Field determination of the water balance of the Areuse River delta, Switzerland

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Abstract The water balance during a period of one year (15 October 1990–15 October 1991) was determined at an experimental site in the Areuse River delta (Switzerland). The groundwater recharge rates were found to be 36% and 33% of total precipitation according to evapotranspiration estimates based on the Primault (1962) and the Penman-Monteith methods, respectively. Variations in the water storage were obtained by weekly measurements with a neutron probe. Observed hydraulic gradients indicated a zero-flux plane between depths of 0.55 and 1.02 m that separated the infiltration zone from the zone of evapotranspiration in all seasons.

Key words unsaturated zone; evapotranspiration; water balance; groundwater recharge; porous media; Neuchâtel; Switzerland

Evaluation in situ du bilan hydrologique dans le delta de l'Areuse, Suisse

Résumé Le bilan hydrologique a été évalué sur une période d'une année (15 octobre 1990–15 octobre 1991) au site expérimental du delta de l'Areuse en Suisse. La recharge de la nappe correspondait à 36 et 33% des précipitations totales, respectivement selon les méthodes d'estimation de l'évapotranspiration de Primault (1962) et de Penman-Monteith. Les teneurs en eau, mesurées chaque semaine à l'aide d'une sonde à neutron, ont permis d'estimer la variation du stock d'eau. Les gradients hydrauliques observés indiquaient un point neutre entre 0,55 et 1,02 mètres, délimitant les zones à flux ascendant et descendant.

Mots clefs zone non saturée ; évapotranspiration ; bilan hydrique ; recharge de la nappe ; milieu poreux ; Neuchâtel ; Suisse

INTRODUCTION

Knowledge of the water budget of soils is important for agriculture and irrigation management, for runoff and peak-discharge forecasts in regions where inundations may threaten densely populated areas, and for understanding the dynamics of pollutant concentrations (e.g. nitrate) in the groundwater. Due to the close link between the hydrological cycle and the atmospheric moisture budget, local and regional feedbacks to climate are strongly affected by the water budget and water resource management of any region where occasional or frequent drought periods occur. In particular, the loss of water to the atmosphere via evapotranspiration and the regional redistribution of water by irrigation can significantly modify local climate (e.g. Zhong & Doran, 1995).

A number of factors affect the recharge of groundwater. Hydraulic conductivity

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and soil porosity, rainfall amounts and their seasonal distribution, the slope of the aquifer, and soil moisture contents, are the most important factors. Of these, hydraulic conductivity and soil porosity have the strongest influence. Other control variables affecting recharge are soil type (increasing clay content decreases recharge; e.g. Athavale *et al*., 1980; Edmunds *et al*., 1992), rainfall (increasing rainfall increases recharge; e.g. Bredenkamp, 1988a,b), and vegetation characteristics (longer growing season or deeper rooting depths decrease recharge; e.g. Nulsen & Baxter, 1986;Zhang *et al.*, 1999).

The estimation or measurement of evapotranspiration under natural conditions at different spatial scales (local, field, regional) and over different time periods (instantaneous, daily, seasonal) is crucial for water conservation and management, in particular for planning the efficiency of irrigation and plant water usage (Stone *et al.*, 1973; Molz, 1981). The knowledge of actual evapotranspiration is also needed for the validation of water transport models (Camillo *et al.*, 1983; Witono & Brucker, 1989). Many methods exist for estimating evapotranspiration. However, it is recognized that the more complicated formula is not necessarily the best one (de Bruin & Stricker, 2000), and that some approaches are not widely known (e.g. Botna & Müller, 1999).

The objectives of this study were to quantify recharge through a highly heterogeneous unsaturated zone of the Areuse River delta, and to compare two methods to estimate evapotranspiration: (a) a relatively simple empirical method by Primault (1962) calibrated for Switzerland, and (b) the Penman-Monteith approach (e.g. Monteith & Unsworth, 1990). The Penman-Monteith equation (Monteith, 1965) is currently the most widely recommended and used method (Allen *et al.*, 1989; Smith *et al.*, 1991) but requires more detailed measurements of driving variables such as net radiation, windspeed, dewpoint temperature, air temperature and vegetation-specific parameters.

Two reasons caused the selection of this location: firstly, its aquifer is an important reserve of water exploited by several communal and municipal water supplies in the region of Neuchâtel, Switzerland; secondly, the aquifer has long been subject to nitrate pollution from agricultural sources. The location of a well in the Areuse River delta was directly influenced by the elevated nitrate concentrations in the area with groundwater quality expected to further decrease with the construction of a sports field in 1989 adjacent to the pumping station. Therefore, in 1990, a new water well was constructed at a new location in the Areuse River delta where nitrate concentrations in the aquifer were lower (Mdaghri-Alaoui *et al.*, 1993).

