Chapter 3

SOIL MOISTURE DYNAMICS IN WATER-LIMITED ECOSYSTEMS

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1. Introduction

The main control exerted by hydrological processes on vegetation in water-limited ecosystems is through the soil water content, which, in turn, results from complex interactions between precipitation, infiltration, evaporation, transpiration, and soil drainage. Most of these processes are state-dependent, in that their rates are functions of the soil water content.

A number of eco-hydrological processes in water-limited ecosystems depend on the soil water content. Soil moisture dynamics affect the occurrence, duration and intensity of periods of water stress in vegetation (Hale and Orchutt, 1987; Smith and Griffiths, 1993; Porporato et al., 2001), with important effects on plant cell turgidity, stomatal conductance, and, in turn, on photosynthesis, carbon assimilation and ecosystem net primary productivity (see Chapter 4). The control of soil moisture on canopy conductance is also of foremost importance in modulating the heat and water vapor fluxes from terrestrial vegetation to the near-surface atmosphere, with important consequences for the moisture content and stability of the atmospheric boundary layer and consequent feedbacks to precipitation and the water cycle (Chapters 5 and 6). Soil moisture exerts an important control also on nutrient cycling (e.g., Linn and Doran, 1984, Skopp et al., 1990; Parton et al., 1998; Porporato et al., 2003), due to its effects on microbial activity, nitrogen leaching and nutrient uptake, as explained in Chapters 11 and 14. Other surface processes affected by the soil water content, include infiltration, runoff, soil erosion and dust emission from dryland landscapes (see Chapter 9).

The study of all these processes and of the impact of possible scenarios of climate variability and change, requires the analysis and modelling of soil moisture dynamics. To this end, a number of models have been developed by different authors to simulate the spatial and temporal patterns of soil moisture in the root zone. This chapter will focus on the temporal dynamics, while Chapter 7 discusses the topography-driven spatial variability of soil moisture.

2. Types of soil moisture models

Soil moisture dynamics are usually studied through a soil water budget accounting for the main inputs and outputs of soil water, namely rainfall infiltration, evapotranspiration, and drainage. The complexity of these models depends on the different levels of detail used in the representation of these fluxes, as well as in the number of soil layers used in the calculation of the soil water storage (Table 1).

A number of models have been developed by crop scientists, agro-meteorologists, and hydrometeorologists to calculate soil water content and evapotranspiration by coupling the soil water budget with the surface energy balance. Some of these models require the simultaneous solution of the water and energy balance equations to determine at the same time evapotranspiration rates and soil water contents. This class of agro-hydrological (e.g., Jones and Kiniry, 1986; Evett and Lascano, 1993; Daamen and Simmonds, 1994) and land surface models (e.g., Sellers et al., 1986; Dickinson et al., 1986) generally requires the measurement of several

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micrometeorological and hydrological variables. In most applications, only limited information is available on vegetation canopy (e.g., the leaf area index, LAI) and a "two layer-two source" approach (e.g., Daamen, 1997; Xue et al, 1996) is often used to model evaporation from soil and leaf surfaces. A major problem in this modeling approach is in properly accounting for both the partitioning of the incident energy between vegetation and soil, and the to possible interactions between soil evaporation and transpiration. In other cases further complexity is added by including a multilayer representation of vegetation canopies (e.g., Norman, 1979; Chen, 1984; Baldocchi and Meyers, 1998).

Table 1: A classification of soil moisture models on the basis of different approaches used in the representation of soil, evapotranspiration, vegetation canopies, infiltration, and precipitation.

Soil	**Multilayer** (e.g., Evett and Lascano, 1993; Sellers et al., 1986; Dickinson et al., 1986; Xue et al., 1996; Daamen and Simmonds, 1994; Parton et al., 1998; Hopmans, 2002; Feddes et al., 2001; Guswa et al., 2002)
	Single Layer (e.g., Budyko, 1958; Rodriguez-Iturbe et al., 1991; 1999; Milly, 1993; Laio et al., 2001a)
Evapotranspiration	**Potential evapotranspiration approach** (e.g., Lai and Katul, 2000; Sellers et al., 1986; Dickinson et al., 1986; Milly, 1993; Rodriguez-Iturbe et al., 1999a; Laio et al., 2001a; Guswa et al., 2002)
	Simultaneous solution of energy and water balance equations (e.g., Evett and Lascano, 1993; Daamen and Simmonds, 1994; Daamen, 1997)
Canopy/Evaporating Surfaces	**1-layer ("big leaf")** with no account for soil evaporation (e.g., Milly, 1993; Rodriguez-Iturbe et al., 1999a; Lai and Katul, 2000; Guswa et al., 2002).
	2-layers: soil and a 1-layer canopy (e.g., Xue et al., 1991; 1996; Daamen and Simmonds, 1994; Evett and Lascano, 1993; Laio et al., 2001a).
	Multi-layer/multi-crop (e.g., Norman, 1979; Chen, 1984; Sellers et al., 1986; Dickinson et al., 1986; Baldocchi and Meyers, 1998).
Infiltration	**Integration of Richards equations** (Sellers et al., 1986; Dickinson et al., 1986; Xue et al., 1996; Evett and Lascano, 1993; Daamen and Simmonds, 1994; Hopmans, 2002; Feddes et al., 2001; Lai and Katul, 2000; Guswa et al., 2002)
	Conceptual model for single layer (Milly, 1993; Rodriguez-Iturbe et al., 1999a; Laio et al, 2001a) or multi-layer soil (e.g., Parton et al., 1998)
Rainfall	**Deterministic/data records** (e.g., Evett and Lascano, 1993; Daamen and Simmonds, 1994)
	Stochastic (e.g., Eagleson et al., 1978a; Milly, 1993; Rodriguez-Iturbe et al., 1999a; Laio et al., 2001a)

Other models (e.g., Lai and Katul, 2000), instead of solving simultaneously the energy and water balance equations, use the concept of potential evapotranspiration, E_p, and express the total (actual) evaporative fluxes, E_{act}, as the by-product between E_p and a function of soil moisture, f(τ), accounting for the reduction in evapotranspiration due to stomata resistance in conditions of limited soil water availability (e.g., Chapter 4),

$$E_{act}=f(\tau)\ E_p, \tag{1}$$

In equation (1) τ is the volumetric soil moisture, i.e. the ratio between water and void volumes in a soil sample. By definition, the rate of potential evapotranspiration does not depend on τ and can be calculated either with an energy balance or with a combination method, such as the Penman-Brutsaert model (e.g., Katul and Parlange, 1992). Thus, the solution of the energy balance (or of the combination) equation occurs independently of the integration of the soil moisture balance, allowing for significant simplifications in the numerical algorithms. The function, f(τ), represents the effect of soil moisture on canopy (and soil) resistance and is usually expressed either through empirical parameterizations (e.g., Jarvis, 1976; Jones, 1992), or through process-based models of the plant physiological response to water stress (e.g., Gao et al., 2002; Daly et al., 2003). More details on the modeling of f(τ) are provided in Section 3.1.

