Ecohydrology of groundwater-dependent ecosystems: 2. Stochastic soil moisture dynamics

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In groundwater-dependent ecosystems, interactions between rainfall, water table fluctuations, and vegetation are exerted through the soil water content. The dynamics of soil moisture, in fact, are strongly coupled to fluctuations of the water table and, together, they control the overall ecosystem dynamics. We propose here a simple process-based stochastic model for the study of soil moisture dynamics at a generic depth, to complement the stochastic model of water table depth presented in the companion paper. The model presented here is based on a local and depth-dependent water balance driven by stochastic rainfall (marked Poisson noise) and accounting for processes such as rainfall infiltration, root water uptake, and capillary rise. We obtain a semianalytical formulation of the stationary probability distribution of soil water content at different depths, which is studied for different values of soil, climate, and vegetation parameters. The probability distributions are used to investigate the ecohydrology of groundwater-dependent ecosystems, including the quantitative description of the vegetation–water table–soil moisture interplay and the probabilistic analysis of root water uptake.


1. Introduction

The soil water content plays a key role in the dynamics of terrestrial ecosystems. Soil moisture, as well as oxygen, nutrients, and light, is an essential resource for life, and it is a crucial variable in the climate-soil-vegetation dynamical system. In particular, the soil moisture controls most of the hydrological and ecological processes occurring on the earth surface and in the shallow subsurface.

A number of hydrologic processes depend on the soil water content, including rainfall infiltration, percolation, runoff generation, capillarity, groundwater redistribution, and pollutant transport. More specifically, soil moisture affects the hydraulic properties of porous media, thus impacting the mechanisms of water transport within the soil. Many biocological processes are also dependent on soil water content, including plant transpiration, nutrient cycling, microbial activity, and nitrogen/carbon dynamics. Owing to the strong dependence of ecological processes on soil moisture, several feedbacks and interactions exist between hydrological processes and ecosystem functioning in groundwater-dependent ecosystems [Rodriguez-Iturbe et al., 2007]. Soil moisture dynamics and their close relation with the water table depth are of paramount importance for many areas of ecohydrology and thus require an appropriate quantitative modeling.

The probabilistic dynamics of soil moisture driven by the stochastic forcing of rainfall have been studied in the recent past for water-limited ecosystems [e.g., Rodriguez-Iturbe et al., 1999, 2001; Rodriguez-Iturbe and Porporato, 2004]. Simple process-based “bucket” [e.g., Laio et al., 2001] or vertically explicit [Laio, 2006] models have been proposed to study the interactions among precipitation, soil water content and vegetation in arid and semiarid ecosystems. These models do not account for the presence of groundwater, which is considered deep enough to exert no influence on the soil water balance.

In contrast, we focus here on environments with a shallow water table, where the groundwater plays a role in the soil water balance of the root zone, and strong interactions exist between precipitation, groundwater, soil moisture and vegetation. Particularly, we model soil moisture dynamics and investigate their dependence on stochastic rainfall and water table fluctuations. The coupling between water table depth and soil moisture dynamics is mainly due to (1) soil-moisture-dependent processes of water redistribution and groundwater recharge, and (2) capillary flux from the water table to the overlying soil. Owing to the coupling induced by these water fluxes, the ecohydrology of ecosystems with a relatively shallow water tables needs to be investigated accounting for the stochastic dynamics of both soil moisture (present paper) and water table depth [Laio et al., 2009].

Groundwater–soil moisture interactions, associated to capillary flux and moisture redistribution, have been studied in the past without accounting for the role of vegetation on the soil water balance [e.g., Eagleson, 1978; Salvucci and...
Table 1. Values of Soil Moisture at Field Capacitya

<table>
<thead>
<tr>
<th>$s_{fc}$</th>
<th>Sand</th>
<th>Loamy Sand</th>
<th>Sandy Loam</th>
<th>Loam Loam</th>
<th>Silty</th>
<th>Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) $k = 5% , T_p$</td>
<td>0.211</td>
<td>0.301</td>
<td>0.428</td>
<td>0.554</td>
<td>0.595</td>
<td>0.801</td>
</tr>
<tr>
<td>(2) $k = 10% , T_p$</td>
<td>0.236</td>
<td>0.331</td>
<td>0.461</td>
<td>0.587</td>
<td>0.629</td>
<td>0.832</td>
</tr>
<tr>
<td>(3) $\psi = -0.33 , \text{MPa}$</td>
<td>0.208</td>
<td>0.286</td>
<td>0.457</td>
<td>0.583</td>
<td>0.659</td>
<td>0.834</td>
</tr>
</tbody>
</table>

*Calculated as (1) in the present paper and Laio et al. [2009], (2) in the work of Laio et al. [2001], and (3) values given by Rawls et al. [1983]. The potential daily evapotranspiration rate, $T_p$, is equal to 5 mm d$^{-1}$. 

Entekhabi, 1994]. Ridolfi et al. [2008], as well, developed a stochastic model for coupled soil moisture–water table dynamics in bare soil conditions. However, in vegetated sites, plant root uptake plays an important role in the local soil water budget. In this paper, we use simple assumptions on root functioning to model the plant uptake from shallow groundwater and from soil moisture in the unsaturated soil. We develop a deterministic model of soil moisture dynamics near saturation (section 2.1), and couple this model with the stochastic water table dynamics described by Laio et al. [2009], and with the vertically explicit model of long-term average soil moisture dynamics proposed by Laio [2006]. This framework provides the steady state probability density function of the soil water content as a function of depth (section 3), which is investigated for various climate-soil-vegetation parameters (section 4.1).

2. Soil Moisture Profiles

[7] Let us consider a soil column extending from the soil surface ($z = 0$, $z$ axis positive upward), to an indefinite (large) depth. Soil properties such as effective porosity, $n$, grain size distribution, $m$, saturated matric potential, $\psi_s$, and saturated hydraulic conductivity, $k_s$, are supposed to be uniform in space and constant in time. The horizontal area of interest is the plot scale, as in the work of Laio et al. [2009], and it is considered reasonably flat; local heterogeneities are not taken into account and topographic gradients are absent. At the plot scale, the overall amount of root biomass is an exponential function of depth with mean $b$, i.e., $r(z) = 1/b \cdot e^{-bz}$, consistently with a number of field observations [Schenk, 2005]. The vertical root profile is assumed constant in time and no biomass growth or reallocation is considered.

