Research papers

Groundwater and unsaturated zone evaporation and transpiration in a semi-arid open woodland

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Abstract

Studies on evapotranspiration partitioning under eddy covariance (EC) towers rarely address the separate effects of transpiration and evaporation on groundwater resources. Such partitioning is important to accurately assess groundwater resources, especially in arid and semi-arid areas.

The main objective of this study was to partition (evaluate separately) the evaporation and transpiration components of evapotranspiration, originated either from saturated or unsaturated zone, and estimate their contributions in a semi-arid area characterized by relatively shallow groundwater Table (0–10 m deep).

Evapotranspiration, tree transpiration and subsurface evaporation were estimated with EC tower, using sap flow methods and HYDRUS1D model, respectively. To set up the HYDRUS1D model, soil material properties, soil moisture, soil temperature, soil matric potential and water table depth were measured in the area. The tree transpiration was sourced into groundwater and unsaturated zone components (0.017 mm d⁻¹ for both) and accounted for only ~6% of the evapotranspiration measured by the EC tower (~0.565 mm d⁻¹), due to the low canopy coverage in the study area (7%). The subsurface evaporation fluxes were also sourced into groundwater and unsaturated zone components using the SOURCE package, and their relative relevance in total evapotranspiration was assessed.

Subsurface evaporation was the main flux year-round (~0.526 mm d⁻¹). During late autumn, winter and early spring time, the unsaturated zone evaporation was dominant, while in dry summer the relevance of groundwater evaporation increased, reaching one third of evapotranspiration, although errors in the water balance closure point still at its possible underestimation. The results show that, in arid and semi-arid areas with sparse vegetation, the often neglected groundwater evaporation is a relevant contribution to evapotranspiration, and that water vapor flow should be taken into account in the calculation of extinction depth.

1. Introduction

In arid and semi-arid areas the scarce water resources are usually stored as groundwater. While the precipitation in the Mediterranean region is decreasing (Gualdi et al., 2013; Mariotti et al., 2015), the demand for the limited groundwater resources is increasing (Scanlon et al., 2006). In such conditions it is imperative to quantify the main water input constraining groundwater resources, i.e. the net groundwater recharge (Chenini and Ben Mammou, 2010). The evapotranspiration of groundwater resources is often underestimated, both because evaporation processes are not yet included in the theory (Zeng et al., 2011a) and because transpiration from roots tapping the water table is not taken into account (Favreau et al., 2009; Miller et al., 2010). The underestimation of groundwater evapotranspiration often results in the overestimation of the net recharge. Therefore it is critical to define accurately groundwater evapotranspiration which may represent a small but relevant reference percentage of total evapotranspiration.

There are various methods to estimate groundwater recharge, but it is difficult to know which one is the most reliable (Scanlon et al., 2002); for routine recharge estimation the best choice is a soil moisture recharge technique, provided all the important physical and physiological processes are represented adequately (Rushton et al., 2006). For example, water infiltrating in the soil after sparse rain events often evaporates completely before
reaching the water table (Tweed et al., 2011); another example is that of the transpiration of water from trees, which can tap the water table with their roots (Favreau et al., 2009; Miller et al., 2010). A good knowledge of evapotranspiration processes is, therefore, fundamental to sustainable agriculture and groundwater management, particularly in water scarce environments.

During dry seasons, mainly in arid and semi-arid locations, high evapotranspiration (ET) quickly depletes the unsaturated zone water, exposing the saturated zone to groundwater evapotranspiration (ETg) or, following the partitioning concept, to the processes of groundwater evaporation (Eg) and groundwater transpiration (Tg). The process of separating saturated from unsaturated zone fluxes is hereafter referred as sourcing.

ET is composed of two different processes, the physical process of evaporation and the biological process of transpiration, the two having different spatial and temporal characteristics (Lubczynski, 2009). ET includes three components: surface evaporation (Es), which includes evaporation from surface water (rivers, lakes and other surface water bodies) and from water intercepted by plants; evaporation of water from below the ground surface (Esw, subsurface evaporation) (Kampf et al., 2005); and transpiration of water by plants (Tu, Scott et al., 2003; Miller et al., 2010), together defined as subsurface evapotranspiration (ETs). The process of separating different components of ET is referred here as partitioning. This study presents an experimental, field based example of such partitioning and sourcing.

In this paper, we followed the terminology of Lubczynski and Gurwin (2005):

\[ ET = E_s + ET_{ss} \]  
(1)

\[ ET_{ss} = E_{ss} + T_s = ET_u + ET_g \]  
(2)

\[ ET = E_s + (E_u + T_u) + (E_T + T_g) \]  
(3)

ETs is groundwater evapotranspiration (ETg = Eg + Tg) where Es and Tg are the respective evaporation and transpiration components depleting groundwater; ETu is the unsaturated water evapotranspiration (ETu = Eu + Tu), where Eu and Tu are the respective evaporation and transpiration components depleting the unsaturated zone (all in L T\(^{-1}\) units). Subsurface evaporation is Ess = Eu + Eg and plant transpiration is Tu = Tg + Tg.

In recent years, interest in ETss partitioning into ETu and Ess has increased due to the development of new monitoring methods, which permit independent measurement of various water fluxes at different time scales and under different climatic conditions (Zhang et al., 2016). Both the Bowen ratio (only in wet conditions) and the eddy covariance (EC) methods (Perez et al., 1999) permit reliable measurement of latent heat flux (and therefore ET) but over relatively small areas and only in specific conditions (e.g. stable conditions for Bowen ratio). Thanks to so called footprint models, the area sampled by both Bowen ratio and EC methods can be determined with large spatial and temporal precision. Besides, is it possible nowadays to estimate Ess from models applying semi-continuous soil moisture and matric potential profile measurements (Kizito et al., 2008), while Es can be estimated from pan evaporation and measurement of tree interception using tipping buckets and gutters placed under a tree canopy (Ghimire et al., 2012; Ghimire et al., 2017) and by interception models. Finally, in situ sap flow measurements (Granier, 1985) can be used to estimate tree transpiration (Tss) when using appropriate sampling, measuring and upscaling techniques (Granier, 1987; Reyes-Acosta and Lubczynski, 2013; Reyes-Acosta and Lubczynski, 2014). In this study we propose to combine all these techniques to determine the contributions of Eg, Ess and Tg to the total ET for different land covers.

The partitioning of ET is particularly effective in landscapes where few plant species are present and individual tree canopies can be identified. In these landscapes Tg, is restricted to trees and can be defined using sap flow measurements, while Eg is the only water output in the bare soil areas outside tree root influence. Tree root influence area is understood as the area where chemical and physical conditions (including water pressure head) are influenced by the presence of roots. Wallace (1997) presents an example of an ET partitioning study in such a landscape; ET was measured by means of EC over an area with a clear heterogeneity of land cover, i.e with patches of woodland and bare soil.

