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SENSITIVITY OF SOIL SURFACE TEMPERATURE IN A FORCE-RESTORE EQUATION TO HEAT FLUXES AND DEEP SOIL TEMPERATURE

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ABSTRACT

The 'force-restore' approach is commonly used in order to calculate the surface temperature in atmospheric models. A critical point in this method is how to calculate the deep soil temperature which appears in the restore term of the 'force-restore' equation. If the prognostic equation for calculating the deep soil temperature is used, some errors in surface temperature calculation and consequently in partitioning the surface energy and land surface water can be introduced. Usually, these errors should appear as a result of incorrect parameterization of surface energy terms in the prognostic equation based on 'force-restore' approach.

In this paper, the sensitivity of the 'force-restore' model for surface temperature to the: (a) changes of soil heat flux; (b) variations of deep soil temperature and (c) changes in soil water evaporation is examined. In addition, the impact of the deep soil temperature variations on partitioning the surface energy and land surface water is discussed. Finally, a new procedure for calculating the deep soil temperature based, on climatological data of soil temperature and its exponential attenuation in the deep soil layers is suggested. All numerical experiments with the LAPS land surface scheme were performed using two data sets, obtained from the micrometeorological measurements over a bare soil at Rimski Šančevi (Yugoslavia), RS, and Caumont (France), HAPEX. Copyright © 1999 Royal Meteorological Society.

KEY WORDS: land surface parameterization; surface processes; force-restore equation; surface temperature; surface energy fluxes; surface water fluxes

1. INTRODUCTION

In atmospheric models of all scales (e.g. regional, mesoscale, small-scale models) the surface temperature is computed from either the energy balance equation at the atmosphere–surface interface in diagnostic form or the balance equation of a thin soil layer in prognostic form. In the diagnostic case, the soil heat flux is parameterized very crudely. One possibility is to consider it as a constant part of the net radiation and the second is that the heat capacity of the earth is supposed to be zero with the ground heat flux also zero. Alternatively, Mahrer and Pielke (1977) have calculated the soil heat flux using a full treatment of soil heat diffusion in a multi-level soil model. Bhumralkar (1975) studied the application of procedures for calculating the surface temperature in the context of a general circulation model. He showed that the foregoing assumption of zero soil heat capacity results in an excessive diurnal range of temperature at the soil surface. He also showed that the heat flux into the soil could be represented by the sum of a temperature derivative term and the difference between surface and deep soil temperature. Blackadar

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(1976) also introduced such an expression with slightly modified coefficients for use in a mesoscale model. Practically, he has established one of the most effective procedures for calculating the surface temperature using a prognostic equation based on the 'force-restore' approach. This 'force-restore' model and its later use (Lin, 1980) and generalization (Dickinson, 1988) are still powerful tools in calculating the surface temperature whose variations in a diurnal range are less extreme than when the assumption of zero heat capacity is made.

In expansion of the force-restore model for calculating the surface temperature, the problem of specifying the deep soil temperature in the restore term has been overshadowed by other ones which followed its development. Namely, in this method, of great importance is how to calculate the deep soil temperature. One possibility is to use a prognostic equation while the second is a proper choice of the temperature which will remain constant through the full period of integration. If the prognostic equation for calculating the deep soil temperature is used, some errors in surface temperature calculation and consequently in partitioning the surface energy and land surface water can be introduced. This possibility does exist during the long-term integration when using some atmospheric models for simulations. Usually these errors appear as a result of the incorrect parameterization of the surface energy terms in the prognostic equation which is based on the 'force-restore' approach.

A very limited number of papers have considered this problem in an integrated way. Some of them just partly take into account the problem of the specification of deep soil temperatures using the force-restore method (Jacobsen and Heise, 1982; Savijarvi, 1992). Despite this fact, there are some approaches towards this (e.g. Acs *et al.*, 1991; Mihailović *et al.*, 1996, 1998). In this paper the following questions will be addressed: (1) What happens to the diurnal surface temperature variations when an error in the soil heat flux is introduced? (2) What is the sensitivity of the force-restore model to the variations of the deep soil temperature? (3) What happens to the diurnal surface temperature variations when an error in the energy terms in the prognostic equation for the deep soil temperature is introduced? (4) How can we specify a constant value for the deep soil temperature in order to avoid possible errors in surface temperature calculations? (5) How do changes in the deep soil temperature part of the restore term affect the partitioning of the: (a) surface energy into sensible and latent heat fluxes and (b) land surface water into water balance components?

2. FORCE-RESTORE APPROXIMATION AND MODEL USED

The equation describing the heat transfer into the soil is:

$$C\frac{\partial T}{\partial t}=\frac{\partial}{\partial z}\lambda_{\mathrm{T}}\frac{\partial T}{\partial z}, \tag{1}$$

where z is soil depth, λ_{T} the thermal conductivity and C the volumetric heat capacity. The ratio λ_{T}/C is the thermal diffusivity. For a soil with uniform thermal diffusivity with respect to depth, the boundary conditions often used are: (a) periodic temperature variation of the surface and (b) no temperature change in the deep soil layers. The solution is a periodic temperature variation which decreases in amplitude and increases phase lag with the depth. If the period of the cycle is τ, then the amplitude ΔT of the wave changes with depth should be expressed as:

$$\Delta T=\Delta T_0\exp\left[-\left(\frac{\pi}{k_{\mathrm{T}}\tau}\right)^{1/2}z\right], \tag{2}$$

where ΔT_0 is the amplitude of the surface temperature and k_{T} is the thermal diffusivity.

The time lag Δt_0 associated with the phase shift across a depth range Δz is

$$\Delta t_0=\frac{\Delta z}{2}\left(\frac{\tau}{\pi k_{\mathrm{T}}}\right)^{1/2}. \tag{3}$$

These equations can be applied to an annual ($\tau = 365$ days) or daily cycle ($\tau = 1$ day). For more realistic non-periodic forcing or non-uniform soil properties, (1) can be solved with a variety of numerical schemes, analytical series expansions or Fourier decomposition (Dickinson, 1988). In order to obtain a more realistic solution for temperature variations in the thermal diffusion equation, changes in soil moisture have to be included. Since the soil moisture depends on precipitation, evaporation, runoff, etc., a fully detailed parameterization of soil processes often becomes quite complicated. In addition to that, this multi-level approach requires more computational time.

