Modelling the groundwater level by water balance: a case study of a Mediterranean karst aquifer of Apulia region (Italy)

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Abstract: *Groundwater is the most easily accessible resource for irrigation so, due to recent climate change with increasing needs of water, aquifers are subject to be depleted at a high rate. This study examines the groundwater level (*GL*) of a well in a rural agricultural area of south Italy (Apulia region), submitted to semi-arid Mediterranean climate. A simple model of groundwater recharge (*R*) is proposed based on the water balance where the term of evapotranspiration and runoff has been deeply investigated.* R *and* GL *are linearly linked by the specific yield* S_y *for karst aquifer. Starting from the analysis of the rainfall regime of the last 39 years the proposed runoff model has been designed in order to take into account periods of successive heavy rainy days, while the water lost by crops has been modelled by the potential evaporation (*PE*). Three models of* PE *were presented; the Penman and the Hargreaves-Samani ones resulted the most suitable. This study showed that the presented model, based on the simplified water balance, can estimate the groundwater* R *at daily scale quite accurately.*
Keywords: *Groundwater recharge, potential evaporation, runoff, rainfall regime analysis, semi-arid climate.*

Riassunto: *L'acqua di falda è la risorsa idrica più facilmente accessibile per l'irrigazione per cui, a causa dei recenti cambiamenti climatici, con gli aumenti delle necessità d'acqua, gli acquiferi sono soggetti a rapido depauperamento. Qui si analizza il livello di falda (*GL*) di un pozzo in una zona rurale del sud d'Italia (regione Puglia), sottoposto a clima semi-arido mediterraneo. Si propone un semplice modello di ricarica (*R*) basato sul bilancio idrico, dove i termini di evapotraspirazione e deflusso superficiale sono stati studiati in dettaglio.* R *e* GL *sono correlati linearmente attraverso il termine di resa* S_y *specifico per un acquifero carsico. A partire dell'analisi del regime pluviometrico degli ultimi 39 anni, un modello di deflusso superficiale è stato progettato per tener conto dei periodi con giorni successivi di pioggia intensa, mentre l'acqua persa dalle colture è stata modellata con l'evaporazione potenziale (*PE*). Sono presentati tre modelli di* PE*; quelli di Penman e di Hargreaves-Samani risultano I più adatti. Questo lavoro mostra che il modello presentato, basato su un bilancio idrico semplificato, può stimare* R *a scala giornaliera con sufficiente accuratezza.*
Parole chiave: *Ricarica falde, evapotraspirazione potenziale, deflusso, analisi del regime pluviometrico, clima semi - arido.*

INTRODUCTION

The 70% of available freshwater at global scale is employed for crop production (Molden, 2007) and irrigated crops in arid and semi-arid zones exert heavy pressure on the available scarce resources (Smedema and Shiati, 2002; Katerji *et al.*, 2008). Among the available resources for irrigation, groundwater is the main, the more reliable and, often, the most easily accessible. So, where groundwater is used for irrigation, aquifers are subject to be depleted at high rate. The rate of depletion is increasing in the last years also due to the lengthening of the irrigation season under climate change scenarios, i.e. the actual perceived and manifest climate modification to wards drought conditions (Linderholm, 2006; Hatfield *et al.*, 2011).

For the above mentioned reasons the groundwater level (*GL*) should be always monitored in agricultural rural areas in order to avoid dangerous depletions. Indeed, many countries have been developed net works of wells' level monitoring so far, to quantify the natural groundwater recharge (*R*)rate in order to improve groundwater resource management. This issue is particularly relevant in arid and semiarid regions, such as the Mediterranean one, where large demands for water supplies are the keyto economic development. The recently large variations of *GL* over years in many parts of arid and semiarid sites suggest to undertake precise and detailed studies for accurately understanding the behaviour of *GL* fluctuations at both spatial and temporal scale,site by site (Ahmadi and Sedghamiz, 2007; Zhou, 2009; Touhami *et al.*, 2013).FollowingHealy and Cook (2002) the changes in water table level (Δ*GL*) is linked to the groundwater recharge *R* by the relation:

$$R = S_y \cdot \Delta GL \quad (1)$$

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Submitted 15 June 2017, accepted 9 January 2018

