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On the importance of accurate depiction of infiltration processes on modelled soil moisture and vegetation water stress

Salvatore Manfreda,[1]* Todd M. Scanlon[2] and Kelly K. Caylor[3]

[1] *Dipartimento di Ingegneria e Fisica dell'Ambiente, Università degli Studi della Basilicata, Potenza I-85100, Italy*
[2] *Department of Environmental Sciences, University of Virginia, Charlottesville, VA 22903, USA*
[3] *Civil and Environmental Engineering, Princeton University, Princeton, NJ 08544, USA*

ABSTRACT

The description of soil moisture dynamics is a challenging problem for the hydrological community, as it is governed by complex interactions between climate, soil and vegetation. Recent research has achieved significant advances in the description of temporal dynamics of soil water balance through the use of a stochastic differential equation proposed by Laio *et al.* (2001). The assumptions of the Laio *et al.* model simplify the mathematical form of the soil water loss functions and the infiltration process. In particular, runoff occurs only for saturation excess, the probability distribution function (PDF) of which is well-represented by a simple expression, but the model does not consider the limited infiltration capacity of soil. In the present work, we extend the soil moisture model to include limitations on soil infiltration capacity with the aim of understanding the impact of varying infiltration processes on the soil water balance and vegetation stress. A comparison between the two models (the original version and the modified one) is carried out via numerical simulations. The limited infiltration capacity influences the soil moisture PDF by reducing its mean and variance. Major changes in the PDFs are found for climates characterized by storms of short duration and high rainfall intensity, as well as in humid climates and in cases where soils have moderate permeability (e.g. loam and clay soils). In the case of limited infiltration capacity, modifications to the dynamics of soil moisture generally lead to higher amounts of vegetation water stress. An investigation of the role of soil texture on vegetation water stress demonstrates that loam soil provides the most favorable condition for plant-growth under arid and semi-arid conditions, while vegetation may benefit from the presence of more permeable soils (e.g. loamy sand) in humid climates. Copyright © 2009 John Wiley & Sons, Ltd.

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INTRODUCTION

Soil moisture is a key variable for ecohydrological modeling (e.g. Eagleson, 1982; Neilson, 1995; Rodriguez-Iturbe *et al.*, 2000). Its evolution in time and space is driven by different processes acting over a variety of scales (e.g. Albertson and Montaldo, 2003; Rodríguez-Iturbe *et al.*, 2006; Manfreda *et al.*, 2007). The severity and persistence of water stress in plants, the outcomes of ecological competition, and the sustainability of vegetation communities are examples of important ecological research questions in which soil moisture dynamic plays a dominant role (e.g. Scholes and Archer, 1997; Porporato *et al.*, 2001; Rodriguez-Iturbe *et al.*, 2001; Sofo *et al.*, 2008). In particular, vegetation water stress is intimately related to relative soil moisture and the length of time that the soil moisture is below a given threshold. The crossing properties of the soil moisture levels are controlled by the drying process and the infiltration inputs into the soil matrix. This last property varies from soil to soil according to the texture and the permeability.

Recent research has achieved significant progress in the description of soil moisture dynamics through the development of a steady-state probability density function of soil moisture within the growing season (Rodriguez-Iturbe *et al.*, 1999; Laio *et al.*, 2001). This approach is based on the steady-state solution of the stochastic differential equation for the soil water balance in which the rainfall represents the stochastic forcing. Although this model necessarily contains assumptions to simplify the mathematical form of the stochastic differential equation used to derive the soil moisture PDF, it represents the most innovative and general method to describe, within a physically-based approach, the soil moisture dynamics. More specifically, this theory is based on the assumption that the infiltration is an additive noise in the soil water balance and it is equal to the rainfall pulse unless the rainfall produces soil saturation. In this case, runoff is produced for saturation excess that is linearly related to the rainfall depth. A different infiltration scheme would introduce a non-linear relationship between the rainfall forcing and the infiltration, which

* Correspondence to: Salvatore Manfreda, Dipartimento di Ingegneria e Fisica dell'Ambiente, Università degli Studi della Basilicata, Potenza I-85100, Italy. E-mail: salvatore.manfreda@unibas.it

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would be too complex to be handled in a theoretical framework.

A significant amount of work has examined this original theory. Some have tested the model with experimental data (e.g. Salvucci, 2001) or numerical analysis (Guswa *et al.*, 2002), while others have evaluated the infiltration scheme or the hypothesis of stationarity of the climatic forcing in order to understand the range of applicability of the theory (e.g. Manfreda *et al.*, 2004; Rigby and Porporato, 2006; Viola *et al.*, 2008). In particular, Rigby and Porporato (2006) compared the model proposed by Laio *et al.* (2001) with the one derived by Eagleson (1978a,b) and Kim *et al.* (1996) over a limited set of soil textures and climatic conditions and observed minimal differences between the two models.

In the present work, the scheme adopted to describe the infiltration process in the soil water balance model is tested over a wide range of climatic conditions and with several different textures in order to define more specifically the range of applicability of the first model and the consequences of a limited infiltration capacity on both the soil moisture PDF and the vegetation stress. Most of the assumptions proposed by Laio *et al.* (2001) have been preserved, but here we also consider runoff production caused by limited infiltration capacity of soil. Saturation excess alone is adequate for the evaluation of the runoff production in some environments, where infiltration capacity of soils is generally much higher than rainfall intensities. However, when the infiltration capacity of the soil becomes similar in magnitude to rainfall intensity, infiltration excess should be taken into account.

We aim to investigate the main differences between the original model proposed by Laio *et al.* (2001) and the modified model proposed herein (Infiltration Excess model). The probability distributions of the soil moisture derived from the two models are compared to determine the effects of the infiltration process schemes on the soil water balance and the vegetation water stress assuming (1) saturation excess runoff, and (2) Infiltration Excess runoff production. The mean and the standard deviation (SD) of the soil moisture obtained with these two different approaches are compared for different climates and soil characteristics. The vegetation water stress is computed using the same theoretical framework traced by Porporato *et al.* (2001).

In the following sections, the conceptual model used by Laio *et al.* (2001) to derive analytically the soil moisture PDF is briefly described. The modifications introduced to account for storm duration and non-linearity in the infiltration process are introduced in the Section on Including the Infiltration Excess Process in the Soil Water Balance. In the Section on Comparison of the Soil Moisture Dynamics, the statistics obtained using these two different infiltration schemes are compared and the implications of the surface control on the soil moisture dynamics are discussed. Finally, in the Section on The Vegetation Water Stress under Two Different Infiltration Schemes, the effects on the dynamic water stress (Porporato *et al.*, 2001) associated with different soil textures and climatic conditions are described.

