Parameterization of a bucket model for soil-vegetation-atmosphere modeling under seasonal climatic regimes

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Abstract. We investigate the potential impact of accounting for seasonal variations in the climatic forcing and using different methods to parameterize the soil water content at field capacity on the water balance components computed by a bucket model (BM). The single-layer BM of Guswa et al. (2002) is employed, whereas the Richards equation (RE) based Soil Water Atmosphere Plant (SWAP) model is used as a benchmark model. The results are analyzed for two differently-textured soils and for some synthetic runs under real-like seasonal weather conditions, using stochastically-generated daily rainfall data for a period of 100 years. Since transient soil-moisture dynamics and climatic seasonality play a key role in certain zones of the World, such as in Mediterranean land areas, a specific feature of this study is to test the prediction capability of the bucket model under a condition where seasonal variations in rainfall are not in phase with the variations in plant transpiration. Reference is made to a hydrologic year in which we have a rainy period (starting 1 November and lasting 151 days) where vegetation is basically assumed in a dormant stage, followed by a drier and rainless period with a vegetation regrowth phase. Better agreement between BM and RE-SWAP intercomparison results are obtained when BM is parameterized by a field capacity value determined through the drainage method proposed by Romano and Santini (2002). Depending on the vegetation regrowth or dormant seasons, rainfall variability within a season results in transpiration regimes and soil moisture fluctuations with distinctive features. During the vegetation regrowth season, transpiration exerts a key control on soil water budget with respect to rainfall. During the dormant season of vegetation, the precipitation regime becomes an important climate forcing. Simulations also highlight the occurrence of bimodality in the probability distribution of soil moisture during the season when plants are dormant, reflecting that soil, it being of coarser or finer texture, can be preferentially in either wetter or drier states over this period.

1 Introduction

Progress has been achieved in advancing scientific knowledge on the interactions within the soil-vegetation-atmosphere (SVA) system and also in developing improved monitoring and modeling technologies. However, there is still little information transferred from the research world to the tools used by public agencies, consultants, and stakeholders responsible for managing land and water resources. On the one hand, hydrologic models developed by scientists are usually very complex and require large amounts of specialized input data that may not be directly available (Höllander et al., 2009). On the other hand, simple models are easy to use, but may be of inadequate accuracy or detail to answer specific management questions, especially in cases of multiple demands upon the SVA system.

With specific reference to soil-water/vegetation interactions, the Richards equation (RE) offers a comprehensive, physics-based description of water movement in the vadose zone and associated hydrologic fluxes, including root water uptake (Lee and Abriola, 1999; Feddes et al., 2001). This model is a parabolic, partial differential equation whose parameters are the soil water retention, \( \theta(h) \), and hydraulic conductivity, \( K(h) \) [or, \( K(\theta) \)], functions, the so-called soil hydraulic properties, that are relationships between matric suction head, \( h \), soil water content, \( \theta \), and hydraulic conductivity, \( K \) (Kutflek and Nielsen, 1994). Solutions of the Richards equation are difficult because of the strong nonlinearity of the functions \( \theta(h) \) and \( K(h) \). One can achieve RE solutions in analytical or semi-analytical terms only for particular types of soils (e.g. linear soils) and specific boundary conditions, instead one has to resort to numerical methods for solving...
accurately most of practical soil-water flow problems (Romano et al., 1998; Šimůnek et al., 2008). Employing RE at laboratory soil core or field plot scales can be done with confidence (Sposito, 1986), but may become unmanageable for hydrologic applications at the scales of hillslopes and catchments chiefly because of the spatial, and sometimes also temporal, variability exhibited by the soil hydraulic properties (Sharma and Luxmoore, 1979; Braud, 1998; Settin et al., 2007), and partly because of theoretical doubts that capillary suction heads and hydraulic gradients are still the driving forces of mass flow and momentum transfer in unsaturated porous media at such larger scales (Beven, 1989; McDonnell et al., 2007). Notwithstanding these limitations, RE is increasingly being used in process-based distributed models of hydrological cycle and is also used at the global scale in Land Surface Models (Kowalczyk et al., 2006; Clark et al., 2008; Manus et al., 2009; Oleson et al., 2010). Moreover, RE solutions are anyhow considered as the best available reference for comparisons to the performances of simplified models to be categorized in terms of their complexity with respect to the number of hydrologic processes employed (Guswa et al., 2002).

The soil water balance bucket model (BM) of Manabe (1969) is the classic example of a lumped model accounting for a simplified description of the major processes evolving in the hydrologic cycle, in which a single soil layer is conceptualized as a bucket receiving and retaining all incident water until its storage capacity is filled. The latter assumption, for example, saves one the need of describing infiltration rate with respect to rainfall intensity through an analytical (usually, empirical) equation. This type of model is particularly efficient to describe land surface processes and thus extensively used when coupled with large scale atmospheric models, including General Circulation Models (GCMs), but may experience problems due to its simple representation of vegetation and estimation of potential evapotranspiration fluxes (Seneviratne et al., 2010). Since the valuable function of soil to transfer water with different rates along the vertical profile is virtually ignored and a direct feedback is described between soil water storage and evapotranspiration, BM is characterized by a rather short soil water storage memory and can hardly describe correctly the system response to rapid time evolution in atmospheric forcing. Yet, in the last decades several studies have shown that bucket models can be efficiently employed for predicting some components of the soil water budget, as they provide results similar to those observed in the field or those simulated with models which are more detailed in the description of the soil water dynamics, at least in soils with specific climatic regimes (Milly, 1994; Guswa et al., 2002; Farmer et al., 2003).

On the wave of studies gaining insights into the potential impacts that land use and climate changes may exert on catchment scale hydrologic response, one challenging question is to obtain efficient predictions while recognizing that water transfer in the SVA continuum is controlled by several non-linear and dynamic interacting processes (Blöschl and Montanari, 2010). The efficiency of simplified models has been tested in areas characterized by stationary condition of the climatic forcing (Milly et al., 2008), which can be observed only in limited zones of the World, such as in the savannas of South Africa (Porporato et al., 2002). Instead, transient soil-moisture dynamics and climatic seasonality play a key role in other environments (Settin et al., 2007), such as those characterized by Mediterranean climates, where rainfall shows an out-of-phase relationship with air temperature, with an alternation of wet-cold and dry-warm periods. The end of wet-cold period corresponds to the beginning of the growing season, when water dynamics are mainly controlled by the evapotranspiration processes, which induce a progressive reduction of the water stored toward the dry-warm season. The soil-moisture dynamics during the growing season is generally far from what can be predicted with the hypothesis of stationarity. Nevertheless, under such complex climatic variability during the year, linking the fluxes and vegetation growth to the variations in soil water contents along the entire soil profile can be a prerequisite to adequately simulate fluctuations in evapotranspiration or other hydrologic fluxes.

