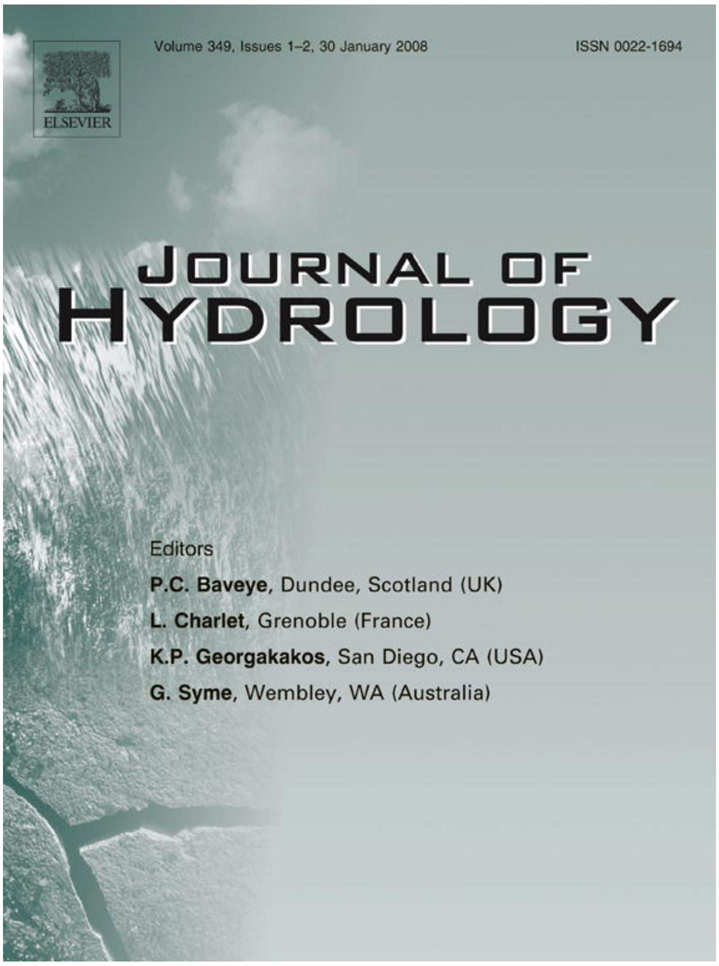
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Riparian zone evapotranspiration estimation from diurnal groundwater level fluctuations

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Summary Riparian vegetation typically has a great influence on groundwater level and groundwater-sustained stream baseflow. By modifying the well-known method by White [White, W.N., 1932. Method of estimating groundwater supplies based on discharge by plants and evaporation from soil — results of investigation in Escalante Valley, Utah — US Geological Survey. Water Supply Paper 659-A, 1–105] an empirical and hydraulic version of a new technique were developed to calculate evapotranspiration (ET) from groundwater level readings in the riparian zone. The method was tested with hydrometeorological data from the Hidegvíz Valley experimental catchment, located in the Sopron Hills region at the western border of Hungary. ET rates of the proposed method lag behind those of the Penman–Monteith method but otherwise the two estimates compare favorably for the day. At nights, the new technique yields more realistic values than the Penman–Monteith equation. On a daily basis the newly-derived ET rates are typically 50% higher than the ones obtainable with the original White method. Sensitivity analysis showed that the more reliable hydraulic version of our ET estimation technique is most sensitive (i.e., linearly) to the laboratory- and/or slug-test derived values of the saturated hydraulic conductivity and specific yield taken from the riparian zone.

Introduction

The diurnal cycle of the climate forcing, such as temperature, solar radiation, and humidity often induces a similar daily fluctuation in the groundwater level of the riparian zone and, especially during drought periods, in the flow rate of the adjacent low-order stream through the mediation of

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the riparian vegetation. This vegetation effect on the groundwater levels and baseflow rates occurs as a result of a daily rhythm in the metabolism of the vegetation modulated by phenological changes through the seasons. Metabolic changes in the vegetation are accompanied by similar changes in transpiration rates. In riparian forests of dense cover, soil evaporation during drought periods is often negligible in comparison with the transpiration rates of vegetation. Several authors have investigated the linkage between riparian transpiration and streamflow rates (Troxell, 1936; Croft, 1948; Tschinkel, 1963; Reigner, 1966; Pörtge, 1996; Lundquist and Cayan, 2002; Loheide et al., 2005; Butler et al., 2007; Boronina et al., 2005) but only a few attempted to estimate the evapotranspiration (ET) rate

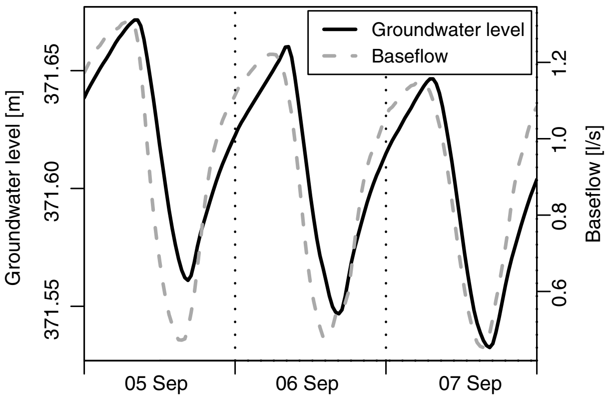

Figure 1 Observed diurnal fluctuations in riparian groundwater level and baseflow values.

of the riparian zone (White, 1932; Bond et al., 2002; Bauer et al., 2004; Nachabe et al., 2005) by using of the observed streamflow, groundwater or soil moisture fluctuations or to provide an analytical description of these signals (Czikowsky, 2003; Czikowsky and Fitzjarrald, 2004).

A typically observable diurnal pattern in groundwater level and streamflow rate is displayed in Fig. 1 for a forested riparian zone in western Hungary. The maxima occur in the morning hours, between 6 and 8 a.m., and the minima in the afternoon, between 4 and 5 p.m. (Gribovszki et al., 2006). Both signals are characterized by sharp lower extrema, but the peak regions of the streamflow signal are more rounded. Notably the two extrema do not overlap perfectly in time, the groundwater extrema lag *behind* those of the streamflow rate by about 1—1.5 h. To our best knowledge no such lag has ever been reported in the literature before. In an accompanying paper by Szilágyi et al. (2007) to this present work the problem is further investigated.