The sap flow technique does not work for grass. Whenever grass is present and active another technique is required to assess grass transpiration, either by modeling (Feddes et al., 1978) or by direct measuring, for example using a gas chamber (Yepez et al., 2005). Sometimes, however, in ET partitioning studies, grass transpiration is lumped together with subsurface evaporation and both regarded as the difference between the EC tower measurements of ET and the sap flow measurements of tree transpiration (Paço et al., 2009).

No groundwater influence on ET rates has been a common assumption in partitioning studies, although groundwater evapotranspiration can be a substantial component of water balance, particularly in dry conditions, which, when not taken into account, can result in underestimation of total ET. In Wilson et al. (2001) the depth at which soil water was supposed to play a negligible role in total ET was set to 0.75 m below the ground surface (b.g.s.), probably due to the wet climate characteristic of the area studied. Baldocchi et al. (2004) showed the typical condition of a semi-arid, open woodland landscape: a poorly developed soil lying on top of a fractured granite bedrock regarded as non-evaporating (no information on water table depth was given in the article); in that case all subsurface evaporation was assumed to come from the unsaturated zone while the groundwater contained in the bedrock was assumed to be non-evaporating. Yaseef et al. (2010) defined bare soil Es as 36% of ET within a year in a semi-arid climate, assuming zero Eg because of the ∼300 m b.g.s. groundwater table.

Williams et al. (2004) was conducted in a periodically irrigated semi-arid area olive orchard (400 trees ha\(^{-1}\); water table depth, soil composition and type of bedrock are not provided). When the top soil layer was irrigated it was moist enough to meet the potential evapotranspiration demand (ETp) without affecting soil moisture in the deeper soil profile; in contrast, when the soil was dry long after irrigation, Ess was assumed to be negligible (ETs = Tss). The study used the isotopic method (Zhang et al., 2010) to partition the ET sources: water evaporated from the soil is depleted in the heavy isotopes compared to the water transpired from leaf surfaces; therefore the analysis of water vapor collected at the EC station gives indication of the individual contributions of Ess and Tss. The tree sap flow measurements (Tss) underestimated the EC measured ETs by 24% in the period before irrigation. The authors assumed that this was due to their sap flow method underestimating the total sap flow of the trees, based on the fact that the isotopic partitioning showed no soil evaporation for the period before irrigation. To overcome the mismatch, the Tu calculation from sap flow measurements was re-calibrated based on the EC measured ET.

The relevance of groundwater evapotranspiration, and hence the importance of sourcing, is supported by recent studies. For example, groundwater uptake from tree tap-roots was studied by Miller et al. (2010) using the water table fluctuation method (Loheide et al., 2005) in an oak savanna located in the western Sierra Nevada foothills. Tree canopy cover was ∼40% of the studied
area and water table was at ~8 m depth b.g.s. Their results showed that groundwater uptake contributed up to 90% of total ET during the dry season. However, the water table fluctuation method applied by Miller et al. (2010) could not distinguish between root groundwater uptake (Tg) and soil groundwater evaporation (Eg). Hence they assumed that Eg was negligible, attributing groundwater discharge entirely to Tg.

Recent studies demonstrated that our understanding of the processes of subsurface evaporation in dry conditions is still incomplete, as it is probably underestimated (Soylu et al., 2011; Smits et al., 2011; Zeng et al., 2011b; Li et al., 2010), mainly due to underestimation of groundwater evaporation (Balugani et al., 2016). It is argued, for example, that vapor flow should be taken into account for a correct estimation of Eg (Saito et al., 2006; Bittelli et al., 2008; Y.J. Zeng et al., 2009). In partitioning studies, the kind of soil model typically used was a bucket model (Miller et al., 2010), but the evaporation extinction depth (the depth at which the evaporation from a water table becomes negligible) should be calculated using transient flow models, as argued by Shah et al. (2007). However, even the HYDRUS1D model used by Shah et al. (2007) did not include heat and vapor flow. Vapor flow through the “almost dry” unsaturated zone can result in even deeper evaporation extinction depths than defined by Shah et al. (2007), meaning that Eg fluxes could be greater than estimated so far. To our knowledge, no studies performed both partitioning and sourcing of ET to investigate the separate contributions of Eg and Tg in the overall water budget, taking into account the heat and water vapor flow in the calculation of subsurface evaporation.

The research questions of this study are: What is the contribution of Eg and Tg to the total ET in the investigated semi-arid, open woodland area characterized by shallow water table? How does Eg change in time? To answer these questions, the following objectives were defined:

1. experimentally partition ET measured with EC method into: (i) Eg, modeled using HYDRUS1D model applying heat and water vapor flow; and (ii) Tg, estimated by upscaling of sap flow measurements;
2. source Eg into Eg and Tg, and source Tg following the protocol of Reyes-Acosta et al. (2015);
3. evaluate how the relative relevance of Eg versus Tg varies in time in a semi-arid open woodland with a shallow groundwater table (Balugani et al., 2016).

These objectives provide the structural sub-headings used in the following Materials and Methods, Results and Discussion sections.

2. Materials and methods

2.1. Study area

To study partitioning and sourcing of ET in semi-arid conditions we selected a 2 × 2 km study area, further referred as the maximum footprint area (MF, Fig. 1), enclosing the maximum extent of all the footprints of the EC tower installed in the center of that area. The MF is characterized by shallow water table and open woodland vegetation. The selected area is located in the northern part of the Sardón Catchment, in the central-western part of the Iberian Peninsula, west of Salamanca, Castilla y León (Spain; latitude: 41.1172°; longitude: −6.1471°). That area has already been investigated by Lubczynski and Gurwin (2005), Mahmoudzadeh et al. (2012), Reyes-Acosta and Lubczynski (2013, 2014), Franch et al. (2014), Hassan et al. (2014), and Reyes-Acosta et al. (2015).

The semi-arid climate of the MF area is typical of the central part of the Iberian Peninsula, with a mean yearly precipitation of 586 mm year−1 (1951–2012, Hassan et al., 2014) and most of the rain events concentrated in spring (March-May, DOY 75–135) and fall (October-December, DOY 285–345); the difference in precipitations between years can be high, with driest years ~300 mm year−1 and wettest years ~900 mm year−1. An average summer (the dry season, June-September, DOY 165–265) is characterized by: −20 °C temperature, ~5 mm d−1 ETp, and ≤20 mm month−1 mean precipitation, while an average winter by: −5 °C temperature, ~0.5 mm d−1 ETp and ≤100 mm month−1 mean precipitation. The study was conducted in the years 2009 and 2010.

