Consistency of Modeling the Water Budget over Long Time Series: Comparison of Simple Parameterizations and a Physically Based Model

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ABSTRACT

This paper investigates the treatment of the soil water budget of two parametric models currently used in atmospheric models for climate studies. Because the parametric models are intended to represent areally averaged behavior, results of the water budget for both models are compared to output of a well-tested physically based, one-dimensional unsaturated flow model with spatially heterogeneous soil hydraulic properties. Computations are performed for the three models using two datasets of soil hydraulic properties and three separate years of daily average meteorological conditions. Neglecting the percolation process in land-surface parameterizations can lead to very unrealistic results in evapotranspiration estimates. Evapotranspiration efficiencies of the parametric models show more rapid time fluctuations compared to the physically based model. Furthermore, it appears that a selected reference set of soil hydraulic properties behaves similar to the areally distributed properties for soil water balance computation in the absence of surface runoff.

1. Introduction

The atmosphere and the land surface are dynamically coupled through exchange processes of energy and water. These processes are governed by the physical properties of soil and vegetation and the atmospheric conditions. The exchange of energy and water are linked through evapotranspiration. According to Budyko (1974), global evapotranspiration from land surfaces amounts to approximately 60% of the global precipitation. The contribution of evapotranspiration to the total net radiative energy transfer from land surfaces to the atmosphere is of the same order of magnitude. Shukla and Mintz (1982) show the large influence of evapotranspiration on climate. Evapotranspiration is either limited by the available energy or the availability of water in the soil. The soil moisture status also affects the surface energy balance through its influence on shortwave albedo and soil thermal conductivity. Simulation experiments with general circulation models (GCMs) and mesoscale climate models confirm the sensitivity of climate to the moisture status of the soil (Mahfouf et al. 1987; Meehl and Washington 1988). Delworth and Manabe (1988, 1989) address the mechanisms through which the surface soil moisture status affects the simulated climate within a GCM and emphasize the effect of persistence of soil moisture anomalies. From the previous work, it is thus clear that a realistic treatment of evapotranspiration and the surface water budget is indispensable in climate modeling.

Combination of the continuity principle with the Darcy flow equation yields the governing partial differential equation for dynamical unsaturated water flow in soils. Numerical solution of this nonlinear equation for transient boundary conditions is computationally intensive, preventing operational use in climate modeling. Hence, it is necessary to describe soil moisture movement in a simplified, yet realistic manner. Many different parameterizations of the soil–vegetation–atmosphere interface have been developed. They range from the simple "bucket" model (Manabe 1969) to complex resistance formulations such as the Simple Biosphere Model (Sellers et al. 1986). These models intend to represent areally averaged behavior of the water budget.

The aim of this study is to investigate the ability of the bucket-type parameterizations (Warrilow 1986) and the force–restore-type parameterizations (Deardorff 1978; Noilhan and Planton 1989) to describe the water budget consistently in time. Both models are outlined later. From the many different available parameterizations, these two were selected because of their fundamentally different treatment of the lower boundary condition, as will be discussed later. Verification using measured data can still not be performed. Although several experiments have been conducted on various spatial scales (André et al. 1986; Sellers et al. 1988), they mainly focused on measuring and inte-

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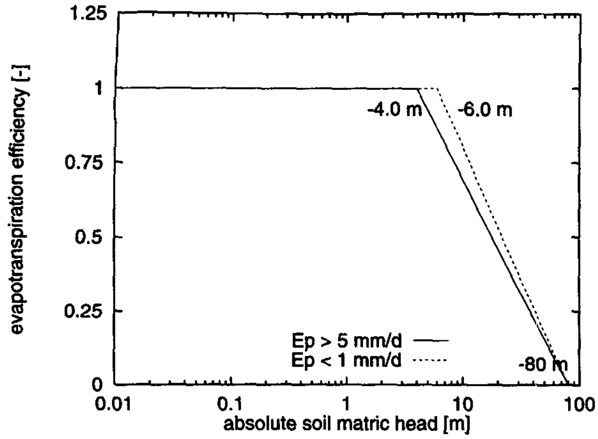

FIG. 1. Evapotranspiration efficiency function $\beta(\psi, E_p)$ after Feddes et al. (1978).

grating the atmospheric fluxes. As indicated before, these fluxes can be strongly affected by the dynamics of soil moisture movement. Nevertheless, dynamical behavior of soil moisture has received only little attention and because large-scale field experiments cover relatively short time periods, long-term effects of dynamical soil moisture behavior are hardly revealed. The long-term dataset of soil moisture at points over Russian territory, introduced by Vinnikov and Yeserkepova (1991), may prove to be very valuable for validation of GCM simulations. However, these point observations do not reveal the influence of the considerable amount of variation of soil properties found on smaller spatial scales.

Therefore, an alternative indirect approach has been adopted as explained hereafter. Given a set of measured soil hydraulic properties within an inhomogeneous region, the mean water budget components can be determined using a detailed multilayer model based on the governing flow equation and a scaling theory for spatially variable hydraulic properties. The ability of this method to realistically simulate the water budget of an inhomogeneous area was shown by Hopmans and Stricker (1989). The reason the measured dataset used in that study is not used here is the presence of groundwater interaction in the considered region, which cannot be accounted for by bucket and "force–restore" type models. Daily average precipitation and potential evapotranspiration, derived from measured standard meteorological time series, provide upper boundary conditions for both the multilayer and parametric models. Given identical upper boundary conditions, the results of the hydrologic schemes can be objectively compared with the multilayer model results. The multilayer model is considered a closer representation of real world phenomena since fewer assumptions regarding the physics of soil moisture processes are made. It is not implied, however, that this approach may serve as a surrogate for field experiments.

The formulations of the multilayer model, the two parametric models, and the scaling theory are presented in section 2. Computations have been carried out for all combinations of the three models with three different meteorological time series (dry, moderate, and wet) and two soil types (sand and loam). The results are discussed in section 3 with respect to the following:

1) The effect of soil heterogeneity on the water budget using the multilayer model.
2) The ability of the parameterizations to capture the dynamics of the water budget.
3) The implications for realistic modeling of the lower boundary.
4) The persistence of soil moisture variability and its effect on atmospheric variability.