The transpiration demand of the vegetation is generally met by the soil moisture of the riparian zone and/or directly by the groundwater. In drought periods the groundwater of the riparian zone used by evapotranspiration is typically replenished via groundwater flow from areas farther away from the stream (Fig. 2) or by so-called induced recharge, when, due to a reversed hydraulic gradient, the groundwater flow is directed from the channel toward the riparian zone. Around the timing of the groundwater level extrema, supply, Q_{net} [LT^{-1}], and demand, ET [LT^{-1}], are in an equilibrium in Eq. (1)

$$\frac{dS}{dt} = Q_{net} - ET \qquad (1)$$

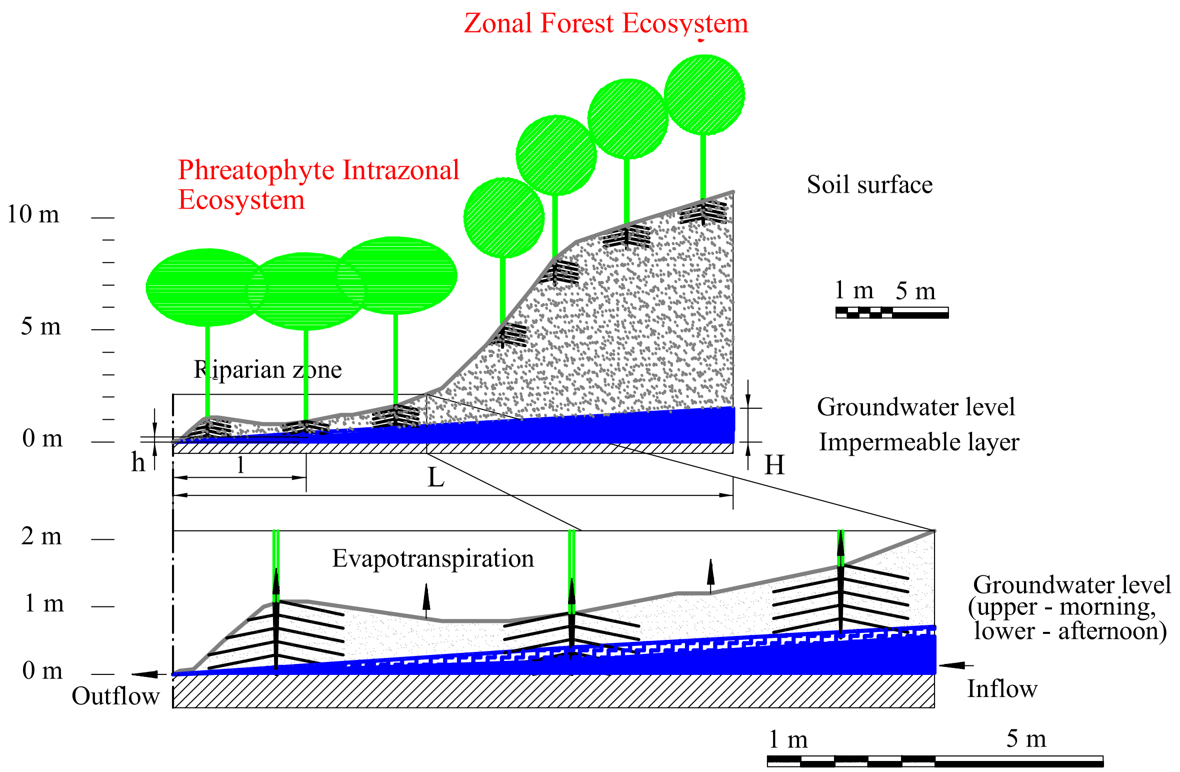

Figure 2 Schematic model of the riparian zone.

where S [L] is the stored water volume per unit area. Around the extrema $dS/dt = 0$, $dS/dt > 0$ and $Q_{net} > ET$ on the rising limb and $dS/dt < 0$ and $Q_{net} < ET$ on the falling limb of the groundwater hydrograph. The ET rate is largest during the day when the groundwater level curve is about the steepest on the descending limb, which is typically close to the radiation maxima. The smallest ET rate however does not necessarily take place when the ascending limb of the groundwater level signal is the steepest, rather, just prior to dawn when vapor pressure deficit is at its diurnal minimum. The minima in the riparian zone groundwater level is accompanied by the steepest hydraulic gradients, so when ET starts to decrease, the steep hydraulic gradient can deliver water to the riparian zone very efficiently, replenishing it fastest right after the occurrence of the groundwater level minimum, thus somewhat (but not entirely) independent of the actual ET rate, at least for a while (for more detail, see the accompanying paper by Szilágyi et al., 2007). This mechanism is especially true in valley settings (Fig. 2), where farther away from the stream the groundwater level is deeper below the surface, thus, being less affected by the diurnal fluctuations in the transpiration rate of vegetation.

White (1932) published a method of estimating riparian ET rates on a daily time step based on fluctuations in the groundwater level. He assumed that during the predawn/dawn hours when ET is negligible, the rate of the observed groundwater level increase is directly proportional to the rate groundwater is supplied to the riparian zone from the neighboring areas. The slope, r [LT^{-1}], of the tangential line drawn to the groundwater level curve in these sections (Fig. 3), multiplied by specific yield value, S_y [–], of the riparian zone, therefore, represents the rate of water supply to a unit area. By extending the tangential line over a 24-h period and taking the difference in groundwater levels, one would obtain an estimate of the total water supply to the unit area over a day. The so-obtained daily rate of water supply must typically be modified by s [L], the difference in the observed groundwater level over the 24-h period, since it rarely happens that the groundwater level returns to the same elevation a day before. The daily ET rate this way is obtained as

$$ET = S_y(24r \pm s) \tag{2}$$

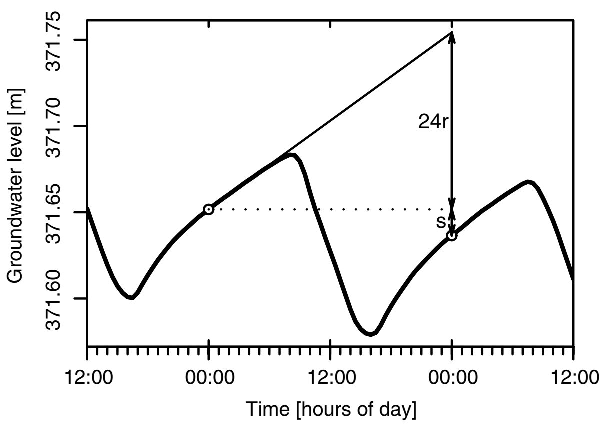

Figure 3 The basic principle of the original White method.

where r is the mean hourly rate of groundwater level increase from midnight to 4 a.m.

Meyboom (1964) suggested a 50% reduction (which he called the readily available specific yield) of the laboratory-derived specific yield value in Eq. (2). A reduction is certainly justified since it takes some time for the drainage to adjust to any new conditions introduced depending on such variables as soil-aquifer type, the thickness of the vadose zone, as well as aquifer and stream geometry (e.g., Szilágyi, 2004).