The morphology of the MF area is typical for the Spanish Meseta: a rugged, savanna-like (dehesa) landscape with gentle slopes (mean slope = 2°; Fig. 2). The local geology is dominated by weathered and fractured granite (Lubczynski and Gurwin, 2005). The land cover consists of open woodland and sparse grasses that are green only for a short period between early spring and early summer (i.e., April-June) and which are predominately consumed by livestock while green so there are minimal senesced grasses present during the remainder of the year. Two tree species are dominant: evergreen oak (Quercus ilex) and broad-leaved deciduous oak (Quercus pyrenaica). Both are able to extract water not only from the unsaturated zone but also from the saturated zone using tap roots (David et al., 2007; Reyes-Acosta and Lubczynski, 2014).

The water table is shallow: 0–3 m b.g.s. in the valleys and 5–10 m b.g.s. in the hills. The water table change has a yearly amplitude of ~2 m (Hassan et al., 2014), shallowest at the end of spring (March-April) and then getting deeper when the rain stops (end of May), reaching a maximum depth at the end of dry summer—beginning of autumn (September-October), when the precipitation events become frequent enough to start to recharge the aquifer again. The area is sparsely populated (~1 p km−2) and the main activity, i.e. non-intensive farming, has low impact on the ground-water levels (Lubczynski and Gurwin, 2005).

The soil sampling campaign demonstrated that the soils in the MF area are the result of the weathering of the underlying granitic rocks. The thickness of the soil changes spatially, thinner in the upland (0.5–1 m) and thicker in the valleys (>3 m). The soil texture does not change spatially in the area; it is essentially sandy (~80%) with some gravel (percentage varies with depth) and silt (~10%).

The remote sensing (Quickbird and WorldView-II images) analysis of the MF area, executed following the protocol described in Reyes-Acosta and Lubczynski (2013) for the whole Sardón catchment, showed that the canopy coverage in the MF study area was ~7%. The two Quercus spp. species had nearly equal canopy coverage; however, Quercus ilex species was predominant in the eastern, elevated part while Quercus pyrenaica in the western, lower part (the Sardón river valley, Fig. 1).

2.2. Partitioning and sourcing framework

To partition and source ET, we: (i) installed an EC tower in the MF area to measure ET; (ii) measured sap flow in the trees to estimate Tp; and (iii) monitored microclimatic conditions, soil moisture, matric potential and water table depth (WTD) to setup, calibrate and validate a soil parameterization in the HYDRUS1D model (Šimunek et al., 2009). The ET measured by the EC tower is a point measurement (taken at the top of the tower); this measurement can be related to actual ET at ground level by a probability density function, called the EC tower’s footprint (Göckede et al., 2004). The footprint of an EC tower changes its size, position and shape in time, depending on wind strength and direction; an example of the EC tower footprint is shown in Fig. 2.

In order to obtain and partition ET, i.e. to estimate Eg and Tg separately and source each of them in the footprint area, we:
1. calculated the EC tower footprint every 30 min (Section 2.3.1);
2. determined the $2 \times 2$ km $MF$ study area (Fig. 1);
3. divided the $MF$ area into two categories (Fig. 2): (i) the bare soil outside the ground projection of tree canopies where we assumed that only $E_s$ was taking place, and (ii) the area corresponding to ground projection of tree canopies where we assumed that only $T_s$ was taking place;
4. upscaled tree sap flow measurements to obtain maps of $TMF$ (estimated $T_s$ in the $MF$ area) every 30 min (Section 2.3.2, following the protocol described in Reyes-Acosta et al., 2015);
5. modeled $E_s$ using the HYDRUS1D model based on soil measuring profiles and water table measurement in the area and upscaled it to obtain maps of $EMF$ every 30 min;
6. added $TMF$ and $EMF$ maps together to obtain $ETMF = EMF + TMF$ and multiplied them by the probability density function of the footprint to obtain an estimate of $ETMF$ in that particular footprint to be compared with EC tower evapotranspiration ($ETec$).

In order to source $E_s$ and $T_s$ to estimate $E_g$, $E_u$, $T_g$ and $T_u$ we:

1. used the SOURCE package (Balugani et al., 2016) and upscaled it to obtain maps of $E_g$ and $E_u$ every 30 min (Section 2.3.3); and
2. sourced $T_s$ using $T_g$ and $T_u$ estimates by Reyes-Acosta et al. (2015), to obtain $T_g$ and $T_u$ maps every 30 min.

As in the partitioning part, the four maps of $E_g$, $E_u$, $T_g$, and $T_u$ were multiplied by the probability density function of the footprint, to obtain estimates of the individual sourced components and compare their sum with $ETec$. The whole process is shown as a flowchart in Fig. 3.

The assumption that tree root water uptake is limited to the ground projection of the canopy area is based on the simplifying assumption that the tree roots are limited to that area (Reyes-Acosta and Lubczynski, 2013). That assumption was justified by direct observation while digging the area for soil sample collection and soil moisture profile preparation. As the tree canopy coverage in the study area was low and the experiment was carried out in dry summer seasons of 2009 and 2010 when grass was dormant and rains rare, the convenient assumption of negligible $E_s$, mainly due to interception, was justified. Also due to the very short vegetative period of the sparse grasses in the area (March-May, DOY 60–140), and continuous cattle grazing, the assumption of negligible grass interception and transpiration is justified.

2.3. Techniques used

2.3.1. Evapotranspiration measured by EC tower

We used an EC tower to measure $ETec$. We installed the EC system on the 10 m high tower located on an elevated point in the northern part of the Sardón catchment, within an oak open woodland landscape (Fig. 1; trees average height is 4–5 m). The system consisted of:

1. CNR1 four components net radiometer (Kipp and Zonen, Delft, The Netherlands).
2. CSAT3 sonic anemometer (Campbell Scientific Inc., Utah, USA).
3. LI7500 gas analyzer (Licor Biosciences, Nebraska, USA).
4. WXT520 “multi weather sensor” for measurements of wind speed, direction, air temperature and humidity (Vaisala Oyj, Helsinki, Finland).

The tower was also equipped with two soil heat flux plates and 19 soil temperature sensors for the energy balance closure. The data were processed with the software AltEddy (www.climatexchange.nl/projects/alteddy/) of the Alterra Institute (Wageningen University, The Netherlands; van der Tol, 2012). The footprint...
When evaluating the peaks of evapotranspiration after rainfall, the question arises how reliable the EC data are during and shortly after rainfall, as heavy rainfall may affect the CSAT3 measurements when droplets block the sound transmission (Munger et al., 2012). According to the LI7500 manual, droplets on the infrared gas analyzer may also hamper the gas concentration measurements.