The soil water balance is usually expressed (e.g., Feddes et al., 2001; Lai and Katul, 2000; Guswa et al., 2002) by a one-dimensional (in the z-direction, with *z* being positive downward) mass-conservation equation for the root zone

$$\frac{\partial \theta(z,t)}{\partial t} = -\frac{\partial q(z,t)}{\partial z} - u(z,t) \tag{2}$$

where *q* is the unsaturated Darcian flux (Richards, 1931) associated with rainfall infiltration and *u* is the water uptaken by plant roots. *u(z)* is related to the actual rate of transpiration using, for instance, root distribution (or density) functions (e.g., Feddes et al., 2001; Hopmans et al., 2002), as explained in Chapter 2. Regardless of the particular method used in the calculations of the actual rates of evapotranspiration, root water uptake depends on soil moisture. Similarly, the infiltration rate is state-dependent, in that it can be expressed as

$$q = -K(\theta)\left[1 + \frac{\partial \psi(\theta)}{\partial z}\right] \tag{3}$$

with *K* (hydraulic conductivity) and | (soil water potential) being both non-linear functions of soil moisture. Appropriate boundary conditions need to be specified at the soil surface to account for soil evaporation and rainfall infiltration, and at the bottom to account for deep infiltration (e.g., Evett and Lascano, 1993). However, because this approach can be at times somewhat cumbersome, simplified models are often used in ecohydrology to calculate the soil water content. More details on the solution of the complete equations (2) and (3) are provided in Chapter (2). A number of analytical solutions of the non-linear infiltration equations are also available; a review can be found in Smith et al. (2002). Here we focus on simplified approaches to soil moisture modeling, involving some approximations in the representation of infiltration, runoff and evapotranspiration.

3. Processes affecting soil moisture dynamics

3.1. EVAPOTRANSPIRATION

The process of evapotranspiration includes soil evaporation, transpiration from the plant stomata, and evaporation of precipitation fractions intercepted by canopy and litter. Chapter 5 describes the dependence of evapotranspiration on solar irradiance, vapor pressure deficit, and wind speed, and explains how the atmospheric evaporative demand varies with these parameters. In arid and semiarid environments evapotranspiration is often limited by soil water availability more than by evaporative demand. Thus, any soil water balance model should include an adequate representation of the dependence of evapotranspiration on soil moisture, for instance, through the function f(τ) (see equation (1)). This function would parameterize the effect of stomata regulation described in Chapter 4.

Thornthwaite and Mather (1955) and Budyko (1958) expressed f(τ) using a piecewise linear function (Figure 1A). According to this model evapotranspiration occurs at a maximum (potential) rate for soil moisture values above a critical level, τ^*, and decreases with τ for $\tau < \tau^*$. This approach is often used in general circulation (e.g., Eagleson, 1982) and ecohydrological (Rodriguez-Iturbe et al., 1999a; Rodriguez-Iturbe and Porporato, 2005) models. In the past, some authors expressed τ^* and a fraction of the moisture content at field capacity (e.g., Eagleson, 1982) while it is now understood (Rodriguez-Iturbe et al., 1999b; Laio et al., 2001a) that τ^* depends not only on soil hydraulic properties but also on vegetation characteristics. It can also be observed that the Thornthwaite-Budyko model is unrealistic for low moisture contents, as it assumes that evapotranspiration becomes zero when the soil is completely dry. Stomatal conductance and transpiration are commonly found to become zero already at the so-called permanent wilting point, τ_w. In the crop science literature τ_w is usually taken as the soil water content corresponding to –1.5 MPa of soil matric potential (i.e., $|_w$=-1.5 MPa). However, dryland vegetation is often found to wilt at water potentials much smaller than –1.5 MPa (e.g., Richter, 1976, pp. 47-49). Thus, in general, τ_w and τ^* depend both on soil texture and vegetation types, as different plants have different strategies to respond to water stress (Chapter 4).

Different models have been suggested to express the function f(τ) in equation (1) (Jarvis et al, 1976; Eagleson et al., 1982; Rodriguez-Iturbe et al., 1999a; Laio et al., 2001a; Daly et al., 2003). Laio et al. (2001a) used a piecewise continuous function (Figure 1B) accounting for soil moisture controls both on transpiration and soil evaporation

$$f(\theta) = \begin{cases} 1 & (\theta^* < \theta < n) \\ \dfrac{\theta - \theta_w}{\theta^* - \theta_w}\left(1 - \dfrac{E_{evap}}{E_p}\right) + \dfrac{E_{evap}}{E_p} & (\theta_w < \theta < \theta^*) \\ \dfrac{E_{evap}}{E_p}\dfrac{\theta - \theta_h}{\theta_w - \theta_h} & (\theta_h < \theta < \theta_w) \\ 0 & (0 < \theta < \theta_h) \end{cases} , \tag{4}$$

with n being the soil porosity and E_{evap} soil evaporation. This model assumes that no soil evaporation occurs below the hygroscopic point, τ_h, and that for $\tau_h < \tau < \tau_w$ soil moisture losses are due only to soil evaporation. Equation (4) will be applied in Sections 4 and 5 to as the evapotranspiration components of a stochastic model of soil moisture.

This section has discussed only possible parameterizations of the effect of soil moisture on the rate of evapotranspiration through stomata regulation. However, the physiological mechanisms controlling stomatal conductance and plant water relations in arid and semiarid ecosystems are more complex (see Chapter 4) and involve a number of other environmental and physiological variables. Plants may close their stomata not only to reduce the transpiration rates in water stress conditions, but also to control the assimilation rates in response to changes is light, enzyme, and CO_2 concentrations (Farquar et al., 1980). Equations (1) and (4) fail to account for the biochemical control on stomatal conductance. On the other hand, commonly used photosynthesis models (e.g., Ball et al., 1987; Collatz et al., 1991) do not account for soil moisture limitations. A unifying, process-based modelling framework can be found in Daly et al. (2003) and Katul et al. (2003), while more empirical approaches - frequently implemented in land surface models (e.g., Xue et al., 1996) - use Jarvis' (1976) formulation to account both for plant hydraulics and biochemical controls on stomatal conductance.

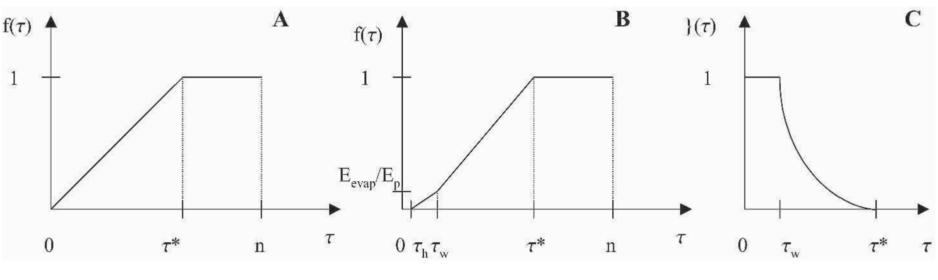

Figure 1. Dependence of evapotranspiration on soil moisture according (A) the Thorntwaite-Budyko model, and (B) the model by Laio et al. (2001a). The (static) stress function (equation (5), Porporato et al., 2001).

3.1.1. Water Stress in Vegetation

The soil moisture levels τ^* and τ_w are indicative of conditions of incipient and severe vegetation water stress, respectively. Thus, models of soil moisture dynamics have been used (e.g., Rodriguez-Iturbe et al., 1999b,c) to determine the impact of different climate, soil and vegetation characteristics on plant water stress. A water stress function, }, was suggested by Rodriguez-Iturbe et al. (1999b) to quantify this impact (Figure 1C). Plants are assumed to be unstressed when soil moisture exceeds τ^*, while the stress is maximum at the wilting point

$$\zeta(\theta) = \left(\frac{\theta^* - \theta}{\theta^* - \theta_w} \right)^q \qquad (\theta < \theta^*) \quad , \tag{5}$$

and}(τ)=0 when $\tau > \tau^*$. A similar formulation was suggested by Sellers et al. (1986) in terms of leaf water potential. The exponent q in equation (5) accounts for the non linear response of vegetation to water deficit. Chapter 4 provides more details on the non-linear physiological processes controlling vegetation response to water deficit. Equation (5), expressing the stress level associated with a certain moisture content, τ, has been defined as "static stress" (Porporato et al., 2001) because it does not provide any information on the overall effect of the temporal dynamics of soil moisture on vegetation. Section 4.1.2 will discuss a stochastic approach to account for the frequency, duration, and intensity of water stress, based on a definition of a suitable "dynamic stress".

