**ABSTRACT:** A two-layer soil water balance model is developed to provide an efficient and robust description of land surface dynamics in response to atmospheric evaporative events. Soil, vegetation, and atmosphere are coupled dynamically under the assumption that soil moisture profiles approximately preserve similarity during simultaneous atmospheric drying and gravity drainage. The exfiltration flux at the land surface in response to the atmospheric evaporative demand is limited by relating the surface resistance to water vapor transfer in the Penman-Monteith equation to the near-surface soil status. In addition, the control of deeper soils on both exfiltration and drainage is expressed by performing a time compression approximation water balance over the entire drying profile and by scaling the obtained exfiltration and drainage fluxes to the near-surface soil layers. The reliability and robustness of the proposed formulation is evaluated with rates of evaporation calculated from measurements of the Bowen ratio and soil moisture data obtained from time domain reflectometry measurements for a bare soil field in the Zwalmbeek catchment (Belgium).

**INTRODUCTION**

Simulations of the land surface response to atmospheric forcing during storm-interstorm sequences are required in catchment-scale hydrology, agricultural engineering, and large-scale meteorology. Atmospheric forcing refers to precipitation intensity during storm events and potential evaporation rate during interstorm periods. The land surface partitions this atmospheric forcing into infiltration, surface runoff, actual evaporation, ground-water recharge, and changes in storage. Soil and canopy wetness introduce memory and nonlinearity into the soil-vegetation-atmosphere (SVA) system. The distributed description of land surface dynamics during atmospheric evaporative events may be carried out by discretizing the land system into a series of noninteracting SVA columns and by using one-dimensional formulations for each of these columns. The evaluation of the scale at which land surface and atmospheric forcing can be assumed as being horizontally homogeneous is a nontrivial task and it is not within the scope of the present work. This work addresses the local scale quantification of the water fluxes and moisture status across the earth-atmosphere interface on a continuous-time basis. In particular, attention is focused on the near-surface soil drying in response to the atmospheric evaporative demand and gravity drainage.

The problem of the simulation of local scale land surface dynamics during evaporative events has been dealt with in various papers. Detailed numerical solutions of the heat and mass transport equations have been tested under laboratory and field conditions [e.g., van Bavel and Hillel (1976), Sophocleous (1979), Milly (1982), Camillo et al. (1983), Higuchi (1984), Passerat de Silans et al. (1989), and Witono and Bruckler (1989)]. These studies have shown that detailed heat and mass flow models coupled with accurate soil surface energy and radiation balances usually give satisfactory results after a calibration phase. However, solution methodologies relying on the standard discretization procedures are computationally intensive and therefore may be poorly suited for spatially distributed analysis and Monte Carlo simulations. In particular, the simulation of the topsoil dynamics in detailed numerical models can incur prohibitive computational expenses, owing to rapid response of this zone to atmospheric forcing, which constrains such models to small grid and time steps. This motivates a need for simplifying the description of the land surface moisture dynamics through the introduction of SVA transfer schemes (SVATSs). This is especially significant in light of the current problems of hydrologic spatial variability, remote sensing, and climate simulation. In each of these areas there is potential for significant theoretical progress by numerical experimentation using large numbers of SVATSs in parallel.

A number of laboratory studies such as those by Youngs (1960), Gardner (1962), Gardner and Hillel (1962), and Gardner and Gardner (1969) suggest that it may be possible to describe the relation of evaporation and drainage to soil water content by relatively simple though somewhat approximate expressions based on flow theory. To reduce the computational effort with respect to detailed numerical solutions of the heat and mass governing equations, a variety of simplified models were developed for local scale investigations [e.g., Black et al. (1969), Gardner (1973, 1974), and Milly (1986), catchment hydrologic simulations [e.g., Famiglietti and Wood (1994), Wigmosta et al. (1994), Flachinger et al. (1996)], and numerical weather predictions [e.g., Sellers et al. (1986) and Dickinson et al. (1993)]. The aim of the present paper is to provide an efficient description of land surface dynamics during atmospheric evaporative events, which can offer an adequate level of generality with the minimum of physically realistic parameterization. Special emphasis has been given to the control volume description of near-surface soil drying, where the concept of moisture profile similarity is required to incorporate the effects of the deeper soils (Salvucci 1997). The linkage of the scheme developed in the present paper to that developed in Orlandini et al. (1996) for the description of land surface dynamics during storm events provides a complete model for land surface hydrologic simulations on a continuous-time basis. The reliability and robustness of the model is evaluated with rates of evaporation calculated from measurements of the Bowen ratio and soil moisture data obtained from time domain reflectometry (TDR) measurements. Experiments were carried out for three periods of several days each at Ghent, Belgium over a bare soil field in the summer of 1994.

**BACKGROUND**

The theoretical basis on which the model is built are the Penman-Monteith formulation, for the evaluation of the potential flux across the land surface in response to the atmospheric evaporative demand, and the theory of water flow in...
unsaturated-saturated soil domains, for the description of the soil profile response to land surface drying and gravitational drainage.