The soil water content is expressed in terms of depth-dependent relative soil moisture, $s(z, t)$ ($0 \leq s \leq 1$), that varies in time, $t$. Soil moisture determines the soil matric potential, $\psi$, and the unsaturated hydraulic conductivity, $k$, through the Brooks and Corey [1964] model, truncated at small values of the hydraulic conductivity,

$$\psi(s) = \psi_s s^{-1/m}; \quad k(s) = \begin{cases} k_s (2 + 3m)/m & \text{if } s < s_{fc}, \\ 0 & \text{if } s \leq s_{fc}, \end{cases}$$

(1)

where $s_{fc}$ is the soil moisture content at field capacity, i.e., the water content held in the soil after gravity drainage [Hillel, 1998]. Operationally, we define $s_{fc}$ as the value of soil moisture at which the unsaturated hydraulic conductivity becomes small (e.g., 5%) compared to the daily rate of potential evapotranspiration, $T_p$ [see also Laio et al., 2001]. From $k(s_{fc}) = 0.05 \, T_p$, one finds

$$s_{fc} = \left( \frac{0.05 T_p}{k_s} \right)^{2/m},$$

(2)

which gives values similar to those obtained with other definitions used in the literature, as shown in Table 1. The operational definition of $s_{fc}$ together with equation (1), leads to neglect fluxes smaller than 0.05$T_p$ in the soil water balance, which is a reasonable simplification.

The soil column is split into three zones according to the local degree of saturation (see Figure 1, left): (1) the saturated zone, where all the voids are filled with water (i.e., $s = 1$), (2) the high-moisture (HM) unsaturated zone, where soil moisture is intermediate between saturation and field capacity (i.e., $s_{fc} \leq s < 1$), and (3) the low-moisture (LM) unsaturated zone, with water content below field capacity.

Figure 1. (left) Example of the soil moisture profile $s(z)$ in a soil column, where low-moisture, high-moisture, and saturated zones are indicated. (right bottom) Time series of daily soil moisture dynamics in a fixed soil layer, with (right top) corresponding rainfall events.
(i.e., $s < s_c$). The surface separating saturated and unsaturated soil lays at depth $y(t)$ (where $t$ is time) from the soil surface and has a soil matric potential equal to $\psi_s$. The surface at zero pressure, i.e., the water table, lays at constant distance $|\psi_s|$ from $y(t)$, that is at a depth $h(y, t) = y(t) + \psi_s$, and has been extensively investigated by Laio et al. [2009].

[10] The soil moisture in the high-moisture unsaturated zone is determined from the balance of rainfall infiltration/redistribution, plant root uptake, and capillary flux from the saturated zone. In the low-moisture zone, instead, the soil water content is small ($s < s_c$) and the hydraulic conductivity is assumed to be negligible (see equation (1)). For this reason, the upward capillary flux does not reach the LM zone and it ceases at a depth $h(y, t)$ above $y$, where the soil moisture equals $s_c$. The surface at field capacity, i.e., the boundary between LM and HM zones, is thus called “zero-flux surface” due to the absence of upward flux across the depth $h$. The zero-flux surface marks the upper limit of groundwater influence on the soil water balance, and the soil moisture dynamics in the LM zone turns out to be independent of the presence of the water table.

[11] When the water table is near the soil surface, the moisture at $z = 0$ is higher than field capacity and there is no low-moisture zone in the soil column. These shallow water table (SWT) conditions occur when the position of the separation surface, $y$, is shallower than a critical depth, $y_c$, corresponding to a soil surface at field capacity. The critical depth is calculated as [Laio et al., 2009]

$$y_c = \psi_s F_1 \left[ \frac{1}{2 + 3m}, 1, 1 + \frac{1}{2 + 3m}, \frac{v^m}{k_s} \right]$$

$$- \psi_s F_1 \left[ \frac{1}{2 + 3m}, 1, 1 + \frac{1}{2 + 3m}, -\frac{v^m}{k_s} \right],$$

(3)

where $\psi_s$ is the soil matric potential at field capacity, obtained by setting $s = s_c$ into equation (1), $F_1[a, b, c, x]$ is the hypergeometric function, and $v^m = \frac{1}{1 + 0.35 - 0.65m}$. In conditions of deep water table (DWT), i.e., $y < y_c$, the depth, $h$, of the zero-flux surface is a function of the position of the separation surface. An approximate relationship between $h$ and $y$ is given by [Laio et al., 2009],

$$h(y) = \begin{cases} 0 & \text{if } y \geq y_c, \\ (1 - A^{1/4})(y - y_c) - A^2(1 - A^{-1/4}) & y < y_c, \end{cases}$$

where $A = \frac{-y_c + \psi_s - \psi_c}{5b + \psi_c - \psi_s}$.

[12] Owing to the fluctuations of $y$ and $h$, the soil moisture dynamics in a given layer at depth $z$ result from the combination of (1) low-moisture phases, with moisture dynamics, $s(z, t) = s(z, t)$ conditional upon $s < s_c$, independent of groundwater dynamics; (2) high-moisture phases, $s(z, t) = s_c(z, t)$ conditional upon $s > s_c$, with moisture dynamics deterministically related to the position of $y$; and (3) saturated conditions, when $s(z, t) = 1$. The decoupling of soil moisture dynamics in the LM and HM zones allows one to analyze the two zones separately, as detailed in the following sections.

### 2.1. Equilibrium Soil Moisture Profiles in the High-Moisture Unsaturated Zone

[13] The soil moisture gradient establishes an upward capillary flux from the saturated zone, which is depleted by plant root uptake while moving toward the soil surface. Above the zero-flux surface, the capillary flux becomes negligible due to the null hydraulic conductivity in the low-moisture zone (see equation (1)). Exfiltration, i.e., the total upward flux leaving the saturated zone, needs thus to be balanced by root uptake in the high-moisture unsaturated zone. From this balance, and considering that root uptake at depth $z$ is $U(z) = T_p F(z)$, we obtain that the net capillary flux, $v(z)$ (with $y < z < h$), can be calculated as

$$v(z) = \begin{cases} 0 & \text{if } y \geq y_c, \\ T_p \int_0^h r(z) dz = T_p \left( 1 - e^{2zh} \right) & \text{if } y < y_c, \end{cases}$$

(5)

where the two cases correspond respectively to the shallow and deep water table conditions.

