SOIL WATER BALANCE IN A SEMIARID COASTAL RANGE: POTENTIAL RECHARGE ESTIMATION

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Introduction

This paper analyses the water balance in the main recharge area of a semiarid range (Sierra de Gádor, Almería, SE Spain). In this area, like in many Mediterranean coastal zones, groundwater demand has rapidly increased due to expansion either in agriculture or in tourism, and its dependence on subterranean water resources. Therefore, an understanding of ground water recharge is necessary to improve the management and sustainable development of groundwater resources (Taylor and Howard, 1996).

Among others, models based on the water balance method are widely used to estimate deep percolation or deep drainage (D) (Fazal *et al*., 2005), although it is necessary to assess all other components of the water balance (runoff, interception, evapotranspiration or infiltration) either with direct measurements or with models. On the other hand, potential recharge can be estimated easily from soil water content (θ) monitoring.

The main objective of this work is to calculate D through (1) a simple approach based on the continuous record of θ, and to compare the D values obtained with those calculated with (2) a more complex approach based on the water balance method which needs to take into account precipitation and actual evapotranspiration (AET).

Material and methods

Sierra de Gádor aquifer consists of a thick series of highly permeable Triassic limestone and dolomites, which underlain by Permian to Triassic metapelites of low permeability, extending under the Campo de Dalías coastal plain (Pulido-Bosch *et al*., 2000). About 70% of the Campo de Dalías surface is covered by greenhouses and the annual rainfall is about 200 mm. Water pumped from Triassic aquifers represents more than twice the estimated recharge (Pulido Bosch *et al*., 2000). A relatively high flat area located between 1200 and 1800m a.s.l. is considered the main recharge area. An experimental site, representative of this recharge area, was established in the *Llano de los Juanes*, a central high plain of about 2 km^2 located at 1600 m a.s.l. (Li *et al*., 2006, Contreras *et al.*, 2008), corresponding to a well-developed karstic plateau (Li *et al.*, 2007), fragmented and slightly tilted to the East. It is an ideal area to measure the variables involved in the water balance and hence to estimate the potential recharge.

To estimate the deep drainage (D) we use two approaches:

1) The first approach estimates D exclusively from daily Θ measurements and assumes that the soil becomes free draining when Θ reaches the field capacity (Θ_{fc}); once this point is reached, excess water drains below the root zone and flows through unsaturated zone towards the regional water table.

$D = \Theta - \Theta_{fc}$

Field capacity (Θ_{fc}) was determined in laboratory when the decreasing rate of Θ was less than 0.001 $m^3 \cdot 30\ min^{-1}$ (Bruno et al., 2006).

2) The soil water balance approach also assumes that the soil becomes free draining when Θ reaches field capacity. However, to determine D with this method, it is necessary to simulate Θ on a daily basis throughout the year (Rushton *et al.*, 2006) starting with a deficit known value of daily Θ and calculate consecutive daily Θ deficits. D is calculated as:

$D = -SMD$ (for $SMD < 0$) and $D = 0$ (for $SMD \geq 0$)

$SMD = SMDpr - P + AET - Ron + Roff$

where SMD is the daily Θ deficit, defined as the amount of water required to bring the soil up to field capacity (Rushton *et al.*, 2006), SMDpr is SMD previous day value, P is daily precipitation, AET is daily actual evapotranspiration, Ron is daily runon, Roff is daily runoff by overland flow or sub-surface flow.

D is assumed to occur on days with a negative Θ eficit. As Θ deficit becomes zero, the soil reaches field capacity and becomes free draining. Consequently, D equals the quantity of excess water from that required to bring the soil to field capacity.

During two hydrological years (2003 - 2004 and 2004 – 2005) AET was measured by the Eddy covariance (EC) method (Swinbank, 1951; Anderson *et al.*, 1984), comprising two turbulent sensors measuring at high frequency 10Hz: a sonic anemometer (CSAT3, Campbell Scientific Inc., USA) measuring the three components of wind speed and a krypton hygrometer (KH20, Campbell Scientific Inc., USA) measuring fluctuations in water vapour density. Measurements were averaged and logged every 30 minutes, and corrections of the water and heat fluxes were made according to air density fluctuations (Webb *et al.*, 1980) and two coordinate system rotations (Kowalski *et al.*, 1997). Hygrometer measurements were also corrected for absorption of radiation by oxygen (Tanner *et al.*, 1993). Rainfall amount and intensity have also been recorded at the site by an automatic 0.20 mm-resolution tipping-bucket rain gauge. Continuous Θ measurements were made with probes installed at 0.06 m depth. All data were averaged every 30 min.

Results and discussion

Rainfall, Θ and AET variations for the hydrological years 2003 – 2004 and 2004 – 2005 are shown in Figure 1 and Table 1. Θ measured at 0.06 m depth, remained high during the autumn and spring periods, decreasing rapidly after May, with a minimum value of about 4% volumetric. Comparing both years, rainfall was more frequent and abundant in 2003-04, with a higher AET, especially from spring end to early summer, which shows the effect of the higher Θ in the moments of larger vegetation activity.