A general objective of the present work is to assess the validity of using a bucket model to simulate the local water balance and associated fluxes assuming a rainfall regime and vegetation characteristics representative of Mediterranean areas. Comparisons are carried out with respect to the results obtained under the same conditions through a RE-based hydrologic model. Under climate, vegetation, and soil conditions relevant to an African savannah site, Guswa et al. (2002) have shown that an appropriate formulation of the single-layer bucket model, with proper account of the relationship between transpiration, \( T \), and degree of saturation in soil, \( s \), can provide results similar to those obtained by a more comprehensive model based on the integration of the vertically discretized Richards equation. At relatively larger time-scale, namely an entire growing season of 200 days, Table 5 of this paper reports some more similarity among the water fluxes as computed by the RE or BM models. However, when moving to a smaller time-scale, namely the daily time-scale, discrepancies between RE model results and those of the BM of Guswa et al. (2002) were about 50% for evapotranspiration flux, \( ET \), and ranged from 50% to 70% over a season for the \( T/ET \) ratio (see, for example, the traces of daily transpiration depicted in Fig. 8 of that paper).

The soil water content at the condition of field capacity, \( \theta_{fc} \) (field capacity, for short), is a key parameter of a bucket-type approach as it represents a threshold water content that controls the processes of surface runoff generation and drainage (leakage) out of the root zone (more precisely, out of the lower boundary of the flow domain). The classic bucket approach is thus unable to manage an infiltration excess runoff (the Hortonian mechanism of runoff generation) and subsurface flow, and handles only a saturation excess runoff (Dunne mechanism). When soil water storage in the bucket exceeds
field capacity, then this excess of water is routed to runoff only, or to both runoff and drainage. Notwithstanding the considerable importance of field capacity in a bucket-type hydrologic approach, its value is still determined using simplified, often very inaccurate techniques (Romano and Santini, 2002). In view of parameterizing a bucket model for effectively computing the soil water budget under environmental conditions representative of a Mediterranean area, a specific objective of this study is to evaluate whether the determination of the field capacity value with an appropriate technique can lead to BM results more in agreement with those offered by a more complex model based on RE.

2 Description of the models

2.1 The SWAP model

SWAP (Soil, Water, Atmosphere and Plant) is a widespread used model in the soil hydrology community to simulate soil water flow in the vadose zone at field scale and during entire growing seasons (van Dam et al., 2008). The vertical soil-water movement is described by the Richards equation with a sink term accounting for the root water uptake, as follows:

$$C(h)\left(\frac{\partial h(z, t)}{\partial t}\right) = \frac{\partial}{\partial z}\left[K(h)\left(\frac{\partial h(z, t)}{\partial z} - 1\right)\right] - S(h)$$

where

$$S(h) = (\alpha(h) \cdot T_{\text{max}})$$

which is written here in its pressure-head form. In Eq. (1), time, t, and vertical coordinate, z (taken positive downward), are the independent variables, whereas suction pressure-head, h, is the dependent variable. Model parameters are the water capacity function, C(h) = dθ/dh, which can be readily obtained from knowledge of the soil water retention function, θ(h), and the hydraulic conductivity function, K[θ(h)]. After having specified the appropriate initial and boundary conditions, the SWAP model solves Eq. (1) numerically using a finite-difference approach.

The following van-Genuchten/Mualem analytical relationships are used as soil hydraulic properties:

$$\theta_s(h) = \frac{\theta_s - \theta_r}{\theta_s - \theta_t} = \left[\frac{1}{1 + (\alpha_{\text{VG}} \cdot h)^n_{\text{VG}}}\right]^{m_{\text{VG}}}$$

$$K(\theta) = K_0 \theta_s^{\tau_{\text{VG}}} \left[1 - \left(1 - \theta_s^{1/m}\right)^{m_{\text{VG}} - 2}\right]$$

where \(\theta_s\) is the volumetric water content at \(h = 0\) (commonly the saturated water content), \(\theta_t\), \(\theta_r\) is the residual water content, namely the water content \(\theta\) for \(h \to \infty\), \(K_0\) is the hydraulic conductivity when \(h = 0\), \(\alpha_{\text{VG}}\) is a scale parameter, whereas \(n_{\text{VG}}, m_{\text{VG}} = 1 - 1/n_{\text{VG}}\), and \(\tau_{\text{VG}}\) are shape parameters. In principle, parameter \(\theta_s\) should be viewed as distinct from the saturated water content, \(\theta_s\), mainly because of possible air entrapment in the soil pores. The term \(s_e(0 \leq s_e \leq 1)\) is the effective degree of soil saturation (simply, effective saturation).

In SWAP, the sink term, \(S(h)\), describes the extraction of water by the root architecture at the point macroscopic scale as a function of suction pressure head, \(h(z, t)\). Specifically, root water uptake rate at a given soil depth, \(z\), is determined by the normalized root density distribution at this depth, \(\beta(z, t)\), and the maximum transpiration rate, \(T_{\text{max}}\), whereas reductions due to water stresses are described by the uptake reduction function, \(\alpha(h)\) (0 ≤ \(\alpha\) ≤ 1), such that

$$S(h) = \beta(z, t) \cdot \alpha(h) \cdot T_{\text{max}}$$

(3a)

Kroes and van Dam (2003). Under conditions of no stress, the sink term represents the maximum uptake rate as follows:

$$S_p = \beta(z, t) \cdot T_{\text{max}}$$

(3b)

For computing evapotranspiration fluxes, SWAP follows a macroscopic approach and uses the Penman-Monteith equation with daily weather data such as air temperature, solar radiation, wind speed and humidity; a minimum value for canopy and aerodynamic resistance is applied. Maximum transpiration rates are computed taking also the evaporation rate of the water intercepted by the vegetation into account. Actual evapotranspiration rates are computed using a reduction factor based on root water uptake and a reduction factor due to maximum soil evaporation flux according to Darcy’s law. The actual transpiration rate is then calculated on the basis of the actual soil water extraction rate over the whole rooting depth. Further details on this module of the model can be found in Kroes and van Dam (2003) and van Dam et al. (2008). Of specific interest for this study is the semi-empirical way with which the uptake reduction function, \(\alpha(h)\), is modeled by SWAP (see Fig. 1). Not accounting for salinity problems, this reduction function can be basically split into three different parts of the diagram (see Eq. 3): Part-I is for \(h_1 \leq h < h_2\) and represents air deficiency; Part-II is for \(h_2 \leq h \leq h_3\) and accounts for a conditions of no water stress; finally, Part-III is for \(h_3 < h \leq h_4\) and reveals water stress conditions. Matric suction head \(h_4\) is the well-known permanent wilting point of the vegetation, \(h_w\).
transpiration, $T$, and leakage, $L$, rates are considered as function of average soil saturation, $s$, only. Losses due to surface runoff are generated only when the bucket is completely full, i.e. at the condition of full saturation in the soil. Only a fraction of the incoming precipitation is able to infiltrate into the soil when the rainfall depth exceeds the storage capacity of the soil profile. Therefore, the relationship $I[s(t), t]=\min[r, nZ_s(1-s)]$ accounts for the dependence of infiltration rate, $I$, on rainfall depth, $r$, and the degree of soil saturation, $s$, in the sense that rainfall infiltration is equal either to rainfall depth or to soil storage capacity, whichever is less. Because of a stochastic representation of precipitation, $P(t)$, Eq. (4) is a stochastic linear ordinary differential equation and should be viewed as a probabilistic model of soil water budget.