SITE AND METHODS

Site description

The experimental site is located at 430 m a.s.l. along the orographical left side of the Areuse River on its delta, 200 m away from the shore of lake Neuchâtel, where measurements were carried out from 15 October 1990 to 15 October 1991. The Areuse River delta has been a focus of several geological and hydrogeological studies in the past decades (Meia *et al.*, 1971; Burger, 1980; Mdaghri-Alaoui, 1990). Its aquifer has a complex and horizontally heterogeneous geological structure, being covered by an approximately 3 m unsaturated soil composed of a 0.2–0.3 m humus layer overlaying a sandy loam. The highly permeable material is organized in localized packages containing coarse sediments varying from sandy gravel to gravel. Additional data from the Areuse River delta aquifer and its geometry were obtained from detailed geophysical investigation of the area in conjunction with a log lithostratigraphic study of the region's wells and piezometers (Mdaghri-Alaoui *et al.*, 1993). The aquifer shows lithological heterogeneities with a complex deltaic structure. It consists of sandy gravel having an average thickness of about 20 m. The substratum of the aquifer is mostly composed of silty clay. The water table varied between 2.7 and 3.2 m in depth.

Variations in water storage were obtained from weekly measurements with a neutron probe. Hydraulic gradients obtained by pressure measurements were used to identify periods with significant infiltration and to locate the zero flux plane separating deep drainage and evapotranspiration. A particle size analysis was carried out because it was expected that this zero flux plane that was seen in the hydraulic gradient data coincides with a textural change in the soil profile.

Soil texture

The soil is composed of recent fluvial deposits from the Areuse River on its delta, without a vertical stratification at the soil pit itself. The whole Areuse delta fills a depression in the basement of the Jura foothills beginning at ˜ 2 km distance from the experimental site. The Quaternary deposits to both sides of the recent river delta are a mixture of fluvio-glacial and fluvial deposits from the last two ice ages (Würm and Riss).

The soil horizon between 0 and 0.36 m is a clay loam soil containing gravel and a few rocks. The soil pH is 7 and the porosity varies between 0.47 and 0.54. Organic matter content varies between 0.04 and 0.05. Between 0.36 and 0.70 m the soil is composed of sandy clay loam (0.36–0.44 m) and sandy loam (0.44–0.70 m) with a very low content of coarse material. The porosity varies between 0.46 and 0.51, with a pH of 7–8. The soil below 0.70 m consists of coarse material, primarily gravel and stones. The soil matrix is composed of fine sand. Thus, in summary three primary interfaces were found which separate the distinct soil horizons: the first is at a depth of 0.08–0.10 m, the second around 0.36 m, and the third and most important for the hydrology of the experimental site at 0.70 m.

Methods

The water budget of a soil column was determined with a combined experimental and modelling approach to quantify the components of the water budget on a weekly basis. The budget equation was used:

$$P - E \pm \Delta S - R = 0 \qquad (1)$$

where P is precipitation, E is evapotranspiration, ΔS is the change of the water stored in the soil column between the groundwater table and the surface, and R is recharge to the groundwater table. All units are in mm water column.

In this one-dimensional simplification it is assumed that lateral flow is negligible. Both surface runoff and subsurface lateral flows can be assumed to be close to zero due to the fact that the slope of the Areuse delta plain is only 1‰ at the surface, and the slope of the water table, as measured in the direction heading towards the lake of Neuchâtel, varies between 1‰ and 6‰. Surface runoff is not normally observed, even during heavy showers.

Precipitation, meteorological and soil moisture measurements Precipitation was measured with a Hellmann-type rain gauge with a resolution of 0.5 mm. Evapotranspiration was modelled with two approaches, an empirical approach by Primault (1962) and the more widely known and frequently used mechanistic approach by Penman and Monteith (Monteith & Unsworth, 1990). Details about the two models are given at the end of this section. Most of the meteorological data needed in both evapotranspiration models were obtained locally from a standard three-cup anemometer and wind vane at 3.5 m height, and a combined air temperature and relative humidity probe (Rotronic, Switzerland, type MP-100F) mounted in a white radiation shield at 2.0 m height. Duration of sunshine and global radiation data were obtained from an automatic weather station run by MeteoSwiss near Neuchâtel, 6 km to the north of the field site.

The vertical profile of the pressure head was measured with tensiometers (SenSym, type LX06005G) at depths of 0.55, 1.02, 1.54 and 2.07 m. A Bell & Howell type BHL 4003 pressure probe was installed in the saturated zone at 3.65 m depth to automatically determine changes in the groundwater table. A model 503DR neutron moisture probe (Campbell Pacific Nuclear, Martinez, California, USA) was used to measure changes in soil moisture. The probe was capable of monitoring moisture to a depth of 3.5 m below the surface. The moisture gauge contained a fast radioactive (americium 241-beryllium) neutron source. For replication and verification of the vertical soil moisture profile, an access tube (MS 63, brass with 40 mm inner diameter and 2 mm tube strength) was inserted into the ground for the neutron probe at a distance of 1 m from the tensiometer profile. The time integration of neutron counts was 32 s and measurements were carried out in 10 cm depth intervals once per week.