2. Model formulations

a. Physically based formulation of unsaturated soil moisture movement

A description of all variables and parameters and their units is given in the appendix. The continuity equation for moisture in a one-dimensional soil column can be written as

$$\frac{\partial \theta(z)}{\partial t} = -\frac{\partial q}{\partial z}, \qquad (1)$$

where θ denotes the volumetric (liquid) moisture content, q the unsaturated water flux, t the time, and z the depth (negative downward).

Steady, one-dimensional unsaturated water flow is described by the flow equation of Darcy,

$$q = -k(\psi)\left(\frac{\partial \psi}{\partial z} + 1\right), \qquad (2)$$

where k is the conductivity of the unsaturated medium and ψ the soil matric head, which is the equivalent of the pressure potential on weight basis relative to atmospheric pressure (where $\psi = 0$). In an unsaturated medium, water present in the pores is held against gravity by capillary forces and ψ is negative accordingly. For a particular soil it holds that the smaller the water content is, the smaller the pores the water resides in, the stronger the capillary forces are and the smaller (i.e., more negative) the soil matric head is. This mechanism is expressed by a $\theta - \psi$ relationship or soil water retention curve $\theta(\psi)$. Equation (2) simply states that water flow in an unsaturated medium is due to a gradient in soil matric head and the gravitational gradient $(\partial z/\partial z)$.

Combination of (1) and (2) yields Richards' equation. Addition of a sink term S that describes plant water extraction results in

$$\frac{\partial\theta(z)}{\partial t}=\frac{\partial}{\partial z}\left[k(\psi)\left(\frac{\partial\psi}{\partial z}+1\right)\right]-S(\psi,E_p) \quad (3)$$

with

$$S(\psi,E_p)=\begin{cases}\beta_S(\psi,E_p)\dfrac{E_p}{-z_r}, & z_r\leqslant z\leqslant 0\\ 0, & z<z_r\end{cases}. \quad (4)$$

Equation (4) simply states that over the depth of the root-zone ($-z_r$) roots extract water proportional to the potential evapotranspiration E_p (discussed hereafter) and an evapotranspiration efficiency coefficient β_S. In general, beyond a certain threshold value of ψ, this efficiency coefficient decreases with decreasing soil matric head since the ability of roots to extract water reduces when the water becomes stronger bonded to the soil particles. Beyond the wilting point, water is so strongly bound to the soil that it no longer can be extracted. The above-mentioned behavior is reflected in Fig. 1, after Feddes et al. (1978). Figure 1 also demonstrates that the threshold value in the model is a function of the magnitude of the atmospheric demand (E_p). Bare soil evaporation is not explicitly taken into account, but is implicit in the evapotranspiration efficiency β_S.

The total evapotranspiration out of the soil is obtained through integration of the sink term over the depth of the rootzone

$$E(\psi,E_p)=\int_{z_r}^{0}S(\psi,E_p;z)dz. \quad (5)$$

To solve (3), the soil hydraulic properties $\theta(\psi)$ and $k(\psi)$ and the initial and boundary conditions must be specified. Realistic parametric descriptions of the soil hydraulic properties are given by Van Genuchten (1980)

$$\Theta(\psi)=\frac{\theta(\psi)-\theta_r}{\theta_s-\theta_r}=(1+|a_g\psi|^n)^{-m} \quad (6)$$

$$k(\Theta)=k_s\Theta^l[1-(1-\Theta^{1/m})^m]^2 \Leftrightarrow$$

$$k(\psi)=k_s\frac{(1-|a_g\psi|^{n-1}[1+|a_g\psi|^n]^{-m})^2}{[1+|a_g\psi|^n]^{ml}}. \quad (7)$$

Here Θ indicates the degree of saturation ($0\leqslant\Theta\leqslant 1$), θ_s the saturated water content ($\psi=0$) or porosity, θ_r the residual water content, k_s the saturated conductivity ($\psi=0$), a_g, n, and l are soil specific parameters and $m=1-n^{-1}$. Equations (6) and (7) reflect the fact that water content and, accordingly, conductivity decrease with soil matric head. This is shown for both the sandy and the loamy soil in Fig. 2. Specification of the parameter values will be discussed in section 2d.

Because of the highly nonlinear nature of (3), analytical solutions are available only for cases with severe restrictions on the $k(\psi)$ and $\theta(\psi)$ relations and the initial and boundary conditions. Numerical solutions are less restrictive and are, therefore, employed more often, although computation time can be large. Here (3) is solved numerically using the SWATRE (soil water flow root extraction) model (Belmans et al. 1983) under the following initial and boundary conditions:

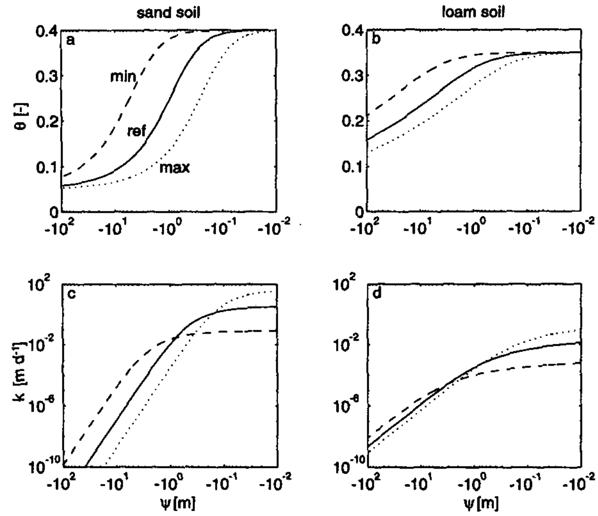

FIG. 2. Hydraulic functions for the sandy soil: water retention (a) and conductivity (c), and for the loamy soil: water retention (b) and hydraulic conductivity (d). Reference curves, $\alpha = 1$ (solid lines), maximum scale factor α (dotted lines) and minimum scale factor (dashed lines).