SOIL MOISTURE MODEL

This model was proposed by Laio *et al.* (2001). Soil water balance may be described through the use of a bucket scheme as first suggested by Manabe (1969). Many others have used the same idea with different aims (e.g. Milly 1994; Kim *et al.*, 1996; Farmer *et al.*, 2003; Porporato *et al.*, 2004). This interpretation is extremely useful because it allows the use of the water balance equation with a finite control volume generally represented by the root zone. Such an assumption is at the core of the model proposed by Laio *et al.* (2001), which is based on the following equation:

$$nZ_r \frac{ds}{dt} = I - ET - L \quad (1)$$

where s is the relative saturation of the soil given by the ratio of the volumetric soil moisture θ (dimensionless) and the soil porosity n (dimensionless); Z_r is the root zone depth (L), I represents the infiltration rate (L T^{-1}), ET the actual evapotranspiration and L the leakage rates (L T^{-1}).

Infiltration, I (L T^{-1}), is interpreted with a simplified scheme particularly useful for analytical purposes. It is assumed equal to the daily rainfall depth, h, if the water deficit $nZ_r(1-s)$ is greater than h and $nZ_r(1-s)$ otherwise. Infiltration assumes the following form:

$$I = \begin{cases} h & h \le nZ_r(1-s) \\ nZ_r(1-s) & h > nZ_r(1-s) \end{cases} \quad (2)$$

where nZ_r (L) represents the soil water content at saturation and h (L) is the rainfall depth. This representation allows for an immediate definition of the infiltration PDF that assumes the same distribution of the rainfall depth (exponential distribution) with an atom probability at the value $nZ_r(1-s)$.

The soil water loss function accounts for two phenomena: evapotranspiration and leakage. Both are described though a deterministic function that depends on the actual value of s. In particular, evapotranspiration assumes four different behaviors conditional to the relative state of the soil moisture:

$$ET(s) = \begin{cases} 0 & s \le s_h \\ \dfrac{s - s_h}{s_w - s_h} E_w & s_h \le s \le s_w \\ E_w + \dfrac{s - s_w}{s^* - s_w}(E_{\max} - E_w) & s_w \le s \le s^* \\ E_{\max} & s \ge s^* \end{cases} \quad (3)$$

where, E_w (L) is the evapotranspiration at the wilting point s_w, $E_{\max}$ (L) is the evapotranspiration at the initial stomata closure s^*, and s_h is the soil water content at which the ET reaches the zero.

The leakage function is described by:

$$L(s) = \begin{cases} 0 & s \le s_{fc} \\ K_s s^c & s > s_{fc} \end{cases} \quad (4)$$

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where K_s is the soil permeability at saturation (L T^{-1}), s_{fc} is the soil moisture content at the field capacity, $c = (2 + 3m)/m$ (dimensionless) is the pore disconnectedness index and m (dimensionless) is the pore-size distribution index. For analytical purposes, Laio *et al*. (2001) modified Equation (4) using an exponential approximation.

An example of the soil water loss function, $\chi(s) = ET(s) + L(s)$, is given in Figure 1 for two specific soil textures. Soil parameters are taken from in Table I, E_w is 0·01 cm/day and $E_{\max}$ is equal 0·45 cm/day.

INCLUDING THE INFILTRATION EXCESS PROCESS IN THE SOIL WATER BALANCE

The previous model has been modified to include a different infiltration mechanism that accounts for the limited infiltration capacity of soil, and also for the effect of rainfall duration in order to provide a more accurate estimation of the soil moisture dynamics. The infiltration is considered as a daily input in the soil matrix, but it is computed as the integral of the Philip's (1960) equation over the rainfall duration. In this scheme, infiltration depends on the rainfall occurrence, intensity, and also duration. This introduces an additional random variable into the soil moisture model represented by the duration of storm events. Consequently, rainfall is described as a Poisson process of rainfall pulses with exponentially distributed and statistically independent total depths and durations. As it will be further addressed in the next paragraphs, the soil moisture distribution seems to be particularly sensitive to this last parameter.

Rainfall forcing

Rainfall is considered as a Poisson process of daily occurrences, where the storm depths are generated according to an exponential distribution $p(h) = 1/\alpha \exp(-h/\alpha)$, where α (L) represents the mean rainfall depth. Similarly, the rainfall arrivals are randomly generated with parameter λ (T^{-1}) representing the mean storms arrivals (Eagleson, 1978c). These two parameters are representative of the local climate and together define the total amount of rainfall during a wet season.

In the second model, each rainfall pulse is assigned a storm duration that is also exponentially distributed with mean duration δ (T). A rainfall pulse is therefore defined by two components that are both relevant: the rainfall depth and its duration. In this analysis, we consider these two variables to be independent. The characteristics of the modeled rainfall process are illustrated in Figure 2, where a sequence of pulses is shown.

In the case of arid climates, the rare rainy storm tends to be extremely short in time and likely with high intensity. Such a condition may inhibit the infiltration process and, at the same time, may increase runoff production while being interpreted as a loss in the soil water budget.

In Figure 3, the probability distributions of the storm durations are drawn for two different rain stations located in arid areas in two different continents. Figure 3a refers to 10 years of hourly rainfall records of station 44 of the Sevillata research area (http://sevilleta.unm.edu/)

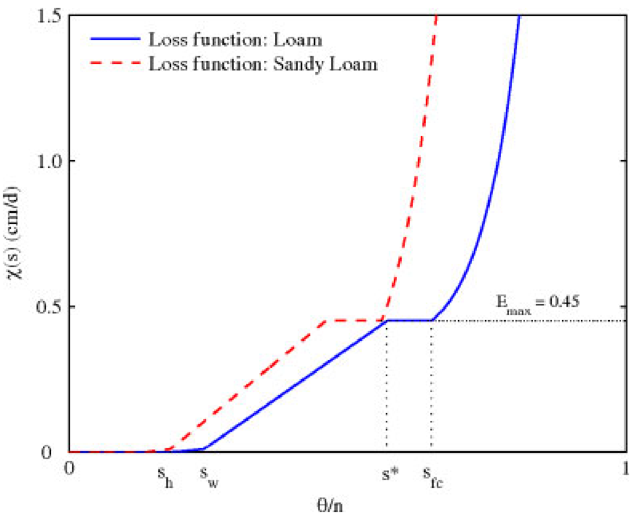

Figure 1. The water loss function given by the sum of evapotranspiration and leakage: $\chi(s) = ET(s) + L(s)$.

