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**Spatial variability of soil surface properties and
consequences for the annual and monthly water balance of a semiarid
environment (EFEDA experiment)**

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1 **Abstract**

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3 During the second phase of the EFEDA experiment (*ECHIVAL* Field Experiment in a
4 Desertification Threatened Area), the spatial variability of the soil water retention and hydraulic
5 conductivity characteristics of layers 2-12 and 17-27 cm depth was characterized. A simplified
6 method, based on particle size distribution and simple infiltration tests was used. It provided these
7 characteristics at the nodes of a 1 km grid over 10x10 km² around the town of Tomelloso (Castilla-
8 La Mancha, Spain).

9 A total number of 78 sample points were used to address the problem of soil surface
10 properties variability and its consequences on the monthly and annual water balance. The *SiSPAT*
11 (Simple Soil Plant Atmosphere Transfer model) 1-D Soil-Vegetation-Atmosphere Transfer (SVAT)
12 model was run with a one-year climatic forcing for the 78 soil profiles until equilibrium was
13 reached. As no runoff was generated, the spatial variability of the water budget components only
14 concerned soil evaporation, transpiration and deep drainage. It was found that i) the choice of the
15 type of boundary condition at the bottom of the soil profile was greatly influencing the final
16 variability, ii) the variability of transpiration was the largest in situations of water stress for the
17 vegetation, iii) soil evaporation was the most sensitive component when plants were well supplied
18 with water.

19 Various aggregation methods of soil surface parameters (use of the arithmetic mean, median
20 of the parameters or parameters associated to the average soil texture of the Clapp and Hornberger
21 (1978) classification) were assessed. The use of median parameters in a single 1-D simulation was
22 found to provide the best agreement with the average of the 78 simulations performed for each grid
23 cell using locally measured soil properties. The use of average soil texture parameters led to a
24 significant bias, especially in the case of water stress.

1. Introduction

In atmospheric and General Circulation Models surface and soil hydraulic properties are often assumed to be homogeneous over meshes of 100 to 10000 km². However, these characteristics and in particular soil hydraulic characteristics are known to be highly variable in space (e.g. Nielsen et al. 1973; Russo and Bressler 1981; Vauclin et al. 1994; Mallants et al. 1996; Lewan 1996; Bell et al. 1980; Haverkamp et al. 1996). The practice of assuming homogeneous properties can lead to errors in the calculation of surface fluxes. As these fluxes constitute the lower boundary condition of most large scale atmospheric and/or climatic models the prediction reliability of these models is strongly reduced (Shao and Henderson-Sellers 1996; Henderson-Sellers 1996). The introduction of some of the variability aspects into General Circulation Models (GCMs) using statistical dynamic approaches has been shown to improve significantly their performance (e.g. Entekhabi and Eagleson 1989; Famiglietti and Wood 1991 and 1992; Avissar 1992 and 1995). A large effort was also dedicated to the definition of "effective" or "aggregated" parameters (e.g. Raupach and Finnigan 1995), supposed to provide the same mean flux as that obtained by resolving explicitly the spatial variability. When soil moisture was fairly homogeneous, Noilhan and Lacarrère (1995) obtained a reasonable agreement between average evaporation fluxes derived from a 3D atmospheric model and the equivalent 1D simulation. When variability of soil hydraulic properties was large some discrepancies could be observed between average fluxes and those derived using average soil parameters (e.g. Braud et al. 1995b; Kim and Stricker 1996; Kabat et al. 1997; Braud 1998; Boulet et al. 1999). Milly and Eagleson (1987) and Kim et al. (1997) showed that the effect of soil parameter variability was largest in the case of runoff generation and that, in general, effective parameters could be more easily defined for total evaporation.

In this paper the regional scale (10x10 km²) relevant for atmospheric models is considered.

First, a data set collected in central Spain in Castilla-La Mancha is presented. Water retention and hydraulic conductivity parameters were measured at the nodes of a 1-km grid over 100 km². For the estimation of soil hydraulic properties, existing methods can be categorized as being either predictive or based on direct experimental measurement techniques. Observations can be made on samples in the laboratory or in-situ at the location of interest. Methods based on direct observations are often difficult to implement and time consuming. Predictive methods employ information on textural and structural properties such as particle size distributions, organic matter and/or dry bulk density in order to estimate the hydraulic properties (e.g. Clapp and Hornberger 1978). The data set presented here uses a simplified in-situ method, aiming at minimizing time and human resources needed to estimate soil hydraulic properties when the number of samples is large. In the following, the soil water retention and the hydraulic conductivity curves are represented by the Brooks and Corey (1964) model:

$$\begin{cases} \frac{\theta}{\theta_s} = \left(\frac{h_{bc}}{h} \right)^\lambda & \text{for } h \leq h_{bc} \\ \theta = \theta_s & \text{for } h_{bc} \leq h \leq 0 \end{cases} \quad (1)$$

$$\frac{K}{K_s} = \left(\frac{\theta}{\theta_s} \right)^\eta \quad (2)$$

where h (m) is the soil water pressure, θ (m³ m⁻³) is the volumetric moisture content and K (m s⁻¹) is the soil hydraulic conductivity. At each location, there are 5 unknown parameters: two shape parameters λ (-) and η (-) defined by the soil textural properties; and three scale parameters i.e. the saturated water content θ_s (m³ m⁻³), the saturated hydraulic conductivity K_s (m s⁻¹) and the scale parameter for the pressure h_{bc} (m), all three strongly related to the soil structural properties (Haverkamp et al. 1998a). The residual water content is assumed to be zero.

The data set was used to study at regional scale the variability of the components of the annual and monthly water budget of a vineyard representing the typical vegetation of the area. Several choices for the lower boundary condition of the soil profiles were considered in the Soil-Vegetation-Atmosphere Transfer (SVAT) model used for this study. Different rules for aggregating surface soil parameters were investigated in order to analyze the feasibility of an effective lumped parameterization.

2. Materials and methods.

a. The study area.

The field measurement program was conducted in the framework of the *ECHIVAL* (European International Project on Climatic and Hydrological Interactions between the Vegetation, the Atmosphere and the Land-surface) program. This program was dedicated to the study of the interaction between the soil, the vegetation and the atmosphere in various regions of the globe. One of these experiment, EFEDA (ECHIVAL Field Experiment in a Desertification Threatened Area), focused on the semiarid environment (Bolle et al 1993). The study area was situated near Tomelloso, in the Castilla-La Mancha region located in central Spain (Fig. 1). An area of 10x10 km² (2° 58' 03" W, 39° 09' 36" N for the top left corner and 2° 51' 46" W, 39° 04' 42" N for the bottom right corner) was selected. The area represented a typical grid element within a GCM. Determination of aggregation/ disaggregation rules for surface parameters and fluxes was the main focus of the experimental and modeling work. The soil of the study area was fairly uniform. However, a field campaign was conducted in June 1994 in order to document the spatial variability of soil hydraulic properties over the whole area. The latter was covered with a 1 km sampling grid mesh (Fig. 1) resulting in 100 measurement locations. The altitude varies from 669 m above sea

level at site A0 (Fig. 1) to 835 m at site G9, increasing in the NW-SE direction.