$$\psi(z)=\psi_0;\quad t=0,\quad z_c\leqslant z\leqslant 0 \quad (8)$$

$$\frac{\partial\psi}{\partial z}=0;\quad t>0,\quad z=z_c \quad (9)$$

$$q=-k(\psi)\left(\frac{\partial\psi}{\partial z}+1\right)=P(t);\quad t>0,\quad z=0 \quad (10)$$

$$E_p=E_p(t);\quad t>0. \quad (11)$$

Condition ψ_0 in (8) implies an initially uniform water content θ_0 in the column of depth $-z_c$. Condition (9) states that at the lower boundary water flow is entirely due to gravity and the percolation flux is equal to the unsaturated conductivity. This corresponds to the lower boundary condition used in the Warrilow model, as is discussed in the next part. Condition (10) represents the infiltration flux boundary. Because of the low rainfall intensities, all precipitation infiltrates and the switching from a prescribed flux to a prescribed (soil matric) head boundary condition does not occur and has therefore been omitted. Condition (11) represents the atmospheric demand at the upper boundary.

The expression of Thom and Oliver (1977) is used to determine E_p:

$$E_p = \frac{\Delta R_n + \rho_a c_p[e_s(T) - e]/r_a}{[\Delta + \gamma(1 + r_s^{\min}/r_a)]\rho_w L_e} \quad (12)$$

with

$$r_a = \frac{4.72}{1 + 0.54u}\left[\ln\left(\frac{z_s}{z_0}\right)\right]^2, \quad (13)$$

where R_n is the net radiation, ρ_a the air density, c_p the dry-air specific heat capacity at constant pressure, e_s the saturated vapor pressure, e the vapor pressure, T the temperature, Δ the slope of the saturation water vapor pressure curve, r_a the aerodynamic resistance, $r_s^{\min}$ the minimal stomatal resistance, γ the psychometric constant, L_e the latent heat of vaporization, u the wind speed, and z_0 the roughness length. All atmospheric variables are evaluated at the screening height z_s, which is 2 m.

When applied over short time periods, (12) may yield values of E_p that could never be maintained by the vegetation, regardless of the availability of soil moisture. Since (12) has been validated for daily values, the various measured meteorological variables have been averaged to yield daily averages. In (12) the soil heat flux is neglected, which on a daily basis is a reasonable assumption.

SWATRE has been used extensively in other studies since its development and has been verified for simulations over longer periods (Hopmans and Stricker 1989). The data of this last study, however, are not suitable for validation of hydrology schemes currently used in climate models since they contain interaction with regional groundwater flow, as mentioned before. This interaction cannot be accounted for, as will be seen hereafter.

b. Formulation of the parametric models

In contrast with the multilayer model, the bucket-type models (Manabe 1969) commonly used in climate modeling do not consider water flow and solely describe the transient behavior of water storage. Two bucket-type models are discussed: the modified version (Noilhan and Planton 1989) of the Deardorff force–restore model (1978) and the model of Warrilow (1986). The two models were selected because of their different lower boundary conditions. These models not only describe phenomena related to the soil water budget, but also of the interfacial layer between soil and atmosphere. The objective of this study, however, is to compare the soil water budget treatment and its possible implications for climate modeling. To isolate the effect of the soil compartment, identical upper boundary conditions (rainfall and potential evapotranspiration) are used in the three models, so the results can be compared in an objective fashion.

1) Warrilow model

In the Warrilow model, the soil is described as a simple reservoir, that can be filled by precipitation and emptied by evapotranspiration and percolation. Unless saturation occurs, all precipitation is assumed to infiltrate.

Evapotranspiration is reduced to less than the potential evapotranspiration rate when the moisture content falls below a critical value. Percolation is described according to the conductivity function of Brooks and Corey (1966), while assuming (9) at the lower boundary. The model can be stated as

$$\frac{d\theta}{dt} = \frac{1}{d}[P - E(\theta) - q_p(\theta)] \quad (14)$$

$$E(\theta) = \beta_W(\theta)E_p = \begin{cases} \left(\dfrac{\theta - \theta_w}{\theta_c - \theta_w}\right)E_p, & \theta < \theta_c \\ E_p, & \theta \geqslant \theta_c \end{cases} \quad (15)$$

$$q_p = k(\theta) = k_s^w\left(\frac{\theta - \theta_w}{\theta_s - \theta_w}\right)^c. \quad (16)$$

As before, θ again denotes the moisture content, θ_c a critical moisture content comparable to the threshold value of ψ in Fig. 1, θ_w the wilting point, P the precipitation, and E the evapotranspiration. The depth of the reservoir is indicated by d and q_p is the percolation flux out of the reservoir.

Because the conductivity functions of Brooks and Corey (16) and van Genuchten (7) differ, k_s^w and k_s in these equations are not the same, as will be explained in section 2d. The boundary conditions, in accordance with (8), (10), and (11), are

$$\theta = \theta_0, \quad t = 0 \quad (17)$$

$$P = \begin{Bmatrix} P(t), \theta < \theta_s \\ 0, \theta = \theta_s \end{Bmatrix} \quad t > 0, \quad z = 0, \quad (18)$$

$$E_p = E_p(t), \quad t > 0, \quad (19)$$

where θ_0 is the initial moisture content, which is given by (6), evaluated at ψ_0 in order to objectively compare the results.

2) Deardorff model

Similar to the Warrilow model, the modified Deardorff model represents the soil as a reservoir. To improve the description near the surface, Deardorff (1978) embedded a top layer, with a depth equal to the depth to which the diurnal soil moisture fluctuation extends (d_1 = 10 cm):

$$\frac{d\theta_t}{dt} = \frac{C_1(\theta_t)}{d_1}\left[P - \frac{1}{10}E(\theta_e)\right] - \frac{C_2(\theta)}{\tau}(\theta_t - \theta) \quad (20)$$

$$\frac{d\theta}{dt} = \frac{1}{d}[P - E(\theta_e)], \tag{21}$$

where

$$C_1(\theta_t) = C_1^{\mathrm{sat}}\left(\frac{\theta_s}{\theta_t}\right)^{(b/2)+1}, \quad C_2(\theta) = C_2^{\mathrm{ref}}\left(\frac{\theta}{\theta_s - \theta + \theta_l}\right) \tag{22}$$

$$\theta_e = \frac{1}{10}\theta_t + \frac{9}{10}\theta. \tag{23}$$

Here θ_t is the volumetric moisture content of the embedded top layer, θ_e an effective moisture content for evapotranspiration, θ_l a small dummy value to prevent a zero division in case the reservoir saturates, and C_1^{sat} and C_2^{ref} are empirical constants.