[14] The soil moisture dynamics in the HM zone $s_c(z, t)$ are modeled, at the daily timescale, as a sequence of equilibrium states and the time dependence is dropped. This simplification relies on the assumption of an instantaneous redistribution (at the daily timescale) of moisture within the soil profile. The assumption is allowed in coarse-grained soils by the large values of hydraulic conductivity, and in fine-grained soils by a field capacity close to saturation, which leads to small moisture gradients within the high-moisture zone. The steady state water balance in a generic layer in the HM unsaturated zone is then described by the Darcy’s law, i.e.,

$$-k(s)(\frac{dv}{dz}) + 1 = v(z),$$

(6)

if $y \geq y_c,$

$$\frac{dv}{dz} = \frac{m}{k(s)} = \frac{m}{k(s_c)}$$

if $y < y_c,$

where $k(s_c) = s_c(z)$ is the steady state soil moisture profile for $s > s_c$. Using equation (1) to express $\frac{dv}{dz}$ in equation (6), one obtains the differential equation

$$\frac{dv}{dz} = \frac{m s_c^{(m+1)/m}}{k(s_c) + v(z)}.$$

(7)

representing the dependence of $s_c(z)$ on the water table depth $y$ and on soil/vegetation properties. Solving equation (7) requires to specify a boundary condition and the capillary flux $v$ in the corresponding position.
Comparison between exact (solid) numerical solution of equation (7) and approximate (dashed) equilibrium soil moisture profiles in the high-moisture zone, for shallow and deep water table conditions ($y = 0.5, 1, 1.5$ and $2$ times the critical depth, $y_c$), in (a, b) loamy sand and (c, d) loam, with $b = 10$ cm (Figures 2a and 2c) and $b = 40$ cm (Figures 2b and 2d).

Since no analytical solutions can be found for the differential equation (7), a numerical solution is computed with a finite difference method. For SWT, i.e., $y > y_c$, the boundary condition $s_y(t) = 1$ is considered. In this case, the first step of the numerical integration is carried out using the exfiltration rate $v(y) = T_p(1 - e^{-b})$ (equation (5)). For DWT, the same boundary condition would require $v(y) = T_p(e^{bh} - e^{-b})$ at the first computational step, with $h = h_y$ approximated by equation (4). However, a more accurate numerical integration for the case of DWT is obtained by imposing the boundary condition $s_y(h) = s_y$ at the zero-flux surface, where $v(h) = 0$. This approach allows one to avoid the approximation error introduced by $h_y$ and allows for an exact numerical solution for $s_y$.

We now seek an explicit analytical function approximating the steady state soil moisture profile in the high-moisture unsaturated zone, $s_y(z)$. This function should (1) have one (SWT) or two (DWT) fixed points ($s_y(t) = 1$ and $s_y(t) = s_y$); (2) correctly describe the soil moisture profile across a broad range of soil types and root profiles; (3) be explicit in $s_y$, in order to allow one to analytically calculate the specific yield, $\beta$, to be described later; and (4) be invertible in $y$ even when $h_y$ (equation (4)) is used, as required for the computation of the probability density functions of $s_y$ (see following section 3).

After extensive explorations, we found that a good approximation valid for a wide range of soils, root profiles, and water table depths, was obtained when the soil matrix potential in the high-moisture unsaturated zone is modeled with a quadratic function. By imposing the fixed points ($\psi_y$ and $\psi_y$) at the extremes of the high-moisture unsaturated zone, one finds

$$
\psi(z) = \begin{cases} 
\psi_y \left[ 1 + \left( s_{c}^{-1/2} - 1 \right) \left( \frac{y - z}{y_c} \right) \right]^2 & \text{if } y \geq y_c, \\
\psi_y \left[ 1 + \left( s_{c}^{-1/2} - 1 \right) \left( \frac{z - y}{y_c} \right) \right]^2 & \text{if } y < y_c.
\end{cases}
$$

Substituting equation (1) into equation (8) we obtain the following soil moisture profiles for the HM zone,

$$
s_y(z) = \begin{cases} 
1 + \left( s_{c}^{-1/2} - 1 \right) \left( \frac{y - z}{y_c} \right)^{-2m} & \text{if } y \geq y_c, \\
1 + \left( s_{c}^{-1/2} - 1 \right) \left( \frac{z - y}{y_c} \right)^{-2m} & \text{if } y < y_c.
\end{cases}
$$

which can be solved for $y$, as required in the mathematical developments presented in section 3. Examples of the correct (numerical) and approximate soil moisture profiles are shown in Figure 2 for different depths of the separation surface. The approximation is overall very good, particularly for deep-rooted vegetation and fine-grained soils.

This explicit soil moisture profile, $s_y(z)$, can be used to calculate the specific yield, $\beta$, which is the volume per unit area of water an aquifer releases or takes into storage (per unit aquifer area) per unit change in water table depth [e.g., Nachabe, 2002]. $\beta$ is obtained from the difference between two integrated soil moisture profiles corresponding to two infinitesimally distant positions of the separation surface between saturated and unsaturated soil, i.e.,

$$
\beta(y) = \begin{cases} 
\int_{y}^{1} \frac{\partial \zeta(z)}{\partial y} ds & \text{if } y \geq y_c, \\
\int_{y_c}^{1} \frac{\partial \zeta(z)}{\partial y} ds, & \text{if } y < y_c,
\end{cases}
$$

where $s_y(0)$ is the soil moisture at the soil surface in SWT conditions. Integrating the approximate profile (equation (9)), one obtains the relations for $\beta$ used in the stochastic model for water table dynamics proposed in the companion paper [Laio et al., 2009].

2.2. Long-Term Average Soil Moisture Profiles in the Low-Moisture Unsaturated Zone

The soil moisture dynamics in the low-moisture zone, $s(z, t) = s'(z, t)$, is decoupled from the groundwater dynamics, thanks to the negligible capillary flux through the surface at depth $h$. Therefore, soil moisture dynamics in the LM zone (conditional upon $s < s_{c}$) can be studied as in the case of water-limited ecosystems [e.g., Laio et al., 2001; Laio, 2006], without considering the presence of the water table. In this case, the local water balance in the LM zone can be written as

$$
n \frac{\partial s'(z, t)}{\partial t} = Q(z, t) - U_{lw}(z, s'),
$$

where $U_{lw}(s', z)$ is the root uptake at depth $z$, and $Q(z, t)$ is the stochastic process of infiltrating events reaching the $z$ layer, detailed hereafter.