Evapotranspiration is a fundamental hydrologic process and most bucket models treat only implicitly the effect that vegetation canopy exerts on evapotranspiration. An interesting feature of the bucketing approach proposed by Guswa et al. (2002) is the separation of soil surface evaporation from transpiration by plants. Transpiration, $T(s)$, and evaporation, $E(s)$, as a function of average soil saturation, $s$, are computed through the following relations:

$$T(s) = \begin{cases} 0 & s \leq s_w \\ \frac{s-s_w}{s^*-s_w} T_{\text{max}}, & s_w < s < s^* \\ T_{\text{max}}, & s \geq s^* \end{cases}$$

$$E(s) = \begin{cases} 0 & s \leq s_h \\ \frac{s-s_h}{s^*-s_h} E_{\text{max}}, & s_h < s < s^* \\ E_{\text{max}}, & s \geq s^* \end{cases}$$

where $s_w$ is the saturation at wilting condition, and $s^*$ is saturation at incipient stomata closure; the term $s_h$ is the hygroscopic saturation, namely the average degree of saturation in soil when soil suction at the soil-atmosphere interface is extremely low (this suction head is often set at a value ranging from 150 \times 10^3 \text{ cm} to 500 \times 10^3 \text{ cm}) and evaporation ceases.

Depletion of soil water due to the uptake by the plant root system is assumed to be governed by the two major mechanisms of atmospheric demand and supply of water in the soil. The model does not account for plant characteristics explicitly. Within this framework, a basic soil parameter is the saturation $s^*$ when soil water starts becoming a limiting factor and plants are no longer able to transpire at the full rate $ET_{\text{max}}$, as the sum of maximum daily evaporation, $E_{\text{max}}$, and maximum daily transpiration, $T_{\text{max}}$ (see Fig. 2 and Eqs. 5–6). When soil water supply is insufficient to meet the atmospheric demand, namely when $s$ is less than $s^*$, the progressive leaf stomatal closure yields a reduction in root water uptake and actual transpiration, $T$, start decreasing from $T_{\text{max}}$ to reach the zero value at wilting point, $s_w$, following a linear pattern. Actual evaporation may not equal the evaporation rate under well-watered conditions, $E_{\text{max}}$, and also decreases linearly from $E_{\text{max}}$, at $s = s^*$, to zero when $s = s_h$. The difference between the two threshold values of $s_w$ and $s_h$ is the definition of evapotranspiration rate under well-watered conditions, $E_{\text{max}}$.

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**Fig. 2.** Loss functions $E(s)$, $T(s)$, and $L(s)$ as average degree of soil saturation, $s$, changes in the bucket model due to evaporation, transpiration, and leakage (redrawn from Guswa et al., 2002).
$s_h$ accounts for the fact that evaporation may proceed even after transpiration has stopped at the wilting point of vegetation (Laio et al., 2001). Therefore, for $s$ less than $s_w$, transpiration ceases and evaporation from soil surface remains the sole active component of the evapotranspiration process. Other features of the BM can be found in the paper by Guswa et al. (2002).

3 Parameterization strategy for evapotranspiration and field capacity

3.1 Homogenizing BM and SWAP evapotranspiration functions

The bucket and SWAP models handle the calculation of evapotranspiration rates in different manner and a sort of equivalence between them should be set in order to make comparisons between model outputs meaningful and more effective.

A first assumption is that the soil is almost completely covered by vegetative canopies so that soil evaporation can be neglected. From a parametric perspective, this is equivalent to assume $s_h = s_w$ in the BM, while for SWAP soil cover fraction, $F_{sc}$, is set at 1.0 and crop factor, $\kappa_c$, is also set at 1.0. We basically consider that transpiration water losses follow two stages, the constant and falling rate stages, but do not distinguish between low or high transpiration demands. In SWAP, the falling-rate phase of uptake, and hence of transpiration, follows a nonlinear, concave-shaped depletion from $T_{\text{max}}$ to zero (de Jong van Lier et al., 2009). In BM, transpiration as a function of average soil saturation in the root zone is modeled by piecewise linear segments as defined by Eq. (5) and shown in Fig. 2. Therefore, to make evapotranspiration losses of the two models comparable, we impose: $\beta(z,t) = 1$ (i.e. uniform root density distribution within the root zone), $s_w = s(h_4)$ (i.e. equivalence of the wilting points) and $h_1 = h_2 \approx 0$ (i.e. the effect of the air deficiency negligible). We also set $s^*$ such as:

$$\int_{s_w}^{s^*} T(s) \cdot ds + T_{\text{max}} \int_{s_w}^{1} \alpha [h(s)] \cdot ds = T_{\text{max}}$$

(7)

where $T(s)$ is defined by Eq. (5), while $\alpha(h)$ is defined by Eq. (3). Equation (7) is equivalent to impose the same average transpiration rate in the soil moisture range from the wilting point to the full saturation condition. Equation (7) is verified if $s^*$ is defined as follows:

$$s^* = 2 \left[ 1 - \int_{s_w}^{1} \alpha [h(s)] \cdot ds \right] - s_w$$

(8)

Figure 3 shows examples of BM and SWAP transpiration functions after setting parameter $s^*$ according to Eq. (8).

3.2 Techniques for determining field capacity

Field capacity is defined as the volumetric water content remaining in a uniform soil profile two or three days after having been completely wetted with water and after free drainage beyond the root zone has become negligible (Soil Sci. Soc. Am., 2008). The field capacity concept was originally introduced for irrigation scheduling purposes under a simplistic view of soil water movement in the SVA system (Veihmeyer and Hendrickson, 1927). It was soon used also by hydrologists who defined it as the maximum quantity of water that can be permanently retained in the soil against the pull of gravity. In catchment hydrology, field capacity is thus considered as a critical threshold parameter of the water-holding capacity of the soil that affects both runoff generation mechanism and evapotranspiration process. As pointed out by Farmer et al. (2003), the field capacity concept is strictly applicable to the unsaturated zone only and therefore the amount of water held at field capacity will vary as the saturated portion of the soil profile increases or decreases.

Apart from some subjectivity related to the definitions of field capacity, further elements of uncertainty are introduced by the commonly used practice to ascribe this parameter to a specific point of the soil water retention characteristic. Field capacity, $\theta_{fc}$, is often estimated on a soil sample in the laboratory as the soil water content at the suction pressure of 1/3 bar (about 3.3 m of suction head), regardless of the specific physical (especially soil texture) and chemical properties of the soil sample. If one would take textural characteristics into account, it is suggested to set the suction head approximately at 1.0 m for coarser sandy soils and at 5.0 m for finer clayey soils, whereas the suction head of 3.0 m (a kind of average between the previous two suction heads) is mostly referred to medium-textured loamy soils. On the other hand, allowing for the definition of field capacity, Meyer and Gee (1999)
suggested $\theta_c$ should be related to the hydraulic conductivity function of a soil and determined from the $K(\theta)$-curve as the water content when $K$ takes on values ranging from $10^{-6}$ cm s$^{-1}$ to $10^{-8}$ cm s$^{-1}$. Some authors have proposed relating the negligible drainage flux at field capacity, $q_{fc}$, to the daily evapotranspiration for the area of interest (Kutílek and Nielsen, 1994), whereas Laio et al. (2001) defined the field capacity point when drainage losses ($K_s$ in practice, under the unit hydraulic gradient assumption) are 10% of the maximum daily evapotranspiration losses, $ET_{max}$.