Based on the study of Nachabe (2002), Loheide et al. (2005) suggested certain guidelines and an equation to obtain S_y as a function of sediment texture, depth to the groundwater table and elapsed time (t) of the drainage. The relationship is based on the Brooks and Corey (1964) model that gives the volumetric water content (θ) as

$$\theta(\Psi) = \theta_R + (\theta_S - \theta_R)\left(\frac{h_a}{\Psi}\right)^{\lambda} \tag{3}$$

where h_a [L] is the air entry pressure, Ψ [L] is the pressure/suction head, λ [–] is the pore-size index, θ_S [–] is the total porosity, and θ_R [–] is the residual water content. With the help of Eq. (3) S_y becomes

$$S_y(t) = \frac{Kt}{\Delta h}\left[\left(\frac{\theta_B - \theta_R}{\theta_S - \theta_R}\right)^{\frac{2+3\lambda}{\lambda}} \left(\frac{\theta_{Surface} - \theta_R}{\theta_S - \theta_R}\right)^{\frac{2+3\lambda}{\lambda}}\right] + (\theta_S - \theta_R)\left(1 - \frac{\theta_B - \theta_R}{\theta_S - \theta_R}\right) \tag{4}$$

where K [LT^{-}1] is the saturated hydraulic conductivity of the soil, Δh is the change in the groundwater surface position over time t, and $\theta_{Surface}$ is the actual water content at the surface, which will depend on the actual depth to the groundwater table. θ_B [–] is an additional parameter of the water content profile defined as

$$\theta_B(t) = \theta_R + (\theta_S - \theta_R)\left(\frac{\Delta h(\theta_S - \theta_R)}{\frac{2+3\lambda}{\lambda}K}\right)^{\frac{1}{\frac{2+3\lambda}{\lambda}-1}} t^{\frac{1}{1-\frac{2+3\lambda}{\lambda}}} \tag{5}$$

Here K is calculated with the help of the Kozeny–Carman relationship as $K_S = B\theta_e^n$, where θ_e is the effective porosity (i.e., total porosity minus the water content at a pressure head of -33 kPa) and B [LT^{-}1] and n [–] are empirical parameters. When K is in cm h^{-}1 then n and B obtain values of 4 and 1058, respectively.

Loheide et al. (2005) demonstrated via numerical modeling experiments that the ET rate given by the White method is not influenced perceptably by the geometry of the vadose zone. Although the White method yields reasonable estimates of daily ET, provided an appropriate S_y value is employed Loheide et al. (2005), it has a weak point. Namely, it assumes that water supply to the riparian zone would happen at a constant rate, observable when ET is negligible (Fig. 3), over the day. As mentioned earlier, this is hardly the case since the hydraulic gradient changes over the course of the day as the riparian zone groundwater level fluctuates. This fluctuation causes a time-varying water supply to the area, since, especially in a valley setting with predominantly horizontal groundwater flow, groundwater levels farther away from the stream (i.e., beyond the riparian zone where depth to the groundwater is larger) typically express much diminished diurnal fluctuations (Fig. 2), thus causing diurnally varying hydraulic gradients between the valley-side and the riparian zone. The same head fluctuations may be true for a deep aquifer case with mainly vertical flow in the riparian zone.

While the White method (and its modification presented below) aims to describe storage changes within the saturated zone only, implicitly it accounts (at least partially) for moisture withdrawal (depending on the depth of the water table) from the vadose zone as well. Vadose zone ET does not come solely from the vadose zone soil moisture because this soil moisture always has a close hydraulic connection with the groundwater table through the capillary fringe, therefore it is indirectly included in the water balance of the groundwater system. Loheide et al. (2005) demonstrated via numerical experiments that water extracted from a 1-m-thick vadose zone shows up in the ET estimates of the White method so that it accounts for 19–23–28% (i.e., for silt, loam, and medium sand, respectively) of the total vadose zone water extraction. This however should not be surprising since moisture extraction from the vadose zone can depress the groundwater table due to a reversed hydraulic gradient.

Shah et al. (2007) performed numerical simulations to partition total ET into vadose zone and groundwater ET. They found that for a water table within half meter of the land surface, nearly all ET came from the groundwater due to the close hydraulic connection between the unsaturated and saturated zones. Depending on the soil type, they also reported a decoupling of the groundwater and unsaturated zone moisture dynamics starting at water table depths between 0.3 and 1.0 m for deep-rooted vegetation.

Hydraulic theory-based modification of the White method for riparian zone ET estimation

Since drought period water supply to the riparian zone is typically regulated by the hydraulic gradients between the background (i.e., away from the stream and the riparian zone) and the area in question, as well as between the same area and the stream, it must be included in the ET estimation method. The lumped version of the mass-conservation equation Eq. (1), can also be written as

$$\frac{dS}{dt} = S_y \frac{dh}{dt} = Q_{net} - ET \tag{6}$$

where h [L] is the groundwater elevation within the control (unit) area. The net water supply, Q_{net}, is defined as the difference between in- (Q_{in}) and outflows (Q_{out}) to it. The latter flow rates (assuming mainly horizontal flow) are formulated by Darcy's equation using the Dupuit approximation (Harr, 1962; Kovács, 1972)

$$Q_{net} = Q_{in} - Q_{out} = \frac{k(H^2 - h^2)}{2d_1(L - l)} - \frac{k\left(h^2 - h_0^2\right)}{2d_1 l} \tag{7a}$$

Here H [L] is the groundwater elevation in the background (where diurnal fluctuations are not apparent) at a distance L [L] from the stream, l [L] is the distance from the control area to the stream, h_0 [L] is the water level in the stream (Fig. 2), k [LT^{-1}] is the (preferably slug-test derived) saturated hydraulic conductivity of the soil-aquifer system, and d_1 [L] is the unit distance along the stream. H, h and h_0 are taken relative to an assumed horizontal impervious layer (Fig. 2) not necessarily at the streambed elevation. Note that the above formulation of Eq. (7a) assumes successive steady-state conditions at each time step of the Q_{net} calculations, which is not strictly true.

When the flow is mainly vertical in the riparian zone (i.e., deep aquifer case) Eq. (7a) can be substituted by Darcy's equation

$$Q_{net} = k_v \frac{H - h}{l'} \tag{7b}$$

where H now is the total ('background') head at a depth of l' below the reference level, and k_v is the saturated hydraulic conductivity value for vertical flow.

Before the application of the hydraulic theory, one has to decide about the location at which the groundwater levels must be observed and its temporal derivatives be computed. As Bauer et al. (2004) and Loheide et al. (2005) demonstrated, the middle part of the riparian zone expresses the least spatial variations and represents average conditions as long as the riparian zone vegetation is fairly homogeneous. They also note that boundary-condition effects (such as a heavily damped signal of diurnal groundwater level fluctuations near the channel) typically die out within a few meters from the stream.

This way the steps of the suggested new ET estimation approach are the following. First the groundwater level record is differenced in time (half-hourly or hourly time steps are convenient) to obtain dh/dt. This new time-series (Fig. 4) can be assumed to be directly proportional to the difference between water supply (Q_{net}) and demand (ET) over the riparian zone. Q_{net} is derived next in two different ways: as an *empirical*, and as a *hydraulic* approach.