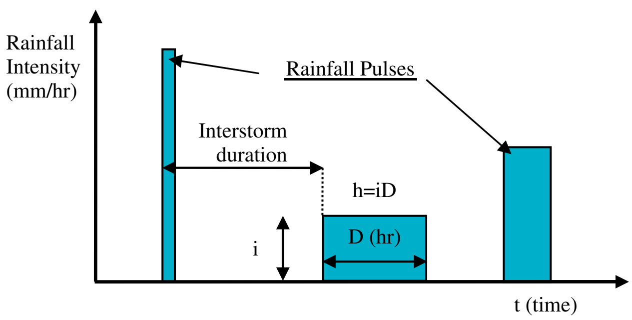

Figure 2. Rainfall scheme of random pulses with random durations and depths mutually independent.

Table I. Soil parameters associated with each of the soil textures taken from Fernandez-Illescas *et al*. (2001) according to the regression equations of Cosby *et al*. (1984).

Soil type	$\psi(1)$ (cm)	c	m	K_s (cm/d)	s_h	s_w	s^*	s_{fc}	n	s_{PAW}
Sand	4·7	9·8	0·30	203·7	0·05	0·07	0·21	0·29	0·37	0·22
Laomy sand	6·4	10·7	0·26	143·2	0·08	0·11	0·28	0·37	0·38	0·26
Sandy loam	13·2	12·0	0·22	61·4	0·14	0·18	0·39	0·50	0·41	0·32
Loam	20·7	14·4	0·18	36·2	0·23	0·28	0·51	0·62	0·43	0·34
Clay	39·1	27·3	0·08	17·3	0·52	0·58	0·77	0·84	0·46	0·26

Parameters s_h, s_w, s^* and s_{fc} correspond to a soil matric potential of −10 MPa, −3 MPa, −0·09 MPa and −0·03 MPa.

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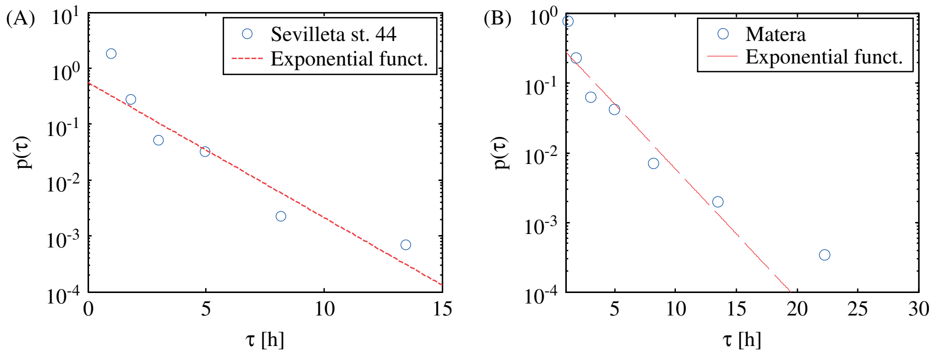

Figure 3. A) Probability distribution of the rainfall storm durations (circles) at the station 44 of the Sevilleta research area (USA), and exponential distribution with mean equal to 1·7 h (dashed line) in semilog plot. B) Similarly, for the station of Matera (Italy), where the mean of the storm duration is equal to 2·3 h.

in New Mexico (Caylor *et al.*, 2005), while Figure 3b refers to a similar dataset recorded at the rain gauge station of Matera, Italy. The probability distributions of both the records approximate an exponential distribution with mean equal to 1·7 h and 2·3 h, respectively. These examples are useful to describe or to give an idea of possible distribution of storm durations in arid climates. These two examples confirm that the storm durations may be very short in time producing a shower of high intensity especially during the growing season.

The assumption of rainfall duration exponentially distributed is widely accepted in the literature (e.g. Eagleson, 1972; Rodriguez-Iturbe *et al*, 1987; Veneziano and Iacobellis, 2002). Nevertheless, the assumption that rainfall depth and duration are exponentially distributed and independent produces a Cauchy distribution (Rigby and Porporato, 2006) that may produce an overestimation of the frequency of rainfall events of high intensity. One can overcome this difficulty by using a probability distribution of durations conditional on rainfall depth. While this may be important for representing actual data, this more advanced treatment of the rainfall statistics was not considered in the present paper for the sake of aiding interpretation of the results.

The limited infiltration capacity of the soil

The method used to calculate the potential and actual infiltration rates are based on Philip's equation (Philip, 1960). This equation requires supplementary information about soil characteristics such as matrix potential curve, pore size distribution index, sorptivity and permeability. It is necessary to remark that infiltration may be also limited by water repellency (e.g. DeBano, 2000; Doerr *et al.*, 2000) that is not considered in the present work.

The relationship between matric potential and relative soil saturation is described by using results of Burdine (1958) and Brooks and Corey (1966): $K(s) = K_s\ s^c$ and $\psi(s) = \psi(1)s^{-1/m}$; where $K(s)$ (L T^{-1}) is the soil permeability and $\psi(s)$ (L) is the matrix potential, while c and m are empirical parameters.

The infiltration capacity of the soil using the Philip's equation is as follows:

$$f(t) \approx \frac{1}{2}S(s_0)t^{-1/2} + A(s_0) \tag{5}$$

where the parameters $S(s_0)$ infiltration sorptivity and $A(s_0)$ gravitational infiltration are respectively defined as $A(s_0) = \frac{1}{2}[K_S - K(s_0)]$ and $S(s_0) = 2n(1 - s_0)\sqrt{\frac{\overline{D}}{\pi}}$, where s_0 is the relative soil moisture at time *0* representing the beginning of rainfall event; $\overline{D}$ is the effective diffusivity of soil.

Following Eagleson (1978c), we express parameters $A(s_0)$ and $S(s_0)$ as a function of the hydraulic soil parameters and the initial soil moisture condition:

$$A(s_0) = \frac{1}{2}K_S\left[1 - s_0^c\right] \tag{6}$$

$$S(s_0) = 2(1 - s_0)\left(\frac{5nK_S\psi(1)\phi_0(m, s_0)}{3m\pi}\right)^{1/2} \tag{7}$$

where m is the pore size distribution index, c is the pore disconnectedness index and $\phi_0(m, s_0)$ dimensionless effective diffusivity.

