Inclusion of Gravitational Drainage in a Land Surface Scheme Based on the Force–Restore Method

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ABSTRACT

This paper presents a simple parameterization of gravitational drainage for land surface schemes describing soil water transfers according to the force–restore method of Deardorff. A one-year time series of observed soil moisture period from HAPEX–MOBILHY (Hydrological Atmospheric Pilot Experiment–Mobilisation du Bilan Hydrique) 1986 revealed the importance of subsurface drainage during the wintertime period. This physical process is accounted for through a Newtonian restore to field capacity when soil moisture is above it. Simulation of the annual cycle of soil moisture by the land surface scheme ISBA (interactions soil biosphere atmosphere) is in this way greatly improved.

1. Introduction

The need for accurate boundary conditions over continental surfaces in atmospheric numerical models is now recognized, both for climate studies and numerical weather prediction (Rowntree 1991; Blondin 1991). Land surface schemes have been greatly improved during the last ten years. The influence of vegetation on surface evaporation is now taken into account by many schemes (Deardorff 1978; Dickinson 1984; Sellers et al. 1986; Abramopoulos et al. 1988). The concept of canopy surface resistance (Monteith 1965) is generally adopted for estimating the effect of plant control on transpiration. These improvements are mostly the result of the availability of new datasets provided by Hydrological Atmospheric Pilot Experiment (HAPEX)-type experiments {HAPEX–MOBILHY [Hydrological Atmospheric Pilot Experiment] 86, FIFE [First Field Experiment] 87, EFEDA [ECHIVAL (European International Project on Climate and Hydrological Interactions between Vegetation, Atmosphere, and Land Surface) Field Experiment in Desertification-Threatened Area] 91, HAPEX–SAHEL 92, BOREAS [Boreal Ecosystem–Atmosphere Study] 94}, which allowed land surface schemes to be checked and calibrated. These field experiments measured various components of the continental hydrological cycle during IOPs (intensive observing periods) over areas compatible with the grid scale of general circulation models. For example, the land surface scheme ISBA (interactions soil biosphere atmosphere) developed by Noilhan and Planton (1989) has been improved from its initial version to better represent forest transpiration (Jacquemin and Noilhan 1990), bare soil evaporation (Mahfouf and Noilhan 1991; Braud et al. 1993), and canopy interception (Manzi and Planton 1994; Mahfouf et al. 1995).

An evaluation of water transfers within the soil is more difficult since it requires data covering longer periods of time (at least 1 year) to obtain significant variations of soil moisture within the layer that interacts directly with the atmosphere. The way to represent them in land surface schemes consists of a discretization of the Richards diffusion equation. At least four layers are required to minimize truncation errors (Mahrt and Pan 1984; Viterbo and Beljaars 1995). Deardorff (1977) proposed a simpler set of equations, taking into account the forcing at the interface (over short timescales), involving the first centimeters of soil and the restore toward deep soil values evolving more slowly with time. This approach, known as the force–restore method, has been generalized by Noilhan and Planton (1989) in order to account for the dependence of transfers on soil texture and moisture.

Mahfouf (1990) proposed a first evaluation of the surface water budget simulated by ISBA during the IOP from HAPEX–MOBILHY (May–July 1986). He demonstrated that when ISBA is forced by observed precipitation and radiation, it produces a correct evolution of soil moisture and surface evapotranspiration.

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TABLE 1. The soil types of the Clapp and Hornberger classification and their main characteristics. The coefficient C_3 is estimated from Eq. (15) with a soil depth of $d_2 = 1$ m.

