partitioning of energy into sensible and latent heat flux is the most sensitive process to changes in vegetation and albedo. The calculations

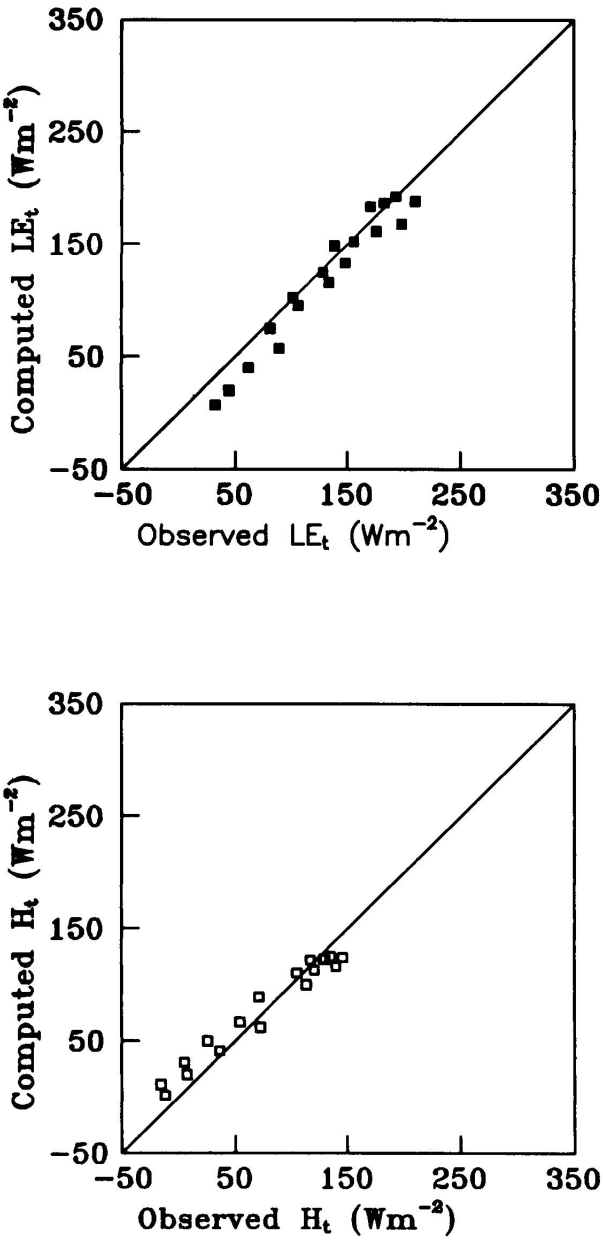

Figure 7. Computed values of surface fluxes plotted against observations using results of simulations from Figure 6 for latent (upper panel) and sensible (lower panel) heat fluxes.

of the latent heat fluxes, as outputs of the sensitivity tests are shown in Figure 8. In all panels, the dashed line represents the values computed with the reference set of vegetation parameters. Careful inspection of this figure clearly indicated that