DOI:10.19199/2018.1.2038-5625.035

where specific yield (S_y), also known as the drainable porosity, is a ratio, less than or equal to the effective porosity, indicating the volumetric fraction of the bulk aquifer volume that a given aquifer will yield when all the water is allowed to drain out of it under the forces of gravity. The range of S_y values for unconfined is between 0.00 (*clay*) and 0.26 (*coarse gravel*) (Healy and Cook, 2002). S_y for confined aquifers generally range from 1 x 10^{-3} to 1 x 10^{-5} (Kozar and Mathes, 2001). In matrix of karst aquifers, S_y is significantly variable: 5.4 x 10^{-5} ÷ 3.6 x 10^{-4} (Baedke and Krothe, 2001); 1 x 10^{-4} ÷ 0.3 with a median of 7 x 10^{-3}(Kozar and Mathes, 2001); 3 x 10^{-3} (Shevenell, 1996). Consequently, the average value is 2 x 10^{-3}.
R rate can be modelled by different approaches, depending on the needed space/time scale of the study objectives (see Scanlon *et al.*, 2002 for a review). One way for estimating groundwater *R* is based on the "indirect" or "residual" approach of the water-budget method, whereby the variables as rainfall (input), evapotranspiration (*ET*, output), runoff (output) capillary rising (output), deep percolation (input) and soil water storage (output) and infiltration (output) are measured or estimated, while groundwater *R* is set equal to the residual of the water budget equation (Sophocleous, 1991; Scanlon *et al.*, 2002; Lo Russo *et al.*, 2003; Cherkauer, 2004; Yeh *et al.*, 2007; Lautz, 2008, among many others). This approach has the advantage to require few assumptions on the equation, but its accuracy depends on the accuracy in the determination of the components in the water budget equation (Scanlon *et al.*, 2002).Moreover, the choice of the time step (at least daily) is crucial for the applicability of this approach also in arid and semiarid regions (Scanlon *et al.*, 2002).
Theoretically, the *GL* of an aquifer, as well as the *ET*, varies at instantaneous scale and, with suitable gauges, could be measured with any resolution. Usually, the hourly scale is considered appropriate for *GL* measurements (Cherkauer, 2004; Gribowski *et al.*, 2010; Cheng *et al.*, 2013, among many others): at this scale, estimation of the so called groundwater evapotranspiration is also possible (White, 1932). Moreover, direct measurements of groundwater fluctuations include supplementary indirect information, like the time during the growth cycle of the plant, the plant type and the moisture availability. For hydrological analysis, the order of cumulated hourly values of evapotranspiration or potential evaporation, runoff and precipitation enable to evaluate water balance at different time scales (day, decades, month, season, year) and to define the water reservoir available in a given area at that scale (Margat, 1992).
Among the terms of the water balance, evapotranspiration is the most difficult to estimate. Actually, many models of evapotranspiration are available in literature (Rana and Katerji, 2000), but, unfortunately, the most accurate evapotranspiration models need weather input variables (solar/net radiation, air temperature and humidity, wind speed) that are not easily supplied in rural areas, where continuous available measurements are, more often, rainfall and air temperature or, mostly, the incident solar radiation.
Furthermore, the problem exists on which quantity should be used in groundwater *R* models to take into account the evapotranspiration variable, thought as loss of water by the system soil-vegetation (Katerji and Rana, 2011). In fact, terms like "potential evaporation" (*PE*), "potential evapotranspiration (*PET*)", "actual evapotranspiration (*ET*)" and "reference evapotranspiration" ET_0 are often misused, even if different models were studied to establish the best one, or the most accurate, or the simplest to be used in the water-budget approach for hydrological purposes (Xu and Singh, 2000; Oudin *et al.*, 2005; Xu and Chen, 2007). *PE* is defined as evaporation from a surface saturated in water (free water at the surface or 100% of humidity on the crop) where loss of water is considered without either the biological control or the control exerted by the vegetation structure (Katerji and Rana, 2011). Thus, this variable is only theoretical and is the amount of water that the atmosphere can retain at given thermodynamic conditions; it does not depend on the vegetation control. The latter feature makes *PE* particularly suitable to be used in water balance modelling at catchment/regional scale, where taking into account all species is not possible.
Another variable modelled in the water balance is the runoff, for which a lot of complex models are available, all requiring many inputs. In general, water table responses are affected by surface and subsurface topography, antecedent soil moisture conditions and rainfall events intensity (Haught and van Meerveld, 2011). Conditions exist in which capillary rising, deep percolation and runoff can be neglected in soil water balance (Rana and Katerji, 2000),even if recent studies have shown that, where the subsoil has impermeable

bedrock flow through, there can be a significant sub-surface runoff component in case of sloped surfaces (Haught and van Meerveld, 2011). The simplest and enough accurate applicative models of runoff for hydrological purposes are based on the existence of a minimum threshold of rainfall above which the runoff is greater than zero and equal to a percentage of the precipitation (Franquin and Forest, 1977; Lhomme and Katerji 1991; Norbiato *et al.*, 2008). These models seem to efficiently simulate the runoff also in arid and semi-arid conditions at daily scale (Lhomme and Katerji, 1991; Rana and Katerji, 2000). Unfortunately, the measured current climate change modified the rainfall regime of large areas (Brunetti *et al.*, 2001; Arora, 2002; Donohoue *et al.*, 2011, among many others) and, consequently, runoff models must be assessed, re-calibrated and updated continuously (McMahon *et al.*, 2011; Du *et al.*, 2016).

Particularly, in the semi-arid areas of south Italy the precipitation regimes of last decades present irregular distribution of rain with drought or wet extreme events, during growing seasons for crops of agricultural interest (Rana *et al.*, 2016; Katerji *et al.*, 2016).

This study examines the *GL* of a well in a rural agricultural area of south Italy, in a site submitted to Mediterranean climate characterised by a hot and dry season in summer and a mild temperature associated to annual rainfall in winter. In fact, at the site the mean annual temperature is 15.6 °C with a mean annual precipitation of 584 mm. Following Holdridge Life Zones (Holdridge *et al.*, 1967; 1971), which classify areas with potential evaporation/mean annual precipitation ratio > 1 as dry climates and areas with ratio < 1 as humid climates, experimental site can be classified as dry since the aridity index is 1.62 on the period 1977-2011.With these characteristics the climate of the site can be considered as semiarid following the discussion in Rana and Katerji (2000) and Allen *et al.* (1998).