Whatever the pre-fixed points of the $\theta(h)$ or $K(\theta)$ functions, soil water content at field capacity is certainly not an inherent soil property, but rather is a process-dependent parameter that may represent only a rough attempt to summarize a dynamic flow process through a single global static coefficient (Romano and Santini, 2002). The redistribution process in soil, and hence the field capacity value, depends on the water content distribution established in the entire soil profile at the end of the preceding infiltration phase, as well as on the boundary conditions of the flow domain. In case of coarser-textured soils with stable aggregates, the field capacity concept is underpinned by experimental evidence since for these types of soils the drainage process is relatively very fast initially, but then its rates decrease drastically due to the abrupt reduction in unsaturated hydraulic conductivities as suction pressure head increases. Therefore, Romano and Santini (2002) argued that a field drainage experiment is the test to be preferred for determining the water content value at field capacity, namely the condition when drainage fluxes become virtually nil during a redistribution process started from an initially saturated soil profile and with no flux from the soil surface. Using the HYDRUS-1D software to simulate field drainage experiments for real soils retrieved from different soil databases, Twarakavi et al. (2009) confirmed the inaccuracy of determining field capacity from pre-fixed points of the water retention curve. They also suggested that the value of 0.01 cm day$^{-1}$ can be considered as a negligible drainage flux being imposed at the lower limit of the rooting zone to meet the condition of field capacity for a wide range of soils. Moreover, these authors developed an empirical analytical relationship to estimate $\theta_c$ from information about the soil hydraulic properties.

Especially for practical applications, when judging on the effectiveness of a method with respect to another (for example, a simplified one) it is important not only to evaluate the discrepancies among parameter values obtained from different methods (parametric evaluation), but also to analyze the outputs of a hydrologic model computed by using different parameterization techniques (functional evaluation). Under soil, vegetation, and climate conditions representative of a Mediterranean area, a focus here is to compare, with respect to the reference model output, the soil-water budget predictions of BM when field capacity is estimated by using either a standard pre-fixed point of the $\theta(h)$ curve or a field transient drainage experiment.

Fig. 4. Scatter plots of field capacity, $\theta_{fc\_drain}$, as determined by the drain-method versus field capacity values obtained from the water retention characteristics by the fix-method, $\theta_{fc\_fix}$, for the three suction heads of 100 cm, 330 cm, and 500 cm. Data points refer to 1087 soil samples grouped according to different textural classes.

4 Soil, vegetation, and climatic characteristics

Analyses were carried out in this study using various soil properties retrieved from the UNSODA database and
Fig. 5. Ternary diagram of the soil texture triangle showing markers whose color depends on the magnitude of variable \( \Delta \text{PAWS}_{\text{max}} \) being the difference between the maximum plant-available soil-water holding capacity when field capacity is alternatively determined using the \( \text{drain} \) or \( \text{fix} \)-method.

Fig. 6. Boxplots of variable \( \Delta \text{PAWS}_{\text{max}} \) for the soils of the fine, medium, and coarse textural classes according to the FAO-HYPRES.

gathered from different zones of the World (Nemes et al., 2001) as well as stored in the database of the Soil Hydrology Laboratory of University of Naples Federico II and mainly collected in various parts of Southern Italy (Ceres et al., 2010). On the whole, we have processed data from 1087 soils. Synthetic transient drainage experiments (referred to as \( \text{drain} \)-method) have been performed to determine the average soil water content, \( \theta_{\text{fc,drain}} \), in the soil profile when the drainage rate is \( q_{\text{drain}} = 0.010 \) cm day\(^{-1} \) at the soil depth \( z_t = 1.0 \) m and imposing a zero flux boundary condition at the soil surface (Romano and Santini, 2002). For the same soils, field capacity (\( \theta_{\text{fc,fix}} \)) was also calculated from the respective water retention functions (referred to as \( \text{fix} \)-method) as the water content at the fixed suction heads of 100 cm, 330 cm, and 500 cm (\( \theta_{\text{fc,fix}} - 100 \), \( \theta_{\text{fc,fix}} - 330 \), and \( \theta_{\text{fc,fix}} - 500 \)).

Using the entire available dataset and for different soil textural classes, the scatter plots of Fig. 4 provide a qualitative assessment of correlations between variables \( \theta_{\text{fc,drain}} \) (horizontal axis) and the three different variables \( \theta_{\text{fc,fix}} \) (vertical axes). A 1:1 line is also drawn in each plot to make it easier detecting possible better agreements among the variables. The scatter plots arranged along the principal diagonal of this kind of matrix diagram (i.e. from the top-left corner to the bottom-right corner of Fig. 4) confirm the higher degree of correlation of \( \theta_{\text{fc,drain}} \) with \( \theta_{\text{fc,fix}} - 100 \) for coarser soils, with \( \theta_{\text{fc,fix}} - 330 \) for medium-textured soils, and with \( \theta_{\text{fc,fix}} - 500 \) for finer soils, respectively (Ceres, 2009). Let \( \text{PAWS}_{\text{max}} = (s_{\text{fc}} - s_w) \times n \times Z_t \) be the maximum plant-available soil-water holding capacity. Then, Fig. 5 illustrates a ternary diagram with relative intensity of each colored dots depending on the magnitude of the difference \( \Delta \text{PAWS}_{\text{max}} = \text{PAWS}_{\text{max,drain}} - \text{PAWS}_{\text{max,fix}} \) when using the \( \text{drain} \) and \( s_{\text{fc,fix}} \) values to compute variable \( \text{PAWS}_{\text{max}} \). This diagram provides a more quantitative evaluation of the differences among the soils employed in this study and helps in finding suitable representative soil samples for the subsequent analyses being carried out in this study. To further proceed in selecting representative soils, we have also grouped all of the 1087 soils according to the classes of the FAO-HYPRES soil textural triangle (Nemes et al., 1999) and computed the boxplots of variable \( \Delta \text{PAWS}_{\text{max}} \) for the soils pertaining to the fine, medium, and coarse textural classes (see Fig. 6). The boxplot for the coarse soil class shows a median approximately of \(-4.3 \) cm, whereas that one for the fine soil class shows a median approximately of \(+1.8 \) cm. With a view to the general aim of this study and based on the results depicted in Figs. 4 to 6, we have considered the soils of the Mediterranean area of Southern Italy and selected a loamy-sand soil as representative of coarse soils and a clay soil as representative of fine soils. The major characteristics of these two soil types are presented in the following (see also Table 1).