Soil type	X_{clay} (%)	w_{sat} ($\text{m}^3\ \text{m}^{-3}$)	K_{sat} $10^{-6}\ \text{m s}^{-1}$	w_{fc} ($\text{m}^3\ \text{m}^{-3}$)	b	C_3
Sand	3.0	0.395	176.0	0.135	4.05	1.705
Loamy sand	6.0	0.410	156.3	0.150	4.38	1.437
Sandy loam	9.0	0.435	34.1	0.195	4.90	0.510
Silt loam	14.0	0.485	7.2	0.255	5.30	0.183
Loam	19.0	0.451	7.0	0.240	5.39	0.196
Sandy clay loam	28.0	0.420	6.3	0.255	7.12	0.206
Silty clay loam	34.0	0.477	1.7	0.322	7.75	0.098
Clay loam	34.0	0.476	2.5	0.325	8.52	0.124
Sandy clay	43.0	0.426	2.2	0.310	10.4	0.139
Silty clay	49.0	0.482	1.0	0.370	10.4	0.112
Clay	63.0	0.482	1.3	0.367	11.4	0.102

Longer times series are needed to evaluate more completely the way in which surface water budget is simulated by land surface schemes. To our knowledge, a dataset including measured soil moisture, precipitation, and surface evaporation does not exist for a 1-yr period. An accurate management of soil moisture evolution is particularly important for climate modeling purposes since models are integrated during several years. This aspect is a priori less important for weather prediction, where the initial state dominates. However, soil moisture is a variable difficult to initialize in operational mode (Mahfouf 1991). A land surface scheme that has a poor physical basis can drift rapidly if not constrained by observations.

One important issue of the long-term management of soil moisture concerns the water fluxes at the bottom of the soil layer. This boundary condition can be handled quite simply in multilayer models either by assuming no flux at the bottom (bedrock) or by setting the vertical gradient of matrix water potential to zero (free drainage condition) (Abramopoulos et al. 1988; Viterbo and Beljaars 1995). A few bulk models (one or two layers) attempt to represent the gravitational drainage. For instance, Warrilow et al. (1986) impose the downward water flux at the bottom of soil layer to hydraulic conductivity. However, this assumption is only valid when soil moisture is close to saturation. Other bulk schemes assume no water flux at the bottom (Deardorff 1977; Noilhan and Planton 1989).

In this paper, we investigate the parametrization of gravitational drainage using the land surface scheme ISBA and a 1-yr time series of soil moisture taken from the HAPEX–MOBILHY database. In section 2, we present the land surface scheme ISBA and the way it has been modified to account for subsurface gravitational drainage. Section 3 describes the dataset used for the evaluation of the scheme and analyzes modeling results.

2. Parameterization of the gravitational drainage

The ISBA scheme predicts the evolution of five prognostic variables for the surface temperature T_s, the daily mean surface temperature T_2, the surface volumetric water content w_g, the bulk volumetric water content w_2, and the water content of an interception reservoir W_r:

$$\frac{\partial T_s}{\partial t} = C_T(R_n - H - LE) - \frac{2\pi}{\tau}(T_s - T_2), \quad (1)$$

$$\frac{\partial T_2}{\partial t} = \frac{1}{\tau}(T_s - T_2), \quad (2)$$

$$\frac{\partial w_s}{\partial t} = \frac{C_1}{\rho_w d_1}(P_g - E_g) - \frac{C_2}{\tau}(w_g - w_{g\text{eq}}), \quad 0 \leqslant w_g \leqslant w_{\text{sat}}, \quad (3)$$

$$\frac{\partial w_2}{\partial t} = \frac{1}{\rho_w d_2}(P_g - E_g - E_{tr}) - \frac{C_3}{\tau}\max[0, (w_2 - w_{\text{fc}})], \quad 0 \leqslant w_2 \leqslant w_{\text{sat}}, \quad (4)$$

and

$$\frac{\partial W_r}{\partial t} = \text{veg}P - E_r - R_r. \quad (5)$$

A list of symbols for this set of equations is given in the appendix. The soil has a depth d_2 including the root zone. Soil water content is characterized by three threshold values. The saturation water content (or porosity) w_{sat} is the maximum amount of water that a given soil can hold. The field capacity w_{fc} is the soil water content at which gravitational drainage effectively ceases. The wilting point w_{wilt} is the soil water content below which it is assumed that plants are unable to pump water from the root zone to stomatal cells. These three quantities depend upon soil texture. In the ISBA scheme, w_{sat} is estimated from Clapp and Hornberger (1978), w_{fc} corresponds to an hydraulic conductivity of 0.1 mm day^{-1}, and w_{wilt} corresponds to a water potential of -15 b. Evaporation is assumed to