The average slope of the area was less than 2%. The area covered mostly cultivated land (92 sites) with the following land-use distribution: vineyards (49 sites); fallow land, generally part of a cereal-fallow rotation scheme (23 sites); barley (11 sites); wheat (5 sites); olive trees (3 sites) and chickpeas (1 site). Only 8 sites were lying at abandoned vineyards (5 sites) and shrubs (3 sites). Soil profiles were composed of an arable top layer with a depth of approximately 30 cm overlying a deep calcareous crust layer down to 4-5 m. Hydraulic properties of the crust layer were characterized using an internal drainage method at site TOM 6 (E4, Fig. 1). Details are given by Haverkamp et al. (1996). A significant proportion of stones and limestone porous crust fragments were encountered in the arable top layer. On the average 35% of the soil surface was covered by stones and rock fragments of 5-10 cm in diameter.

b. Derivation of water retention and hydraulic conductivity parameters:

1) SHAPE PARAMETER λ OF THE RETENTION CURVE:

The particle size distribution can be modeled using the following function (Haverkamp et al. 1998a) :

$$F(d) = \left[1 + \left(\frac{d_g}{d} \right)^N \right]^M \quad \text{with} \quad N = \frac{2}{1-M} \quad (3)$$

where d (m) is the particle-size diameter, d_g (m) is the particle-size scale parameter and M (-) the particle-size shape parameter.

M can be directly identified from particle size distribution data when such data are available. If only the soil texture is available (i.e. clay, silt and sand content), Zammit (1999) proposed a “cartography” of this parameter within the soil texture triangle based on the analysis of a 750 soil database called GRIZZLY (Haverkamp et al. 1998c) .

The shape parameters λ of the water retention curve can be derived from the knowledge of soil texture and, more specifically, from the knowledge of the values for M and N . The following relationship was fitted by Braud et al. (2002) on data from the GRIZZLY soil data base (Haverkamp et al., 1998c):

$$\lambda = MN(0.382 + 0.0469MN) \quad (4)$$

and was used in the present study for the estimation of parameter λ .

2) SATURATED VOLUMETRIC WATER CONTENT θ_s :

The soil porosity ε (–) can be determined from the soil dry bulk density ρ_d (g cm⁻³) and soil particle density ρ_s ($\rho_s = 2.65$ g cm⁻³) through:

$$\varepsilon = 1 - \frac{\rho_d}{\rho_s} \quad (5)$$

The saturated soil volumetric water content θ_s (m³ m⁻³) can be related to the porosity using the following relationship proposed by Haverkamp et al. (1998c):

$$\theta_s = \varepsilon 2^{\left(\frac{\lambda}{2+\lambda} - M\right)} \quad (6)$$

3) SHAPE PARAMETER h_{bc} OF THE WATER RETENTION CURVE:

Zammit (1999) showed that predictive methods based on textural analysis failed to predict this parameter. The latter is more related to soil structure and therefore to local effects (Haverkamp et al. 1998a). Consequently, the most reliable procedure to determine h_{bc} seemed to be through direct measurement of matric potential and water content or indirectly from simple infiltration tests (Braud et al. 2002). Infiltration is very sensitive to this parameter because h_{bc} dominates the retention behavior in the vicinity of saturation. It seems therefore advisable to make these measurements near saturation if only one field campaign can be conducted. If several campaigns

can be realized different parts of the retention curve must be covered. Least square errors optimization techniques on the data pairs (matric potential, volumetric water content) available can subsequently be used to derive h_{bc} assuming that the other parameters are known.

4) SHAPE PARAMETER η OF THE HYDRAULIC CONDUCTIVITY CURVE:

The derivation of the shape parameter for the hydraulic conductivity curve η can be performed by introducing a tortuosity factor p , given by:

$$M = \frac{\lambda}{2 + \lambda}(1 + p) \quad (7)$$

and related to η by (Haverkamp et al., 1998a):

$$\eta = \frac{2}{\lambda} + 2 + p \quad (8)$$

5) SATURATED HYDRAULIC CONDUCTIVITY K_s :

Like the scale parameter h_{bc} , models based on texture fail to predict the structure related scale parameter K_s (Zammit 1999). The most reliable way to get an estimation of this parameter was thus to perform in-situ experiments. One of the simple ways, which do not require too much time, consists of artificially wetting the soil under a constant positive pressure head. Then the Green and Ampt (1911) approach can be used to characterize the infiltration process (Hillel 1980). Assuming uniform initial volumetric soil water content, θ_o ($\text{m}^3 \text{m}^{-3}$), the Green and Ampt model leads to the following relationship between the vertical cumulative infiltration I_{ID} (m) and the corresponding infiltration time t_{inf} (s):

$$I_{ID} = K_s t_{inf} + (\theta_s - \theta_o)(h_{surf} - h_f) \log \left[1 + \frac{I_{ID}}{(\theta_s - \theta_o)(h_{surf} - h_f)} \right] \quad (9)$$

where h_f (m) is the wetting front suction, h_{surf} (m) is the constant head at the soil surface. The saturated hydraulic conductivity can be estimated from this expression. The suction at the wetting

front can be calculated by (10). Indeed, under positive head infiltration, the sorptivity is an intrinsic property of the soil and its value should not depend on the model chosen to represent the soil water retention and hydraulic conductivity curves. Therefore expressions obtained using the Brooks and Corey model and the Green and Ampt models must be equal leading to Eq. (10) (Haverkamp et al., 1998b; Braud et al., 2001).

$$h_f = \frac{2\lambda\eta(\lambda\eta-1) + \lambda(2\lambda\eta-1) h_{bc}}{(\lambda\eta-1)(\lambda\eta-1+\lambda)} \frac{h_{bc}}{2} \quad (10)$$

c. Experimental work and derivation of the water retention and hydraulic conductivity curves.

At each location over the sampling grid, measurements were carried out at two depths: 2-12 cm depth for layer 1 and 17-27 cm depth for layer 2, not situated on the same profile but at two adjacent spots. In June 1994, when the experiment was conducted, no rainfall had fallen since several months. Therefore, the surface layer was very dry and an artificial wetting was required. A constant head water supply device ("bucket method") consisting of a single-ring infiltrometer (a bottom free plastic bucket of 18 cm in diameter) and a 2-L plastic bottle acting as a Mariotte flask were used. Appropriate wetting (pressure range and wetting front depth) were obtained after infiltration of 1 to 2.25 L of water.