The *GL* measurements were analysed with the main purpose of producing a sufficiently simple and accurate model for calculating groundwater recharge *R* by using easily available agro-meteorological input variables. The water balance is the chosen approach to model *R*. The objectives of this study are: (i) to analyse all terms of the water balance model, also in relation to the measured current change in the precipitation regime; (ii) to study the performances of different models of water lost by the crop-soil system at regional scale as potential evaporation; (iii) to formulate a simple *R*model based on the water balance approach to obtain enough accurate estimations of *GL*.

MATERIALS AND METHODS

In this section at first the experimental site is described, then all terms of the proposed water balance model are illustrated in details. Particular attention is paid here to the descriptions of the three model of water loss by the system soil-crops as potential evaporation, *PE*; then three *PE* models are compared to have the best performances of *GL*. Finally the adopted water balance model is presented.

The experimental site

The experimental and theoretical work has been finalized to be applicable to a karst aquifer in an environment submitted to semiarid climate of the Mediterranean region. More precisely, the well is monitored by a regional network managed by a public institution (Regione Puglia) and is placed in the rural territory of Rutigliano, south Italy (40°59′33′′ N; 17° 2′1′′ E , altitude 147 m a.s.l.). *GL* was measured by an automatic sensor (Sensor TechnikSirnach AG, Germany). The monitored well falls in the karst aquifer of the Murgia characterized by an area of 6843 $km^2$where the groundwater flows to the sea in perpendicular direction to the coast line, with piezometric gradients ranging from 0.1% to 0.5% (Cotecchia, 1993).The monitored well is located in karst areas where aquifer is confined (Coteccha *et al.*, 1999) and soil is composed by fractured limestone and limestone-dolomite rocks covered with small thick of colluvial deposits. The monitored well falls into a groundwater recharge area and intercepts water table at a depth of about 100 meters, whose variations are poorly influenced by marine intrusion for effect to the high distance from the coast line (about 30 km). The recharge area is crossed by a dryland river system (Giotta watershed) where the runoff is present only during very intense meteorological events. The runoff absence for large time intervals is caused by high hydrological losses due to infiltration through micropores and karstic forms that characterize the "*soil - subsoil*" system in the study area. For this reason, the groundwater *R* is mainly guaranteed by rainwater.

The measurement period was between 2nd November 2009 and 28 June 2011 (604 days), covering one complete irrigation season in

the spring-summer 2010 and part of the next irrigation season in the spring 2011.
All agro-meteorological variables to calculate water balance terms were measured above the reference grass meteorological station in the CREA-AA experimental field (40° 59′ 33.45′′N, 17° 1′ 57.69′′E, 147 m a.s.l.). The sensors and the characteristics of the reference grass field were accurately described in Rana *et al.* (1994) and Rana *et al.* (2012).The soil at this site is predominantly clay and silty (41% and 26%, respectively); soil depth is up to 0.60-0.70 m because of a cracked rocky layer which limits root development, ensuring optimal drainage of excess water (Campi *et al.*, 2009). The climate of the site in terms of minimum and maximum air temperature and precipitation, at monthly scale in the period 1977-2011 is shown in Tab. 1.
The specific yield for this aquifer (S_y in Eq. 1) was taken equal to 1.72 x 10^{-3}, obtained as the mean value of S_y for rock matrix, fracture and conduit that characterize the subsoil in karst areas (Shevenell, 1996). This value is coherent with the mean value of Sy obtained from various studies conducted in the karst environment (Baedke and Krothe, 2001; Kozar and Mathes, 2001; Shevenell, 1996).
In these particular areas the water movement in subsoil occurs through the rock matrix micropores, the fractures caused by tectonic processes and the conduit due to the karst phenomena.

Month	Tmin (C)	Tmax (C)	P (mm)
January	4.09	12.08	57
February	3.97	12.24	64
March	5.94	14.95	58
April	8.38	17.90	47
May	12.68	23.33	34
June	16.27	27.47	28
July	19.08	29.81	20
August	18.96	29.90	23
September	16.13	26.34	56
October	12.45	21.88	60
November	8.45	17.20	70
December	5.13	13.14	67

Tab. 1 - Mean monthly values of minimum, maximum air temperature and cumulated precipitation in the period 1977-2015 in the experimental site.
Tab. 1 - Valori medi mensili di temperature dell'aria minima, massima e precipitazioni nel periodo 1977-2011 nel sito.

The rainfall regime and the air temperature increasing

The recent precipitation measurements identified the Mediterranean region as one of the most prominent "hot-spots" in future climate change rainfall projections (Gao *et al.*, 2006; Giorgi and Lionello, 2008; Espadafor *et al.*, 2011). In particular, for Italy, many studies confirm a negative precipitation trend (i.e. Brunetti *et al.*, 2001), steeper in the central and southern areas than in northern ones (Rana *et al.*, 2016; Katerji *et al.*, 2016). Variations in total precipitation can be caused by change in the frequency of precipitation events, or in the intensity of precipitation per event, or a combination of both. Since precipitation has great impact on the *GL* and, particularly, on the amount of water discharged by runoff, the variations in heavy precipitation are particularly important to address valuable modelling of the runoff term in the water balance for *R* modelling (Zhang *et al.*, 2008; Beven, 2012). Therefore, the rainfall regime of the site has been analysed in detail, considering a 39-years long series (1977-2015) of precipitation measured in the experimental field. All the presented precipitation data were measured at hourly scale by mechanical rain gauges (SIAP, Bologna, Italy, various models in the 39-years period), annually calibrated against electronical rain gauges (TEXAS Inst., USA, different models).
First, standard elementary statistics (mean, minimum, maximum) of the annual precipitation series were calculated and normality was evaluated via both a numerical and a graphical method, the Shapiro-Wilk test (Shapiro and Wilk, 1965) and the Quantile-Quantile (Q-Q) plot method (Wilk and Gnanadesikan, 1968), respectively. The time series resulted not normal (data not shown) from the numerical and graphical analysis and this result was expected because time series of precipitation very often suffer, at all time scale, of