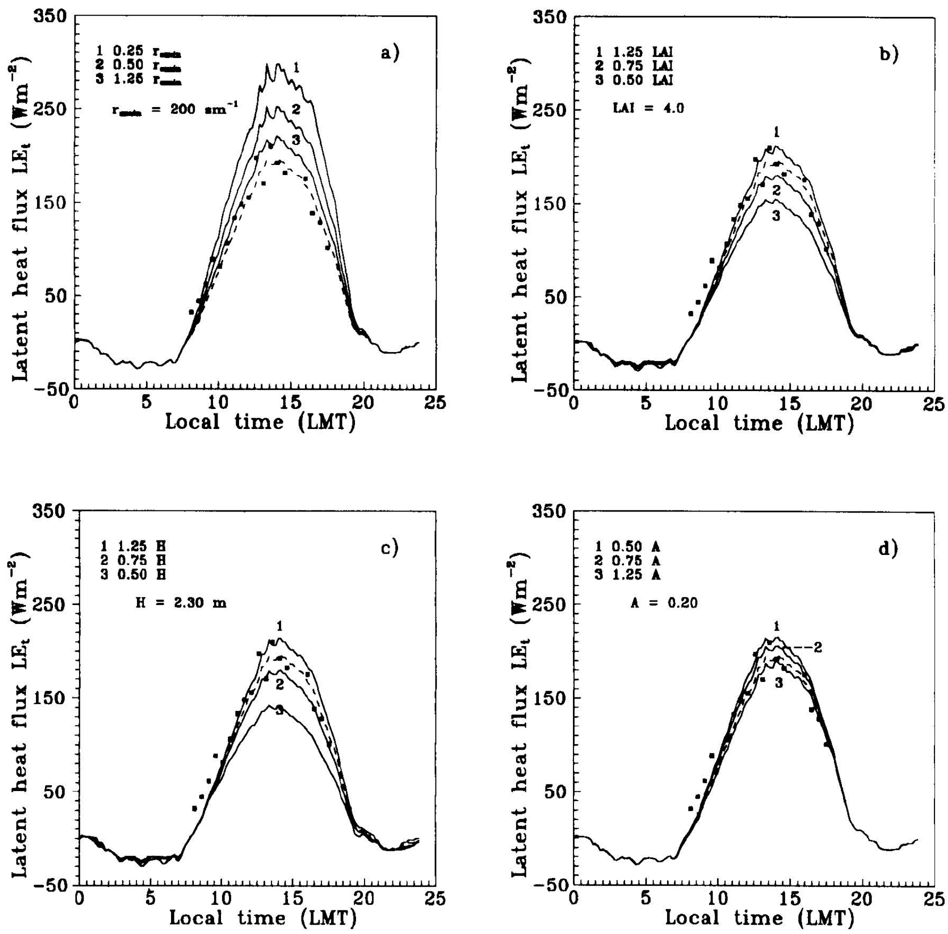

Figure 8. Results of sensitivity tests performed by LAPS for different values of: a) minimum stomatal resistance, $r_{s\min}$; b) leaf area index, LAI; c) canopy height, H and d) albedo A, using PL89 maize data set. The dashed line represents latent heat fluxes simulated with the reference set of parameters. Black squares are observations.

the LAPS is most sensitive to the variations of the stomatal resistance while the smallest variability in the latent heat flux outputs is achieved when the albedo has been changed. Thus, the latent heat flux took the maximum values: 196 W m^{-2}, 221 W m^{-2}, 253 W m^{-2}, and 298 W m^{-2}, when the minimum stomatal resistance, $r_{s\min}$, was: 200 s m^{-1}, 150 s m^{-1}, 100 s m^{-1}, and 50 s m^{-1}, respectively. The changes in minimum stomatal resistance lead to an error in estimation of the water vapor which is transpired into the atmosphere, up to 52%. These results show the importance of this physical parameter.

The runs with various leaf area index (LAI) values showed that this is another vegetation parameter with considerable influence on the latent heat flux. For example, the maximum fluxes computed are: 160 W m^{-2}, 181 W m^{-2}, and 208 W m^{-2} when the leaf area index was: 0.5 LAI, 0.75 LAI and 1.25 LAI, respectively. LAI = 4.0 is the reference value. Following these values, it can be seen that the error in the computed latent heat fluxes introduced by changes of the leaf area index, is around 24%.

The LAPS scheme showed an interesting response to changes of the canopy height H. Looking at panel *c* of Figure 8 it is apparent that the maximum latent heat fluxes are distributed in intervals (143 W m^{-2} for 0.5 H, 180 W m^{-2} for 0.75 H and 214 W m^{-2} for 1.25 H, where H = 2.30 m is the reference canopy height) which are larger than the interval of their distribution when the leaf area index is changed (48 W m^{-2}). Computing the percent of the latent heat flux changes, as a result of the canopy height variations, it was found that it could reach 36% of the reference state.

Finally, the LAPS scheme is less sensitive to the albedo than to other chosen parameters. The maximum latent heat fluxes obtained by the scheme are: 215 W m^{-2}, 207 W m^{-2} and 189 W m^{-2} for values of 0.5 A, 0.75 A and 1.25 A, respectively, where A = 0.2 is the reference albedo. The amount of its deviations from the reference state was 13%.

Since the net radiation module is critically important to the simulation of the evaporation, sensitivity tests were made for variations in albedo, leaf area index and canopy height. The net radiation fluxes as the outputs of these sensitivity tests are presented in Figure 9. In this figure, it is shown that the net radiation module is most sensitive to changes of the albedo. The net radiation reached the maximum values: 449 W m^{-2}, 421 W m^{-2} and 365 W m^{-2}, when the albedo was: 0.50 A, 0.75 A and 1.25 A, respectively. The net radiation module exhibits less sensitivity to the canopy height. According to the panel b of the Figure 9, the net radiation fluxes are distributed in the interval (379 W m^{-2} for 1.25H, 389 W m^{-2} for 0.75H and 396 W m^{-2} for 0.50H). It coincides with the fact that the taller vegetation absorbs more radiation than the shorter one. The LAPS net radiation module has the smallest sensitivity to the variations of the leaf area index. In this case the maximum net radiation fluxes are: 385 W m^{-2}, 390 W m^{-2} and 394 W m^{-2} for values of 0.5 LAI, 0.75 LAI and 1.25 LAI, respectively.