The physical properties of the loamy-sand soil are as follows: bulk density of 1.198 g cm\(^{-3} \), and USDA percentages of sand, silt, and clay content equal to 73.78 %, 23.27 %, and 2.95 %, respectively. For the clay soil, the physical properties are as follows: bulk density of 1.348 g cm\(^{-3} \), and USDA percentages of sand, silt, and clay content equal to 21.20, 31.80, and 47.00, respectively. For these two differently textured soils, undisturbed soil cores were subjected to evaporation experiments to obtain the respective parameters of the soil water retention, \( \theta(h) \), and hydraulic conductivity, \( K(s_e) \), through an optimization technique (Romano and Santini, 1999). The optimized hydraulic parameter values of Eqs. (2) are reported in Table 1.
Results from the simulation runs are as follows: $s_{fc, \text{drain, LS}} = 0.510$ for the loamy-sand soil, and $s_{fc, \text{drain, Cl}} = 0.830$ for the clay soil. Instead, adopting the simplified method (referred to as fix-method) of estimating the degree of soil saturation at the field capacity from pre-fixed points of the soil water retention function (see Table 1 for the water retention parameters), we obtained the following values: $s_{fc, \text{fix, LS}} = 0.670$ for the loamy-sand soil at the suction head of 100 cm, and $s_{fc, \text{fix, Cl}} = 0.790$ for the clay soil at the suction head of 500 cm. The choice of these two suction heads for computing $s_{fc, \text{fix}}$ for the coarser soil and the finer soil stems from the results we have presented in Fig. 4.

For both hydrologic models, the control volume is an active soil profile with a depth of 1.00 m. Therefore, for the bucket model the depth of rooting zone, $Z_r$, is 1.00 m. The porosity values are $n_{LS} = 0.447$ for the loamy-sand soil and $n_{Cl} = 0.426$ for the clay soil. The product $nZ_r$ represents the so-called active soil depth, which is the volume per unit surface area available for water storage.

For the SW AP model, the lower limit of the root zone is set at a depth of 1.00 m and the entire soil profile was split in three portions accounting for the numerical discretization of the flow domain: the first uppermost portion of 0.10 m in thickness comprises 10 compartments; the second interposed portion of 0.40 m in thickness comprises 8 compartments, and the third lowermost portion of 0.50 m in thickness comprises 5 compartments. The hydraulic properties of the two different uniform soils are identified by the parameters of Table 1.

Precipitation and transpiration from vegetation are usually out of phase in Mediterranean-type climates being characterized by warm and dry summers, during which vegetation is often under stress conditions, followed by colder and wet winters where precipitations are predominant and air temperatures are relatively mild. With reference to a hydrologic year starting on 1 November, we assumed vegetation dormancy (Dormant Vegetation Phase, DVP) during a wetter rainy period lasting 151 days (from 1 November to 31 March), and the subsequent vegetation regrowth (Regrowth Vegetation Phase, RVP) during a drier and rainless period lasting 214 days (from 1 April to 31 October). During the wetter season, rainfall volume exceeds evapotranspiration losses, thus soil “bucket” tends to be filled close to the field capacity. During the drier season, in spring vegetation starts leafing out and exhausting soil water storage by transpiration, and when summer is coming transpiration is regulated by the leaf stomatal closures as atmospheric demand increases more and more.

Viola et al. (2008) suggested that in a Mediterranean ecosystem soil moisture dynamics can be effectively investigated by representing the climatic forcing during the year as a sequence of a wet and a dry period, each characterized by a stationary rainfall regime. We considered a Mediterranean woodland ecosystem comprising the following deciduous species: *Quercus pubescens* Willd., *Acer campestris* L., and *Fraxinus ornus* L. For this woody area, values of maximum evapotranspiration, $ET_{\text{max}}$, are set constant during each season: at 0.46 cm day$^{-1}$ during the RVP season, and 0.20 cm day$^{-1}$ during the DVP season (Pumo et al., 2008).

To evaluate the impact of the seasonal evolution of rainfall intensity on soil moisture dynamics, the occurrence of precipitation in time, $P(t)$, has been probabilistically described using a Poisson Rectangular Pulse (PRP) model on a daily time scale with probability of occurrence equal to $\lambda(t) dt$ ($dt = 1$ day). The Poisson process of arrival rate $\lambda$
Table 3. SWAP hydrologic parameters.

| Soil texture      | Vegetative phase | $h_1 = h_2$ (cm) | $h_3$ (cm) | $h_4 = h_w$ (cm) | $Z_r$ (cm) | $\text{ET}_{\text{max}}$ (cm day$^{-1}$) | $F_{\text{sc}}$ (–) | $\kappa_c$ (–) |
|-------------------|-------------------|------------------|------------|------------------|------------|-----------------------------------------|------------------|--------------|
| Loamy-sand, LS    | RVP               | 1.0              | 300        | 16 000           | 100        | 0.46                                    | 1.0              | 1.0          |
|                   | DVP               | 1.0              | 300        | 16 000           | 100        | 0.20                                    | 1.0              | 1.0          |
| Clay, CI          | RVP               | 1.0              | 1500       | 16 000           | 100        | 0.46                                    | 1.0              | 1.0          |
|                   | DVP               | 1.0              | 1500       | 16 000           | 100        | 0.20                                    | 1.0              | 1.0          |

Table 4. Relative soil moisture at field capacity, $s_{fc}$, for BM and soil-water storage capacity, $W_{\text{ini}}$, in the system at the start of the simulation run.

| Soil texture      | Method             | $s_{fc}$ (–) | $W_{\text{ini}}$ (cm) |
|-------------------|--------------------|--------------|------------------------|
| Loamy-sand, LS    | drain (drainage exp.) | 0.510        | 22.80                  |
|                   | fix (from Eq. 2a at $h = 100$ cm) | 0.670        | 29.95                  |
| Clay, CI          | drain (drainage exp.) | 0.830        | 35.36                  |
|                   | fix (from Eq. 2a at $h = 500$ cm) | 0.790        | 33.65                  |

(in number of storms per day) does not account for the temporal structure within each rainfall event, whereas the distribution of storm depth is exponential with a mean depth of $\eta$ (in cm per storm). Following Pumo et al. (2008), the selected PRP parameter values are: $\lambda_{\text{RVP}} = 0.195$ day$^{-1}$ and $\eta_{\text{RVP}} = 0.595$ cm for the RVP period; $\lambda_{\text{DVP}} = 0.493$ day$^{-1}$ and $\eta_{\text{DVP}} = 0.701$ cm for the DVP period.

4.1 Overview of input parameters for BM and SWAP – initial and boundary conditions

To summarize the parameterization strategy employed for the bucket model and the Richards equation based SWAP model discussed in the previous sections, Tables 2, 3, and 4 report the various parameter values selected for the two different models (Tables 2 and 3, respectively) and for the two differently textured soils (the loamy-sand soil and the clay soil, respectively).

In view of one specific objective of the present study, Table 4 reports the values of relative soil moisture at the condition of field capacity for the loamy-sand and the clay soils when adopting the drain-method or the fix-method (Romano and Santini, 2002).