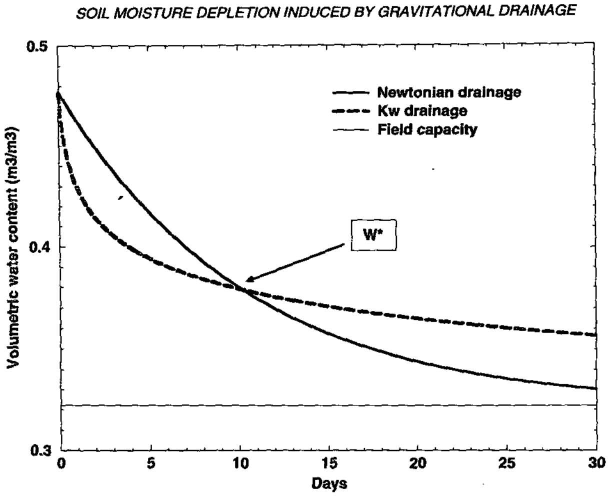

FIG. 1. Time evolution of the bulk soil water content as a result of gravitational drainage for silty clay loam. Starting from saturation, the soil moisture depletion if computed from Eq. (13) (solid line) and from Eq. (14) (dotted line). The *e*-folding time given by the intercept of the two curves (defining the value of w^*) is about 10 days. The horizontal line indicates the value of the field capacity.

take place at the potential rate when soil moisture is between $w_{\rm fc}$ and $w_{\rm sat}$, while transpiration ceases below $w_{\rm wilt}$. Thus, the water holding capacity of the soil layer expressed in meters is $d_2(w_{\rm fc} - w_{\rm wilt})$.

For short-term integrations, Noilhan and Planton (1989) have assumed no water flux at the lower boundary of the soil layer. In this article, a revision of this model in which deep drainage by gravity is added for annual timescales is described. Since water content at field capacity $w_{\rm fc}$ defines qualitatively the influence of gravitational drainage, its seems reasonable to include a loss of water at the bottom of the soil layer as long, as the mean water content w_2 is above field capacity. The C_3 coefficient in Eq. (4) characterizes the rate at which the water profile is restored to the field capacity. This quantity must depend upon soil texture to account for the increase in water transfer as soil texture becomes coarser.

In the forthcoming discussion, we assume that the sink of water over a soil column d_2 is produced only by gravitational drainage at the bottom of the layer. If the evolution of w_2 is given by a Newtonian restore term, it can be written as

$$\frac{\partial w_2}{\partial t} = -C_3 \frac{w_2 - w_{\rm fc}}{\tau}. \qquad (6)$$

We then want to derive C_3 from soil textural properties. The evolution of the local volumetric water content η, according to the Richards diffusion equation, is:

$$\rho_w \frac{\partial \eta}{\partial t} = -\frac{\partial W_s}{\partial z}, \qquad (7)$$

where $W_s(z)$ is the water flux at depth z. Integrating Eq. (7) from $z = 0$ to $z = d_2$ leads to

$$\rho_w \int_0^{d_2} \frac{\partial \eta}{\partial z}(z)dz = W_s(z = 0) - W_s(z = d_2). \qquad (8)$$

A zero flux is assumed at the surface $[W_s(z) = 0]$, while $W_s(z = d_2)$ is given by Darcy's law:

$$W_s = \rho_w K_w \frac{\partial}{\partial z}(\psi + z), \qquad (9)$$

where K_w is the hydraulic conductivity, ψ is the soil matrix potential, and z is the gravitational potential.