Heusinkveld et al. (2008) quantified the effects of dew on the LI7500 sensor by comparing measurements of two devices, one heated and one not heated, as in the case of our instrument. Although we took care to minimize the effects of droplets by mounting the LI7500 tilted, the applied quality filtering still resulted in missing evapotranspiration data in wet conditions. For our analysis it was important to assess whether such data gaps affected the estimate of the daily average ET. If failure of the EC technique coincided with a high evapotranspiration rate, then the daily average $ET_{ec}$ underestimated real ET. To verify whether this was the case, first we analyzed the fraction of half-hourly data that was flagged as high quality versus the number of hours and days since last rainfall. This procedure showed how the quality of the data was correlated with sensor wetness. Second, we assessed how sensor wetness was correlated to $ET_{ec}$, by plotting $ET_{ec}$ versus hours since last rainfall. The two correlations together indicated how reliable daily $ET_{ec}$ were.

### 2.3.2. Tree transpiration estimated with sap flow

To quantify the contribution of tree transpiration to ET, we took sap flow measurements during the months of August and September in 2009 and in 2010 using the optimization method described in Reyes-Acosta and Lubczynski (2014), which combines thermal dissipation probes (TDP) measurements (Granier, 1985) with heat field deformation (HFD) measurements (Nadezhdina et al., 1998; David et al., 2007). The optimization method consisted of three steps: (i) obtaining the TDP sap flow measurements using a standard TDP system (UP Gmbh, Germany) in cyclic switching power mode and correcting them using the cyclic heat dissipation (CHD) method (Lubczynski et al., 2012; Reyes-Acosta et al., 2012) to eliminate the natural thermal gradient (NTG) bias; (ii) correcting the TDP measurements for night flow using the HFD measurements (ICT international, Armidale, NSW, Australia); and (iii) correcting the sap flow data for radial and circumferential...
heterogeneity using multiple HFD measurements (Reyes-Acosta and Lubczynski, 2014).

To assess the sources of water transpired by trees (\( E_{\text{p}} \) or \( E_{\text{t}} \)), we applied the methodology described by Lubczynski (2009) to experimental data described in Reyes-Acosta et al. (2015). Constant doses of deuterated water were injected during 24 h into the groundwater using a network of piezometers drilled around the trees, following Brooks et al. (2002). This setup maximized the plume shape and the chemical dispersion through the soil. During and after the deuterated water injection, we used sap flow sensors to monitor \( T_{\text{sa}} \) and collected xylem water, soil water and groundwater samples to assess the sources' contributions to \( T_{\text{sa}} \) (Reyes-Acosta et al., 2015).

To upscale the transpiration estimates for the MF area we: (i) established the upscaling functions for each tree species using the procedure explained in Reyes-Acosta and Lubczynski (2013) after digital mapping of the trees in the study area using a QuickBird image as reference; (ii) upscaled the 30 min sap flow measurements into whole-tree scale and to the investigated footprint area; (iii) extrapolated the results of the sap flow measurement campaign temporally into the whole dry summer period applying an artificial neural network model (Liu et al., 2009); and (iv) sourced sap flow \( T_{\text{sa}} \) into \( T_{\text{u}} \) and \( T_{\text{g}} \) using deuterium tracing experiments (Reyes-Acosta and Lubczynski, 2011; Reyes-Acosta et al., 2015) for the whole dry summer period (according to the measurements, performed during dry summer campaigns of 2009 and 2010, Reyes-Acosta and Lubczynski, 2011).

2.3.3. Evaporation modeled with HYDRUS1D

To model bare soil \( E_{\text{b}} \), we characterized spatially the soil properties, the water table depth and the soil moisture in the MF area from 2009 to 2015. 31 locations were sampled at multiple depths to characterize the soil texture variability. Also double infiltrometer tests were conducted in 7 locations to determine the vertical hydraulic conductivity in the MF area. To investigate the groundwater, we measured water table levels in 6 shallow piezometers installed within the MF area and 4 others outside the MF area but close to its borders (Fig. 1; one of them penetrated the bedrock). To measure soil water content at three locations, multisensor Hydrolprobes (Stevens Water Monitoring Systems, Inc.), were installed at four different depths (25, 50, 75, 100 cm b.g.s.). We also installed Decagon matric potential sensors at two different depths in each of the three above-mentioned profiles at 25 and 75 cm b.g.s. and polymer tensiometers (van der Ploeg et al., 2008) at 15 cm b.g.s. in two of the three profiles. In addition to the EC tower, a standard, 2 m high \( E_{\text{p}} \) weather station was available as a backup at distance of ~100 m (Fig. 1).

We modeled the soil water fluxes in the monitored profiles with the HYDRUS1D code (Simunek et al., 2008) using the soil physical parameters, the measured water levels and the measurements from \( E_{\text{p}} \) weather station. We used the HYDRUS1D version which solves the equation for the coupled flow of heat, liquid water and vapor water (Saito et al., 2006) in order to account for evaporation by water vapor flow to the surface. The use of a 1D model for the unsaturated zone is justified due to the low relevance of the lateral water flow along the small topographic gradients in the study area.

In each of the three HYDRUS1D simulated soil moisture profiles, we set the measured WTD as bottom boundary condition, the measurements from the weather station (net radiation, air humidity, air temperature, precipitation, wind speed, soil temperature) as top boundary condition and we applied the soil physical parameters as determined from the soil samples and from the infiltrometer tests. The model was calibrated against the soil temperature, soil matric potential and soil moisture monitored in the three profiles (for more information about the calibration, refer to Balugani et al., 2011). During the calibration period, the soil was completely bare (no grass present). Finally, in 2011 we validated the HYDRUS1D model performing an infiltration experiment on one of the monitored profiles.

We then used the validated HYDRUS1D parameters to model \( E_{\text{w}} \) for the various water table depths and depths to bedrock in the area. First, we created maps of water table depth (bottom boundary condition for the model) and soil depth (depth of the modeled profile) with 1 m\(^2\) precision. Soil hydraulic properties were found to be homogeneous in the MF area, so no map was created for them. Second, we grouped every pixel in the area with the same soil depth and WTD in different sets and we ran HYDRUS1D for each of these sets, using: (i) the measured evaporative conditions as upper boundary conditions for all the sets; (ii) the soil depth in that set for the model geometry; (iii) the same soil hydraulic properties for all the sets; and (iv) the WTD maps from Hassan et al. (2014) to determine the bottom boundary condition of each set.

