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Four Decades of Progress in Monitoring and Modeling of Processes in the Soil-Plant-Atmosphere System: Applications and Challenges

Assessment of the water balance in an Alpine climate: Setup of a micrometeorological station and preliminary results

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Abstract

An experimental field campaign was performed during summer 2012 in an Alpine environment (Cividate Camuno, Oglio river basin, 274 m a.s.l.) in order to assess the water balance of an anthropized soil at the local scale. A micrometeorological station equipped with traditional sensors coupled with eddy covariance apparatus and TDR was installed in order to measure precipitation, evapotranspiration and soil—water content at different depths. The soil water properties were determined after field and laboratory investigations. Here the preliminary results of the campaign are presented and discussed, focusing on the assessment of the evapotranspiration and of the water exchanges in the vadose zone as ones of the major problems of water imbalance.

Keywords: Alpine environment; Water balance; Anthropized soil; Eddy covariance; FAO Penman—Monteith; Evapotranspiration; Percolation and groundwater contribution.

1. Introduction

The knowledge of the water balance at the local scale is of crucial importance for the hydrological science, both in order to deepen the knowledge of the variety of phenomena which are involved, and in view of designing water management practices. Moreover in case of climate variability, a processes—based approach is necessary in view of predicting the water balance based on climatic scenarios. However assessing the water balance is still a major challenge due to the uncertainties which are involved in the

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measurement of some of the terms of the balance, viz the evapotranspiration and the groundwater exchange. In the last decades the application of the eddy covariance technique joint to the use of TDR apparatus opened the possibility of continuously measuring both the evapotranspiration and the soil—water content, so that their use became quite a common practice to assess the water balance [1], also in mountain areas [2,3,4]. Anyway the application of the eddy covariance technique, particularly in areas with complex orography, is still affected by uncertainties which are evident in the energy imbalance problem [5,6], so that a comparison of the results with traditional methods, such as the FAO Penman—Monteith model of evapotranspiration [7], is still required in order to assess the underestimation of the eddy covariance fluxes [8]. Also the groundwater exchange can be hardly measured, by means of lysimeters [9] or coupling tensiometer measurement with the knowledge of the soil—water constitutive laws [10], but in most of the cases it is indirectly estimated. In mountain environments, furthermore, soils are typically non—mature and both the soil and the ground—water table are shallow, so that the hydrologically active soil layer is often not evident and an important role in the water balance is played by the groundwater exchange.

With the aim at contributing to deepen the knowledge of the water balance partitioning in a mountain environment, a micrometeorological station also equipped with eddy covariance sensors and a TDR apparatus with multiplexer, was installed in the Oglio river basin (Central Italian Alps) during summer 2012, in an area which is very important for hydropower production and irrigation. In this paper the first results of the estimated water balance are reported. After describing the case study site (Section 2) and recalling the major difficulties in the estimate of the evapotranspiration fluxes and of the groundwater exchange (Section 3), the results are presented and discussed (Section 4).

Table 1. Summary of the instrumentation installed during the field campaign (Legend: u (m s^{-1}) planar wind speed positive Northward, v (m s^{-1}) planar wind speed positive Westward, w (m s^{-1}) vertical wind speed positive upward, T_{son} (°C) sonic temperature, [H_2O] (mmol m^{-3}) water vapour concentration, [CO_2] (mmol m^{-3}) carbon dioxide concentration, P_{atm} (kPa) atmospheric pressure, T (°C) environmental temperature, T_{air} (°C) air temperature, RH (-) relative humidity, $SW\downarrow$, $SW\uparrow$, $LW\downarrow$, $LW\uparrow$ (W m^{-2}) respectively are short—wave incoming and reflected radiations and long—wave incoming and outgoing radiations, h (mm) rainfall depth, T_{soil} (°C) soil temperature, H_g (W m^{-2}) ground heat flux, θ (cm^3 cm^{-3}) soil—water content. (§) : The device registers the time of each tipping).