Dimensionless effective diffusivity can be written as a function of the pore-size distribution index and the initial soil moisture s_0 (Bras, 1990):

$$\phi_0(m, s_0) = \frac{3\pi}{10(1 - s_0)}\left(\frac{m}{1 + 4m} + \frac{m^2 s_0^{1/m+4}}{(1 + 4m)(1 + 3m)} - \frac{m^2 s_0}{1 + 3m}\right) \tag{8}$$

The total infiltration during a storm event with duration D may be computed by dividing the rainfall pulse in two intervals according to the magnitude of the infiltration capacity with respect to the rainfall intensity, i. The two rates are equal when $t_0 = \frac{S(s_0)^2}{4\,(i - A(s_0))^2}$. Nevertheless, the rainfall intensity, i, is lower than the infiltration rate when $t < t_0$ consequently the soil's infiltration capacity decreases slower than $f(t)$ and the actual ponding time is higher than t_0. Eagleson (1978c) shows that, under the assumption $i \gg A_0$, the actual ponding time can

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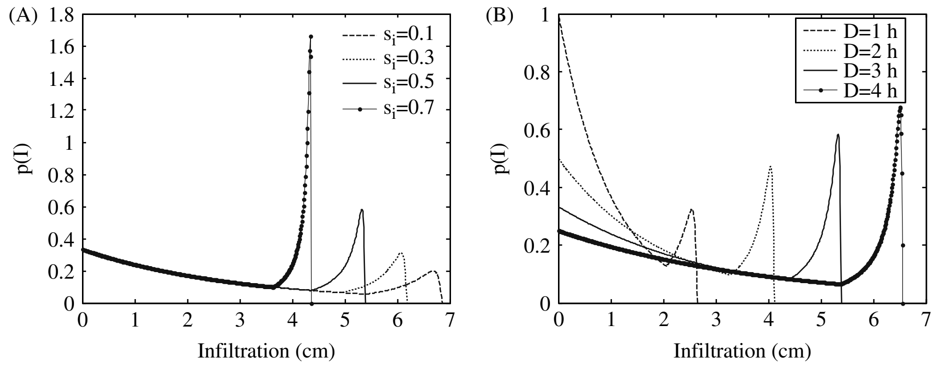

Figure 4. Probability density function of the infiltration into the soil using the hypothesis of limited infiltration capacity for different values of initial soil moisture (A), and for rainfall pulse of 1, 2, 3 and 4 h duration (B). Parameters used are $\lambda = 0{\cdot}30$ event/day, $\alpha = 1{\cdot}0$ cm/day, $\delta = 3$ h (only in A), $s_i = 0{\cdot}5$ (only in B) and the soil texture is loam (Table I).

be assumed equal to $2t_0$ in order to account for the surface saturation effect. The described value for the ponding time produces a slight overestimate with respect to the exact solution obtained from the equation $\int_0^{t_p} i dt = \int_0^{t_0} f(t)dt$.

Consequently, the rainfall occurring in a duration lower than $2t_0$ will saturate the soil surface and infiltrate gradually into the soil after the rain event. It follows that the infiltration due to a rainfall event of intensity, i, and a duration, D, may be computed as

$$I = \begin{cases} iD & D > 2t_0 \\ 2it_0 + \int_{t_0}^{D-t_0} \left(\frac{1}{2}S(s_0)t^{-1/2} + A(s_0)\right) dt & D \le 2t_0, \end{cases} \quad (9)$$

which leads to

$$I = \begin{cases} iD & D \le 2t_0 \\ 2it_0 + A(s_0)(D - 2t_0) + S(s_0)\sqrt{D - t_0} - S(s_0)\sqrt{t_0} & D > 2t_0. \end{cases} \quad (10)$$

The infiltration equation given above allows the derivation of the probability distribution of infiltration into the soil under the hypothesis of rainfall pulses of assigned duration and total depths exponentially distributed. An example of these distributions is given in Figure 4 where one can appreciate how the PDF of the infiltration deviates from the original exponential function of the rainfall. The graphs describe the PDFs of the infiltration, I, assuming different soil moisture states (Figure 4a) and different storm durations (Figure 4b) for a loamy soil under a specific climate.

In the present work, the rainfall pulses are assumed to be characterized by random durations and depths. Consequently, a numerical approach was used in order to derive the PDFs of the soil moisture under several climatic conditions and soil types. The soil parameters adopted in the present work are taken from Fernandez-Illescas *et al.* (2001) that derived such values according to the univariate regression equations defined by Cosby *et al.* (1984) and are summarized in Table I. A visual description of the range of water content for the considered soil textures is given in Figure 5. In this graph, the plant-available soil water (PASW) is also depicted along with the most significant hydraulic soil parameters (hygroscopic point, wilting point, field capacity and porosity). PASW increases as one proceeds along the scale of soil-texture from sand to clay reaching a maximum in loam soil. This fact is particularly relevant for the dynamic of vegetation that may benefit from a higher available water storage capacity.

The two models considered herein differ only in the description of the infiltration input in the soil matrix. In particular, the Infiltration Excess model exploits Equation (10) to account for the limited infiltration capacity of soil. Nevertheless, saturation excess occasionally may occur also in this model because of the limited storage capacity of the soil and also because infiltration is computed as the integral over the rainfall duration and it is added to the soil water balance as an instantaneous input in the soil water balance.

COMPARISON OF THE SOIL MOISTURE DYNAMICS

Both models have been simulated numerically, adopting for both the same time series of rainfall depths in order to avoid problems in the inter-comparison of the results. Furthermore, the soil water losses have been computed using the analytical expression of the soil moisture decay during a dry phase obtained from the soil water balance differential equation [see Equation (20) in Laio *et al.* (2001)], thereby avoiding errors associated to with the time discretization.

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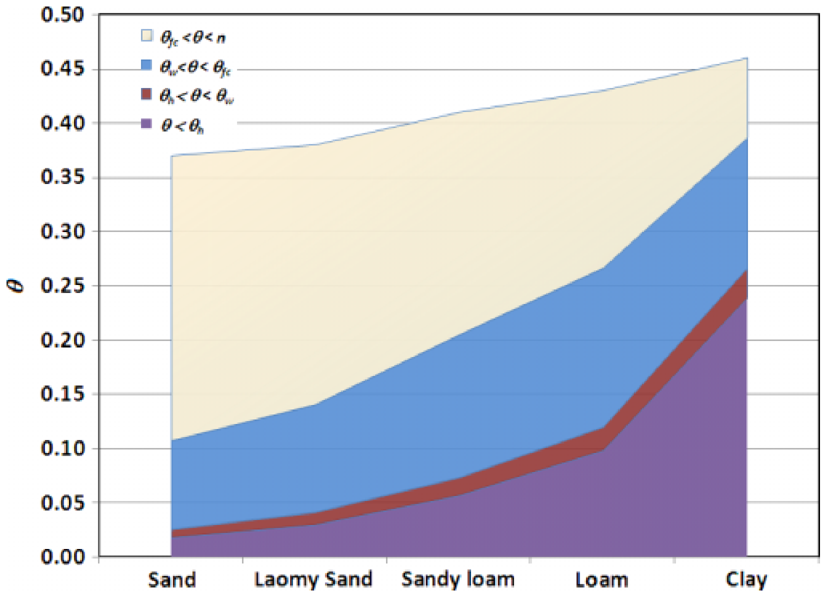

Figure 5. Range of water contents for five different soil texture: soil porosity n, field capacity (θ_{fc}), wilting point (θ_w), hygroscopic content (θ_h) and the plant-available soil water (PASW $= \theta_{fc} - \theta_w$).