Due to the fact that most of the surface temperature variations occur within a shallow layer near the surface Bhumralkar (1975) and Blackadar (1976) suggested a two-layer approximation where a shallow slab of soil is bounded below by a thick constant temperature slab (Garratt, 1992). In fact, they have developed a slab on a two-layer model of the soil, based on the energy balance for a near-surface layer of thin depth. The corresponding prognostic equation for calculating the surface temperature T_0 has the form:

$$C_0 \frac{\partial T_0}{\partial t} = R_0^{\text{net}} - H_0 - \lambda E_0 - C\left(\frac{k_{\text{T}}\omega}{2}\right)^{1/2}(T_0 - T_{\text{d}}), \tag{4}$$

where R_0^{net} is the net radiation of the surface, H_0 is the sensible heat flux, λE_0 is the latent heat flux at the surface, λ is the heat of vaporization, C_0 is the heat capacity per unit area, $\omega = 2\pi/\tau$, and T_{d} is a deep soil temperature at the substrate heat reservoir. This prognostic equation is frequently referred as the force-restore method. This is due to the forcing by the $R_0^{\text{net}} - H_0 - \lambda E_0$ part which has been modified by the restoring term $C[k_{\text{T}}\omega/2]^{1/2}(T_0 - T_{\text{d}})$ that includes T_{d}. This term restores T_0 exponentially towards T_{d} if the forcing term has been removed. In practice, the heat capacity per unit area C_0 is often calculated from the expression $C_0 = 0.95C(k_{\text{T}}/2\omega)^{1/2}$ (Blackadar, 1976; Zhang and Anthes, 1982; Mihailović, 1991). When the T_0 is calculated, then the soil heat flux G_0 can be obtained from the equation:

$$G_0 = C\left[\frac{k_{\text{T}}\omega}{2}\right]^{1/2}(T_0 - T_{\text{d}}). \tag{5}$$

The numerical model used in this study is LAPS (Land–Air Parameterization Scheme), developed by the Faculty of Agriculture at the University of Novi Sad, and partly by the Center for Meteorology and Environmental Modelling, University of Novi Sad. This scheme is designed as a software package which can be run as a part of an atmospheric model or as a stand-alone model.

The processes parameterized in the LAPS are divided into three parts fully describing: subsurface thermal and hydraulic processes; bare soil transfer processes and canopy transfer processes. The BARESOIL module for the prediction of surface temperature and soil moisture in three layers, of the LAPS used for the present experiments has been comprehensively described by Mihailović *et al.* (1993), Mihailović (1996), Mihailović and Ruml (1996), and Mihailović and Kallos (1997); thus in this section, a short identification of the module structure governing equations and the hydrological part will be given.

The prediction of the surface temperature was made by using Equation (4). The thermal diffusivity k_{T} in this equation is calculated by using the approximate formulation for a loam soil proposed by de Vries (1963) which was later adopted by Wilson *et al.* (1987) and Mihailović *et al.* (1992).

The parameterization of the volumetric soil moisture content is based on the concept of the three-layer model, which is already supported by Sellers *et al.* (1986) and Mihailović (1991). The governing equations take the form:

$$\frac{\partial \vartheta_1}{\partial t} = \frac{1}{D_1}\left(P_1 - Q_{1,2} - \frac{1}{\rho_{\text{W}}}E_0 - R_0 - R_1\right), \tag{6}$$

$$\frac{\partial \vartheta_2}{\partial t} = \frac{1}{D_2}(Q_{1,2} - Q_{2,3} - R_2), \tag{7}$$

$$\frac{\partial \vartheta_3}{\partial t} = \frac{1}{D_3}(Q_{2,3} - Q_3 - R_3), \tag{8}$$

where ϑ_i and D_i ($i = 1, 2, 3$) are the volumetric soil moisture content and the thickness of the ith layer, respectively, and ρ_w is the density of water, P_1 is the infiltration of precipitation, E_0 is the rate of evaporation from the soil surface, $Q_{i,i+1}$ is the water flow between i and $i+1$ soil layer, Q_i ($i = 3$) is the gravitational drainage from the bottom soil layer, R_0 is the surface runoff and R_i is the subsurface runoff from the ith soil layer. Equation (4) is solved by an implicit backward method, while Equations (6)–(8) are solved using an explicit time scheme. The calculations of fluxes and other terms in Equations (4)–(8), based on papers by Businger *et al.* (1971), Clapp and Hornberger (1978), Deardorf (1978) and some original solutions (Mihailović *et al.*, 1993, 1995a,b; Mihailović and Rajković, 1994; Mihailović and Kallos, 1997) are completely described in these papers.

The surface runoff R_0 is computed as:

$$R_0 = P_1 - \min(P_1, K_s), \tag{9}$$

while the subsurface runoff, R_i, for each soil layer is calculated using the expressions:

$$R_1 = F_{1,2} - \min(F_{1,2}, K_s), \tag{10}$$

$$R_2 = F_{2,3} - \min(F_{2,3}, K_s), \tag{11}$$

$$R_3 = F_3 - \min(F_3, K_s), \tag{12}$$

where K_s is the hydraulic conductivity at saturation.

At the end of every time step, Δt, the variable, Γ_i is calculated as

$$\Gamma_i = \frac{D_i}{\Delta t}[\vartheta_i^k + A_i \Delta t - \vartheta_c], \tag{13}$$

where ϑ_i^k is the volumetric soil moisture content at the beginning of the time step, A_i represents the terms of the right-hand side of Equations (6)–(8) and ϑ_c is a wet reference parameter like field capacity or volumetric soil moisture content at saturation, depending on soil texture. If the condition $\Gamma_i > 0$ is satisfied, then the Γ_i becomes runoff, which is added to corresponding subsurface runoff, R_i. Consequently, at the end of the time step, the calculated value of the volumetric soil moisture content ϑ_i^{k+1} takes the value ϑ_c (Mihailović *et al.*, 1998).