The net rainfall infiltration is modeled as a marked Poisson process with rate $\lambda_0$ and rainfall depths exponen-
tially distributed with mean \( \alpha \), i.e., \( \mathcal{P}(\lambda_0, \alpha) \). Every rainfall event infiltrates (at the daily timescale) as an instantaneous piston flow that initially saturates soil layers, and then drains them to field capacity, while excess water percolates down to deeper layers. The water content in soil layers not reached by the wetting front remains unmodified. Assuming that rainfall always find the same long-term average soil moisture in the soil, \( \bar{s}'(z) \), the Poissonian structure of wetting events is preserved, with good approximation for shallow layers [see Laio, 2006]. The sequence of wetting events at generic depth \( z < h \) (equation (11)) is then a state-dependent marked Poisson process with rate \( \lambda(z, \bar{s}'_m(z)) \) and exponentially distributed depths with mean \( \alpha \), i.e., \( Q(z, \alpha) = \mathcal{P}(\lambda(z, \bar{s}'_m(z)), \alpha) \). The interarrival rate at depth \( z \) is given by

\[
\lambda[z, \bar{s}'_m(z)] = \lambda_0 \exp\left[\frac{nz(s_k - \bar{s}'_m(z))}{\alpha}\right],
\]

where \( \bar{s}'_m(z) \) is the mean above \( z \) of the long-term average soil moisture, conditional upon \( z > h \), i.e., \( \bar{s}'_m(z) = \frac{1}{z-h} \int_0^z \bar{s}'(w) \, dw \).

[21] The root uptake in the low-moisture zone can be expressed as a function of depth \( z \) and depends on the root density and soil moisture at that depth,

\[
U_m(s', z) = T_p \rho(s') r(z) = \frac{s'(z) - s_w}{s_k - s_w} r(z),
\]

where \( \rho(s') \) is the water stress function, which is assumed to vary linearly from 0 at the wilting point, \( s_m \), to 1 at field capacity, \( s_k \) [e.g., Laio, 2006].

[22] Inserting equation (13) in the local water balance of the low-moisture zone (equation (11)), taking the long-term average and setting \( \frac{d\bar{s}'_m(z)}{dz} = 0 \), one has the long-term average steady state water balance of the generic \( z \) layer of depth \( dz \), i.e.,

\[
\lambda[z, \bar{s}'_m(z)] n [s_k - \bar{s}'(z)] dz = T_p \frac{\bar{s}'(z) - s_w}{s_k - s_w} r(z) dz.
\]

The term on the left-hand side is the average incoming flux, given by the rate, \( \lambda[z, \bar{s}'_m(z)] \), of infiltration events reaching the \( z \) layer multiplied by the average fraction of water storage available for gravity drainage, \( n(s_k - \bar{s}'(z)) dz \).

[23] Combining equation (14) with equation (12), one obtains

\[
\lambda(n s_k - \bar{s}'(z)) \exp \left[ \frac{ns_k - s_k}{\alpha} \right] \exp \left[ \int_0^z \frac{1}{s_k - s_w} \frac{\bar{s}'(u)}{\alpha} du \right] = T_p \frac{s_k(z) - s_w}{s_k - s_w} b \exp^{s'_{0}z/b},
\]

Taking the natural logarithm of equation (15), and its derivative with respect to \( z \), we find

\[
n s_k - \frac{1}{b} - \frac{n s_k}{\alpha} = \left[ \frac{1}{\bar{s}'(z) - s_w} + \frac{1}{s_k - \bar{s}'(z)} \right] \frac{d\bar{s}'(z)}{dz}.
\]

Equation (16) above can be rearranged and integrated to give

\[
z = \frac{b \alpha}{n b (s_k - s_w) - \alpha} \log \left[ \frac{\bar{s}'(z) - s_w}{s_k - s_w} + b \log \left[ \frac{s_k - \bar{s}'(z)}{s_k - s_0} \right] \right] - \frac{n b^2 (s_k - s_w) - \alpha}{n b (s_k - s_w) - \alpha} \log \left[ \frac{n b (s_k - s_0) - \alpha}{n b (s_k - s_0) - \alpha} \right],
\]

where the boundary condition \( \bar{s}'(0) = \bar{s}'_0 \) has been used. The value of \( \bar{s}'_0 \), i.e., the long-term average soil moisture at the soil surface, can be found by setting \( z = 0 \) in equation (15), finding

\[
\bar{s}'_0 = s_k - \frac{T_p (s_k - s_w)}{T_p + nb \lambda_0 (s_k - s_w)}.
\]

By solving numerically equation (17) one finds the explicit long-term average soil moisture profile, \( \bar{s}'(z) \), in the low-moisture zone. The numerical profile is necessary to calculate the depth-dependent rate of rainfall infiltration, \( \lambda(z, \bar{s}'_m(z)) \), using either equation (12) with the definition of \( \bar{s}'_m(z) \), or equation (14) with an exponential root profile, i.e.,

\[
\lambda[z, \bar{s}'_m(z)] = \frac{T_p (\bar{s}'(z) - s_w) e^{z/b}}{nb (s_k - s_w) (s_k - \bar{s}'(z))}.
\]

The rate of infiltration events reaching the zero-flux surface is \( \lambda(h, \bar{s}'(h)) \), which thus represents the rate of recharge events (i.e., events contributing to the water table fluctuations) in DWT conditions. This is a key point in the modeling of stochastic water table dynamics presented in the companion paper [Laio et al., 2009].

2.3. Variability of the Long-Term Average Moisture Profiles in the Low-Moisture Zone

[24] The long-term average soil moisture profile in the low-moisture zone, \( \bar{s}'(z) \), can have two different behaviors: it can be either increasing or decreasing with depth. The type of behavior depends on the model parameters and can be studied through the sign of \( \frac{d\bar{s}'(z)}{dz} \) using equation (16). For an exponential root distribution and a consequently monotone soil moisture profile, we find that the long-term average profile increases with depth if

\[
n b (s_k - \bar{s}'_0) - \alpha < 0,
\]

where \( \bar{s}'_0 \) is defined in equation (18).