Simulations are carried out over a wide range of climatic conditions using different soil textures whose characteristics are described in Table I. For the sake of brevity, the graphs reported herein refer only to the four different soil textures that represent common soil types. The temporal window of simulation is 50 years in order to obtain sufficiently stable results. An example of a 2-year run is given in Figure 6, where the soil moisture evolution in time for the two proposed schemes is depicted. The paths of the two soil moisture models are almost the same. The Infiltration Excess model slightly deviates from the path of the model based only on Saturation Excess mechanism when the relative saturation of the soil gets higher and especially when intense rainfall occurs.

In Figure 7, eight examples of PDFs referred to both the original model by Laio *et al.* (2001) and the model

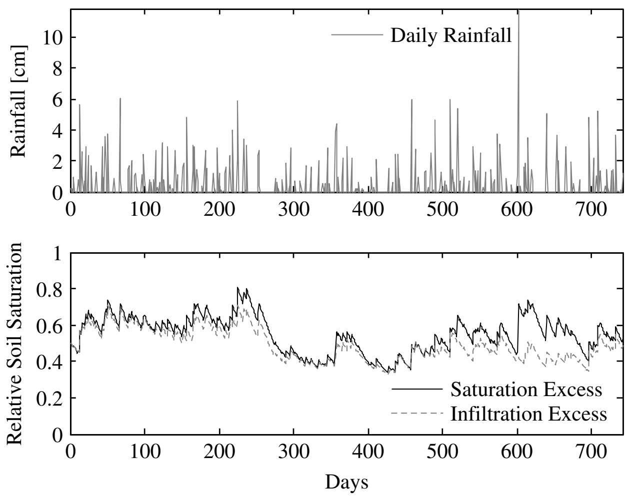

Figure 6. An example of soil moisture dynamics driven by stochastic rainfall obtained using two schemes for the infiltration process. Parameters used for the simulation are $\lambda = 0{\cdot}30$ event/day, $\alpha = 1{\cdot}5$ cm/day, $\delta = 3$ h, $Z_r = 30$ cm, E_w is 0·01 cm/day, $E_{\max}$ is equal 0·45 cm/day and the soil texture is loam (see Table I).

with Infiltration Excess are plotted in order to compare the soil moisture dynamics in two different rainfall regimes and for different soil textures. The two rainfall regimes refer to an arid climate with parameters $\alpha = 1{\cdot}0$ cm/day, $\lambda = 0{\cdot}1$ event/day and $\delta = 3{\cdot}0$ h and to a humid regime with rainfall parameters $\alpha = 1{\cdot}5$ cm/day, $\lambda = 0{\cdot}30$ event/day, and $\delta = 3{\cdot}0$ h. As a general remark, the differences between the two models were negligible in the case of sandy soil and obviously become more relevant for less permeable soils such as loam and clay (Figure 7F and H). In the arid climate (Figure 7A, C, E and H), it is clear that the two distributions are practically identical in all the soil texture types; while in humid conditions (Figure 7B, D, F and G), the differences between the two distributions are more significant especially with regard to the right tail of the probability distribution and for the less permeable soils like loam and clay.

With the aim to provide a quantitative comparison between the two simulation schemes, the mean and the SD of the soil moisture have been estimated using different soil textures and different values of the rainfall parameters (α and λ). These parameters characterize the local climate conditions that are assumed to vary from severe arid ($\alpha = 0{\cdot}1$ and $\lambda = 0{\cdot}1$) to extremely humid condition ($\alpha = 1{\cdot}5$ and $\lambda = 0{\cdot}4$). Furthermore, the comparison is carried out for three different values of the mean storm duration δ (1·5, 3·0, 5·0 h). The differences between the resulting soil moisture PDF obtained by the two models are summarized in Figures 8 and 9. In particular, the graphs describe the relative change in mean and SD of the soil moisture obtained with the first model (Saturation Excess runoff mechanism) with respect to the second one (Infiltration Excess and Saturation Excess runoff mechanism) which expresses in percentage difference. This relative change, Δ, is generally positive for