**Atmospheric Evaporative Demand**

The main goal of evaporation research applied to catchment hydrology is to find methods for calculating the value of evaporation under any given conditions to estimate the amount of water lost from the catchment. The publication of Penman’s combination equation in 1948 was a landmark in studies of evaporation from field crops (Penman 1948). Penman calculated the potential rate of evaporation from standard meteorological data but he recognized that actual evaporation would be less than the potential amount if there was a shortage of soil water. Monteith (1965) generalized the Penman equation by specifying two resistances to vapor transfer: from evaporating surfaces (into the stomatal cavities of vegetation) to the air within a canopy \( r_c \) and through the air to the reference height \( r_a \). The Penman-Monteith equation can be written in the form

\[
p_a L_e = \frac{\Delta (Q_s - Q_e)}{\Delta + \gamma (1 + r_a/r_c)} + \rho_a c_p \left[ e_s(T) - e(T) / r_a \right]
\]

where \( \rho_a \) = density of air; \( L_e \) = latent heat of vaporization per unit mass of water; \( e \) = evaporation rate; \( \Delta \) = rate of change of saturated vapor pressure with temperature; \( Q_s \) = soil heat flux density (equal to the net radiation flux density minus the reflected shortwave and long-wave solar radiation and the long-wave radiation emitted by the SV body because of its temperature); \( Q_e \) = soil heat flux density (equal to the net energy advected by moving water minus the change in system energy); \( \rho_a \) = density of air; \( c_p \) = specific heat of air at constant pressure; \( e_s(T) \) = saturated vapor pressure at screen temperature; \( T \) = temperature; \( e \) = vapor pressure at screen height; \( r_a \) = aerodynamic resistance to water vapor transfer; \( \gamma \) = psychrometer constant

\[
\gamma = \frac{c_p \rho_a}{0.622 L_e}
\]

\( p_a \) being the ambient atmospheric pressure at the ground surface; and \( r_c \) = surface resistance to water vapor transfer. Latent heat represents the amount of heat exchanged required for inducing the state change per unit mass of substance and can be expressed as a function of temperature by \( L_e = 597.3 - 0.577 T \), where \( L_e \) is given in calories per gram and \( T \) is given in degrees Celsius. The saturated vapor pressure \( e_s \) in Pascals at screen temperature \( T \) in degrees Celsius and \( \Delta = de/dT \) (in Pascals per degrees Celsius) are obtained on the basis of the empirical equation

\[
e_s(T) = 611 \exp \left( \frac{17.27 T}{237.3 + T} \right)
\]

The actual vapor pressure \( e \) in Pascals at screen height is derived from the wet-bulb \( T_w \) and dry-bulb \( T_d \) temperatures measured by an aspirated psychrometer as \( e(T_w, T_d) = e_s(T_w) - 6.66 \times 10^{-4} \rho_a (T_w - T_d) \). The aerodynamic resistance

\[
r_a = \left( \frac{\ln \left( \frac{z_a}{d} \right)^2}{\left[u(z) h^2 \right]} \right)
\]

where \( z_a \) = roughness length; \( d \) = zero plane displacement; \( u(z) \) = wind speed at height \( z \); and \( k \) = von Karman constant (\( k = 0.41 \)) is calculated by assuming that \( z_o \) is given by \( z_o = 0.10 h \) (Szeicz et al. 1969), where \( h \) = height of the crop and the zero plane displacement is given by \( d = 0.75 h \) (Kung 1961).

In the Penman-Monteith equation the two main processes that govern evaporation are separated. The radiation term describes the rate of heat input necessary to change liquid water into vapor when the atmosphere is water saturated, and the ventilation term describes the ease with which the vapor is carried away from the evaporating surface because of the deficit of vapor pressure with respect to saturation conditions. The surface resistance \( r_c \) introduced by Monteith (1965) with reference to vegetated land surfaces generalizes the Penman equation to those cases for which vapor pressure at the surface of the dry leaf is less than the saturation vapor pressure at the leaf temperature and liquid water changes to vapor inside the leaves. Potential evaporation \( e_p \) can be defined as the evaporation calculated with \( r_c = 0 \) in (1). This is equivalent to assuming that the surface acts as if its vapor pressure were the saturated vapor pressure at the surface temperature. For surfaces of small \( z_o \), \( r_a \) is greater than \( r_c \) and \( e_p \) is a useful measure of the maximum evaporation to be expected. However, for forests \( z_o \) is large, \( r_c \) is less than \( r_a \) and \( e_p \) is very large and is not near the actual evaporation unless the canopy is wetted by rain. Therefore \( e_p \) is not a very useful value for forests and does not correspond to evaporation without limiting soil water supply. The evaporation that would occur if soil water supply were not limiting can be called “unstressed evaporation.” This is the evaporation that would occur if stomatal openings were not affected by plant water potential and can be calculated with \( r_c = r_{\text{min}} \) in (1). Unstressed evaporation does depend both on the response of the stomata to ambient humidity, height, and temperature and on the canopy properties of \( z_o \) and leaf area index, but it does not depend on soil, root, and internal resistance parameters (Federer 1979).