3. THE DEEP SOIL TEMPERATURE AND FORCE-RESTORE EQUATION

3.1. *Data sets, boundary and initial conditions used*

In order to examine the sensitivity of the force-restore equation to the deep soil temperature some experiments were performed using the model described previously. For this purpose, the data from the sites of Rimski Šančevi (Yugoslavia) and Caumont (France) were used.

The experimental site Rimski Šančevi (45.33°N, 19.5°E), altitude of 84 m, on a chernozem soil is located in the northeastern part of Yugoslavia. A description of its structure and distribution was given by Mihailović *et al.* (1993). Some of its hydraulic properties are listed in Table I. In this study, we used data sets collected during a large measurement campaign which examined the exchange processes of heat, mass and momentum above bare soil, winter wheat, and soybean planted surfaces during the growing season in the period 1981–1985. The selected cases are referred as RS0603 for 3 June 1981, RS0527 for 27 May 1982, RS0604 for 4 June 1982, RS0611 for 11 June 1982, RS0624 for 24 June 1982, and RS0706 for 6 July 1982. The soil temperature was measured using platinium resistance thermometers located at 0.02-, 0.05-, 0.1-, 0.2-, 0.3-, 0.5-, and 1-m depths. More details of measuring the surface fluxes and surface temperature can be found in Mihailović *et al.* (1995a,b) and Mihailović and Kallos (1997). In all the RS data sets, the atmospheric boundary conditions at the reference height $z_r = 2$ m were derived from measurements of global and reflected radiation, cloudiness, precipitation, wet and dry-bulb temperatures, and average wind speed over 1 h intervals. Then the measured values were interpolated to the beginning

Table I. Soil properties of a chernozem soil where the experimental site Rimski Šančevi is located

	Symbol	Value
Type: Chernozem soil		
Density	ρ_s	1290 kg m^{-3}
Ground roughness length	z_0	0.01 m
Hydraulic properties		
Saturated soil 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
Dry soil volumetric heat capacity	C_s	1344×10^3 J m^{-3} K^{-1}
Dry soil thermal conductivity	λ_s	0.3145 J m^{-1} s^{-1} K^{-1}
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_0	0.97

of each time step which was 600 s in this study. The thicknesses of the soil layers were: $D_1 = 0.1$ m, $D_2 = 0.3$ m, and $D_3 = 0.6$ m. The initial conditions for the volumetric soil moisture contents (ϑ_1, ϑ_2, ϑ_3) and the surface temperature (T_0), are given in Table II for all the cases. The deep soil temperature (T_d) was used either as a single value or as a value calculated from a prognostic equation, depending on the experiment used. The initial condition for the atmospheric pressure was always the same—1016 hPa. The HAPEX data set in its present form was prepared by Shao *et al.* (1994).

The data were obtained from a HAPEX-MOBILHY experiment during 1986 at Caumont (SAMER No. 3, 43°41′N, 0°06′W, mean altitude 113 m) on the loam soil. Some of the soil properties are listed in Table III. Detailed information on the SAMER network and the site can be found in Goutorbe (1991) and Goutorbe and Tarrieu (1991). If data at Caumont were missing, measurements from neighbouring meteorological stations were used. The reference data set which was used for investigating the impact of different deep soil temperatures to the partitioning of the land surface water, was the one representing the water budget for the first 4 months (days 0 to 120). This was generated by using the observed weekly root zone (1.5 m), the soil moisture content, the cumulative precipitation and the evaporation estimated by the Penman–Monteith formula. The estimated evaporation was 149.6 mm for the period indicated. The total precipitation was 368.5 mm. The available observations from the first 4 months show very little changes in total soil moisture. The total root zone water content change was estimated at -22.3 mm while the

Table II. A list of initial conditions for the six cases at Rimski Šančevi used in the numerical simulations

Dates	ϑ_1	ϑ_2	ϑ_3	T_0	Time interval
June, 1981					
3	0.15	0.22	0.23	295.25	500–400*
May, 1982					
27	0.20	0.21	0.22	289.35	500–400*
June, 1982					
4	0.16	0.23	0.26	291.75	500–400*
11	0.12	0.18	0.20	292.25	500–400*
24	0.18	0.20	0.21	292.85	500–400
July, 1982					
6	0.18	0.18	0.18	292.55	500–400*

'*' denotes the next day.

Table III. Soil properties of a loam soil where the experimental site Caumont is located

	Symbol	Value
Type: Loam		
Density	ρ_s	1500 kg m^{-3}
Ground roughness length	z_0	0.01 m
Hydraulic properties		
Saturated soil moisture potential	Ψ_s	-0.300 m
Saturated hydraulic conductivity	K_s	0.4×10^{-5} m s^{-1}
Clapp–Hornberger's constant	B	5.66
Dry soil volumetric heat capacity	C_s	1214×10^3 J m^{-3} K^{-1}
Dry soil thermal conductivity	λ_s	0.2719 J m^{-1} s^{-1} K^{-1}
Field capacity	ϑ_{fc}	0.36 m^3 m^{-3}
Volumetric soil moisture content at its saturation	ϑ_{si}	0.47 m^3 m^{-3}
Photometric properties		
Emissivity	ε_0	0.97

generated runoff plus drainage was 241.1 mm during the 4 months. The atmospheric forcing data: downward short- and long-wave radiation, precipitation, air temperature, wind speed, atmospheric pressure and specific humidity at 2 m, were available at 30 min intervals for 1 year continuously. The time step in the experiment with the HAPEX data set was also 1800 s. The thicknesses of the soil layers were: $D_1 = 0.1$ m, $D_2 = 0.4$ m, and $D_3 = 1$ m. The initial conditions for the volumetric soil moisture contents were: $\vartheta_1 = 0.32$ m^3 m^{-3}, $\vartheta_2 = 0.33$ m^3 m^{-3} and $\vartheta_3 = 0.35$ m^3 m^{-3}. For the deep soil temperature a single value of 280 K was used.