In the *empirical* approach the maximum of Q_{net} for each day was calculated by selecting the largest positive time-rate of change value in the groundwater level readings such

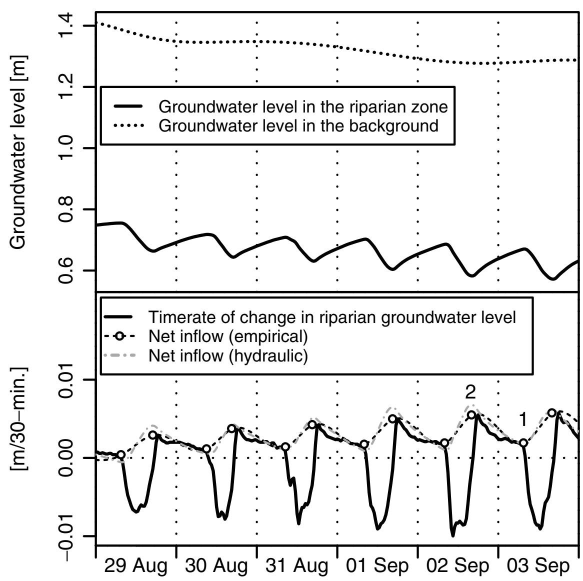

Figure 4 Measured groundwater levels within the riparian zone, calculated time-rate of change in the measured groundwater level values, and the Q_{net} water supply estimates, divided by S_y for better comparison.

as $Q_{net} \approx S_y\, dh/dt$, while the minimum was obtained by calculating the mean of the smallest time-rate of change in h taken in the predawn/dawn hours. The averaging is necessary in order to minimize the relatively large role of measurement error when the changes are small. The resulting values of the Q_{net} extrema in Fig. 4 then were assigned to those temporal locations where the groundwater level extrema took place. It was followed by a spline interpolation of the Q_{net} values to derive intermediate values between the specified extrema. Most probably the resulting empirical maxima are somewhat smaller than the corresponding actual maximum supply rates by virtue of the ET term being unaccounted for in Eq. (6) in this empirical method. At the same time, the estimated minima are somewhat larger than the actual minimum supply rates, due to the necessary averaging and due to observational evidence that groundwater levels reach their maxima somewhat later, i.e., between 6 and 8 a.m. in the summer. However, the dh/dt values of this period (i.e., between 6 and 8 a.m.) should not be used because by that time ET may have already become significant, thus leading to increased dh/dt rates, groundwater level values not affected by ET were chosen from the predawn/dawn hours for analysis.

The *hydraulic* approach estimates Q_{net} from Eqs. (6) and (7a) or (7b). In case of a predominantly horizontal flow the k, h, and l values are typically known from measurements, but the values of H and L need to be determined. The latter distance is largely determined by the width of the riparian zone. In case of a predominantly vertical flow situation (i.e., deep aquifer case) the k_v and h values are known typically from measurements, but the values of H and l' may be estimated again. In lieu of measurements, l' can be taken equal to the thickness of the aquifer below the stream.

The corresponding H values can then be obtained from Eq. (6) in combination of Eq. (7a) or (7b), realizing that in the predawn/dawn hours ET is close to zero in Eq. (6), as

$$H = \sqrt{2(L-l)\left(\frac{S_y d_1}{k}\frac{dh}{dt} + \frac{h^2 - h_0^2}{2l}\right) + h^2} \qquad (8a)$$

$$H = \frac{S_y}{k_v}\frac{dh}{dt}l' + h \qquad (8b)$$

which, thus, yields an estimate for the 'background' groundwater elevation/head each day. Note that during a drought period even this background groundwater elevation changes from day-to-day along a typically slow recession curve (Fig. 4) for the horizontal flow scenario. Similar slow changes can be expected for H in the deep aquifer case as well. To obtain intermediate H values, again a spline interpolation was employed in Fig. 4. The subsequent Q_{net} values over the day are then obtained from Eq. (7a) or (7b) by making use of the interpolated H values. In the present study it was possible to check the accuracy of the H estimates by comparing them to well readings along the valley slope (Fig. 5), and a good agreement was found with Eq. (8a).

If one has information of the groundwater flow direction within the riparian zone, which often times deviates significantly from a direction perpendicular to the stream, then the l and L distances must be taken along that direction of the groundwater flow. This situation is investigated in more detail later during the sensitivity test of the method.

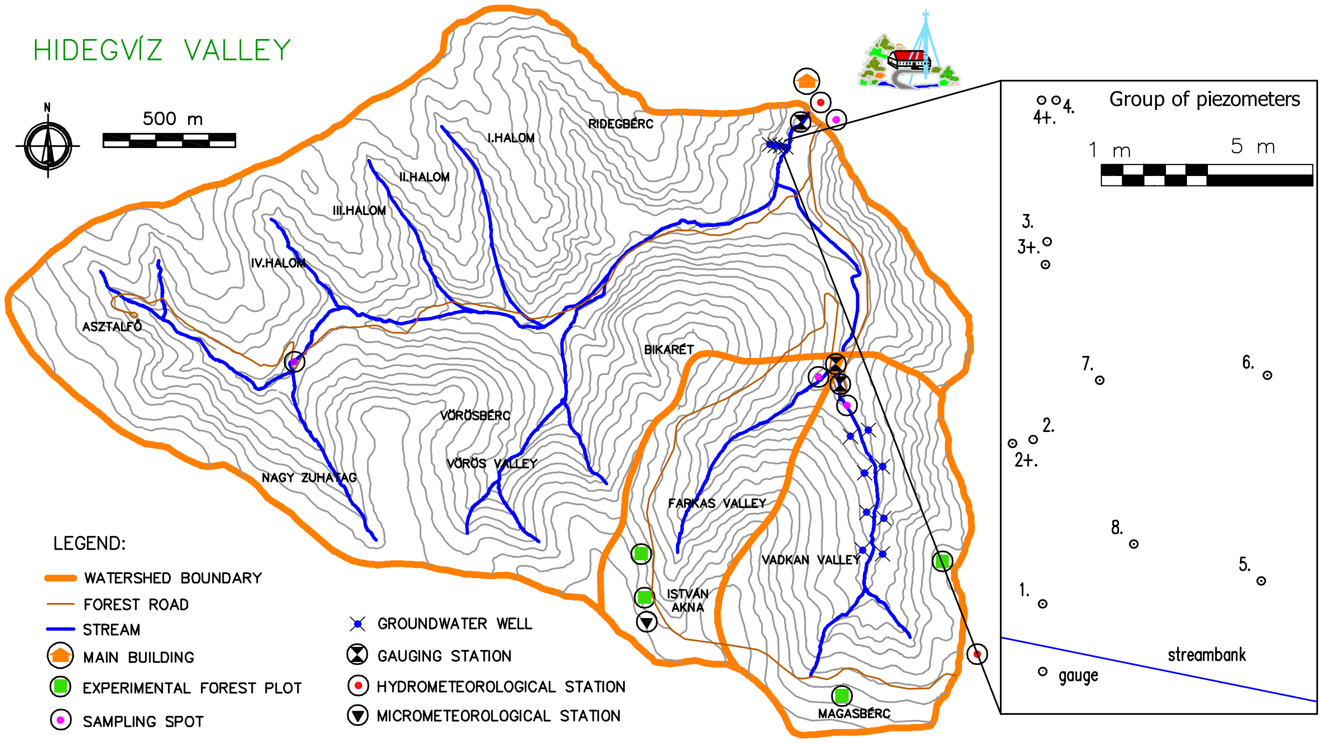

Figure 5 The experimental catchment and the location of the groundwater wells as well as the micrometeorological station employed in the study.

Finally, the ET rates, characteristic of the riparian zone, can be obtained by rearranging Eq. (6) as

$$ET = Q_{net} - S_y \frac{dh}{dt} \quad (9)$$

For the present ET estimation method the importance of a continuous record of high accuracy groundwater level measurements at a high temporal resolution cannot be stressed enough because differentiation of the groundwater level record may invoke large errors in the resulting ET estimation whenever the original groundwater level measurements are inaccurate. In order to reduce this uncertainty, the application of a low-pass numerical filter (smoother) is recommended. Care must however be taken not to oversmooth the data because it can lead to loosing important details about the nature of the diurnal fluctuations. A recommended approach is to collect measurements at the largest possible frequency and apply a filter accordingly. For example, if one would like to use 30-min data for analysis then the sampling interval should be at least 10 min.