Stomatal resistance depends on a number of factors. The effect of light on stomatal opening differs little among species: stomata are nearly fully open in one-tenth of full sunlight. On sunny days, it decreases rapidly at sunrise and increases at sunset, remaining at some minimum value all day if water supply to the leaf is adequate. This minimum resistance \( r_{\text{min}} \) exists when neither water nor light limit stomatal opening. The significant variable for characterizing the water status of leaves is leaf water potential that depends on the soil water potential through the transport mechanism of vegetation (van den Hoënt 1948). For bare soil land surfaces, \( r_c \) can be assumed to represent the resistance to water vapor transfer of the drying topsoil (van de Griend and Owe 1994). With this extension, the Penman-Monteith equation is applicable to all types of land surface (bare soil included) but its use requires that the surface resistance \( r_c \) be measured or related to simple attributes of the soil and crop in the same way that the aerodynamic resistance \( r_a \) can be estimated from land surface roughness or crop height and wind speed. From the pioneering work of Russell (1980), increasing attention has been devoted to the relationship between \( r_c \) and some measure of soil moisture status [e.g., Mahfouf and Noilhan (1991); Crago and Brutscater (1992), van de Griend and Owe (1994), and Daamen and Simmonds (1996)]. To estimate losses by evaporation from catchments, the land surface dynamics can be described by considering the evaporative demand to incorporate the main features of atmospheric forcing and by limiting this evaporation demand to soil according to the land surface resistance and soil profile dynamics.

**Soil Moisture Dynamics**

In virtually all studies of the unsaturated zone, the fluid motion is assumed to obey the classical Richards equation. This equation may be written in several forms, with either pressure head \( \psi \) or moisture content \( \theta \) as dependent variables. The constitutive relationship between \( \psi \) and \( \theta \) allows for conversion of one form of the equation to another. The one-dimensional Richards equation with pressure head as the dependent variable can be written as
\[ \sigma(\psi) \frac{\partial \psi}{\partial t} = \frac{\partial}{\partial z} \left[ K_r(\psi) \frac{\partial (z + \psi)}{\partial z} \right] \]  

(5)

where \( \sigma(\psi) = \frac{\partial \theta}{\partial \psi} \) = specific moisture capacity; \( t \) = time; \( z \) = vertical coordinate (positive upward); and the hydraulic conductivity is expressed as a product of the conductivity at saturation \( K_r \) and the relative conductivity \( K_r(\psi) \). This equation is obtained easily by incorporating the expression of the vertical Darcian flux for saturated-saturated domain,

\[ q_r = -K_r(\psi) \frac{\partial (z + \psi)}{\partial z} \]  

(6)

into the continuity equation \( \partial \theta / \partial t + \partial q_r / \partial z = 0 \). The constitutive relationships introduced by Brooks and Corey (1966) can be used to describe the nonlinear dependencies of \( \theta, K_r \), and \( \sigma \) on \( \psi \)

\[ \theta(\psi) = \begin{cases} \theta_0 + (\theta_s - \theta_0) \left( \frac{\psi}{\psi_s} \right)^n & \text{if } \psi \leq \psi_s \\ \theta_0 & \text{if } \psi > \psi_s \end{cases} \]  

(7)

\[ K_r(\psi) = \begin{cases} \left( \frac{\psi}{\psi_s} \right)^2 & \text{if } \psi \leq \psi_s \\ 1 & \text{if } \psi > \psi_s \end{cases} \]  

(8)

\[ \sigma(\psi) = \begin{cases} \frac{\eta(\theta_0 - \theta_0)}{\theta_0} \left( \frac{\psi}{\psi_s} \right)^{n-1} & \text{if } \psi \leq \psi_s \\ 0 & \text{if } \psi > \psi_s \end{cases} \]  

(9)

where \( \theta_0 \) = residual moisture content; \( \theta_s \) = saturated moisture content; \( \psi_s \) = saturated soil matrix potential; and \( \eta \) = a pore-size distribution index. With the foregoing constitutive equations, hysteresis effects on moisture redistribution in the soil profile are not taken into account.

Vapor fluxes and thermal effects are neglected in (5). Philip and de Vries (1957) noted that during soil-limited evaporation the sites of phase change occur lower in the soil, leaving a dry layer in which the dominant moisture transport mechanism is vapor flow along moisture and temperature gradients. At first glance the presence of this layer may appear to invalidate the application of methods based on liquid transport. However, the evaporation rate still can be estimated by modeling the liquid flow regime beneath this layer, so long as vapor can be transported through the dry layer at a rate greater than or equal to the rate of the liquid flow below. The applicability of simplified single-phase isothermal descriptions depends not on the presence or absence of a dry layer but rather on which process is limiting (Salvucci 1997). The work presented here assumes that the evaporation rate is limited ultimately by liquid flow and thus that a simple single-phase isothermal process model description is adequate.