Simulation runs have been carried out using one time-series of synthetic daily rainfall records generated stochastically for a 100-year-long period, obtained by combining two different Poisson Rectangular Pulse (PRP) models, for the regrowth (RVP) and dormant vegetation (DVP) phases, respectively. Table 2 reports the PRP parameters for the RVP and DVP periods whereas, as an example, Fig. 7 shows for a generic year the rainfall sequences generated with the PRP model for these two seasons of the hydrologic year.

In order for the SWAP lower boundary condition to be irrelevant or not to affect too much the simulated soil-water balance in the first 1.0 m of rooting depth, the lower end of the flow domain is set at a depth of $z_{\text{bot}} = 2.0$ m from soil surface. The lower boundary condition is specified as the Neumann condition of free drainage, namely the unit gradient of the total hydraulic head [i.e. $\nabla (h - z) = -1$] applied at $z_{\text{bot}}$. Field capacity can be a representative value for the soil water content at the end of the wet season and, for both models, in the first year simulations start when the systems are at the condition of field capacity in soil at the beginning of what we assumed being the growing season (1 April). Therefore, the starting condition is the degree of soil moisture $s_{fc}$ for the BM model, and the suction head $h_{fc}$, corresponding to $\theta_{fc}$, for the SWAP model. In particular, for the BM model the initial soil-water storage capacity, $W_{\text{ini}} = s_{fc} \times n \times Z_r$. 

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represents a relatively wet condition for the soil-vegetation system and the relevant values are reported in Table 4 for the two soil types and the two different techniques to determine soil moisture at field capacity. To lessen or remove the effects of the initial conditions on the intercomparison results, the first year of simulation is taken as a sort of spin-up time and removed from the subsequent analyses. Therefore, the actual initial condition is the state of the system at the end of this first year of preliminary simulation. Apart from other forcing variables, a spin-up time for soil moisture varies with the thickness of soil profile: the deeper the soil depth is, the longer the soil moisture reaches its equilibrium state. By running several simulations with different initial conditions, we observed that the results were independent from the initial condition after the first year.

4.2 Performance measures

We have selected the mean error (ME) and the root-mean-square error (RMSE) as performance measures, which are computed using the following equations:

\[
ME = \frac{1}{N} \sum_{i=1}^{N} \text{err}_i
\]

\[
RMSE = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (\text{err}_i)^2}
\]

where \(N\) is the total number of data and \(\text{err}_i = (v_{SWAP}^i - v_{BM}^i)\) represents the deviation between the generic variable, \(v_{SWAP}^i\), computed by the RE-based SWAP model (assumed as a reference) and the corresponding variable, \(v_{BM}^i\), computed by the BM model. The ME statistic reveals the presence of biases (a positive value means that on average the bucket model underestimates) and is a measure of accuracy. The RMSE statistic is a commonly used measure of precision. Both ME and RMSE are dimensional indices and their best values are zero. A feature of RMSE statistic is that it tends to emphasize larger values in a series, whereas lower values are virtually neglected.

5 Results of the numerical experiments and discussion

Using the parameterization strategy outlined in Sect. 3, for the loamy-sand and the clay soils described in Sect. 2.3 we have compared the BM’s hydrologic responses to those of the RE-based SWAP model when in BM the degree of soil saturation at field capacity, \(s_{fc}\), is either determined from a field drainage experiment (\(\text{drain-method}\)) or estimated as a prefixed point of the soil water retention characteristic (\(\text{fix-method}\)).

Simulations refer to a well vegetated landscape, and therefore transpiration is assumed to be dominant to evaporation within the evapotranspiration processes (i.e. \(E_{max} \approx 0\)). A preliminary check was carried out to ensure that the generated daily rainfall rate, \(r\), never exceeds the saturated hydraulic conductivity, \(K_s\), so as to avoid that the occurrence of surface runoff might have led to difficulties when comparing the BM and RE-based SWAP performances, and hence to have better similarity among the responses offered by these two hydrologic models.

The selected performance indices are reported in Table 5 and show relatively good agreements between the BM and RE outputs, with ME and RMSE absolute values ranging around an order of some percent unit. The worst accuracy (ME = 3.1 %) and precision (RMSE = 12.0 %) for the bucket model occur when field capacity of the loamy-sand soil is determined by the fix-method. Both these largest ME and RMSE values refer to the simulation of leaching rates evolving during the dormant vegetation period and are indication of certain underestimations and wider spread of the loss fluxes computed by BM. Based on the ME index, parameterizing \(s_{fc}\) of the loamy-sand soil with the \(\text{drain-method}\) (see Table 5) leads almost always to the least biased responses. This occurs not only for the entire hydrologic year, but also for both RVP and DVP periods. The type of biases with respect to the reference SWAP model depends on the type of variable considered. BM systematically overestimates transpiration rate and underestimates leaching rate. In the case of soil moisture the bias behavior is different since there is an underestimation for the \(\text{drain-method}\), but an overestimation when the \(\text{fix-method}\) is used. Only for the DVP period of soil moisture variable, the bias in absolute terms is more favorable when the \(\text{fix-method}\) is used to estimate field capacity. In terms of RMSE (see Table 5), the determination of \(s_{fc}\) with the \(\text{drain-method}\) increases the precision of BM responses significantly over the entire hydrologic year: for the loamy-sand soil, there are reductions in RMSE values of about 57.1 % for soil moisture, 33.3 % for transpiration rate, and 44.6 % for draining rate. When splitting the hydrologic year into the two different periods, the reductions in RMSE values are still evident over the RVP period for all of the three variables, but for the DVP period the advantage of the \(\text{drain-method}\) is much less evident for transpiration rate.

These comparisons among the index values help in understanding under what circumstances the different methods of parameterization of field capacity can lead to better or poorer performances of the bucket model. With respect to soil texture, overall the water balance in coarser soils feels more the positive effect of determining the field capacity value through a more rigorous method such as the \(\text{drain-method}\). When considering seasonality, the dry season of vegetation regrowth (RVP period) appears more sensitive than the wet dormant season (DVP period) to the \(\text{drain-method}\) of parameterizing field capacity soil moisture.

Probability density functions (PDFs) of relative soil moisture, \(\rho(s)\), obtained using the two hydrologic models and for the two sub-periods of the hydrologic year are depicted in

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Table 5. Performance indices for the loamy-sand (LS) and clay (Cl) soil.