All instruments listed above were attached to a data logger (DataTaker, model DT100) and average values recorded every 30 min.

The empirical evapotranspiration model by Primault (1962) Primault (1962) proposed an empirical approach to estimate evapotranspiration from short-cut grassland in Switzerland at elevations between 300 and 1200 m a.s.l.,

$$E = c\left[A\frac{103 - rH}{100}\left(t_s + 2t_p\right) + B\right] \tag{2}$$

where total evapotranspiration E (mm) during a period of t_p days is a function of the average relative humidity rH (%) of this period, and the total duration of sunshine t_s (h). Three parameters correct for variations in the altitude (A and B) and the season (c). Variations in the altitude are given by:

$$A = -0.12 + 0.00306\,h - 2.83\cdot 10^{-6}\,h^2 + 9.45\cdot 10^{-10}\,h^3 \tag{3}$$

where h is the altitude in m a.s.l. This parameter reflects that increasing precipitation at increasing altitudes is expected to lead to increasing evapotranspiration when the solar forcing and atmospheric conditions are kept constant. The value of A varies between 0.57 and 1.11. The height-dependent offset B can be approximated by:

$$B = 0.5387 - 0.0003263\,h - 6.525 \cdot 10^{-7}\,h^2 \tag{4}$$

It varies between 0.38 at 300 m and –0.79 at 1200 m and only plays a minor role. In contrast, the seasonal variation c significantly influences E as is described by:

$$c = -0.5068 \cdot \sin\left(\frac{2p}{365} DOY + 0.5593\right) - 0.0711 \cdot \left(\frac{4p}{365} DOY + 0.6112\right) + 0.6271 \tag{5}$$

where DOY is day of year. The coefficient c adjusts the evapotranspiration estimates for temperature effects. Such a correction is necessary since air temperature (or net radiation) is not a driving variable in this empirical model. Primault (1962) only published tables and figures of the coefficients. Equations (3)–(5) were therefore fitted to Primault's graphical data using the least-squares technique. The average relative humidity rH for a period of several days is not an actual physical measure, but instead a statistical value that is easily derived from standard meteorological measurements. Unweighted averaging of the half-hourly measurements was used to obtain the average rH of all weekly periods. Primault's (1962) model is supposed to be used at this or longer time steps.

The Penman-Monteith evapotranspiration model The Penman-Monteith equation was used to compute evapotranspiration from half-hourly data in energy flux density units (W m^{-2}),

$$\lambda E = \frac{s \cdot (R_n - G) + c_p \rho_a \dfrac{e_{sat} - e}{r_a}}{s + \gamma \dfrac{r_a + r_c}{r_a}} \tag{6}$$

where λ is the latent heat of vaporization (approximated by 2501000 – 2370T in J kg^{-1}, where T is temperature in °C), E is the water vapour flux (kg m^{-2} s^{-1}), s is the rate of change of saturation vapour pressure with temperature, R_n is the net radiation, G is the ground heat flux, c_p is the specific heat of air at constant pressure, r_a is the density of air, e_{sat} and e are the saturation vapour pressure and actual vapour pressure, respectively, r_a and r_c are the aerodynamic and canopy resistances, respectively, and γ is the psychrometric constant. For details see Monteith & Unsworth (1990).

Saturation water vapour pressure e_{sat} was approximated by Stull (1988):

$$e_{sat}(T) = 611.2 \cdot \exp\left[\frac{17.67 \cdot (T - 273.16)}{T - 29.66}\right] \tag{7}$$

where e_{sat} is in Pa, and air temperature T is in K. Water vapour pressure e was obtained for each interval of 30 min by multiplying e_{sat} at measured T with measured relative humidity, $e = e_{sat}(T) \times rH/100$. Net radiation R_n was obtained from global radiation measurements R_s at Neuchâtel using the regression approach

$$R_n \approx (1 - a)R_s + L_* \tag{8}$$

where a is the short-wave albedo, and L_* is the long-wave radiation balance (Table 1) (cf. also Kalma *et al.*, 2000). Estimates for a and L_* were derived from measurements published by Eugster (1994) and Eugster & Hesterberg (1996), respectively, from a litter meadow at comparable altitude in Switzerland. Ground heat flux G was estimated (Stull, 1988) to be $0.1R_n$ when $R_n > 0$ (daytime conditions), and $0.5R_n$ when $R_n = 0$. Aerodynamic resistance r_a was derived from wind speed measurements and a best estimate for the roughness length z_0 (Table 1) taken from Panofsky & Dutton (1984):

$$r_a = \frac{\left(\ln \frac{z-d}{z_0}\right)^2}{k^2 \bar{u}(z)} \tag{9}$$

where z is the measuring height (10 m), d is the zero plane displacement (˜ 0.1 m), k is the von Kármán constant (0.40), and $u(z)$ is the wind speed at height z (m s^{-1}). The chosen value for z_0 takes into account that the measurement site is situated in a heterogeneous landscape with a larger effective roughness than the one of the local vegetation.