**MODEL FORMULATION**

The essential parameters of the formulations described earlier are incorporated into the SVATS developed in this study to provide an integrated representation of the SVA continuum at the elemental scale described by digital elevation data and terrain attributes (Fig. 1). The response of vegetation and soil to the atmospheric forcing is calculated through the application of a simple vegetation cover water balance model and a two-layer soil water balance model based on the time compression approximation (TCA) concept. As reported in Salvucci (1997), this approximation first was applied to problems involving the wetting of soils by Sherman (1943) and has been used successively in various forms for both wetting and drying [e.g., Gardner and Hillel (1962), Gardner and Gardner (1969), Smith and Hebert (1983), and Milly (1986)]. The approximation results from the observation that soil-limited evaporation (infiltration) rates at a given time depend primarily on the cumulative evaporation (infiltration) depth up to that time and only in a secondary fashion on the details of the evaporation (infiltration) history. The TCA uses the exfiltration (infiltration) capacity to estimate the soil-limited transport that would occur if the soil first underwent a period of atmosphere-limited evaporation (infiltration). A review of this approximation, its accuracy, and its use in water balance modeling recently was provided by Salvucci and Entekhabi (1994). Applicability and performance of the proposed parameterization in response to storm events have been considered in a previous paper (Orlandini et al. 1996). The response of the land surface to atmospheric drying and gravity drainage during interstorm periods is considered in the present work to provide a complete description of land surface dynamics on a continuous-time basis.

Actual evapotranspiration is considered as a phenomenon associated with the soil and canopy moisture depletion responses to atmospheric potential demand. It is assumed that the atmospheric evaporative demand \( e_p \) is obtained by (1) with \( r_s = 0 \), and it is set to zero at night. Actual evaporation from wet canopy and soil is expressed dynamically as a function of \( e_p \) and the moisture status of each subsystem. If the atmospheric evaporative demand is not satisfied completely by the wet canopy, the residual rate acts as forcing of the SV system and the water flux released by this system is limited by the surface resistance to vapor transfer \( r_s \) in (1) and by the exfiltration capacity of the unsaturated soil profile (soil-limited evaporation). The surface resistance to water vapor transfer \( r_s \) is allowed to range from a minimum \( r_{s,\text{min}} \) to a maximum \( r_{s,\text{max}} \) depending on the degree of saturation of the upper near-surface soil layer. Exfiltration capacity and gravity drainage are calculated by means of a TCA procedure on the basis of the assumption of geometric similarity of the soil moisture profile during the drying event. Dynamically, the model computes a series of steady-state flux densities over short time intervals and recalculates storages at the end of each interval for use in the next interval.
Vegetation Cover Water Balance

The water balance of the vegetation cover is expressed by the continuity equation

\[
dS \frac{dt}{d} = p - e, -p_s
\]

where \( S \) = actual water content of the cover; \( t \) = time; \( p \) = gross inflow precipitation; \( e \) = actual evaporation from the cover; and \( p_s \) = net outflow precipitation (ground level precipitation). On the basis of the works by Rutter and Morton (1977) and Massman (1980), ground level precipitation and actual evaporation from the wet cover are expressed through the storage-outflow relationships \( e, e_r, (S/C)^{e_r} \) and \( p_s = p(S/C)^{e_r} \), respectively, where \( C \) is the potential retention of the cover, \( e \) is the wet surface potential evaporation, and \( e_r \) and \( v \) are parameters characteristic of the vegetation cover. Eq. (10) is solved numerically during storm and interstorm events by applying a simple explicit Euler scheme (like that described in the next paragraph for the soil water balance).

Soil Water Balance

The formulation of a control volume water balance for the near-surface soil during evaporation and drainage events is a nontrivial task because at the land surface the term in (6) related to the soil matrix gradient (diffusional term) may be important and the direct evaluation of this term may be obtained only by solving numerically the Richards equation in (5) along the soil profile on the basis of well-defined initial and boundary conditions. Neglecting near-surface soil matrix gradients during drying conditions may lead to excessive control volume drainage outflows and premature control volume emptying. In the proposed formulation the lack in the knowledge of the soil moisture is filled with the assumption of geometric similarity of the vertical soil moisture profile introduced by Salvucci (1997), that is

\[
\Theta(z,t) = \begin{cases} 
\Gamma(z/L(t)) & \text{if } -L < z \leq 0 \\
\Theta_0 & \text{if } z \geq 0 
\end{cases}
\]

where \( \Theta = (\Theta_0 - \Theta_0)/(\Theta_0 - \Theta_0) \) = degree of soil saturation; \( \Gamma(z) \) = some (undetermined) function describing the shape of the drying profile; \( L(t) \) = scaling length of the profile and may be thought of approximately as a penetration depth of the drying front; and \( \Theta_0 \) = (uniform) initial condition of soil saturation. As noted in Salvucci (1997), (11) “is not meant to imply that the drying front is sharp [as in the interpretation of Green and Ampt (1911) wetting fronts] but only that the profile preserves geometric similarity.” This assumption is shown by Salvucci (1997) to be reasonable through comparison with a numerical solution of the Richards equation in (5) for homogeneous soils under a wide range of soil types and initial condition moisture saturations.