Device	Manufacturer	Elevation/Depth (m)	Sampling frequency	Variables
Sonic anemometer	Gill WindMaster 3 Axes	3.32	20 Hz	u, v, w, T_{son}
Gas analizer	LiCOR Li—7500	2.98	20 Hz	[H_2O], [CO_2], P_{atm}, T
Thermohygrometer	Lastem	2.07	1 min^{-1}	T_{air}, RH
Radiometer	Kipp & Zonen	1.91	1 min^{-1}	$SW\downarrow, SW\uparrow, LW\downarrow, LW\uparrow$
2 Rain—gauges	Hobo	0	(§)	h
3 Soil thermometers	Lastem	- 0.05, - 0.10, - 0.15	1 min^{-1}	T_{soil}
Heat plate exchanger	Lastem	- 0.075	1 min^{-1}	H_G
TDR with 4 probes	Campbell Sci	- 0.05, - 0.15, - 0.22, - 0.375	5 min^{-1}	θ

2. The case study

2.1. The test site and the setup of the station

The Valle Camonica, of glacial origins, is the main valley of the Oglio river basin (Central Italian Alps), which ranges from 187 m a.s.l. at Sarnico, at the closure of Lake Iseo, to 3539 m a.s.l. of the Adamello massif. Lake Iseo is still an important reservoir for irrigation, and moreover, due to the high elevation ranges, the water resources of the valley have been deeply exploited for hydropower production as well [11,12]. According to the definition proposed by Bandini [13] the climate ranges from typical

Alpine sublitoranean pattern in the lower part of the valley, to the continental one in the higher part [14]. Here an experimental site was chosen in the valley floor at Cividate Camuno (274 m a.s.l.). The site is a flat rectangular lawn (95 m in the NE—SW direction and 70 m in the SE—NW direction) surrounded by complex and heterogeneous orography, with a slope Eastward and a street Northward. During the field campaign the loan was mainly covered by alfalfa (*Medicago sativa*), wild carrot (*Daucus carota*) and yarrow (*Achillea millefolium*), with a maximum height ranging from 0.6 to 0.7 m. A micrometeorological station, placed at the center of the lawn, was equipped with traditional devices coupled with eddy covariance (EC) apparatus and TDR with four probes placed at different depths in the soil (Table 1). Data were stored by two Campbell dataloggers (CR1000 and CR3000). The fetch between the station and the surrounding obstacles was 50 m (NE) from a country house, 40 m (SE) from the slope, and 15 m (S) from the solar panels installed to supply power. Data were collected from 9 July 2012 until the end of September. Here the data from 17 July to 22 August will be presented and discussed.

2.2. Characterization of the soil physical properties

The lawn is located nearby a hydropower plant placed in an artificial cavern, built in 1939, and the soil developed over the scrap which was produced by the excavation of the cave. The parent rock is therefore mainly composed by marl fragments of anthropic genesis. The topsoil, investigated by means of a 0.5 m—deep dig, is non mature technosol characterized by three horizons (A, C, D). The observed A and C horizons are respectively 9 cm and 29 cm thick. Two sub—horizons, C1 and C2, were detected, as the upper one (C1, 13 cm thick) is characterized by more roots presence and less skeletal fraction than the lower one (C2, 16 cm thick). The quite fast development of A horizon is in agreement with other technosols reported in the literature [15]. Two infiltration field tests with a single ring infiltrometer were performed to determine the soil hydraulic conductivity at saturation of layer A and C1 as a fitting parameter of the infiltration curve [16]. The soil water content of natural and modified conditions was determined in the laboratory over both loose and core samples. Moreover laboratory tests were performed in order to determine the porosity (ϕ), the volumetric water content after long imbibition (θ_w), the soil—water retention relationships (SWRRs) by means of the Richards apparatus, the organic matter content (x_o) and the grain size distribution curve. The SWRRs were interpolated by means of the van Genuchten model with the usual constraint $m = 1 - 1/n$. The relevant physical properties of hydrologic interest are reported in Table 2 for each horizon.