Table 1 Parameter estimates for the Areuse River delta near Neuchâtel used in the Penman-Monteith evapotranspiration model. See text for details.

Parameter	Description	Value
a	short-wave albedo	0.20
L_*	long-wave radiation balance	–25 W m^{-2}
$r_{c,min}$	minimum canopy resistance for H_2O	75 s m^{-2}
b_{rs}	radiation where $r_c = 2\ r_{c,min}$	300 W m^{-2}
z_0	aerodynamic roughness length	0.045 m

Canopy resistance r_c was estimated with the rectangular hyperbolic approximation by Turner & Begg (1973):

$$r_c = r_{c,\min} + b_{rs} \cdot r_{c,\min} / R_s \tag{10}$$

where minimum resistance $r_{c,min}$ was estimated from the value given for NO_2 by Eugster & Hesterberg (1996) for relatively moist conditions, and was converted to a value for H_2O by dividing it by the ratio of the two diffusion coefficients D_{H2O}/D_{NO2} = 1.6 (Erisman *et al.*, 1994). The parameter b_{rs} indicates the radiation level where r_c equals twice the minimum canopy resistance. Its value was also taken from Eugster & Hesterberg (1996) (Table 1). For global radiation values R_s < 10 W m^{-2} a maximum r_c of 10 000 s m^{-1} was defined (Wesely, 1989).

Change in soil water storage The water storage S (m) between depths z_1 and z_2 ($\Delta z = z_2 - z_1$) was computed per unit area according to:

$$S = \frac{q(z_1) + q(z_2)}{2} \cdot \Delta z \tag{11}$$

where θ is the water content (m^3 m^{-3}) measured at depths z_1 (bottom) and z_2 (top), respectively, of the layer under consideration.

The weekly change in soil water storage ΔS between the beginning of a week t_1 and the beginning of the next week t_2 for a layer of thickness Δz is given by:

$$\Delta S = S(t_2) - S(t_1) \tag{12}$$

Finally, total water storage was estimated by integrating vertically over the total unsaturated zone and temporally over the full time period under consideration.

Computation of the hydraulic gradients The hydraulic gradient $\Delta H/\Delta z$ is calculated by Darcy's law (Darcy, 1856):

$$\frac{\Delta H}{\Delta z} = \frac{H(z_2) - H(z_1)}{z_2 - z_1} \tag{13}$$

where H is the hydraulic head in meters, $H = \Psi - z$, Ψ is the pressure head (m), and z is depth (m). The term $\Delta H/\Delta z > 0$ indicates a contribution to evapotranspiration while $\Delta H/\Delta z < 0$ indicates infiltration, and no flux separation exists if $\Delta H/\Delta z = 0$. The tensiometer measurements were used to determine $H(z)$ at four depths in the soil.

RESULTS AND DISCUSSION

Change in soil water storage

Time series of water storage in different soil layers are shown in Fig. 1 together with precipitation and the water table depth. The variation in soil water storage shows four principal forms of behaviour:

(a) In the uppermost 0.7 m, which is composed of fine material, the variations correlate highly with the precipitation input. Each precipitation event (Fig. 1(c)) directly increases the soil water storage in this layer (Fig. 1(a)), except for some short periods in winter where precipitation is received as snow and the surface is snow covered and partially frozen (this was only the case in February 1991; after three weeks the snow melted, which led to an increase in water storage in early March which was not related to the concurrent precipitation events).

(b) Between 0.8 and 2.4 m the water storage variation during winter is highly correlated with the precipitation, while in summer, the precipitation events have no direct influence on the water storage at this depth. Starting in April, the water storage in the layers below 1.1 m slightly decreases over the summer independently of precipitation inputs (Fig. 1(b)).

(c) Between 2.6 and 3.0 m the water storage (Fig. 1(d)) is highly influenced by the height of the groundwater table (Fig. 1(f)) which is high in summer and lower in winter, thus reversed to what one might expect for a flat surface.

(d) Below 3.2 m the water storage varies very little (Fig. 1(e)) because the water table is above 3.2 m in all seasons, except for cold periods with frozen ground or snow cover (February 1991).

In general, in winter (October 1990–March 1991), the soil water storage above 2.4 m depth is largely determined by the precipitation input whereas the subsoil below is primarily linked to the groundwater table. Conversely, in summer, the soil water storage declines steadily over time with a much less direct influence of precipitation events.