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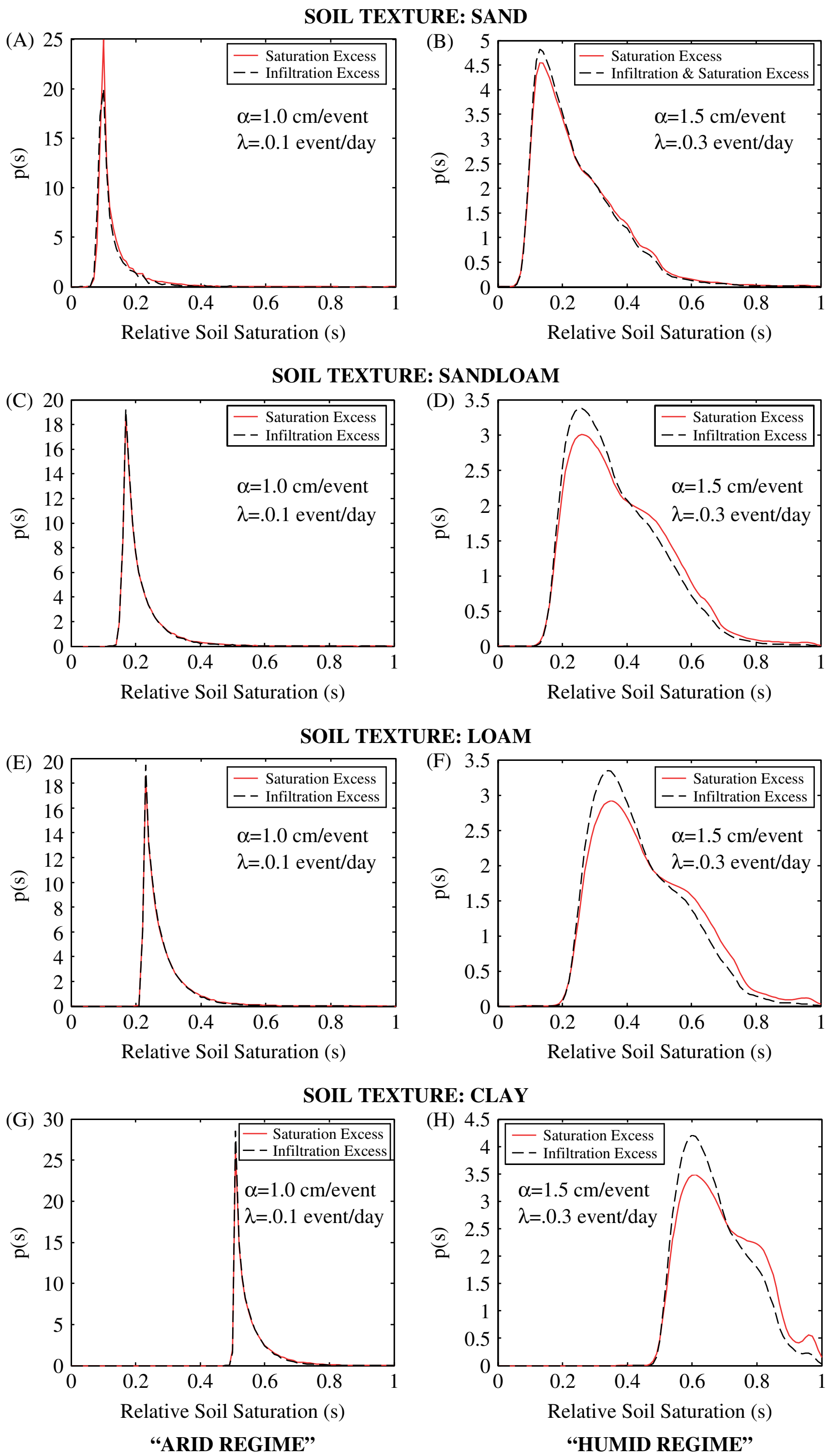

Figure 7. Comparison of the soil moisture PDFs obtained with the two soil moisture models for two different climatic conditions (on the columns) and four different soil textures (on the rows). Soil parameters are taken from Table I and other parameters are the same as of Figure 5.

the mean and the SD of the soil moisture that are overestimated by the first model. The challenge is to understand when and where such over or underestimations are important. A first attempt to define this range is made here assuming a significant threshold value $\Delta = \pm 5\%$ and 10% and these limits are depicted in the graphs with a continuous line with two distinct colors (red and magenta).

The comparison of the two schemes allows us to understand the implications that the choice of an infiltration scheme has on the soil moisture under different climatic conditions. The analyses provided the following results: (1) the errors in the estimation of the mean of the first model depend on the mean rainfall rate ($\alpha\lambda$), while the SD seems to be more markedly influenced by the rainfall intensity (controlled by the parameters α and δ);

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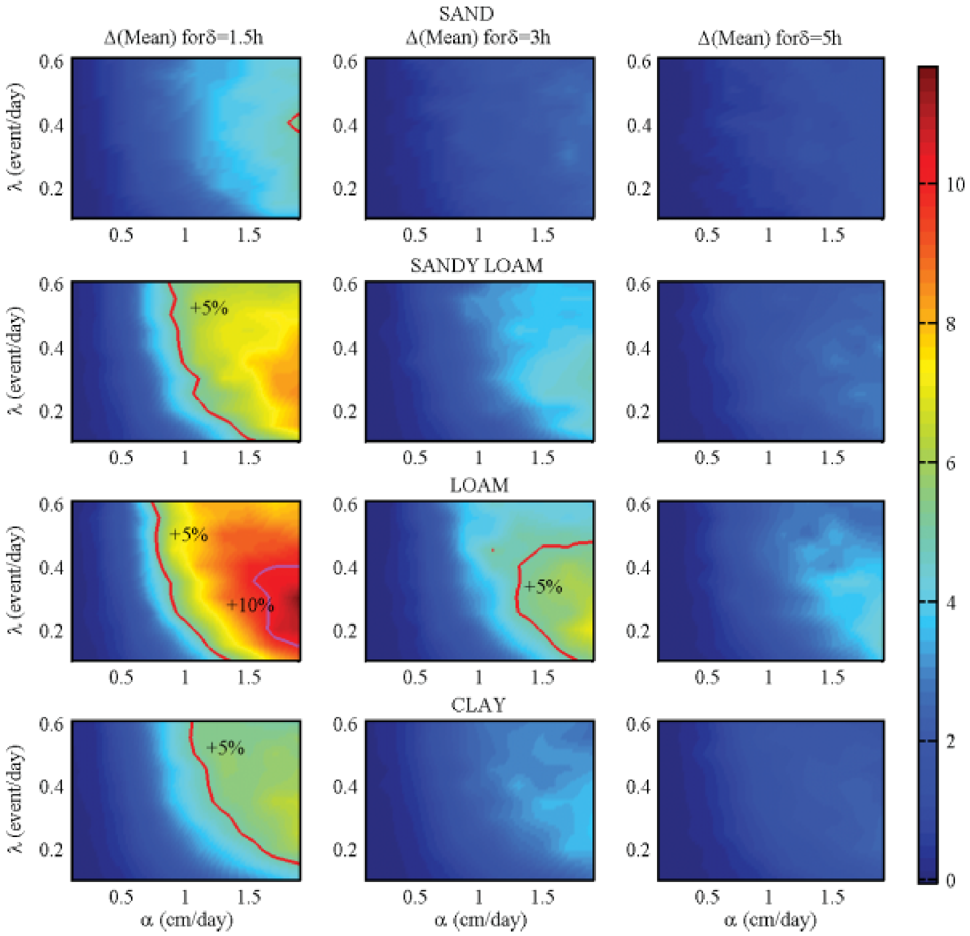

Figure 8. Percentage differences (Δ) between the mean soil moisture obtained using the original model by Laio *et al.* (2001) and modified version with limited infiltration capacity. The Saturation Excess model provides an accurate estimate of the mean with errors always lower than the 18%, the errors decrease with the increase of the mean storm duration values δ. From the left to the right the mean storm duration varies from 1·5, 3 to 5 h. Adopted parameters for the simulations are the same as of Figure 5, and the soil parameters for each texture are taken from Table I.