lack of assumptions required for the application of classical parametric analyses. Then, because continuous climatic time series are crucial for studying observed climate variability and trend, especially at local and regional scales (Easterling *et al.*, 1996), the presence of potential breakpoints was investigated. A nonparametric method was used for detecting any distributional change within an independent sequence of data without making any distributional assumptions (Matteson and James, 2014). The search for breakpoints did not give any result (data not shown).

The anomalies of annual and seasonal values of total precipitation (*TP*, mm/year), number of wet days (*WD*, % number of days/year) and precipitation intensity (*PI*, mm/wet days/year) were calculated. In particular, *WD* is the number of wet days with measured precipitation greater than 0.2 mm (the minimum measurable rainfall by rainfall gauges) and *PI* is average rain amount per rainy day. Furthermore, it is important to know whether the change in precipitation frequency is due to a change in the number of days with heavy precipitation or with light precipitation. Thus, also the anomaly of the number of high rainy days (*HPD*, % number of high rainy days/year) was calculated. HPD is defined as the number of days in the year when precipitation is greater or equal to 25 mm (Forest, 1984). This last index has relevant importance in the parameterization of the runoff model. On these quantities the following techniques were applied for highlighting trends: (i) Theil-Sen trend line estimation (so called Theil-Sen approach - TSA) (ii) Ljiung-Box autocorrelation test, (iii) Man-Kendall (MK) test for trend.

The actual air temperature increasing measured in the experimental site (Katerji *et al.*, 2016) has impact on evaporation through the vapour pressure deficit, the slope of the saturation pressure curve and the psychrometric constant (see for example Rana and Katerji, 1998). Therefore, all these variables are included in the proposed potential evaporation models to take into account the temperature rising.

Runoff

Assuming that in consecutive heavy rainy days, starting from the second one, the runoff is independent by the threshold value P_0, in this work the following model for runoff has been used:

$$if\ P_t \leq P_0 \rightarrow R_{off} = 0 \tag{2a}$$

$$if\ P_t > P_0 \& P_{t-1} \leq P_0 \rightarrow R_{off} = \beta(P_t - P_0) \tag{2b}$$

$$if\ P_t > P_0 \& P_{t-1} > P_0 \rightarrow R_{off} = n\beta(P_t - P_0);\quad if\ R_{off} > P_t \rightarrow R_{off} = P_t \tag{2c}$$

where P_t represents the rain at the time t and n is the number of consecutive heavy rainy days; moreover, the conventional value P_0 = 25 mm and β=0.15(Forest, 1984) were confirmed for the region.

Evapotranspiration

Evapotranspiration is an instantaneous process (Monteith 1965; Perrier, 1975; Rana *et al.*, 1994 among many others), so it is theoretically correct to express *ET* at hourly scale. The combination Penman-Monteith model represents a basic general description of the evaporative process from a vegetative surface (Monteith, 1973; Rana *et al.*, 1994), but all input needed variables should be measured on the crop, with evident insurmountable practical problems. Therefore, *ET* is often calculated by a more operational model as

$$ET = K_c ET_0 \tag{3}$$

with K_c crop coefficient, specific for each crop and phenological stage, and ET_0 reference *ET*, i.e. *ET* measured above a reference crop (Rana *et al.*, 1994; Allen *et al.*, 1998).In this case, daily expressions for ET_0 can be found in Allen *et al.* (1998) and (2006). Other accurate operational *ET* models, at hourly and daily scales based on variables measured above reference grass without using K_c can be found in Rana *et al.* (2012). This latter models need calibration crop by crop.

Due to the inherent practical difficulty of estimating an accurate time-variable K_c (Lautz, 2008; Lazzara and Rana, 2010), *ET* by the Penman-Monteith model and *ET* by its operational versions for all crops cultivated in the given area, it is necessary to find an alternative greatness which validly represents the crop actual evapotranspiration in the recharge modelling represented by the water balance at territorial scale. Actually, for this purpose, *ET* is the water lost by the crops and soils and restored into the soil by irrigation with water supplied by the well or by precipitations. Therefore, it is reasonable to search for a greatness linked to the crop water requirements. The water requirement of a crop is directly linked to the evaporative demand of the atmosphere (Katerji and Rana, 2011), i.e. how much water the atmosphere is able to retain at given thermodynamic conditions and supposing that the water is already available in the soil or that the soil is water free. This greatness exactly responds to one of the definitions of potential evaporation, *PE* (Thornthwaite, 1948; Anon, 1956; Perrier, 1974; Katerji and Rana, 2011). Furthermore,the sensitivity analysis carried out by Finch (1998) on the parameters having the greatest influence on simplified water balance model for estimating groundwater *R*, founds that the recharge estimates resulted relatively insensitive to the vegetation canopy parameters for short vegetation, such as the land cover type of the studied area. Therefore, the *PE* modelling has been employed for the analysis in this study (Donohoue *et al.*, 2010) and three *PE* models are analysed in this following:(I) Penman, (II) Priestley-Taylor, (III) Hargreaves-Samani.