The first term on the right-hand side of (20) can be considered as a force term. The second term characterizes the diffusivity of water in the soil (restore term). Since in this study surface runoff is not present because of the low enough average daily rainfall intensities, the effect of the embedded top layer is minimal. For the SWATRE and the Warrilow model, evapotranspiration can, respectively, be written in the form of $E = \beta E_p$. In the Deardorff model on the other hand, evapotranspiration is controlled by an aerodynamic and a stomatal resistance, the latter dependent on the soil moisture status

$$E(\theta_e) = \frac{r_a}{r_a + r_s(\theta_e)} E_w \tag{24}$$

with r_a according to (13) and

$$r_s(\theta_e) = \max\left[200\left(\frac{\theta_w}{\theta_e}\right)^2, r_s^{\min}\right] \tag{25}$$

$$E_w = \frac{r_s^{\min} + r_a}{r_a} E_p, \tag{26}$$

where E_w is the potential evapotranspiration of a wet vegetation ($r_s = 0$) and $r_s^{\min}$ the minimum stomatal resistance of dry, well watered vegetation. From (24) and (26) it is clear that when the evapotranspiration is climate controlled ($r_s = r_s^{\min}$), the actual evapotranspiration is equal to E_p according to (12), satisfying the requirement of identical boundary conditions. In (20) the fraction of total evapotranspiration extracted from the embedded top layer represents bare soil evaporation.

Equations (20) and (21) are solved subject to the initial and boundary conditions (17)–(19). The initial value of θ and θ_t are set equal to θ_0.

3) MEANING OF SOIL MOISTURE IN THE MODELS

The main difference between the multilayer model and the parametric models discussed above is that the last two are based only on the continuity equations (14) and (21), whereas the first model includes flow inside the model domain. In the parametric models, the soil moisture state greatly depends on the specification of the reservoir depth d, which specifies the amount of water available for the various fluxes, but also the response of the system to external forcing. Therefore, the soil moisture state serves as a manipulable variable and comparison with the soil moisture state of the multilayer model or even validation using measured data does not make sense. Hence, any comparison or validation must be solely on the percolation and evapotranspiration fluxes.

c. Stochastic treatment of spatial heterogeneity of soil hydraulic properties

Besides treating the physics of soil moisture processes in a simplified way, hydrologic schemes in climate models generally do not account for spatial variation of the soil properties they use. They rather assume effective values for the entire large-area grid. Sharma and Luxmoore (1979) show the effect spatially variable soil hydraulic properties may have on the water budget. To account for spatially variable soil physical properties in the multilayer model, a stochastic approach is used whereby random fields of soil hydraulic properties are constructed within a certain soil type.

The stochastic approach is based on the similar media concept of Miller and Miller (1956), who derived scaling relations for geometrically similar soils. Soils are considered geometrically similar when they differ only with respect to their internal length scale and are therefore of equal porosity. This is illustrated in Fig. 3. Since soils with larger internal length scales have larger pores, it is intuitive that they are less able to capillary bind water. On the other hand, the conductivity in case of complete saturation is higher. It was proved by Miller and Miller (1956) that for two similar soils with identical moisture contents, the soil matric head scales inversely proportional and the conductivity quadratically with the ratio of the two length scales, α

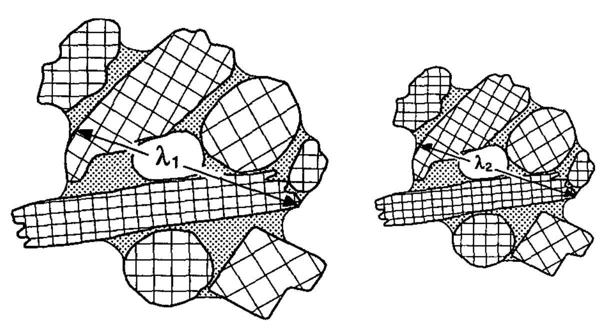

FIG. 3. Similar media (after Miller and Miller 1956). Here λ indicates the internal length scale of the soil.

TABLE 1. Cumulative amounts of precipitation, potential evapotranspiration, and model simulated actual transpiration over the growing season (days 91–273) for the different combinations of soil properties and meterological boundary conditions. The unit for all is millimeters.

				SWATRE			Deardorff	
Soil	Year	P	E_p	Ref.	Stoch.	Warrilow	No. perc.	Perc.
Sand	1976	197	504	222	230	234	272	248
	1978	337	373	300	319	327	373	346
	1982	244	444	267	285	279	405	295
Loam	1976	197	504	247	241	232	266	264
	1978	337	373	341	339	318	356	342
	1982	244	444	307	304	281	329	313

$$\psi_i = \alpha_i^{-1} \psi_{\text{ref}} \tag{27}$$

$$k_i = \alpha_i^2 k_{\text{ref}}. \tag{28}$$

Here the subscript "ref" indicates a reference soil and α_i the ratio of the internal length scale for soil i and the length scale of the reference soil ($\alpha_i = \lambda_i \lambda_{\text{ref}}^{-1}$).

The above-mentioned behavior is reflected in Figs. 2a–d, in which the range of soil hydraulic properties used in this study are depicted. The conductivity at saturation ($\psi = 0$) is largest for the maximum scale factor, but decreases more rapidly with increasing absolute soil matric head. This last phenomenon is due to the smaller capillary tension that can be maintained, leaving a smaller fraction of the soil available for flow.

In general, however, soil types or soil samples are not strictly geometrically similar since they do not obey the requirement of identical values of θ_s. Therefore, departing from the original theory, scaling has been performed using the degree of saturation (6). Many studies have shown the ability of this method to characterize spatial variability of soil hydraulic properties (e.g., Warrick et al. 1977; Hopmans 1987).