Close to saturation, $\psi < z$, and $W_s = \rho_w K_w$ is a reasonable approximation. By definition,

$$w_2 = \frac{1}{d_2} \int_0^{d_2} \eta(z)dz. \qquad (10)$$

Finally, we write

$$\frac{\partial w_2}{\partial t} = -\frac{K_w}{d_2}. \qquad (11)$$

Eq. (11) can be integrated in time starting from saturation $w_{\rm sat}$ and assuming the following relationship between the hydraulic conductivity K_w and the volumetric water content w_2:

$$K_w = K_{\rm sat}\left(\frac{w_2}{w_{\rm sat}}\right)^{2b+3}. \qquad (12)$$

The water content at time t is given by

$$w_2(t) = w_{\rm sat}\left[1 + (2b + 2)\frac{K_{\rm sat}}{d_2 w_{\rm sat}} t\right]^{-1/(2b+2)}. \qquad (13)$$

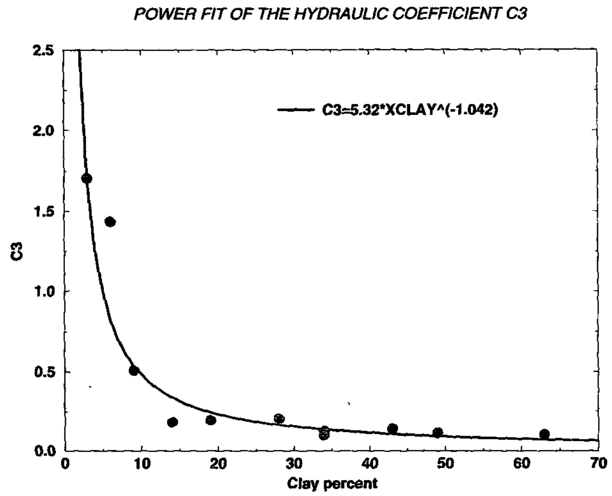

FIG. 2. Variation of the coefficient C_3 versus the percent of clay (Table 1). The solid line corresponds to the best-fit curve [Eq. (16)].

Integration of Eq. (6) provides the volumetric water content w_2^* for $t = \tau / C_3$ starting from saturation for a Newtonian restore:

$$w_2^* = w_{\mathrm{fc}} + \frac{w_{\mathrm{sat}} - w_{\mathrm{fc}}}{e}. \qquad (14)$$

Replacing this value in Eq. (13) provides an estimate of the coefficient C_3:

$$C_3 = \frac{\tau(2b + 2)K_{\mathrm{sat}}}{d_2 w_{\mathrm{sat}}[(w_2^* / w_{\mathrm{sat}})^{-2b-2} - 1]}. \qquad (15)$$

Textural properties (w_{sat}, K_{sat}, b, X_{clay}) are taken from a statistical analysis of Clapp and Hornberger (1978) for samples from 11 soil types (Table 1). The time constant C_3 increases for finer soil textures. For a soil depth of 1 m, sand is restored in 14 h toward field capacity and clay is restored in 10 days. As an example, according to Eqs. (6) and (11), the evolution of w_2 is displayed on Fig. 1 for silty clay loam. In order to derive continuous relationships for ISBA [see Noilhan and Lacarrère (1995)], the textural dependency of this coefficient can be fitted by a power law with the fraction of clay X_{clay} (Fig. 2)

$$C_3 = \frac{5.32 X_{\mathrm{clay}}^{-1.042}}{d_2}. \qquad (16)$$

This simple expression can be used when only the fraction of the clay content is known.

In summary, since the force restore model cannot be used with the vertical gradient of ψ, we impose a Newtonian restore toward field capacity to represent gravitational drainage. However, Eq. (11) remains valid close to saturation, and it is used to estimate the e-folding time of soil moisture depletion, starting from saturation, as a function of soil textural properties.