The Penman (1948) combination method is usually considered as the most physically based model to accurately estimate *PE* (see Oudin et al, 2005, and Katerji and Rana, 2011, for theoretical and practical considerations and overview). Penman model (PE_P) can be written in its original version as

$$PE_P = \frac{1}{\lambda}\frac{\Delta R_n + \gamma D f(u)}{\Delta + \gamma} \tag{4}$$

with Δ the slope of the saturated vapour pressure curve (kPa C^{-1}), R_n the net radiation, γ the psychometric constant (kPa C^{-1}), λ the latent heat of vaporization of water (MJ kg^{-1}), *D* the vapour pressure deficit (kPa) and *f(u)* wind function, analogous to aerodynamic resistance, but suitable for free water surfaces equal to

$$f(u) = 1 + 0.54u \tag{5}$$

u is the wind speed. It was amply demonstrated that the Penman model is the most suitable to well estimate the potential evaporation in Mediterranean environment and, specifically, in the present experimental site (e.g., Rana *et al.*, 1994; Katerji and Rana, 2011), so it was considered the reference method, even if it is the most complex since it requires many meteorological input variables.

Since all needed variables to calculate *PE* by the Penman model are not always available in rural sites, here two further formulas to calculate potential evaporation are proposed.

The Priestley-Taylor model ($PE_{E\text{-}T}$, Priestley and Taylor, 1972) has only the net radiation as input and is particularly suitable at regional scale:

$$PE_{P-T} = \frac{1}{\lambda}\alpha_{P-T}\frac{\Delta R_n}{\Delta + \gamma} \tag{6}$$

where $\alpha_{P\text{-}T}$ is a constant, equal to 1.26 in the original version, but that can be calibrated for each environment: in thissite it was taken equal to 1.16 (Rana, 1998).

The Hargreaves-Samani model ($PE_{H\text{-}S}$, Hargreaves and Samani, 1982,) has only the air temperature and the extra-terrestrial solar radiation (R_a) as input. This model was found to be also usable for modelling the potential evaporation (Fennessey and Vogel, 1996; Hargreaves and Allen, 2003) and can be written as:

$$PE_{H-S} = \frac{R_a}{\lambda}\alpha_{H-S}(T_{max} - T_{min})^{0.5}(T_a + \beta_{H-S}) \tag{7}$$

where $\alpha_{H\text{-}S}$ and $\beta_{H\text{-}S}$ are coefficients equal to 0.0023 and 17.8, T_{max}, T_{min} and T_a are maximum, minimum and mean air temperature in the day, respectively. R_a is calculated following Allen *et al.* (1998).

The other water balance terms: capillary rising, deep percolation, soil water storage

The amount of water transported upwards by capillary rise (C_r) depends on the soil type, the depth of the water table and the wetness of the root zone(Erickson and Stefan, 2007). In arid and semi-arid regions, the term C_r in the soil water balance could show problems of correct evaluation. In fact, if the soil system is closed (i.e. shallow soils or groundwater depth becomes 2-3 m), C_r can be considered negligible (Kahlown *et al.*, 1998). Vice versa, if the system is open (deep soils or soils with a surface water table), C_r cannot be neglected (Katerji *et al.*, 1984). In the present case, since the soil is shallow (see following description of the soil in the experimental site) C_r can be assumed to be zero, but we want to further underline that it could be necessary to estimate C_r in other sites.

The drainage term *Dr* represents the lateral flow processes in the saturated zone or towards adjacent waterways (Pirastru and Nieddia, 2011). Lateral discharge is mainly influenced by aquifer permeability and thickness. The impact of the lateral groundwater flow on water-level during a recharge event is generally ignored, which could underestimate the actual recharge rate

(Cai and Ofterdinger, 2016). In karstic contexts, variations in the *GL* are influenced from lateral flow by means of karst conduits which are strongly heterogeneous (Camarasa-Belmonte, 2016). However, the degree fracturing in the rock can be considered scarce and discontinuous in the study area (Cotecchia *et al.*, 1999). Considering the temporal time scale of study and the subsoil characteristics, lateral flow is hypothesized to be negligible.

The soil moisture storage (*S*) is defined as the total amount of water that is stored in the soil within the plants' root zone. The change in soil moisture storage (ΔS) is the amount of water that is being added to or removed from what is stored. ΔS falls between 0 and the field capacity on the basis of permanent wilting point (%) and root zone depth (m).In open soil systems, when the conditions are favourable for root system development, it is almost constant, also when soil humidity decreases considerably, because of an appreciable contribution of the non-rooted soil layer to the water balance. In closed soil systems (pots or shallow soils, for example) Δs is variable following the soil humidity and it begins to decrease appreciably for values of soil water reserve 60-70% of available soil water to transpiration (Tardieu and Katerji, 1991). However, in groundwater hydrology a change in groundwater level corresponds to an equivalent change in water storage and, hence, recharge (Sophocleous, 1991). Therefore, also the term ΔS in the soil water balance is directly incorporated in groundwater level observations if a suitable time scale is used to model the recharge *R*. In fact, the water storage into the soil layer above the water table depends on the time employing by the water to infiltrate which, at its turn, depends on the soil characteristics (Cherkauer, 2004). At the given time scale the water contained into the soil has two ways of escaping: the deep percolation toward the groundwater aquifer and the evaporation toward the atmosphere, favoured by the capillary rising, therefore the variation of water storage ΔS is incorporated either in the recharge itself by the ground water level or in the potential evaporation term. In conclusion, change in soil moisture storage is negligible over extended periods of time (Singhal and Gupta, 2010).