By virtue of (6) and (7) and the scaling relations (27) and (28), it can be derived that

$$a_{g,i} = \alpha_i a_{g,\text{ref}} \tag{29}$$

$$k_{s,i} = \alpha_i^2 k_{s,\text{ref}}. \tag{30}$$

With known values $a_{g,\text{ref}}$, $k_{s,\text{ref}}$, n, and l, random fields of soil hydraulic properties can be constructed through sampling of the probability density functions of α and θ_s, respectively, $f(\alpha)$ and $g(\theta_s)$.

Assuming vertical uniform, spatially independent soil hydraulic characteristics, no lateral soil moisture flow and independence of the density functions, the areally mean actual evapotranspiration is obtained as

$$\mathrm{E}[E(t)] = \frac{1}{N} \sum_{i=1}^{N} E_i(\alpha_i, \theta_{s,i}; t), \tag{31}$$

where $E(\alpha_i, \theta_{s,i}; t)$ is computed by the multilayer model and N is the number of realizations within the random field. It appeared that for $N > 30$ the mean was sufficiently determined. Using the technique described by Clausnitzer et al. (1992), the (scaled average) reference soil water retention and hydraulic conductivity functions and the set of corresponding scale factors have been determined for two datasets of measured hydraulic properties. From literature it is known that values of α and θ_s within a soil type are, respectively, lognormally and normally distributed (e.g., Warrick et al. 1977; Hopmans and Stricker 1989), which indeed is the case for the soils used in this study.

d. Specification of parameters and meteorological time series

The conditions considered here assume an areally uniform grass vegetation that covers the soil completely. The value for $r_s^{\min}$ in (12) is set to 65 s m^{-1}, a value often used for grass. The total depth for the multilayer model, z_c, is taken as 100 cm, divided into 25 layers, increasing in thickness with depth. The depth of the root zone, $-z_r$, is 30 cm, consistent with the observed rooting depth of grass in the Dutch area where the meteorological time series were collected. Data of the years 1976, 1978, and 1982, respectively, dry, wet, and moderate provide upper boundary conditions. Daily average rainfall intensities are low enough to prevent surface runoff for all models. The cumulative values of precipitation and potential evapotranspiration for the time series are given in Table 1.

The soil hydraulic properties used in this study originate from samples taken near Lubbon (sand) and Castelnau (loam) in the HAPEX–MOBILHY study area (André et al. 1986). For each sample, the multistep outflow method (van Dam et al. 1994) was used for determining the soil hydraulic parameters in (6) and (7). Here θ_s was measured separately and $g(\theta_s)$ was determined using the measured values. Applying the scaling technique mentioned before, the values $a_{g,\text{ref}}$, $k_{s,\text{ref}}$, n, l and the probability density function $f(\alpha)$ were determined. Given the small variation of θ_r values between the samples, the average value of θ_r has been

TABLE 2. Van Genuchten parameters for the reference soil.

Soil	$a_{g,\mathrm{ref}}$ (m^{-1})	n	$k_{s,\mathrm{ref}}$ (m day^{-1})	l	θ_r
Sand	1.90	1.72	3.75	2.5	0.05
Loam	1.03	1.18	0.043	2.5	0.01

TABLE 3. Mean and standard deviation of the distribution functions of ln(α) and θ_s.

	ln(α)		θ_s	
Soil	μ	σ	μ	σ
Sand	−0.31	0.80	0.40	0.07
Loam	−0.28	0.72	0.35	0.05

used. The respective parameter values are given in Table 2. The means and standard deviations of the probability density functions $f(\alpha)$ and $g(\theta_s)$ are summarized in Table 3. In Fig. 2, the curves corresponding to the reference values of the parameters ($\alpha = 1$) and the mean values of θ_s are depicted for both soils. To prevent unrealistic soil hydraulic properties, the distribution functions were first truncated and subsequently normalized such that all probability mass lies inside the finite interval $\mu_x - 2\sigma_x \leqslant x \leqslant \mu_x + 2\sigma_x$, where $x \in \{\alpha, \theta_s\}$. The effect of the scale factor on the $k(\psi)$ and $\theta(\psi)$ curves is indicated in Fig. 2 for the minimum and maximum value of possible α values. The initial conditions were set to $\psi_0 = -2.5$ m in accordance with the observed profiles of soil matric head in the area where the meteorological measurements were taken.

The parameter values used in the parametric models are derived from the reference soil hydraulic characteristics. The values of θ_c and θ_w in (15), (16), and (25) were determined using (6) evaluated at, respectively, $\psi = -5.0$ m and $\psi = -80$ m. This coincides with the ψ values used in the β_S function (Fig. 1). Values for k_s^w and c in (16) are derived through optimization from (7) in the range where percolation is important (near saturation). Calibration of the constants C_1^{sat} and C_2^{ref} in (22) is performed according to the method described by Noilhan and Planton (1989). The constant b was determined according to Clapp and Hornberger (1978). The depth of the soil reservoir for both parametric models was set at 50 cm. Purposely, this depth was not taken the same as the depth of the rootzone used in the multilayer model. Deardorff (1978) explicitly suggests a value of 50 cm and Noilhan and Planton (1989) indicate that d is deeper than the depth of the root zone, which is 30 cm as mentioned previously. Warrilow suggests a slightly deeper reservoir; however, for objective comparison with the Deardorff model the same value was chosen. The parameter values of the Warrilow and the Deardorff model are specified in Table 4.