3.2. INFILTRATION AND RUNOFF

Infiltration is the water flux from the surface into the ground. It depends on rainfall rates and soil hydraulic properties, and can be calculated as $I=-q(z=0)$, with q given by equation (3) for $z=0$. As water infiltrates into the ground, surface soil moisture increases, leading to changes both in hydraulic conductivity and in water potential gradients. If the rainfall rate is greater than the saturated hydraulic conductivity, after some time the surface soil layer reaches saturation, water potential gradients at the surface tend to zero, and the maximum infiltration rate – known as infiltration capacity, I_c – tends to the saturated hydraulic conductivity, K_s (see equation (3)). Thus, the soil infiltration capacity decreases with time and tends asymptotically to K_s (Figure 2, white square boxes). I_c can be calculated with a number of analytical models as a function of the soil hydraulic properties (Green and Ampt 1911; Philip, 1957; 1969; Smith and Parlange 1978; See also Eagleson, 2002, p.176, and Smith et al., 2002, for a review).

Thus, if the rainfall rate exceeds the infiltration capacity, once the surface layer reaches saturation, infiltration is equal to I_c, and the rainfall excess generates runoff. This mechanism of runoff production is known as "Hortonian runoff" or "infiltration excess runoff" (Horton, 1933). Conversely, if the soil surface is fairly permeable and the rainfall is less than K_s, the precipitation rate is smaller than the infiltration capacity and no infiltration-excess runoff is generated. However, when a relatively shallow soil is underlain by an impervious surface (e.g., a bedrock, a clay layer, or the water table), only a limited amount of water can be stored in the soil and runoff may be generated by "saturation excess" (Dunne and Black, 1970; Dunne, 1978). While Hortonian runoff is observed in watersheds with low soil infiltration capacity, saturation-excess is typical of areas with limited soil storage capacity.

In a review of a number of case studies around the world, Dunne (1978) concluded that Hortonian runoff is typical of some arid and semiarid regions with thin vegetation cover, and of catchments disturbed by overgrazing or agriculture. Conversely, in regions with relatively dense vegetation saturation-excess is the dominant mechanism of runoff generation. Subsequent studies have confirmed that vegetation density plays an important role in determining the dominant mechanism of surface runoff production. Casenave and Valentin (1992) studied the ability of soils to generate infiltration-excess surface runoff and classified a number of tropical soils on the basis of vegetation cover, surface crusting, and other soil hydraulic properties. As expected, runoff was found to increase with decreasing vegetation densities and in the presence of soil crusting. Field studies have also shown that, due to the strong spatial heterogeneity of soil surface characteristics, the spatial patterns of overland flow are in general not uniform. Moreover, in areas with poor drainage networks, surface runoff collects into pools or "water holes", contributing to the spatial heterogeneity of surface moisture.

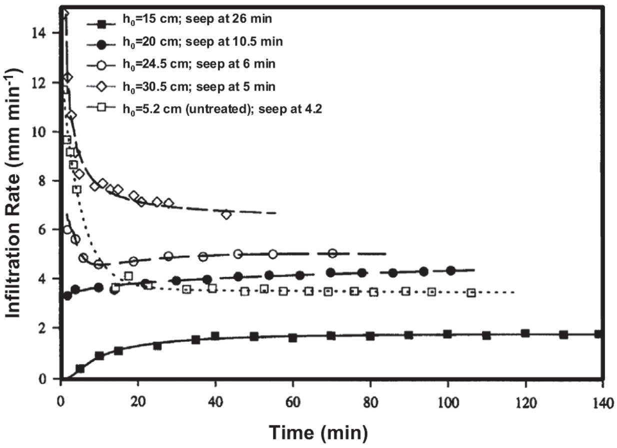

Figure 2. Infiltration rate as a function of time for untreated soils (white square boxes) and for soils treated with hydrophobic substances (see Section 3.3 for more details on hydrophobicity). The water-entry pressure head of the hydrophobic soil was h_p=| $_p$/γ_w=8.4 *cm and different depths,* h_0, *of ponding water were used. Taken from Letey (2001), published with permission from John Wiley & Sons (©John Wiley & Sons).*

3.3. EFFECTS OF DISTURBANCES ON INFILTRATION AND RUNOFF

3.3.1. The effect of logging and wood harvesting

A number of studies on watersheds around the world have shown how the removal of vegetation increases the water yield. Bosch and Hewlett (1982) reviewed 94 catchment studies on the effect of deforestation and forest harvesting on surface runoff: the removal of forest vegetation was found to consistently increase runoff, due to the loss of soil infiltration capacity and to the lower fraction of annual precipitation evapotranspired by vegetation (soil evaporation is usually smaller than plant transpiration). The effects of forest harvesting on runoff tend to last for several years in semiarid environments, due to the slow rates of forest regeneration. Both the plantation of commercial timber vegetation (afforestation) and the natural regrowth of forests after clearcut (reforestation) have effects that are opposite to deforestation: soil infiltration capacity increases due to the growth of roots and forest floor (see Table 2). Moreover evapotranspiration increases, due to canopy and litter interception, and to the ability of trees to tap deeper water.

The effect of savanna, grassland, and shrubland vegetation on infiltration and runoff has also been studied by several authors (e.g., Mainguet, 1999) who showed how vegetation removal resulting from overgrazing and poor land management decreases soil infiltration capacity and increases soil erosion.

Table 2: Infiltration capacity with different vegetation covers (data from Lull, 1964).

Surface cover	**I_c (mm h^{-1})**
Undisturbed forest floor	60
Forest floor without litter and humus	49
Yearly burned forest floor	40
Pasture	29

3.3.2. The impact of fires

The effect of fires on infiltration (Table 2), runoff, and water yield has important hydrologic implications (Krammer and DeBano, 1965). The post-fire increase in runoff and soil erosion was initially attributed to loss of infiltration capacity due to rainsplash and soil compaction. In addition, fires were believed to decrease surface soil permeability by clogging the soil pores with ashy particles (DeBano, 2000). Krammer and DeBano (1965) and DeBano (1966) showed that the decrease in infiltration capacity subsequent to fire is in large part associated with water repellency developed by the fire at the soil surface or at shallow depths. Organic compounds of chaparral and other vegetation types are volatilized by the fire and transported downwards into the soil by the strong temperature gradients existing through the soil profile. These gases condensate at a certain depth (of only a few centimeters), developing a hydrophopic coating around the soil particles (e.g., DeBano, 2000). This effect depends on the fire regime (Chapter 16), in particular on fire temperature, as repellency is observed (e.g., DeBano, 2000; Doerr et al., 2000) to develop neither with relatively low (e.g., $T<175\ ^{o}$C) nor with high temperatures ($T>300\ ^{o}$C). The organic compounds released by the fire affect the physical-chemical properties of the grain surfaces: in particular, the contact angle, ϖ, formed by the air-water interface with the soil grains becomes greater than 90°, causing a positive capillary pressure head, $|_{p}$, in correspondence to the hydrophobic layer[1] (e.g., Letey, 2001). Thus a water drop reaching a water repellent surface is not drawn into that surface but sets on it. However, after some time water infiltrates through the hydrophobic layer, presumably due to a decay of the organic coating of the soil grains. This explains (see equation (3)) why in the presence of soil hydrophobicity infiltration capacity increases through time (Figure 2), while in wettable soils (section 2.1) both theory and observations suggest a decrease in infiltration capacity (Letey, 2001).