During the wetter hydrological year (2003-04), for 108 days the average Θ corresponded to a water tension between –33 kPa and –1500 kPa, indicating that there was no significant gravitational water or, if there was, this would only be for short periods of less than one day. Average daily Θ was wetter than –33 kPa on 31 days, and gravitational water movement was then possible. For the hydrological year 2004-05, the daily average Θ never exceeded the soil water content at –33 kPa at a daily scale, suggesting that deep percolation, if it occurs, is limited to periods of hours or less. Table 1 shows the annual distribution of the three variables.

The use of both approaches to calculate the deep drainage (D) gave rise to the following results:

Approach 1 (considers only soil moisture field data and assumes no drainage when Θ is lower than Θ_{fc}): Θ_{fc} assessed from Θ measurements during the wet season was 37% (vol) at 0.06 m depth. The estimation of D was 237 mm $year^{-1}$ (47% of the total rainfall) for hydrological year 2003 – 04. For hydrological year 2004 – 05 no D was generated because Θ never reached Θ_{fc} .

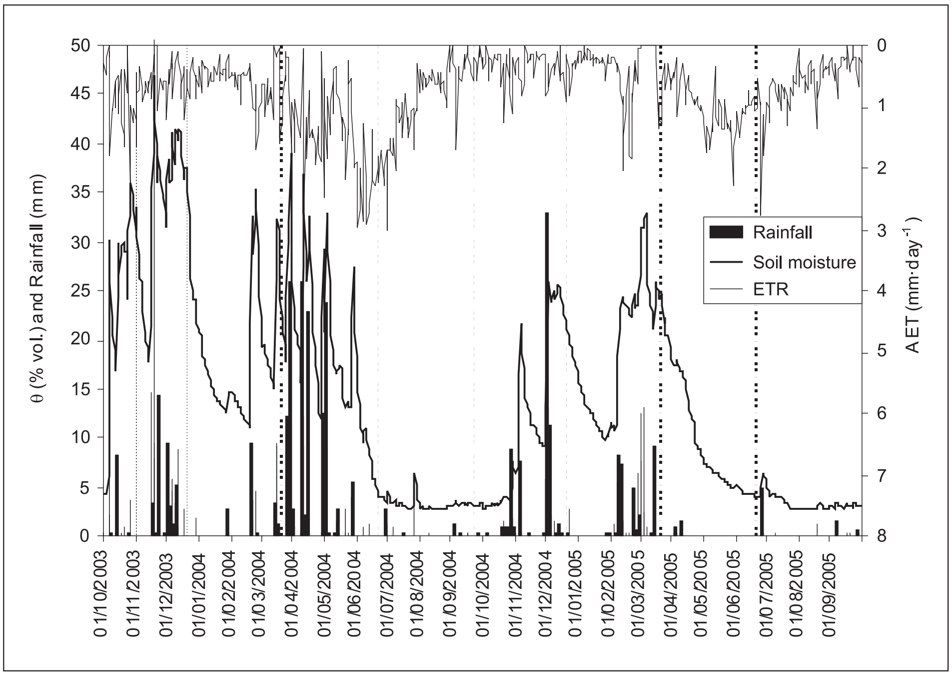

Figure 1. θ measured at 0.06 m depth, AET and rainfall for hydrological years 2003 – 04 and 2004 – 05 at the Llano de los Juanes experimental site

Table 1. Monthly distributions and annual rainfall (P), measured evapotranspiration (AET) and mean soil moisture for hydrological years 2003-04 and 2004-05

Month	2003-04			2004-05		
	P (mm·month^{-1})	AET (mm·month^{-1})	θ (% vol.)	P (mm·month^{-1})	AET (mm·month^{-1})	θ (% 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 sum (mm·y^{-1})	507.28	326.49		209.91	214.92	
Annual mean (mm·y^{-1})		0.89	16.97		0.59	10.75

Approach 2 (uses rainfall, runoff and AET data): The deep percolation (D) calculated by this approach was 181 mm year^{-1} for the hydrological year 2003-04, representing 36% of the total rainfall. For the hydrological year 2004-05 the estimated D was 8 mm year^{-1}. This small value of D can be explained because the 2004-05 hydrological year was very dry (Fig. 1 and Table 1) with AET representing more than 100% of the total rainfall (210 mm) and therefore, with D very reduced or absent.

The comparison of both approaches showed that using only data from 2003–04 hydrological year, the total annual D calculated with approach 1 was significantly greater than when calculated from approach 2 (about 50 mm). Moreover, the temporal patterns of D obtained with both approaches are quite different (see