\( E_{\text{b}} \) calculated by HYDRUS1D was subsequently sourced into \( E_{\text{g}} \) and \( E_{\text{u}} \) using a post-processing package called SOURCE (for the algorithm used, refer to Balugani et al., 2016). The SOURCE package analyzes the output files of HYDRUS1D model containing information about soil moisture and water fluxes at all nodes in the modeled profile and fluxes at the model boundaries. This information is then used to make a water balance for the soil profile to determine \( E_{\text{g}} \) and \( E_{\text{u}} \). Therefore, the definitions of \( E_{\text{g}} \) and \( E_{\text{u}} \) in the SOURCE package are formulated from a water balance point of view: if \( E_{\text{u}} \) is a loss of water stored in the profile, then it is investigated which zone of the soil (the saturated or the unsaturated zone) was affected by that loss. In other words, \( E_{\text{g}} \) is defined as “the evaporative flux corresponding to the decrease of water stored in the saturated zone (groundwater) due to loss of water vapor at ground surface” while \( E_{\text{b}} \) is defined as “the evaporative flux at the ground surface corresponding to the decrease of water stored in the unsaturated zone” (Balugani et al., 2016).

3. Results

3.1. Partitioning of ET

3.1.1. Evapotranspiration, transpiration and evaporation estimates

During the 2009–2010 study period we observed typical patterns of dry summer and winter rainfall and \( E_{\text{p}} \), as described in Section 2.1; the yearly precipitation for the two study years differed widely: ~300 mm \( \text{y}^{-1} \) in 2009 and ~700 mm \( \text{y}^{-1} \) in 2010. In the dry summers the rainfall events were rare, but, if present, of high intensity (e.g. the two events at the end of the dry summer 2009 and 2010, both of ~50 mm \( \text{d}^{-1} \), Fig. 4), while \( E_{\text{p}} \) was high, up to ~7 mm \( \text{d}^{-1} \) at the peak of the dry summer 2010. During winter, the rainfall events were more frequent but less intense, while \( E_{\text{t}} \) conditions were low (~2 mm \( \text{d}^{-1} \), Fig. 4), due to low temperatures, low radiation and high relative humidity (not shown).

The tree transpiration had the same magnitude in the two dry summers of 2009 and 2010 (Fig. 5, also Reyes-Acosta and Lubczynski, 2013) despite of differences in rainfall and WTD. The very low value of the upscaled, dry summer tree transpiration (\( T_{\text{sp}} = 0.03 \text{ mm d}^{-1} \)) is due to the low tree density (canopy coverage ~7%). During dry summers, \( E_{\text{t}} \) was higher and the unsaturated zone was mostly dry, but the tree water demand was satisfied by groundwater uptake (Reyes-Acosta et al., 2015).

The soil moisture dataset from one of the three measured profiles is shown in Fig. 6 together with the WTD recorded in a piezometer placed at ~3 m distance; all other datasets showed similar patterns. During the dry summer periods (DOY 200–300 of year 2009, 19 July to 27 October, and DOY 150–260 of year...
2010, 30 May and 17 September) the soil moisture remained low, decreasing slowly while drying. At that time there was no recharge and the water table was slowly decreasing due to the combined effect of groundwater outflow to the stream and groundwater evapotranspiration ($ET_g$). The first rain events started to moisten the soil profile already around DOY 280 (19 July) of year 2009 and DOY 260 (17 September) of year 2010, until moisture exceeded field capacity and groundwater recharge took place (DOY 357 of year 2009–23 December); at that time the soil moisture at 50, 75 and 100 cm b.g.s. reached a value close to saturation due to the high water table, while the soil moisture at 25 cm b.g.s. only occasionally showed an increase due to infiltrating rain.

3.1.2. Comparison of $ET_{ec}$, $E_{EMF}$ and $T_{MF}$

The time series of $ET_{ec}$, $E_{EMF}$ and $T_{MF}$ for the year 2009 and 2010 are shown in Figs. 5 and 7. Rainy days show poorer quality data and higher $ET_{ec}$ than dry days, but the problem of poor quality data was limited to the first 90 min after rainfall (for the procedure followed to check data quality, refer to Sections 2.3.1 and 4.2.1).

$E_{EMF}$ followed the trend of $ET_{ec}$ (Fig. 7), especially in the winter period, so they are described together. In general, after $ET_{ec}$ and $E_{EMF}$ reached a maximum in April 2010 (DOY 110–120, 20–30 April, Fig. 7), they started to decrease again in late spring, reaching values of less than $1\text{ mm d}^{-1}$ in the first week of September (DOY 240–250, diamond symbols in Fig. 7), while $ET_p$ continued to increase significantly up to 23 August 2010 (DOY 235, Fig. 4). In 2010 $ET_{ec}$ and $E_{EMF}$ started to increase again with the first rain events during the end of the dry summer (see Fig. 7 for DOY 260–270, 17–27 September 2010).

Both $ET_{ec}$ and $E_{EMF}$ followed the pattern described by Hillel (1998), i.e. just after a rain event the soil was wet and $ET_{ec}$ and $E_{EMF}$ rates were determined by $ET_p$ conditions (first stage evaporation, Hillel, 1998), e.g. in Fig. 7 after the frequent rain events in the period between DOY 30 and 110 2010 (30 January – 20 April); later on, the first centimeters of soil became dry and evaporation dropped steadily (second stage evaporation, again in Fig. 7 DOY 90 and 190 2010 or 31 March – 9 July). During late spring and summer (May-October, DOY 130–280), when the rain events were
sparse and of high intensity, the $ET_{ec}$ and $EMF_{ss}$ differed: (i) during the first stage evaporation $EMF_{ss}$ was higher than $ET_{ec}$ (e.g. in Fig. 7, the period between DOY 150 and 180 in 2010, 30 May and 29 June); and (ii) in the late, asymptotic part of the second stage of evaporation in dry summer, when the long periods without precipitation events allowed the soil to dry up, resulting in low and almost steady evaporation rates (Fig. 7, period between DOY 90 and 200 of 2010, 31 May and 19 July, during this period $EMF_{ss}$ was lower than $ET_{ec}$).

$TMF_{ss}$ values were very low with respect to overall $ET_{ec}$ (Fig. 7, the thick dashed line at the bottom of the plots). As explained before, this was due to the low tree canopy coverage in the $MF$ area. $TMF_{ss}$ estimated for the late spring-summer period was at its maximum because of large moisture availability and maximum $ET_p$ in that period, allowing both tree species, $Q. ilex$ and $Q. pyrenaica$, to transpire at the potential rate.