The effects of water withdrawals on the *GL* have been neglected because wells for drinking or industrial use are not present. It was also assumed that the wells for irrigation practices are not used during precipitation events.

Soil water balance, recharge modelling and the temporal scales of the process

The rainfall that reaches the surface ground may be partially discharged into episodic streams as runoff and/or partially infiltrated into the ground. The latter further percolates into groundwater aquifers storing in subsurface. The soil stores infiltrated water to become soil moisture, and then it recharges the aquifer if the soil is saturated. The rise of stored water determines deeper percolation and *ET*. Nevertheless, during rainless period soil releases slowly the water as subsurface flow (Chow *et al.*, 1988), but in karst areas the precipitated water can reach the aquifer also by channels and rock breaks. This results in a rapid and significant increase of *GL* at the precipitation events (Cai and Ofterdinger, 2016) depending on the amount and intensity of rain (Jukic and Denic-Jukic, 2009) as well as the structure of the karstic forms (width, orientation, filling material). Moreover, farmers pump water from the well for irrigating, to restore the soil water lost by the crops as actual *ET*.

In summary, by considering all terms (in mm) as input, with the sign +, and output, with the sign -, *R* (mm) of groundwater aquifer at daily scale can be written as follows:

$$R = P - R_{off} - ET - Dr - C_r \pm \Delta S \quad (8)$$

where *P* is the rainfall (here also indicated as precipitation).

In this study, the *GL* and weather variables were measured at hourly scale, but, for practical purposes, the daily scale has been used to formulate the groundwater *R* model. Measurements at daily scale of all variables were averaged starting from 24 hourly values, precipitation and radiation at daily scale were of course calculated as the sum of hourly values.

Considering the above-mentioned argumentations on each terms of Equation 8, in steady state conditions, approximated in natural ecosystems over a daily time scale (Cherkauer, 2004), the modelled recharge *R* in this study (R_m) was calculated through the following simplified equation:

$$R_m = P - PE - R_{off} \quad (9)$$

P was directly measured, the potential evaporation term, *PE,* was modelled either by the Eqs. (4), (6) or (7) and the runoff term, R_{off}, was modelled by

Eqs. (2). In this study the measured GL of the first day was considered as an offset to be added to the next measurements, so GL was equal to 0 at the beginning and, successively, could be positive or negative.
Finally, the modelled groundwater level (GL_m) is taken as:

$$GL_m = S_y \cdot R_m \tag{10}$$

RESULTS AND DISCUSSION

In this section the performances of the proposed models are illustrated together with the discussion around the justification of the chosen modelling approach for the terms of the water balance**.**
The daily runoff term is often modelled considering that the rainfall produces surface runoff when it is greater than a threshold P_0 by an amount estimated using a percentage of daily P (Lhomme and Katerji, 1991; Albergel *et al.*, 1991):

$$if\ P > P_0 \rightarrow R_{off} = \beta(P - P_0) \tag{11a}$$

$$if\ P \leq P_0 \rightarrow R_{off} = 0 \tag{11b}$$

In Mediterranean region Forest (1984), Lhomme and Katerji, (1991) and Rana and Katerji (2000) found that the values P_0 = 25 mm and β=0.15 adequately described the measured runoff at daily scale. However, the analysis of the rainfall regime in the experimental site (see following) shows that the above formulation is inappropriate mainly due to the increasing of the heavy rainfall events, therefore its formulation was upgraded to take into account the actual rainfall regime (Eqs. 2).

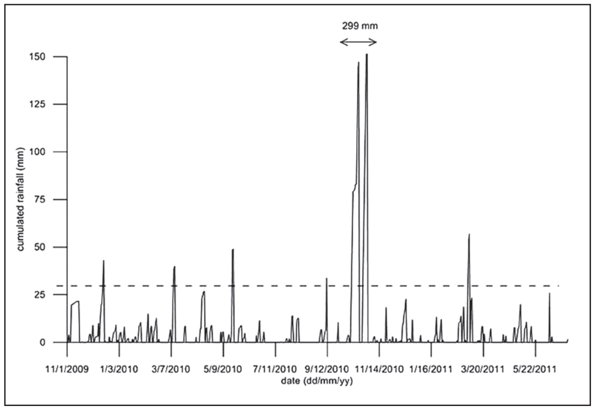

Fig. 2 - The daily cumulated precipitation values over successive rainy days.
Fig. 2 - Valori di precipitazione giornaliera cumulata.