3.2. Sensitivity of the surface temperature calculated using the 'force-restore' method upon the deep soil temperature

The soil heat flux at the interface, G_0 depends on a number of factors, e.g. solar radiation and surface forcing, time of the day, soil type, soil moisture content and other physical properties. In atmospheric model calculations, the soil heat flux can be neglected but this is not always appropriate since it may become a significant part of the net radiation. In addition, its inaccurate calculation can introduce an error in calculating the surface temperature and, consequently, can affect the accuracy of the model. In order to illustrate the impact of the soil heat flux on the surface temperature calculation, a numerical experiment was performed. For this purpose, the model described in the previous section with the field data set RS0624 was run. The error in the soil heat flux was simulated by its linear changes in the range from -50% to 50% upon the reference values of the latent heat flux. Errors in the numerical simulation of the surface temperature produced by the error in the soil heat flux were calculated. They were quantified by the root-mean-square error (RMSE) and mean-absolute error (MAE). The calculated values of RMSE and MAE are presented in Figure 1. From this figure, it is evident that the largest RMSE and MAE values of the surface temperature are 0.57 and 0.51°C, respectively, obtained when the error artificially introduced in the soil heat flux was -50% of its reference value. It can be also seen that there is a slight asymmetry for both statistical parameters with the errors which are larger in the case when the soil heat flux takes lower values upon its reference state.

The impact of the errors in the soil heat parameterization on calculating the surface temperature is further investigated by analysing the diurnal variations of the surface temperature, latent and soil heat fluxes, obtained with the two introduced errors in the soil heat flux, i.e. -50% and 50% upon its reference state. The integrations were performed with the RMS0624 data set representing a hot day (maximum of the air temperature was 27.9°C). During this day, a low amount of water in the top soil layer was just above the wilting point (0.17 m^3 m^{-3} for a chernozem soil), while the other two layers were well supplied by water (Table II). The partitioning of the net radiation, for this case, into corresponding fluxes expressed through the percentages of their daily sum was, 48% for the latent heat flux, 37% for the sensible heat flux and 15% for the soil heat flux. In this experiment the model was running with the deep soil temperature T_d calculated from the prognostic equation:

$$C_0 \frac{\partial T_d}{\partial t} = (R_0^{net} - H_0 - \lambda E_0)/\sqrt{365\pi}, \tag{14}$$

which also was used by Xue *et al.* (1991). In the following paragraphs, this method for calculating the deep soil temperature will be referred to as TDX.

The diurnal variations of the surface temperature and fluxes simulated for the RS0624 data set are presented in Figure 2. The runs were performed for the reference state when there were no changes in the soil heat flux, and for two cases when the errors of -50% and 50% upon their reference state were introduced. As is seen in Figure 2(a), when the soil heat flux is restricted to half of its reference value, the diurnal course of the surface temperature predicted by Equation (4) follows the reference one up to noon. In the afternoon hours, the predicted temperatures are systematically higher than those obtained by the reference data set. These differences are not greater than 1.5°C. This behaviour of the surface temperature course can be explained by the fact that the artificial restriction of the soil heat flux, results in an increase in the latent and sensible heat fluxes (Figure 2(b) and (c)). In addition, this redistribution of the surface energy enhances evaporation from the bare soil and consequently causes the heating of the adjacent layer of air. In contrast to this, when the amount of the soil heat is increased by 50%, the simulated surface temperatures become lower than those obtained by the reference experiment. The response to the significantly increased soil heat flux is that the surface becomes cooler and therefore the evaporation rate and the sensible heat flux becomes lower. This simple experiment illustrates a significant sensitivity of the force-restore equation, and consequently the calculated surface temperature, to the changes in the restore term. An important source of errors in the soil heat flux term of Equation (4) could be addressed to the specification of the deep soil temperature T_d. A very limited number of papers is devoted to this problem including physical aspects and sensitivity tests. A short elaboration related to this problem can be found in Acs *et al.* (1991) and Mihailović *et al.* (1996). In order to better understand the response of the restore term in changes of the deep soil temperature, the following simulations were performed. Several runs were executed for T_d growing for the temperature increment of 0.5°C in the interval from 15 to 40°C. For each T_d, the RMSE of the surface temperature was calculated. The RMSE was calculated using the soil temperatures measured by resistance thermometers at a depth of 2 cm for the reference state (Acs *et al.*, 1991; Mihailović *et al.*, 1995a). The results obtained are plotted in Figure 3. As seen in this figure a minimum in the RMSE of surface temperature occurs, regardless of the data set used. This shows that it is always possible to find such a value for the deep soil temperature, T_d, for which the RMSE of the daily course of the surface temperature has a minimum (in the following text it will be referred as the TDR method). There is no general conclusion which could be derived about the regularity of occurrence of this

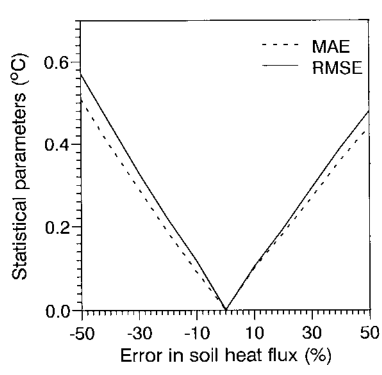

Figure 1. Statistical parameters for differences between the simulated and reference values of T_0 which results from -50% to 50% errors imposed on the originally calculated soil heat flux