### 3.2. Sourcing of $EMF_{ss}$ into $E_g$ and $E_u$

The time series of the sourced components of $EMF_{ss}$ show (Figs. 8, 2009) that: the $E_g$ was above the $E_u$ during the long period of drought in between the two rain events, when the groundwater table continuously declined (Fig. 6); the $E_u$ was below the $E_g$ during rain events and shortly after because water that infiltrated in the top soil started to evaporate quickly before reaching the “deepest drying front” (the depth at which infiltrating water moves downward faster than it evaporates, Y. Zeng et al., 2009). Hence, the water infiltrated after the dry summer rain events did not recharge the water table. During the wet winter and spring of 2010, when the soil moisture was above field capacity due to frequent rains, the agreement between $ET_{ma}$ ($ET_{ma} = E_{ma} + TMF_{ma}$) and $ET_{ec}$ was very good with only 1% difference. The $EMF_{ss}$ was the main contributor to $ET_{ec}$ (69%) while the $E_g/EMF_{ss}$ ratio was 0.35 and the $T_{ma}/TMF_{ma}$ ratio was 1. The tree transpiration was a minor contributor to $ET_{ec}$, 6%. In the dry summer 2010 (the “wet year” with rainfall of ~700 mm y$^{-1}$ rainfall), $ET_{ma}$ was higher than $ET_{ec}$ (0.685 mm d$^{-1}$ and 0.518 mm d$^{-1}$ rainfall, respectively), $EMF_{ss}$ was, again, the main contributor to $ET_{ec}$ (93%) while the evaporation and transpiration sourcing ratios were the same as in 2009; the tree transpiration was similar (6% of $ET_{ec}$) as well.

In a similar analysis carried out between 30 January (DOY 30) and 7 October (DOY 280) 2010 (Table 1), we excluded the tree transpiration due to the lack of sap flow measurements in winter and spring periods and also the days when the quality of the EC measurements was low. The results show: generally higher abso-
lute values of evaporative fluxes than in dry summer; the $E^\text{MF}_s$ lower than $E^\text{TSS}_s$ by 24%; and an $E^\text{MF}_s/E^\text{TSS}_s$ ratio (0.22) significantly lower than comparable ratios in the dry summer, emphasizing lower $E^\text{MF}_g$ contributions to $E^\text{ETMF}_s$.

The uncertainty associated with $T^\text{MF}_m$ and $E^\text{MF}_s$ are ~16% of $E^\text{ETMF}_s$. Error in $T^\text{TSS}_s$ was less than 1% (Reyes-Acosta and Lubczynski, 2013) while for the evaporation estimates, the uncertainty in WTD and soil hydraulic properties resulted in an error of ~15% of the total measurement. This value was calculated using the standard deviation of the mean values for the soil hydraulic properties (which are calculated using all the samples taken in the area) and the uncertainty in the water table level calculated by the distributed model of Hassan et al. (2014): the HYDRUS1D model was run for all possible soil hydraulic properties and WTD combinations (e.g. maximum hydraulic conductivity, lower WTD) and we used the results to establish the uncertainty of the $E^\text{TSS}_s$ calculation.

### 4. Discussion

#### 4.1. ET partitioning in semi-arid areas

In arid or semi-arid areas, especially those covered by open woodlands, the tree $T_u$ component of ET, next to climatic and water availability constraints, is highly dependent on tree species type, density, and size of canopy area, i.e. tree characteristics detectable from space (Lubczynski, 2009; Reyes-Acosta and Lubczynski, 2013). As “open woodland” can have a canopy coverage ranging from 5% to even 80% (Anderson et al., 1999), the $T_u$ component of ET can vary substantially between different land covers. However, in an open woodland, trees never get “crowded”, and are limited in their growth by water or nutrient availability and competition between different individuals.

In the MF study area, with canopy coverage of 7%, and with a “shallow” water table, $T^\text{MF}_s$ accounted for only 6% of $E^\text{TSS}$. For comparison, the Table 2 presents a summary of the results of various partitioning studies performed in open woodlands with climate similar to the MF study area, but with different canopy densities. Yaseef et al. (2010), who directly measured $E_u$ and $T_u$ in an area with Mediterranean climate (Israel) with ~60% canopy coverage, found yearly values for $E_u$ and $T_u$ of 33.8% and 42.3% of total ET, respectively. In an olive field in Morocco with high tree density (400 stems ha$^{-1}$, no information on canopy coverage), Williams et al. (2004) estimated $E_u$ after irrigation as high as ~20–30% of $E^\text{TSS}$ and $T_u$ as 80–70% ($E_u$ was assumed negligible). In these two examples, a higher tree density and canopy coverage compared to our study resulted in a larger contribution of $T_u$ to ET (although $E_u$ still represented ~1/3 of ET), suggesting $T_u$ dependence on canopy coverage.

However, there are studies that seem to contradict the relationship between $T^\text{MF}_s$ and canopy coverage. For example, Paço et al. (2009), with a canopy coverage of 21%, found $T^\text{MF}_u$ = 100% of ET (and $E_u = 0$) during a dry summer period with a setup similar to the one presented in this study (the area studied was the maximum footprint area for a 28 m high EC tower). Their finding of $E_u = 0$ might have been due to experimental design. They measured ET with an EC system and estimated $T_u$ with sap flow measurements while $E_u$ was assumed to be $E_u = ET - T_u$. To estimate $T_u$ they used the Granier sap flow measurement method (Granier, 1985) which, in sparse tree environments, is highly vulnerable to overestimation errors if the analysis does not account for: (i) natural temperature gradient (Lubczynski et al., 2012; Reyes-Acosta et al., 2012); and (ii) radial variability of the sap flow as discussed by Reyes-Acosta and Lubczynski (2014). Paço et al. (2009) did not explain whether they dealt with these potential problems. The overestimation of $T_u$ could explain the underestimation of $E^\text{MF}_s$.

It’s interesting to compare the contributions of dry summer $E_u$ and $T_u$ fluxes, i.e. normalized $E^\text{MF}_u$ ($E^\text{MF}_u$) and normalized tree transpiration ($T^\text{MF}_u$, i.e. tree water uptake divided by corresponding canopy area) respectively. The $E^\text{MF}_u$ was, on average (taking both years), ~0.6 mm d$^{-1}$, while $T^\text{MF}_u$ of Q. ilex and Q. pyrenaica were 0.83 mm d$^{-1}$ and 1.19 mm d$^{-1}$ respectively, as measured by Reyes-Acosta and Lubczynski (2013). $T^\text{MF}_u$ was higher than $E^\text{MF}_u$, however, as the MF area covered by bare soil (93%) was much larger than the canopy coverage area. This resulted in much higher overall $E^\text{MF}_u$ than $T^\text{MF}_u$.