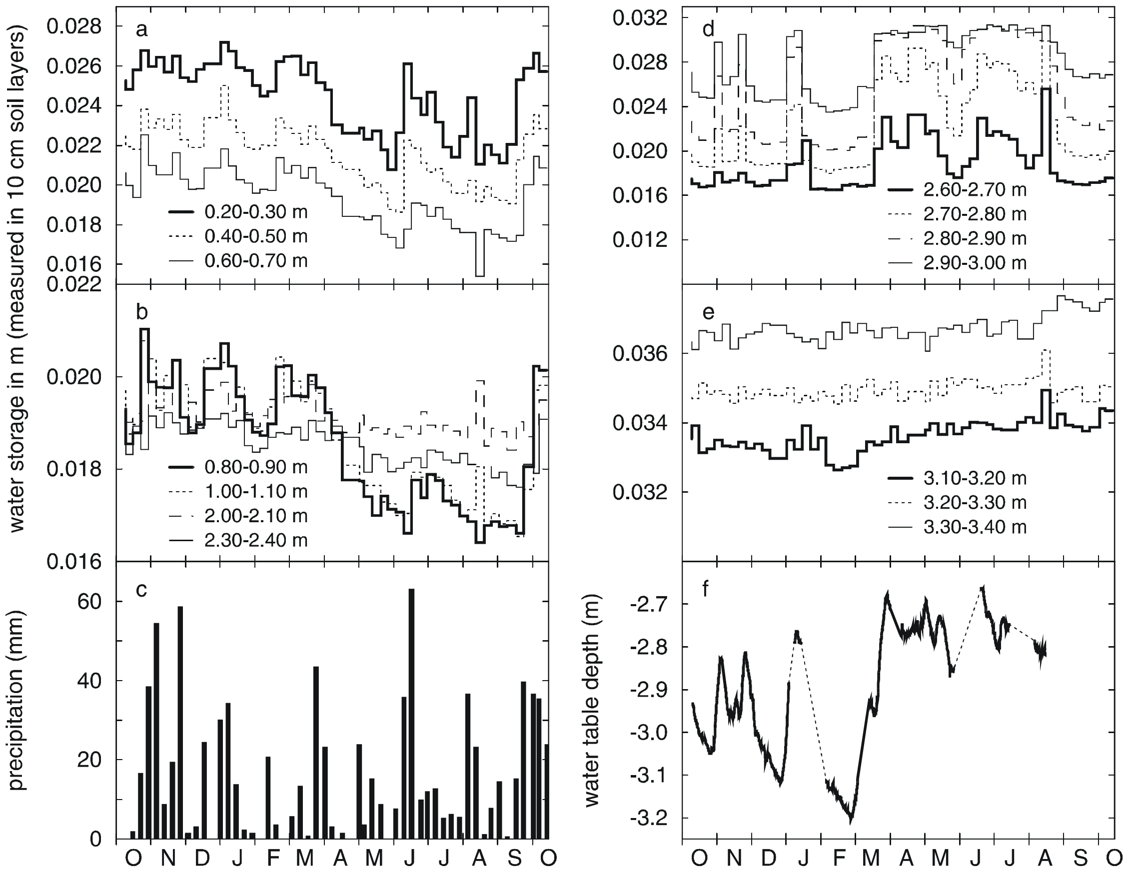

Fig. 1 Time course of the soil water storage (a, b, d, e) derived from neutron probe measurements using equation (11). Soil layers above 2.5 m (a, b) follow more closely the precipitation pattern (c) whereas the variations in the lower soil layers (d, e) follow the groundwater table depth (f).

Hydraulic gradients

Hydraulic gradients were measured from 15 October 1990 to 12 August 1991. After this period, the tensiometer measurements had to be discarded due to deterioration of the instruments which was influencing the measurements. However, since these measurements were not used for computing the annual water balance, this did not affect the overall goal of this project.

From October 1990 to April 1991 all hydraulic gradients are negative across the whole soil profile investigated (Fig. 2), indicating dominant net infiltration during this period. In summer, the hydraulic gradients between the surface and 0.55 m depth fluctuate around zero, indicative of periods with predominantly net evapotranspiration intermixed with periods of net infiltration. Conversely, below 0.55 m the gradients are always negative showing that the water losses contribute to recharging the groundwater body. This finding is consistent with the particle size distribution in the soil profile which showed two zones with differing texture. In fact, the lower zone is composed of coarser gravel than the upper zone, and the water transfer across the interface located at 0.7 m depth between the two zones is determined by the hydraulic conductivity.