Table 2. Soil physical properties of hydrologic interest (Legend : ϕ porosity, ρ_b bulk density, θ_w volumetric water content after long imbibition, Ψ_1 , n van Genuchten fitting parameter, K_s hydraulic conductivity at soil saturation, x_o organic matter content, x_{sk} skeletal fraction, *n.a.d.* not available data).

Horizon	Depth (cm)	ϕ (cm³/cm³)	ρ_b (g/cm³)	θ_w (cm³/cm³)	Ψ_1 (kPa)	n (-)	K_s (m s^{-1})	x_o (%)	x_{sk} (%)	USDA textural class
A	0—9	0.53	1.24	0.47	21.6	1.56	1.19E-05	10.1	44.6	Silty loam
C1	9—22	0.44	1.49	0.44	20.2	1.45	5.58E-06	7.8	44.6	Loam
C2	22—38	0.50	1.32	0.37	25.6	1.41	5.58E-06	6.1	59.2	Silty clay loam
D	Below 38	*n.a.d.*	*n.a.d.*	*n.a.d.*	16.0	1.53	*n.a.d.*	6.2	88.6	Silty loam

3. The mass imbalance problem on an anthropized soil in an Alpine environment

Let us consider a control volume **B** at the interface between the upper soil layers and the atmospheric surface layer. If the soil is horizontal, the water balance (WB) over a time window Δt is described by the following formulation of the mass conservation principle:

$$P - ET - G = \Delta S\,, \tag{1}$$

where P [L] is the total precipitation per unit area entering the control volume **B** during Δt, ET [L] is the total evapotranspiration, G [L] is the total flux at the bottom of **B** and ΔS [L] is the variation of the soil—water stored in **B** during Δt. The precipitation is taken positive if entering **B**, while ET and G are taken positive if outgoing from **B**. When the water balance is numerically or experimentally assessed, an imbalance ε [L], due to experimental or modeling uncertainties, is typically found at each time step Δt. The equation *(1)* is therefore rewritten as:

$$P - ET - G = \Delta S + \varepsilon\,. \tag{2}$$

The water balance is therefore closed in a weak sense if the expectation $E[\varepsilon]$ is null over the whole investigated time period. In the followings the major sources of imbalance, related to the estimate of the evapotranspiration and of the bottom fluxes are presented.

3.1. Evapotranspiration

The eddy covariance technique (EC) allows to directly estimate the evapotranspiration fluxes, by means of the measurements of the turbulent fluxes of energy and mass exchanges between the land and the atmospheric surface layer. The energy and mass fluxes are calculated by estimating the covariance between high frequency measurements (greater than 10 Hz) of the vertical wind speed component and a generic scalar variable [17]. Particularly, sensible (H_S) and latent heat fluxes (H_E), being the latter directly related to the evapotranspiration by means of the latent heat of evaporation λ = 2.45 MJ kg^{-1}, are estimated as covariance between the vertical wind velocity and the temperature and the water vapor, respectively. Anyway in mountain environments, the hypotheses of steady—state turbulent conditions and null expectation of the vertical wind velocity are often not fulfilled, due to complex orography and drainage winds. Therefore, in order to assess the reliability of the estimated fluxes, an important role is played by the footprint analysis which allow to determine their representative source area [18]. However, an energy imbalance is often observed [6,19] between the available energy (calculated as the net contribution between the net radiation R_n, the ground heat flux H_g and the variation ΔH of the stored heat in the control volume) and the sensible and latent heat fluxes, i.e.:

$$R_n - H_g - \Delta H = H_S + H_E + \varepsilon_H\,, \tag{3}$$

being the residual ε_H typically positive. Therefore the residual ε_H of the energy balance is regarded to as a reliable index of the accuracy of the EC method, being the EC more accurate the lower is its value [7,20]. At present it is not clear whether the imbalance is due to an underestimation of latent heat or sensible heat fluxes, or both, so that the evapotranspiration measured with EC technique can be considered an inferior boundary of the actual evapotranspiration. Accordingly, we estimated an upper boundary of ET at hourly scale from the closure of the energy balance. In order to do so, it was assumed that the energy imbalance ε_H is shared between latent and sensible heat fluxes at the same ratio of the measured H_S and H_E [6]. Defining the Bowen ratio Bo as the ratio between the measured sensible and latent heat fluxes:

$$\mathrm{Bo} = \frac{H_S}{H_E}\,, \tag{4}$$

the upper value H_E' of latent heat flux is obtained from the available energy as in the following equation:

$$R_n - H_g - \Delta H = H_E'\left(1 + \mathrm{Bo}\right). \tag{5}$$

As a term of comparison of the measured evapotranspiration fluxes, we estimated the actual evapotranspiration of the investigated crop by means of FAO Penman—Monteith method (FAO—PM) [21]. The method incorporates both low frequency micrometeorological data and physiological parameters of the vegetation. According to this method the actual evapotranspiration ET is given by:

$$ET = K_C \cdot K_r \cdot ET_0, \tag{6}$$

in which ET_0 is the potential evapotranspiration for a reference crop; K_C is a crop coefficient, which values are tabulated and depends on the crop type, variety and stage of development; $K_r \leq 1$ is a water stress coefficient depending on the available soil water content. According to [21, p.74], the reference evapotranspiration ET_0 at hourly time step is given by the following equation:

$$ET_0 = \frac{0.408 \cdot \Delta\left(R_n - H_g\right) + \gamma \dfrac{37}{T_{air} + 273} u_2\left(e_s - e_a\right)}{\Delta + \gamma\left(1 + 0.34 u_2\right)}, \tag{7}$$

where ET_0 is in mm hour^{-1}, R_n is again the net radiation at the grass surface (MJ m^{-2} hour^{-1}), H_g the soil heat flux (MJ m^{-2} hour^{-1}), T_{air} the mean air temperature (°C), Δ the saturation slope of the vapor pressure curve at T_{air} (kPa °C^{-1}), γ the psychrometric constant (kPa °C^{-1}), $e_s(T_{air})$ the saturation vapor pressure at T_{air} (kPa), e_a the average hourly actual vapor pressure (kPa), u_2 the average wind speed at 2 m of height over the crop (m s^{-1}). The approach of using the FAO—PM method as a term of comparison of the estimated ET fluxes with eddy covariance is quite common [7] anyway it is very sensitive to the estimate of the crop coefficient K_C which is affected by uncertainty. As an example, K_C for an alfalfa covered crop is expected to range from 0.5 to 1.2 during the maximum growing period, depending on the management practices, while K_C for a grazing pasture is expected to range from 0.75 to 1.05 during the same period. For the investigated crop we chose K_C ranging from 0.75 to 0.95 as average values among alfalfa and grazing pasture ones.

3.2. Groundwater exchange

The soil–water balance in temperate climates is sensitively characterized by the water exchange between soil and groundwater. During summer, due to high evapotranspirative requirements and precipitation shortage, the hypothesis of null exchange between soil and groundwater can be realistic, thus the vadose zone behaving as a capillary barrier. However, in an anthropized and mountain environment, as the soil and the water table depth are shallow, both percolation and water rise from the water table can happen and meaningfully affect the water balance [22]. As a physically based approach, three main different hypotheses on the nature of the water exchange at the bottom of the soil control volume can be introduced: (*i*) a null exchange rate, (*ii*) a pure percolation with unitary gradient of the total hydraulic potential and (*iii*) a percolation or water rise driven by the local gradient of water potential. The latter is the complete physical representation of the phenomenon based on the Darcy law, but requires both the soil—water constitutive laws at the bottom of the control volume to be known [10], and a reliable estimate of the series of soil—water potential at the same depth. In this sense a common approach, in order to bridge the difficulties related to the field representativeness of the laboratory—measured constitutive laws, is to determine the effective soil—water constitutive laws after inverse simulation [23], or statistical assimilation of field series [9]. In this work, once that it was assessed the range of variation of the measured evapotranspiration data, the groundwater exchange was investigated by means of the cumulated water balance. Equation *(2)* can in fact be rewritten in the following integrated form:

$$\int(G+\varepsilon)\,\mathrm{d}t=\int(P-ET-\Delta S)\,\mathrm{d}t\,. \tag{8}$$

If the expectation of the residuals ε is null, the average derivative ∂G of the cumulative, i.e.