Thus, fire occurrences have important ecohydrological implications because the increase in runoff an of the associated erosion of the soil surface redistribute water and nutrients (e.g., Schlesinger et al., 1990) while the heterogeneity of burnt areas partly contributes to the emergence of patchy patterns of vegetation. A more thorough analysis of vegetation patterns and fire regimes can be found in Chapters 15 and 16.

3.4. VERTICAL REDISTRIBUTION OF SOIL MOISTURE BY TREE ROOTS

The physiological processes controlling water flow through the soil-plant-atmosphere continuum described in Chapter 4 suggest that transpiration is the main process of water loss by vegetation. However, some evidence exists that at night, when the stomata are closed, the root system may offer a preferential pathway for water flow through the soil column and contribute to the transfer of moisture between different soil layers (Burgess et al., 1998). For example, in rainless periods - when most of the root-zone soil moisture is concentrated in deep soil layers - roots may transfer water from the deep soil and release it in the dry shallow layers. Known as "hydraulic lift" (Richards and Caldwell, 1987), this effect is driven by water potential gradients between different parts of the soil profile. The occurrence of hydraulic lift has been documented for a number of different species and ecosystems (Caldwell et al., 1998, Richards and Caldwell, 1987; Burgess et. al., 1998; Horton and Hart, 1998; Ludwig et. al., 2002). The ecohydrological significance of this phenomenon in arid and semiarid ecosystems is due to its positive impact on neighboring shallow rooted species, as well as to the possible enhancement of plant nutrient uptake from nutrient-rich shallow soil layers (Caldwell and Richards, 1989; Dawson, 1993).

The opposite process has been also observed: soon after rain, when the soil surface is moister than the deep soil, night-time downward fluxes (probably due to soil water potential gradients) can occur through the root system of some species (Burgess et al., 1998; Schulze et al.,

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[1] In fact, using the pore-capillary tube analogy, the water-entry pressure in the capillary tube is $|_{p}=-2\varphi\cos\varpi/(r\upsilon g)$, with φ being the surface tension of water, r the capillary tube radius, υ the water density, and g the gravitational acceleration.

1998; Smith et al, 1999). By means of this "reverse flow" (or "downward siphoning") trees and shrubs can subtract water and nutrients from the reach of shallow-rooted species.

4. Simplified models of soil moisture dynamics

One of the simplest models of soil water balance considers only one layer of soil (e.g., Budyko, 1958) and provides estimates of depth-average soil moisture through the equation (e.g., Rodriguez-Iturbe et al., 1999a)

$$Z_r \frac{d\theta}{dt} = I(\theta, t) - E(\theta) - L(\theta) \cdot \tag{6}$$

with Z_r being the depth of the root zone, I the rainfall infiltration, E the rate of evapotranspiration, and L the rate of drainage at the bottom of the soil layer. Rainfall infiltration is generally assumed to occur instantaneously and to be limited only by soil storage capacity. Thus, rainfall infiltration is equal either to the storm depth or to the soil storage capacity, $(n-\tau)Z_r$, whichever is less. Losses of water from the control volume are due to evapotranspiration and deep drainage. Evapotranspiration is generally calculated with equation (1), with a suitable function, $f(\tau)$, expressing the effect of soil moisture limitation (e.g., equation (4)). Drainage losses, $L(\tau)$, are assumed to be zero when soil moisture is below the so-called field capacity, τ_{fc}, representing the maximum moisture content at which water can be retained within the ground against gravity drainage. When $\tau > \tau_{fc}$ $L(\tau)$ can be expressed as equal to the (unsaturated) hydraulic conductivity, which in turn is a function of soil moisture (Chapter 2). For instance the following function has been used by Laio et al. (2001a)

$$L(\theta) = \frac{K_s}{e^{\frac{\beta}{n}(n-\theta_{fc})} - 1} \left[e^{\frac{\beta}{n}(\theta-\theta_{fc})} - 1 \right] \qquad (\theta_{fc} < \theta \leq n) \tag{7}$$

with K_s being the saturated hydraulic conductivity and ϵ a parameter depending on soil texture.

This model, sometimes referred to as "bucket model", has been widely used to determine surface boundary conditions in atmospheric models (Budyko, 1958), to study the feedbacks between the soil surface and the near-surface atmosphere (e.g., Rodriguez-Iturbe et al., 1991; 1998; Porporato et al., 2000; D'Odorico and Porporato, 2004), as well as in ecohydrological models of plant available water at the daily time scale (Rodriguez-Iturbe et al., 1999a,b,c).

The main limitation of this model is due to the use of only one soil layer. This approach does not allow for a calculation of the time needed by water to infiltrate to relatively deep soil layers and could lead to unrealistic soil moisture estimates at subdaily time scales. However, the comparison with the results of a more complex model based on the integration of Richards equation (1931) has shown (Guswa et al. (2002)) that that the two models provide similar results at the daily time scale. This is particularly true when roots are assumed to be able to extract more water from wet portions of the root zone to compensate for the lower uptake from the drier parts of the soil column. In some cases the single-layer ("bucket") model has been modified by adding multiple layers. A simplified multilayer model of the water balance uses equation (1) for each layer, considering as main moisture input to the lower layers the drainage from the overlying soil (Parton et al., 1998). Studies at relatively short (subdaily) time scales or in deep soils still require the numerical integration of Richards' (1931) equation through the soil profile (see Chapter 2).

4.1. A STOCHASTIC APPROACH

4.1.1. Probability distribution of soil moisture

In many ecohydrological applications the intermittent and unpredictable stochastic nature of rainfall occurrences and amounts suggests using a probabilistic approach in the modeling and characterization of precipitation and soil moisture regimes (Eagleson, 1978b). While several stochastic models of precipitation have been developed in the last few decades (e.g., Todorovic and Woolhiser, 1975; Eagleson, 1978a), the stochastic modeling of soil moisture is much more recent and, to date, only a few probabilistic models of soil water dynamics have been studied in detail, providing analytical solutions of the stochastic soil water balance equation (Rodriguez-Iturbe et al., 1999a; Laio et al. 2001a). These models use the simplified framework described in Section 3, where the soil water balance is expressed by equation (6). Losses due to evapotranspiration and drainage are modeled as deterministic functions of soil moisture (i.e., equations (4) and (7), respectively), while precipitation is expressed as a sequence of Poissonian occurrences of rate (e.g., Eagleson et al., 1978a), with each rainfall event having a random depth with exponential distribution of mean δ. Thus the rainfall regime is characterized by the two parameters, o and δ, representing the average storm frequency and depth. Eagleson (2002) calculated these parameters for a number of locations across the continental U.S. The decomposition of the rainfall regime into a sequence of storm events separated by intersorm periods is of foremost importance to the modeling of soil moisture because the timescales of variability of the soil water content are in general of only a few days, depending on depth, soil texture, and climate. Moreover the effect of climate change on the rainfall regime can be manifested (e.g., Knapp et al., 2002) either in changes in the number of storms occurring during the rainy season (i.e., of o) or in the their size (i.e., of δ). The effect of these changes on the soil water balance in dryland ecosystems is discussed in Section 4.1.3.