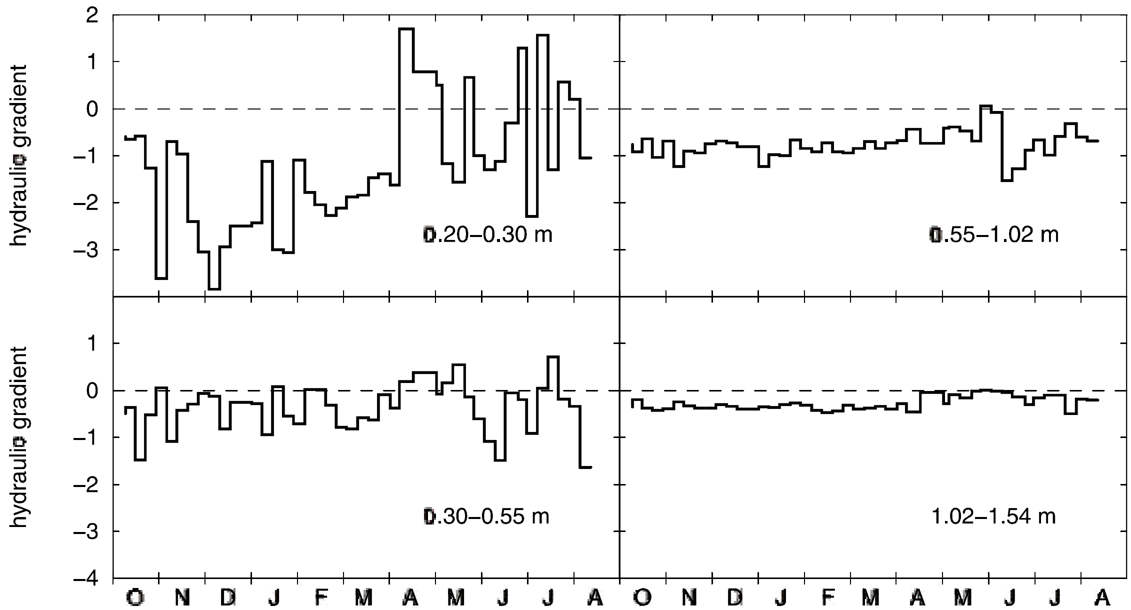

Fig. 2 Time course of the hydraulic gradient of four soil layers at depths ranging from 0.20 to 1.54 m between 15 October 1990 and 15 August 1991.

This textural examination of the soil allows one to complete the observations on water storage: the decrease in water storage observed during summer below 0.55 m corresponds to the infiltration towards lower layers during a time where precipitation has little effect on groundwater recharge. Thus, the water stored below this interface is not used for evapotranspiration and appears to be below the typical rooting depth of the vegetation (˜ 0.4 m).

Groundwater recharge

The weekly groundwater recharge R was calculated using Equation (1) in the rearranged form $R = P - E \pm DS$. The two methods used for estimating evapotranspiration E revealed the sensitivity of R to this component of the water budget. The disadvantage of the Penman-Monteith equation (Equation (6)) is that it only applies to active vegetation and is therefore not necessarily the best estimate during winter conditions with snow and frozen ground.

In contrast, Primault's (1962) empirical fit for seasonality takes this into account for Switzerland, where the equation was developed, and is therefore also applicable during winter conditions. However, the Primault method requires cross-validation with an independent method to ascertain that the empirical parameters used in equations (3)–(5) are valid at a given experimental site.

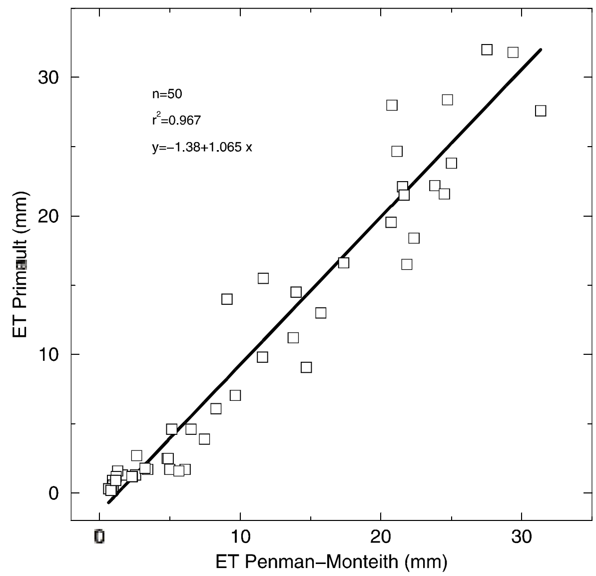

Fig. 3 Regression between the weekly evapotranspiration values obtained with the empirical approach by Primault (1962) and the values from the more physically-based Penman-Monteith approach.

Despite these important differences in approaches, the results obtained for the Areuse River delta were highly correlated ($r^2 = 0.967$, Fig. 3). The total evapotranspiration of the 12 months measuring period as was estimated with the Primault approach was 3% less than the one obtained with the Penman-Monteith approach. From this it can be concluded that the Primault method is a valuable and simple approach for longer-term studies where only standard meteorological measurements are available without energy balance data, the availability of which is essential for the use of the Penman-Monteith approach and which is one of the most limiting factors for its application. Therefore, without direct measurements of evapotranspiration (e.g. using the eddy correlation method) it cannot be conclusively said whether the Penman-Monteith method is a good reference for such a comparison during those periods, since low evapotranspiration values were mostly found during the cold season (Fig. 4) where vegetation is not very active.