$$\partial G\equiv\frac{\Delta\int G\mathrm{d}t}{\Delta t}=\frac{\Delta}{\Delta t}\int(P-ET-\Delta S)\,\mathrm{d}t\,, \tag{9}$$

is positive if the time interval Δt is characterized by percolation and negative if it is characterized by groundwater contribution.

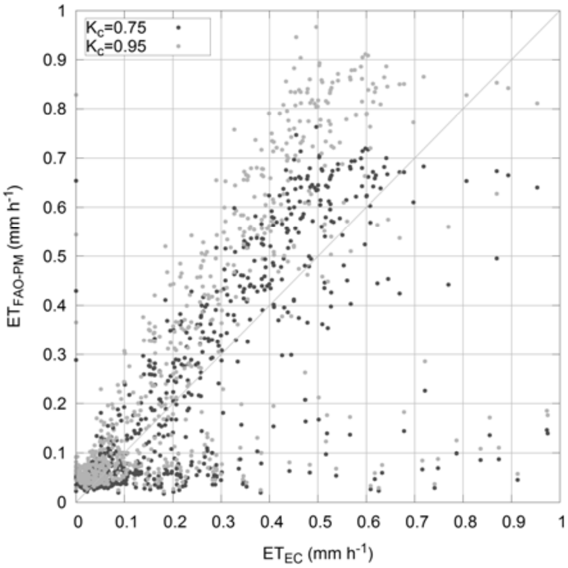

Fig. 1. Comparison between the estimated evapotranspiration fluxes by means of eddy covariance technique (ET_{EC}) and FAO—PM model ($ET_{FAO—PM}$) for two different values of the crop coefficient K_C.

4. Results and discussion

The results about the estimation of the evaporative fluxes, the groundwater exchanges and the water balance from 17 July to 22 August will be presented and discussed. The planar wind speed components (u and v), collected by the sonic anemometer and averaged over one minute, allowed to detect the wind directions regime. An alternation between two main regimes was observed: an Eastern katabatic wind blowing daily since 8 pm until 5 am with a speed between 1 and 2 m s^{-1} and a local SW wind, called *Òra*, recently observed also by Valerio et al. [24] which rises the valley from Lake Iseo in the afternoon with higher speed variability, ranging from 1 to 5 m s^{-1}. A less frequent drainage condition is represented by a wind that blows during the evening from NNE with a speed between 1 and 5 m s^{-1}.

The turbulent data were processed by means of the EddyPro software [25] to obtain hourly fluxes. Following [6,26] the main corrections applied were: (*i*) threshold of the 10% of allowance for the missing values, (*ii*) cross—wind corrections, (*iii*) linear detrending with time constant of 5 minutes, (*iv*) maximization of the covariance, (*v*) the corrections of the WPL theory for the density fluctuations [27], and (*vi*) the planar fit method with no velocity bias [28] over eight equally spaced sectors. The footprint model suggested by Hsieh et al. [29] allowed to estimate the representative area under different stability conditions and to check that most of the computed fluxes are representative of the experimental site. The obtained sensible and latent heat fluxes were compared with the available energy (the storage in the soil was computed according to [30]). A residual ε_H of 43% of the available energy was obtained on average for hourly fluxes and with the above introduced corrections. The obtained slope of $H_S + H_E$ vs. $R_n - H_g - \Delta H$ is 0.52, smaller than that obtained in other investigations in Northern Italy floodplain [31] but in agreement with values observed in other mountain ranges [32]. Accordingly, from equation *(5),* the percentage correction of the latent heat with respect to the measured flux, results $(H'_E - H_E)/H_E = 76\%$. A close value is found for the sensible heat flux, according to the hypothesis of the Bowen ratio correction.