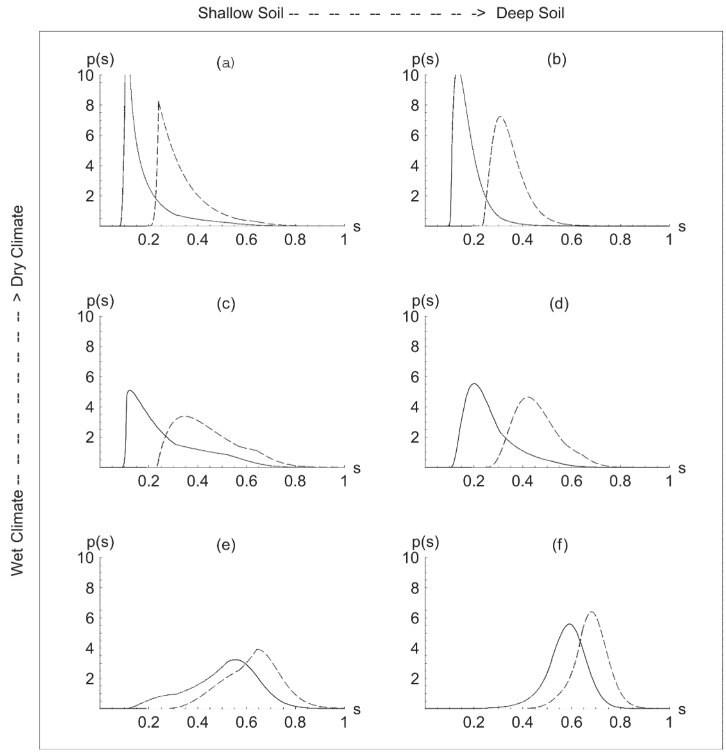

Figure 3. Examples of pdf's of soil moisture for different type of soil, soil depth, and mean rainfall rate. Continuous lines refer to loamy sand, dashed lines to loam. Left panels correspond to rooting depth of 30 cm, right panel to 90 cm. Top, center, and bottom graphs have a mean rainfall rate λ of 0.1, 0.2, and 0.5 d^{-1} respectively. Common parameters to all graphs are α=1.5 cm, and E_{max}=0.45 cm/d. After Laio et al. (2001).

The stochastic soil water balance studied by Rodriguez-Iturbe et al. (1999a) and Laio et al. (2001a) assumes that evapotranspiration changes are due only to soil moisture fluctuations, while E_p and E_{evap} are constant. This assumption is plausible for a fairly steady summer season, as long as equation (6) is considered at daily (or longer) time scales and diurnal fluctuations in potential evapotranspiration are ingored.

When rainfall infiltration is modeled as a stochastic process, equation (6) becomes a stochastic differential equation and its solution provides the probability density function of soil moisture as a function of rainfall, vegetation and soil parameters. Analytical solutions of the stochastic soil water balance equations were determined by Milly (1993; 2001), Rodriguez-Iturbe et al. (1999a), and Laio et al. (2001a). Figure 3 shows an example of probability distributions of relative soil moisture, s ($s=\tau/n$), for different soil and precipitation parameters. As the rainfall parameters, o (average storm frequency) and δ (average storm depth), increase, the probability distribution of soil moisture shifts towards wetter conditions. Moreover, coarser soils are consistently found to be drier than fine-texture soils, while the soil depth is found to affect the variance of soil moisture fluctuations, with the shallow soils having broader probability distributions of soil moisture. Information on the probability distribution of soil moisture is of foremost importance in ecohydrology, in that it allows, for instance, the estimation of the probability that vegetation is under stress during the growing season, as discussed in the following sections.

4.1.2. Duration and frequency of water stress in vegetation

The probabilistic framework presented in Section 4.1 has been used to calculate the level-crossing statistics of soil moisture. In particular, analytical expressions have been obtained to calculate (Ridolfi et al., 2000a; Porporato et al., 2001) the average duration and frequency of temporal intervals in which soil moisture is below a given critical level. In particular, these statistics were studied with respect to soil moisture levels of incipient (τ^*) and severe water stress (τ_w) and interesting non-linearities were found in the dependence on soil climate, and vegetation parameters.

These results were used by Porporato et al. (2001) to characterize the stress conditions associated with different regimes of soil moisture: a "dynamical water stress" function was defined as an indicator of the overall stress of vegetation throughout the growing season

$$\overline{\chi} = \begin{cases} \left(\dfrac{\overline{\zeta}'\overline{T}_{\theta*}}{kT_{seas}} \right)^{1/\sqrt{\overline{n}_{\theta*}}} & if \quad \overline{\zeta}'\overline{T}_{\theta*} < kT_{seas} \\ 1 \qquad otherwise & \end{cases} \tag{8}$$

where T_{seas} is the growing season length, $\overline{T}_{\theta*}$, is the average length of an excursion below τ^*, $\text{n}_{\tau*}$ is the number of intervals with $\tau<\tau^*$, and $\overline{\zeta'}$ is the average value of the static stress, }, (equation (5)) conditioned to } being greater than zero. The theoretical considerations justifying this formulation of the stress function can be found in Porporato et al. (2001) along with the analytical expressions of the dynamical stress as a function of plant, soil, and rainfall parameters. An example of the application of plant dynamic stress is presented in the following section.

4.1.3. The inverse texture effect.

The framework presented in the previous sections allows for the investigation of the effect of soil properties and rainfall regime on vegetation water stress. In particular, Laio et al. (2001b), used the concept of dynamic water stress to study the suitability of different soil textures to the growth of vegetation under different rainfall regimes. This approach allowed these authors to explain the "inverse soil texture effect" introduced by Noy-Meir (1973, p. 37), that "the same vegetation can occur at lower rainfall on coarse soils than it does on fine ones. The balance point between the advantage of coarser texture and its disadvantage occurs somewhere between 300 and 500 mm rainfall". Thus different combinations of climate and soil properties can lead to similar soil moisture regimes and explain the existence of the same vegetation with lower seasonal precipitation on coarser soils.

As an example of the inverse texture effect Laio et al (2001b) applied their stochastic soil moisture and water stress models to the shortgrass steppe in north-central Colorado. The recruitment patterns of *Bouteloua gracilis*, the dominant species in this steppe, is significantly affected by soil texture (Lauenroth et al., 1994). The dynamic water stress, $\overline{\chi}$, was used to express the combined effect on vegetation of static stress, and duration and frequency of water stress periods. Figure 4 shows the dynamic stress calculated for different combinations of soil textures in the USDA soil texture triangle. In the case of figure 4a the climate is relatively dry and the lowest stress is associated with coarse textures, suggesting a better fitness of *B. gracilis* on sandy soils. Conversely, in the case of a relatively wet climate, the same shortgrass would perform better in a fine soil. Laio et al. (2001b) showed also that in this area the point at which coarse soils become better than fine soils corresponds to an annual rainfall of about 370 mm, in agreement with the range of values indicated by Noy-Meir (1973). A similar study was proposed by Fernandez-Illescas et al. (2001) for different combinations of rainfall and vegetation conditions in savannas sites in southern Texas.

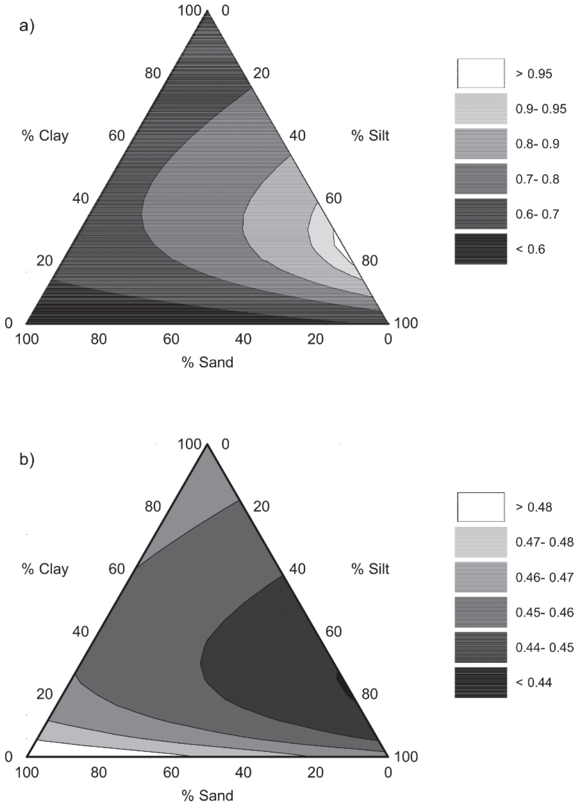

Figure 4. Dynamic water stress on the soil texture triangle for Bouteloua gracilis (a) under a relatively dry climate, α=0.576 cm and λ=0.17 d^{-1}; and (b) under a relatively wet climate, α=0.674 cm and λ=0.28 d^{-1}. After Laio et al. (2001).