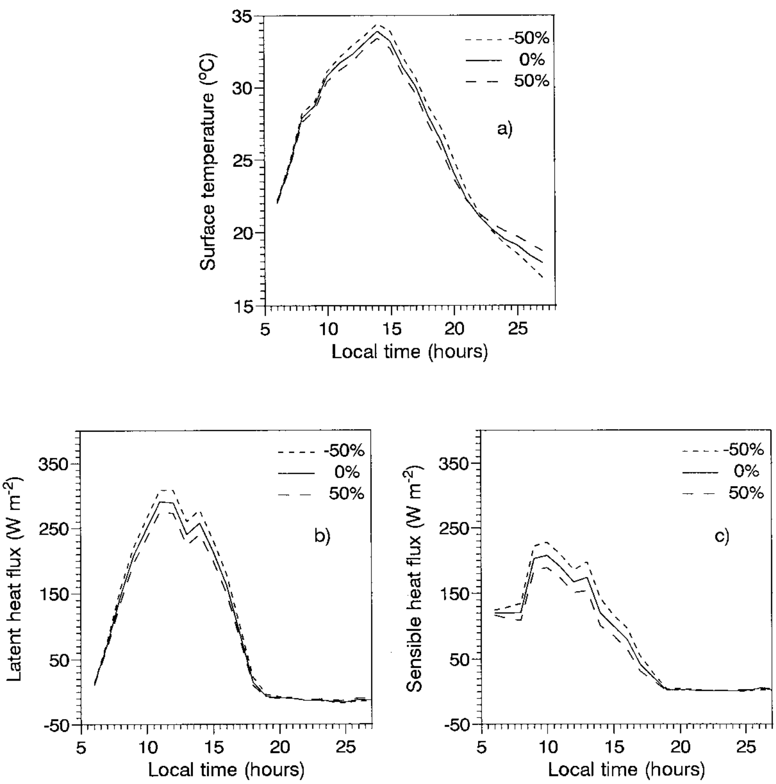

Figure 2. Diurnal variations of the: (a) surface temperature; (b) latent heat flux and (c) sensible heat flux simulated by using the RS0624 data set

minimum as well as the procedure for obtaining the corresponding value of T_d. The only comment which should be made about the sensitivity of the surface temperature, T_0, to the changes of the deep soil temperature, T_d, is that there is some value of T_d for which the partitioning of surface energy in the force-restore Equation (4) is the most proper, so that the calculated values of T_0 exhibit the smallest deviations from the observations.

3.3. The 'force-restore' method for surface temperature and specifying the deep soil temperature

The 'force-restore' approach is considered as a very powerful tool where a prognostic equation for temperature is used in order to reproduce exactly the response to periodic heating and uniform thermal properties of the soil. In this method, of great importance is the method for calculating the deep soil temperature which appears in the restore term. One possibility is to use a prognostic equation while the second is a proper choice of the temperature that will remain constant through the full period of integration.

If the prognostic equation for calculating the deep soil temperature is used, some errors in the surface temperature calculation, and consequently, in partitioning the surface energy and land surface water can be introduced. This possibility exists constantly during the long-term integration, with an atmospheric model when this method is applied. Usually, this error appears as a result of incorrect parameterization of the surface energy terms in the prognostic equation based on the 'force-restore' approach. In various parameterization schemes, besides that of Equation (14), the following prognostic equation for calculating T_d is used:

$$\frac{\partial T_d}{\partial t} = \frac{1}{\tau}(T_0 - T_d), \tag{15}$$

where τ is equal to 86400 s. This equation, which was initially used by Noilhan and Planton (1989), will be denoted as the TDN method.

The error in the energy terms of the Equation (14) is usually introduced by the incorrect parameterization of the evaporation from bare soil regardless of the approach used (i.e. *alpha*, *beta* or threshold method). The subject of parameterizing the evaporation from bare soil in atmospheric models has been comprehensively elaborated by numerous authors (Kondo *et al.*, 1990; Mahfouf and Noilhan, 1991; Lee and Pielke, 1992; Mihailović *et al.*, 1995a; Mihailović and Kallos, 1997 among others). They have recommended some approaches for the calculation of bare soil evaporation for use in atmospheric models. In addition, Mihailović and Rajković (1994) and Mihailović *et al.* (1995a) have discussed the impact of the use of different bare soil evaporation schemes on the variability of surface temperature in land–air parameterization. Also, they have indicated the occurrence of a large error in the calculation of the surface temperature when the evaporation is incorrectly parameterized.

In order to estimate the order of magnitude of the error in calculating T_d, introduced by the incorrect parameterization of bare soil evaporation in Equation (14), the previous simulations were repeated with