[25] The condition in equation (20) can be interpreted in terms of a critical mean rooting depth, \( b_0 \),

\[
b_0 = \frac{T_p \alpha}{n (s_k - s_w) (T_p - \lambda_0 \alpha)}.
\]

When \( b < b_0 \) (\( b > b_0 \)), the long-term average soil moisture profile, \( \bar{s}'(z) \), increases (decreases) with depth, as shown in Figure 3. When \( b = b_0 \), the long-term average soil moisture profile is uniform, i.e., \( \bar{s}'(z) = \text{const} = \bar{s}'_0 \) (equation (18)). This uniform soil moisture profile has been postulated by Laio et al. [2006] as resulting from a strategy of optimal root allocation in semiarid ecosystems. In this paper, however, we investigate soil moisture dynamics in more general climate/vegetation conditions. Thus we do not set the mean root depth to \( b_0 \) but consider \( b \) as a free parameter.

[26] As shown in Figure 3, if \( b \) is larger than \( b_0 \), \( \bar{s}'(z) \) decreases with depth and approaches \( s_m \) in the deepest soil layers. This pattern is due to the fact that deep roots extract...
water from deep soil layers, which are reached by few wetting events, leading to dry long-term soil conditions. Conversely, if $b$ is smaller than $b_0$, the profile $s'(z)$ increases with depth and tends to $s_f$. In these conditions most of the roots are shallow but incapable of capturing all of the infiltrating rainfall, a fraction of which drains to deeper layers, thereby maintaining wetter soil conditions.

The condition in equation (20) can also be interpreted in terms of critical rainfall parameters, for a fixed root distribution (see Figure 4). A rainfall increase induces a shift in the long-term average moisture content of deeper soil layers, even though the surface values do not change sensibly. In particular, we observe that $s_0' \text{ (equation (18))}$ depends only on the mean interarrival time, $l_0$, of rainfall events and not on the mean rainstorm depth. In fact, every (net) rainfall event wets the soil surface, regardless of the amount of precipitation. The moisture at the soil surface thus depends only on the average waiting time between two rainfall events, i.e., $\lambda_0$.

3. Probability Density Functions

[28] The steady state soil moisture profiles in the HM zone and the long-term average soil moisture profiles in the LM zone presented in the previous section allow us to determine the steady state probability density function (pdf) of soil moisture at generic depth. This probability distribution also depends on the position, $y$, of the separation surface and the zero-flux surface, $h$. Equations defining the pdfs of $y$ and $h$ are thus briefly recalled from the companion paper by Laio et al. [2009], before deriving the depth-dependent pdfs of soil moisture.

3.1. Probability Distribution of the Depths $y$ and $h$

[29] The stochastic soil water balance expressing groundwater dynamics involves fluxes to and from the saturated

Figure 3. Examples of (left) different vertical root distributions and (right) the corresponding long-term average soil moisture profiles in the low-moisture zone (conditional upon $s' < s_f$). In Figure 3 (right), the soil is a loamy sand, vegetation has a mean root depth $b = 30 \text{ cm}$, and rainfall parameters are $\lambda_0 = 0.2 \text{ d}^{-1}$ and $\alpha = 1.5 \text{ cm}$. The limit value of the mean root depth, $b_0$, is $49 \text{ cm}$.

Figure 4. Impact of rainfall parameters (left) $\lambda_0$ and (right) $\alpha$ on the long-term average soil moisture profile in the LM zone, conditional upon $s' < s_f$. Soil and vegetation are the same as in Figure 3, while rainfall parameters are $\alpha = 1.2 \text{ cm}$ (Figure 4, left) and $\lambda_0 = 0.2 \text{ d}^{-1}$ (Figure 4, right).
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\[ \frac{\partial y}{\partial t} = f(y) + g(y)\xi(y, t), \]  
\[ f(y) = \begin{cases} 
\frac{k_i(y_0 - y - \psi_c) - T_p}{n[1 - s_i(0)]} & \text{if } y \geq y_c, \\
\frac{k_i(y_0 - y - \psi_c) - T_p e^{\alpha h}}{n[1 - s_i(0)] + B(\frac{\partial y}{\partial t}) - 1} & \text{if } y < y_c;
\end{cases} \]

\[ g(y) = \begin{cases} 
1 & \text{if } y \geq y_c, \\
\frac{1}{n[1 - s_i(0)] + B(\frac{\partial y}{\partial t}) - 1} & \text{if } y < y_c;
\end{cases} \]

\[ \xi(y, t) = \begin{cases} 
\mathcal{P}(\lambda_0, \alpha) & \text{if } y \geq y_c, \\
\mathcal{P}(\lambda_0, \alpha \exp\left[\frac{\psi_c - y}{s_i(0)}\right], \alpha) & \text{if } y < y_c.
\end{cases} \]

Here \( y_0 \) is the depth of external water body, \( k_i \) is the constant of proportionality for the lateral flow, \( s_i(0) \) is the soil moisture at the soil surface in SWT conditions, and

\[ B = \frac{1 - 2m}{1 - \gamma_c} \left(\frac{1 - \gamma_c}{1 - \gamma_c} \right) + 2ms_i\gamma_c. \]

[30] All parameters in equation (23) above are either constant, or can be represented analytically as functions of \( y \) or \( h \), with \( h \) being, in turn, a function of \( y \) through equation (4). Therefore equation (22) represents an univariate (in \( y \)), stepwise-continuous first-order stochastic differential equation, where the random forcing is a state-dependent multiplicative compound Poisson noise. Equation (22) can be solved under steady state conditions to determine the probability density function (pdf) of the variable \( y \). The solution for the case of a Poisson process with constant rate, \( \lambda \), and zero mean is given by Van Den Broeck [1983].

[31] In the case of a generic state-dependent noise with rate \( \lambda(y) \), and a noise average \( \lambda(y) \), the probability density function of \( y \) can be expressed [D’Odorico and Porporato, 2004; Porporato and D’Odorico, 2004] as

\[ p_f(y) = \frac{C}{f(y)} \exp\left[ -\int_0^y f(u) + \alpha \lambda(y) g(u) \, du \right]. \]  

where \( C \) is a normalization constant obtained by setting \( \int_{-\infty}^0 p_f(y) \, dy = 1 \).