| Soil moisture | Transpiration rate | Leaching rate |
|---------------|--------------------|--------------|
|               | Year | RVP | DVP | Year | RVP | DVP | Year | RVP | DVP |
| LS, drain-method |     |     |     |     |     |     |     |     |     |
| ME            | 1.0E−02 | 6.2E−03 | 1.7E−02 | −6.7E−03 | −1.0E−02 | −2.1E−03 | 9.9E−03 | 8.7E−03 | 1.2E−02 |
| RMSE          | 2.4E−02 | 2.2E−02 | 2.7E−02 | 4.8E−02 | 6.2E−02 | 1.1E−02 | 4.6E−02 | 2.4E−02 | 6.5E−02 |
| LS, fix-method |     |     |     |     |     |     |     |     |     |
| ME            | −1.3E−02 | −1.6E−02 | −8.1E−03 | −1.8E−02 | −2.9E−02 | −2.1E−03 | 1.8E−02 | 9.7E−03 | 3.1E−02 |
| RMSE          | 5.6E−02 | 5.9E−02 | 5.2E−02 | 7.2E−02 | 9.4E−02 | 1.1E−02 | 8.3E−02 | 3.1E−02 | 1.2E−01 |
| Cl, drain-method |     |     |     |     |     |     |     |     |     |
| ME            | 9.1E−03 | 7.1E−03 | 1.1E−02 | −9.0E−04 | −1.9E−04 | −2.1E−03 | 8.3E−03 | 6.9E−03 | 9.0E−03 |
| RMSE          | 1.9E−02 | 1.9E−02 | 2.0E−02 | 4.2E−02 | 5.2E−02 | 1.1E−02 | 3.9E−02 | 1.7E−02 | 5.8E−02 |
| Cl, fix-method |     |     |     |     |     |     |     |     |     |
| ME            | 9.4E−03 | 7.2E−03 | 1.3E−02 | −9.4E−04 | −2.0E−04 | −2.0E−03 | 8.4E−03 | 7.0E−03 | 1.0E−02 |
| RMSE          | 2.0E−02 | 1.8E−02 | 2.2E−02 | 4.1E−02 | 5.3E−02 | 1.0E−02 | 4.1E−02 | 1.8E−02 | 5.9E−02 |

The shapes of the $p(s)$ functions for single-layer bucket models are positively skewed and narrow distributions, with relative soil moisture contents that vary in a small range around the peak and show quite long tails starting soon after the relative soil moisture point $s^*$ of incipient stomata closure. The modal value located close to the wilting point corresponds to the preferred state over the main part of the RVP, whereas the succeeding long tail located in the right side of the plot is representative of the fast transition from a wet state at the end of the dormant season toward a nearly dry stationary state. During this vegetative (re-)growing period, vegetation transpiration dominates the water balance and controls soil moisture contents in the active soil zone, which are consistently low. Consequently, leakage becomes a negligible process. For the fraction of the hydrologic year corresponding to the wetter DVP period (Fig. 5b and 5d), the fluctuation of precipitation and storage capacity of soil are perceived to dominate over the smaller amount of transpiration and to exert a remarkable control on soil moisture variability over time. Most of the soil water stored in the system results in unstressed evapotranspiration. The different precipitation regime and transpiration characteristics of this period basically make the $p(s)_{DVP}$ functions to be broader than those of the previous case and more concentrated around $s$ values near the field capacity point, $s_{fc}$.

To better frame the results depicted in Fig. 8, a quantitative but simple analysis can be carried out for identifying under which conditions the soil moisture at field capacity, $s_{fc}$, can play a more dominant role in the computation of soil water balance. This analysis is particularly relevant to the objective of the present study since this role can have different features depending on the season (e.g. the RVP phase) of the entire hydrologic year one is dealing with. In single-layer bucket-type models, soil moisture depletes almost immediately due to the evapotranspiration process and is replenished when precipitation occurs. Allowing for seasonality, depletion and replenishment of soil moisture can be faster or slower depending on the specific exchanges between atmosphere and soil during the considered period of the year. An evaluation
of the response timescales of soil moisture in the root zone can help in gaining a better understanding of the model comparisons presented in this study.

The evolution of the dry RVP season depletes soil moisture from the rooting zone as precipitation falls only over a small fraction of this period. The depletion time indicates how long it would take the maximum plant-available soil-water holding capacity, \( \text{PAWS}_{\text{max}} = (s_{\text{fc}} - s_w) \times n \times Z_r \), of the soil-vegetation system to dry out subject to the net outflow, \( R_d = ET_{\text{max}} - P \), between maximum evapotranspiration and mean precipitation rates \( (P = \eta \cdot \lambda) \), that remains constant and not considering other fluxes (Beljaars et al., 1996; Viola et al., 2008). In a similar way, we can define a replenishment timescale of the soil bucket during the DVP season assuming that the term \( R_r = P - ET_{\text{max}} \) remains constant throughout that period. Accordingly, the depletion time, \( t_d \), and the replenishment time, \( t_r \), give an indication of average moisture residence time in the soil and are calculated as follows:

\[
\begin{align*}
    t_d &= \frac{\text{PAWS}_{\text{max}}}{R_d}, \\
    t_r &= \frac{\text{PAWS}_{\text{max}}}{R_r}.
\end{align*}
\]

These two timescales are in Table 6, which also reports the ratios \( \rho_d \) and \( \rho_r \) that relate the depletion and replenishment times to the durations of the RVP (214 days) and DVP (151 days) seasons, respectively. The values of ratio \( \rho_d \) are fairly similar among them and show that depletion times are approximately 25% or 20%, depending on the specific soil type, of the total length of the dry RVP season and this situation, from the one hand, justifies the presence of the extended tails in the soil moisture probability density functions, \( p(s) \) (see Fig. 8a and b). On the other hand, however, it indirectly suggests that the amount of water stored in soil is relatively small for most of this dry period and hence field capacity becomes a parameter that can exert a more limited influence on the computation of soil water balance.

Looking at the wet DVP season, the \( \rho_r \) values clearly indicates that now soil moisture stored in the bucket capacity is high over periods commensurate to the length of the DVP season. Therefore, one would expect that field capacity will play a more relevant role when computing the soil water balance during a wetter season. Consequently, and in a way consistent with the results depicted in Fig. 8, the greater the discrepancies between field capacity values as determined by
the drain-method or the fix-method (and this happens for the loamy-sand soil; see plots 5a and 5b), the worse appear the comparisons between the BM and RE-SWAP models. This discussion provides a more quantitative perspective of our results and also reinforces the significance of the performance indices reported in Table 5. By dwelling on relative soil moisture over the DVP season, we have a reduction in RMSE between the two methods of about 48.1% for the loamy-sand soil and of about 9.1% for the clay soil. Instead, for the leaching rates computed over the DVP season there are RMSE reductions of about 45.8% and 1.7% for the LS and Cl soils, respectively. As expected, transpiration rates over the RVP period occur for the loamy-sand soil and when adopting the fix-method to parameterize the bucket model. Compared to the SWAP reference model, the performances of the bucket model are acceptable only when one refers to the RVP period, when there is a rapid increase of the saturation degree of the soils considered, but this is also partly due to the narrow temporal persistence around preferential states and hence a bimodality in the probability density function of relative soil moisture time-series. In this case, the variability of precipitation imposed during the DVP season, and specifically the frequency of storm occurrence, can result in the bimodal shape of the \( p(s) \) distributions depicted in Fig. 8. Over all, the soil moisture probability distributions in the plots 5b and 5d show that during the DVP period the soil, being of coarser or finer texture, can preferentially be in either wet states close to field capacity, or relatively drier states closer to the incipient stomata closure. In other words, the bimodality observed in the DVP is representative of two dominant states: one during the transition stage from the dry season to the wet season, when there is a rapid increase of the saturation degree above the stress value threshold; the other corresponding to a stationary wet state.