#### 4.2. Sourcing of $E^\text{TSS}_s$ and $T^\text{MF}_s$ into groundwater and unsaturated zone components

In the two dry summers of the years 2009 and 2010, trees used the same amount of water from the shallow root system ($T_u$) and from the roots tapping the water table ($T_s$). The tree transpiration ($T_u$) did not change much during the late spring-summer period, even when the soil dried up and the water table dropped considerably (more than 1 m). Also Miller et al. (2010) showed that ground-water uptake by oak trees, measured using WTD fluctuation technique, did not change much between May-Nov 2008, while the water table dropped ~1 m. The low values of $T^\text{MF}_s$ and $T^\text{MF}_u$, both ~0.016 mm d$^{-1}$, were due to very low tree density, so it did not affect the water budget significantly. The decline of water table observed typically from June to September (~DOY 150–270, Fig. 6) was directly due to lateral groundwater outflow and $E^\text{MF}_u$ while $T_u$ had negligible impact.

In the MF area, grass had a negligible contribution to ET. The grass was present only shortly in early spring (March-May, DOY 60–140), i.e. when $T^\text{MF}_s$ was low, and died quickly during late spring, when the soil dried up. Besides, the grass in the study area was sparse, probably due to the poor development of the soil (regolith or entisols) and to intensive cattle grazing. For all these reasons, we assumed a negligible effect of grass transpiration on yearly $E^\text{TSS}_s$ (Table 1) and no effect on dry summer $E^\text{TSS}_s$. The $E^\text{MF}_u$ in
the area under the trees was assumed negligible due to the tree shade effect on soil surface temperature and net incoming radiation.

The diffusion of water vapor increased $E_{\text{MF}}$, especially during the dry summer period. In the dry summer of 2009 and 2010, the dryness of the top unsaturated zone thermally insulates the groundwater from changes in temperature at the ground surface, creating high temperature gradients in the top unsaturated zone. In the MF study area, for a clear-sky summer day at midday, the dry top soil temperature difference between the surface and 15 cm b.g.s. was 10°C and 18°C between the surface and 25 cm b.g.s., reaching stability below ~1 m b.g.s. (the temperature at 1 m changed only seasonally, but not daily). The high temperature gradient is a driving factor for water vapor flow and influences the flow of water in the unsaturated zone (Y. Zeng et al., 2009); in this study, the vapor flow by diffusion enhanced the overall evaporation of both infiltrating water and groundwater.

When taking into account the whole year (DOY 30–280 for year 2010, Table 1), $E_{\text{M}}$ is the most relevant source of $E_{\text{MF}}$. This is because the rain input was concentrated in the wet, cold part of the year (from October–November to April, DOY ~290–100), when the $ET_{p}$ was low and water input high, so that the unsaturated zone was moist enough to meet the potential evapotranspiration demand ($E_{a} = ET_{p}$). The absolute amount of water evaporated this way ($E_{a} = ET_{p}$) during the wet part of the year is higher than the absolute amount of water evaporated as $E_{a}$ during the dry summer.

4.2.1. Problems estimating the first and late second stages of evaporation

The larger HYDRUS1D estimates of the first stage evaporation as compared to $ET_{ec}$ ($E_{\text{MF}}$ right after rain, Section 3.1.2 and Fig. 7) can be either due to the low quality data collected by the tower in the first 90 min after rain event (Section 3.1.2) or by the HYDRUS1D model setup. In relation to the EC tower measurements, the quality of the data collected during these 90 min has been analyzed following Foken et al. (2005). The analysis shows that 84% of the data collected in the first 90 min after rainfall is to be considered “good” quality and ~60% of “very good” quality (categories 1–6 and 1–3 in Foken et al., 2005). This means that the problem of droplets on the gas analyzer had only a small effect on the daily evaporation rate calculated from the EC data, because the duration of the instrument failure was always shorter than the sampling time.

If the EC measurements taken after rain events are correct (Fig. 7), then the problem should be in the HYDRUS1D simulations. One possibility could be the frequency we selected for atmospheric variables: we passed to the HYDRUS1D model hourly data (we could not use smaller time steps for the rainfall input because the acquired field measurements had an hourly frequency), and this means that the water reaching the soil was averaged over the whole hour. If all the rain fell in few minutes of an hour, as it is often the case during dry summer in the MF area, this would result in a precipitation rate higher than soil infiltration capacity and so in some run-off, implying that not all rain could infiltrate in the subsurface and evaporate from there.

The substantially lower $E_{\text{MF}}$ than $ET_{ec}$ in the late second stage evaporation (that is, in “dry” periods, Section 3.1.2 and Fig. 7) is more difficult to interpret. Possible explanations are: (i) underestimation of the hydraulic conductivity of the soil; or (ii) wrong prediction of spatially distributed WTD applied to the upsampling of $E_{\text{MF}}$. Out of these two, the first is unlikely due to the large amount of hydraulic conductivity measurements taken and relatively low variability of these measurements, while the second is analyzed below.

To investigate the $E_{\text{MF}}$ uncertainty resulted by eventually incorrect definition of the WTD, we compared several averaged WTDs to fit the EC tower observation (with the assumption of negligible $T_{\text{ss}}$). We simulated $E_{\text{MF}}$ with HYDRUS1D in four sub-periods corresponding to the months of year 2010 in which groundwater level was recessing (DOY 90–120; 120–150; 150–180; 180–210 or 31 March–30 April; 30 April–30 May; 30 May–29June; 29 June–29 July), using as fixed bottom boundary conditions different WTDs (Fig. 9), e.g. for sub-period one we ran the HYDRUS1D model with WTD fixed at 1.0 m, 1.5 m, 2.0 m, 2.5 m, 3 m and 5 m; we repeated the procedure for sub-period two and so on. We then compared the medians of $E_{\text{MF}}$ (for various WTDs) and the median of $ET_{ec}$ (Fig. 9 and Table 3) as an indicator of the fit between HYDRUS1D simulation and the EC tower measurements; we used the median because it is a robust statistic and gives a general idea of the late second stage evaporation value, avoiding the overestimation of the peaks. The median values are also shown in Table 3.