5. Some Concluding Remarks

In this work, the sensitivity of the land surface parameterization scheme, LAPS, was tested and discussed. The tests were made for the parameterization of bare soil processes, ground roughness length, vegetation parameters (stomatal resistance, leaf area index and canopy height) and albedo. This is one of various schemes which have been designed for parameterizing the energy and momentum transport from the surface, on hourly to annual scales. In general, these schemes should be sub-divided into schemes for: climate, numerical weather prediction, region-

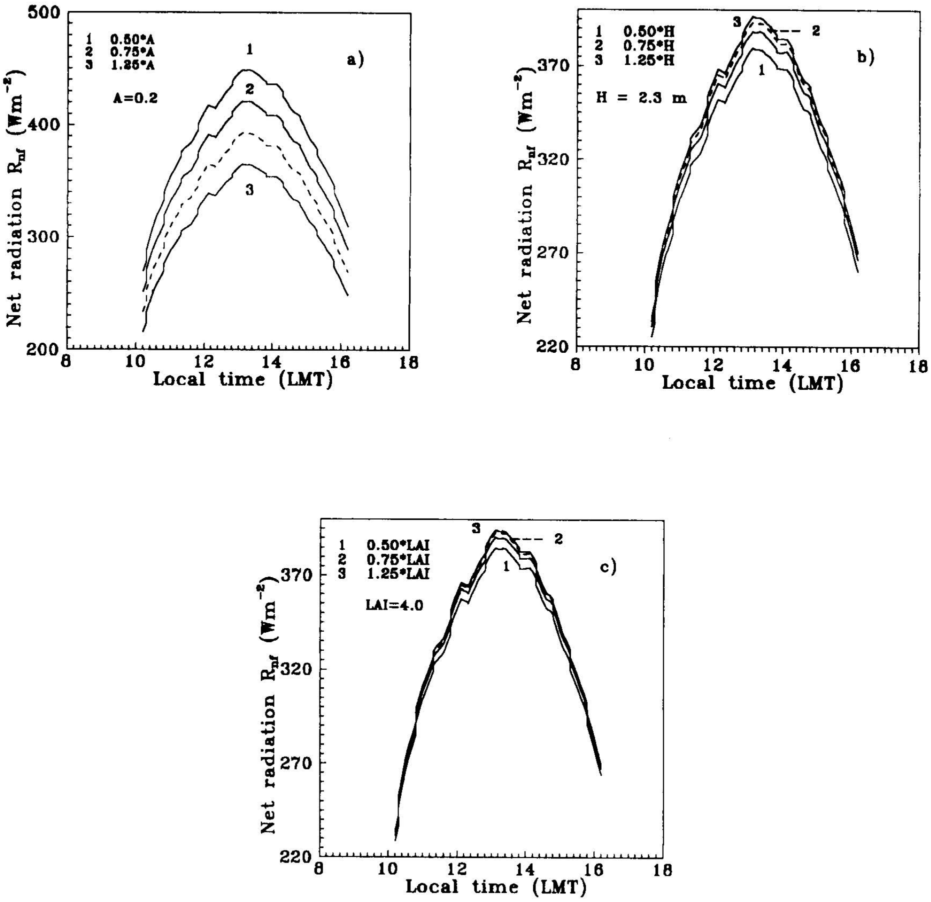

Figure 9. Results of sensitivity tests of the LAPS net radiation module for different values of: a) albedo, A; b) canopy height, H and c) leaf area index, LAI.

al/mesoscale, hydrological and ecological models. According to the design of this scheme, it can be considered as one applicable to meso and larger-scale models. Computationally, it is very efficient and should be used in model simulations where the thermal circulations are considered as very important.

In order to investigate the sensitivity of the LAPS scheme, time integrations with the bare soil and maize data sets were performed. The major conclusions derived from the sensitivity tests using the two data sets from Rimski Šančevi (Yugoslavia) and De Sinderhove (The Netherlands) are as follows:

- LAPS is very sensitive to the choice of the bare soil parameterization scheme. There are very large differences in the simulations of diurnal variations of the latent heat flux, regardless of the "α" or "β" method used. None of them was able to properly reproduce the diurnal variations of the latent heat flux. They either considerably overestimate or underestimate the observed diurnal course of the latent heat flux over a bare soil. It seems that the "α" evaporation scheme incorporated in the LAPS scheme suggested by Mihailović et al. (1993) and the "β" one (Deardorff, 1978) could be considered as a satisfactory choice but they are still far from the desirable quality.

- The LAPS scheme was found to be sensitive to the changes of the ground roughness length and the drag coefficient, as it was expected. The variation of the ground roughness length in the interval from 0.001 to 0.1 m (0.003–0.019 for drag coefficient) introduces, on average, an error of about 35% of the daily latent heat flux of the reference state. Corresponding error introduced in computing the ground temperature varies, on average, from 2.2 °C to 3.6 °C. The smallest error (1.4 °C) was obtained when the value of 0.005 m was prescribed for the ground roughness length.
- The LAPS scheme is extremely sensitive to changes of stomatal resistance. The changes in minimum stomatal resistance should lead to an error in estimation of the water vapor pressure which is transpired into the atmosphere, up to 52% in comparison with the results obtained by the reference data set. The changes in canopy height and leaf area index should produce variations of about 35% and 24%, respectively, in the computed latent heat fluxes. It was found also that the LAPS scheme showed the lowest sensitivity to the albedo variations (13%.).

Finally, it is worth mentioning that before its implementation into an atmospheric model (regional/mesoscale, weather prediction or general circulation model), there is a need for intensive tests on the performance of the hydrological segment of the LAPS land surface scheme.

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