The rainfall regime

The results of the rainfall regime trend analyses at annual scale in the experimental site are reported in Fig. 1 in terms of TSA slope and intercept and MK trend test p-value. *TP*, *WD* and *PI* (Fig. 1a, 1b and 1c) have a tendency towards a decrease, which is more marked for *TP* and significant at 0.05 for *TP* and *PI* and at 0.1 for *WD*. Instead, *HPD* shows a very light and not significant decreasing trend (p=0.513), it is high after the year 2004 (Fig. 1d), particularly during summer periods (data not shown) and merits to be further investigated. In fact, the increasing of extreme rainfall events has strong influence on the cumulate rainfall amount in the site and, consequently, it must be taken into account when the runoff term is estimated.
Going in more detail, in Fig. 2 the daily cumulated precipitation values over successive rainy days (rainfall greater than 0.2 mm) is shown focusing attention in the experimental period between

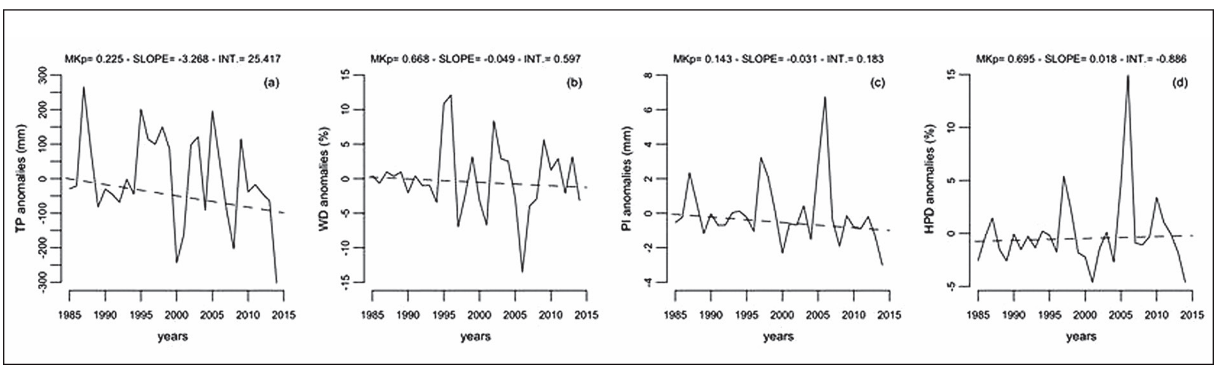

Fig. 1 - The trends of total precipitation (TP), number of wet days (WD), precipitation intensity (PI), and number of high rainy days (HPD)in terms of TSA slope and intercept and MK trend test p-value.
Fig. 1 - Trends della precipitazione totale (TP), del numero di giorni piovosi (WD), dell'intensità di precipitazione (PI) e del numero di giorni molto piovosi (HPD).

November 2009 and June 2011. Each drawn value of precipitation is obtained as

$$P = \sum_{i=1}^{n} P_i$$

where n is the number of successive days when the daily precipitation is $P_i \geq 0.2$ mm/d.This graph shows that the threshold of 25 mm (Forest, 1984) was exceed several times (5) in isolated peaks of one-two days and 2 times, reaching very high values of about 150 mm of rain for each period of 10 and 6 days respectively, in early October 2010, being equal to 299 mm in two weeks. These results have been used for modify the algorithm of the runoff model, by taking into account periods of successive heavy rainy days.

The performances of potential evaporation, runoff and recharge models

The performances of the three proposed potential evaporation models have been evaluated for selecting the best one to be employed in the groundwater R_m modelling by simplified water balance approach (Eq. 9). In Fig. 3 the cumulate values of the daily PE_P, $PE_{P\text{-}T}$ and $PE_{H\text{-}S}$ are shown for the considered period.These results show that $P_{H\text{-}S}$ model had the same path and final value as the PE_P model, the values of the cumulated evaporation being 1523 and 1550 mm, respectively. Conversely, the $PE_{P\text{-}T}$ model overestimated the potential evaporation with respect to the Penman model, despite it was locally calibrated; this over estimation is about+11%.

In the same Fig. 3 also the path of the modelled runoff is reported as cumulated values over the experimental period. The sudden increase around early October 2010 corresponds to the heavy rainy period already pointed out in the rainfall regime analysis and wouldn't be detected with the simple runoff model expressed by the Eqs. (11), without the improvement introduced by Eqs. (2), which takes into consideration the strong runoff caused by successive heavy rainy days.

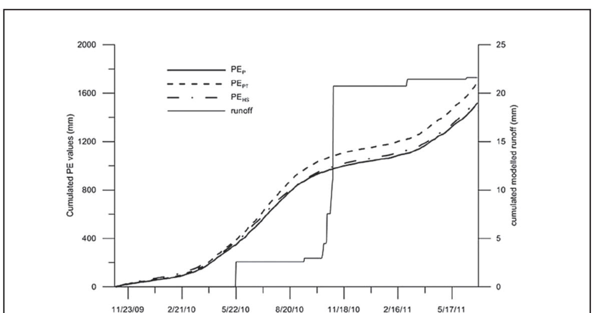

Fig. 3 - Cumulate values of the daily potential evaporation by the Penman (P), Priestley-Taylor (P-T) and Hargreaves-Samani (H-S) models together with the cumulated values of modelled runoff.

Fig. 3 - Valori cumulativi di evotraspirazione potenziata calcolati mediante i modelli di Penman (P), Priestley-Taylor (P-T) e Hargreaves-Samani (H-S) insieme ai valori simulati di deflusso superficiale.