After drainage for a known time, the volumetric water content θ_{TDR} was determined through Time Domain Reflectometry (TDR) (Topp et al., 1980) using a Tektronik 1052 C cable tester. The soil matric potential h was measured simultaneously using a hand held micro-pressure transducer tensiometer developed by Division of Soils, CSIRO, Townsville, Australia. Besides these in-situ measurements of volumetric water content and water pressure, soil samples were collected at the same depths for dry bulk density and gravimetric water content W_{ap} determination. An auger with a volume of about 500 cm³ was used to sample approximately the same soil volume as that sensed by the TDR probe.

1 A team of five people accomplished the fieldwork in two weeks (seven sites per day).
2 Average time taken per soil site was 67 minutes (20 minutes for location of site and transportation,
3 24 minutes for water infiltration and 23 minutes for actual soil measurements and sampling). This
4 makes this type of soil characterization campaign accessible even for large areas, because the
5 human and time investment is relatively low.

6 The collected soil samples were also used for particle size distribution analysis, combining
7 sieving and sedimentation techniques for the determination of the following particle-size fractions:
8 very coarse sand (2000-1000 μm), coarse sand (1000-500 μm), medium sand (500-200 μm), fine
9 sand (250-100 μm), very fine sand (100-50 μm), coarse silt (50-20 μm), medium-fine silt (20-2 μm)
10 and clay ($< 2 \mu\text{m}$) (Gee and Bauder 1986). The percentage of stones and crust fragments in the
11 coarse soil fraction ($> 2000 \mu\text{m}$) was determined before the particle-size analyses.

12 It has been mentioned that a large proportion of stones and crust fragments were present in
13 the soil samples. Furthermore, it was observed that the crust fragments were also holding water.
14 Therefore the dry bulk density derived from θ_{TDR}/W_{ap} was only an apparent value and was different
15 from the fine soil bulk density ρ_d , needed in Eq. (5) used to derive the porosity. This bias was
16 obvious given the substantial number of sites having large values of apparent dry bulk density
17 (higher than 2 g cm^{-3} in some cases). A simple correction (Haverkamp, unpublished work) was thus
18 developed to obtain firstly the fine soil volumetric water content θ , obviously lower than θ_{TDR} , and
19 secondly the fine soil dry bulk density ρ_d . Data discussed below will be the corrected values.

20 Parameters of Eq. (3) were fitted on the experimental fine soil particle size distribution
21 functions. Then the shape parameter λ was deduced from Eq. (4) and introduced in Eq. (6) to
22 determine the saturated water content θ_s . Once λ and θ_s were known for each layer, the scale
23 parameter h_{bc} was calculated by introducing the only data pair (h, θ) available in Eq. (1). It was not
24 possible to use the two layers to optimize the scale parameter, because the saturated water content

of both layers were found to be statistically significantly different (Haverkamp et al. 1996) and each layer had to be treated separately, which of course, reduces the robustness of the procedure.

For the hydraulic conductivity curve at each location, the shape parameter η was evaluated using Eqs. (7) and (8). The saturated hydraulic conductivity was derived from (9), assuming a value of the surface suction h_{surf} of 1 cm. The initial soil moisture content θ_o was not available. It was assumed that, given the long time since the last rainfall, steady state over the whole area had been reached. A constant arbitrary value $h_o=10^7$ cm, representative of dry surface conditions, was thus assumed at each point and the corresponding water content calculated using (1). θ_o values ranged between 2 and 10 %. It was checked that the results on K_s were not very sensitive to the choice of this value for h_o . Furthermore, the measured infiltrated height was representative of the 3D infiltration, whereas the 1D value must be used in (9). The following correction was applied (Haverkamp et al. 1994):

$$I_{1D} = I_{3D} - \frac{\gamma}{r_d} 2(h_{surf} - h_f) K_s t_{inf} \quad (11)$$

where $\gamma=0.7$ and r_d is the radius of the infiltrometer. This leads to the following equation, solved iteratively to obtain K_s , once h_f had been estimated using (10).

$$K_s t_{inf} = \frac{I_{3D} - (\theta_s - \theta_o)(h_{surf} - h_f) \log \left[1 + \frac{I_{3D}}{(\theta_s - \theta_o)(h_{surf} - h_f)} - \frac{\gamma}{r_d} \frac{2K_s t_{inf}}{(\theta_s - \theta_o)} \right]}{1 + \frac{2\gamma(h_{surf} - h_f)}{r_d}} \quad (12)$$

The experimental protocol used during the EFEDA experiment and reported in this paper was a first attempt to develop such a large scale strategy for soil hydraulic properties derivation. Since this early work, the method has been refined and theoretically justified. An update of the procedure, known as the Beerkan method, is provided in Braud et al. (2002) and could be used in practice.

Fig. 2 shows the retention curves and hydraulic conductivity curves drawn using (1) and (2) and the values as calculated from the method described above for all the measurement points. A large

scatter can be observed showing the variability of those curves at this scale. Table 1 summarizes the statistical properties of the different parameters. The spatial correlation of the data was also examined through variogram analysis. No spatial correlation was detected at the scale of the measurements (1 km), justifying therefore the assumption of statistical independence of the soil profiles used in the subsequent modeling work. At the field scale (4 ha), Vandervaere (1995) found correlation lengths of about 150 m for soil dry bulk density and hydraulic conductivity. The coefficients of variation (8-10% for porosity and 20-30% for hydraulic conductivity) were much smaller than those reported in Table 1 calculated at the regional scale.

d. The SiSPAT model.

An extensive description of the SiSPAT model can be found in Braud et al. (1995c) and Braud (2000) and only a brief summary is presented here. SiSPAT is a vertical 1-D model. The driving forces are climatic time series of air temperature and humidity, wind speed, incoming solar and long-wave radiation and rainfall. In the soil, coupled heat and mass transfer equations, derived from the Richards (1931) equation are solved for temperature T and matric potential h . They include both liquid and vapor transfers as formulated by Philip and De Vries (1957) or Milly (1982). The model deals with vertically heterogeneous soils. The upper boundary conditions are obtained by the solution of the soil-plant-atmosphere interface, modeled as a two source model of heat and vapor, where bare soil and vegetation are considered separately (Deardorff, 1978), as formulated by Shuttleworth and Wallace (1985) and Taconet et al. (1986). Five equations can be written: energy budget over bare soil and vegetation, continuity of the sensible and latent heat fluxes through the canopy and continuity of the surface flux at the soil surface. Leaf temperature T_v , canopy temperature T_{av} , canopy specific humidity q_{av} , soil surface temperature T_I and surface matric potential h_I can thus be calculated. The resolution of this module provides the surface soil

1 heat and masses fluxes and the surface matric potential h_I and temperature T_I . In the case study,
2 surface matric potential and temperature were used as the upper boundary condition of the soil
3 module for all the simulations because it proved to be numerically more stable. When rainfall
4 exceeds the infiltration capacity of the soil, saturation of the surface occurs. The matric potential at
5 the surface is set to zero and the runoff is calculated from the mass budget equation.