Application of the ET estimation method to a small experimental catchment in Hungary

Both the *empirical* and *hydraulic* versions of the proposed ET estimation method were tested at a small (drainage area is 6 km^2) experimental watershed (Fig. 5) in the Sopron Hills of western Hungary.

The geology of the catchment is made up of strongly unclassified crystalline bedrock deposited in the tertiary (Miocene) period, along with fluvial sediments deposited in five distinct layers. On the surface only the two upper layers of the latter appear. Over the slopes and hilltops the so-called Felsőtödl Gravel Formation is found in a 10–50 m-thick layer. This layer contains coarse gravel and fine loam as well, so it is strongly unclassified. In the valley bottoms, a finer-grained layer, the Magasbérc Sand Formation appears, which is a good aquifer, giving rise to perennial streams (Kisházi and Ivancsics, 1985).

The riparian zone vegetation in the valleys is a typical phreatophyte intrazonal ecosystem dominated by alder (*Alnus glutinosa* (L.) *Gaertn.*). The mean height of the young- to middle-aged riparian forest stand is about 15 m with a mean trunk diameter (at a height of 1.3 m) of 13 cm. Leaf area index (LAI) of this forest stand was 7.4.

The area enjoys a sub-alpine climate, with daily mean temperatures of 17 °C in the summer, and 0 °C in the winter, and with an annual precipitation of 750 mm, late spring and early summer being the wettest and fall the driest seasons (Danszky, 1963; Marosi and Somogyi, 1990).

The depth to the groundwater in the riparian zone varies between 60 to 90 cm during typical drought periods. Consequently, the root system of the trees is in direct contact with the saturated zone, or at least the capillary fringe throughout the year. Following Shah et al. (2007), the decoupling of the groundwater dynamics from the vadose zone in the soil of our experimental site was found to start at a depth of 0.8–0.9 m, therefore almost all year long the total ET is very close to groundwater ET.

The groundwater measurements for the study took place at the north-eastern corner of the catchment in a well denoted by 2+ in Fig. 5, situated in the middle of an approximately 20-m wide riparian zone of the west bank of the stream. Groundwater levels (h) in that well were recorded by a pressure transducer at a 10-min sampling interval and with an accuracy of 1 mm. The well was dug with an 80-mm drill. The PVC well casing has a diameter of 63 mm, screened at the bottom 1 m, starting 25 cm below the surface. The space between the casing and the wall of the borehole is filled with coarse sand.

The parameter values derived for the study are listed in Table 1. The values of H were calculated by Eq. (8a) as a function of the measured h values in the well. For each day, the S_y values were estimated by Eqs. (3)–(5) based on the groundwater extrema values and the elapsed time between them. The required Brooks and Corey model parameter values in Table 1 were obtained by the help of laboratory-derived water-retention curves using samples taken from the location of the well. The parameter values of Eq. (3) were then adjusted by trial-and-error until a favorable match was obtained with the sample retention curves. In the calculation of the S_y value no hysteresis effect could be taken into consideration because we had only the drying curve of the soil water characteristics. The effective porosity value (θ_e) required by the Kozeny–Carman equation was also derived from the laboratory samples. As a result, S_y changed between 0.039 and 0.103, with a median value of 0.071, in the study period. Note that the Kozeny–Carman equation, when employing laboratory samples, yielded a saturated hydraulic conductivity value ($K = 2 \cdot 10^{-7}$ m s^{-1}) which is a magnitude smaller than the slug-test derived values in Table 2. This is not surprising considering the increasing importance of preferential flow with growing scale (e.g., Brutsaert and Nieber, 1977; Szilágyi et al., 1998).

Representative drought periods for the analysis were chosen from 2005. The ET estimates produced by the current method were compared with those of the Penman–Monteith method (Allen et al., 1998) at a 30-min resolution, and of the original White method, Eq. (2), on a daily basis. Although the present method calculates mainly

Table 1 Study site parameters for the ET calculations

	k^a (m/s)	l (m)	L (m)	h_0 (m)	λ (–)	θ_S (–)	θ_R (–)	θ_e (–)	B^b	n^b (–)
Employed (median) value	$1.8 \cdot 10^{-5}$	9.4	40	0.23	0.32	0.379	0.035	0.091	$2.94 \cdot 10^{-3}$	4
Observed range	$1.1 \cdot 10^{-6} - 2.9 \cdot 10^{-4}$	–	–	–	–	–	–	–	–	–

[a] Determined from 16 slug-tests (Schwartz and Zhang (2003)) and validated by inverse modeling (Kovács and Szanyi (2005)) against groundwater level readings of the surrounding piezometer nest.
[b] From Maidment (1993) (m/s).

Table 2 Sensitivity analysis of the ET estimates to parameter values of the methods applied for the growing season of 2005

Methods and parameters	May	June	July	August	September	October
	ET (mm/day)					
Penman–Monteith (PM)	6.41	6.04	11.13	7.56	7.34	2.84
1.2 · PM	7.69	7.25	13.36	9.07	8.81	3.41
Original White method	5.44	5.51	6.81	5.78	5.21	2.37
Empirical method	6.49	7.97	11.33	9.42	8.74	3.12
h_0 = 0.07 m, d_0 = 0 m, median k	6.26	7.05	8.56	8.05	7.24	2.73
h_0 = 0.57 m, d_0 = 0.5 m, median k	7.06	8.52	11.39	10.31	9.72	3.38
h_0 = d_0 = 0.5 m, median k	7.06	8.52	11.39	10.31	9.72	3.38
h_0 = 0.7 m, d_0 = 0.5 m, median k	7.06	8.52	11.39	10.31	9.72	3.38
h_0 = 0.37 m, d_0 = 0.3 m, median k	6.76	7.86	10.49	9.34	8.61	3.15
Flow direction 40° to the stream, h_0 = −0.03 m, l = 14.6 m, d_0 = 0.5 m[a], median k (probably close to reality)	6.70	7.82	10.14	9.28	8.52	3.05
Flow direction 40° to the stream, h_0 = −0.23 m, l = 14.6 m, d_0 = 0.5 m[a], median k (likely steeper streambed gradient than reality)	6.70	7.82	10.14	9.28	8.52	3.05
L = 20 m, h_0 = d_0 = 0.3 m	7.26	8.85	12.19	10.74	10.24	3.61
L = 110 m, h_0 = d_0 = 0.3 m	6.60	7.55	9.93	8.87	8.07	2.99
$k = 1.1 \cdot 10^{-6}$ m/s (minimum), h_0 = d_0 = 0.3 m	6.18	6.41	7.85	7.34	6.31	2.43
$k = 2.9 \cdot 10^{-4}$ m/s (maximum), h_0 = d_0 = 0.3 m	29.58	47.92	78.61	62.07	67.16	20.66

For median k value see Table 1. d_0 means the depth to datum under the streambed.

[a] Here the d_0 and h_0 values are related to different cross-sections of the stream due to the downstream component of the groundwater flow path. When the reference level is taken at d_0 = 0.5 m below the streambed at the stream cross-section closest to the well, the streambed (therefore the stream-water level, h_0) may drop below that reference level downstream where the groundwater flow path intersects the stream.

groundwater ET, the estimated ET values are very close to total ET rates obtained by the Penman–Monteith equation.

