Schemes for Parameterizing Evaporation from a Non-Plant-Covered Surface and Their Impact on Partitioning the Surface Energy in Land–Air Exchange Parameterization

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ABSTRACT

The correct simulation of the sensible and latent heat fluxes from a non-plant-covered surface is very important in designing the surface scheme for modeling the processes in the land–air exchange. However, using different bare soil evaporation schemes in land surface parameterization, an error in partitioning the surface fluxes can be introduced.

In parameterization of evaporation from a non-plant-covered surface in resistance representation, the α and β approaches are commonly used in corresponding formulas where the α and β are functions of soil water content. The performance of different schemes within these approaches is briefly discussed. For that purpose six schemes, based on different dependence α or β on volumetric soil moisture content and its saturated value, are used.

The latent and sensible heat fluxes and the ground temperature outputs were obtained from the numerical tests using the foregoing schemes. The tests were based on time integrations by the bare soil parameterization scheme using real data. The datasets obtained over the experimental site in Rimski Šančevi, Yugoslavia, on chernozem soil were used.

The obtained values of the latent and sensible heat fluxes and the ground temperature were compared with the observed values. Finally, their variability was considered using a simple root-mean-square analysis.

1. Introduction

We are the witnesses of growing demand for more precise land–air modeling schemes and their incorporation in different-scale meteorological models for modeling the atmospheric motions, the hydrological cycle, and the quality control of the air and soil. Several representations of the surface schemes have already been circulated in various meteorological models (Deardorff 1978; Dickinson et al. 1986; Sellers et al. 1986; Mihailović and Jeftić 1994). In these models, authors have chosen different model variables and parameterizations to express the feedback between weather or climate changes and land surface processes.

A vast number of papers have been published on sensitivity studies associated with the importance of (i) various soil and plant parameters and (ii) parameterization schemes chosen as the representative in a land–air scheme (for a recent review, see, e.g., Wilson et al. 1987; Mihailović et al. 1992; and Mihailović et al. 1993).

It has been increasingly apparent that the surface energy fluxes from bare soil, simulated by the model, and their correct partitioning into the sensible and latent heat fluxes are strongly sensitive to the evaporation scheme used in a land–air scheme. According to Avissar and Pielke (1989), the parameterization of the Bowen

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ratio is of prime importance in achieving accurate simulations with mesoscale modeling. Thus, an inaccurate parameterization of the latent heat flux can seriously disturb this ratio, that is, the energy partitioning between the sensible and the latent heat at the surface and, consequently, the accuracy of the model.

It can be said that the collection of the offered bare soil evaporation schemes used in the land–air parameterization is impressive, starting from the well-known Philip's (1957) approach to the recently designed schemes that have been reviewed by Mihailović et al. (1993). They, either in α or β formulation, exhibit large variations in the amount of the simulated latent heat flux as is shown in Mahfouf and Noilhan (1991). Regardless of which of these schemes is used, an error in the latent and sensible heat fluxes is always introduced. That further produces the error in the partitioning of energy between the sensible and the latent heat fluxes at the surface and computing the ground temperature or other prognostic model variables.

This paper is concerned with the analysis of the impact of various schemes for evaporation from a non-plant-covered surface on partitioning the surface energy (i.e., the variability of the Bowen ratio using in situ data). A short review and discussion of the performances of the chosen schemes are presented in section 2. In section 3 we describe the model for computing the surface fluxes using real datasets and different schemes. Finally, the results of time integrations and corresponding discussion are presented in section 4.

2. An overview of schemes for evaporation from a non-plant-covered surface

One of the main processes in the air–land energy exchange is surface evaporation. This process is controlled by atmospheric conditions and surface soil wetness, which is affected by water transport within the soil. The latent heat flux LE_g from a non-plant-covered surface can be written in the form

$$LE_g = [e_*(T_g) - e_r]\frac{\rho c_p}{\gamma}\frac{h_s}{r}, \qquad (1)$$

where E_g represents the evaporation rate, L is the latent heat for evaporation, $e_*(T_g)$ is the saturation vapor pressure at the ground temperature T_g, e_r is the vapor pressure of the air at the reference height z_r, ρ is the density of air, c_p is the specific heat of the air, γ is the psychrometric constant, r is equal to r_a or $r_s + r_a$ depending on whether the α or β method is applied, r_s is the surface resistance, r_a is the aerodynamic resistance between the ground and z_r, 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 to estimate the soil surface humidity of water vapor pressure; that is,

$$h_s = \begin{cases} [\alpha e_*(T_g) - e_r][e_*(T_g) - e_r]^{-1}, \\ \qquad\qquad \alpha \text{ formulation}, \\ \beta, \quad \beta \text{ formulation} \end{cases} \qquad (2)$$

where α and β are the functions of soil wetness. Notice that the α acts like the relative humidity of the air adjacent to the water in the soil pore, although this is not strictly accurate (Wetzel and Chang 1987).

Table 1 summarizes different schemes in α and β formulations used in numerical tests along with recent new proposed forms. This collection of various formulas is practically the same as Table 1 in Mihailović et al. (1993). 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

$$\alpha = \exp\left(\frac{\psi_g g}{R_w T_g}\right), \qquad (3)$$

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. Numerous authors have followed this approach (Sasamori 1970; Nappo 1975; Camillo et al. 1983; Sellers et al. 1986; Mihailović et al. 1990). However, this scheme produces somewhat unrealistic results, especially when the upper soil layer is dry. This fact is considered in more details by Kondo et

TABLE 1. Collection of various formulas for α and β and their combination. 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 the gravity, ψ_g is the soil water potential at the surface, R_w is the gas constant for the water vapor, and T_g is the temperature at the ground.

Investigator	Abbrev	Method and formula
Barton (1979)	BA	$\beta = \begin{cases} 1.8\vartheta/(\vartheta + 0.3), & \vartheta \leq 0.375 \\ 1, & \vartheta > 0.375 \end{cases}$
Deardorff (1978)	DE	$\beta = \min(1, \vartheta/\vartheta_{fc})$
Lee and Pielke (1992)	LP	$\beta = \begin{cases} \frac{1}{4}\left[1 - \cos\left(\frac{\vartheta\pi}{\vartheta_{fc}}\right)\right]^2, & \vartheta \leq \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$
Philip (1957)	PH	$\alpha = \exp\left(\frac{g\psi_g}{R_w T_g}\right)$
Jacquemin and Noilhan (1990)	JN	$\alpha = \begin{cases} \frac{1}{2}\left[1 - \cos\left(\frac{\vartheta\pi}{\vartheta_{fc}}\right)\right], & \vartheta \leq \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$
Mihailović et al. (1993)	P2	$\alpha = \begin{cases} 1 - \left(1 - \frac{\vartheta}{\vartheta_{fc}}\right)^2, & \vartheta \leq \vartheta_{fc} \\ 1, & \vartheta > \vartheta_{fc} \end{cases}$

al. (1990). They showed that the α factor depends strongly on the ratio of the specific humidity of the air and specific humidity of the air saturated 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), who found that for different soil textures 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_g = 293 K and Ψ_g = −160 m, α takes values of 0.989. Figures 1a and 1b show the α and β factors for a chernozem soil computed for the different α and β formulations listed in Table 1.