6 The incoming energy is partitioned between bare soil and vegetation through a shielding
7 factor σ_f (Deardorff 1978; Taconet et al. 1986). In the soil, a root extraction model, based on
8 Federer (1979) is included. For each soil layer, a soil-root and a root-leaf resistance are put in
9 series. The moisture extraction in layer j is proportional to the water potential difference between
10 the leaf h_f and the soil h_j . The leaf water potential is calculated by assuming steady state at each
11 time step and that total moisture extraction is equal to the transpiration calculated from the
12 atmospheric conditions. The leaf water potential controls the water stress function of the stomatal
13 resistance, which also depends on the incoming radiation and vapor pressure deficit. Iterative
14 procedures are used to solve the various modules of the model. They are described in detail in
15 Braud (2000).

16
17 *e. Method used to assess surface fluxes variability and effective parameters soundness, in response*
18 *to surface soil parameters variability.*

19 At the end of the data processing, 78 sample points were available with estimation of both
20 the retention and hydraulic conductivity curves for layers 2-12 and 17-27 cm depth. 22 sample
21 points had to be removed from the analysis because the infiltrated volume was not recorded at the
22 beginning of the field work and the saturated hydraulic conductivity could not be determined for
23 these points. These data were used to investigate the influence of soil surface hydraulic properties
24 spatial variability on surface fluxes, using the SiSPAT SVAT model. The methodology was the
25 following:

(i) One year of half-hourly values of climate forcing (air temperature, humidity and wind speed at two meters, solar and long-wave radiation and rainfall) typical of semi-arid Spain had been generated by the 1-D version of the UK Meteorological Office general circulation model (Lean 1992) in the framework of the SLAPS II (Spatial Variability of Land Surface Processes) (Dooge et al. 1994). This data set was used as the forcing of the SiSPAT SVAT model. Daily values are presented in Fig. 3. The rainfall and the other climate forcing variables were assumed to be homogeneous over the whole area, in order to focus the study on soil surface properties influence.

(ii) For the same reason, one vegetation type, vineyard, typical of the area of Tomelloso was chosen. On the 100 km² studied area, vineyard was only covering 43 % of the surface but Sene (1996) reported values of 70 to 80 % for the whole EFEDA zone. Plant parameters necessary to run the SiSPAT model had been calibrated by Braud et al. (1995a). The use of these parameters, including plant resistance, maximum root density, and parameters describing the stomatal conductance, allowed to reproduce quite well one month of observations of surface fluxes measured in 1991 during the first EFEDA experiment. The root mean square error on latent and sensible heat flux were 37 W m⁻² and 25 W m⁻² respectively. These figures lie in the range of errors reported in an intercomparison study of various surface schemes (Linder et al. 1995) performed using the same data set. A drill showed that the vineyard roots could reach a depth of 3 m (Santa Olalla Manas 1994), a soil profile of 4 m was chosen with a rooting depth of 3 m. A plausible one year time series of Leaf Area Index (*LAI*) was defined for a vineyard cultivation, based on information given in Sene (1996). It also appears on Fig. 3.

(iii) It was assumed that the whole area could be represented by independent soil profiles, with vertical heat and water transfers. This hypothesis was reasonable as the water table was located below 10 m depth (Sene 1996) and no spatial correlation between parameters was found at the scale of the study. Such an hypothesis was also retained in the work of Peck et al. (1977). 78 soil profiles were defined and the water retention and hydraulic conductivity curves measured at the grid points

were assigned to one of them for layers 0-20 and 20-50 cm. No data about the variability of soil properties below the 17-27 cm layer were available. Therefore the properties for the 50-400 cm layers were kept identical for all soil profiles and assigned a value measured for the calcite crust encountered at that depth (Haverkamp et al. 1996). The saturated hydraulic conductivity of that layer was measured using suction infiltrometers leading to $K_s=2.78 \times 10^{-6} \text{ m s}^{-1}$ (Vandervaere, personal communication). This value was consistent with a value of $K_s=4.63 \times 10^{-6} \text{ m s}^{-1}$ used by Sene (1996) in a water balance study performed in the same region. Another trial was also performed with a value of the saturated hydraulic conductivity divided by 10.

(iv) For the 78 soil profiles, the same atmospheric forcing was applied during two or three years in order to obtain equilibrium of the solution. In analogy with climate modeling, the word “equilibrium” means that the soil water storage at the beginning of the simulation was equal to the soil water storage at the end of the simulation (i.e. after one year). This ensured that model results did not depend on the initial moisture conditions and that various scenarios could be compared with each other (Dooge et al., 1994). The same vegetation characteristics and the same lower boundary condition (sinusoidal annual cycle for temperature and either a constant value of the matric potential or gravitational drainage for the moisture) were retained for all 78 soil profiles.

(v) Then the mean water budget was calculated at the annual and monthly time scales by averaging the results of the 78 profiles, each profile representing 1/78 of the whole area. The variability of those components of the water budget was assessed. A comparison with a single simulation using lumped parameters derived using various methods was conducted. Simulations using lumped parameter values associated with the average texture, as given by the Clapp and Hornberger classification were also performed. The differences in the water budget obtained with the various aggregation scenari and the average of the 78 profiles outputs could therefore be quantified

Milly and Eagleson (1987), followed by Kim et al. (1997) proposed an analytical framework

for the derivation of effective parameters, based on reasonable assumptions about the probability density functions of the soil parameters. Such a generality was not sought in the present paper, which aimed only to test the relevance of possible parameter aggregations readily available from measured data.

3. Results:

a. Analysis of the annual water balance.

The assumptions retained in the modeling approach constrained greatly the modeled annual water balance. Due to equilibrium, annual change in water storage was zero. Rainfall was the same for all the soil profiles, because it was chosen to focus the analysis on soil surface properties influence. Calculated runoff was zero in all the cases, because the soil saturated hydraulic conductivity was very high (more than 120 mm/day) as compared to rainfall intensity (less than 20 mm/day). Therefore, the variability of the water balance was reduced to a balance between total evaporation and deep drainage. Deep drainage was proportional to the hydraulic conductivity at the bottom of the soil column. Soil characteristics of the lower layer were the same for all the soil profiles. Therefore, the lower boundary condition choice and the value of the saturated hydraulic conductivity of this layer mainly governed deep drainage. Table 2 shows that if a constant matric potential was assumed at the bottom of the soil profile, capillary rises were generated (upwards flux) whereas the flux was directed downward if gravitational flux was assumed. In the latter case, drainage at the end of the first year decreased from 146 to 35 mm when the bottom saturated hydraulic conductivity was divided by ten. When equilibrium was reached, variability of drainage was low (Tables 4 to 6). As a consequence, a change in surface soil hydraulic properties mainly affected the partition of total evaporation between bare soil evaporation and transpiration by the plants.