3. Simulation of the annual cycle of soil moisture

a. Description of the HAPEX–MOBILHY dataset

The HAPEX–MOBILHY experiment described in André et al. (1986, 1988) took place in southwestern France and documented the various components of the surface water budget over a 1° × 1° domain. Most of the instrumental network was run during a three-month IOP (May–July 1986). Soil moisture was monitored during a 2-yr period from 1985 to 1986 (Goutorbe et al. 1989), and the network of automatic weather stations recorded during all of 1986. From these data, a 1-yr time series of atmospheric forcing data was built, which started on 1 January 1986 at the site of Caumont (SAMER station 3), consisting of soya crop over loamy soil (Goutorbe 1991). The dataset was prepared to assess soil moisture simulation in current land surface schemes and to increase understanding of soil moisture parameterization. This was the aim of the RICE and PILPS workshop (14–25 November 1994 at the Climatic Impacts Centre, Macquarie University, Sydney) (Shao et al. 1994). The series includes atmospheric parameters (wind speed, relative humidity, temperature, surface pressure at 2 m), precipitation, and radiative parameters (downward longwave radiation, downward shortwave radiation), given a 30-min sampling. Missing data have been completed from neighbouring stations.

b. Presentation of the modeling results

The annual cycle of soil moisture was simulated over a 1.6-m depth with the surface scheme ISBA, using atmospheric and radiative forcing presented previously. Input parameters characterizing the soil and the vegetation are reported in Table 2 for those that are constant and in Table 3 for those having a seasonal evolution. Soil texture was analyzed, and threshold values (w_{fc} and w_{wilt}) correspond to extremes of observed soil moisture curves. Surface albedo was deduced from upward and downward solar radiation measurements, whereas the evolution of the soya crop properties (vegetation cover, leaf area index, roughness length) has been obtained from height measurements of the vegetation canopy (Bessemoulin et al. 1987). Initial soil moisture is taken from observations on 1 January 1986.

Two simulations were performed, one corresponding to the original scheme (OLD) and another accounting for gravitational drainage (NEW). The annual cycle of simulated soil moisture were compared with weekly neutron sounding over the whole soil column (1.6 m), as shown in Fig. 3. The necessity for including a factor C_3 different from zero is clearly shown. Without a drainage term, rainfall accumulates in the soil during the first four months where the atmospheric demand for surface evaporation is weak (low radiative energy). In the OLD experiment, soil moisture at the end of spring is much too high. Even though evaporation occurs at potential rate during May and June, soil moisture only reaches a level compatible with observations by the middle of September. On the contrary, during winter, the NEW experiment simulates an almost constant soil moisture content close to field capacity, which is in better agreement with observations. During the depletion phase, soil moisture decreases more rapidly in the NEW experiment than in observations, but the recharge phase in autumn is well reproduced. During the IOP, the land surface scheme simulates an accu-

TABLE 2. Input parameters for soil and vegetation properties.

Soil type	Loam
Wilting point w_{wilt} (m^3 m^{-3})	0.20
Field capacity w_{fc} (m^3 m^{-3})	0.32
Minimum stomatal resistance $R_{s\min}$ (s m^{-1})	150
Albedo	0.20
Emissivity	1.00

TABLE 3. Monthly values of vegetation parameters having a seasonal cycle.

Month	1	2	3	4	5	6	7	8	9	10	11	12
Leaf area index ($m^2\ m^{-2}$)	0	0	0	0	1	3	3	3	3	0	0	0
Vegetation cover (%)	0	0	0	0	50	90	90	90	90	0	0	0
Heat roughness length (cm)	1	1	1	1	5	15	15	15	15	1	1	1
Momentum roughness length (cm)	0.1	0.1	0.1	0.1	0.5	1.5	1.5	1.5	1.5	0.1	0.1	0.1

mulated evaporation of 118 mm from 28 May to 3 July for the NEW experiment, compared to 123 mm measured by the SAMER station [balance energy method, see Goutorbe (1991)]. The various terms of the water balance simulated in the annual mean are: total evaporation of 629 mm and accumulated drainage of 227 mm for total precipitation equal to 856 mm. Approximately 26% of the water input is lost by gravitational drainage. The total evaporation amount simulated is consistent with estimates coming from a simple surface model run operationally at Météo-France from weather station reports (Choisnel et al. 1987).