Hughes et al. (2001) found the Penman–Monteith method to be one of the most reliable in estimating evapotranspiration from densely vegetated surfaces. The Penman–Monteith method derives ET as

$$\mathrm{ET} = \frac{\Delta(R_0 - S) + \rho c_p \mathrm{VPD} r_a^{-1}}{L_v[\Delta + \gamma(1 + r_c r_a^{-1})]} \tag{10}$$

where the Penman–Monteith ET is in mm day^{-1}, L_v, is the latent heat of vaporization (MJ kg^{-1}), Δ is the slope of the saturation vapour pressure curve (kPa °C^{-1}), γ is the psychrometric constant (kPa °C^{-1}), R_0 is the net radiation (MJ m^{-2} day^{-1}), VPD is the vapour pressure deficit (kPa), S is the soil heat flux and temporary storage of energy into the tree itself (MJ m^{-2} day^{-1}), ρ is the air density (kg m^{-3}), c_p is the specific heat of moist air (kJ kg^{-1} °C^{-1}), r_a is the aerodynamic resistance (s m^{-1}), and r_c is the bulk canopy resistance (s m^{-1}).

Data required by the Penman–Monteith method were obtained from a micrometeorological station which, however, is not situated in the valley but rather on a hillslope, 1.9 km to the south from the riparian zone studied (Fig. 5). Because the tree canopy is 10–15 m above the ground in the study catchment, soil heat flux contributions to the available energy for the canopy were considered negligible. The temporary storage of energy into the tree trunks and limbs was estimated at 5% of the solar radiation (Goodrich et al., 2000). Goodrich et al. (2000) also recommend to employ a high r_c value (i.e., 5000 s m^{-1}) for the nighttime period so as to extinguish ET at night. But this probably is not realistic because nocturnal sap flow measurements indicate that nighttime ET could be as high as 10–25% of the daily totals (Gazal et al., 2006). As a consequence, positive ET values during the night, as our estimates suggest (Figs. 6 and 7), can be quite realistic. Seasonal changes in the r_c value were calculated based on LAI measurements (r_c = 200/ LAI [Allen et al., 1998]) over the growing season, and here it was also assumed Goodrich et al. (2000) that before foliation and after defoliation r_c = 1000 s m^{-1}.

Note that the Penman–Monteith ET values maybe significantly different from the riparian ET estimates for objective reasons and not only for possible deficiencies in the proposed method because the meteorological tower is at a considerable distance from the riparian zone and has a significantly more exposed location on a hillslope. The PM method estimates total ET (i.e., groundwater plus vadose zone ET) and not only the portion of ET that derives from the groundwater. Furthermore, groundwater and soil moisture conditions maybe quite different on a hillslope where the tower is located from the ones characteristic of the riparian zone within a valley setting. All said, the Penman–Monteith estimates still represent ''real-world'' ET values for a well-watered vegetation that can be used as a bench-mark to compare the estimates of the proposed method with.

In the 2005 growing season (May–October) there were altogether 100 days (Fig. 6) that could be included in the analysis. Days with less than 2–3 mm of precipitation did not present any problem for the ET estimation because these light rain events cannot produce any measurable groundwater recharge due to the high interception losses characteristic of these forests. In a prolonged (i.e., longer than 5 days) drought period even a 5 mm rain (e.g., May 30, July 25, and September 26 in Fig. 6) will not disturb the present ET estimation method, however, a mere 3 mm of rain can affect it if it takes place not long after a previous larger precipitation event. Large rain events can affect the

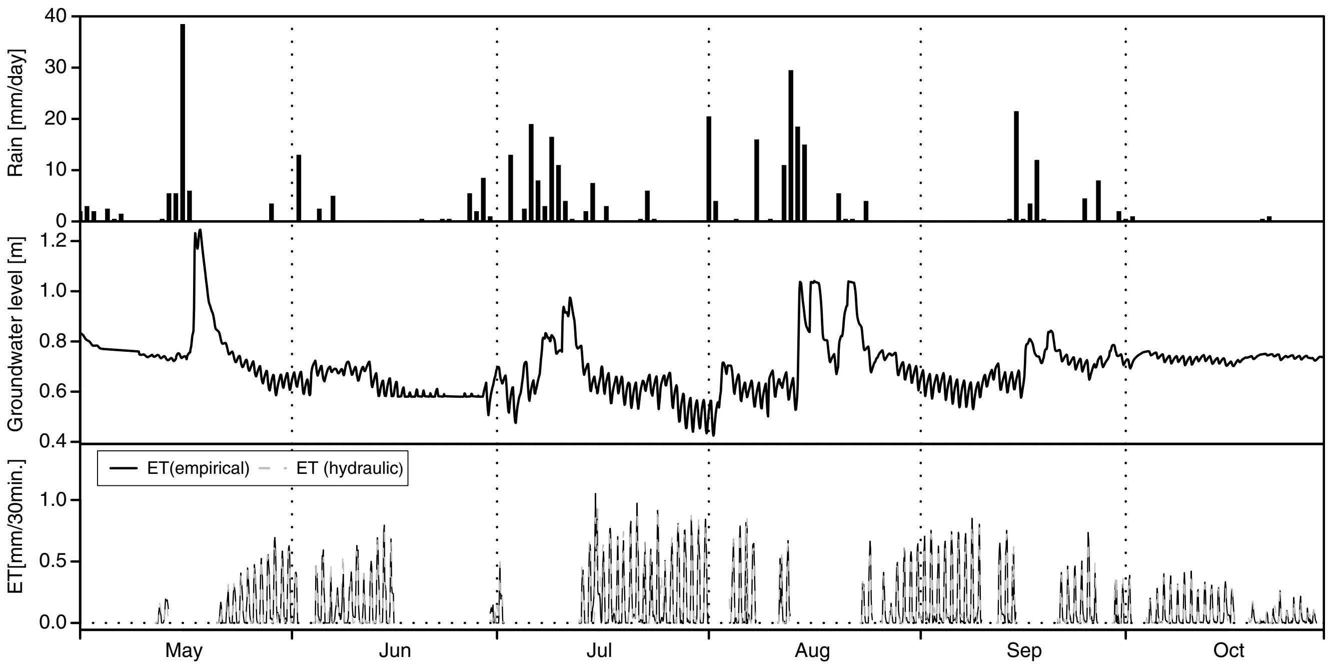

Figure 6 Precipitation, groundwater level measurements and calculated ET rates for the 2005 growing season.

present ET estimation method for up to 2 days, so those periods were excluded from the analysis. It was observed that the *empirical* version of the present method is more sensitive to the disturbing effects of precipitation than the *hydraulic* one, as well as to periods with little diurnal change in the groundwater levels. While the *hydraulic* version functioned well even in these periods, for the sake of comparison between the two versions, such days were excluded from the subsequent analysis. In addition, a two-week period at the end of June was also excluded from the analysis due to instrumentation problems.

Results and discussion

Thirty-minutes ET estimates by the present method are compared with the Penman–Monteith estimates in Fig. 7. The former yields higher ET rates during the nights, as explained above, so for comparison, only the daily values should be considered.