With the foregoing similarity assumption the exfiltration capacity of the drying profile can be expressed through the TCA concept as

\[
f_s = \frac{S_e^2}{2(F_s + G)}
\]

where \( S_e \) = desorptivity; \( F_s \) = cumulative exfiltration depth at the land surface in response to atmospheric (variable) drying; and \( G \) = cumulative drainage depth at the drying front \( z = -L \) in response to gravity forces (Salvucci 1997). As expressed in Salvucci (1997), (12) is obtained under the assumption that during drying conditions exfiltration capacity can be expressed by means of the Darcian flux equation in (6) where the diffusional term \(-K_sK_0^0/d\) is dominant with respect to the gravitational term \(-K_sK_0^0/d\), as it reasonably can be assumed for near-surface soils. The desorptivity \( S_e \) can be evaluated on the basis of the soil hydraulic properties and initial (uniform) moisture saturation \( \Theta \), [e.g., Eagleson (1978), Parlane et al. (1985)]. In the present study the expression given by Eagleson (1978) is used

\[
S_e = 2(\Theta_0, (\Theta_0)^{0.5} \left[ \frac{\phi_1 \psi K_0^0}{\eta \pi} \right]^{0.5})
\]

where \( \Theta_0 \) and \( \Theta_0 \) = initial and land surface boundary conditions of medium saturation; \( \delta = (1 + 2n)/n \); and \( \phi_0 \) = a dimensionless parameter defined by the equations

\[
\phi_0 = \frac{1.85}{(\Theta_0 - \Theta_0)^{0.85}} \int_{\Theta_0}^{\Theta_0} \Theta_0 \Theta_0, \Theta_0^0.85 d\Theta
\]

The initial condition of soil saturation \( \Theta_0 \) in (13) is expressed by

\[
\Theta_0 = \frac{Z_{up}}{Z_{up} + Z_{low}} \Theta_{up} + \frac{Z_{low}}{Z_{up} + Z_{low}} \Theta_{low}
\]

where \( \Theta_{up} \) and \( \Theta_{low} \) = average initial soil moisture saturations of the upper and lower layer, respectively. The land surface boundary condition during interstorm events is assumed to be

\[
\Theta_0 = 0
\]

The actual exfiltration flux at the land surface may be expressed as

\[
f_s = \min(e_r, f_s)
\]

where \( e_r \) = residual atmospheric evaporative demand; and \( f_s \) = exfiltration capacity. Following Wigmosta et al. (1994), \( e_r \) is given by

\[
e_r = (\psi - e_r) \frac{\Delta + \gamma}{\Delta + \gamma(1 + r/r_s)}
\]

where the control of land surface drying is incorporated by varying the surface resistance to water vapor transfer \( r \), with the upper layer moisture status \( \Theta_{up} \) as proposed by Mahfouf and Noilhan (1991) and further investigated by Daamen and Simmonds (1996), that is

\[
r_s = A \exp(B \Theta_{up})
\]

where \( A = r_{max} \) and \( B = -\ln(r_{max}/r_{min}) \). The gravity flux at the drying front \( z = -L \) may be expressed on the basis of (6), where the Brooks and Corey (1996) relationships are incorporated to estimate the unsaturated conductivity and the soil matrix gradient is neglected, that is

\[
g = K_0 \Theta_0^{2.5 - \psi n/\psi}
\]

where \( \Theta_0 \) = initial soil moisture saturation along the unsaturated profile as given by (15).

Water balances of the upper and lower soil layers are carried out by scaling the exfiltration and the drainage fluxes as obtained from the foregoing TCA calculation. For the upper soil layer one can write

\[
Z_{up} \frac{d\Theta_{up}}{dt} = -f_s - g_{up}
\]

where

\[
f_s = f_s Z_{up} / L
\]

and

\[
g_{up} = g Z_{up} / L
\]
whereas the lower layer water balance is expressed by

\[ Z_{\text{low}} \frac{d \theta_{\text{low}}}{dt} = -f_{\text{low}} - g_{\text{low}} \]

(24)

where

\[ f_{\text{low}} = f \cdot Z_{\text{low}} / L \]

(25)

and

\[ g_{\text{low}} = g \cdot Z_{\text{low}} / L \]

(26)

The scaling length \( L \) in (22), (23), (25), and (26) represents the drying front depth and is estimated on the basis of the \( \theta(\psi) \) constitutive equation in (7) as if the pressure head profile in the unsaturated soil was maintained hydrostatic (\( \varepsilon + \psi = \text{constant} \)) during the drying event, with a minimum threshold of \( (Z_{\text{up}} + Z_{\text{low}}) \). Under these assumptions one can obtain

\[ L(t) = \max \{ |\psi| [(\Theta_{\text{up}}(t)]^{-1/\gamma}, Z_{\text{up}} + Z_{\text{low}} \} \]

(27)

Eqs. (22), (23), (25), and (26) are based on the observation that water drains almost equally from all layers during drying events as reported in Prill et al. (1965) (on the basis of laboratory studies of drainage from various sands) and in Richards et al. (1956) (on the basis of drainage experiments in the field). This evidence perhaps may be interpreted, at least in part assuming that during drying events the soil matrix is drained by preferential paths in the soil domain and both upward (exfiltration) and downward (drainage) fluxes that may occur in these preferential paths do not interact significantly with the lateral soil matrix. Although the assumption employed to derive (27) that pressure head profile is maintained hydrostatic during drying events is not verified generally in a strict manner, it allows us to incorporate the broad effects of near-surface soil moisture status \( \Theta_{\text{up}} \) and soil hydraulic properties (\( \psi, \eta \)) in the dynamic estimation of the drying soil depth \( L(t) \). This approximation is introduced in the present work as a first possible step in the purpose of building a simplified control volume model iteratively and accurate evaluations of errors induced by deviations of soil moisture profiles from hydrostatic conditions will constitute the next logical step. The minimum threshold \( (Z_{\text{up}} + Z_{\text{low}}) \) for the drying front depth \( L(t) \) in (27) is necessary not to obtain unrealistic overestimates of exfiltration \( (f_{\text{up}} + f_{\text{low}}) \) and drainage \( (g_{\text{up}} + g_{\text{low}}) \) fluxes from the two considered soil layers with respect to those provided by (17) and (20), respectively, when the capillary fringe is shallower than the bottom of the lower soil layer. These are circumstances in which the problem of excessive drainage fluxes in control volume soil water balances mentioned in the first paragraph of the present section is relatively unimportant.