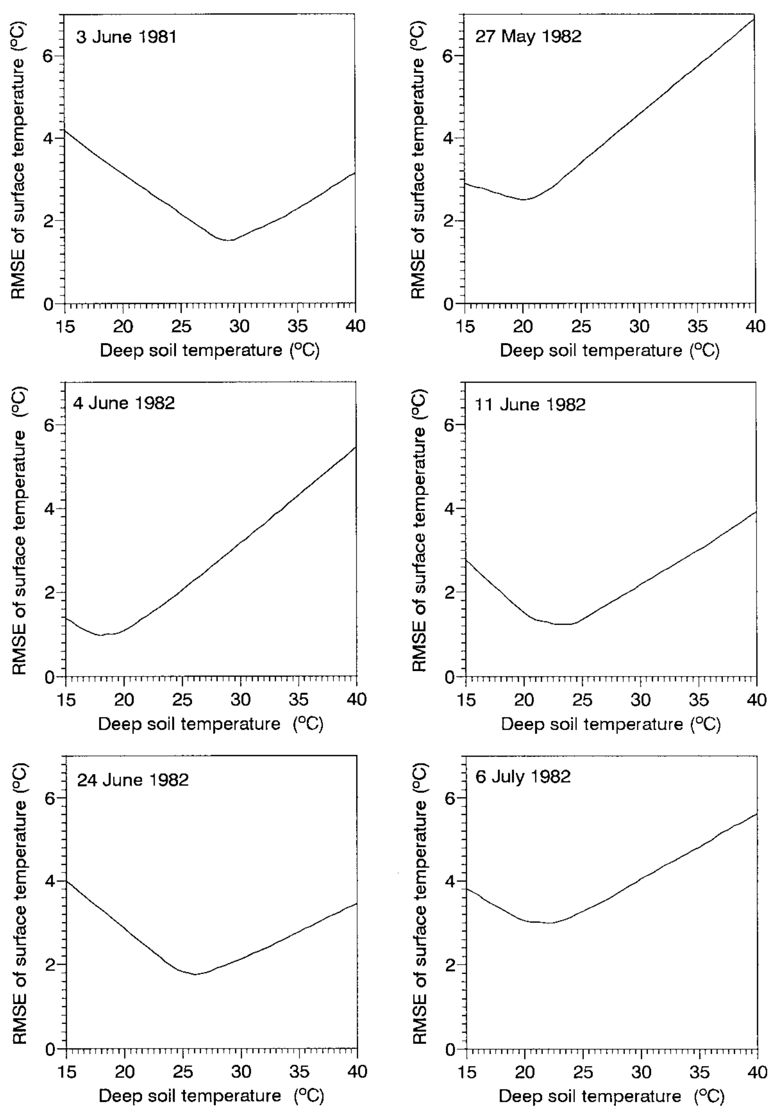

Figure 3. RMSE of T_0 calculated from Equation (4) as a function of different values of deep soil temperature T_d

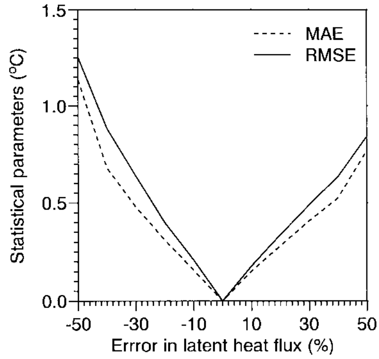

Figure 4. Statistical parameters for differences between the simulated and reference values of T_0 which results from -50% to 50% errors on the originally calculated latent heat flux in Equation (14)

the introduction of an artificial error in the latent heat flux which was linearly changed from -50% to 50% and using the RS0624 data set. The calculated values of RMSE and MAE are shown in Figure 4. In this figure it can be seen that the largest RMSE and MAE values of the surface temperature are 1.29 and 1.14°C, respectively. An asymmetry for both statistical parameters is also evident. The larger errors occurred in the case where the latent heat flux takes lower values from the reference state.

Instead of calculating the T_d from either Equations (14) or (15), it should be specified as a simple value which will be kept constant during the period of integration. This value can be derived considering the annual course of the soil temperature and its change with the depth. Following the solution of the equation for the heat diffusion and the expression (2), the daily mean temperature T_z, for a Julian day J, at a depth z, can be written in the form:

$$T_z = T_a + \Delta T_0 \, e^{-\gamma_a z} \sin[\omega_a(J-110) - \gamma_a z], \tag{16}$$

where T_a is the mean annual temperature, ΔT_0 is half of the mean annual amplitude at 2 cm, $\gamma_a = (\omega_a/2K_T)^{1/2}$, $\omega_a = 2\pi/\tau_a$ and $\tau_a = 365$ days. The value of 110 days corresponds to the date of April 20 when T_z takes the value of T_a for the physical conditions of a chernozem soil at Rimski Šančevi. Let us mention that the date (expressed in Julian days) of zero gradient in Equation (16) varies from year to year and from location to location. Following the 'force-restore' approach, the depth z should be replaced with the depth:

$$d = \left[\frac{2\lambda}{\omega_d C}\right]^{1/2} \ln\frac{1}{\beta}, \tag{17}$$

where $\omega_d = 2\pi/\tau_d$ and $\tau_d = 1$ day and β is the ratio of amplitude of depth temperature to that of surface temperature. The quantity defined by Equation (17) is referred to as the damping depth (Sellers, 1972) since the amplitude is reduced to $e^{-\ln 1/\beta}$. After some simple algebra we reach the expression for T_d in the form:

$$T_d = T_a + \Delta T_0 \, e^{-\sqrt{\omega_a/\omega_d}} \sin\left[\omega_a(J-110) - \sqrt{\frac{\omega_a}{\omega_d}}\right]. \tag{18}$$

By replacing the numerical values of ω_a and ω_d in this equation, a simple expression for the mean daily temperature T_d can be written in the form:

$$T_d = T_a + 0.949 \, \Delta T_0 \sin[0.0172(J-110) - 0.0523]. \tag{19}$$

Equation (19) can be slightly modified and applied for different soil types and geographical locations. In the following text, this method for calculating T_d will be referred to as the TDG. For practical calculation of T_d, it is necessary to know only the climatological values of T_a and ΔT_0 which are available from the climatological archives all over the world. It is worth noting that the value of T_d, obtained by the foregoing method, is constant during 1 day of integration. During a long period of integration its changes follow the annual course at the depth chosen according to Equation (17). In order to check the validity of Equation (19), the surface temperature was calculated by using T_d, as the mean daily value of the soil temperatures at the damping depth when $\beta = 1$. The damping depth, estimated by using Equation (17) and the physical properties of chernozem soil listed in Table I, was 0.11 m. The corresponding temperatures at this depth were obtained by the interpolation of the soil temperatures measured hourly at depths of 10 and 20 cm (Mihailović *et al.*, 1993). In the following text this control specification of the deep soil temperature will be referred to as the TDD method.