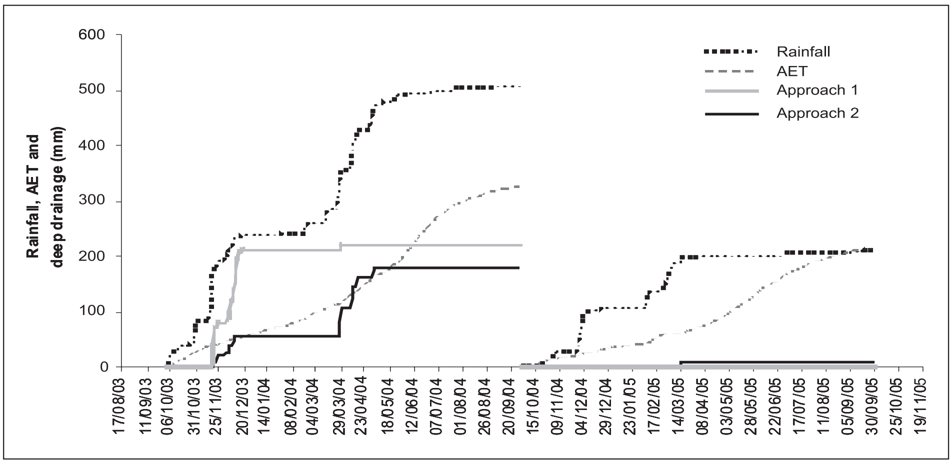

Figure 2. Rainfall, AET and deep drainage (D) in mm calculated with approach 1 and 2 on the hydrological years 2003–04 and 2004–05

Fig. 2). The most striking difference is that in approach 1, D is concentrated in autumn and winter (94% of total drainage or recharge), while approach 2 estimated 66% of the total drainage accounted in spring. This behaviour of approach 1 can be explained both because Θ rarely exceeded Θ_{fc} in spring, and because soil water depletion is faster in spring than in autumn and winter due to higher AET rates in spring (Figure 1).

Hence, the analysis of daily Θ and AET after spring rainfalls and the comparison of daily AET and daily soil water losses along the year confirmed the occurrence of deep drainage in spring. When D is assessed considering only Θ evolution (approach 1), results are very sensitive to field capacity estimates, probably due to the presence of rock fragments and the soil depth at which Θ is measured.

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References

Anderson, D.E., Verma, S.B. and Rosenberg, N.J., 1984: *Eddy correlation measurements of CO_2, water vapour and sensible heat fluxes over a crop surface*. Boundary-Layer Meteorology, 29: 167-183.

Bruno, R.D., da Rocha, H.R., de Freitas, H.C., Goulden, M.L. and Miller, S.D., 2006: *Soil moisture dynamics in an eastern Amazonian tropical forest*. Hydrological Processes, 20: 2477-2489.

Contreras, S., Boer, M.M., Alcalá, F.J., Domingo, F., García, M., Pulido-Bosch, A. Puigdefábregas, J. 2008: *An ecohydrological modelling approach for assessing long-term recharge rates in semiarid karstic landscapes*. Journal of Hydrology, 351(1-2): 42-57.

Fazal, M.A., Imaizumi, M., Ishida, S., Kawachi, T. and Tsuchihara, T., 2005: *Estimating groundwater recharge using the SMAR conceptual model calibrated by genetic algorithm*. Journal of Hydrology, 303: 56-78.

Kowalski,A.S., Anthoni, P.M., Vong, R.J., Delany, A.C. and Maclean, G.D., 1997: *Deployment and Evaluation of a System for Ground-Based Measurement of Cloud Liquid Water Turbulent Fluxes*. Journal of Atmospheric and Oceanic Technology. 14(3): 468-479.

Li, X.-Y., Contreras, S. and Solé-Benet A., 2007: *Spatial distribution of rock fragments in dolines: A case study in a semiarid Mediterranean mountain-range (Sierra de Gádor, SE Spain)*. Catena, 70(3): 366-374.

Li, X.-Y., González, A. and Solé-Benet, A., 2005: *Laboratory methods for the estimation of infiltration rate of soil crusts in the Tabernas Desert badlands*. Catena, 60(3): 255-266.

McMillen, R.T., 1988: *An eddy correlation technique with extended applicability to non-simple terrain*. Boundary-Layer Meteorology, 43: 231-245.

Pulido-Bosch, A., Pulido-Lebouf, P., Molina, L., Vallejos, A. and Martín-Rosales, W., 2000: *Intensive agriculture, wetlands, quarries and water management. A case study (Campo de Dalías, SE Spain)*. Environmental Geology, 40(1-2): 163-168.

Rushton, K.R., Eilers, V.H.M. and Carter, R.C., 2006: *Improved soil moisture balance methodology for recharge estimation*. Journal of Hydrology, 318: 379–399.

Tanner, B.D., Swiatek, E. and Green, J. P., 1993: *Density fluctuations and use of the krypton hygrometer in surface flux measurements*. In: Allen, R.G. and Neale, C.M.U., Proc. Management of Irrigation and Drainage Systems. ASCE. Park City, UT: 105-112.

Taylor, R.G. and Howard, K.W.F., 1996: *Groundwater recharge in the Victoria Nile basin of east Africa: support for the soil moisture balance approach using stable isotope tracers and flow modelling*. Journal of Hydrology, 180: 31-53.

Swinbank, W.C., 1951: *The measurement of vertical transfer of heat and water vapor by eddies in the lower atmosphere*. Journal of Meteorology, 8(3): 135 - 145.

Webb, E.K., Pearman, G.I. and Leuning, R., 1980: *Correction of flux measurements for density effects due to heat and water vapour transfer*. Quarterly Journal of the Royal Meteorological Society, 106: 85-100.