Cross-correlation analysis of the 30-min ET values between the Penman–Monteith and the present method show a peak (r = 0.85–0.93) generally at a separation distance of 60–90 min (larger lag time typical at the start and even at the end of the vegetation period) with the new method's values lagging behind those of the Penman–Monteith approach. Extra long (180 min) lags can be detected in the beginning of May, in the middle and end of October. 60–90 min lags can also be found between the groundwater level and discharge response to increased or decreased ET demand, therefore a difference between the local and overall hydraulic gradients can cause the lag, as it is explained in the accompanying paper by Szilágyi et al. (2007). On the other hand, the lag can be the consequence of the delayed water transport mechanism in the trees because the trunk of the trees can store a relatively high amount of water, and this storage capacity allows some difference between the time of the transpiration and absorption of water from the soil. In consequence, the smaller the water transport (i.e., ET) compared to the stored amount of water in the trees and/or the harder it is to absorb moisture from a drying soil, the larger the lag.

The *hydraulic* version of the new method occasionally yields zero or negative ET values in the evening hours (7–9 p.m.) in summer period when the groundwater level bounces back fast after intensive mid-day ET rates. This occurs because the groundwater flow is described by three control points, out of which only one is located strictly within the riparian zone (the other one is at the bank of the stream, while the third one [the background value] is outside the domain ET exerts its diurnal influence upon). During fast groundwater recovery, following intensive ET rates in the riparian zone the dS/dt term in Eq. (6) is positive but occurs with a negative sign when obtaining the ET value, thus if the groundwater supply rate to the area is underestimated as a result of the too few control points then the resulting ET rate will become negative. This kind of error typically happens at the end of extended drought periods of the summer.

Fig. 8 displays the estimated ET rates on a daily basis, obtained by summing the 30-min values over the days. Both versions of the proposed method yield significantly larger daily rates than the original White approach. These differences (as percentage of the White method's monthly value) from May to October, are 24, 43, 54, 62, 65, 33 for the *hydraulic* and 19, 45, 66, 63, 68, 32 for the *empirical* versions, respectively. The explanation lies in the different assumptions the two methods employ. The original White approach assumes a constant rate of groundwater supply to the riparian zone throughout the day estimated when ET is close to zero in the predawn/dawn hours, when the supply rate is diminished due to a deflated hydraulic gradi-

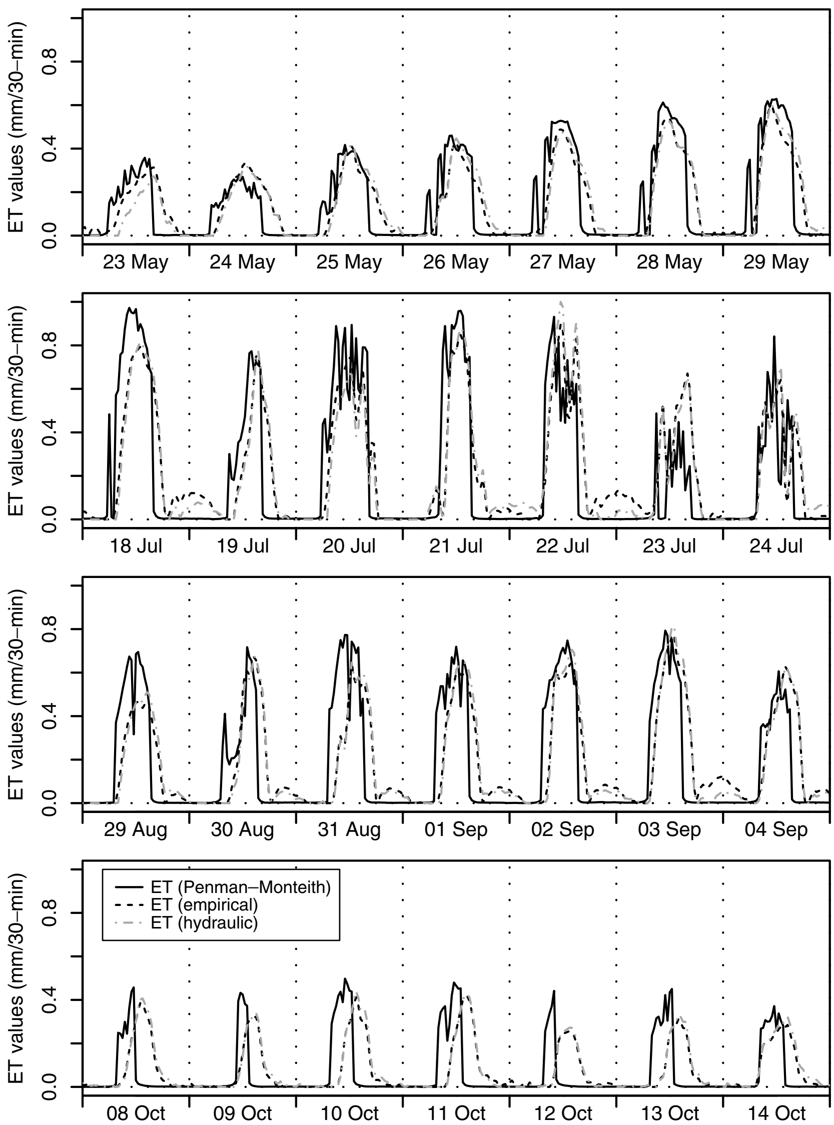

Figure 7 Comparison of the 30-min ET estimates with those of the Penman—Monteith method for some selected growing season periods.

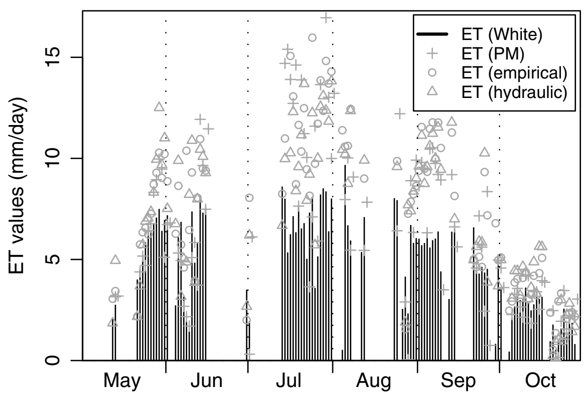

Figure 8 Daily ET rates by the original White approach and Penman—Monteith (PM) method, as well as by the empirical and hydraulic versions of the present ET calculation method for the 2005 growing season.

ent. In contrast, the proposed method accounts for (even if sometime incorrectly) this diurnal change in the hydraulic gradient which has a maximum when ET is most intensive (in fact, a bit later due to the earlier mentioned delay in the groundwater response to changes in ET rates) and minimum in the morning hours.

Loheide et al. (2005) concluded in their numerical study that the White method gave reliable riparian zone ET estimates. However, the ET rates they applied in their model were rather small (1 mm d^{-1}), in fact, almost a magnitude smaller than what was observed in our experimental catchment in Hungary, therefore leading to relatively small diurnal groundwater level fluctuations (<1.5 cm), which thus would not cause a significant change in the overall hydraulic gradients throughout the day. Indeed, with such low ET rates there is hardly any difference in the estimates between the original White approach and our proposed method. However, when the daily ET rate (estimated by the Penman—Monteith method) is about 10 mm a day, as in our experimental watershed, the difference between the two methods is significant.