4.1.4. The effect of interannual climate fluctuations

Climate variability affects terrestrial vegetation both through changes in temperature and rainfall regimes. In the case of arid and semiarid ecosystems in warm regions, changes in rainfall patterns are expected to have the most important effects on vegetation. These effects can be quantified by studying how soil moisture dynamics would change under different rainfall scenarios. This type of analysis was carried out by Rodriguez-Iturbe et al. (1999a) and Laio et al. (2001a) in a sensitivity analysis of their models with respect to the rainfall parameters (Figure 3).

However, the assessment of the effect of interannual climate fluctuations on vegetation is a different problem. In addition to the effect of daily hydrologic fluctuations, arid and semiarid regions are also characterized by strong year-to-year rainfall variability (Chapter 1), due anomalies in the large-scale atmospheric circulation. For example, in these regions differences in only a few rainstorm occurrences in course of the rainy season may significantly affect the total annual rainfall. D'Odorico et al. (2000) and Ridolfi et al. (2000b) studied the effect of climate fluctuations on average soil moisture, $\iota\,\theta\,|$ (or $\iota\,s\,|$), as well as on the duration, T_{τ^*}, and number, n_{τ^*}, of water stress periods during the growing season. Interannual fluctuations of the rainfall regime were modeled as fluctuations in the rainfall parameters, o (average storm frequency), and δ (average storm depth), in the stochastic soil moisture model of Section 4.1.1. Thus, o and δ were treated as random variables and their distributions were determined through the analysis of several decades of daily precipitation records (e.g., Figure 5). Each growing season was characterized by different rainfall parameters sampled from these distributions and the soil moisture statistics ($\iota\,\theta\,|$, T_{τ^*}, and N_{τ^*}) were calculated. The probability distributions of average seasonal soil moisture, stress duration and frequency were then calculated as derived probability density functions of the distributions of o and δ.

With this framework it was possible to address the important question of whether the nonlinearities embedded in the soil moisture dynamics would enhance or damp fluctuations in the rainfall regime. By comparing the coefficient of variation (standard deviation/mean) of T_{τ^*} with those of o and δ, Ridolfi et al. (2000b) were able to show that in most cases these nonlinearities enhance the interannual fluctuation of water stress, and hence the effect of disturbance exerted by climate fluctuations on arid and semiarid ecosystems.

The probability distributions of average seasonal soil moisture were calculated by D'Odorico et al. (2000) for different soil types and rainfall fluctuation parameters. Figure 6 shows an example of probability distribution or average seasonal soil moisture (in terms of relative soil moisture, $s = \theta/n$): a well defined bimodality exists with relatively strong rainfall fluctuations, while it disappears when the fluctuations become weaker. Thus, for some values of the coefficient of variation of o and δ, these distributions may exhibit bimodal behavior, suggesting that the system tends to select two preferential states and to switch between them due to the interannual fluctuations in rainfall parameters. The two preferential states correspond to "dry" and "wet" average seasonal soil moisture conditions. The implications of the emergence of bimodal behavior is of foremost importance for ecosystem dynamics because it implies that the system is more likely to be found in two states that are far form the long-term average, while the long-term average conditions occur with the lowest probability. Moreover, the bimodal behavior enhances the likelihood of occurrence of dry conditions and the effect of disturbance of climate fluctuations on terrestrial ecosystems.

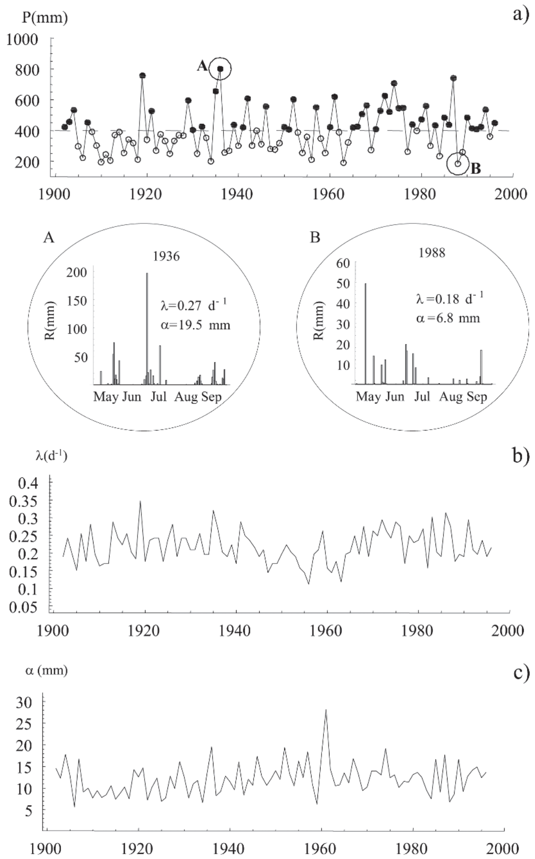

Figure 5. Analysis of the rainfall regime during the growing season at Luling (Texas) based on daily precipitation data: (a) time series of total seasonal rain (May 1^{st}-Sept. 30^{th}. (b) Time series of the estimated rate of storm arrivals, λ. (c) Time series of the average storm depth, α. After D'Odorico et al. (2000a).

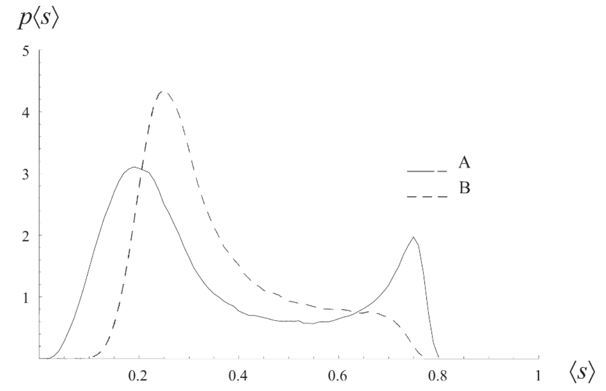

Figure 6. Probability density function of the average soil moisture during the growing season. The parameters for soil and vegetation are as follows: n=0.43; Z_r=1.40 m; K_s=9.5 10^{-6} m/s; s_1=0.8; s^*=0.36; E_{max}=3.2 mm/day. The rainfall is characterized by $\langle\alpha\rangle$=12.4 mm/storm and $\langle\lambda\rangle$=0.21 d^{-1}, with coefficients of variation: (A) $CV[\alpha]$=0.45; $CV[\lambda]$=0.23. (B) $CV[\alpha]$=0.22; $CV[\lambda]$=0.11. After D'Odorico et al. (2000).

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6. References

Baldocchi, D., and T. Meyers, (1998). On using eco-physiological, micrometeorological and biogeochemical theory to evaluate carbon dioxide, water vapor and trace gas fluxes over vegetation: a perspective. Agric. For. Meteorol., 90(1-2), 1-25.

Ball, J.T., Woodrow, I.E., Berry, J.A. (1987). A model predicting stomatal conductance and its contribution to the control of photosynthesis under different environmental conditions. In J. Biggins (Ed.), Progress in Photosynthesis Research, Martinus-Nijhoff Publishers, Dordrecht, The Netherlands, pp. 221-224.

Bosch, J. M., and J. D. Hewlett (1982). A review of catchment experiments to determine the effect of vegetation changes in water yield and evapotranspiration, J. Hydrol. 55, 3-23.