[32] The probability density function of \( h \) can be determined as a derived distribution of \( p_f(y) \),

\[ p_f(h) = p_f[y(h)] \left(\frac{dh}{dy}\right)^{-1}. \]  

with \( y(h) \) obtained from equation (4). An atom of probability appears in the distribution of \( h \) at \( h = 0 \), with associated mass, \( p_{h,0} \), corresponding to the probability of being in SWT conditions, i.e., \( P_{h,0} = \int_0^0 p_f(y) \, dy \).

The properties of the steady state probability distributions of \( y \) and \( h \) are investigated in the following sections.

3.2. Probability Density Functions of Soil Moisture

[33] The probability density function of soil moisture at a given depth \( z \) results from the combination of three different contributions, with relative weights, \( W_i, i = 1, 2, 3 \). These contributions correspond to the three states that may occur at depth \( z \): LM conditions (\( h < z \leq 0 \), with weight \( W_{S,3}(z) \)), HM conditions (\( y < z \leq h \), with weight \( W_{S,2}(z) \)), and saturated conditions (\( z \leq y \), with weight \( W_{S,1}(z) \)).

[34] The three weights are obtained from the cumulative density functions of \( y \) and \( h \), as sketched in Figure 5. The probability for a soil layer at depth \( z \) to be saturated equals the probability that \( y \geq z \), i.e.,

\[ W_{S,1}(z) = \int_z^0 p_f(u) \, du, \]  

with \( p_f(u) \) given by equation (24). The probability for the same layer to be in the high-moisture unsaturated zone depends on the probability distributions of both \( y \) and \( h \), i.e.,

\[ W_{S,2}(z) = \int_z^0 p_f(h) \, dh - W_{S,1}(z), \]  

with \( p_f(h) \) given by equation (25). Finally, in order to have a total weight equal to one, the probability associated with low-moisture unsaturated conditions is

\[ W_{S,3}(z) = 1 - \int_z^0 p_f(u) \, du = 1 - W_{S,1}(z) - W_{S,2}(z). \]
The three portions of the overall probability distribution of $s$ at the generic depth $z$ are calculated separately as follows. The probability for the $s$ layer to be saturated is an atom of probability centered at the value $s = 1$, whose associated probability mass is $W_s(s, z)$. The probability density function of soil moisture in a layer of the HM zone is obtained as a derived distribution of $y$. In fact, using equation (4) to express $h(y)$ in equation (9), one obtains the relation between $y$ and the soil moisture profile in the HM zone, $s_s(z)$. Solving this relation for $y$ gives

$$y(s_s, z) = \begin{cases} 
  z - y_c \left( \frac{s_0^{1/2m} + 1}{s_c^{1/2m} + 1} \right) \\
  a_1 - \sqrt{a_1^2 - 4a_0a_2} \left( 2a_2 \right) \\
  z + \left( \psi_c - \psi_s \right) \left( \frac{s_0^{1/2m} - 1}{s_c^{1/2m} - 1} \right)
\end{cases}$$

where $a_0$, $a_1$ and $a_2$ are

$$a_0 = D \psi_c^2 - (A^{1/4} - 1) \left( \frac{s_0^{1/2m} - 1}{s_c^{1/2m} - 1} \right) y_c + \left( \frac{s_0^{1/2m} - 1}{s_c^{1/2m} - 1} \right) z,$$

$$a_1 = -2D \psi_c + A^{1/4} \left( \frac{s_0^{1/2m} - 1}{s_c^{1/2m} + 1} \right) - s_0^{1/2m} + 1,$$

$$a_2 = D,$$

with $D = \left( \frac{A^{1/4 - 1} \psi_c^2 (1 - s_0^{1/2m})}{s_0^{1/2m} + 1} \right)$ and $A = \frac{-\psi_c + y_c - \psi_s}{-\psi_c + y_c + \psi_s}$ (as in equation (4)). Equation (29) is then used to obtain the derived probability density function of soil moisture, $p_{S,2}(s_s, z)$, conditional upon the HM state of layer at depth $z$, i.e.,

$$p_{S,2}(s_s, z) = \int \frac{y(s_s, z)}{W_s(z)} \, ds_s,$$

where $W_s(z)$ is the normalizing constant for a unitary area in the interval $(s_c, 1)$.

The probability density function of soil moisture in a layer of the LM zone (i.e., $s(z) = s(z)$ conditional upon $s < s_c$), can be calculated as by Laio [2006], without dependence on the water table (see section 2.2). This function reads

$$p_{S,3}(s'|z) = \frac{n(s'; z, s_c)}{r(z) T_p} \left( \frac{s}{s_c - s} \right)^{s_c / (s_c - s)} \, e^{-s / (s_c - s)}$$

where $r(z)$ is the exponential vertical root distribution.

In summary, the procedure for the computation of the pdf of soil moisture at a given depth, $z^*$, is the following:

1. The long-term soil moisture profile in the low-moisture zone (section 2.2) is used to compute the probability density function of the position of separation surface, $y$, and zero-flux surface, $h$ (equations (24) and (25)).

2. In turn, these pdfs determine the weights of the three soil moisture states at depth $z^*$, i.e., $W_s(z^*)$ (equation (26)), $W_{S,2}(z^*)$ (equation (27)), and $W_{S,3}(z^*)$ (equation (28)).

4. Results and Discussion

The model presented in the previous sections allows one to evaluate the stationary probability distribution of soil moisture at different depths, as a function of climate, vegetation and soil properties. Since the proposed model is based on assumptions and simplifications, one should study the outcome as a source of general information, conveyed by main features of the pdfs, i.e., mean, variance, shape, limits. Details such as jumps and discontinuities (especially in the pdfs of $s$ and $y$) do not reflect reality, but depend on the reasonable and consistent approximations introduced, which are discussed where necessary.

We now apply this model to investigate the soil moisture dynamics for different parameter values. In particular, we show examples of soil moisture pdfs, of long-term average profiles of soil moisture and pdfs of root uptake in different environmental conditions.