1) INFLUENCE OF LOWER BOUNDARY CONDITION CHOICE:

Table 2 provides the terms of the mean annual water budget (average over the 78 profiles) for the three choices of lower boundary condition (Case 1: constant matric potential (with $K_s=2.78 \times 10^{-6} \text{ m s}^{-1}$); Case 2: gravitational flow with $K_s=2.78 \times 10^{-6} \text{ m s}^{-1}$; Case 3: gravitational flow with $K_s=2.78 \times 10^{-7} \text{ m s}^{-1}$) and the two/three years of the simulation.

Table 2 shows that quite different results were obtained according to lower boundary condition choice for the water flow equation. In the case of constant matric potential, large capillary rises were simulated (almost 200 mm/year). This figure was very high and one might suspect that a lower boundary condition defined with a value of the matric potential evolving through the season would be more realistic. Unfortunately, no data were available to define such an annual course. When the gravitational flow was considered, the choice was not very satisfactory because only downward fluxes were allowed and capillary rises were systematically excluded. Furthermore, the value of the saturated hydraulic conductivity $K_s=2.78 \times 10^{-6} \text{ ms}^{-1}$ (chosen consistently with earlier published work on this area) at the bottom led to large values of drainage, especially for the first year and then, few water remained in the soil profiles for the second year. At the end of the second year equilibrium was not reached. However, it was not possible to run the model for a third year, because of model divergence, associated with a very low water content in the soil profiles. Therefore, another trial was done by dividing the value of the saturated hydraulic conductivity of the lower layer by ten in order to avoid too much drainage.

Boundary condition choice was crucial for such a long-term study and results in terms of water balance were very different according to this choice. When matric potential was constant at the bottom of the soil profile, total evaporation was larger than rainfall, due to almost 200 mm of capillary rise. With a gravitational flow, total evaporation was almost equal to rainfall, because deep percolation was very small, once equilibrium was reached. It was also interesting to see that,

lower boundary condition choice mainly affected plant transpiration, which was greatly reduced in the gravitational flow case, whereas bare soil evaporation was almost the same. The change in mean water storage implied that less water was available for transpiration in deeper layers, whereas bare soil evaporation was more linked to surface soil moisture, and therefore to rainfall time evolution (see also discussion in 3.b).

Finally, it seems that for such a study, the ideal choice for the lower boundary condition would be an imposed value of the matric potential evolving with time, provided that such a time evolution could be defined. This is in fact very difficult in practice, because such an evolution depends on rainfall time evolution and redistribution of water within the soil profiles, i.e. on model results themselves. Another choice could be to use a very deep soil profile, where the lower boundary condition could be defined by the water table. However, in this case, the determination of soil hydrodynamic properties for deeper layers becomes a problem because these layers are poorly known and a coupling with a 2-D groundwater model could be required.

The problem posed by the specification of the soil lower boundary condition is more crucial for models based on the Richards (1931) equation than for reservoir models, not resolving explicitly the diffusion equation within the soil (e.g., the ISBA model of Noilhan and Planton, 1989). The predictive potential of Richards equation based models is therefore limited. Boone and Wetzel (1996) and Lee and Abriola (1999) also reported sensitivity of such models, used within General Circulation Models, to the lower boundary conditions representation and to the soil discretization.

Nevertheless, results obtained using the SiSPAT SVAT model are worth discussing because some common features can be extracted, independently of the lower boundary condition choice.

2) EVALUATION OF SEVERAL AGGREGATION RULES FOR SURFACE SOIL HYDRAULIC PROPERTIES:

The average water budget was compared with one 1-D run conducted with aggregated

1 parameters. Two choices were tested. "*aagg*" means arithmetic mean for λ and θ_s and geometric
2 mean for h_{bc} and K_s , and corresponds to the median of the four parameters, according to their fitted
3 probability density function (see Table 1). "*aaaa*" means arithmetic mean for all the four
4 parameters. Water budget values derived with parameters calculated from the Clapp and
5 Hornberger (1978) classification for the average soil texture of the two surface layers are also
6 given. Corresponding parameter values are summarized in Table 3. Table 4, 5 and 6 provide the
7 statistical analysis of water budget annual components at the end of the second or third year.

8
9 For the first lower boundary condition choice (imposed constant matric potential value), the
10 variability in the water balance components was less than 10% and was maximum for bare soil
11 evaporation and deep drainage. Transpiration was not affected very much because, due to capillary
12 rise, vegetation was always well supplied with water. On the other hand, in the second and third
13 cases (gravitational flow), a large amount of water left the soil profiles through percolation and the
14 average water content could become lower than the wilting point. Vegetation water stress appeared
15 for some soil profiles and transpiration variability was the highest (31 to 37%) whereas bare soil
16 evaporation showed a lower variability than in the first case (7%). Contrasts in variability between
17 Case 1 and 2 or 3 resulted from the steady state being reached. At the end of the first year (when
18 steady state was not reached), the water balance components' variability was similar for the three
19 lower boundary condition choices, except for deep percolation (not shown). Note also that although
20 transpiration and bare soil evaporation variability might be large, total evaporation variability was
21 always very small, due to compensation effects.

22 When looking at the results in terms of aggregated parameters, median values for the four
23 soil parameters ("*aagg*") led to the closest agreement between the 1-D run with the aggregated
24 parameters and the 78 soil profiles simulations average in the three test cases. The arithmetic mean
25 led to a larger bias in terms of partition between bare soil evaporation and transpiration (average

scale parameter h_{bc} and saturated hydraulic conductivity K_s were larger than median values used in the previous case because the probability density functions were lognormal). In general, transpiration was underestimated with the aggregated parameters and bare soil evaporation overestimated. Due to compensation effects, the bias on total evapotranspiration was however small (less than 5%, except for run CH in Case2 where it reached 15%). The bias on evaporation and transpiration was small using median values of the parameters (less than 5%), larger using the arithmetic mean (between 5 and 30%) and very large using mean parameters derived from the Clapp and Hornberger classification, especially in the gravitational flow case (Case 3) where it reached 55%. In this case, predicted transpiration was half the true value.