Budyko, M.I. (1958). Heat balance of the earth's surface. U.S. Weather Bureau, Washington, D.C.

Burgess, S.S.O., Adams, M.A., Turner, N.C., Ong, C.K. (1998). The redistribution of soil water by tree roots, Oecologia, 115, 306-311.

Caldwell, M.M. and J.H. Richards (1989). Hydraulic lift water efflux from upper roots improves effectiveness of water uptake by deep roots, Oecologia, 79, 1-5.

Caldwell, M.M., T.E. Dawson, and J.H. Richards (1998). Hydraulic lift: consequences of water efflux from the roots of plants. Oecologia, 131, 151-161.

CasenaveA. and C. Valentin (1992). A runoff capability classification system based on surface feature criteria in semiarid areas of West Africa, J. Hydrol., 130, 231-249.

Chen, J. (1984). Uncoupled multi-layer model for transfer of latent and sensible heat flux densities from vegetation, Boundary Layer Meteorol., 28, 213-225.

Collatz, G.J., Ball, J.T., Grivet C., Berry J.A. (1991). Physiological and environmental regulation of stomatal conductance, photosynthesis and transpiration: a model that includes a laminar boundary layer, Agric. For. Meteorol. 54, 107-136.

Daamen, C.C. and L.P. Simmonds (1994). Soil, water, energy, and transpiration – a numerical modelof water and energy fluxes in soil profiles and sparse canopies, Department of Soil Science, University of Reading.

Daamen, C.C. (1997). Two-source model of surface fluxes for millet fields in Niger, Agric. For. Meteorol., 83, 205-230.

Dawson, T.E. (1993). Water sources of plants determined from xylem-water isotopic composition: perspectives on plant competition, distribution, and water relations, in J.R. Ehleringer, A.E. Hall, and G.D. Farquhar, G.D. (Eds.) Stable Isotopes and Plant Carbon/Water Relations, Academic Press, pp. 465–496.

DeBano, L.F. (1966). Formation of non-wettable soils involves heat transfer mechanism, USDA Forest Service Research Note PSW-132, 8pp.

DeBano, L.F. (2000). The role of fire and soil heating on water repellency in wildland environments: a review, J. Hydrol., 231, 195-206.

Dickinson R.E., A. Henderson-Sellers, PJ Kennedy, MF Wilson (1986). Biosphere-atmosphere transfer scheme (BATS) for the NCAR community climate model. NCAR Technical Note TN-275 + STR, 69 pp.

D'Odorico, P., L. Ridolfi, A. Porporato and I. Rodriguez-Iturbe (2000). Preferential states of seasonal soil moisture: the impact of climate fluctuations, Water Resour. Res., 36(8). 2209-2219.

D'Odorico P., and A. Porporato (2004). Preferential states in soil moisture and climate dynamics. Proc. Natnl. Acad. Sci. U.S.A, 101: 8848-8851, 10.1073/pnas.0401428101.

Doerr, S.H., R.A. Shakesby, and R.P.D. Walsh (2000). Soil water repellency: its causes, characteristics and hydrogeomorphological significance, Earth Sci. Rev., 51, 33-65.

Dunne, T and R.D. Black (1970). An experimental investigation of runoff prediction in permeable soils, Water Resour. Res., 6(2), 478-490.

Dunne, T. (1978). Field studies of hillsope flow processes, in M.J. Kirkby (Ed.) Hillslope Hydrology, J. Wiley& Sons, New York, pp. 227-293.

Eagleson, PS. (1978a). Climate, soil, and vegetation: The distribution of annual precipitation derived from observed storm sequences, Water Resour. Res. 14(5), 713-721.

Eagleson, PS. (1978b). Climate, soil, and vegetation: A simplified model of soil moisture movement in the liquid phase, Water Resour. Res. 14(5), 722-731.

Eagleson, PS., (1982). Land surface processes in atmospheric general circulation models, Cambridge University Press, Cambridge, UK.

Evett, S.R. and R.J. Lascano (1993). ENWATBAL.BAS: a mechanistic evapotranspiration model written in compiled BASIC, Agron. J., 85(3), 763-772.

Farquhar, G.D., von Caemmerer S., Berry J.A. (1980). A biochemical model of photosynthetic CO_2 assimilation in leaves of C_3 species. Planta 149, 78-90.

Feddes R.A., H. Hoff, M. Bruen, T. Dawson, P. de Rosnay, P. Dirmeyer, R. B. Jackson, P. Kabat, A. Kleidon, A., Lilly, and A. J. Pitman (2001). Modeling root-water uptake in hydrological and climate models, Bull. Am. Met. Soc., 82(12), 2797-2809.

Fernandez-Illescas CP, A. Porporato, F. Laio, and I. Rodriguez-Iturbe (2001). The ecohydrological role of soil texture in a water-limited ecosystem, Water Resour. Res., 37 (12): 2863-2872.

Gao, Q., P. Zhao, X. Zeng, X. Cai, and W. Shen (2002). A model of stomatal conductance to quantify the relationship between leaf transpiration, microclimate and soil water stress, Plant, Cell and Environment, 25, 1373-1381.

Green W.H. and C.A. Ampt (1911). Studies on soil physics, I. The flow of water and air though soils. J. Agric. Sci., 4, 1-24.

Guswa, A.J., M.A. Celia, and I. Rodriguez-Iturbe (2002). Models of soil moisture dynamics in ecohydrology: A comparative study, Water Resour. Res., 38 (9): Art. No. 1166.

Hale, M. G. and D. M. Orchutt (1987). The Physiology of Plants Under Stress, J. Wiley & Sons, New York.

Hopmans, J.W., and K.L.Bristow (2002). Current capabilities and future needs of root and nutrient uptake modeling, Advances in Agronomy, 77, 104-175.

Horton, R.E. (1033). The role of infiltration in the hydrologic cycle, Trans. Am. Geophys. U., 14, 446-460.

Horton, J.L., S.C. Hart (1998). Hydraulic lift: a potentially important ecosystem process. Tree 13 (6), 232-235.

Jarvis, P.G. (1976). The interpretation of the variations in leaf water potentialand stomatal conductance found in canopies in the field, Phil. Trans. R. Soc. Lond. B 273, 593-610.

Jones, H.G. (1992). Plants and microclimate, Cambridge University Press, 428 pp., Cambridge.

Jones, C.A. and J.R. Kiniry (1986). CERES-Maize: A Simulation Model of Maize Growth and Development, Texas A&M University Press, 194 pp.

Knapp, A. K., P. A. Fay, J. M. Blair, S. L. Collins, M. D. Smith, J. D. Carlisle, C. W. Harper, B. T. Danner, M. S. Lett, and J. K. McCarron (2002). Rainfall variability, carbon cycling and plant species diversity in a mesic grassland, Science, 298, 2202-2205.

Katul, G.G., and M.B. Parlange (1992). A Penman-Brutsaert model for wet surface evaporation, Water. Resour. Res., 28(1), 121-126.

Katul, G.G., R. Leuning and R. Oren (2003). Relationship between plant hydraulic and biochemical properties derived from a steady-state coupled water and carbon transport model, Plant, Cell and Environment, 26, 339-350.

Krammers, J.S. and L.F. DeBano (1965). Soil wettability: a neglected factor in watershed management, Water Resour. Res., 1, 283-286.

Lai, C-T, and G. Katul (2000). The dynamic role of root-water uptake in coupling potential and actual transpiration, Adv. Water Resour., 23, 427-439.

Laio F., A. Porporato, L., Ridolfi, and I. Rodriguez-Iturbe (2001a). Plants in water-controlled ecosystems: Active role in hydrological processes and response to water stress: II Probabilistic soil moisture dynamics, Adv. Water Res., 24 (7), 707-723.