Following the considered methods, the outputs of the calculated surface temperature were compared with the observed ones for the six chosen cases, with 132 observed values. This amount of data could be considered as adequate in order to check the reliability of the various methods for calculation of the surface temperature by the force-restore equation, using different approaches for specifying the deep soil temperature. Consequently, its deviation from the observed values could be an important indicator for the accuracy of the method followed. In Figure 5 the observed values of the surface temperature are plotted

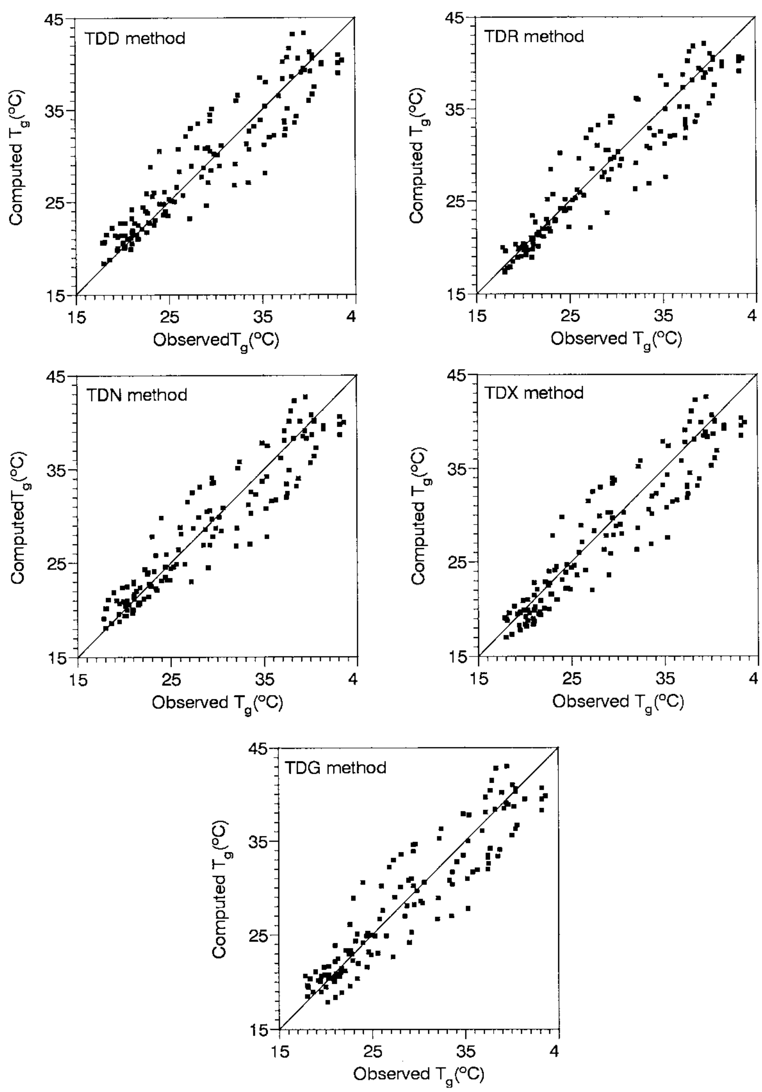

Figure 5. Values of surface temperature calculated by different methods for specifying the deep soil temperature, plotted against the observed ones at Rimski Šančevi

against the values calculated by the different approaches for calculating the deep soil temperature. Looking at the panels of this figure, it can be clearly seen that the scatter around the diagonal is approximately the same for all the methods and slightly less for the TDR method. From the same figure it is also apparent that the deviations of the simulated values from the observations are less in the region of the lower values of the surface temperature. This behaviour of the calculated T_0 should be addressed to the performance of the bare soil evaporation scheme used in Equation (4).

In order to quantify the differences between the considered approaches for calculating the deep soil temperature, the simple statistical parameters MAE and RMSE have been calculated. The calculated values of the statistical parameters MAE and RMSE of the surface temperature are shown in Figure 6. The TDD and TDG approaches have the highest values of the RMSE at 2.7°C. The TDX and TDN methods gave slightly lower values (i.e. 2.5 and 2.4°C, respectively), than the other two schemes. The lowest errors in RMSE of the surface temperature are achieved by the TDR method (2.3°C). Concerning the MAE parameter it should be noted that the TDD and TDG approaches gave the same values (− 2.2°C). The TDX and TDN methods also gave similar results (2.1 and 2.0°C, respectively). The lowest values of the MAE for the surface temperature are obtained when the TDR method is applied. The comparison of the TDG method with the control method TDD, derived from measured values of soil temperature at the damping depth, indicates that there is almost no difference between them. In addition, the values of the surface temperature calculated by using the force-restore equation are comparable with those obtained by the control method TDD. Similar results are also achieved when the deep soil temperature is calculated from the prognostic Equations (14) and (15).

Obviously, some errors can be introduced into the calculations of the deep soil temperature, depending on the method applied. These errors should affect the estimation of T_d which is used in the partitioning of the surface energy into sensible and latent heat fluxes and the land surface water into water balance components. In order to establish the sensitivity of Equation (4) to the changes of the deep soil temperature and their impact to the partitioning of the surface energy into sensible and latent heat fluxes, several simulations have been performed with the RS0624 data set. In these simulations, T_d has been varied from 15 to 40°C with a 1°C increment. With these values of T_d the corresponding sensible and latent heat fluxes were calculated. In addition to that, the statistical parameters MAE and RMSE were also calculated. The reference state in these tests were the surface fluxes obtained by running the model with the value of T_d corresponding to the minimum shown in Figure 3 for each data set. The calculated statistical parameters, concerning the surface energy partitioning test, are shown in Figure 7. As is seen in this figure, the sensible and latent heat fluxes exhibit deviation from the reference state when the deep soil temperature changes linearly. It is also seen that the errors in calculating the sensible heat fluxes are greater than those for the latent heat fluxes.