4. Conclusions

The land surface scheme ISBA (Noilhan and Planton 1989), based on the force–restore method (Deardorff 1977), has been modified to account for gravitational drainage at the bottom of the soil layer. A simple relaxation toward field capacity is proposed, with a time constant depending upon soil texture. With this modification, the scheme is capable of reproducing the observed evolution of soil moisture during a 1-yr period at the location of Caumont in southwestern France (HAPEX–MOBILHY 86) (soya crop over loamy soil). In particular, during the winter season, soil moisture within 1.6 m remains at a constant value close to field capacity, although rainfall is important.

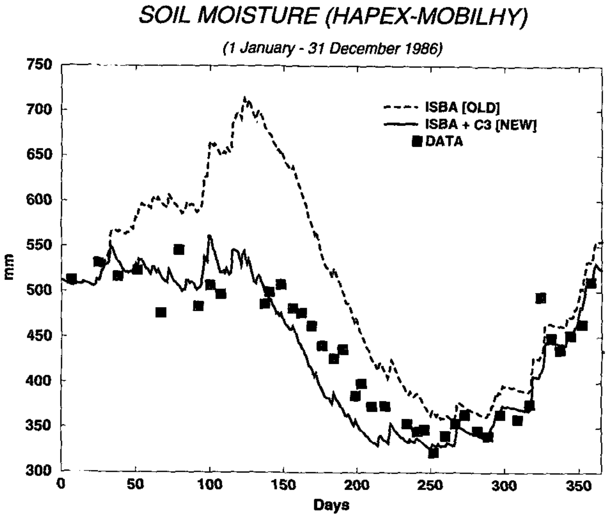

FIG. 3. Annual cycle of total soil water content (mm) in the top 1.6 m simulated by ISBA with (solid line) and without drainage (dotted line) and in comparison with HAPEX data (solid squares).

This study emphasizes the important role played by observed datasets in improving land surface parameterization schemes. Field experiments such as HAPEX, FIFE, and BOREAS have collected huge amounts of data during monthly intensive observing periods. Surface evapotranspiration has then been compared to model estimates. However, the link between soil moisture and evaporation can only be checked during longer periods of time. Moreover, the link between surface processes and deep hydrology is not well known. Future field experiments {GCIP [GEWEX (Global Energy and Water Cycle Experiment) Continental-Scale International Project] LAMBADA–BATERISTA} will promote long-term measurements related to all the components of the continental hydrological cycle. This appears to be the only way to improve our understanding of the processes involved and to better represent them in numerical models.

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APPENDIX

List of Symbols

b	slope of the retention curve
C_T	surface transfer coefficient for heat
C_1	soil transfer coefficient for moisture
C_2	soil transfer coefficient for moisture
C_3	soil transfer coefficient for moisture
d_1	superficial soil depth
d_2	total soil depth
E	total evaporation rate, $E = E_g + E_{tr} + E_r$
E_g	bare-soil evaporation rate
E_{tr}	transpiration rate
E_r	evaporation rate from the interception reservoir (interception loss)
H	sensible heat flux
K_w	hydraulic conductivity
K_{sat}	hydraulic conductivity at saturation
L	latent heat of vaporization
P	precipitation rate
P_g	flux of liquid water reaching the soil surface

R_n	net radiation
R_r	runoff from the interception reservoir
T_s	surface temperature
T_2	daily mean surface temperature
veg	fraction of vegetation cover
w_{fc}	volumetric water content at field capacity
w_g	surface volumetric water content
w_{geq}	surface volumetric water content when gravity balances the capillarity forces
w_{sat}	volumetric water content at saturation
w_{wilt}	volumetric water content at wilting point
w_2	mean volumetric water content
W_r	water content of the interception reservoir
$W_{r\max}$	maximum water content of the interception reservoir
η	local volumetric water content
ψ	soil matrix water potential
ρ_w	density of liquid water
τ	restore constant of 1 day

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