Laio F., A. Porporato, C. P., Fernandez-Illescas, and I. Rodriguez-Iturbe (2001b). Plants in water-controlled ecosystems: Active role in hydrological processes and response to water stress: IV Discussion of real cases, Adv. Water Res., 24 (7), 745-762.

Lauenroth, W.K., O.E. Sala, D.P. Coffin, and T.B. Kirchner (1994). The importance of soil water recruitment of Bouteloua gracilis in the shortgrass steppe, Ecol. Appl., 4(4), 741-749.

Letey, J. (2001). Causes and consequences of fire-induced soil water repellency, Hydrol. Process., 15, 2867-2875.

Linn, D.M., and J. W. Doran (1984). Effect of water-filled pore space on carbon dioxide and nitrous oxide production in tilled and nontilled soils, Soil Science Society of America Journal, 48, 1267–1272.

Ludwig, F., T.E. Dawson, H., Kroon, F., Berendse, and H. H. T., Prins, (2002). Hydraulic lift in Acacia tortolis trees on and East African savanna. Oecologia 134, 293-300.

Lull, H.W. (1964). Ecological and silvicultural aspects. In V.T. Chow (Ed.), Handbook of applied hydrology, McGraw-Hill, New York, pp. 1 (6) 6-30.

M. Mainguet and M. Mainquet (1999). Aridity: Droughts and Human Development, Springer.

Milly, P.C.D. (1993). An analytic solution of the stochastic storage problem applicable to soil water. Water Resour. Res., 29, 3755-3758.

Milly, P.C.D. (2001). A minimalist probabilistic description of root-zone soil water, Water Resour. Res., 37(3), 457-463.

Norman, J.M. (1979). Modelling the complete canopy crop. In B. Barfield and J. Gerber (Eds.), Modifications of the aerial environment of crops, ASAE, St. Joseph, MI, pp. 249-277.

Noy-Meir, I. (1973). Desert ecosystems: environment and producers, Ann. Rev. Ecol. Systemat., 4, 25-51.

Parton, W.L., M. Hartman, D., Ojima, and D. Schimel (1998). DAYCENT and its land surface submodel: description and testing, Global and Planetary Change, 19: 35-48.

Peugeot C.M. Eteves, S. Galle, J.L. Rajot, and J. P. Vandervaere (1997). Runoff generation processes: results and analysis of field data collected at the East Central Supersite of the HAPEX-Sahel experiment, J. Hydrol., 188-189, 179-202.

Philip, J. R. (1975). Numerical solution of equations of the diffusion type with diffusivity concentrationdependent. Austral. J. Phys., 10, 29-42.

Philip, J. R. (1969). Theory of infiltration. In V. T. Chow, Advances in hydroscience, 5: 215-296. Academic Press, New York.

Porporato, A., P. D'Odorico, L. Ridolfi, and I. Rodriguez-Iturbe (2000). A spatial model for soil-atmosphere interaction: Model construction and linear stability analysis. Journal of Hydrometeorology 1:61-74.

Porporato, A., F. Laio, L., Ridolfi, and I. Rodriguez-Iturbe (2001)., Plants in water-controlled ecosystems: Active role in hydrological processes and response to water stress. III: Vegetation water stress. Adv. Water Resour., 24: 725-744.

Porporato, A, P. D'Odorico, F., Laio, and I. Rodriguez-Iturbe (2003). Soil moisture controls on the nitrogen cycle I: Modeling scheme, Adv. Water Res., 26, 45-58.

Richards L.A. (1931). Capillary conduction of liquids through porous medium, Physics (1), 318-333.

Richards, J.H. and M.M. Caldwell (1987). Hydraulic lift: substantial nocturnal water transport between soil layers by Artemisia tridentata roots. Oecologia 73, 486–489.

Richter, H. (1976). The water status in plants: Experimental evidence. In O.L. Lage, L. Kappen, and E-D Schulze (Eds.), Water and Plant Life: Problems and Modern Approaches, Springer-Verlag Berlin.

Ridolfi, L., P. D'Odorico, A. Porporato and I. Rodriguez-Iturbe (2000a). Duration and frequency of water stress in vegetation: an analytical model, Water Resour. Res., 36(8), 2297-2307.

Ridolfi, L., P. D'Odorico, A. Porporato and I. Rodriguez-Iturbe (2000b). Impact of climate variability on the vegetation water stress, J. Geophys. Res. (Atmospheres), 105(D14), 18013-18025.

Rodriguez-Iturbe, I., D. Entekhabi, and R. L. Bras (1991). Nonlinear dynamics of soil moisture at climate scales: 1. Stochastic analysis. Water Resour. Res., 27(8): 1899-1906.

Rodriguez-Iturbe, I, P. D'Odorico and A. Rinaldo (1998). A possible self-organizing dynamics for land-atmosphere interactions, J. Geophys. Res., 103(D18), 23,071-23,077.

Rodriguez-Iturbe, I., A. Porporato, L., Ridolfi, V., Isham, and D.R. Cox (1999a). Water balance at a point: the role of climate, soil and vegetation. Proc. R. Soc. Lond. A, 455, 3789-3809.

Rodriguez-Iturbe, I, P. D'Odorico, A. Porporato and L. Ridolfi (1999b). Tree-Grass coexistence in savannas: the role of spatial dynamics and climate fluctuations, Geophys. Res. Lett., 26(2), 247-250.

Rodriguez-Iturbe, I., P. D'Odorico, A. Porporato and L. Ridolfi (1999c). On the spatial and temporal links between vegetation, climate and soil moisture, Water Resour. Res., 35(12), 3709-3722.

Schulze E.D., M.M. Caldwell, J. Canadell, H.A. Mooney, R.B. Jackson, D. Parson, R. Scholes, O.E. Sala, and P. Timborn (1998). Downward flux of water through roots (i.e. inverse hydraulic lift) in dry Kalahari sands, Oecologia, 115, 460-462.

Sellers, P.J., Y. Mintz, Y.C. Sud, and A. Delcher (1986). A simple biosphere model (SiB) for use with general circulation models, J Atmos. Sci., 43, 505-531.

Skopp, J., M.D. Jawson, and J.W. Doran (1990). Steady-state aerobic microbial activity as a function of soil water content, Soil Sci. Soc. Am. J., 54, 1619-1625.

Smith, J.A.C., and H. Griffiths (1993). Water deficits: plant responses from cell to community, Bios, Oxford.

Smith, D.M., N.A. Jackson, J.M. Roberts, and C.K. Ong (1999). Reverse flow of sap in tree roots and downward siphoning of water by Grevillea robusta, Functional Ecology 12, 256-264.

Smith, R.E. and J-Y Parlange (1978). A parameter-efficient hydrologic infiltration model, Water Resour. Res., 14(3), 533-538.

Smith R.E., K.R.J. Smettem, P. Broadbridge, and D.A. Woolhiser (2002). Infiltration Theory for hydrologic applications, American Geophysical Union, Washington, D.C.

Thornthwaite, C. W., and J. R. Mather (1955). The water balance. In: Publications in Climatology, Drexel Institute of Technology, 8(1), 1-104.

Todorovic, P. and D.A. Woolhiser (1975). A stochastic model of n-day precipitation, J. Appl. Meteorol., 14, 17-24.

Xue, Y., P.J. Sellers, J.L. Kinter, and J. Shukla (1991). A simplified biosphere model for global climate studies, J. Clim., 4, 345-364.

Xue, Y., F.J. Zeng, and C.A. Schlosser (1996). SSib and its sensitivity to soil properties – a case study using HAPEX-Mobilhi data, Global and Planetary Change., 13, 183-194.