4.1. Probability Distributions for Different Soil, Climate, and Vegetation Parameters

We first investigate the case of a deep water table with small fluctuations (see Figure 6). This is the case, for example, of a deep external water body and a coarse-grained soil associated with a relatively dry rainfall regime. The water table and the zero-flux surface are for most of the time below the bulk of the root zone (considered as the zone containing 95% of the root profile) and the moisture dynamics in shallow layers are not affected by the ground-water. As a consequence, the pdfs of soil moisture in shallower soil layers (Figure 6, $z_1$ and $z_2$) show the same behavior obtained by Laio [2006, Figure 3] in the case of water-limited ecosystems. Deeper layers exhibit higher soil water contents, reflecting the intermittent presence of the
Figure 6. Probability density functions of (top) the positions of the separation between saturated and unsaturated soil, $y$ (continuous line), and the zero-flux surface, $h$ (dashed line), and (bottom) probability density functions of soil moisture at different depths. The soil is a loamy sand (parameters as in the work of Rawls et al. [1983]); rainfall parameters are $\lambda_0 = 0.3 \, \text{d}^{-1}$ and $\alpha = 2 \, \text{cm}$, vegetation parameters are $T_p = 0.5 \, \text{cm/d}$ and $b = 30 \, \text{cm}$; while $k_l = 3.7 \cdot 10^{-3} \, \text{d}^{-1}$ and $y_0 = 3 \, \text{m}$.

Figure 7. Probability density functions of (top) $y$ (continuous line) and $h$ (dashed line), and (bottom) probability density functions of soil moisture at different depths. Soil is a loamy sand, and the external water table depth is $y_0 = 2 \, \text{m}$; other parameters are as in Figure 6.
discontinuity at \( y_c \), due to the discontinuity in the specific yield \( \beta \) (equation (10)). This discontinuity also affects the derived probability distribution of soil moisture in the high-moisture zone, as shown in Figure 8, \( z_1 \) and \( z_2 \). In even deeper soils, the atom of probability at saturation grows, until it reaches a unit value at the deepest water table position.

4.2. Overall Long-Term Average Soil Moisture Profiles

In section 2.2 we introduced the long-term average soil moisture profile in the low-moisture zone, \( \bar{s}(z) \), as the result of soil moisture dynamics conditional on \( s < s_c \). The overall long-term average soil moisture profile is the result of the combination of three states (low moisture, high moisture, and saturated) occurring intermittently in a fixed soil layer. This profile is the expected value of the piecewise probability density function as a function of depth \( z \), which reads

\[
\bar{s}(z) = W_{S,1}(z) + W_{S,2}(z) \int_{s_c}^{1} s p(s,z) \, ds + W_{S,3}(z)s(z). \tag{34}
\]

The overall profile of expected values, \( \bar{s}(z) \), can be compared with the long-term average low-moisture profile \( \bar{s}(z) \) (equation (17)), which is the benchmark case without groundwater, under the same climatic conditions. Figure 9 shows the comparison between the two profiles for the cases presented in Figures 6, 7, and 8.

The two profiles overlap in the shallow soil layers, where only rainfall and root uptake play a role in the water balance (Figures 9a and 9b). At depths \( z \) where the presence

Figure 8. Probability density functions of (top) \( y \) (continuous line) and \( h \) (dashed line), and (bottom) probability density functions of soil moisture at different depths. Soil is a loam (soil parameters as in the work of Rawls et al. [1983] and \( k_i = 7.9 \cdot 10^{-4} \text{ d}^{-1} \), \( y_0 = 2 \text{ m} \), and other parameters are as in Figure 6.

Figure 9. Expected average soil moisture profiles in groundwater-dependent (solid line) and independent (bold line) ecosystems, with the standard deviation (dashed line) corresponding to the case with water table. Model parameters are the same (a) as in Figure 6, (b) as in Figure 7, and (c) as in Figure 8.
of the zero-flux surface above \( z \) becomes more frequent, the influence of the water table increases, and the two profiles split, as the groundwater-dependent profile tends to saturation, i.e., to \( s = 1 \). The smaller the variance of the pdf of \( y \) and \( h \) is, the more abrupt is the shift between the groundwater-dependent profile and the no-groundwater profile. If the water table is shallow for most of the time and the pdf of \( h \) has an atom of probability at \( z = 0 \), the two long-term profiles are already separated at the soil surface (Figure 9c). The groundwater-dependent pdfs also have the maximum standard deviation at depths where the gradient of average soil moisture with the depth is larger.

By comparing the overall long-term soil moisture profiles, one can highlight the role of different parameters on the soil water balance. An example is the variability of the mean profile with the rainfall regime (Figure 10): more abundant precipitation lead to shallow water table, shallower permanent saturated soil conditions, and wetter soil surface. Another example, shown in Figure 11, highlights the strong effect of different mean root depths, \( b \), on the long-term water balance. Roots concentrated in the upper portion of the soil column maintain a relatively dry soil surface and allow shallow layers beneath the bulk of the root zone to have permanently saturated conditions. Conversely, deep roots contribute to lower the soil moisture along the whole soil column, keeping the permanently saturated conditions only in deep soil layers. Because the deepest mean root depth used in Figure 11 is larger than the value \( b_0 \) (equation (21)), the long-term soil moisture profile initially decreases with depth as discussed in section 2.3, before increasing due to the influence of the water table.

Notice that the curves shown in Figure 11 differ only for the vertical distribution of roots and not for parameters directly related to inflow or outflow (e.g., rainfall or lateral flux). The different water storages outlined by these profiles are determined by a different total plant water uptake in the three cases driven by the different vertical distribution of roots. A high root density in the upper soil layers quickly depletes the water available, establishing dry conditions that favor plant water stress. The assumption of noncooperative rooting system and the lack of deeper roots do not allow for groundwater uptake. Thus, despite the presence of a shallow groundwater, the total water uptake is low. An evaluation of total root uptake (see details in section 4.3) in the case of Figure 11 with \( b = 10 \) cm gives a mean value of 0.35 cm/d, versus a maximum value of \( T_p = 0.5 \) cm/d. Conversely, roots with a deeper vertical distribution (higher \( b \)) can take full advantage of the moisture supplied both by groundwater and precipitation, thereby avoiding severe water stress conditions. In fact, when \( b = 50 \) cm (Figure 11) the average value of the total root uptake is equal to 0.47 cm/d.

### 4.3. Probability Density Function of Root Uptake

The probability density function of root water uptake is a useful indicator of the health status of vegetation, of the long-term level of water stress, and of the main source of water for plants (e.g., groundwater or rainfall). The pdf of root uptake can be evaluated from the probability distribution of the separation surface, \( y \), zero-flux surface, \( h \), and soil moisture at every soil depth.