Eqs. (21) and (24) are solved numerically during the interstorm event by applying two levels of discretization (Fig. 2). With reference to the upper soil layer at the first level the interstorm simulation period \([0, T]\) is divided into \( M \) intervals \([m = 1, \ldots, M]\) and at the \( m \)th time interval the average value of the atmospheric input variables are considered. At the second level of discretization the generic time interval \([m-1, m]\), \( m \Delta T \) is divided into \( K \) subintervals \((k = 1, \ldots, K)\), where \( \Delta T = m \Delta T \), so that (21) can be solved numerically from the initial condition \( \theta_{\text{up}}^0 = \theta_{\text{up}}^{n-1} \). The mean values of the fluxes \( f_{\text{up}} \) and \( g_{\text{up}} \) are taken to be

\[ f_{\text{up}}^m = \frac{1}{\Delta T} \sum_{k=1}^{K} \frac{1}{2} (f_{\text{up}}^{n-1} + f_{\text{up}}^{n}) \Delta T \]

(29)

\[ g_{\text{up}}^m = \frac{1}{\Delta T} \sum_{k=1}^{K} \frac{1}{2} (g_{\text{up}}^{n-1} + g_{\text{up}}^{n}) \Delta T \]

(30)

### REAL CASE APPLICATION

The SVATS developed in this paper is tested against rates of evaporation calculated from measurements of the Bowen ratio and soil moisture data obtained from TDR measurements. Field data were collected by the Laboratory of Hydrology and Water Management of the University of Ghent (Belgium) and refer to a bare soil plot in the Zwalmbeek catchment. Atmospheric variables and soil moisture measurements were available for three periods of several days each among which the August 9–17, 1994 period was selected to present the capabilities of the proposed formulation. The parameter values used in the present study are reported in Table 1.}

### TABLE 1. SVATS Simulation Parameters

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1)</td>
<td>(2)</td>
</tr>
<tr>
<td>( h ) (m)</td>
<td>0.10</td>
</tr>
<tr>
<td>Leaf area index</td>
<td>1.00</td>
</tr>
<tr>
<td>( C ) (mm)</td>
<td>0.10</td>
</tr>
<tr>
<td>( \mu )</td>
<td>0.15</td>
</tr>
<tr>
<td>( \nu )</td>
<td>3.00</td>
</tr>
<tr>
<td>( r_{\text{min}} ) (m/s)</td>
<td>24</td>
</tr>
<tr>
<td>( r_{\text{min}} ) (m/s)</td>
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</tr>
<tr>
<td>( \theta_{\text{e}} )</td>
<td>0.16</td>
</tr>
<tr>
<td>( \theta_{\text{i}} )</td>
<td>0.40</td>
</tr>
<tr>
<td>( \eta )</td>
<td>0.675</td>
</tr>
<tr>
<td>( \psi_{\text{r}} ) (m)</td>
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</tr>
<tr>
<td>( K_{\text{i}} ) (mm day(^{-1}))</td>
<td>57.0</td>
</tr>
<tr>
<td>( Z_{\text{up}} ) (m)</td>
<td>0.10</td>
</tr>
<tr>
<td>( Z_{\text{low}} ) (m)</td>
<td>0.60</td>
</tr>
</tbody>
</table>

The initial condition \( \theta_{\text{up}}^0 = \theta_{\text{up}}^{n-1} \). The mean values of the fluxes \( f_{\text{up}} \) and \( g_{\text{up}} \) are taken to be

\[ f_{\text{up}}^m = \frac{1}{\Delta T} \sum_{k=1}^{K} \frac{1}{2} (f_{\text{up}}^{n-1} + f_{\text{up}}^{n}) \Delta T \]

(29)

\[ g_{\text{up}}^m = \frac{1}{\Delta T} \sum_{k=1}^{K} \frac{1}{2} (g_{\text{up}}^{n-1} + g_{\text{up}}^{n}) \Delta T \]

(30)
parameters (h, leaf area index, C, μ, and ν) are set here to reproduce the surface roughness and storage capacity of the bare soil. Surface resistances (rmin and rmax) are estimated as fitting parameters on the basis of simulated and measured actual cumulative evaporation depths. Soil hydraulic properties (θup, θmax, η, ψs, and Ks) are obtained from detailed laboratory water release curve analysis and are typical of clay soils. Structural parameters (Zup and Zlow) are set to provide a useful representation of the near-surface soil dynamics for both interstorm and storm events (Orlandini et al. 1996).