(2) the Saturation Excess scheme may produce significant overestimation of the soil moisture variance, while overestimation of the mean values is always minor; (3) the relative changes in the SD and mean are most pronounced for a loam soil texture. This last result is due to the changes in s_h with the soil texture. In particular, this parameter represents a lower limit for the soil moisture that reduces its variance. In case of clay soils s_h assumes a particularly high value limiting the range of variability of the soil moisture and also the relative changes between the two simulation schemes. Under those conditions the relative changes for clay in the mean and the SD of soil moisture are lower than those measured for loam.

The inclusion of infiltration excess in the model does not lead to significant differences in the PDFs of soil moisture in the case of arid climate with small amount of rainfall. However, increasing mean rainfall intensity may cause the inhibition of infiltration, thereby reducing accuracy of the original model even in arid conditions. Differences between the two schemes are essentially due to the overestimation of the infiltration term in the water balance equation. Limited infiltration capacity may reduce the mean soil moisture value, but its major effect is observed on the SD of the soil moisture. Of particular interest is the fact that the soil moisture is highly sensitive to the mean storm duration. In fact, a reduction in duration strongly increases the runoff production and affects the mean and variance of soil moisture (Figures 8 and 9).

THE VEGETATION WATER STRESS UNDER TWO DIFFERENT INFILTRATION SCHEMES

The dynamic water stress of vegetation was evaluated under the two different schemes to characterize the possible implications of limited infiltration capacity on the vegetation state. To this end, we referred to the theoretical scheme proposed by Porporato *et al.* (2001). In the following, we report the main concept used to derive the dynamic water stress index, but for reason of brevity we do not include the details.

The so-called ‘static’ water stress, ζ, measures the state of stress of the plants as a function of the relative saturation of soil (Porporato *et al.*, 2001):

$$\zeta(t) = \begin{cases} 1 & if \quad s(t) \le s_w \\ \left[\dfrac{s^* - s(t)}{s^* - s_w}\right]^q & if \quad s_w \le s(t) \le s^* \\ 0 & if \quad s(t) > s^* \end{cases} \tag{11}$$

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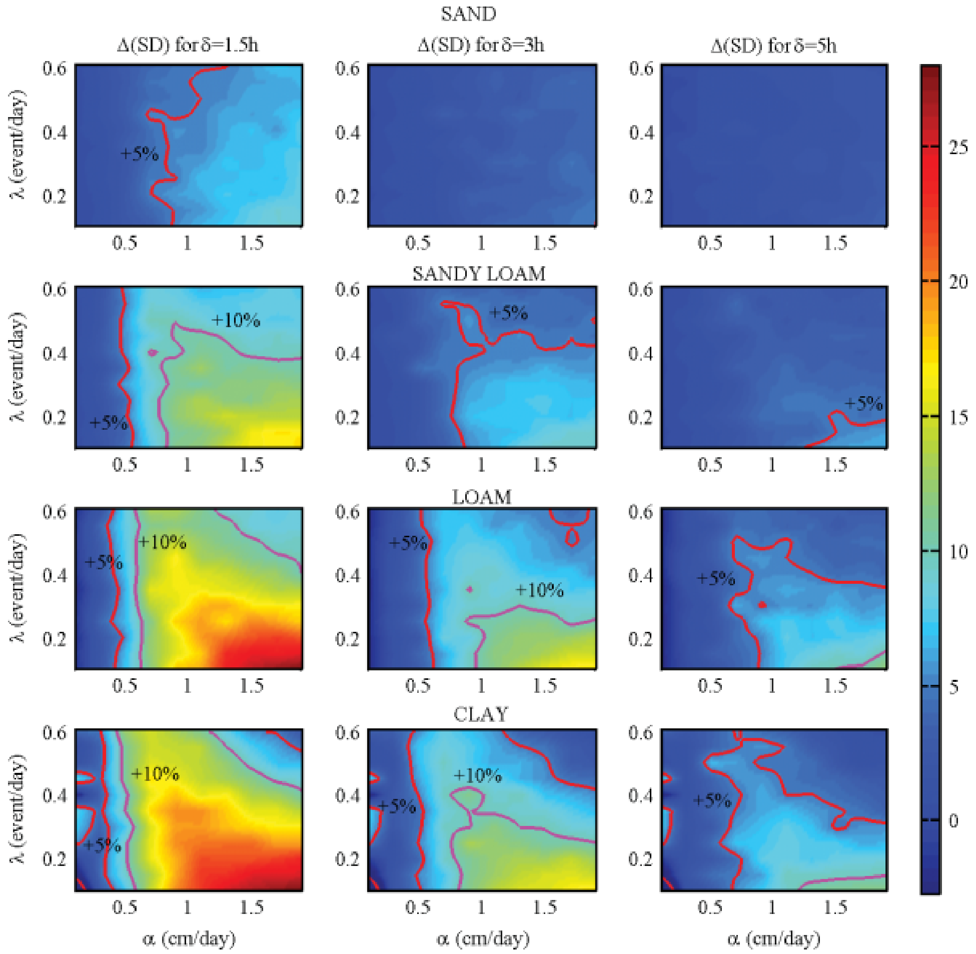

Figure 9. Percentage differences (Δ) between the SD of soil moisture obtained using the original model by Laio *et al.* (2001) and modified version with limited infiltration capacity. The graphs highlight the presence of large areas where the Saturation Excess scheme produces an overestimation of the SD up to 30% for humid climates with short storm durations. Adopted parameters for the simulations are the same as of Figure 5 and the soil parameters for each texture are taken from Table I.

where the exponent q accounts for the non-linear relationship between the plant stress and the soil water content. The static stress does not account for the temporal dynamic of soil moisture; for this reason Porporato *et al.* (2001) introduced the two new stochastic variables: T_ξ the length of the time intervals in which the soil moisture is below a threshold ξ, and the number n_ξ of such intervals during the growing season. The frequency of up crossing is defined, in Porporato *et al.* (2001), as

$$\nu_\xi = \lambda' \int_{s_h}^{\xi} e^{-\gamma(\xi-u)} p(u) du = \rho(\xi) p(\xi) \tag{12}$$

The mean number of upcrossing during a growing season of length T_{seas} is readily obtained from the rate of occurrence, ν_ξ, as

$$\overline{n}_\xi = \nu_\xi T_{seas} = \rho(\xi) p(\xi) T_{seas} \tag{13}$$

The dynamic water stress index is a measure of vegetation stress obtained combining the mean intensity, duration and frequency of periods of water deficit. It is defined as

$$\overline{\theta} = \begin{cases} \left(\dfrac{\overline{\zeta}' \overline{T}_{s*}}{k T_{seas}} \right)^{1/\sqrt{\overline{n}_{s*}}} & if \quad \overline{\zeta}' \overline{T}_{s*} < k T_{seas} \\ 1 & otherwise \end{cases}, \tag{14}$$

where the mean time duration of the soil condition below the ξ is

$$\overline{T}_\xi = \frac{P(\xi)}{\rho(\xi) p(\xi)}. \tag{15}$$

In this case, we adopted the following parameters for the simulations: a parameter k equal 0·75, $q = 2$ and a threshold level $\xi = s^*$.