Table 2 The hydrological budget of the Areuse River delta for the period 15 October 1990–15 October 1991 using two different estimation for estimating evapotranspiration.

	mm		%	
Precipitation	865		100	
Evapotranspiration	532*	564†	62*	65†
Change of water stored in the soil	19.6		2	
Groundwater recharge	313*	281†	36*	33†

* empirical approach by Primault (1962).
† Penman-Monteith approach

About one third (33–36%) of the annual precipitation contributes to groundwater recharge depending on the evapotranspiration estimation method (Table 2). Evapotranspiration accounts for 62–65% of the water budget, and the remaining 2% were found in ΔS (Table 2). These results are in agreement with those obtained by Christe

(1990) who measured 769 mm precipitation from 19 October 1989 to 25 August 1990 at the same location, whereof 32% contributed to groundwater recharge, 65% were used by evapotranspiration, and 3% contributed to the change in soil water storage.

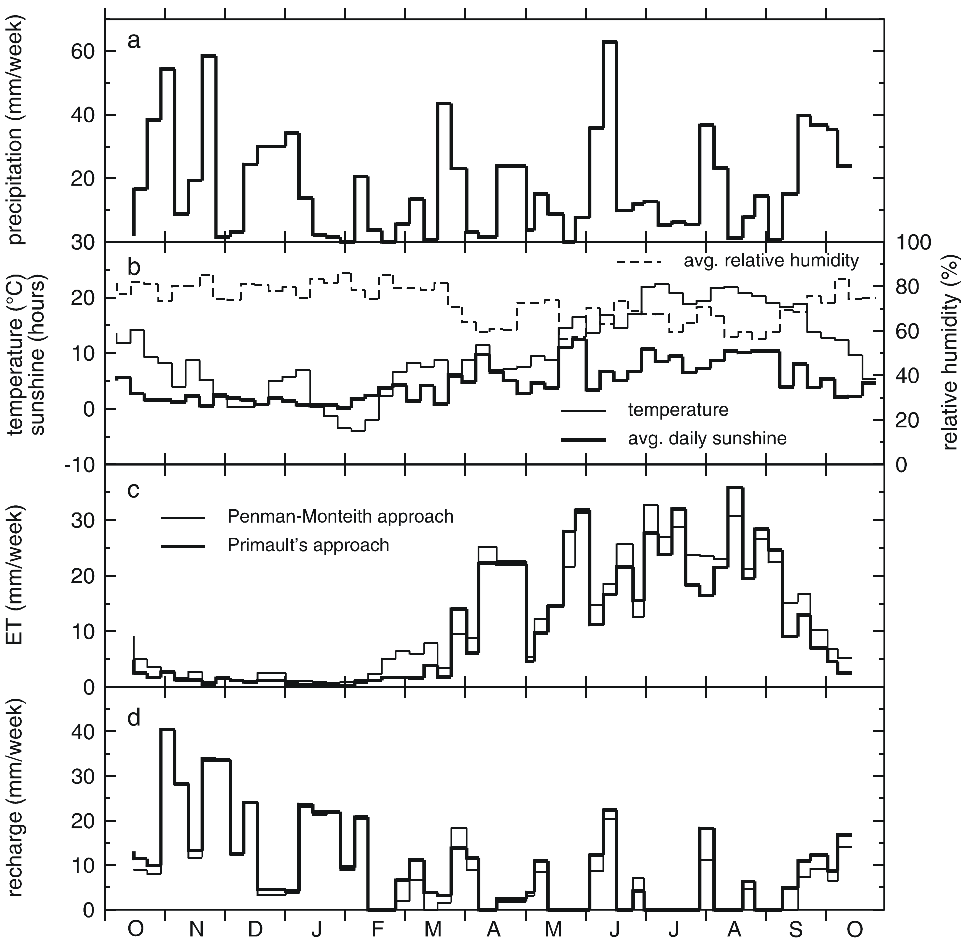

Fig. 4 Water balance of the Areuse delta between 15 October 1990 and 15 October 1991 (weekly sums or averages): (a) precipitation sum; (b) averages of temperature, relative humidity and daily sunshine duration; (c) evapotranspiration sum; (d) recharge. Evapotranspiration was estimated by two approaches, hence there are also two variants for the recharge estimates.

The total precipitation of 865 mm measured during our study period is very close to the 852.1 mm registered at the automatic weather station of the MeteoSwiss in Neuchâtel. However, it should be noted that the year of our measurements was moderately dry and only yielded 89% of the long-term annual precipitation as measured at the Neuchâtel weather station during the 30-year period 1961–1990.