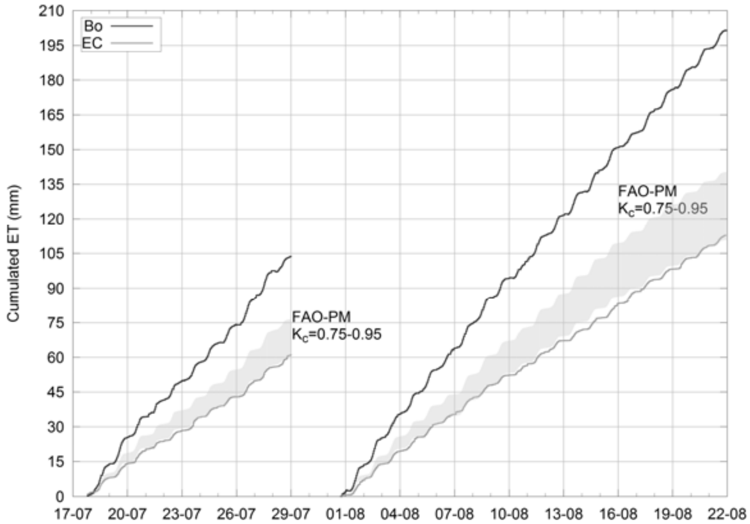

Fig. 2. Comparison between the cumulated ET estimated with EC technique, Bowen ratio correction and FAO—PM model, for different values of the crop coefficient K_c.

In Figure 1 the comparison between the estimated evapotranspiration fluxes by means of eddy covariance technique (ET_{EC}) and FAO—PM model ($ET_{FAO—PM}$) is presented for two different values of the crop coefficient K_C. It can be observed that FAO—PM model tends on average to overestimate the EC—measured fluxes of *ET*. As pointed out in Section 3.1, the EC estimate is regarded to as a lower boundary of the actual evapotranspiration, and the Bowen ratio correction as an upper one. During the investigated period the Bowen ratio method provided a correction of measured *ET* of 76%. The cumulated evapotranspiration, estimated both by means of FAO—PM model and by EC and Bowen ration correction, is represented in Figure 2 for two periods of the field campaign. The FAO—PM model provides an interval of values which is contained within the values estimated by means of EC and Bowen ratio correction. An *a posteriori* estimate of the crop coefficient provides K_C values ranging from 0.76 to 1.34 for EC and Bowen ratio correction estimates of *ET*, respectively. On average an *ET* flux of 5.2 and 9.2 mm d^{-1} was estimated by means of EC and Bowen ratio correction. In Figure 3 the term $G + \varepsilon$ as defined by Equation *(8)* is cumulatively plotted for the two continuous periods in which EC—based *ET* data are available. The soil control volume is closed at three different depths, respectively considering only the layer A, or the layers A and C1, or all the soil layers A, C1 and C2. The evapotranspiration fluxes are provided both by EC and with the Bowen ratio method. The figure shows that during most of the investigated periods the groundwater contributes to the water balance, as the average time derivative defined by equation *(9)* is often negative for both the *ET* estimates, being positive during rainy days, as the evapotranspiration is small for both the estimates. The behavior of the series A—C2 on 6 August 2012 anyway enlightens a source of uncertainty which is strictly related to the underestimation of the real precipitation by rain gauges in case of intense rainfalls [33,34,35]. The dropdown of series A—C2, in fact, results from $P < \Delta S$, being *ET* negligible during the precipitation. As the local orography is such that the surface and hypodermic runoff is negligible, such a pattern is a signal of underestimation of the precipitation.