3. Results and discussion

Computations have been carried out for the three models, the three micrometeorological time series, and the two soil types. As the soil moisture state affects evapotranspiration most severely during the growing season (days 91–273), the comparison of the models is mainly performed on evapotranspiration efficiencies during the growing season.

a. *Effect of soil heterogeneity*

The effect of heterogeneity on the water budget is analyzed by comparison of the deterministic (reference soil) and the average stochastic computations (31) obtained with SWATRE. Table 1 shows that the cumulative evapotranspiration for the reference sandy soil is slightly smaller than for the stochastic computations and that for the loamy soil the results compare very well. The graphs in Fig. 4 show the day-to-day course of the efficiency factor E/E_p for the reference soil and the averaged results of the stochastic computations. The temporal behavior of E/E_p directly reflects the influence of the soil moisture status on evapotranspiration. In the range where the percolation process is important in the water budget (near saturation), the $k(\psi)$ values of the loamy soil are orders of magnitude smaller than of the sandy soil (Fig. 2). Since percolation depends on the hydraulic conductivity, soil spatial heterogeneity has a much greater effect for the sandy soil than for the loamy soil. This induces larger differences in soil water availability between the realizations for sandy soil. Hence, the differences between the E/E_p values for the reference and the stochastic application of SWATRE are more evident for the sandy soil than for the loamy soil, as can be seen in Figs. 4a–f. From the results it can be concluded that the reference soil hydraulic properties describe the average water budget of the sto-

TABLE 4. Parameter values for the Warrilow and Deardorff model.

	Warrilow			Deardorff				Both	
Soil	θ_c	k_s^w (m day^{-1})	c	C_1^{sat}	C_2^{ref}	b	d_r (m)	θ_w	d (m)
Sand	0.13	1.82	8.0	0.2	6.0	3.0	0.1	0.06	0.5
Loam	0.25	0.01	14.0	0.4	0.8	6.0	0.1	0.15	0.5

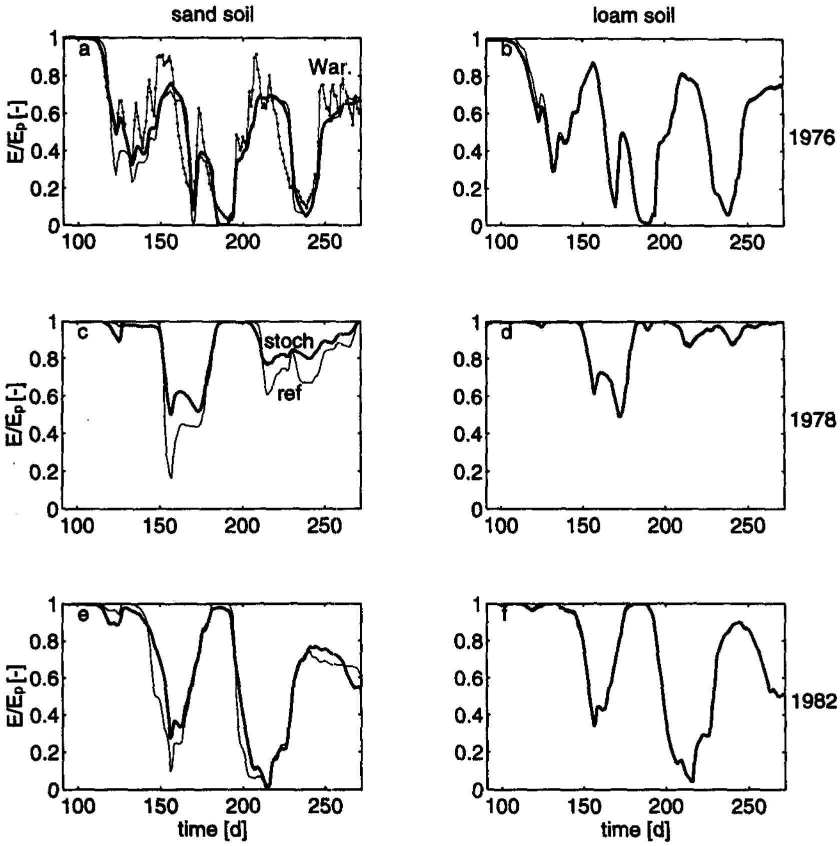

FIG. 4. Mean evapotranspiration efficiency (E/E_p) versus time for the mean stochastic results (thick lines) and the reference curves (thin lines) during the growing season. (a) Sand soil, 1976, (b) loam soil, 1976, (c) sand soil, 1978, (d) loam soil, 1978, (e) sand soil, 1982, and (f) loam soil, 1982. Line with dots in (a) indicates Warrilow model.

chastic problem very well for the loamy and reasonable well for the sandy soil. Given the degree of spatial heterogeneity of the soil properties, the fact that the (scaled average) reference properties are so effective in describing the transient spatially averaged water budget in the absence of runoff is remarkable. For the sandy soil, these results are confirmed by Kim and Stricker (1996) using a half-hourly meteorological forcing derived from a slightly different climate.

b. Comparison between the parametric models and the multilayer model

Comparison of the averaged stochastic results from the previous section with the results of the parametric models indicates the ability of the parametric models to describe the soil water budget of an inhomogeneous region. Any disagreement between the averaged stochastic results and the results of the parametric models should be attributed more to the different treatment of the physics rather than to a neglect of spatial variability of the various parameters. This because of the good agreement between the deterministic and averaged stochastic results in the previous section and the fact that the reference soils were used for the specification of the parametric models. As stated before, comparison of the soil moisture state between the multilayer and the parametric models does not make sense. Similar to the previous section the evapotranspiration efficiencies are compared since these essentially capture the influence of the soil moisture state on the fluxes.

In Table 1 the values of the cumulative evapotranspiration are given for the three growing seasons. The Warrilow model shows only small differences with the stochastic SWATRE results for the sandy soil and somewhat more for the loamy soil. Regarding the daily evapotranspiration in Figs. 5a,b, however, it appears that differences are considerable, especially for the sandy soil. Scatter increases with higher values of E/E_p. The different temporal behavior of the two models is indicated in Fig. 4a for the sandy soil and the meteorological conditions of 1976 and is explained hereafter. The Warrilow model behaves more erratically than the multilayer model, indicating its faster response