During the summer period 1 April–31 August, 306 mm of precipitation were collected, which corresponds to 35% of the total precipitation collected during the whole experimental period (15 October 1990–15 October 1991). The evapotranspiration loss computed with the Primault and the Penman-Monteith approach corresponds to 135 and 134%, respectively, of the precipitation collected between 1 April and 31 August, whilst the fraction of precipitation available for groundwater recharge is 35 and 33%, respectively, during this period. In both cases, this corresponds to roughly 12% of total rainfall measured during the full period 15 October 1990–15 October 1991. This is a rather high fraction given the fact that groundwater recharge in summer is only seen during periods with high precipitation input. Thus, evapotranspiration revealed to be the dominant component in the water balance of the Areuse River delta during the warm season.

For comparison, almost the same precipitation amount was measured between 15 October 1990 and 15 January 1991 (303 mm). However, the fraction used for evapotranspiration was much lower (5% or 8% according to the Primault and the Penman-Monteith approach, respectively, corresponding to 2–3% of the annual precipitation). As a consequence, the recharge accounted for 80% or 77%, respectively, during this period. In winter the precipitation events and amounts were generally more evenly distributed over time, and their intensities were less variable than in summer.

Due to the fact that the groundwater body is regenerated primarily during winter, whereas the highest demand for tap water from the surrounding communities is in summer, it is difficult to realize a sustainable water resource management in the Areuse River delta region without having direct and long-term measurements of evapotranspiration. The long-term experience obtained from the nitrate concentration measurements conducted between 1979 and 1989 showed that it took almost a decade to let the concentrations drop back to acceptable levels after a ban of slurry and fertilizer applications to this area. Over this period a decrease of the nitrate concentration from 40 mg l^{-1} in 1981 to 15 mg l^{-1} in 1989 was achieved (Mdaghri-Alaoui, 1998). This indicates that evapotranspiration measurements should cover at least a similar time period to assess the sustainability of the water resource management in this area.

CONCLUSIONS

The water balance over a one-year period (15 October 1990–15 October 1991) was evaluated via weekly measurements of soil moisture using a neutron probe, and by estimating evapotranspiration from standard meteorological data. Additionally, the hydraulic gradients were investigated using tensiometer measurements.

The fraction of precipitation contributing to groundwater recharge was estimated to be 33–36%. Both the soil water storage determined from neutron probe measurements and the hydraulic gradients derived from tensiometer measurements revealed the same time periods with significant infiltration into the soil which tend to be more frequent in winter. The calculations of water recharge gave satisfactory results, except during the short period with snow cover, which is not adequately taken into account by the empirical water balance. In winter the soil water storage above 2.4 m depth is positively correlated with the precipitation input whereas the subsoil below is primarily linked to the groundwater table.

The observed summer trend of decreasing water storage between the surface and 0.70 m depth is due to a net infiltration that contributed to groundwater recharge despite the frequent short periods with net evapotranspiration losses from this zone, while no evapotranspiration was observed below this depth. This observation was consistent with the particle size analysis which showed a significant change in soil texture from fine material on the top of a coarse and highly permeable soil horizon below 0.70 m.

The evapotranspiration as determined with the Primault (1962) approach was only 3% below the one obtained with the Penman-Monteith approach, indicating that Primault's (1962) empirical approach is an easy-to-use and efficient method to estimate evapotranspiration in Switzerland.

Despite the unambiguous behaviour of infiltration and groundwater recharge during the experimental period the results clearly showed that for long-term water resource management a much longer observation period would be needed to include the unknown influences of inter-annual variability especially in winter precipitation.

While the vertical heterogeneity of the soil was addressed by the experimental set-up chosen for this study, the uncertainty associated with horizontal spatial heterogeneity of the unsaturated soil remains unknown. Because a high density network of soil pits is non-realistic due to its high costs and the heavy disturbance of the environment, it is suggested to combine numerical modelling with longer-term measurements at one site to obtain a reliable data basis for water resource management in this area.

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Field determination of the water balance of the Areuse River delta, Switzerland

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Corrigenda

Page 751, between Eqs. (6) and (7): λ is the latent heat of vaporization; ρ_a is the density of air; γ is the psychrometric constant.

Page 752, bottom: θ is the water content.

Page 753, top: soil water storage is ΔS (not $?S$); layer of thickness is Δz (not $?z$). The hydraulic gradient is $\Delta H/\Delta z$ (not $?H/?z$); the hydraulic head is $H = \Psi - z$, where Ψ is the pressure head.

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Page 751, typographic error in Eq. (5): the second term in this equation is lacking the sin before the brackets; the correct equation should read

$$\begin{aligned} c &= -0.5068 \cdot \sin\left(\frac{2\pi}{365}\mathrm{DOY} + 0.5593\right) \\ &\quad -0.0711 \cdot \sin\left(\frac{4\pi}{365}\mathrm{DOY} + 0.6112\right) + 0.6271\ . \end{aligned} \tag{5}$$

The correct equation lead to the following form of c: