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Temporal dynamics of soil water balance components in a karst range in southeastern Spain: estimation of potential recharge

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Temporal dynamics of soil water balance components in a karst range in southeastern Spain: estimation of potential recharge

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Abstract This paper analyses the temporal dynamics of soil water balance components in a representative recharge area of the Sierra de Gádor (Almeria, southeastern Spain) in two hydrological years. Two approaches are used to estimate daily potential recharge (PR): Approach 1 based on deriving PR from the water balance as the difference between measurements of rainfall (P) and actual evapotranspiration (E) obtained by eddy covariance; and Approach 2 with PR obtained from the dynamic pattern of the soil moisture (θ) recorded at two depths in the site's thin soil (average 0.35 m thickess). For the hydrological year 2003/04, which was slightly drier than the 30-year average, E accounted for 64% of rainfall and occurred mainly in late spring and early summer. The PR estimated by Approach 1 was 181 ± 18 mm year^{-1} (36% of rainfall), suggesting an effective groundwater recharge in the study area. In the unusually dry hydrological year 2004/05, E was about 215 mm year^{-1}, close to the annual rainfall input, and allowing very little (8 ± 12 mm year^{-1}) PR according to Approach 1. Estimation of PR based on Approach 2 resulted in PR rates lower than those found by Approach 1, because Approach 2 does not take into account the recharge that occurs through preferential flow pathways (cracks, joints and fissures) which were not monitored with the θ probes. Moreover, using Approach 2, the PR estimates differed widely depending on the time scale considered: with daily mean θ data, PR estimation was lower, especially in late spring, while θ data at 30 min resolution yielded a more reliable prediction of the fraction of total PR resulting from the downward movement of soil water by gravity.

Key words potential recharge; evapotranspiration; soil moisture; soil water balance; Mediterranean; Sierra de Gádor

Dynamique temporelle des composantes du bilan hydrique du sol dans une chaine karstique du sud-est de l'Espagne: estimation de la recharge potentielle

Résumé Ce travail analyse la dynamique temporelle des composantes du bilan hydrique du sol dans une zone de recharge représentative de la Sierra de Gador (Almeria, SE de l'Espagne) pendant deux années hydrologiques. Deux approches ont été utilisées pour estimer la recharge potentielle journalière (RP): dans l'approche 1, RP est calculée à partir de la différence entre les mesures de pluie (P) et d'évapotranspiration réelle (E) obtenues par la méthode de la covariance turbulente; dans l'approche 2, RP est obtenue à partir de l'humidité du sol (θ) mesurée à deux profondeurs dans les sols peu profonds du site (épaisseur moyenne de 0.35 m). Pendant l'année hydrologique

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2003/04, légèrement plus sèche que la moyenne des 30 années précédentes, E atteint 64% des précipitations et se répartit essentiellement à la fin du printemps et au début de l'été. La RP estimée par l'approche 1 vaut 181 ± 18 mm an^{-1} (36% des précipitations), suggérant qu'il existe une recharge de nappe effective dans le site d'étude. Pendant l'année hydrologique 2004/2005, inhabituellement sèche, E vaut approximativement 215 mm an^{-1}, proche de la pluviosité annuelle, engendrant une faible RP (8 ± 12 mm an^{-1}) selon l'approche 1. L'approche 2 donne des estimations de RP inférieures à celles de l'approche 1, parce que l'approche 2 ne considére pas la recharge à travers les chemins d'écoulement préférentiels (fractures, joints et fissures) non mesurée par les sondes d'humidité du sol. De plus, l'estimation de RP avec l'approche 2 diffère largement selon l'échelle temporelle considérée: avec des donnés journalières moyennes de θ, l'estimation de RP est inférieures, en particulier à la fin du printemps, tandis que les donnés de θ de résolution 30 minutes engendrent une prévision plus fiable de la fraction de la RP liée à la circulation descendante gravitaire de l'eau du sol.

Mots clefs recharge potentielle; évapotranspiration; humidité du sol; bilan hydrique du sol; Méditerranée; Sierra de Gádor

1 INTRODUCTION

Apart from precipitation (P), evapotranspiration (E) is the largest component in the terrestrial hydrological budget (Brutsaert, 1982). At a given site, E depends largely on the availability of water and energy, and thus is site-specific. In Mediterranean regions, due to the scarcity of water resources, reliable evaluation of water lost through evapotranspiration is very important (Rana & Katerji, 2000). Despite the attention that these ecosystems are receiving (Detto *et al.*, 2006), knowledge of E rates in Mediterranean mountain areas continues to be poor. This is due to the difficulty of evaluating E accurately because all the weather variables range over very large intervals and any variation in one parameter immediately influences all the other variables that are mutually linked (Rana & Katerji, 2000), and also due to land cover heterogeneity. Its measurement becomes especially relevant in the recharge areas of high permeability karst landscapes where long-term surface runoff can be assumed negligible and potential recharge (PR) calculated as the difference between precipitation and evapotranspiration (Contreras *et al.*, 2008).

There are many sophisticated models and indirect methods for assessing PR, defined here as the downward drainage of atmospheric water and water percolation bypassing the soil root zone. Approaches that have been used successfully elsewhere include the soil water balance, chemical and isotopic mass balances, hydrodynamical methods based on Darcy's law, Richards' equation, and "tipping-bucket" models (Kendy *et al.*, 2003; Scanlon *et al.*, 2006). Often it is not easy to apply such models. In some cases this is because data are laborious and expensive to obtain (e.g. radioisotopes and artificial tracers in the vadose zone); in others it is due to the time-consuming and difficult nature of measuring functional relationships such as those needed to apply one-dimensional unsaturated flow models based on the Richards' flow equation, complicating the estimate of PR (Lerner *et al.*, 1990).

The water balance method is a widely-used, indirect technique for estimating PR (Fazal *et al.*, 2005). Water balance models have been developed on various time scales (e.g. hourly, daily, monthly and yearly) and with varying degrees of complexity (Xu & Singh, 1998). For a reliable estimate of PR via the water balance, all other components of the water balance (runoff, interception, evapotranspiration and infiltration) must be known. Since direct measurement of these components, in most cases, is not possible, they must be estimated, often from rainfall and other meteorological data (temperature, net radiation or saturation water vapour pressure deficit), soil properties and plant cover characteristics. Apart from the data requirement of water balance methods, large errors can occur when the accounting period is less than 10 days (Howard & Lloyd, 1979). Moreover, the soil water balance technique is more likely to fail in Mediterranean areas than in temperate environments because of strong seasonal contrasts in precipitation and potential evapotranspiration (Gee & Hillel, 1988; Allison *et al.*, 1994) and the heterogeneity of surface cover and soils.

In other cases, PR is estimated from soil water content (θ) change monitored using automated data logging systems (Thomsen & Hougård, 1995; Delin *et al.*, 2000, 2007; Delin & Herkelrath, 2005). Recently-developed non-destructive techniques for automatic high-frequency, volumetric measurement of soil water content, such as time-domain-reflectometry (TDR), make it possible to estimate PR more accurately, and to estimate and validate soil water balance model components (Ladekarl, 1999; Ladekarl *et al.*, 2005).

In many Mediterranean coastal regions, such as the semi-arid coastal fringe of southeast Spain, water demand has rapidly increased due to tourism and expansion of intensive agriculture, consequently increasing dependence on groundwater resources, especially when surface water resources fail. A good example is the Campo de Dalías coastal plain (Almería, SE Spain) whose deep aquifers, recharged through rainfall infiltration occurring in the Sierra de Gádor (or Gador mountain range) are overexploited for irrigation of over 26 000 ha of greenhouses (Campra *et al*., 2008) and the growing water demand by coastal tourism (Pulido-Bosch *et al*., 2000). Understanding the water balance and especially the groundwater recharge in these areas is critical to optimal management and sustainable development of groundwater resources (Taylor & Howard, 1996).

This work was carried out at the Llano de los Juanes field site (Sierra de Gádor, southeastern Spain), a relatively flat area at about 1600 m a.s.l., where fissured and fractured outcrops or denuded soil permit rapid infiltration, with a goal of estimating the PR and analysing the temporal dynamics of E, P, θ and PR during two contrasting hydrological years. Two methods for finding detailed temporal information on the PR were applied: Approach 1, based on continuous calculation of the water balance through the measurement of actual evapotranspiration (E) by the eddy covariance technique, and Approach 2 based solely on soil moisture (θ) measurements, recorded at two soil depths for one and a half years, for which all water in excess of that required to bring the soil to field capacity is assumed to be PR. This study has two main objectives: (a) to present valuable information on the temporal dynamics of the soil water balance components in this type of substrate, and (b) to estimate the potential groundwater recharge, which is critical to the optimal management and sustainable development of groundwater resources in such areas.

2 MATERIAL AND METHODS

2.1 Site information

Field studies have been conducted since September 2003 at Llano de los Juanes, an experimental research site located in the sector recognized by several authors (Vallejos *et al*., 1997; Vandenschrick *et al*., 2002; Contreras *et al*., 2008) as the main recharge area of the Sierra de Gádor. The Sierra de Gádor is located in southeastern Spain just west of the city of Almeria (Fig. 1). A mountain range with a maximum altitude of 2242 m a.s.l., it consists of a thick series of high permeability (fractures and fissures) Triassic

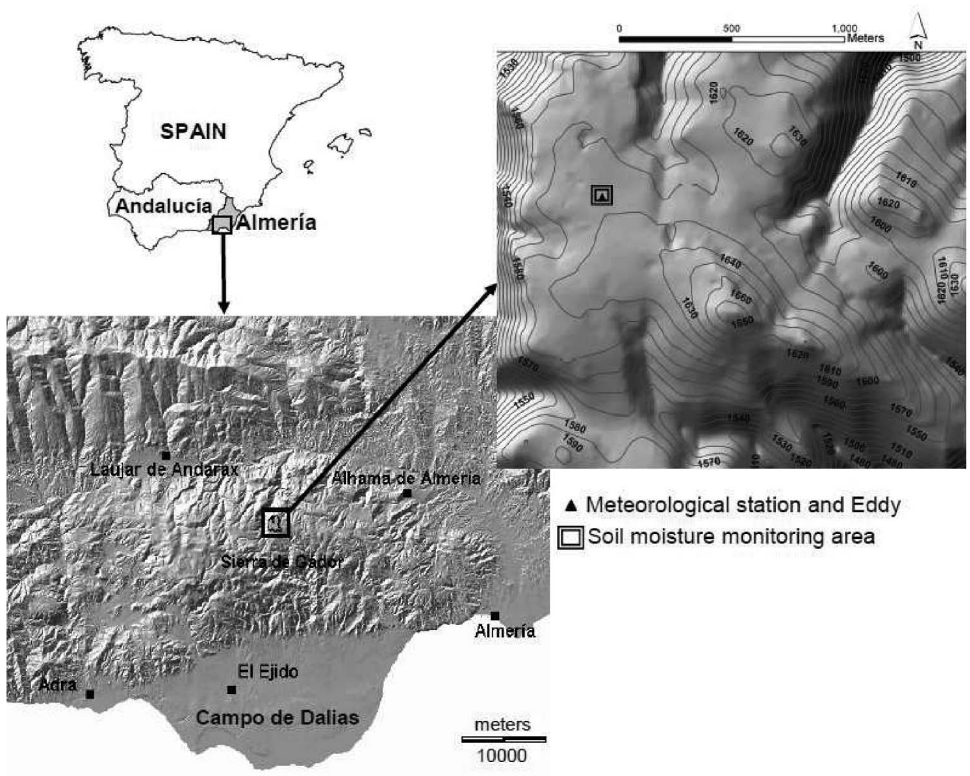

Fig. 1 Location of the study area. The location of the eddy covariance tower, soil moisture sensors and meteorological station is marked on the map compiled from a DEM (10-m resolution) of Llano de los Juanes.

limestones and dolomites with intercalated calcschists. Underlying these materials, low permeability Permian metapelites belonging to the Gádor and Félix tectonic units of the Alpujarride Complex appear (Pulido-Bosch *et al*., 2000; González-Asensio *et al*., 2003; Li *et al*., 2007). The Triassic carbonate aquifers continue under the Campo de Dalías.

In the Campo de Dalias, where about 70% of the total surface is covered by greenhouses, annual rainfall is about 200 mm and the groundwater pumping from the aquifer, 135×10^6 m^3 year^{-1} (González Asensio *et al*., 2003), represents more than twice the recharge estimated by several different water administration agencies (González Asensio *et al*., 2003). In recent years, several research papers concerning recharge of this aquifer system have been published (Pulido-Bosch *et al*., 2000; Vandenschrick *et al*., 2002; Alcalá *et al*., 2007; Frot *et al*., 2007; Contreras *et al*., 2008).

Analysis of the stable isotopic signature of the groundwater enabled identification of the most favourable sectors for rainfall infiltration and the altitudinal distribution of recharge rate (Vallejos *et al*., 1997; Vandenschrick *et al*., 2002; Alcalá *et al*., 2007). A relatively flat area at between 1200 and 1800 m a.s.l., with much karst development and denuded soils that permit rapid infiltration, was recognized as the main recharge area, in which Llano de los Juanes is located.

The climate of the Gador range varies from semi-arid to subhumid Mediterranean, with highly irregular rainfall. The climate has strong altitudinal gradients in annual precipitation and temperature. The mean annual precipitation is 260 mm in Alhama de Almeria (520 m a.s.l.) and ~650 mm on the summit (2246 m a.s.l.) (Contreras, 2006).

The 2-km^2 Llano de los Juanes experimental site is located at 1600 m a.s.l. A detailed description of the site is given by Li *et al*. (2007; 2008) and Contreras *et al*. (2008). It is a quite flat (Fig. 1), well-developed, fragmented limestone plateau (Li *et al*., 2007), rising slightly to the east. The soil is classified as a Lithic Haploxeroll–Lithic Ruptic Argixeroll complex (Oyonarte, 1992). Soils are thin (0.30–0.35 m) with scattered stones at the surface and throughout the profile, and are composed predominantly of silt (40–59%) and clay (21–68%) with clay illuviation in the subsurface horizon (Delgado *et al*., 2003). Soil organic matter content is relatively high, ranging from 1.3 to 7.3%. The mean bulk density is 1.1 g cm^{-3}, and the unsaturated hydraulic conductivity averages 1.5–1.6 mm h^{-1} and 12–194 mm h^{-1} for hydraulic heads of 120 mm and 30 mm, respectively. The soil surface is covered by a mosaic of patchy dwarf perennial shrubs and grasses, rock outcrops and bare soil with patches of rock fragments interspersed with vegetation. *Genista pumila* (Vierh) ssp. *pumila* predominates among woody shrubs and *Festuca scariosa* (Lag.) Hackel in the herbaceous stratum.

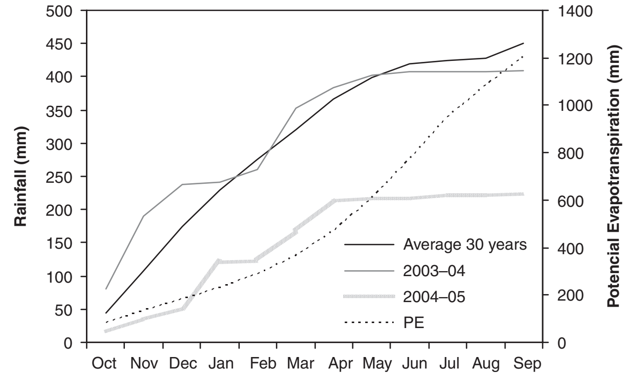

Fig. 2 Comparison of cumulative precipitation during the study period (hydrological years 2003/04 and 2004/05) with the last 30 years average, and the cumulative average potential evapotranspiration (PE), obtained by applying the Hargreaves & Samani (1982) method with interpolation (Contreras, 2006).

La Zarba, the closest long-term weather station (10.4 km from the field site), at 1219 m a.s.l, has recorded a mean annual rainfall of 463 mm over the last 30 years, with a high inter-annual coefficient of variation (38.3%). The highest total annual rainfall recorded was 1289 mm in hydrological year 1962–1963 and the lowest was 179.5 mm in 1980–1981. The hydrological year 2003/04, analysed at the experimental site for this study, was slightly drier than the 30-year average at the La Zarba station and 2004/05 was unusually dry (Fig. 2).

The mean annual precipitation at Llano de los Juanes is 537.5 mm, determined by multiple regression of data from 20 nearby stations over a 30-year period (Contreras, 2006), though the recorded mean in the 3 years from 1 September 2003 to 31 August 2006 was only 378 mm year^{-1}. Rainfall intensities recorded during the study period in Llano de los Juanes were generally low, with a maximum rainfall intensity in 5 min (I_{5min}) of 78.1 mm h^{-1}. Only 12% of the total recorded rains exceeded 1 mm, and had a maximum I_{5min} greater than 15 mm h^{-1}.

2.2 Field measurements

Evapotranspiration was measured by an eddy covariance system (Anderson *et al*., 1984; Verma *et al*., 1986). The eddy covariance system measured the turbulent fluxes of water vapour (E, or its equivalent the latent heat flux, LE), temperature (sensible heat flux, H) and momentum. The eddy covariance system included a three-dimensional sonic anemometer (CSAT3, Campbell Scientific Inc., USA) for measuring wind speed in three dimensions and a krypton hygrometer KH20 (CSAT3, Campbell Scientific Inc., USA) for measuring water vapour density fluctuations. Air temperature and humidity were measured by a thermo-hygrometer (HMP 35C, CSI, USA). Data were acquired and stored by a datalogger (CR23X, CSI). Means, variances, and co-variances of 10-Hz data were calculated and stored every 30 min. Eddy flux corrections for density perturbations (Webb *et al*., 1980) and coordinate rotation (McMillen, 1988; Kowalski *et al*., 1997) were carried out in post-processing. Hygrometer measurements were corrected for absorption of radiation by oxygen according to Tanner *et al*. (1993).

Net radiation (Rn) was measured with a radiometer (NR Lite, Kipp and Zonen, Delft, The Netherlands). The soil heat flux (G) was calculated as the sum of the average flux measured with two soil heat flux plates (HFT-3, REBS, Seattle, Washington, USA) at a depth of 0.08 m, and the heat stored in the layer of soil above the plates (Fuchs, 1986; Massman, 1992).

The amount and intensity of rainfall were also recorded at the site by an automatic 0.2-mm-resolution tipping-bucket raingauge (model 785M, Davis Instruments Corp., Hayward, California, USA). Llano de los Juanes occasionally receives some precipitation as snow that can saturate the raingauge. In order to determine the error caused and the proportion of water input as snow, we collected snowfall with a petrol barrel (0.57-m diameter and 1.5-m high) open at the top. All snow was collected after each event, melted and measured. During the 2003/04 hydrological year there were four snowfall events which represented less than 5% of the rainfall of that year and a negligible difference with precipitation measured by the gauge. During 2004/05 there was one snowfall event, and the cold weather maintained the snow cover for several weeks, but the total water volume was nonetheless negligible in comparison with total rainfall.

Soil moisture was measured every 30 min using capacitive probes, SBIB (self-balanced impedance bridge) (Vidal, 1994; Vidal *et al*., 1996). The technical characteristics of SBIB probes, their calibration as well as successful results with replicates of the same prototype used in different soils, are detailed in different publications (Puigdefábregas *et al*., 1996, 1998, 1999; Vidal *et al*., 1996; Domingo *et al*., 1999, 2001; Cantón *et al*., 2001, 2004). Beginning in July 2003, six probes were installed in a 10 × 10 m area the eddy tower at a depth of 0.06 m. Moreover, at the end of 2003, six additional θ probes were installed in the same place at two depths (three at a depth of 0.06 m and three more at 0.25 m depth), such that θ data at two depths are available only from 12 February 2004 to 31 August 2005. The soil depths for measuring θ were selected to monitor the two soil horizons described at the soil site (A and Bt). No probes could be installed any deeper because the soils are very shallow (0.30–0.35 m) and contain some rock fragments. Soil moisture data are expressed as a volumetric percentage (%vol.). The soil in which the probes were installed consisted of 25.3 ± 3.8% clay and 54.9 ± 3.4% silt, with a gravel content of, on average, 22 ± 4% in the A horizon, and 53.9 ± 1% clay, 32.9 ± 1.5% silt and a mean gravel content of 7 ± 0.3% in the Bt horizon. Medium and fine roots are common in the A horizon but rare in the Bt horizon. The drainage class is well drained. The bulk density of the A horizon is 1.05 ± 0.01 g cm^{-3}.

2.3 Evapotranspiration data processing

Data filtering of the E measurements was performed in order to eliminate errors in measured fluxes. First, to avoid data measured in periods with low turbulence, when eddy covariance measurements are not reliable, a friction velocity threshold of 0.2 m s^{-1} was established, below which data were discarded. The friction velocity is calculated from covariances of the components of the wind speed measured by the sonic anemometer. Moreover, night-time data, with Rn lower than 10 W m^{-2} (to include dusk and dawn moments), were also eliminated as turbulent transport is diminished during night-time stable atmospheric conditions. Although this filtering accounted for most of the error peaks found in the E data, we also eliminated LE data greater than 800 W m^{-2}, according to the limit values proposed by CARBOEUROPE (European network of eddy covariance systems).

In order to ensure that the measured E data were accurate, different analyses were made. First, an analysis of the energy balance closure for the period studied was made. The energy balance closure at the ecosystem scale can be considered when the available energy (Rn − G) equals the sum of the latent and sensible heat fluxes (LE + H). The regression between (Rn − G) and (LE + H) indicated that for our data, turbulent fluxes accounted for 79% of the available energy ((LE + H) = 0.79(Rn − G), $R^2 = 0.86$, $n = 10\ 890$). Considering that with eddy covariance measurements a 100% energy balance closure is never reached (Baldocchi, 2003), our data showed an energy balance closure that is in the range reported by most FLUXNET sites (Wilson *et al.*, 2002). Moreover, we performed a footprint analysis to verify that the source area of the measured fluxes originated from well within the eddy covariance fetch. Using the Flux-Source Area model (Schmid, 1994) we estimated that the maximum up-wind distance from the eddy covariance of the point of maximum source was 27 m, and the maximum distance to the source areas far end was 114 m. The study site includes several hundred metres of homogeneous surface upwind of the eddy covariance tower in every direction, and the footprint analysis showed that the source area of the measurements was within that homogeneous surface.

Due to data filtering and measurement failures, approximately 60% of data were missing from our 30-min E database. For the purpose of this work, we did not consider night-time data as its contribution to the annual water balance is negligible. The resulting diurnal E database had 10% of data missing. Gap filling was done according to the energy balance closure (LE = 0.79(Rn − G) − H). Gap filling of Rn, G and H was done by averaging the immediate neighbouring data. Where E gaps were of several days, as E will be used on a daily to annual time scale, the daily gaps were filled by averaging a time interval of 4 to 7 days (Falge *et al.*, 2001).

A comparison of the E data from two eddy covariance towers separated by 20 m at the same field site showed differences of around 0.083 mm d^{-1} (RMSE) and 1% (Bias) (Were *et al.*, 2009).

2.4 Potential recharge estimation

2.4.1 Approach 1 The principle of PR estimation by the soil water balance method is that a soil becomes free-draining when the moisture content of the soil reaches the field capacity. Excess water then drains through the soil, becoming potential recharge below the root zone. To determine when the soil reaches this field capacity, daily soil moisture conditions throughout the year must be simulated (Rushton *et al.*, 2006). This value is found from E, P and runoff (R) field data, and it is unnecessary to measure θ. Approach 1 was applied in daily time steps to estimate PR_t (PR on day t) in Llano de los Juanes in hydrological years 2003/04 and 2004/05 as follows:

$$\begin{aligned} &\mathrm{PR}_t = \theta_t^d\ (\text{for } \theta_t^d < 0) \quad \text{and} \\ &\mathrm{PR}_t = 0\ (\text{for } \theta_t^d \geq 0) \end{aligned} \tag{1}$$

$$\theta_t^d = \theta_{t-1}^d - P_t + E_t - R\_on_t + R\_off_t \tag{2}$$

where θ_t^d is the daily soil moisture deficit on day t, i.e. θ_t^d is the depth of water required to bring the soil to field capacity. This concept of a soil moisture deficit is useful for calculations; however, it makes no assumptions about the variation in moisture content with depth (Rushton *et al.*, 2006). P_t is daily precipitation on day t and E_t is daily evapotranspiration on day t. R_on_t is daily runon and R_off_t is daily runoff by overland or subsurface flow on day t. The slope gradient at the study site is very slight, favouring infiltration, and according to Li *et al.* (2008, 2010) infiltration rates are very high, while both R_on and R_off are negligible. The variable θ_{t-1}^d is θ_t^d on day $t-1$. The initial θ_{t-1}^d value (corresponding to 30 September 2003) was found by calibrating the model (Approach 1), i.e. running the model until the sum of daily PR obtained by Approach 1 coincided with the annual PR obtained as the residual between annual rainfall and annual E. This calibrated initial θ_{t-1}^d represents the depth of water required to bring the soil (from a soil water content of about 0%) to field capacity. The initial θ_{t-1}^d obtained as explained above was 131 mm.

Potential recharge is assumed to occur when θ_t^d is negative. As θ_t^d is reduced to zero, the soil reaches field capacity and becomes free-draining. Consequently PR is the water in excess of the amount required to bring the soil to field capacity.

2.4.2 Approach 2 Potential recharge is estimated with daily mean field θ data assuming that soil becomes free-draining when θ reaches the limiting value which is the field capacity (θ_{FC}). The soil moisture in the soil profile at time t, expressed in mm of water and taking into account θ field data at both depths, and averaging all probes at each depth is $\theta(t)$, which was calculated for the soil depth, using data from both horizon A (0 ± 0.12 m) and horizon Bt (0.12 ± 0.35 m) based on the sensor readings and using the following:

$$\theta(t) = \theta_{h1} z_{h1} 10 + \theta_{h2} z_{h2} 10 \tag{3}$$

where θ_{h1} and θ_{h2} are the average measured water content (%vol.) at time t in the A and Bt horizons respectively, and z_{h1} and z_{h2} are the thickness (in m) of the A and Bt horizon, respectively.

If $\theta(t) > \theta_{FC}$ then PR $= \theta(t) - \theta_{FC}$ and $\theta(t)$ is set to θ_{FC} until the end of the recharge event, where θ_{FC} is also expressed in mm. This approach was implemented at two time resolutions: Approach 2a at the daily scale, using daily mean θ, and Approach 2b at an event scale, examining θ field data at 30-min intervals (θ_{30min}) during and after rainfall events, with recharge events defined as beginning when θ reached θ_{FC} and lasting until it decreased below θ_{FC}. The maximum θ_{30min} recorded in the soil profile during the recharge event was used to estimate PR as the residual of this value and θ_{FC}.

Field capacity (θ_{FC}) was determined from θ records during wet periods after selected rainfall events, considering values recorded at night and from 24 h after soil saturation. The selected rainfalls were those following periods with low air temperature and a pressure vapour deficit close to zero. Night records of θ_{30min} after those rainfall events were examined until drainage had decreased such that θ was steady and the rate of change of θ was less than 0.001 $m^3\ m^{-3}\ 30\text{-}min^{-1}$, using the Bruno *et al*. (2006) procedure. Consequently, θ_{FC} is not a daily mean, but rather a θ_{30min} value which remained steady for at least one night. In addition, θ_{FC} was determined in the laboratory from fine-earth water retention at −33 kPa in Richards plates.

As θ at two depths (0.06 m and 0.25 m) was available only from 12 February 2004 to 31 August 2005, the PR estimations using Approach 2 (both 2a and 2b) were only calculated for this period. Approach 2 is a single-store approach, because it does not take into consideration the transfer of water between individual soil layers.

2.4.3 Sensitivity analysis A sensitivity analysis was performed to evaluate changes in PR from uncertainties in E and θ field measurements due to their spatial variability. Rainfall uncertainty was not included in the analysis because the area is relatively small (<1 km^2) and flat, i.e. there are no orographic variations. During the assessment, one parameter was varied at a time. For the analysis, E was varied within the RMSE (0.083 mm d^{-1}) value obtained from the comparison of the E data from two eddy covariance towers (see Section 2.3). Regarding θ, for Approach 2a we took into account the daily standard deviation of daily θ which varied between 0.1 and 5.2 %vol. at 0.06 m depth, and 0.1 and 4.4%vol. at 0.25 m depth. For each day of the study period, 12 February 2004 to 31 August 2005, at both depths, the corresponding daily standard deviation, obtained from these θ records was added, or subtracted when examining negative deviations, to each daily mean θ (obtained as an average of the available θ data from the soil water probes at each depth, nine at 0.06 m and three at 0.25 m). For Approach 2b, the same calculations were performed but using the mean maximum θ_{30min} of each recharge event and ranging within ± one standard deviation.

3 RESULTS

3.1 Temporal water balance dynamics

The annual variation in the water balance is summarized in Table 1, which shows monthly totals and means for hydrological years 2003/04 and 2004/05. A strong seasonal variation in E was observed at Llano de los Juanes. The monthly E budgets in Table 1 show that the monthly rates varied by one order of magnitude, from 6.9 mm $month^{-1}$ in January 2005 to 66 mm $month^{-1}$ in June 2004. The maximum monthly E rate did not coincide with the maximum monthly rainfall, as expected in a climate with strong seasonal contrast. In fact, that month (June 2004) total rainfall was only 6.3 mm. The highest E rates were measured in late spring and the first half of summer, when little soil water is stored (less than 10%), but the radiative energy input is very high. May to July recorded 43% of the total E in hydrological year 2003/04, which was characterized by a rainy spring, and 45% in 2004/05, when annual rainfall was less than half and the spring was dry. The highest monthly E recorded was 66 mm (2003/04) in June and 39.4 mm in May (2004/05). When these rates are examined on a weekly time scale, the high E rates recorded in summer stand out in both years; the second highest weekly E rate, i.e. 14 mm $week^{-1}$ in 2003/04 and 10 mm $week^{-1}$ in 2004/05, was measured at the beginning of July. The maximum daily E rates for the respective years were 3 mm d^{-1} on 30 June 2004 and 2.8 mm d^{-1} on 26 June 2005. Annual cumulative E in 2003/04 was 326.5 mm with a mean of 0.89 mm d^{-1}, and 214.9 mm with a mean of 0.59 mm d^{-1} in 2004/05. The annual E was 64% of total rainfall in the wet year monitored (2003/04), whereas in the drier year, annual E was equivalent to the total annual rainfall.

Table 1 Monthly distribution and annual budget of rainfall (P), measured evapotranspiration (E) and average soil moisture for hydrological years 2003/04 and 2004/05 at Llano de los Juanes.

	2003/04			2004/05		
	P ($mm\ month^{-1}$)	E ($mm\ month^{-1}$)	Soil moisture (%vol.)	P ($mm\ month^{-1}$)	E ($mm\ month^{-1}$)	Soil moisture (%vol.)
October	80.55	24.42	23.58	17.70	11.47	3.84
November	109.84	19.16	30.42	16.27	8.60	11.87
December	47.60	16.40	34.16	70.17	12.91	22.27
January	2.85	13.43	15.40	0.81	6.85	12.19
February	19.53	20.49	18.53	45.77	18.54	20.38
March	92.14	23.22	22.49	45.97	11.05	24.77
April	88.68	41.41	22.92	3.66	22.41	13.75
May	50.65	35.11	18.84	0.00	39.36	6.21
June	6.31	65.99	7.59	5.09	34.49	4.65
July	6.51	38.36	3.58	0.20	23.57	3.68
August	0.41	17.57	2.85	1.22	16.27	2.89
September	2.24	10.92	3.28	3.05	9.39	3.16
Annual total ($mm\ year^{-1}$)	507.28	326.49		209.91	214.92	
Daily mean		0.89 ($mm\ d^{-1}$)	16.97 %vol.		0.59 ($mm\ d^{-1}$)	10.75 %vol.

The daily θ pattern in the study area is shown in Fig. 3 along with rainfall and E during hydrological years 2003/04 and 2004/05, where θ at the surface horizon is observed to remain quite high throughout the autumn and spring, and dry up quickly after May. In the wetter hydrological year (2003/04), for 108 days mean θ (at 0.06 m depth) corresponded to a water tension of between −33 kPa and −1500 kPa, indicating that there was no significant gravitational water or, if there was, it was only for short periods of less than one day. On 31 days, daily mean θ (at 0.06 m) was wetter than θ_{FC}, i.e. the soil water content at −33 kPa (θ_{FC} at 0.06 m was 32% in the field and 36% from laboratory fine earth data), and gravitational water movement from the surface horizon was then possible. However, at 0.25 m depth, for the period from 12 February to 30 September 2004, only 4 days exhibited mean θ greater than θ_{FC} at 0.25 m (23.5 %vol.

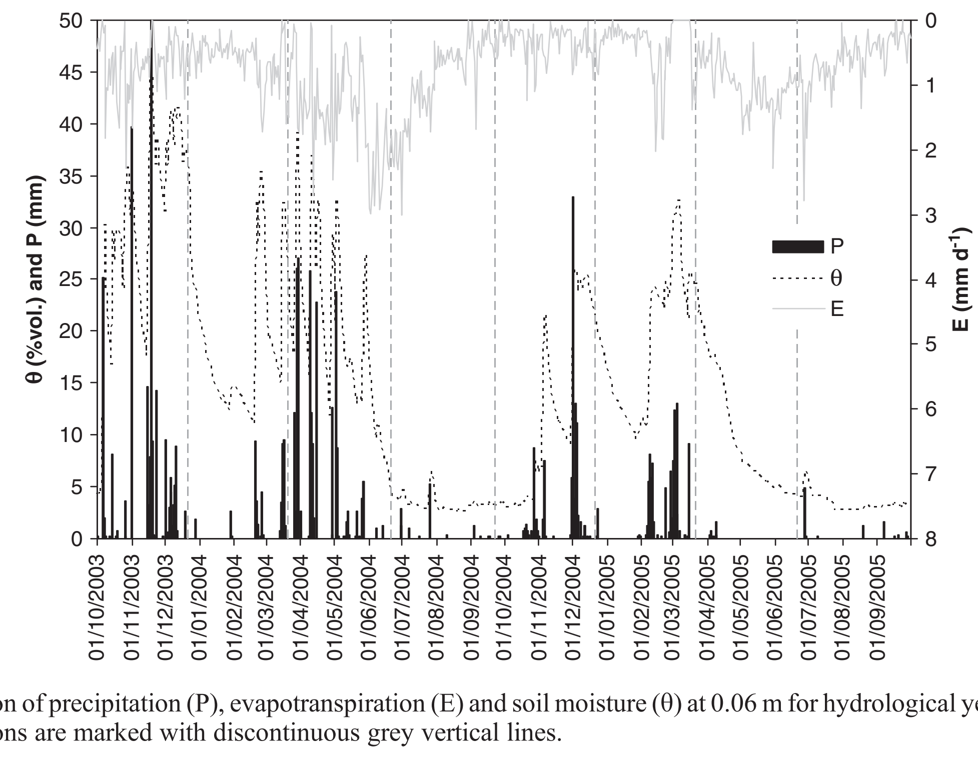

Fig. 3 Evolution of precipitation (P), evapotranspiration (E) and soil moisture (θ) at 0.06 m for hydrological years 2003/04 and 2004/05. Seasons are marked with discontinuous grey vertical lines.

estimated from θ field data, and 29.6 %vol. from laboratory fine earth data). In the 2004/05 hydrological year, daily mean θ at both depths (0.06 and 0.25 m) never exceeded the soil water content at −33 kPa, suggesting that deep percolation, if any, was limited to short periods of hours or less. Moreover, the availability of water was lower that year because the daily average θ corresponded to water tension of between −33 kPa and −1500 kPa on 71 days. The 2003/04 hydrological year had little rainfall all winter, explaining the drier regime that season, in which daily θ at 0.06 m depth was less than the permanent wilting point on 61 days, whereas in the 2004/05 hydrological year, even though it was much drier (507 mm year^{-1} in 2003/04 *vs* 209 mm year^{-1} in 2004/05), daily θ in the surface horizon was less than the permanent wilting point for only 48 days in winter. The average daily θ for hydrological year 2004/05 was 17 %vol., which is quite high (*versus* 10.8 %vol. in 2004/05). The standard deviation of daily θ at 0.06 m ranges from 0.1 to 5.2 %vol., with a mean standard deviation of 2.3 %vol., and at 0.25 m depth ranges from 0.1 to 4.4 %vol. with 2.0 %vol. mean standard deviation.

3.2 Estimation of potential recharge

3.2.1 Approach 1 In hydrological year 2003/04, the total PR, as calculated by Approach 1 (using rainfall and E field data), delivered 181 mm year^{-1}, 36% of total annual rainfall. The 2004/05 hydrological year was unusually dry (Fig. 2) with an annual rainfall of only 210 mm, and E exceeded 100% of the total rainfall, so we can infer that PR is very low or non existent, and in fact it was only 8 mm year^{-1} when calculated by Approach 1. Figure 4 shows the cumulative daily PR estimated by Approach 1 for both hydrological years.

3.2.2 Approach 2 Applying Approach 2a (based on daily mean field θ data exclusively) for the period with θ data available at two depths for the first hydrological year (12 February 2004 – 30 September 2004), the estimated PR was 18 mm, substantially lower than the 126 mm estimated by Approach 1 for this period. For hydrological year 2004/05, Approach 2a estimated that PR was equal to zero.

When Approach 2 is implemented at the rainfall-event scale (Approach 2b) using the maximum θ in 30 min (θ_{30min}) instead of considering daily mean values, the predicted PR is 57 mm for the period from 12 February 2004 to 30 September 2004, greater than the PR estimated by Approach 2a for the same period but still lower than the 126 mm estimated by Approach 1 for that period. For the hydrological year 2004/05, Approach 2b estimated 2 mm, while Approach 2a yielded 0 mm and Approach 1 estimated 8 mm. Using Approach 2b, PR represents 22% of the rainfall for the period from 12 February to 30 September with six recharge events, a number coinciding with the recharge events counted by Approach 1. Figure 5 compares the cumulative PR estimated by both methods, for the available period during hydrological year 2003/04, distinguishing between Approach 2a and 2b. It can be observed that Approach 2a always predicts lower PR and the underestimation is enhanced, especially in late spring, predicting in some cases no PR after large rainfalls which actually yielded PR.

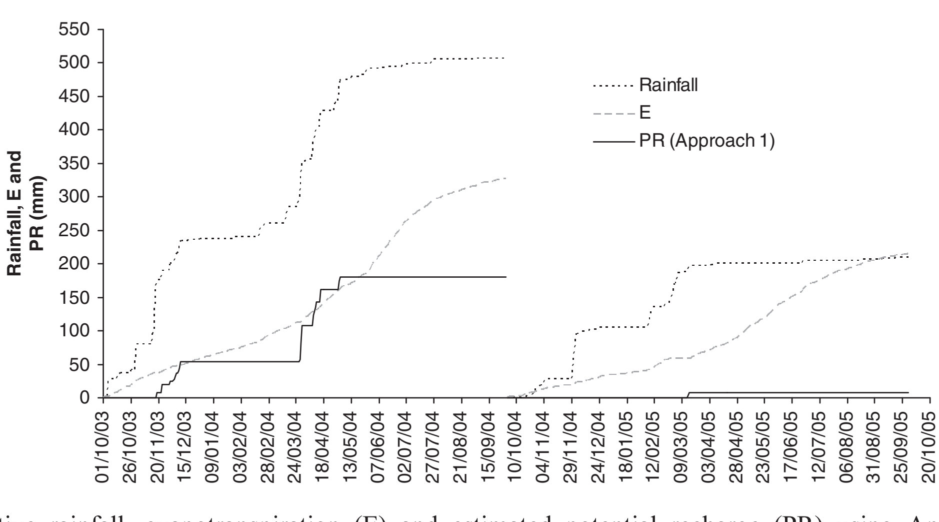

Fig. 4 Cumulative rainfall, evapotranspiration (E) and estimated potential recharge (PR) using Approach 1 for the hydrological years 2003/04 and 2004/05.

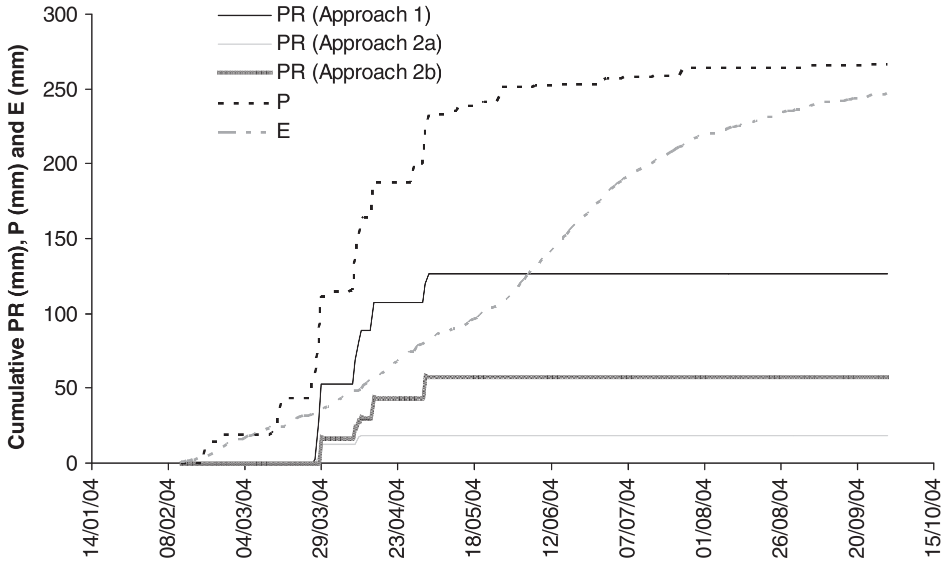

Fig. 5 Cumulative rainfall (P), evapotranspiration (E) and estimated potential recharge (PR) using approaches 1, 2a and 2b for the period 12 February–30 September 2004. Cumulative P – E appears smaller than cumulative PR (Approach 1) because of the effect of rainfalls occurring earlier in the season (i.e. autumn 2003 and early winter 2003/04).

Table 2 shows the different PR amounts estimated with approaches 1, 2a and 2b, for three representative events during the spring of 2004. As can be observed in Table 2, the use of daily mean θ yields a lower PR estimation, especially in late spring. Only three recharge events were estimated using Approach 2a θ (Fig. 5), all in early spring.

3.2.3 Sensivity analysis Results of the sensitivity analysis (Table 3) shows that PR estimation using Approach 1 ranged between 32% and 39% of rainfall for the hydrological year 2003/04 and between 0% and 4% for 2004/05. However, PR estimations using Approach 2 were affected by a much higher uncertainty whether using 2a daily mean θ, or 2b maximum θ_{30min}.

Table 2 Values of potential recharge (expressed in mm) estimated by approaches 1 and 2 for three rainfall events during the spring of 2004.

Event	Rainfall (mm)	Approach 1 PR (mm)	Approach 2a PR (mm)	Approach 2b PR (mm)
27–29 March 2004	67.9	53.03 ± 5.4	12.86 ± 8.1	16.44 ± 10.8
11 April 2004	9.2	8.06 ± 0.2	0.89 ± 4.1	2.67 ± 2.5
2–3 May 2004	32.5	19.16 ± 1.5	0 ± 0	13.65 ± 2.5

Note: Mean daily θ data and maximum θ_{30min} were used to estimate PR when implementing approaches 2a and 2b, respectively.

Table 3 Sensitivity analysis for changes in E and θ spatial variability. PR estimates are expressed in mm and the percentage of total rainfall for the period considered appears in brackets.

		Hydrological year		Period
		2003/04	2004/05	12 February–30 September 2004
E	E + RMSE	163.1 (32)	0 (0)	114.7 (43)
	E – RMSE	198.2 (39)	19.9 (4)	138.4 (52)
Daily θ	θ + SD		40.8 (19)	67.2 (25)
	θ – SD		0 (0)	3.4 (1.3)
Maximum θ_{30min}	θ_{30min} + SD		5.7 (3)	87.3 (33)
	θ_{30min} – SD		0 (0)	32.2 (12)

Note: The changes in annual PR for hydrological year 2003/04 cannot be processed with Approach 2, because θ at two depths are not available for the entire period.

Approach 2a showed high uncertainty, with changes of 20% for both hydrological years in the percentage of rainfall that PR represented; the prediction of Approach 2b had lower uncertainty in the drier year (2004/05).

4 DISCUSSION

4.1 Dynamics of water balance components

The Sierra de Gádor mountain range is the main recharge area of the Sierra de Gádor-Campo de Dalías aquifer system. In this coastal region, temporal water balance patterns show a high inter-annual rainfall variability of around 30–40%, similar to that observed in other Mediterranean mountains (Farmer *et al.*, 2003; Wilcox *et al.*, 2003; Scanlon *et al.*, 2006), and the variability in rainfall is transferred to the rest of the water balance components. Although all the water balance components' patterns are typically Mediterranean, the high E rates recorded in summer are striking. From 1 June to 31 July 2004, E was 32% of the annual total, even though θ for the total soil profile (0.35 m) was low, suggesting that most of the E is from transpiration. Thus, plants compensate for the lack of water in shallow soil layers by extracting water via their roots from cracks, fissures and pipes that store soil deposits in the carbonate bedrock, during dry periods. Mean θ at 0.06 m during this period was 5.5 %vol., and only exceeded 10 %vol. on the first five days in June, and at 0.25 m only reached 9 %vol. In nearby areas with similar climate, the maximum E occurs in May (Domingo *et al.*, 1999). The seasonal pattern of variation in θ, high in spring and late autumn and characteristically low in summer when plants actively take up soil water, is well known (Henninger *et al.*, 1976; Ehrenfeld *et al.*, 1997; Gehrels *et al.*, 1998).

4.2 PR estimation and its uncertainty

The soil water balance method is one of the most widely-used, indirect procedures for estimating PR, and more often than not, the water balance components are estimated rather than measured. In this study, the two complementary procedures applied to estimate daily PR use direct daily field measurement of the water balance components (P, E and θ), which is quite unusual, given the complications of measuring these components.

4.2.1 ***Approach 1*** PR estimated by Approach 1, in the unusually dry hydrological year 2004/05, was very low, 8 mm year^{-1}, and in hydrological year 2003/04 PR was 181 mm year^{-1}, accounting for 36% of rainfall, and suggesting an effective recharge to the aquifer. Considering that the hydrological year was slightly drier than the 30-year average (at the closest La Zarba weather station, Fig. 2) we can assume that PR represents nearly 40% of total precipitation in most years. This coincides with the results found for the same area by Contreras *et al.* (2008), who applied a water balance method based on the annual hydrological equilibrium hypothesis, and by Alcalá *et al.* (2007) who applied an atmospheric chloride mass balance, sampling the chloride content of recharge in local springs. Both estimates assumed negligible runoff, an assumption later supported by Frot *et al.* (2008) by monitoring water harvesting systems. In hilly aquifers, like Sierra de Gádor, effective or real aquifer recharge may be less than PR due to losses induced by surface seepage, groundwater pumping in shallow perched aquifers, diversions of interflow to harvesting systems, lateral transfers to rivers and other aquifers, etc. In Sierra de Gádor, there are small areas of surface seepage and local springs (Vallejos *et al.*, 1997; Alcalá *et al.*, 2007), as well as, evidence of diffuse lateral aquifer discharges to the Adra and Andarax river valleys (García-López *et al.*, 2009). Lowered piezometric levels, induced by pumping, complicate assessment of the effective aquifer recharge through simple equal-discharge spring flow methods or hydrodynamical techniques based on water-table fluctuations. Measured effective aquifer recharge data suitable for validating our results are not available, and no groundwater numerical modelling has been performed to validate potential *vs* effective aquifer recharge functions.

The sensitivity analysis showed that differences in E measurements can lead to PR estimate errors (Approach 1) of about ±4%.

4.2.2 Approach 2 Approach 2 is strongly affected by the time scale. When applied at a daily time step (Approach 2a), before rainfall starts, θ could be relatively low and so the daily mean θ remains below θ_{FC} for the first day of the rainfall event and no PR is predicted by Approach 2a. The use of daily mean θ limits the prediction, not only for the first day of the recharge event, but also for recharge events occurring in spring when soil water depletion is faster than in autumn and winter (Fig. 3). This is due to the higher temperatures and E rates recorded in spring, as shown in Fig. 6, which compares soil water depletion and cumulative E in spring and autumn for two similar events (23 mm of rainfall and similar antecedent θ). Daily mean θ is therefore lower in spring. After some rainfall events lasting several hours, θ exceeded θ_{FC}, and deep drainage occurred for several hours; however, the soil dried out rapidly and daily mean θ dropped to below θ_{FC} so no PR was calculated using Approach 2a.

The PRs estimated by approaches 1 and 2 were quite different for the study period during hydrological year 2003/04. Approach 2 (both 2a and 2b) estimated lower PR than Approach 1, probably because the PR estimated from Approach 2 does not account for recharge by preferential flow through cracks, joints and fissures, which is not monitored by θ probes, but which is detected at the plot scale (Approach 1). For Approach 1, E is recorded by eddy covariance in a plot of shrubs representative of the study site ecosystem which integrates all the spatial heterogeneity, while for Approach 2, θ is measured locally. Nevertheless, approaches 2a and 2b use mean values for θ (daily or maximum θ_{30min}) from nine probes at 0.06 m depth, the horizon with greater variability, and three at 0.25 m depth; although these values may be considered as representative of the plot, Approach 2 cannot measure the occurrence of preferential vertical flows. Therefore, Approach 1 can be regarded as a plot-scale estimate of total PR, while Approach 2 is more representative of diffuse PR. The approaches can be considered as complementary but neither can validate the other. The differences between the approaches may be accentuated in fine textured soils under very dry conditions, when shrinkage cracks can be abundant and the proportion of preferential flow in the total PR can be especially high (Keese *et al*., 2005).

Moreover, approaches 2 (2a and 2b) are unable to detect continuing recharge when measurable θ changes are absent. Once θ_{FC} is exceeded, the soil profile probably continues to drain slowly, because of the lower saturated hydraulic conductivity of the Bt horizon, though this slow drainage can still generate PR, as found by Kendy *et al*. (2003) for silty clay loam and clay loam horizons in deep soil in the North China Plateau. As a consequence, PR estimations based on θ data are expected to be lower than the total PR estimated by Approach 1. Furthermore, another factor must be considered to understand the results of Approach 2: in spring Approach 2a underestimates PR compared to Approach 1 (Table 2) because its temporal resolution is insufficient to track the rapid soil water loss favoured by higher temperatures and E rates recorded in spring (Fig. 3 and 6), missing hourly-scale events when θ exceeds θ_{FC} following rainfall, and so not yielding any PR. The PR underestimation

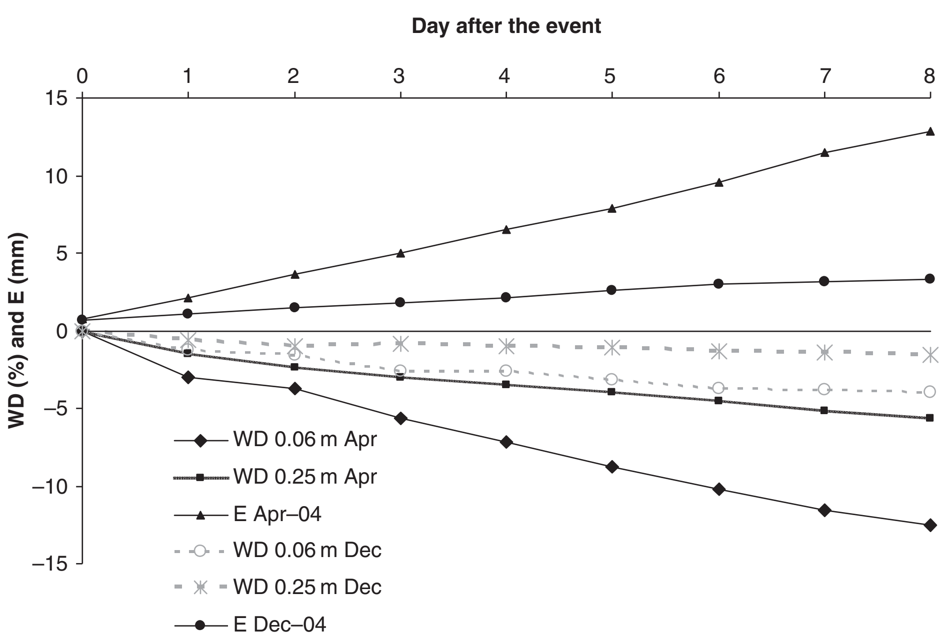

Fig. 6 Daily evapotranspiration (E) and soil water depletion (WD) at 0.06 and 0.25 m depth, after two rainfall events that occurred in December and April.

was more accentuated in late spring when E rates were higher. The use of maximum $\theta_{30\text{min}}$ to estimate PR (Approach 2b) for every recharge event increased PR more than threefold for the period from 12 February 2004 to 30 September 2004. Several authors (Howard & Lloyd, 1979; Taylor & Howard, 1996) have pointed out the influence of time scale in estimating PR using the soil water balance method, insisting on the need for a daily time step because longer (10-day or monthly) intervals can lead to significant underestimation of PR. Our results also show the importance of time scale and, furthermore, provide evidence of underestimation of PR when using daily time steps, at least during periods when E and soil water loss rates are very high in estimations performed using mean daily θ data (Approach 2). Taking account of time scale is particularly important in semi-arid Mediterranean environments with multiple short wetting and drying cycles. Future work should explore the performance of Approach 2a in wet autumns and winters with PR, which was not possible during the present work.

Comparison of daily E and soil water losses on days without rainfall from 12 February 2004 to 7 July 2005 (when θ was recorded at two depths) (Fig. 7) supports PR in spring of hydrological year 2003/04. In summer, daily E values are higher than daily θ losses, as most summer data are above the 1:1 line, suggesting capillary rise from deeper layers. As θ decreases, most E is from transpiration, and plants have to compensate the lack of water in the soil profile by exploring cracks and fissures in the bedrock. In winter, E was generally low (34 mm all winter, 2004/05), and θ losses were therefore usually greater than E. In autumn, daily θ losses were usually higher than the corresponding daily E rates, and in spring, on many days E was higher than the corresponding loss of θ, as expected for this growing season. According to phenological data from Serrano-Ortiz *et al.* (2007), maximum photosynthetic activity in tussock grassland occurs in June and July,

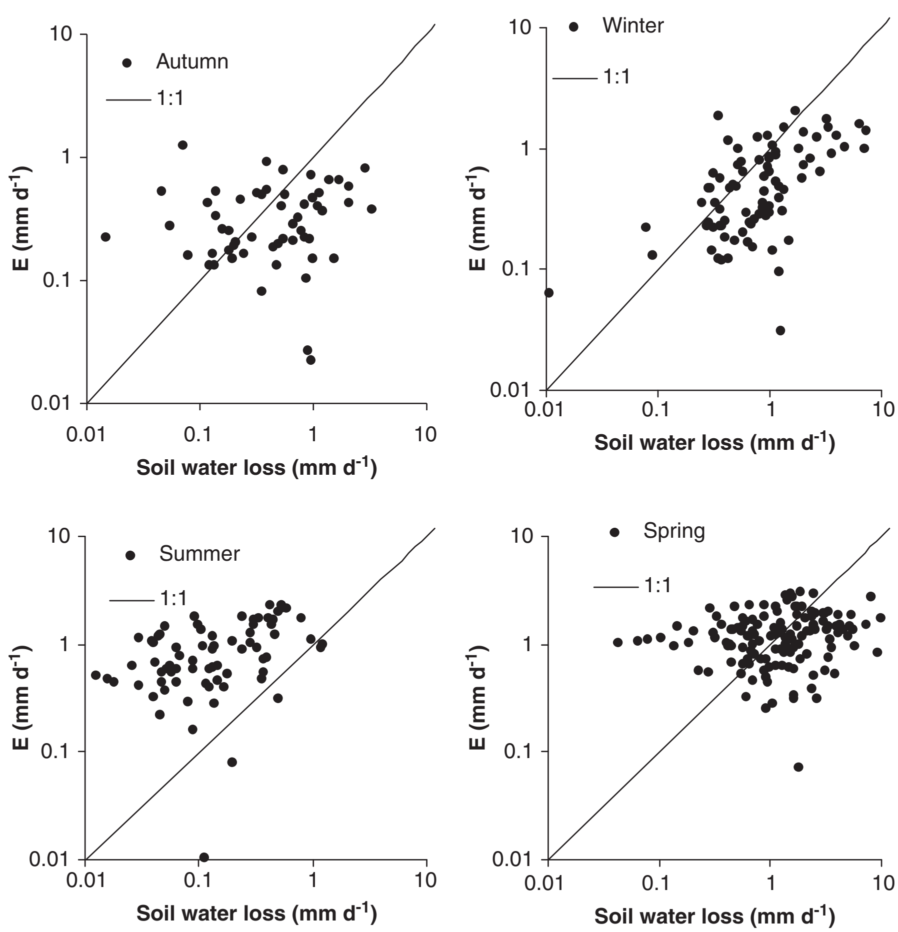

Fig. 7 Relationship between daily rates of evapotranspiration (E) and loss of water (for the soil profile, i.e. 0.35 m depth) during hydrological year 2003/04.

decreasing in August. Evapotranspiration was associated with that phenological activity, coinciding with the results in other tussock grasslands (Ramírez *et al*., 2007). However, in spring there are also many days with θ losses much greater than the corresponding E rates. The data corresponding to θ losses >3 mm are all two, three or even four days after rainfall events in spring 2004 when PR was estimated by Approach 1.

The sensitivity analysis indicates that PR estimates using approaches 2a and 2b suffer from high uncertainty due to the spatial variation of θ. The high variability in surface cover, which is a mosaic of shrub, rock fragments and grass (Li *et al*., 2008) – characteristic of the study site – configures a complex matrix of structures acting as sources (bare patches) and sinks (vegetated patches) of soil water (Calvo-Cases *et al*., 2003), which also affect θ heterogeneity and PR. In general, vegetation type also significantly impacts recharge rate (Lerner *et al*., 1990; Keese *et al*., 2005). However, at the study site Li *et al*. (submitted) found no significant differences in wetting front depths between the different plant species studied (i.e. *T. serpylloides*, *H. spinosa*, *G. pumila* and *F. scariosa*), or in associated unsaturated hydraulic conductivity, as reported by Li *et al*. (2008). Moreover, variability in soil properties – for example, stone content and distribution in the soil and root abundance and distribution – is also important in controlling recharge. Hence, approaches 2a and 2b showed PR estimation errors of about ±20% when mean daily θ was used (Approach 2a) and ±11% when using maximum θ_{30min} (Approach 2b). The lower uncertainty of Approach 2b *vs* 2a may be characteristic for this type of environment; the standard deviation of θ is lower for dry and saturated soil conditions (very well represented by maximum θ_{30min}) than for intermediate conditions (daily mean θ after some rainfalls could be closer to intermediate conditions than to soil saturation conditions) as occurs for other Mediterranean environments (Llorens *et al*., 2003).

Most water balance models use the soil water content at field capacity (θ_{FC}) (Fazal *et al*., 2005). Our results show that θ_{FC} estimated from laboratory data was higher than field θ_{FC} during rainy periods, when θ_{FC} was presumably reached. Pachepsky *et al*. (2001) compared soil water retention in fine-textured soils in both the field and laboratory, and found that water content was substantially less in the field than in the laboratory. This discrepancy between field and laboratory θ_{FC} has also been observed by other authors in different environments (Field *et al*., 1985; Shuh *et al*., 1988; Bork & Diekrügger, 1990; Ladekarl, 1999), and is attributed to sample disturbance, scale effects, or inadequate representation of large pores in the laboratory samples. In our case, the use of laboratory data should predict a lower PR. This implies significant uncertainty and ambiguity in the PR data, demonstrating the need to give more consideration to the effects of soil parameters (e.g. θ_{FC} or gravel content) when applying the water balance model or pedotransfer functions for PR prediction in the laboratory, or in water balance models in general.

5 CONCLUSIONS

Evapotranspiration was quantified by eddy covariance for two hydrological years (2003/04, a nearly average year, and 2004/05, a rather dry one) in the main recharge area of a limestone mountain range in southeastern Spain. In 2003/04, E represented 64% of rainfall and in the drier year (2004/05) was over 100% of rainfall. In both years, E occurred predominantly in late spring and summer, coinciding with maximum photosynthetic activity of the vegetation cover. Nevertheless, the E rates in the first half of summer are surprisingly high considering the relatively low θ in the soil profile, indicating extraction of water by plants from deep cracks and fissures in the bedrock.

Our results suggest a recharge contribution from Llano de los Juanes to the aquifer system in hydrological years with near-average precipitation conditions. The magnitude of effective aquifer recharge induced by PR is unknown.

The PR underestimation of Approach 2, even taking into account the uncertainty caused by spatial variations in θ, showed that in Llano de los Juanes an important fraction of PR occurs in preferential flows through the abundant cracks, joints and fissures, which is difficult to detect using θ probes installed in the soil. Moreover, the application of Approach 2a (using daily mean θ) significantly underestimates PR in late spring, with no PR estimation for some events, showing that the high E rates led to lower mean daily θ resulting from faster soil water depletion each day. Analysis of daily θ and E after spring rainfall events, and comparison of daily E and daily soil water losses during the year, confirmed PR in spring. Using maximum θ_{30min}, a higher PR is predicted, especially when E is very high and the daily mean θ lower, as occurs during late spring, though the estimated PR rates are still less than those predicted by Approach 1, confirming that preferential flow is a significant factor affecting recharge at this site.

The application of Approach 2 also demonstrated that PR estimation based on field θ data differs widely depending on the time scale considered: with daily mean θ data the total PR resulting from the process of downward movement of soil water by gravity is underestimated, especially in late spring, while the use of maximum θ_{30min} yielded a more reliable prediction. These results provide evidence for the known influence of time scale on the PR estimation from soil moisture data, emphasizing that in semi-arid Mediterranean environments with multiple short wetting/drying cycles even a daily scale may be insufficient temporal resolution to detect brief but sizeable PR events.

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