The computed average moisture content time series for the upper soil layer (Zup = 0.10 m) θup are compared with TDR soil moisture measurements at 5 cm depth θ5 in Fig. 3, where the ground level precipitation p, also is reported to allow the distinction between storm and interstorm events in the simulation period. The agreement between simulated and observed data also is shown in Fig. 4, in the form of a scatter plot. Hourly mean evaporation was calculated by applying the energy balance method where the Bowen ratio concept is incorporated. Because the sensible-heat term Qc of the energy balance equation is difficult to measure, it commonly is related to the energy used in evaporation Qe as Qc = RQe, where the Bowen ratio R is expressed as

\[
R = \frac{c_p \rho_a}{0.622 L_c} \frac{T(z) - T(z_0 + d)}{\epsilon(z) - \epsilon(z_0 + d)}
\]

where \(\epsilon_{br}\) is the rate of evaporation calculated from measurements of the Bowen ratio R. The computed actual evaporation time series (\(e_{comp} = e_r + f_e\)) is compared with the evaporation obtained from Bowen ratio measurements (\(e_{br}\)) in Fig. 5, in which the atmospheric evaporative demand (\(e_r\)) also is reported. Cumulative evaporation depths are plotted in Fig. 6, where \(E_{br}\), \(E_{nrm}\), and \(E_{comp}\) denote the integrals of the \(e_{br}\), \(e_{nrm}\), and \(e_{comp}\) evaporation rates, respectively.

By using the Penman-Monteith equation with estimation of the energy budget, the surface resistance to water vapor transfer can be obtained. On rearrangement of the Penman-Monteith equation, the surface resistance \(r_s\) is given by

\[
r_s = \left( \frac{\Delta R}{\gamma} - 1 \right) r_v + \frac{\rho_a c_p}{\gamma(Q_e - \bar{Q}_e)} [\epsilon_r(z) - \epsilon(z)](1 + R)
\]

where \(r_v\) is the resistance to vapor transfer to the atmosphere, \(\rho_a c_p\) is the heat capacity at constant pressure, \(Q_e\) is the latent heat flux, and \(\gamma\) is the psychrometric constant. The values of surface resistance restricted to those time intervals of the simulation period, in which (1) \(e_r > 0\) (daylight-time intervals), (2) \(p = 0\) (interstorm time intervals), and (3) the actual evaporation rate \(f_e\) is not limited by the exfiltration capacity \(f_{es}\), are calculated through (33) \([r_s(br)]\) and are plotted in Fig. 7 together with values obtained from the model relationship in (19) \([r_s(mod)]\).

To determine the accuracy with which soil hydraulic properties and surface resistance to water vapor transfer for the developed SVATS must be evaluated, a sensitivity analysis is carried out based on the simulated values of topsoil moisture content and actual cumulative evaporation depth. For instance, the sensitivity of the SVATS to saturated hydraulic conductivity \(K_s\) is shown in Fig. 8 by plotting simulated against measured topsoil moisture content for different values of \(K_s\). The sensitivity to minimum and maximum surface

FIG. 3. Comparison between Time Series of Computed Average Moisture Contents of Upper Soil Layer θup and TDR Soil Moisture Measurements at 5 cm Depth θ5.

FIG. 4. Comparison between Model Computations of Average Moisture Content at Upper Soil Layer θup and TDR Measurements of Soil Moisture Content at 5 cm Depth θ5.
FIG. 5. Comparison between Time Series of Computed Actual Evaporation Rates $e_{\text{comp}}$ and Actual Evaporation Rates Calculated from Bowen Ratio Measurements $e_{br}$

FIG. 6. Comparison between Model Computations of Actual Cumulative Evaporation Depth $E_{\text{comp}}$ and Actual Cumulative Evaporation Depths Calculated from Bowen Ratio Measurements $E_{br}$

FIG. 7. Comparison between Values of Surface Resistance Obtained from Field Measurements through (33) $r_s(br)$ and Those Obtained from Model Relationship in (19) $r_s(mod)$

SUMMARY AND DISCUSSION

A simple SVXTS has been developed to describe the local scale land surface dynamics during interstorm drying events. This scheme was linked to the storm water balance formulation developed in Orlandini et al. (1996) and tested against evaporation rates calculated from Bowen ratio measurements and soil moisture data obtained from TDR measurements. On the basis of these data, the control volume description of near-surface soil drying appeared to be a nontrivial task. Data such as that of Prill et al. (1965) (obtained from laboratory studies of drainage from various sands) and Richards et al. (1956) (for drainage in the field) demonstrated that the soil matrix gradient term in the flow equation in (6) is relatively small during a large fraction of the drainage process and that water drains almost equally from all layers above the initial wetting depth in a uniform profile. However, this evidence is not reproduced by simply neglecting the soil matrix gradient term in (6) as reported in many published formulations [e.g., Famiglietti and Wood (1994); and Wigmosta et al. (1994)], because this assumption leads to excessive control volume outflows and premature control volume emptying. In the developed formulation the near-surface moisture dynamics was described as part of a TCA-based water balance for the entire drying profile on the basis of the soil profile similarity assumption given in (11). Dynamically, variable estimates of the drying front depth were obtained from near-surface soil moisture contents on the basis of the $\theta(\psi)$ constitutive equation in (7) as if the pressure head resistance to water vapor transfer $r_{\text{min}}$ is shown in Fig. 9 by plotting simulated actual cumulative evaporation depths for different values of $r_{\text{min}}$, namely, 10, 40, 70, and 100 m$^{-1}$ s. It is remarked here that results shown in Figs. 3–7 are obtained allowing only one model parameter, namely, the surface resistance to water vapor diffusion $r_s$ to be calibrated. Simultaneous reproduction of measured actual evaporation rates and near-surface soil moisture contents with a single fitting parameter is a nontrivial task that hardly can be achieved (by means of simplified control volume formulations) without introducing the concept of geometric similarity of soil moisture profiles.
surface resistance to water vapor transfer appeared clearly in Fig. 7, where the recalculated values of topsoil saturation are maintained during the entire event. This referred to a particular field situation in which high degrees of it also should be noted that the available measurements refer to a particular field situation in which high degrees of soil saturation ($\Theta_s > 0.75$). Additional tests also are needed to validate the assumption that the drying front pressure head profile is maintained hydrostatic during the drying event.