Therefore in the case study where no runoff was simulated, median values used as aggregated soil surface parameters led to a satisfactory simulation of the water balance with one "equivalent" 1-D simulation. This result was consistent with studies reported by Milly and Eagleson (1987), Kim and Stricker (1996), Kim et al. (1997), Braud (1998), Boulet et al. (1999). They showed that runoff occurrence was triggering the effect of the spatial variability of soil properties on surface fluxes. Estimation of the median value for the parameters remains however an unsolved problem, given the large variability of some of them (see Fig. 4) requiring a large sample for achieving a robust estimation.

b. Analysis of the monthly water balance.

Results are summarized on Fig. 4 for the second year of Case 1 (imposed matric potential at the bottom), Fig. 5 for the second year of Case 2 (gravitational flow with saturated hydraulic conductivity of $2.78 \times 10^{-6} \text{ m s}^{-1}$) and Fig. 6 for the third year of Case 3 (gravitational flow with saturated hydraulic conductivity of $2.78 \times 10^{-7} \text{ m s}^{-1}$). The mean monthly water balance of the 78 soil profiles $\pm$ one standard deviation are shown as well as the monthly water balance obtained with the 3 choices of aggregated parameters. Rainfall time evolution is also shown. Bare soil evaporation

1 appeared mainly related to rainfall whereas transpiration time evolution was more related to LAI
2 evolution (significantly different from zero between April and September). Tables 7, 8, and 9
3 provide the monthly coefficients of variation for total evaporation, bare soil evaporation and
4 transpiration for the last simulation year and the three lower boundary conditions choices.

5 Bare soil evaporation variability was maximum in May (32 %), a month with no rainfall
6 following a month with a large rainfall. This was especially true for Case 1 (constant matric
7 potential at the bottom of the soil profile), where transpiration variability was very small, because
8 the vegetation was well supplied with water, due to the capillary rise. For Cases 2 and 3,
9 transpiration variability was maximum (up to 66 %) during the dry period because of plant water
10 stress. Note that in Case 3, where the lack of water was larger, the coefficients of variation were
11 also the highest. For total evaporation, the monthly coefficient of variation did not exceed 30 %,
12 showing once again that compensation effects occurred between bare soil evaporation and
13 transpiration. In the gravitational flow case, deep drainage variability was very small and was much
14 larger in the imposed matric potential case (not shown).

15 When looking at the results obtained with the aggregated parameters, "*aagg*" and "*aaaa*"
16 choices almost always fell in the $\pm$ one standard deviation interval whereas the Clapp and
17 Hornberger results were often outside this interval, especially for transpiration in the dry period. At
18 the monthly scale, aggregated runs performed using parameters median values led to the best
19 agreement obtained with the 78 soil profiles average.

20

21 **4. Discussion and conclusions:**

22

23 A methodology to derive soil hydraulic properties on a large area was presented. It was
24 shown that soil water retention and hydraulic conductivity curves at a large number of sample
25 points could be derived using simple measurements and the variability could be characterized at the

1 regional scale. Since the early EFEDA campaign in 1994, the methodology has been improved and
2 is fully described in Braud et al. (2002).

3 The numerical study conducted using this data set focused on the influence of variability of
4 surface hydraulic properties on the annual and monthly water budget. Rainfall and vegetation cover
5 were the same for all the soil profiles. Results were discussed at equilibrium when the annual
6 change in water storage was zero and no runoff was generated, due to high values of the hydraulic
7 conductivity. Consequently, total evaporation and deep drainage variability was less than 10%,
8 regardless of the lower boundary condition, whereas the choice of the lower boundary condition
9 resulted in quite different scenarios for bare soil evaporation and transpiration. When the
10 combination of lower boundary condition and soil hydraulic properties generated water stress for
11 the vegetation on some soil columns, transpiration variability over the whole area was large. When
12 the lower boundary condition ensured that no water stress occurred for the vegetation, transpiration
13 variability over the whole area was small whereas that of bare soil evaporation variability was
14 higher. At the monthly time scale, differences observed between the various cases were enhanced,
15 leading to a very large variability of transpiration in case of water stress. A limitation of the study is
16 the lack of interaction between surface condition, the atmosphere and vegetation growth. The use of
17 a vegetation growth model coupled with an atmospheric model would be the next step to investigate
18 further interaction between soil surface properties variability and surface fluxes.

19 The study has shown that using median parameter values in a 1-D run was leading to a good
20 agreement with the 78 profiles associated with the measured surface properties average. The use of
21 arithmetic means or worse, of the Clapp and Hornberger classification was leading to a serious bias,
22 especially in case of water stress for the vegetation. Peck et al. (1977) found that for a forest cover,
23 an averaging procedure of soil parameters, based on the scaling theory of Miller and Miller (1956)
24 was providing a good agreement between average fluxes and fluxes calculated for the equivalent
25 medium. However, their period of study was much shorter than in our case (only a few months) and

1 the effect of initial conditions on the results was not considered.

2 When examining the consequences in terms of GCM modeling, it must be stressed that the
3 use of texture derived parameters (such as the Clapp and Hornberger classification) can lead to
4 significant bias in simulated partition between soil evaporation and transpiration, although total
5 evaporation might be correctly simulated. These results were obtained using an assumption of
6 homogeneous rainfall and vegetation characteristics and no runoff was simulated. The bias reported
7 in this paper was therefore certainly a lower bound. An intercomparison of SVAT models used by
8 climate modelers and hydrologists showed that the runoff term was responsible for the largest
9 differences between models and that when rainfall variability was taken into account, simulated
10 runoff was considerably modified (Dooge et al. 1994).

11 Given the large soil properties spatial variability, and therefore the large sample needed to
12 obtain a representative value, the experimental effort needed to get the median value is the major
13 obstacle to the derivation of such values for large areas. The methodology presented in the first part
14 of this paper represents the smallest effort achievable at the present time, but is still difficult to
15 apply on a large number of regions. Future research should probably focus on the derivation of
16 representative set of soil parameters at the scale at which the modeling is conducted, provided that
17 these unit represent homogeneous areas. As an illustration, Soria et al. (2001) showed that the
18 fluxes calculated for an equivalent medium of two soils exhibiting the same structural parameters
19 and different textural parameters could not match the reference flux deduced from a full 2D model
20 of water transport, whereas the average flux was well represented by the weighted average of the
21 individual vertical fluxes. These results plead for parameterisation of sub-grid processes as weighed
22 averages of individual components (combination of soil type and vegetation cover for instance)
23 referenced as the mosaic approach (Koster and Suarez 1992). This seems more promising than
24 increasing model complexity, assuming homogeneous surfaces. Information on soil hydraulic
25 properties will probably still remain necessary. A research line which requires further development

1 (namely in order to restrain the number of independent parameters) is certainly the inversion of
2 SVAT models in order to match fluxes estimated at the scale of interest which could be provided by
3 remote sensing. The availability of evaporation fluxes at the scale of the model could be of great
4 help to reach this goal and for GCMs validation, but such remote sensing estimation are still not
5 reliable enough.. Assimilation of remote sensing within SVAT or hydrological models data could
6 also be a promising way of achieving this goal, provided the number of parameters to be estimated
7 remains sufficiently small to obtain robust estimations.