The following four days were dry, so that:

$$\partial G = \frac{\Delta}{\Delta t}\int\left(-ET-\Delta S\right)\mathrm{d}t\,, \tag{10}$$

and the water balance is strongly affected by the different estimates of *ET*. In fact the estimate obtained with the Bowen ratio correction evidences that all the soil control volumes are characterized by groundwater contribution as it was in the previous dry days. On the other hand the EC estimate contributes to depict a scenario in which only the layer A is affected by water content rise from below, while the control volume A—C1 seems to be able to supply all the water needed for the evapotranspiration, as the average derivative ∂G is null over that period. The control volume A—C2 is instead still percolating as the average derivative ∂G is positive for some days.

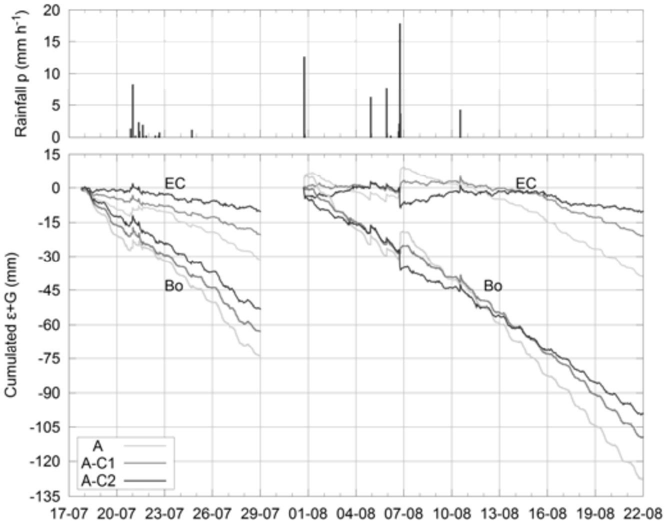

Fig. 3. Cumulated values of $G + \varepsilon$ as defined by equation *(5)* for two periods for which continuous eddy covariance measurements are available. The soil control volumes are closed respectively at the bottom of layer A (line A), C1 (line A—C1) and C2 (line A—C2). Evapotranspiration is estimated both with eddy covariance (EC) and Bowen ratio (Bo) method.

The investigated soil is therefore characterized both by percolation and groundwater contribution along its whole profile. A modeling approach based on the first two hypotheses introduced at Section 3.2, i.e. the hypothesis of null groundwater exchange and of percolation with unitary gradient, is not realistic and cannot properly describe the actual soil behavior. On average, during dry days, the layer A experiences a groundwater contribution of 3.6 mm d^{-1} for EC estimate of *ET*, and of 8.0 mm d^{-1} for estimate of *ET* provided by Bowen ratio correction. If compared with the estimated *ET* fluxes, this means that a meaningful fraction of the evapotranspiration is supplied by groundwater contribution.

5. Conclusions

A micrometeorological station equipped with eddy covariance devices and TDR apparatus was installed in a mountain environment in order to investigate the soil water balance components of an anthropized soil. According to previous investigations it was recognized that the eddy covariance technique underestimates the energy balance. Thus the Bowen ratio correction was applied in order to get an upper boundary for the measured evapotranspiration. The FAO Penman—Monteith estimates of the evapotranspiration, for a range of realistic crop coefficients, were found to be comprised between values estimated with the eddy covariance technique and those corrected by means of the Bowen ratio. Therefore the measured evapotranspiration values were considered reliable limits of the value of the actual evapotranspiration and were used to assess the water balance of the upper soil layers. An average evapotranspiration flux of 5.2 and 9.2 mm d^{-1} was estimated by means of the eddy covariance technique and of the Bowen ratio method, respectively. A meaningful groundwater contribution was detected for all the soil layers, during most of the dry days, and for both the minimum and the maximum estimated values of evapotranspiration. Particularly, during dry days, the estimated groundwater contribution to the layer A is about 3.6 mm d^{-1} if the evapotranspiration is estimated with eddy covariance technique, and of 8.0 mm d^{-1} if Bowen ratio—corrected values are used. The groundwater contribution, therefore, supplies to a meaningful fraction of the evapotranspiration.

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