Root uptake takes place in the low-moisture, high-moisture, and saturated zones. Since uptake from the high-moisture unsaturated zone is balanced by the exfiltration from the saturated zone, one can quantify the amount of groundwater, \( U_{\text{wt}}(h) \), extracted by the plant as the sum of

![Figure 10. Expected average soil moisture profiles (in black) and their standard deviations (in gray) for a groundwater-dependent ecosystem, with different rainfall regimes (\( \lambda_0 \)) in a loamy sand. Other parameters are \( \alpha = 2 \) cm, \( T_p = 0.5 \) cm/d, \( b = 30 \) cm, \( y_0 = -3 \) m, \( k_l = 3.7 \cdot 10^{-3} \) d\(^{-1} \).](image1.png)

![Figure 11. Expected average soil moisture profiles (in black) and their standard deviations (in gray) for a groundwater-dependent ecosystem with different types of vegetation (mean root depth) in a loam. Other parameters are \( \lambda_0 = 0.25 \) d\(^{-1} \), \( \alpha = 1.5 \) cm, \( T_p = 0.5 \) cm/d, \( y_0 = -2 \) m, \( k_l = 7.9 \cdot 10^{-4} \) d\(^{-1} \).](image2.png)
The density functions of groundwater uptake in the three cases analyzed in Figures 6, 7, and 8, are shown in Figure 12 with dotted lines. The atom of probability at 0.5 corresponds to the probability of being in SWT condition, where the plant does not suffer from water stress and root uptake from groundwater equals $T_p$. 

[53] The root water uptake in a generic layer of the low-moisture zone at depth $z$ is a function of the local soil moisture, through equations (13). One can thus compute the inverse function $s'(U_{lm, z}) = s_w + \frac{U_{lm}[s(z)]}{T_p[z]} (s_{fc} - s_w)$ and determine the $z$-dependent probability distribution of root uptake conditional upon $s_w < s' < s_{fc}$ as a derived distribution of $p_{S_{lm}}(s[z])$ (equation (32)). The resulting pdf is given by

$$ p_{U_{lm}}(u(z)) = \frac{n\lambda[z, s_m'(z)][s_{fc} - s_w]}{T_p^2 r(z)^2} \left( \frac{U_{lm}[s(z)]}{T_p r(z)} \right)^{s'(z) - s_{w}} e^{-\sum_{l=1}^{m} n\lambda[z, s_m'(z)][s_{fc} - s_w] u(z)^{s'(z) - s_{w} - 1}}, $$

which, however, cannot be analytically translated into the probability distribution of the total uptake from the low-moisture zone, $U_{lm, tot}$, due to the correlation existing in root uptake at different depths $z$ and to the stochastic nature of both water stress and the boundary $z = h$ of the low-moisture domain. To circumvent these issues, one can compute the depth-dependent long-term average root uptake, $U_{lm, tot}(z)$, by substituting into equation (13) the long-term average soil moisture, $s'(z)$ (from the numerical inversion of equation (17)). This quantity can then be integrated over the stochastic domain $0 \leq z \leq h$, to give the average root uptake from the low-moisture zone, as

$$ U_{lm, tot}(h) = \int_0^h \frac{s'(z) - s_w}{T_p} r(z) dz. $$

This relation can be numerically computed and solved for $h$ to give the corresponding probability density function as a derived distribution of $p_{U}(h)$ (equation (25)). Examples of pdfs of the average root uptake from the low-moisture zone are shown in Figure 12 with dashed lines. The atom of probability at $U_{lm, tot} = 0$ corresponds to the probability of being in SWT conditions (i.e., without LM zone).

[54] As a last point, we can determine the total root uptake from the whole soil column, $U_{tot}$, as a sum of the two components, $U_{lm, tot}(h)$ and $U_{h}(h)$. $U_{tot}(h)$ can be determined numerically, and the corresponding probability density function is obtained as a derived distribution of $p_{U}(h)$. Examples are shown in Figure 12 (solid lines): the shift of the pdfs toward the maximum value ($T_p = 0.5$ cm/d) of root uptake indicates a decrease in plant water stress from case 1 to case 3. In addition, Figure 12c remarks that the pdf of the total root uptake is far from being the convolution of the pdfs of $U_{lm, tot}(h)$ and $U_{h}(h)$. This is due to the correlation existing between the two variables, regardless of the fact that assumptions of noncooperative root functioning and negligible upward flux at $z = h$ would support the indepen-

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**Figure 12.** Probability density functions of root uptake from groundwater ($U_{w}$, dotted line), average root uptake from the low-moisture zone ($U_{lm, tot}$, dashed line), and total root uptake ($U_{tot}$, solid line) in the three cases shown previously: (a) case 1 (Figure 6), (b) case 2 (Figure 7), and (c) case 3 (Figure 8).
dence between root uptake above and below the zero flux surface.

5. Concluding Remarks

[55] In this paper, a probabilistic framework has been proposed to model the stochastic dynamics of soil moisture in a generic soil layer in groundwater-dependent environments. Processes such as rainfall infiltration, root water uptake and capillary flux are taken into account in the local water balance, along with the stochastic groundwater dynamics investigated in the companion paper by Laio et al. [2009].

[56] The proposed framework enables the investigation of the stationary probability distributions of moisture content in every soil layer, under different climatic and environmental conditions. The key role of a fluctuating water table on the soil moisture dynamics is highlighted by the analysis of the probability density functions of soil moisture (Figures 6, 7, and 8) and long-term average soil moisture profiles (Figure 9). Vegetation is found to crucially interact with the groundwater dynamics, and the mean root depth strongly affects the water table depth and the moisture distribution along the soil profile (see Figure 11). In turn, hydrologic conditions determine the rate of root uptake and plant transpiration through water stress mechanisms (Figure 12), with a consequent impact on vegetation health and a number of ecosystem processes such as plant growth and interspecies competition. In particular, the probability density functions of root uptake presented in this paper enable the study of plant water stress, of the occurrence of soil saturation, and vegetation strategies of root allocation and flood resistance.

[57] In conclusion, the semianalytical model proposed here gives an insight into the role of a shallow water table and moisture level on ecosystem processes forced by stochastic precipitation. The method enables a quantitative analysis of the complex and intertwined dynamics of vegetation and hydrologic conditions in groundwater-dependent ecosystems.

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