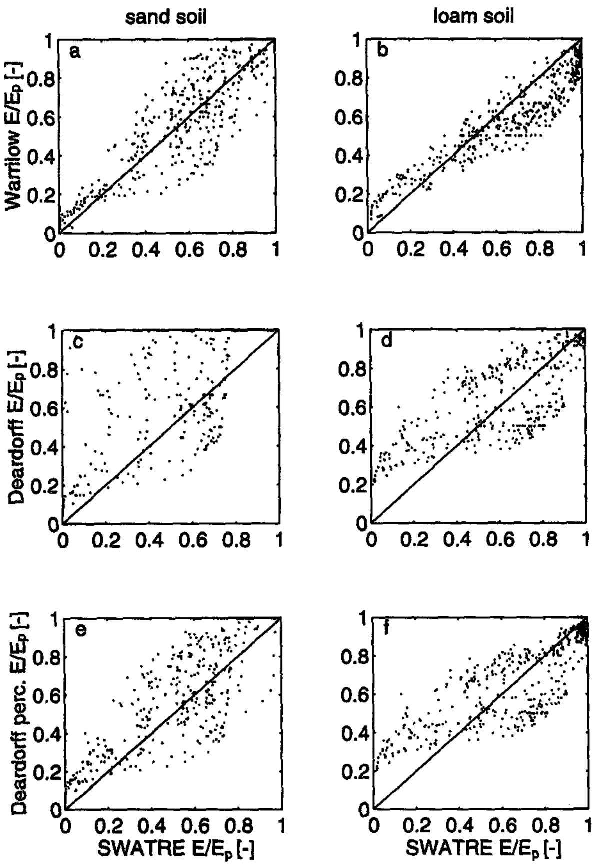

FIG. 5. Scatter diagrams of the evapotranspiration efficiency (E/E_p) of the parametric models versus the SWATRE mean stochastic results for the three years 1976, 1978, and 1982. (a) Warrilow sandy soil ($\Re$ = 0.18), (b) Warrilow loam soil ($\Re$ = 0.15), (c) Deardorff sand soil ($\Re$ = 0.40), (d) Deardorff loam soil ($\Re$ = 0.21), (e) Deardorff with percolation term sand soil ($\Re$ = 0.21), (f) Deardorff with percolation term loam soil ($\Re$ = 0.20).

to changing boundary conditions. This can largely be explained by the transient behavior of the flow processes in the multilayer model in contrast with the instantaneous redistribution of moisture in the Warrilow model. For instance, in the Warrilow model, precipitation is redistributed directly over the entire column whereas in the multilayer model a transient wetting front propagates through the profile. Another example of the smooth behavior of the multilayer model with respect to time is the incorporation of upward capillary flow that prevents an abrupt falling of the ratio E/E_p with time, as can be clearly seen in Fig. 4a.

A good measure to express the differences of daily values between the two models is the root-mean-square error $\Re$. Values of 0.18 and 0.15 were found for, respectively, sand and loam. After averaging of the ratio E/E_p over 10 days these values decrease only to 0.15 and 0.13, indicating that the scatter due to the absence of soil moisture flow in the Warrilow model does not largely average out over a 10-day period. This would result in a persistent difference in the partitioning of available energy between the two models.

For the sandy soil, the cumulative evapotranspiration values of the Deardorff model compare poorly with those of the stochastic multilayer model. The Deardorff model overestimates evapotranspiration greatly (see Table 1 and Fig. 5c), indicating a systematically too large water storage. This can almost entirely be explained by the absence of a percolation term in the parameterization. For the loamy soil, this effect is not present since percolation is less important and a better result is obtained (Fig. 5d). However, especially for the loamy soil, the Deardorff model gives higher efficiencies for dry conditions than the multilayer model. This is caused by the different methods both models employ for the reduction of potential transpiration: (4) and (5) for the multilayer model, and (24) and (25) for the Deardorff model. In the latter case, evapotranspiration is still possible at values below θ_w, where more moisture is available for transpiration for the loamy soil than for the sandy soil (Fig. 2b).

After addition of a percolation term similar to (16), for the sand soil, the Deardorff and the Warrilow models qualitatively show the same scatter in efficiency values (see Figs. 5a,e). As for the Warrilow model, averaging of the ratio E/E_p over 10-day periods only slightly reduces the scatter ($\Re$ values of 0.19 and 0.18 for the sand and loam soil, respectively). The improved description near the surface in the Deardorff model has hardly any effect in this study because surface runoff does not occur.

c. *Implications for modeling the lower boundary*

Given the sensitivity of the water budget to the lower boundary condition, the question arises what condition is realistic for use in GCMs. For one-dimensional parameterizations, a no-flow lower boundary can be regarded as the lower limit. Since the conditions below the reservoir are generally unknown, condition (9), which indicates vertical flow due to gravity only, may be considered realistic assuming the water table resides deep. However, water lost through percolation at topographically high positions may reappear in the surface layer at topographically lower positions because of saturated lateral flow. The areally averaged percolation is in that case smaller than one-dimensional gravity-driven percolation. The Deardorff and the Warrilow model can, therefore, be regarded as the two limiting cases and a realistic description of the lower boundary must be in between these two.

As an example, the Simple Biosphere Model (Sellers et al. 1986) uses as lower boundary condition an expression similar to (16) multiplied by $\sin\omega$, where ω is the terrain slope angle. Strictly interpreted, this term represents lateral unsaturated gravity-driven flow. Note that for completely flat terrain ($\omega = 0$) it becomes a no-flow boundary similar to the Deardorff model and for a completely tilted terrain ($\omega = \pi$) it becomes identical to the Warrilow lower boundary. The parameter ω can thus be interpreted as a parameterization of the areally effective lower boundary condition and thus reflects the effect of lateral redistribution of soil water. In short, defining a realistic lower boundary condition for parameterizations to be used in GCMs essentially means that two- or three-dimensional flow processes need to be incorporated in a one-dimensional model and the need for further research is evident.

d. Persistence of soil moisture anomalies

Delworth and Manabe (1988) used a closed bottom bucket representation of the land surface to investigate the persistence of soil wetness. They state that the soil layer acts as an integrator of precipitation forcing: short timescale precipitation with a white noise spectrum is transformed into a red response spectrum with most of its variance concentrated in the lower frequencies. Accordingly, since the soil moisture state greatly influences the partitioning of available energy over the surface heat fluxes, low-frequency soil wetness variability generates low-frequency atmospheric variability. It is therefore of interest to elaborate on how the results presented before affect the decay of soil moisture anomalies—that is, deviations from the long-term mean state.

In a closed bucket a positive anomaly can be erased only by evapotranspiration. Inclusion of percolation will therefore shorten the timescale of soil moisture anomalies, especially for conductive soils. Delworth and Manabe (1989) note that soil moisture fluctuations in regions or seasons with a small value of the ratio of potential evapotranspiration over precipitation (E_p/P) hardly affect atmospheric variability since the actual evapotranspiration is almost always potential. In regions or seasons with very high values of E_p/P, the persistence of soil moisture anomalies, and consequently anomalies of the surface heat fluxes, is low. Inclusion of percolation will decrease the persistence of soil wetness and atmospheric variability. For regions or seasons with small values of E_p/P this implicates that the influence of the soil water state on atmospheric conditions will increase because evapotranspiration will be "soil controlled" more frequently.

The improvement of the Deardorff model after insertion of a percolation term may serve as an illustration of the smaller persistence of a wet anomaly due to percolation, and the resulting decrease of evapotranspiration due to increased soil control. For the sandy soil and the meteorological conditions of 1982, the difference between the cumulative evapotranspiration values of the Deardorff model with and without percolation term (110 mm, Table 1) is almost entirely balanced by the cumulative amount of percolation prior to the start of the growing season (96 mm). The period prior to the growing season (day 1 to 90) has an E_p/P ratio of 0.33 and evapotranspiration cannot balance the precipitation. During the growing season, with its E_p/P ratio of 1.82 (Table 1), the influence of percolation is much smaller (14 mm).

4. Conclusions

The reduction of the evapotranspiration rate to less than the potential rate, as predicted by the multilayer SWATRE model (Belmans et al. 1983), the adapted Deardorff model (Noilhan and Planton 1989), and the Warrilow model (1986) were compared, using sandy and loamy soil types and three different years of daily average micrometeorological data. Surface runoff did not occur because of the low precipitation intensities, resulting from the daily averaging. The following conclusions are drawn:

1) The scaled mean soil hydraulic properties describe the average water budget and the reduction of evapotranspiration of the stochastic problem very well for the loamy soil and reasonable well for the sandy soil. Differences between the stochastic results of the multilayer model and the results of the parametric models can, therefore, mainly be attributed to the different parameterization of the physics.

2) Since the Deardorff model neglects percolation, reduction of evapotranspiration is systematically underestimated compared with both the multilayer model and the Warrilow model. Insertion of a percolation term results in a model behavior similar to the Warrilow model. For realistic situations, the lower boundary conditions used in the two parametric models may be regarded as the lower and upper limiting case, respectively.

3) Because flow of soil moisture is not considered, both the Deardorff and the Warrilow models show large variations in daily and 10-day average evapotranspiration efficiency when compared to the multilayer model. Efficiencies over longer time periods (up to months) show of course less variation. This leaves an open question to atmospheric modelers about the timescale over which the water budget, and by that the partition of energy, must be consistent and realistic.

4) Validation of the results using measurements could not be performed. Nevertheless, the multilayer model has proven its value in simulating observed phenomena. Therefore, the possible effects of the assumptions regarding the physics and spatial variation of soil properties frequently made in hydrological schemes for use in climate models can be addressed. Nevertheless, the need for long-term monitoring of soil moisture and water budget components and collection of information about spatially variable surface properties is emphasized.

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APPENDIX

List of Symbols

a_g (m^{-1})	shape parameter
b	empirical constant
C_1	empirical constant
C_1^{sat}	empirical constant
C_2	empirical constant
C_2^{ref}	empirical constant
c	empirical constant
c_p (J kg^{-1} K^{-1})	dry-air specific heat capacity
d (m)	depth of reservoir
d_1 (m)	depth of top layer
E	expectation operator
E (m s^{-1})	evapotranspiration
E_p (m s^{-1})	potential evapotranspiration
E_w (m s^{-1})	wet crop transpiration
e (mb)	vapor pressure (at z_s)
e_s (mb)	saturated vapor pressure
$f(\alpha)$	probability density function α
$g(\theta_s)$	probability density function θ_s
k (m s^{-1})	conductivity
k_s (m s^{-1})	saturated conductivity
k_s^w (m s^{-1})	saturated conductivity Warrilow
L_e (J kg^{-1})	heat of vaporization
l	shape parameter
m	$1 - n^{-1}$
N	number of Monte Carlo realizations
n	shape parameter
P (m s^{-1})	precipitation
q_p (m s^{-1})	percolation
q (m s^{-1})	unsaturated water flux
R_n (J m^{-2} s^{-1})	net radiation
r_a (s m^{-1})	aerodynamic resistance
r_s (s m^{-1})	stomatal resistance
r_s^{min} (s m^{-1})	minimal stomatal resistance
S (s^{-1})	root extraction
T (K)	temperature (at z_s)
t (s)	time
u (m s^{-1})	wind speed (at z_s)
z (m)	depth (negative downward)
z_c (m)	depth of the column
z_r (m)	depth of the root zone
z_s (m)	screening height (2 m)
z_0 (m)	roughness length
α	scale factor
β_S	transpiration efficiency SWATRE
β_W	transpiration efficiency Warrilow
γ (mb K^{-1})	psychrometric constant
Δ (mb K^{-1})	de_s/dT
Θ	degree of saturation
θ (m^3 m^{-3})	volumetric moisture content
θ_c (m^3 m^{-3})	critical moisture content
θ_e (m^3 m^{-3})	effective moisture content
θ_l (m^3 m^{-3})	small value to prevent zero division
θ_r (m^3 m^{-3})	residual moisture content
θ_s (m^3 m^{-3})	saturated moisture content
θ_t (m^3 m^{-3})	moisture content top layer
θ_w (m^3 m^{-3})	wilting point moisture content
θ_0 (m^3 m^{-3})	initial volumetric moisture content
λ (m)	soil internal length scale
ρ_a (kg m^{-3})	air density
ρ_w (kg m^{-3})	water density
τ (s)	diurnal period
ψ (m)	soil matric head
ψ_0 (m)	initial soil matric head
$\Re$	root-mean-square error

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