Finally, Fig. 9 displays the estimated daily mean ET rates by month. Note that these values account predominantly for dry days only. The largest difference between the methods is found in July when ET rates and, thus, groundwater dynamics are most intensive. And conversely, the different methods give the most uniform ET estimates in May and October when the amplitude of the diurnal ET fluctuation is typically small, thus leading to limited diurnal changes in the overall hydraulic gradients within the riparian zone.

Among the two versions of the present ET method, the *hydraulic* approach yields higher ET estimates in the damped diurnal ET amplitude months (May and October), while in the intervening period the *empirical* approach produces the larger ET estimates. The reason lies in the above-mentioned property of the *hydraulic* version of underestimating high groundwater supply rates to the riparian zone. Even with this known error in the *hydraulic* version, it probably produces more reliable ET estimates within the day than the *empirical* one, because (a) the shape of its ET-curve is closer to the shape of the Penman—Monteith derived diurnal ET signal, and; (b) its net inflow values in Fig. 4 reproduce better the time-rate of change in groundwater levels through time, and so are probably closer to reality, than those of the empirical one, which come from a curve fitting of a spline interpolation method, thus are somewhat detached from physical reality between the measurements. The empirical method on the other hand, is more suitable for defining a lower limit (>0) for the ET rate (see its description above), thus it can help in calibrating the hydraulic version.

In comparison of the present ET estimates (3.2—10.5 mm d^{-1}) with other study results, it can be stated that these values typically represent the high-end values of those estimates. For example, Tóth (2007), based on groundwater level readings of piezometer nests, found summer riparian ET rates of 2—12 mm d^{-1} for the same experimental catchment in a very shallow groundwater environment (groundwater depths were between 0.2 and 1.3 m from the surface, therefore the calculated ground water ET fraction is very close to the total ET rate). Around the world, applying different measurement and estimation techniques, Bauer et al. (2004) obtained riparian ET rates of 0.06—4.3 mm d^{-1} for mixed (trees, shrub, and grasses) vegetation of variable density in Botswana, where continuous groundwater level readings of piezometers were used for the estimates (groundwater depths varied between 2 and 3 m from the surface, therefore the calculated groundwater ET fraction may be comparable to vadose zone ET). Butler et al. (2007) obtained ET rates of 2.9—9.3 mm d^{-1} also for mixed vegetation type based on continuous groundwater level readings (groundwater depths were between 0.3 and 3.4 m from the surface, therefore calculated groundwater ET rates were close to total ET when groundwater levels were close to the surface and were smaller than total ET when groundwater levels were deeper). Nachabe et al. (2005) calculated monthly average total ET rates of 1.5—3.5 mm d^{-1} for a pasture and 1.5—6.3 mm d^{-1} for a low-lying forest with the help of continuous soil moisture profile measurements in Florida. Gazal et al. (2006) found 2—7 mm d^{-1} for a semi-arid cottonwood forest; Goodrich et al. (2000) obtained 4—8 mm d^{-1} also for mixed vegetation, while Hughes et al. (2001) found 2—6 mm d^{-1} for a temperate salt marsh in Australia. For the last three experiments sap flow measurements and micrometeorological methods were used for calculating total ET. Unfortunately, important vegetation characteristics (such as LAI) cannot always be deduced from these studies. Notwithstanding, the present riparian ET estimation method seems to yield realistic values, especially, when one considers the ready access of vegetation to the groundwater, the abundance of available energy in the growing season considered, as well as a large value of LAI, all combined with a favorable match with the Penman—Monteith ET values as control.

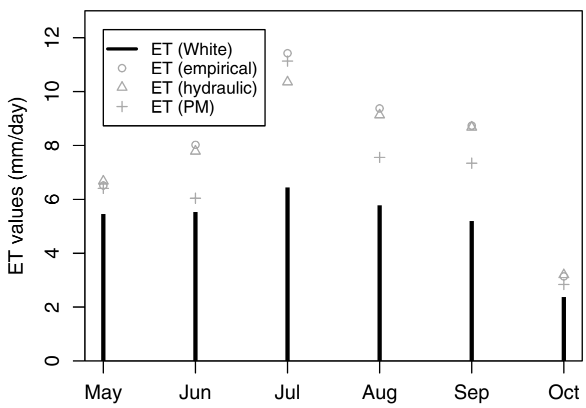

Figure 9 Daily mean ET rates by month of the original White approach and Penman—Monteith (PM) method, as well as of the empirical and hydraulic versions of the present ET estimation method for the 2005 growing season.

Sensitivity analysis of the *hydraulic* method can be summarized (Table 2) as follows. (a) The method is least sensitive to the value of h_0, i.e., the water-level in the stream. Between the two extremes of $h_0 = 0$ m and $h_0 = 0.2$ m (flood level), assuming the reference level is at the channel-bed elevation of the stream, the resulting daily ET values differed only in the third decimals. Thus any effect of diurnal fluctuation in the stream level (which was less than 1 cm in this catchment) to the ET estimate is negligible. (b) The method is only slightly sensitive to the choice of the L parameter. Even an about 5-fold change in its value affected the daily ET estimates by a mere 20%. (c) The method is moderately sensitive to the elevation of the datum. With every 0.1 m lowering of the datum (within the 0—0.5 m interval, which seems realistic in our case), the daily ET estimates increased by only 3—5%. (d) The method is similarly sensitive to the angle formed by the main groundwater flow direction and the stream. Based on simultaneous readings of well-water levels, this angle in the study catchment is about 40° downstream. Accounting for this, mainly by the enlarged value of l, lead to a 5—14% decrease in the daily ET values. (e) Finally, the ET estimates depend most strongly, i.e., linearly (see Eqs. (7a), (7b) and (9) on the values of the specific yield, S_y, and the saturated hydraulic conductivity, k. Thus, the key of reliable ET-rate estimates with the present method lies in the accuracy of the estimated field-scale values of the hydraulic parameters, k and S_y. The observed changes in the ET estimates during the sensitivity analysis, reported in Table 2, occurred more markedly during the period of intensive ET and groundwater dynamics (July—September), and less so in May and October.

In summary the following can be stated. The current ET estimation method is a modified version of the original White method (1932). It considers the growing season diurnal fluctuations of the riparian zone groundwater levels and can fairly well estimate the daily ET rates from high frequency samples (10-min or finer) of the groundwater level in a single well. Interestingly, the sub-daily ET-rate estimates are typically delayed by a few hours in comparison with Penman—Monteith derived ET values. The new method has two versions (i.e., *empirical* and *hydraulic*). The *hydraulic* version requires field-scale values of the saturated hydraulic conductivity (k) of the riparian zone as well as time-varying estimates of the specific yield (S_y) beside the high frequency groundwater level readings and the distance (l) of the groundwater well to the stream along the direction of the main riparian groundwater flowpath. The accuracy of the ET estimates is most sensitive (i.e., linearly) to the effective field-scale value of k and S_y. In the absence of the reliable field-scale value of k and/or when l is unknown, the *empirical* version of the proposed ET estimation method is recommended to be applied which yields ET estimates comparable to those by the hydraulic version.

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