As shown in Figs. 8 and 9, simulations may be highly sensitive to soil hydraulic properties and surface resistance to water vapor transfer and thus they must be evaluated in spatially distributed simulations with adequate accuracy (on the basis of the available digital terrain attributes) to provide a reliable description of land surface dynamics during interstorm drying events. Although a more comprehensive model validation is needed, especially to test the roles of surface resistance and exfiltration capacity in extreme soil drying conditions, the results obtained in the present work suffice to warrant a reliable model behavior under humid topsoil conditions, such as those normally offered by the land surface during short interstorm periods. This is important to improve the simulation of land surface dynamics in response to sequences of storms with short interstorms, on one hand, and encourages the progress of the work for improving long-term interstorm simulations on the other hand.

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APPENDIX I. REFERENCES


APPENDIX II. NOTATION

The following symbols are used in this paper:

- $C$ = potential retention of vegetation cover (L);
- $c_p$ = specific heat of air at constant pressure (L$^2$ T$^{-2}$ K$^{-1}$);
- $d$ = zero plane displacement (L);
- $E$ = cumulative evaporation depth (L);
- $e$ = evaporation rate (L T$^{-1}$);
- $e_{av}$ = actual evaporation rate from vegetation cover (L T$^{-1}$);
- $e_p$ = potential evaporation rate (atmospheric evaporative demand) (L T$^{-1}$);
- $F_c$ = cumulative exfiltration depth (L);
- $f_c$ = actual exfiltration rate (L T$^{-1}$);
- $f_{ex}$ = exfiltration capacity (L T$^{-1}$);
- $f_i$ = actual infiltration rate (L T$^{-1}$);
- $G$ = cumulative drainage depth (L);
- $g$ = drainage flux (L T$^{-1}$);
- $K_r$ = relative soil water conductivity (L T$^{-2}$);
- $K_s$ = saturated soil water conductivity (L T$^{-2}$);
- $k$ = von Karman constant (dimensionless);
- $L$ = drying front depth (L);
- $L_e$ = latent heat of vaporization per unit mass of water (L$^2$ T$^{-2}$);
- $p$ = gross precipitation rate (L T$^{-1}$);
- $p_a$ = ambient atmospheric pressure at ground surface (M L$^{-1}$ T$^{-2}$);
- $q$ = Darcian flux (L T$^{-1}$); 
- $R$ = Bowen ratio (dimensionless);
- $r_a$ = aerodynamic resistance to water vapor transfer (L$^{-1}$ T);
- $r_s$ = surface resistance to water vapor transfer (L$^{-1}$ T); 
- $S$ = water content of vegetation canopy (L);
- $S_d$ = soil desorptivity (M T$^{-2}$);
- $T$ = temperature (K, time (T)); 
- $T_d$ = dry-bulb temperature (K);
- $T_e$ = wet-bulb temperature (K);
- $t$ = time (T);
- $Z$ = soil control volume depth (L);
- $z$ = vertical ascending space coordinate (L);
- $z_0$ = roughness length (L);
- $\Gamma(\cdot)$ = function describing shape of moisture profile (dimensionless);
- $\gamma$ = psychrometer constant (M L$^{-1}$ T$^{-2}$ K$^{-1}$);
- $\delta$ = rate of change of saturated vapor pressure with temperature (M L$^{-1}$ T$^{-2}$ K$^{-1}$);
- $\varepsilon$ = vapor pressure at screen height (M L$^{-1}$ T$^{-2}$);
- $\varepsilon(T)$ = saturated vapor pressure at screen temperature $T$ (M L$^{-1}$ T$^{-2}$);
- $\eta$ = pore-size distribution index (dimensionless);
- $\theta$ = volumetric soil water content (dimensionless);
- $\theta_r$ = residual volumetric soil water content (dimensionless);
- $\theta_s$ = saturated volumetric soil water content (dimensionless);
- $\Theta$ = reduced soil moisture content (dimensionless);
- $\mu$ = parameter for canopy water balance (L$^{-1}$);
- $\nu$ = parameter for canopy water balance (L$^{-1}$);
- $\rho_a$ = density of air (M L$^{-3}$);
- $\rho_n$ = density of water (M L$^{-3}$);
- $\sigma$ = specific moisture capacity (L$^{-1}$);
- $\psi$ = soil matrix potential (L); and
- $\psi_s$ = saturated soil matrix potential (L).