With respect to the two different methods to determine parameter \( s_{fc} \), the soil moisture PDFs of Fig. 8 help in identifying under what circumstances might it be appropriate to use the fix-method to parameterize the bucket model. Compared to the SWAP reference model, the performances of the bucket model are acceptable only when one refers to the RVP period (plots 8a and 8c). It is interesting to note that this behavior occurs independently from the textural properties of the soils considered, but this is also partly due to the narrow shape of the soil moisture probability distributions over the considered regrowing season. In accordance to the performance indices of Tables 1 and 2 and as one would expect, the larger discrepancies between SWAP and BM during the RVP period occur for the loamy-sand soil and when adopting the fix-method for estimating \( s_{fc} \). Instead, the plots 5b and 5d pertaining to the dormant and wetter season (DVP period) overall show rather poor performances of the bucket model especially when \( s_{fc} \) is estimated by the fix-method and for the coarser soil (Fig. 5b).

In general terms, determining the field capacity soil moisture by the traditionally-proposed technique of a prefixed point of the water retention function results in discrepancies with respect to the reference RE-based SWAP model of different extents depending on the soil type and the period of the hydrologic year. Errors become quite unacceptable when

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**Table 6.** Characteristic time scales of the bucket model during the DVP and RVP periods computed for the loamy-sand (L-S) and the clay (Cl) soils.

| Soil texture | Method | PAWS\(_{\text{max}}\) (cm) | RVP (214 days, \( R_d = 0.344 \text{ cm d}^{-1} \)) | DVP (151 days, \( R_e = 0.146 \text{ cm d}^{-1} \)) |
|-------------|--------|----------------|----------------|----------------|
|             | \( t_d \) (days) | \( \rho_d \) (%) | \( t_t \) (days) | \( \rho_t \) (%) |
| LS          | drain\(^a\) | 15.24 | 44.3 | 20.7 | 104.7 | 69.3 |
|             | fix\(^b\)  | 22.39 | 65.1 | 30.4 | 153.8 | 101.9 |
| Cl          | drain\(^a\) | 15.76 | 45.8 | 21.4 | 108.3 | 71.7 |
|             | fix\(^c\)  | 14.06 | 40.9 | 19.1 | 96.6 | 63.9 |

\(^a\) drain = \( s_{fc} \) defined by the drainage experiment; \(^b\) fix = \( s_{fc} \) from Eq. (2a) at \( h = 100 \text{ cm} \); \(^c\) fix = \( s_{fc} \) from Eq. (2a) at \( h = 500 \text{ cm} \).
Fig. 9. Daily values of the 5th, 50th, and 95th quantiles for the loamy-sand soil over the entire hydrologic year as obtained from the 99 years of simulation run. Red lines refer to the SWAP model used as a reference, while blue lines refer to BM when field capacity is parameterized either with the drain-method (plots on the left-hand side) or with the fix-method (plots on the right-hand side): (a) and (b) refer to relative soil moisture content, $s$; (c) and (d) refer to daily transpiration fluxes, $T$; (e) and (f) refer to daily drainage (leakage) fluxes, $L$.

One would capture soil moisture dynamics in coarser soils and in the rainiest season of a Mediterranean climatic area.

All the above comments and discussions are also clearly reflected in the representation of Fig. 9 showing the 5th, 50th, and 95th quantiles of relative soil moisture content, $s$, over the entire hydrologic year as obtained from the 99 years of simulation run. The various plots of this figure pertain to the loamy-sand soil only as all the results discussed before have clearly shown that the main discrepancies when BM is differently parameterized occur for coarser textured soils. The left plots of Fig. 9 (namely 9a-c-e) depict the simulation results when field capacity of BM is determined by the drain-method; the right plots (9b-d-f) represent the simulation results when field capacity of BM is estimated by the fixed method.
in Fig. 3. In general, the time variations of transpiration fluxes as computed by BM during RVP are in agreement with those from the Richards equation only when soil moistures start attaining lower values and close to the critical moisture value $s^*$. The drier are moisture conditions in this soil, the better BM transpiration fluxes follow the time fluctuations of the RE-computed fluxes. Overall, the discussed outcome again reinforces the need and the importance of accounting for local seasonal conditions to evaluate the effectiveness of using one modeling scheme of a soil-vegetation system with respect to another. Comparisons appear to be more satisfactory during the DVP period even when the values of $s$ are relatively low. As expected, larger deviations of the BM losses from leakage occur for the loamy-sand soil and during the colder and rainier season straddling over the end of the DVP period and the beginning of the RVP period. The BM modeling scheme tends to underestimate leakage losses over both the RVP and DVP seasons (see also the positive ME values in Table 5 for this loamy-sand soil), with larger discrepancies with respect to RE-SWAP occurring over the DVP season since leakage becomes important under high $s$-values. It is worth mentioning again that DVP is a relatively wetter season, with greater soil moisture values almost throughout the period and hence more prone to failure in the model intercomparison because of the differently estimated values for the field capacity parameter.

6 Conclusions

Bucket models are widely used tools to represent land surface hydrology, usually at regional scales, mainly because they account for soil water changes in a relatively simple way. However, since soil water is a key regulator of primary hydrologic processes, a careful parameterization of these capacitance models is a crucial phase. We have shown that the commonly used approach to estimate the field capacity at a specific point of the water retention curve (usually at the soil water content at the suction head of 3.3 m) might lead in general to poorer predictions of the various terms contributing to the soil-water budget, particularly for coarse-textured soils. Allowing for the key role that field capacity plays in bucket models, it is advisable that this parameter is determined via a field drainage experiment (drain-method).

This suggestion is particularly relevant if one would employ a bucketing approach for modeling soil water budget in areas with a climatic regime characterized by marked seasonality. Our study has also emphasized the benefit of looking at the parameterization problem also by carrying out functional evaluations. Through numerical simulations and intercomparisons between the bucket model (BM) and Richards’ equation based SWAP model (RE-SWAP), one can see that larger discrepancies should be expected when dealing with coarser soils and field capacity for this soil type is parameterized using the water retention curve (fix-method). Of specific
interest and a novelty, as far as we are concerned, is the possibility to further frame this outcome while considering a seasonality in the input climatic forcing. At least for the cases considered in this study, the bucket model with field capacity as determined by the fix-method provides some relatively good predictions preferably when considering finer textured soils and when soil-water balance refers to a drier season of the hydrologic year and vegetation is in under a regrowing stage (RVP period). It is important to point out, however, that the way with which we have parameterized both BM and SWAP is such so as to make boundary conditions and transpiration functions as similar as possible and to perform effective intercomparisons. Finally, analyses of the probability density functions of relative soil moisture content have highlighted the possible occurrence of bimodality in the shape of curves during the dormant vegetation phase as result of the combination of a dry-to-wet transition and a wetter state in the soil. A discussion on this bimodality is far from the purpose of this paper and it can be the subject of a subsequent study.

Appendix A

Abbreviations

BM, bucket model; RE, Richards’ equation; SWAP, Soil Water Atmosphere Plant model; DVP, dormant vegetation phase; RVP, regrowth vegetation phase; ME, mean error; RMSE, root-mean-square error.

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