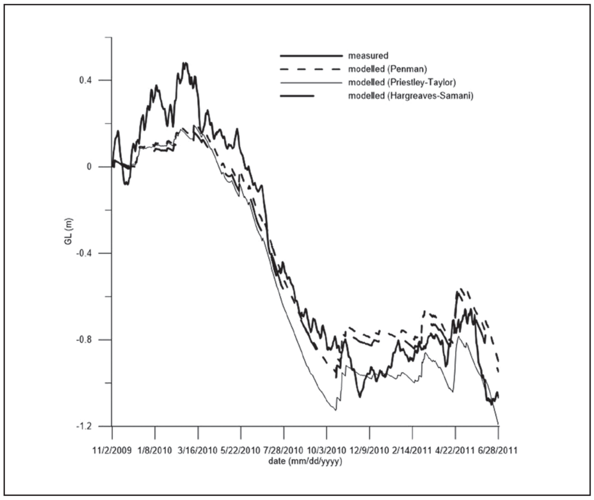

Fig. 4 - Comparison among dynamics of measured and modelled groundwater levels (GL) calculated with the three tested formulations of potential evaporation (see text for description).

Fig. 4 - Confronto tra i valori misurati e simulati di livello delle acque sotterranee determinati utilizzando i tre modelli per il calcolo dell'evapotraspirazione potenziale.

In Fig. 4 the comparison between measured and modelled GL is reported at daily scale. The modelled GL has been calculated by Eq. 10 with R_m obtained using PE_P, $PE_{P\text{-}T}$ and $PE_{H\text{-}S}$ models. The path of GL shows a general increase during winter and early spring of 2009-2010, followed by a long decrease during the irrigation season of 2010 (from March to September) and a gentle further increase until April 2011. In this case, the irrigation season started later than the previous one (March 2010), because of the water stored in the soil during the strong rainy period of fall and winter 2010-2011.

In terms of dynamics, the modelled GL_m showed acceptable performances for all adopted potential evaporation models, enough accurately following the measured GL. The linear regressions between measured and modelled GL are reported in Fig. 5 at daily scale for all PE models, with high values of r^2; the GL modelled by the Priestley-Taylor model shows the highest slope (0.92) with respect to the Penman and Hargreaves-Samani (both slopes equal to 0.77).The intercepts are

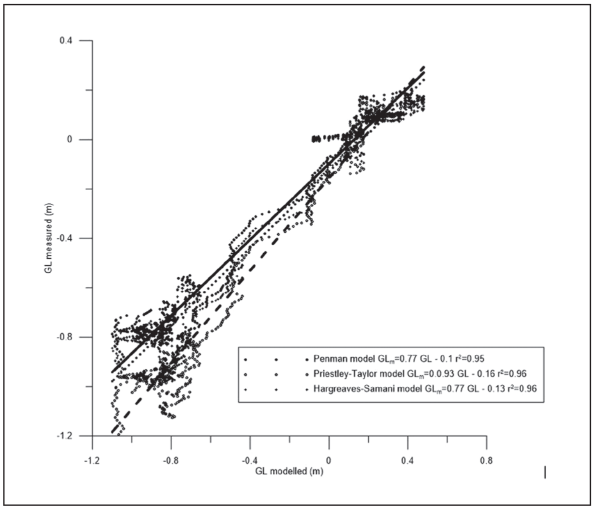

Fig. 5 - Linear regression between modelled and measured groundwater level (GL) at daily scale for the three tested potential evaporation models (see text for description).
Fig. 5 - Regressione lineare tra valori simulati e misurati di livello delle acque sotterranee a scala giornaliera con tre modelli di evapotraspirazione potenziale.

similar for all *PE* models. The best performance of the Priestley-Taylor model confirms that it is suitable for regional application because the parameter $\alpha_{P\text{-}T}$, which compensates for the lack of an aerodynamic term included in the Penman equation, can be specifically estimated for a site, as in the present case (Castelvì *et al.*, 2001).

CONCLUSIONS

In this study, to estimate the groundwater level of a karst Mediterranean aquifer a simple model of groundwater recharge was presented. The groundwater level is obtained by the product of the specific yield, taken equal at 1.7 x 10^{-3}, which is a value find by Shevenell (1996) for a karst aquifer: it is a particularly low value (Healy and Cook, 2002) given the peculiar structure of a karst subsoil with three types of structures, conduit, fracture and matrix, for the water storage (Shevenell, 1996). The model was based on the water balance approach, in a simplified form, and was validated using experimental measurements of groundwater levels of a well monitored in a semi-arid zone submitted to Mediterranean climate. The simplification of the water balance consisted in neglecting the terms of lateral drainage, capillary rising and moisture storage terms; therefore, the groundwater level was obtained by the algebraic sum of rainfall (measured), runoff (modelled) and potential evaporation (modelled).
A detailed analysis of the rainfall regime in the last 39years in the experimental area showed that, while annual amount of precipitation tends to decrease, the number of wet days with heavy rain tends to increase and that huge quantity of rain can be cumulated over short periods of 1-10 days. This fact conduced to design a runoff model which can take into account these extreme events when the groundwater recharge has to be estimated.
Furthermore, the study demonstrated that the loss of water by an agricultural area due to evapotranspiration can be estimated by the Priestley-Taylor potential evaporation model, having only the net radiation as input.
The results showed that the such modelled *GL*is well able to estimate the measured groundwater level in the experimental site of south Italy at daily scalequite accurately and can be used for applicative purposes in analogous pedo-climatic conditions.

ACKNOWLEDGEMENTS

This work was supported by Apulia Region (Italy) – "PO 2007-2013 azione 1.2.4 Azione a sostegno dei partenariati regionali per l'innovazione". The analysis of the precipitation regime was supported by the National Research Project "Collezioni E-A-OR". Funding: Ministry of Agriculture, Food and Forestry Policies (MiPAAF), Italy.

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