For establishing the sensitivity of the 'force-restore' equation to changes of the deep soil temperature and partitioning the land surface water, several runs have been performed using the HAPEX data set. In

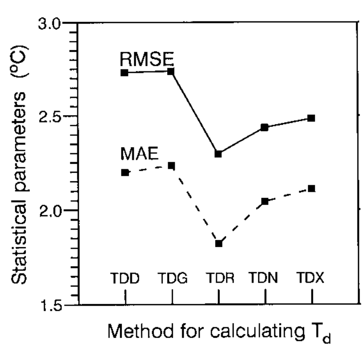

Figure 6. Statistical parameters for differences between the observed and simulated values of the surface temperature calculated by different methods for estimation of the deep soil temperature T_d

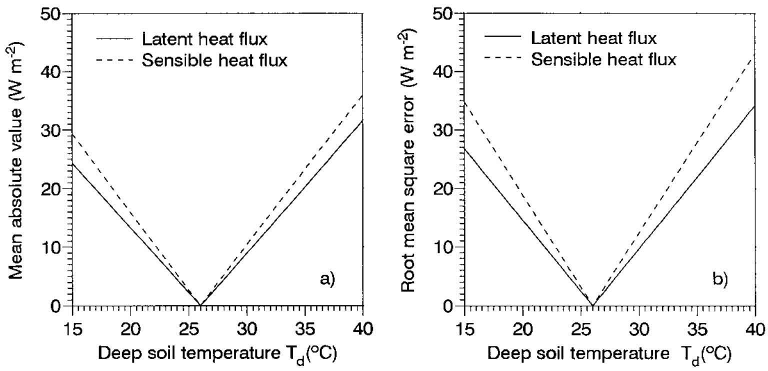

Figure 7. Statistical parameters MAE (a) and RMSE (b) for differences between the simulated and reference values of surface fluxes over a bare soil at Rimski Šančevi depending on T_d in Equation (4)

these runs, T_d has been varied from 2 to 20°C using a 1°C increment. The cumulative values of the water balance components were calculated. The cumulative values of soil water change used in these tests are obtained from the soil moisture budget equation, derived from the Equations (6)–(8), which can be written as:

$$\frac{dM}{dt} = P_r - E - R - D, \tag{20}$$

where M is the mass of water in a soil column of unit area, P_r is the precipitation, E is the evapotranspiration or bare soil evaporation, R is the runoff representing the loss of soil water through horizontal flow and D is the drainage describing the loss of water in the vertical direction. All these quantities are expressed as mm s^{-1}. The integration was performed for the first 120 days of 1986 at Caumont (France).

The calculated values of the water balance components as the function of T_d are presented in Figure 8. It is seen from this figure, that there is a decrease of cumulative values of runoff and drainage. These changes are grouped into two domains; the first one is for the temperature domain 2–17°C, where the slope of runoff and drainage line is steeper than it is in the second interval of temperatures extending from

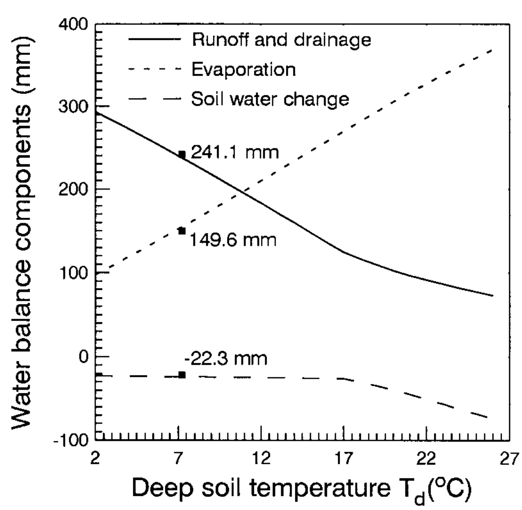

Figure 8. Partitioning of the land surface water into water balance components during the 4 months of integrations over a bare soil at Caumont (France) depending on T_d in Equation (4). Black squares indicate the observations

17 to 27°C. It means that, in the first domain, the water leaves the soil column horizontally and vertically, such that there is a decrease of 11.7 mm when deep soil temperature is increased by 1°C. It is also worth mentioning that there are no variations in the soil water change when the deep soil temperature is below 17°C; when T_d is below this value the soil water change quickly decreases. From Figure 8 it is seen that for $T_d = 7$°C, all three simulated water balance components are very close to observed values (151.8 mm for the evaporation, 239.5 mm for the horizontal and vertical water flows and -22.9 mm for the water change). The value of 7°C was used for T_d in calculating the water balance components and soil water change during the 120-day integration.

4. CONCLUSIONS

In this study, the 'force-restore' equation was used in order to calculate the surface temperature. At the same time, the deep soil temperature in the restore term of this equation was specified in different ways. Data sets used for the numerical experiments were taken from the micrometeorological measurements at Rimski Šančevi (Yugoslavia) and Caumont (France), for 6 and 120 days, respectively. Based on the results presented, we can summarize the conclusions as follows.

(i) The force-restore model used for the calculation of the surface temperature is sensitive to the errors introduced in the restore term.
(ii) It is always possible to find a minimum value of the deep soil temperature in the restore term for which the partitioning of the surface energy, in the force-restore equation, is the most proper so that the calculated values of the surface temperature exhibit the smallest deviations from the observations.
(iii) The use of a prognostic equation in order to specify the deep soil temperature can introduce an error in the calculations of the surface temperature which has an order of magnitude of 1°C.
(iv) The suggested method for calculating the constant value of the deep soil temperature can be applied in the restore term of the 'force-restore' equation.
(v) This method was found to be reliable enough especially for long-term integration.
(vi) The changes of the deep soil temperature significantly affect the partitioning of the surface energy as well as the partitioning of the land surface water.
(vii) Finally, the advantages as well as limitations of the considered methods for specifying the deep soil temperature in the 'force-restore' equation can be summarized as follows. The use of the TDR and TDN methods is restricted by the fact that there is not an easy and universal way to find T_d. Also, this T_d is not necessarily and physically the same T_d as in the other methods. In calculating T_d by the TDX method, apparently, the partitioning of energy is more or less arbitrary. It actually varies daily, seasonally and with location and soil moisture content. The TDD method can be implemented in the land surface parameterization or other calculations when the daily or annual soil temperature data sets are available. In the method suggested (TDG) we have to know the date of the zero soil temperature gradient (20 April in this study). This method is very useful for initialization in different atmospheric models.

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