In the first sub-period (DOY 90–120, corresponding to April 2010) the median value of $ET_{ec}$ is very similar to the median value of $E_{\text{MF}}$ for an averaged WTD of 3 m b.g.s. (Table 3 and Fig. 9). In the second sub-period (DOY 120–150 ~ May), the $ET_{ec}$ median value is

<table>
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<tr>
<td>Baldocchi et al. (2004)</td>
<td>Two EC towers (at 23 m and 2 m above ground), bucket soil model</td>
<td>Semi-arid, mediterranean climate (California) open woodland (oak)</td>
<td>40%</td>
<td>underestory (soil + grass) $ET_{ec}$ = 40–60% of overstory ET in winter and 10% in dry summer</td>
<td>n.r.</td>
</tr>
<tr>
<td>Williams et al. (2004)</td>
<td>EC tower, isotopic partitioning, sap flow (heat ratio method), bucket soil model</td>
<td>Mediterranean climate (Morocco), olive orchard plantation</td>
<td>21%</td>
<td>$E_{a} + \text{grass } T = 44%$</td>
<td>n.r.</td>
</tr>
<tr>
<td>Paço et al. (2009)</td>
<td>EC tower, heat dissipation sap flow method, GWD monitoring, interception measurements</td>
<td>Mediterranean climate (Portugal), open oak woodland</td>
<td>40%</td>
<td>during dry periods, $T_{g} - 90%$ of $ET_{ec}$</td>
<td>3.5–6</td>
</tr>
<tr>
<td>Miller et al. (2010)</td>
<td>Same as Baldocchi et al. (2004), plus sap flow (heat ratio) and GWD fluctuation for $T_{g}$</td>
<td>Same as Baldocchi et al. (2004)</td>
<td>60%</td>
<td>$E_{a} = 0%$, $E_{a} \sim 30%$, $T_{g} \sim 40%$</td>
<td>300</td>
</tr>
<tr>
<td>Yaseef et al. (2010)</td>
<td>Microclimatic station, sap flow (heat pulse velocity and heat dissipation methods), soil CO$_{2}$ flux chamber</td>
<td>Mediterranean climate (Negev desert), open woodland of Aleppo Pine</td>
<td>20–30%</td>
<td>$E_{a} \sim 60%$, $T_{g} - T_{a} \sim 7%$</td>
<td>3–10</td>
</tr>
<tr>
<td>This study</td>
<td>EC tower, microclimatic station, soil transient flow model, thermal dissipation and heat field deformation sap flow methods, WTD measurements</td>
<td>Semi-arid climate (Spain), open woodland (oak)</td>
<td>20–30%</td>
<td>$ET_{ec}$</td>
<td>~8</td>
</tr>
</tbody>
</table>
close to the $E_{\text{MF}}^{\text{ss}}$ median value for the 2 m b.g.s. WTD. In the third sub-period (DOY 150–180 / June) the $E_{\text{TEC}}$ has a median value also close to the $E_{\text{MF}}^{\text{ss}}$ median value for the 3 m b.g.s. WTD. However, our measurement-based estimates for the averaged water table in the MF area, based on the observation taken from the piezometers and on the prediction from an independent hydrogeological model of the catchment area of which the MF area is a subset (Francés et al., 2015) (Section 2.1), were deeper, i.e. 3.9 m depth on DOY 180 (year 2010) and 5.1 m depth on DOY 233 (year 2010). The difference between the WTD required to be fed to HYDRUS1D to match the EC tower measurements and the WTD actually observed in the area is so large that the problem of disagreement between HYDRUS1D and EC tower estimates (Fig. 7) could not be explained by uncertainty of the WTD used as an input to HYDRUS1D. The reason of the low estimates of the late second stage evaporation is discussed in further detail in Section 4.3.

The low late second stage evaporation estimates from our HYDRUS1D model is the reason for the 24% mismatch between $E_{\text{TEC}}$ and $E_{\text{MF}}^{\text{ss}}$ for the extended period (DOY 30–280) in year 2010 (Table 1). However, the mismatch between $E_{\text{TEC}}^{\text{ss}}$ and $E_{\text{MF}}^{\text{ss}}$ in the dry summer 2010 (Table 1) is likely also due to the small amount of the EC collected data in this period. The absence of sap flow measurements to estimate $T_{\text{ss}}$ outside of dry summer (Table 1) is not critical because $T_{\text{ss}}$ is negligible anyway.

### 4.3. The relevance of $E_{g}$ in semi-arid climate

In semi-arid areas, particularly during dry seasons, $E_{g}$ can be an important water loss for an aquifer (Lubczynski, 2000; Lubczynski, 2009; Lubczynski, 2011). However, the $E_{g}$ can be easily overlooked due to the relatively small rates as compared to rainfall, and difficulties in predicting its dependence on WTD. Our findings indicate that, in dry summers in the MF area, ~30% of $ET$ originated from groundwater ($E_{g}$), especially from $E_{g}$, even when the average WTD was at 5.8 m b.g.s. When the unsaturated zone was dry, there was little liquid contact between saturated zone and ground surface, so that capillary flow, even if present, was very small. However, the temperature gradients explained in Section 4.2.1 were high enough to trigger $E_{g}$, not only by liquid fluxes but also by water vapor flow, reaching its maximum in late dry summer (Fig. 8 of this article, see also Zeng et al., 2009; Zeng et al., 2011b). Even while taking into account capillary flow and only diffusive water vapor flow in the HYDRUS1D model, the simulated $E_{g}$ underestimated $E_{g}$. That underestimation, mainly in the late second stage $E_{g}$, seems to be due to an evaporation process not accounted for by HYDRUS1D. The correct estimate of the whole water vapor flow is very important for $E_{g}$ estimates in the dry summer because of much higher permeability of a very dry porous media to gas flow as compared to a wet porous media (Saito et al., 2006; Shokri et al., 2008). The temperature gradients (Section 4.2.1) and their daily variations (40 degrees at the ground...
by defining the relationship between $E_{st}$, particularly during long dry seasons typical for arid and semi-arid areas. The inclusion of water vapor in the determination of $E_g$ extinction depth should be carefully considered to avoid strong underestimation of $E_{st}$, e.g. due to incorrect soil modeling, which could eventually cover possible errors in the sap flow estimates of $T_s$. A wrong estimation of $E_g$ can lead, for example, to incorrect water vapor input to the atmosphere in climatic models of semi-arid and arid areas around the globe and to incorrect calibration of aquifer saturated hydraulic conductivity in groundwater modeling, resulting in underestimation of total water loss from an aquifer and related overestimation of groundwater resources. The low estimates of $E_g$ obtained in the simulations with HYDRUS1D highlight the need for an improvement of the knowledge related to gas flow processes and implementation of that knowledge in the subsurface hydrological models to improve accuracy of $E_{st}$ estimates, particularly in dry conditions by defining the relationship between $E_g$ and WTD in different hydrogeological conditions.

5. Conclusion

Areas with sparse vegetation, abundance of bare soil, “shallow groundwater” and dry climatic conditions are prone to substantial groundwater evaporation ($E_g$), particularly during long dry seasons. The presence of these unaccounted gas flow processes could explain the late second stage underestimation of $E_{st}$ as a consequence of underestimation of $E_g$. However, to solve that problem, a more focused study on identifying different gas flow processes considering the relation between $E^T$ and specific water table depths is required.

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