8

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14

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Fig. 2: Left: Retention curves of the 78 soil samples and Right: Hydraulic conductivity curves of the 78 soil profiles for Layer 2-12 cm. Points are the measurements made at each location.

Fig. 3: Daily value of the atmospheric forcing. Top panel: Air temperature ($^{\circ}\text{C}$), specific humidity ($10^{-3} \text{ kg kg}^{-1}$), wind speed (m s^{-1}). Bottom panel: Incoming solar and long-wave radiation (W m^{-2}), rainfall (mm) and Leaf Area Index (10^{-3} LAI) (-).

Fig. 4: Monthly evolution of the mean value $\pm$ one standard deviation (error bars) of the 78 soil profiles for total evaporation, bare soil evaporation, transpiration and deep drainage (positive for percolation and negative for capillary rises). Monthly values of rainfall are also shown together with the monthly evolution of the components of the water balance for the case "aagg" (open triangle) or "aaaa" (full square) averaging and the case where the soil hydraulic properties of the surface were derived from the Clapp and Hornberger classification (open diamond). Case 1: imposed matric potential at the bottom of the soil profiles (second year).

Fig. 5: Monthly evolution of the mean value $\pm$ one standard deviation (error bars) of the 78 soil profiles for total evapotranspiration, bare soil evaporation, transpiration and deep drainage (positive for percolation and negative for capillary rises). Monthly values of rainfall are also shown together with the monthly evolution of the components of the water balance for the case "aagg" (open

1 triangle) or "aaaa" (full square) averaging and the case where the soil hydraulic properties of the
2 surface were derived from the Clapp and Hornberger classification (open diamond). Case 2:
3 gravitational flow with saturated hydraulic conductivity of $2.78 \times 10^{-6} \text{ m s}^{-1}$ (second year).

4

5 Fig. 6: Monthly evolution of the mean value $\pm$ one standard deviation (error bars) of the 78 soil
6 profiles for total evapotranspiration, bare soil evaporation, transpiration and deep drainage (positive
7 for percolation and negative for capillary rises). Monthly values of rainfall are also shown together
8 with the monthly evolution of the components of the water balance for the case "aagg" (open
9 triangle) or "aaaa" (full square) averaging and the case where the soil hydraulic properties of the
10 surface were derived from the Clapp and Hornberger classification (open diamond). Case 3:
11 gravitational flow with saturated hydraulic conductivity of $2.78 \times 10^{-7} \text{ m s}^{-1}$ (third year).

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Table 1: Basic statistics of secondary soil hydraulic parameters for layer 1 (2-12 cm) and layer 2 (17-27 cm): minimum, maximum, mean, standard deviation, coefficient of variation CV, number of observations, n_{obs} , type of distribution tested with the χ^2 -test (N=normal, LN=lognormal), value of the probability p-test of the χ^2 -test. Values with * means rejection at the 10% level, ** means rejection at the 5% level. Values in brackets are the corresponding values for the logarithm of the variable.

Table 2: Components of the mean water budget (in mm) for the two/three years of simulation and the three lower boundary condition choices. Deep drainage was positive for percolation and negative for capillary rises. Total rainfall was 352 mm in all the cases and runoff was zero. Evaporation of intercepted rainfall was not given in this table but it was of the order of 7 mm in all the cases.

Table 3: Values of the parameters for the three 1-D cases. "aaaa" means arithmetic mean for all the four parameters. "aagg" means arithmetic mean for λ and θ_s and geometric mean for h_{bc} and K_s . and corresponds to the median value of the four parameters. "Mean CH" represents the Clapp and Hornberger derived parameters for the average soil texture over the whole area (class 3 for both layer).

Table 4: Case 1: constant matric potential at the bottom of the soil profile for the second year. Statistics of the annual water balance calculated by using measured surface hydraulic properties in the 78 soil profiles. Deep drainage was positive for percolation and negative for capillary rises. Mean values obtained through one 1D run using either the "aagg" or "aaaa" averaging or the soil surface hydraulic properties derived from the Clapp and Hornberger classification for the average soil texture are also given.

Table 5: Case 2: gravitational flow at the bottom of the soil profile with $K_s=2.78 \times 10^{-6} \text{ ms}^{-1}$ for the second year Statistics of the annual water balance calculated by using measured surface hydraulic

properties in the 78 soil profiles. Deep drainage was positive for percolation and negative for capillary rises. Mean values obtained through one 1D run using either the "aagg" or "aaaa" averaging or the soil surface hydraulic properties derived from the Clapp and Hornberger classification are also given.

Table 6: Case 3: gravitational flow at the bottom of the soil profile with $K_s=2.78 \times 10^{-7} \text{ ms}^{-1}$ for the third year. Statistics of the annual water balance calculated by using measured surface hydraulic properties in the 78 soil profiles. Deep drainage was positive for percolation and negative for capillary rises. Mean values obtained through one 1D run using either the "aagg" or "aaaa" averaging or the soil surface hydraulic properties derived from the Clapp and Hornberger classification for the average soil texture are also given.

Table 7: Case 1: constant matric potential at the bottom of the soil profile for the second year. Monthly coefficients of variation (%) of total evaporation (EVT) bare soil evaporation (BSE) and plant transpiration (TR).

Table 8: Case 2: gravitational flow at the bottom of the soil profile with $K_s=2.78 \times 10^{-6} \text{ ms}^{-1}$ for the second year. Monthly coefficients of variation (%) of total evaporation (EVT) bare soil evaporation (BSE) and plant transpiration (TR)).

Table 9: Case 3: gravitational flow at the bottom of the soil profile with $K_s=2.78 \times 10^{-7} \text{ ms}^{-1}$ third year. Monthly coefficients of variation (%) of total evaporation (EVT) bare soil evaporation (BSE) and plant transpiration (TR).