Different dynamic water stress values have been derived for the two models obtaining, as expected, differences that are strictly dependent on the climatic conditions. These differences can be observed in Figure 10 where the dynamic water stress of vegetation is described as a function of the soil texture and for several climatic conditions. The general pattern displays several differences among the two infiltration schemes. In particular, the dynamic water stress reaches generally higher values in the case of limited infiltration capacity and these differences are more marked under more humid conditions. It is interesting to remark that the minimum stress of vegetation is generally observed in loam textures, but when the limited infiltration capacity is taken into account these differences among soil texture collapse back to being minimal and apparently sandy-loam, loamy-sand,

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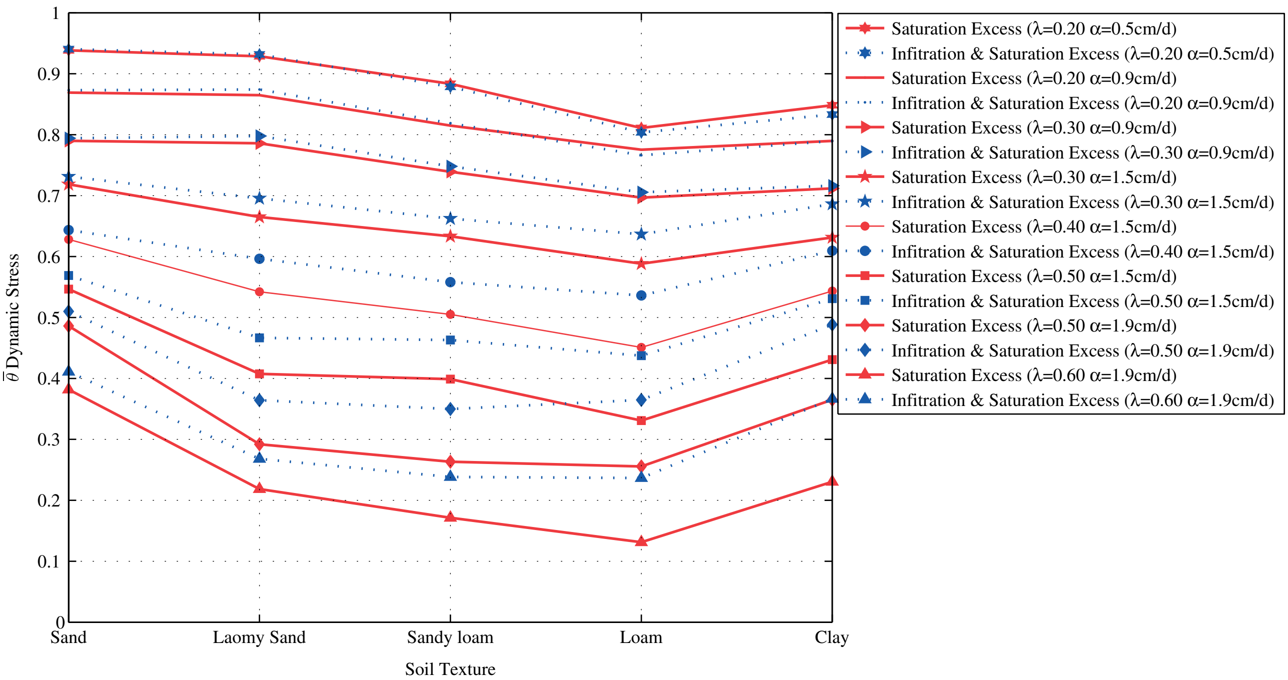

Figure 10. Dynamic water stress obtained using two infiltration schemes for different soil textures and climatic conditions with $Z_r = 30$ cm, $q = 2$, $k = 0{\cdot}75$, $T_{seas} = 160$ gg and $\xi = s^*$.

and loam behave in the same way with respect to the vegetation water stress.

As a general remark, we observe that the characterization of infiltration can have greater outcomes on the distribution/dynamics of the ecological state variable (in this case plant water stress) than it has on the hydrological state variable (soil moisture). This is likely because ecological processes are non-linear in response to soil moisture (i.e. q = 2 in the water stress formulation), so shifts in the soil moisture PDF are magnified into larger shifts in the water stress PDF. While we show this explicitly for plant water stress, the generality holds for other critical processes that are soil moisture dependent such as biogeochemical processes, which also vary nonlinearly with soil moisture (e.g. denitrifcation). This is an important 'novel' result, as it connects the characterization of hydrological processes to ecologically relevant outcomes.

CONCLUSIONS

The model proposed by Laio *et al.* (2001) provides a reliable representation of the soil moisture dynamics over a wide range of climatic conditions. The infiltration process is well-characterized especially in the case of highly permeability soil such as sandy soils. It is necessary to remark that for less permeable soils the hypothesis to neglect the surface control on the infiltration capacity may produce a significant overestimation of the mean and variance of the soil moisture especially in climates characterized by storms of high intensity and short durations. Significant differences in the variance of the soil moisture may produce change in the crossing properties of the soil moisture process also influencing the vegetation water stress with important implications for ecohydrological models. In fact, the vegetation water stress tends to be underestimated by the original model in humid climates. The presence of a limited infiltration capacity also shows that loam texture is more favorable for vegetation, while moving to more humid climates loamy-sand and sandy-loam may be suitable as well as loamy soils for the vegetation growth. Moreover, the characterization of hydrological dynamics has a greater impact on the ecological state variables, such as plant water stress, than the hydrological one, which has important consequences on modelers who seek to use hydrological models to explain an ecological phenomenon.

The results of this paper need to be interpreted also considering that the analyses presented refer to a soil water balance at the point scale where redistribution mechanisms have been neglected. This hypothesis implies that the results of these models can be applied to flat landscapes or to arid climates where soil moisture redistribution does not take place. The rainfall regimes in arid climates tend to be characterized by storm of short duration producing events of high intensity. In those cases, the soil moisture scheme developed by Laio *et al.* (2001) is consistent when dealing with more permeable soils, while more attention needs to be paid to infiltration excess in less permeable soils and in settings where the rainfall intensity is particularly high.

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