A vast number of authors have tried to overcome this deficiency of Philip's formula, resulting in a whole class of new formulas in either the α or the 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 formulas 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); it simply assumes some relationship between β and ϑ/ϑ_{fc} for the volumetric soil moisture content ϑ below the field capacity ϑ_{fc}.

3. The model used in time integrations

The numerical model used in the present study is the land surface model (BARESOIL) incorporated into the land–air parameterization scheme UNICOS (Mihailović and Jeftić 1994), which is designed for application in different-scale atmospheric modeling.

The biophysical background of the scheme is based on a single-layer approach commonly used by numerical modelers (Henderson-Sellers et al. 1993). It predicts the foliage temperature, the ground temperature, the water amount stored on the leaves, and the soil moisture content in the three layers. The prognostic equations for the foliage and the ground temperatures are solved using two budget equations for the vegetative cover and the bare soil, respectively. The radiation in the scheme is parameterized according to McCumber (1980). The flux calculations are performed using their resistance representation.

The soil moisture content is taken into account via the three equations, including the terms that represent the canopy extraction of moisture from the first and second soil layers. The complete parameterization of these terms are done following van der Honert (1948) and Clapp and Hornberger (1978).

The amount of the water stored on the leaves is calculated from the equation used by Deardorff (1978), while the transpiration of the remaining part of the leaves is parameterized using the stomatal resistance. The stomatal resistance is calculated considering its dependence upon atmospheric factors (photosynthetically active radiation, vapor pressure deficit of the atmosphere, and air temperature) and upon available water in the soil (Dickinson et al. 1986). Further details of the scheme are reviewed by Mihailović and Jeftić (1994).

The BARESOIL model is based on a model for the prediction of the soil moisture in the three layers as developed by Mihailović (1991) and Mihailović et al. (1993). The prediction of the ground temperature was made using the force–restore method (Bhumralker 1975) and starting from the energy budget equation

$$C_g \frac{\partial T_g}{\partial t} = R_{ng} - H_g - LE_g - \left(\frac{\omega C \lambda}{2}\right)^{1/2} (T_g - T_m), \tag{4}$$

where R_{ng} is the absorbed net radiation, H_g is the sensible heat flux, LE_g is the latent heat flux, ω is the an-

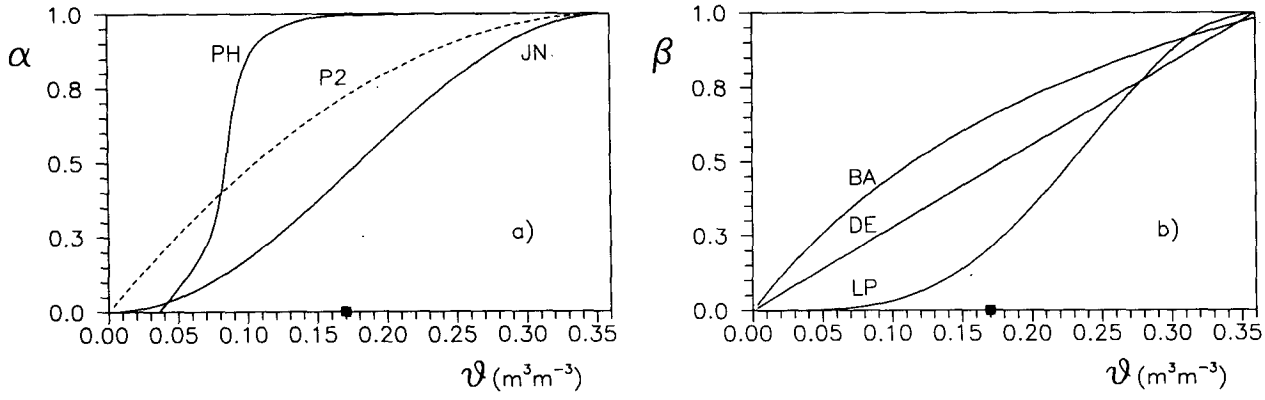

FIG. 1. The α and β factors for a chernozem soil for the different (a) α and the (b) β formulations listed in Table 1. The wilting point for a chernozem soil (■) is marked on the bottom.

gular velocity of the earth, C is the volumetric heat capacity of soil, λ is the thermal conductivity, T_g is temperature in the upper 2 cm of soil, and T_m is the average daily temperature in the 2-cm surface soil layer. In (4), C_g represents the bulk heat capacity per unit area of the upper 2 cm of soil, which is parameterized according to Zhang and Anthes (1982). In the last term of (4), instead of daily and layer-averaged temperatures in the upper 2-cm depth of soil T_m, the average daily soil temperature at 2-cm depth was taken for T_m (Mihailović 1991; Ács et al. 1991). The volumetric heat capacity C of the surface, which depends on the volumetric soil moisture content, is computed following de Vries (1963). The thermal diffusivity K_t is calculated following the approximate formula for a loam soil by de Vries (1963), which is later used in experiments by Wilson et al. (1987) and Mihailović et al. (1992). The thermal conductivity λ was calculated as $\lambda = K_t C$.

The net radiation R_{ng} at the ground accounts for the contributions of the solar radiation (R_s) and the thermal radiation from the atmosphere (R_L) absorbed by the ground. It also accounted for the component of the solar radiation ($\alpha_g R_s$) reflected from the ground. The net radiation is therefore defined as follows:

$$R_{ng} = (1 - \alpha_g)R_s + R_L - \epsilon_g \sigma T_g^4. \quad (5)$$

Here α_g and ϵ_g are the albedo and emissivity of the ground, and σ is the Stephan–Boltzmann constant. The albedo α_g is considered by taking into account the variability with volumetric soil moisture content (McCumber 1980). For the thermal radiation from the atmosphere R_L we used the expression proposed by Staley and Jurica (1972)

$$R_L = [n + (1 - n)0.67(1670q_r)^{0.08}]T_r^4, \quad (6)$$

where n is the cloud fraction, T_r is the air temperature at the reference level z_r, and q_r is its corresponding specific humidity.

The sensible heat flux between the soil surface and the surrounding air, H_g, is

$$H_g = \frac{T_g - T_r}{r_a} \rho c_p. \quad (7)$$

For the latent heat flux LE_g we used expression (1) with different formulations of h_s, including the α and the β functions and their combinations listed in Table 1.

The aerodynamic resistance r_a was calculated as the integral of the inverse value of the transfer coefficient K_m over a specified length; for example,

$$r_a = \int \frac{1}{K_m} dz. \quad (8)$$

Under nonneutral conditions, this integral takes the form

$$r_a = \int_{z_r}^{z_g} \frac{\Phi_h}{k u_* z} dz, \quad (9)$$

where u_* is the friction velocity, z_g is the roughness length of the ground, k is the von Kármán constant taken to be 0.41, and Φ_h is the correction factor for the transport of heat (Businger et al. 1971), where u_* and Φ_h are assumed to be independent of air temperature. For neutral conditions, this integral can be solved analytically and yields

$$r_a = \frac{1}{k^2 u_r} \ln^2\left(\frac{z_r}{z_g}\right), \quad (10)$$

where u_r is the wind speed at the reference level z_r. Further, under stable conditions, integrations of (8) can also be made analytically. Then

$$r_a = \frac{1}{k u_*}\left[\ln\left(\frac{z_r}{z_g}\right) + 4.7\Delta\xi\right], \quad (11)$$

where $\Delta\xi$ is defined as

$$\Delta\xi = \frac{z_r - z_g}{L_0}, \quad (12)$$

where L_0 is the Monin–Obukhov (1954) length. Taking into account the definition of the Monin–Obukhov length, the expression for wind, and (12), we come to a second-order expression for $\Delta\xi$, eliminating L_0 and u_*. The integration of (9) was executed numerically. For the calculation of $\Delta\xi$, an iterative procedure is needed until equality

$$R_{id} = \Delta\xi \frac{I_h}{I_m^2} \quad (13)$$

is reached. In the last expression, I_h and I_m are the integrals of the correction factors Φ_h and Φ_m, respectively, and R_{id} is a bulk Richardson number expressed as a difference

$$R_{id} = \frac{g(z_r - z_g)(T_g - T_r)}{u_r^2 T_{\text{abs}}}, \quad (14)$$

where T_{abs} is the absolute air temperature, here used as 300 K. The surface resistance r_s in (1) is calculated using the empirical expression given by Sun (1982)

$$r_s = 33.5 + 3.5\left(\frac{\vartheta_1}{\vartheta_{s1}}\right)^{-2.38}, \quad (15)$$

where ϑ_1 is the volumetric soil moisture content at the top layer and ϑ_{s1} its value at saturation.

The soil moisture content was parameterized using three equations for three soil moisture storages following Sellers et al. (1986):

TABLE 2. A list of initial conditions for the four cases used in the numerical tests. The variables ϑ_1, ϑ_2, and ϑ_3 are the volumetric soil water content in the three model layers from the top to the bottom, respectively, and T_g is the ground temperature. The observed values for the latent heat fluxes are derived by the method described in section 4. The stars refer to the next day.

Dates	ϑ_1	ϑ_2	ϑ_3	T_g	Time interval (LT)	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

$$\frac{\partial \vartheta_1}{\partial t} = \frac{1}{D_1}\left(P - Q_{12} - \frac{1}{\rho_w} E_g\right), \tag{16a}$$

$$\frac{\partial \vartheta_2}{\partial t} = \frac{1}{D_2}(Q_{12} - Q_{23}), \tag{16b}$$

$$\frac{\partial \vartheta_3}{\partial t} = \frac{1}{D_3}(Q_{23} - Q_3), \tag{16c}$$

where ϑ_i is the volumetric soil moisture content of the ith layer, D_i is the thickness of the ith soil layer, ρ_w is the density of water, P is the infiltration of precipitation, E_g is the rate of evaporation from the soil surface, $Q_{i,i+1}$ is the water flow between i and $i + 1$ soil layers, and Q_i ($i = 3$) is the gravitational drainage from the bottom soil layer. The water flow between adjacent layers is given by

$$Q_{i,i+1} = \bar{K}\left(2\frac{\psi_i - \psi_{i+1}}{D_i + D_{i+1}} + 1\right) \tag{17}$$

and

$$\bar{K} = \frac{D_i K_i + D_{i+1} K_{i+1}}{D_i + D_{i+1}}, \tag{18}$$

where K_i is the effective hydraulic conductivity of the soil layer, which was determined from the empirical formula (Clapp and Hornberger 1978)

$$K_i = K_{si} W^{2B_i+3}, \tag{19}$$

where K_{si} is the hydraulic conductivity at saturation of the ith soil layer. Furthermore, Ψ_i is the soil moisture potential of the ith layer, obtained by the formula (Clapp and Hornberger 1978)

$$\psi_i = \psi_{si} W_i^{-B_i}, \tag{20}$$

where Ψ_{si} is the soil moisture potential at saturation, B_i is an empirical of the ith soil layer, and $W_i = \vartheta_i / \vartheta_{si}$ is the soil moisture wetness of the ith soil moisture. The gravitational drainage rate from bottom was calculated by

$$Q_i = K_{si} W^{2B_i+3} \sin x, \tag{21}$$

where x is the mean terrain slope angle, estimated in this study to be 3° (Vučić 1964).

4. Numerical tests

We have undertaken a series of numerical tests using in situ data with a primary aim to examine the impact of the foregoing described schemes on partitioning the energy in the latent and sensible heat fluxes. We have chosen datasets corresponding to four cases (Table 2) in which the evaporation from a bare soil was calculated by the Bowen ratio method. In this section, we present the procedure of computing the latent heat fluxes from the gradient measurements over the experimental site and the boundary and initial conditions used in the numerical tests.

a. Experimental site description

The Department of Meteorology at the Institute for Crops and Vegetables of the University of Novi Sad has a micrometeorological experimental site at Rimski Šančevi near Novi Sad in northern Yugoslavia (45.33°N, 19.50°E, 84 m above mean sea level) on a chernozem soil of the loess terrace of southern Bačka. A description of its structure and its distribution was given by Mihailović et al. (1993).

The annual mean temperature at the site is 10.9°C, and the average annual precipitation for the period 1948–82 was 603 mm with the maximum in the summer. These data were taken from the neighboring (a distance of 500 m from the soil site) standard meteorological station, Rimski Šančevi, which has been in operation in the Yugoslavian Meteorological Network since 1948.

During the 1981–85 period, a gradient measurements program was carried out. The main objective of that work was to examine the exchange processes of heat, mass, and momentum above bare soil, winter wheat, and soybean. The bare soil area was 50 m × 50 m.

b. Data collecting for the gradient measurements

Except for the thermometers set in the ground, the rain gauge, and the measuring instruments in the thermometer screen, all sensors were fixed to a minitower.

Wind speed was calculated by cup anemometers. Temperature was measured using platinum resistance thermometers (a measurement accuracy 0.1°C). 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, one for exposure from above and one for exposure from below, were used for measurements in the 0.3–2.5-μm shortwave spectral range (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 to 1-m depth by gypsum blocks manufactured in the Department of Meteorology (Mihailović 1983).

From these measurements the latent heat flux was estimated by the Bowen ratio method starting from the energy budget equation

$$R^o_{ng} - H^o_g - LE^o_g - G^o = 0, \tag{22}$$

where R^o_{ng} is the absorbed net radiation, H^o_g is the sensible heat flux, LE^o_g is the latent heat flux, and G^o is the soil heat flux. Hereafter, the superscript or subscript o will refer to observed values and values derived from them. Equation (22) can be transformed to

$$LE^o_g = \frac{R^o_{ng} - G^o}{1 + \beta_w}, \tag{23}$$

where $\beta_w = H^o_g/LE^o_g$ is the Bowen ratio. All terms in Eq. (23) are measured independently or computed from independently measured quantities. Net longwave radiation R^o_{nl} was determined using Ångström's formula (Ångström 1916):

$$R^o_{nl} = [\epsilon_{go}\sigma T^4_{ro}(0.180 + 0.250 \times 10^{-0.094e_{ro}}) + 4\epsilon_g\sigma T^3_{ro}(T_{go} - T_{ro})](1 - 0.70n), \tag{24}$$

where ϵ_{go} is the emissivity of the ground, T_{go} is the ground temperature, and T_{ro} and e_{ro} are the air temperature and vapor pressure at 2 m, respectively. The constants used in Eq. (24) are those introduced by Boltz and Falckenberg (1949). They are assumed to be appropriate for calculating the net longwave radiation during both day and night. Then, the net radiation R^o_{ng} was calculated as

$$R^o_{ng} = R^o_s - R^o_g - R^o_{nl}, \tag{25}$$

where R^o_s and R^o_g are the measured global and reflected solar radiation.

The soil heat flux was estimated from gradient measurements of soil temperature using Ceytin's method (Ceytin 1953; Vereshnin et al. 1959).

The soil heat flux G_o into or out of the soil was determined under the equation

$$G_o = C\left[\frac{Z\Delta T}{\Delta t} + K\left(\frac{dT}{dz}\right)_Z\right], \tag{26}$$

where C is the volumetric heat capacity of the soil, Z is the depth of the daily soil temperature change (e.g., it was 0.30 m for a chernozem soil in this study), $\Delta T/\Delta t$ is the time rate of temperature change averaged for the soil layer from the surface to Z, K is the thermal diffusivity, and $(dT/dz)_Z$ is the vertical temperature gradient existing at the depth Z.

Both terms in Eq. (26) may be written in the finite-difference form. Thus,

$$G^o = \sum_{i=1}^{N} C_i\left(\frac{\Delta T}{\Delta t}\right)_i (Z_i - Z_{i-1}) + \frac{C_N K_N (T_N - T_H)}{H - Z}, \tag{27}$$

where N is the total number of depths in the layer Z meters below the surface, while T_H is the temperature at the depth H, which, depending on soil textures, varies from 0.5 to 1 m.

The time rate of temperature change $(\Delta T/\Delta t)_i$ can be expressed in the form

$$\left(\frac{\Delta T}{\Delta t}\right)_i = [(T_i + T_{i-1/2} + T_{i-1})' - (T_i + T_{i-1/2} + T_{i-1})''], \tag{28}$$

where T_i is the temperature at ith depth and $T_{i-1/2}$ is the temperature in the middle of the ith layer. The prime and double prime denote the values of temperature at the beginning and the end of the time step Δt, respectively. The thermal diffusivity K_N in (27) is estimated following Kolmogorov (1950):

$$K_N = \frac{0.0000274(z_N - z_{N-1})^2}{\ln^2 S} \tag{29}$$

with quantity S defined as

$$S = \frac{(T_{N-1,1} - T_{N-1,3})^2 + (T_{N-1,2} - T_{N-1,4})^2}{(T_{N,1} - T_{N,3})^2 + (T_{N,2} - T_{N,4})^2}, \tag{30}$$

where subscripts 1, 2, 3, and 4 enumerate the values of the soil temperature measured at the beginning of the successive 6-h intervals, during 24 h, starting from 0600 LT in the morning.

Measuring the temperature at chosen depths and using expressions (28)–(30), we can get all quantities that are needed in (27) for calculating the soil heat flux. Dependence of the volumetric heat capacity C_i on the soil moisture in the ith soil layer is considered according to de Vries (1963). In these computations, we used 1) $Z = 0.30$ m, 2) $H = 0.50$ m, 3) the four depths located at 0.05, 0.1, 0.2, and 0.3 m, and 4) $\Delta t = 3600$ s.

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

In some case, mainly near sunset, we have obtained incorrect results when applying 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 become 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 3.

TABLE 3. Hydraulic constants of the chernozem soil of the loess terrace of southern Bačka.

Type	Density (ρ_s)	Ground roughness length (z_g)
Chernozem soil	1290 kg m^{-3}	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

c. *Boundary and initial conditions*

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. The saturated water vapor pressure e_* at any temperature T was calculated by the empirical formula (Goudriaan 1977)

$$e_*(T) = 6.11 \exp\left[\frac{17.4(T - 273.16)}{T - 34.16}\right], \quad (31)$$

where T is in kelvins.

Distribution of the soil layers was $D_1 = 0$–0.1 m, $D_2 = 0.1$–0.4 m, and $D_3 = 0.3$–1 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 2. The soil temperature T_m was assumed to be equal to 301.45 K for all cases. The initial condition for the atmospheric pressure was always the same—101.6 kPa.

Equation (4) was solved using an implicit backward-differencing scheme, while an explicit scheme was applied for solving Eqs. (16a)–(16c).

5. Results and discussion

As we have already mentioned above, we concentrated in this paper on the following question: How do the performances of the chosen schemes influence (a) partitioning the surface energy into latent and sensible heat fluxes and (b) computation of the ground temperature? Certain aspects of subject (a) have been extensively examined and elaborated on in a vast number of papers (Kondo et al. 1990; Mahfouf and Noilhan 1991; Lee and Pielke 1992; Mihailović et al. 1993; Kondo and Saigusa 1994). Mostly, they compared various formulations of evaporation over bare soil, also including the proposed ones, in order to establish the performance of the considered schemes. For example, Mahfouf and Noilhan (1991) tested them against detailed measurements of water and energy exchanges over loamy bare ground. Similarly, Mihailović et al. (1993) performed some tests but using in situ data for a single day.

Following the above works, in some segments, we compared the calculated outputs of the latent and sensible heat fluxes against the observed ones for the four chosen cases with 66 available observed values. All values were, mostly, obtained for the dry conditions when the top soil volumetric moisture content was around the wilting point (Table 2). This number of data could be statistically significant for making more reliable conclusions about partitioning of the surface energy obtained using different approaches. Furthermore, the correct simulation of this process in numerical modeling is strongly related to calculation of the ground temperature. Consequently, its deviation from the observed values could be an important indicator for numerical modelers in estimating the quality of the scheme. Due to the evident absence of this information, a simple root-mean-square analysis of the predicted ground temperature was included. Figures 2–4 present the observed values of the latent and sensible heat fluxes and the ground temperature plotted against the values calculated by the α and β schemes.

Figure 2 shows values of latent heat flux computed by the considered schemes plotted against the observed ones. This figure indicates that the BA, PH, DE, and P2 schemes overestimate the observed values of the latent heat flux, as can be readily seen from the concentration of the squares to the left of the diagonal on the corresponding panel. In contrast to them, the LP and JN schemes simulate evaporation from a bare soil that is systematically smaller than observed in the real world, which is presented on the corresponding panels as the higher concentration of the squares to the right of the diagonal. Further inspection of this figure points

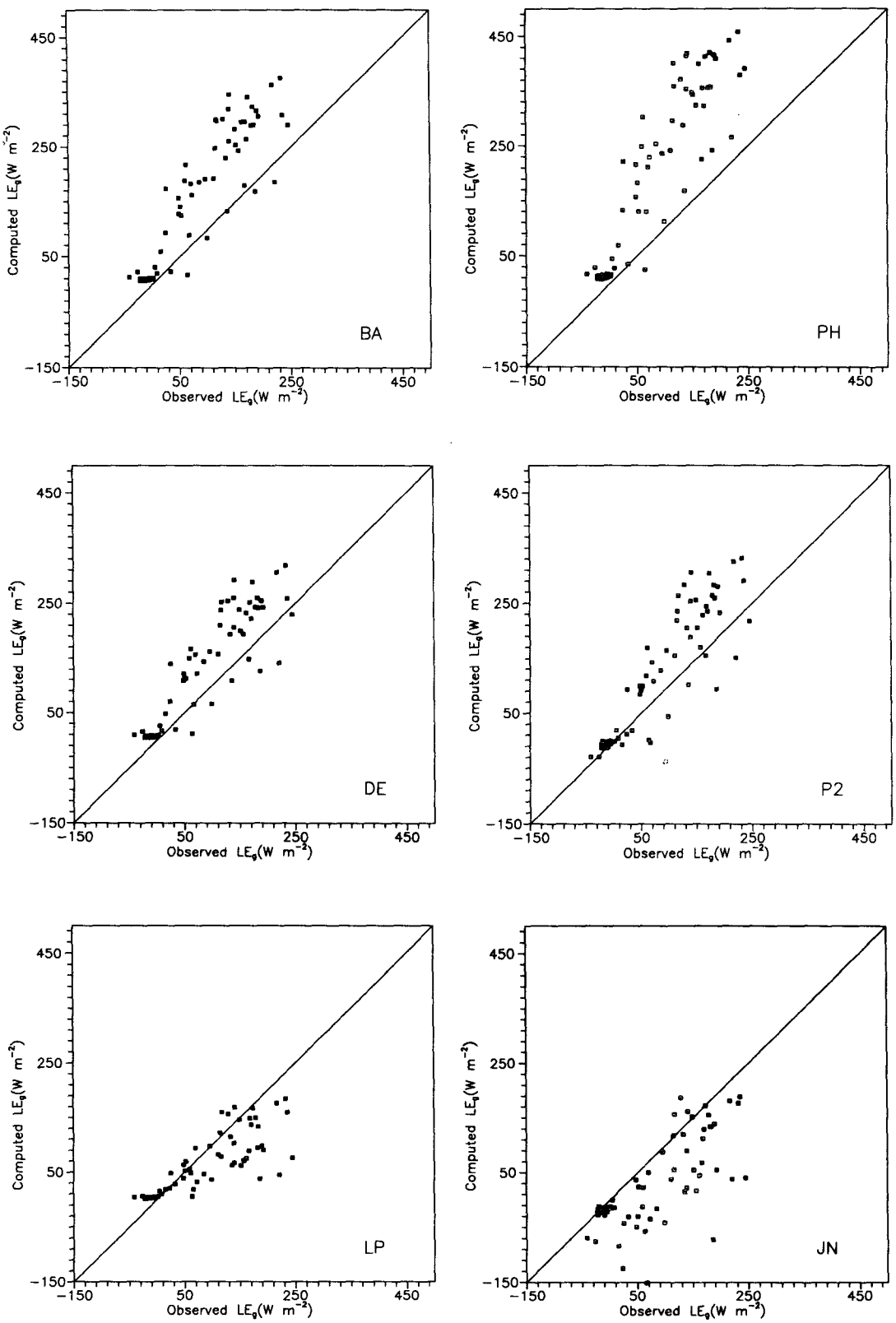

FIG. 2. Values of latent heat flux computed by the α and β schemes plotted against values observed over a chernozem soil at Rimski Šančevi for the four chosen cases (Table 2) used in the numerical tests.

out that the degree of overestimation is not equally distributed among the schemes that give higher values than the observed ones. Thus, the PH scheme practically overestimates all observed values of the latent heat flux. Moreover, that overestimation is more pronounced in comparison with the BA, DE, and P2 schemes. Although, the BA scheme mostly overestimates the observed values of the latent heat flux, the values obtained by this scheme have a tendency toward the values that are lower than in the case when the PH scheme is applied. In the group of schemes that underestimate the observed values of the latent heat flux (LP and JN), a distinction between them can also be made. Namely, the values of the latent heat fluxes computed by the LP scheme are more evenly distributed around the diagonal than they are distributed in the case of application of the JN scheme. Moreover, the JN scheme has a more emphasized tendency toward the lower values.

Regarding the comparison of the sensible heat fluxes displayed in Fig. 3, we can note that values computed by all schemes, except the PH scheme, generally overestimate the observed heat fluxes. In contrast to them, the PH scheme has a very balanced concentration of squares around the diagonal. The BA, DE, and P2 schemes have a concentration of squares toward the higher values with a similar order of magnitude for the overestimated values. In contrast to them, the LP and JN schemes give a concentration of squares with considerable scattering toward the higher values.

Apparently, partitioning the net radiation into the latent and sensible heat fluxes highly determines the variations in the ground temperature. This fact has to be strongly kept in mind when designing the surface schemes. To establish differences between the foregoing schemes in partitioning surface energy into the sensible and latent heat fluxes, we plotted values of the ground temperature computed by the BARESOIL model using the considered schemes against observed ones. These plots are presented on panels in Fig. 4. Looking at the panels with the PH and BA ground temperature outputs, it is clear that they mostly underestimate the observed values of the ground temperature. It coincides with the manner of partitioning the energy by these schemes. Since they remarkably overestimate the observed latent heat flux, it implies that the process of evaporation prevails (Fig. 2). Furthermore, it results in cooling the land surface during the day; thus, the ground temperature predicted by the model is lower than it is in reality. However, the LP and JN schemes simulate evaporation in a manner that restricts more intensive exchange of the latent heat flux between the land surface and the air. Consequently, their use in the surface schemes introduces overestimation of predicted values of the ground temperature, which can be seen in corresponding panels in Fig. 4. It seems that the P2 and DE schemes better simulate evaporation from a non-plant-covered surface compared to other considered ones. They properly balance the sensible and the latent heat fluxes, which results in a more realistic ground temperature predicted by the surface scheme. For both schemes, the squares on DE and P2 panels are concentrated around the diagonal, which indicates a better agreement between the predicted and observed values of the ground temperature.

To quantify the differences between the considered schemes in partitioning the surface energy, we have computed the root-mean-square error (rmse), defined as

$$\text{rmse} = \left[\sum_{i=1}^{n} (Y_i^c - Y_i^o)^2/N\right]^{1/2}, \qquad (32)$$

where Y_i^c and Y_i^o are the computed and observed quantity for the ith value, and N is the total number of observations. The computed values of the rmse of the latent heat fluxes, the sensible heat fluxes, and the ground temperature are presented in Fig. 5. The PH scheme has the highest value of the rmse for the latent heat flux (148.1 W m^{-2}), while the LP scheme has the lowest rmse (53.8 W m^{-2}). All other schemes with their rmse are placed in between these two in the range (96.1–53.8 W m^{-2}). The values of rmse monotonically go down except for the DE and P2 schemes, which are very close (67.1 and 65.0 W m^{-2}, respectively). The values of the rmse of the sensible heat flux go up monotonically from the PH (21.3 W m^{-2}) under the BA (31.6 W m^{-2}) to the JN scheme (114.5 W m^{-2}). The DE and P2 schemes again have the values of the rmse that are close to each other (53.2 and 49.6 W m^{-2}), while the LP scheme gives a higher value of the rmse (96.6 W m^{-2}). Looking at the panel on the bottom of Fig. 5, we can see that the largest error in the ground temperature is made when the BA scheme is employed. It seems that this scheme, among all considered schemes, is partitioning the energy by introducing the largest error in computing the ground temperature (rmse = 4.69°C). The PH and LP schemes introduce smaller errors (4.18° and 4.16°C, respectively) that are nearly equal, although they have quite different partitioning of energy (see middle and upper panels in Fig. 5). Regardless of the large error that is introduced in the sensible heat flux by the JN scheme, this approach has a smaller rmse (3.65°C) than the foregoing schemes. The best results in predicting the ground temperature are achieved by the DE and P2 schemes (rmse = 2.77° and 2.74°C, respectively), which have the most proper balance between the sensible and latent heat fluxes.

The results presented above are synthesized in the three-dimensional plot (Fig. 6). This plot shows the behavior of the considered schemes in computing the latent and sensible heat fluxes and the ground temperature. Apparently, the DE and P2 schemes are de-

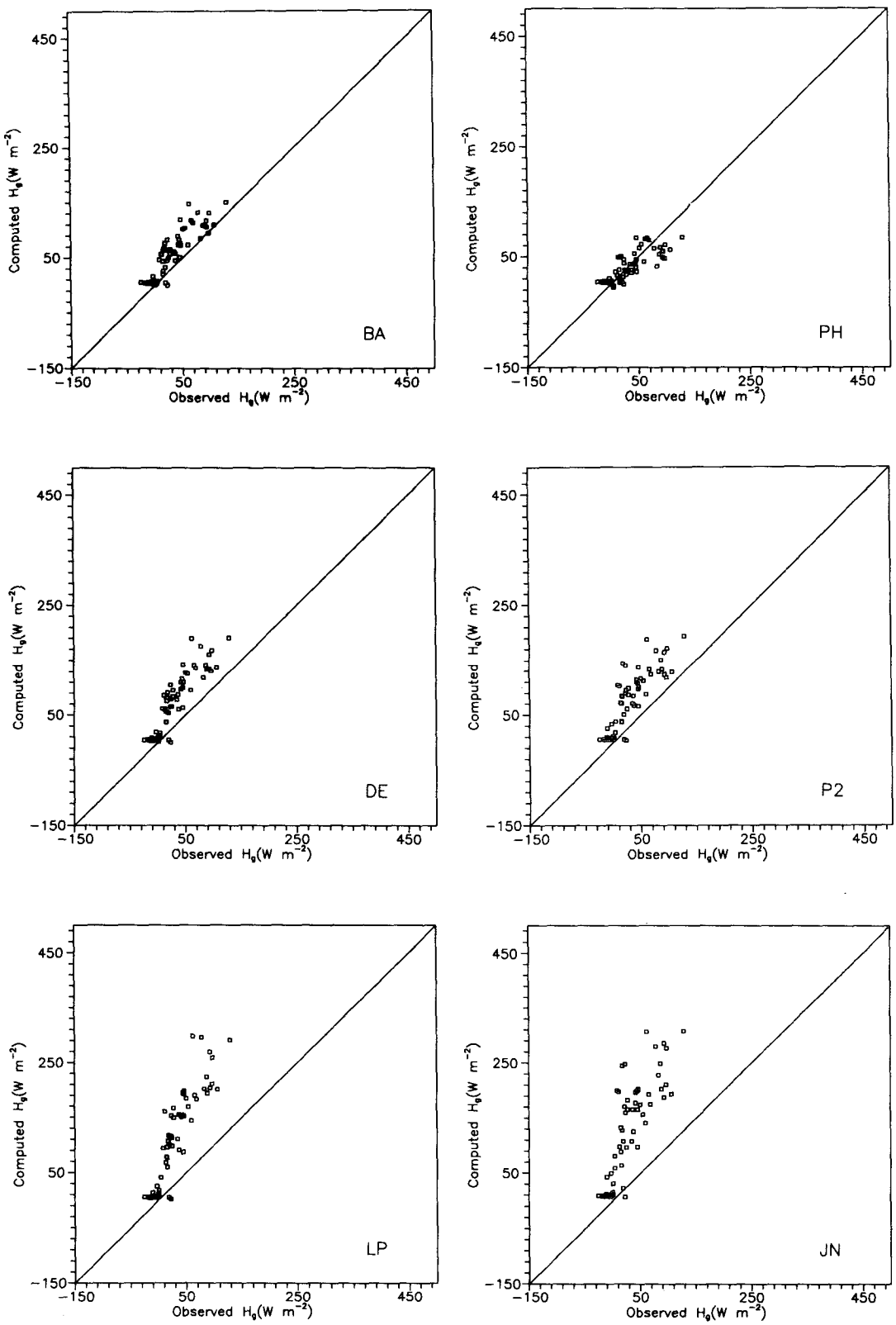

FIG. 3. Values of sensible heat flux computed by the α and β schemes plotted against values observed over a chernozem soil at Rimski Šančevi for the four chosen cases (Table 2) used in the numerical tests.

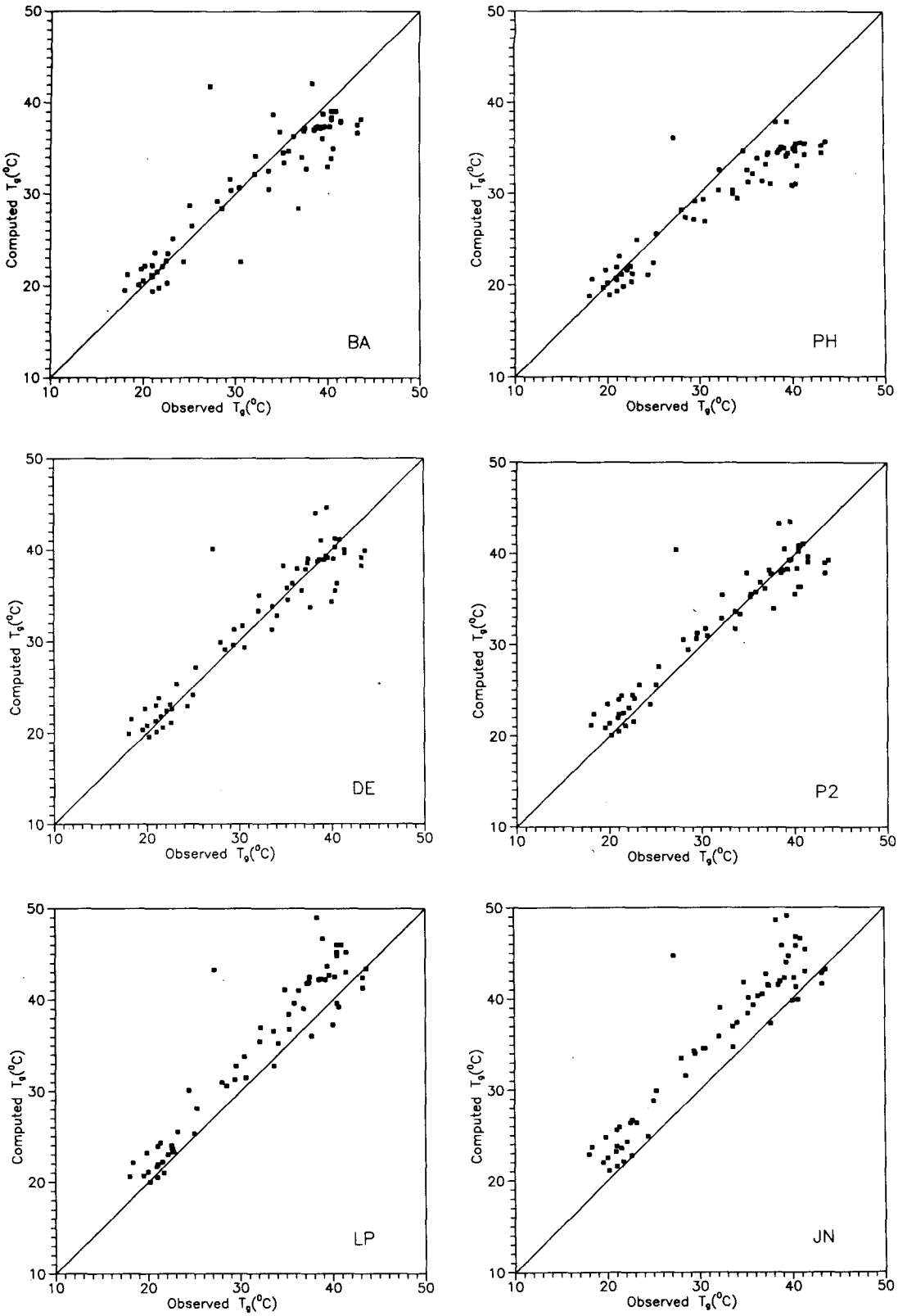

FIG. 4. Values of ground temperature computed by the α and β schemes plotted against values observed over a chernozem soil at Rimski Šančevi for the four chosen cases (Table 2) used in the numerical tests.

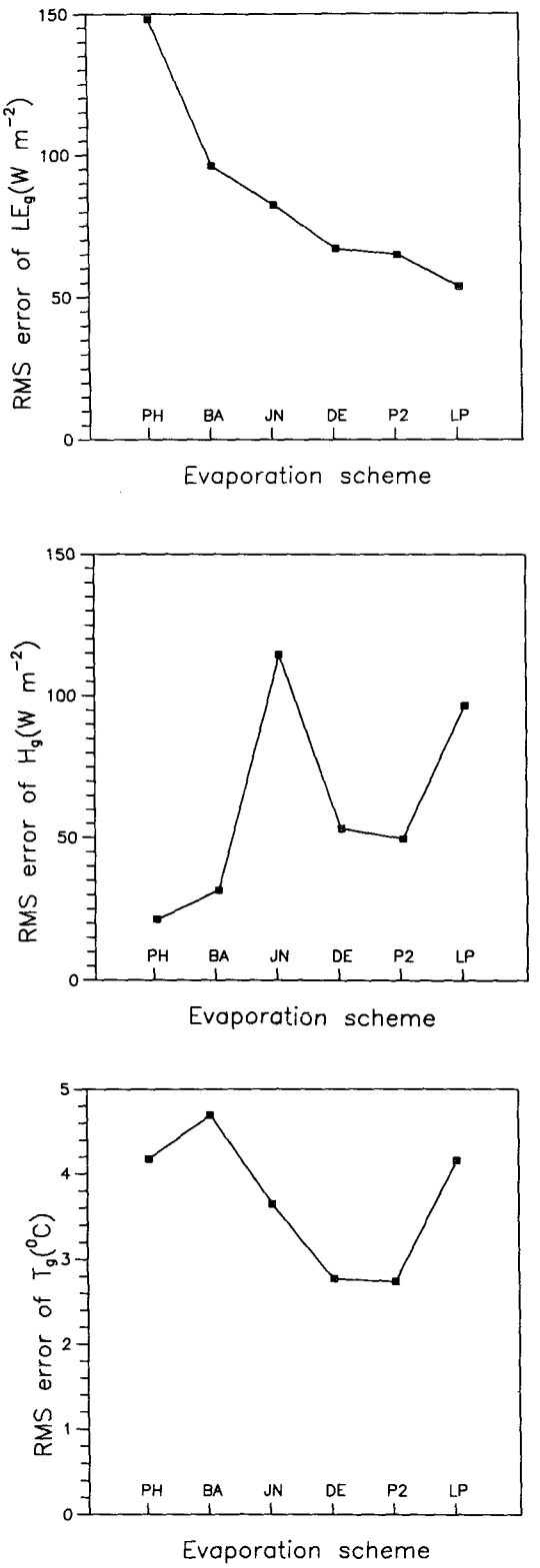

FIG. 5. The root-mean-square errors (rmse) between simulated and observed values of the latent heat flux (upper panel), sensible heat flux (middle panel), and ground temperature (lower panel).

signed in such a manner that they minimize the error in partitioning of surface energy between the latent and sensible heat more than other schemes. Consequently, it results in obtaining more accurate values of the ground temperature. These values are much less than the observed when the other schemes reviewed in this study are applied. However, as can be seen from this plot, these two schemes still introduce large errors in calculating the surface fluxes (66 W m^{-2} for the latent heat flux and 50 W m^{-2} for the sensible heat flux, in average). These suggest that a higher level of sophistication in designing the schemes has to be applied. Looking at the plot, it can be seen that the JN scheme does not reduce the error in the predicted values of the ground temperature as the DE and P2 schemes do. However, this error is still smaller than the error introduced by the PH, LP, and BN schemes. In this three-dimensional plot, the LP and PH schemes take the places that are located diametrically. It is a consequence of the fact that the LP scheme calculates latent heat fluxes much better than the sensible heat ones. In contrast to that, the PH scheme gives more correct values for the sensible heat flux. However, both schemes have exactly the same rmse in calculating the ground temperature, which is still smaller than in the case when the T_g is calculated by the BA scheme.

6. Concluding comments

We have considered the schemes for parameterization from non-plant-covered surfaces and their impact on partitioning of surface energy. We have chosen for comparison datasets for 4 days in which the hourly values of evaporation from the particular soil type (chernozem soil of the loess terrace southern Bačka) were derived from the gradient measurements over the experimental field using the Bowen ratio method. Following the results shown, we can summarize the conclusions as follows:

- All chosen schemes differently deviate from the observations. The scheme proposed by Philip (1957) shows the largest deviations with respect to the observed latent heat fluxes. The smallest variations are obtained by the scheme proposed by Lee and Pielke (1992).
- All considered schemes mostly overestimate the observed values of the sensible heat flux. The smallest variations from the observations give the PH scheme, while the largest ones are obtained by the JN scheme.
- The P2 and DE schemes show results that are very close to each other. Apparently, for both fluxes, a balance between underestimations and overestimations is produced only within these schemes.
- The BARESOIL model used in this study predicts a diurnal course of the ground temperature in the most correct manner with the DE and P2 schemes incorporated in it. Use of other schemes introduces larger errors. Presumably, it is determined by the fact that

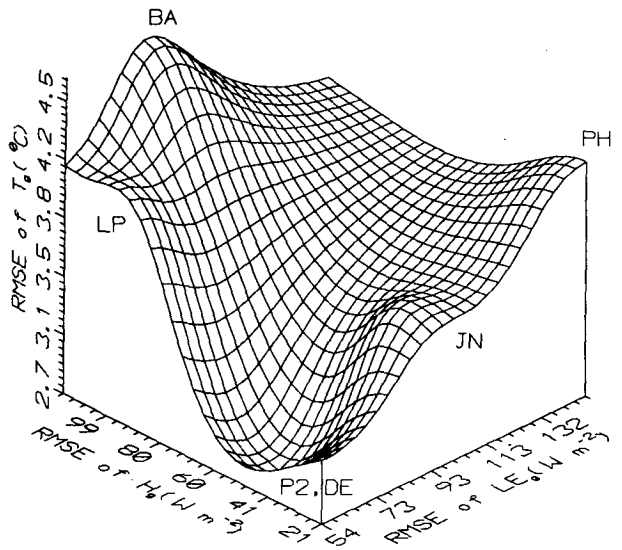

FIG. 6. Three dimensional representation of functional dependence: rmse of ground temperature T_g on the rmse of latent and sensible heat fluxes, LE_g and H_g, respectively.

the DE and P2 schemes reproduce partitioning the energy into the sensible and latent heat more correctly than other schemes.

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