1 Table 1

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	Layer	Minimum	Maximum	Mean	Standard deviation	CV (%)	n_{obs}	Type of distribution	p-test value of the χ^2 -test
Dry bulk density ρ_d (g cm ⁻³)	1	0.94	1.56	1.27	0.16	12.5	100	N	0.10
	2	0.86	1.69	1.32	0.20	14.9	98	N	0.05*
Soil porosity ε	1	0.41	0.65	0.52	0.06	11.5	100	N	0.10
	2	0.36	0.68	0.50	0.07	14.8	98	N	0.05*
Saturated water content θ_s (cm ³ cm ⁻³)	1	0.39	0.63	0.50	0.06	11.6	100	N	0.19
	2	0.34	0.66	0.48	0.07	15.1	98	N	0.08*
Shape parameter of the particle size distribution M	1	.102	.196	.138	.018	13.0	100	N	0.47
	2	.100	.194	.136	.018	13.2	98	N	0.44
Shape parameter of the Brooks and Corey retention model λ	1	.119	.243	.165	.026	15.7	100	N	0.51
	2	.113	.269	.165	.027	16.5	98	N	0.42
Shape parameter of the Brooks and Corey model η	1	10.15	15.60	12.88	1.19	9.2	100	N	0.85
	2	9.61	16.13	12.90	1.27	9.8	98	N	0.06*
Scale parameter of the particle size distribution d_g (mm)	1	177.6	1774.0	629.9	272.3	43.2	100	LN	0.03**
	2	160.8	1526.0	660.7	287.8	43.5	98	LN	0.28
Scale parameter of the Brooks and Corey retention model $-h_{bc}$ (cm)	1	0.52 (-.66)	52.8 (3.97)	12.4 (2.12)	10.7 (0.96)	86.5 (45.1)	99 (99)	LN	0.36
	2	0.19 (-1.67)	76.8 (4.34)	20.0 (2.46)	17.7 (1.26)	88.8 (51.3)	93 (93)	LN	0.02**
Saturated hydraulic conductivity, K_s (cm s ⁻¹)	1	.00014 (-8.84)	.0075 (-4.89)	.00151 (-6.84)	.00129 (.883)	85.5 (12.9)	83 (83)	LN	0.51
	2	.00005 6 (-9.79)	.0083 (-4.78)	.00122 (-7.28)	.00149 (1.102)	122.3 (15.1)	79 (79)	LN	0.23

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1 Table 2

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	Total evaporation	Deep drainage	Change in water storage	Bare soil evaporation	Transpiration
First year constant h	574	-45	-177	336	231
Second year constant h	549	-192	-5	319	223
First year gravitational $K_s=2.78 \cdot 10^{-6} \text{ ms}^{-1}$	549	146	-343	334	208
Second year gravitational $K_s=2.78 \cdot 10^{-6} \text{ ms}^{-1}$	410	6	-64^(*)	303	97
First year gravitational $K_s=2.78 \cdot 10^{-7} \text{ ms}^{-1}$	564	35	-246	332	224
Second year gravitational $K_s=2.78 \cdot 10^{-7} \text{ ms}^{-1}$	424	12	-82	310	107
Third year gravitational $K_s=2.78 \cdot 10^{-7} \text{ ms}^{-1}$	378	4	-29	306	64

3 ^(*) Equilibrium was not reached at the end of the second year. However, runs on a third year were not possible because
 4 of a divergence of the model due to very small water content within the soil profile.

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6 Table 3

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	Layer	λ	$\theta_s \text{ (cm}^3 \text{ cm}^{-3}\text{)}$	$-h_{we} \text{ (cm)}$	$K_s \text{ (cm s}^{-1}\text{)}$
"aaaa"	1	0.163	0.504	11.8	0.00151
	2	0.162	0.486	11.7	0.00111
"aagg"	1	0.163	0.504	7.86	0.00105
	2	0.162	0.486	9.97	0.00066
"Mean CH"	1	0.204	0.435	21.8	0.00341
	2	0.204	0.435	21.8	0.00341

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Table 4

	Total evaporation	Deep drainage	Change in water storage	Bare soil evaporation	Transpiration
Mean	549	-192	-4	319	223
Median	553	-195	-4	322	224
Minimum	497	-139	-8	244	204
Maximum	571	-216	-1	359	246
Standard deviation	17	17	1	27	11
Coefficient of variation (%)	3	9	-36	9	5
1D run "aagg"	550	-193	-4	319	224
1D run "aaaa"	562	-204	-5	335	220
1D run "Mean CH"	569	-216	-1	358	204

Table 5

	Total evaporation	Deep drainage	Change in water storage	Bare soil evaporation	Transpiration
Mean	410	6	-63	303	97
Median	410	6	-63	305	94
Minimum	393	4	-80	240	46
Maximum	423	9	-46	339	175
Standard deviation	8	1	8	23	30
Coefficient of variation (%)	2	14	13	7	31
1D run "aagg"	403	6	-56	303	92
1D run "aaaa"	402	6	-55	323	72
1D run "Mean CH"	352	6	-59	302	43

1 Table 6

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	Total evaporation	Deep drainage	Change in water storage	Bare soil evaporation	Transpiration
Mean	378	4	-29	306	64
Median	378	4	-29	309	61
Minimum	374	4	-39	245	29
Maximum	388	6	-25	338	136
Standard deviation	3	0.4	3	21	24
Coefficient of variation (%)	1	8	10	7	37
1D run "aagg"	378	4	-29	307	63
1D run "aaaa"	376	4	-28	322	47
Mean Clapp and Hornberger	375	4	-25	339	28

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1 Table 7

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	J	F	M	A	M	J	J	A	S	O	N	D
<i>EVT</i>	12.7	17.3	7.7	10.2	17.1	3.2	3.3	6.6	4.1	1.0	9.9	8.2
<i>BSE</i>	12.9	17.8	8.1	10.8	32.6	8.9	7.5	9.9	3.4	2.8	12.2	8.6
<i>TR</i>	5.8	5.6	1.3	2.2	0.9	1.6	5.2	8.9	7.4	3.7	4.3	1.3

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6 Table 8

	J	F	M	A	M	J	J	A	S	O	N	D
<i>EVT</i>	12.8	17.6	5.9	10.7	11.1	12.5	32.8	27.8	6.9	2.3	6.5	3.7
<i>BSE</i>	13.0	18.2	6.2	11.3	22.2	4.6	4.3	6.4	2.4	1.8	11.9	5.8
<i>TR</i>	5.8	5.5	1.1	2.6	1.7	24.5	54.0	52.3	34.4	25.0	44.1	40.6

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10 Table 9

	J	F	M	A	M	J	J	A	S	O	N	D
<i>EVT</i>	12.0	16.7	5.3	8.4	13.4	21.4	29.1	15.8	4.3	1.8	9.9	7.1
<i>BSE</i>	12.2	17.5	5.7	9.0	20.8	3.5	3.6	6.0	2.2	1.9	12.8	7.9
<i>TR</i>	6.2	6.1	1.7	3.1	16.1	50.9	66.5	38.5	26.3	20.6	32.7	14.1

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