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Numerical Validation of a Diurnal Streamflow-Pattern-Based Evapotranspiration Estimation Method

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Abstract – The evapotranspiration (ET) estimation method by Gribovszki et al. (2010b) has so far been validated only at one catchment because good quality discharge time series with the required high enough temporal resolution can probably be found at only a handful of watersheds worldwide. To fill in the gap of measured data, synthetic groundwater discharge values were produced by a 2D finite element model representing a small catchment. Geometrical and soil physical parameters of the numerical model were changed systematically and it was checked how well the model reproduced the prescribed ET time series. The tests corroborated that the ET-estimation method is applicable for catchments underlain by a shallow aquifer. The slope of the riparian zone has a strong impact on the accuracy of the ET results when the slope is steep, however, the method proved to be reliable for gentle or horizontal riparian zone surfaces, which are more typical in reality. Likewise, errors slightly increase with the decrease of riparian zone width, and unless this width is comparable to the width of the stream (the case of a narrow riparian zone), the ET estimates stay fairly accurate. The steepness of the valley slope had no significant effect on the results but the increase of the stream width (over 4m) strongly influences the ET estimation results, so this method can only be used for small headwater catchments. Finally, even a magnitude change in the prescribed ET rates had only a small effect on the estimation accuracy. The soil physical parameters, however, strongly influence the accuracy of the method. The model-prescribed ET values are recovered exactly only for the sandy-loam aquifer, because only in this case was the model groundwater flow system similar to the assumed, theoretical one. For a low hydraulic conductivity aquifer (e.g. clay, silt), root water uptake creates a considerably depressed water table under the riparian zone, therefore the method underestimates the ET. In a sandy, coarser aquifer the flow lines never become vertical even bellow the root zone, so the method overestimates the ET rate, thus the estimated ET values need to be corrected. Luckily the prescribed and estimated ET rates express a very high linear correlation, so the correction can be obtained by the application of a constant, the value of which solely depends on soil type.

numerical model / baseflow / groundwater/ riparian vegetation

Kivonat – A vízhozamok napi ingadozásán alapuló párolgásbecslési módszer vizsgálata numerikus modellezéssel. A vízfolyásmenti vegetáció (elsősorban erdők) párolgásának becslésére Gribovszki et al. (2010b) kifejlesztett eljárást, amely a vízfolyások vízhozamának napi ingadozásán alapul eddig csak egy vízgyűjtő lefolyási adatain sikerült tesztelni. Sajnos a módszerhez szükséges nagy időbeli felbontású, pontos vízhozammérések csak igen kis számban fellelhetőek és ezek

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hozzáférhetősége is kérdéses. Egy vízgyűjtő hidrológiai körfolyamatának numerikus modellezés azonban lehetőséget nyújt szintetikus vízhozam idősorok előállítására. A vizsgálat során a minta vízfolyás egy átlagos keresztszelvényében 2D numerikus modellezést végeztünk. A modell-paramétereket egy kezdeti értékhez képest szisztematikusan változattuk és vizsgáltuk a beadott párolgási idősor visszanyerhetőségét. Az eredmények alapján a kimunkált módszer a patakmederben vagy nem sokkal alatta elhelyezkedő vízzáró réteg, ún. sekély víztartók esetén alkalmas a becslésre. A vízfolyás menti zóna lényeges (10% fölötti) meredeksége jelentősebb eltéréseket is okozhat a párolgásbecslésben, annak akár közel vízszintes volta azonban csak kis mértékben módosítja az eredményeket. Ugyancsak erősen befolyásolja a módszer eredményeit a vízfolyásmenti zóna keskeny volta, ugyanakkor a jelentősebb szélessége nincs ilyen mértékű hatással a számításra. A vízfolyásmenti zónán kívüli terep meredeksége ugyancsak nem bír meghatározó befolyással a módszerre. A vízfolyásmeder szélességének növelése, mintegy 4m-es szélességnél már erősen befolyásol, így az új metódus csak kisvízfolyások felső szakaszain alkalmazható. Ugyancsak kis befolyással bír a párolgás nagyságrendjének változatása a becslésre. A leginkább befolyásolják a módszer eredményeit a talajfizikai paraméterek. A vízfolyásmenti zóna környezetében kialakuló áramkép és a módszer hipotetikus áramképének megfelelősége a gyökérzónában sandy clay loam fizikai féleségű talajok estében nyújt pontos eredményt. Az alacsony vízvezetőképességű (pl. agyag, iszap) víztartó esetében a vízfolyásmenti zóna alatt egy jelentős depressziós tölcsér alakul ki, a módszer alulbecsli a párolgást. A homokos, durvább szövetű talajoknál pedig a gyökérzónában az áramlási vonalak nem válnak függőlegessé, hanem közel vízszintesek maradnak (mivel a szükséges vízigény szinte mindig kielégíthető az utánpótlódással), így a módszer felülbecsül. Ez utóbbi esetekben nem teljesül a módszer egy-egy alapfeltételezése, tehát korrekcióra szorul. Szerencsére a modellben beadott és az új módszerrel kapott párolgási értékek lineáris korrelációja igen magas, így a korrekció egy talajtípustól függő konstans szorzó bevezetésével megtehető.

numerikus modell / alapvízhozam / talajvíz / vízfolyásmenti vegetáció

1 INTRODUCTION

Small streams and the neighbouring riparian zone groundwater table generally have characteristic diurnal fluctuations in the baseflow period (*Figure 1.* Gribovszki et al. 2010a). Almost all researchers consider evapotranspiration as the primary inducing factor of the growing-season diurnal signal in streamflow and in shallow groundwater level, and some apply the observed diurnal fluctuations for evapotranspiration estimation methods (White 1932; Meyboom 1964; Reigner 1966; Bauer et al. 2004; Nachabe et al. 2005; Gribovszki et al. 2008; Loheide 2008).

A new, baseflow-fluctuations based groundwater ET estimation algorithm is described by Gribovszki et al. (2010b) but validated only at one catchment *(Figure 1)*. The significant advantage of the method over other existing ones is that it requires only very basic geometric properties (i.e., length and width) of the riparian zone and the stream channel width.

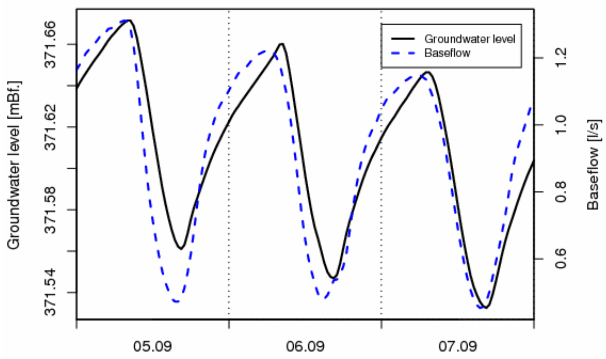

Figure 1. ET-induced diurnal fluctuations in groundwater level and baseflow rates, Hidegvíz Valley experimental catchment near Sopron, Hungary, 2005

2 DERIVATION OF THE BASEFLOW-BASED ET ESTIMATION

The water table within the riparian zone is generally close to the surface and vegetation water uptake during rainless periods comes directly or indirectly from the groundwater. This water use may depress the groundwater table which thus, via an enlarged hydraulic gradient, induces an enhanced seepage from the valley side (and not rarely from the stream as well) toward the riparian zone. *Figure 2* schematically depicts the daily change of the riparian-zone groundwater table and the general groundwater flowpaths in a rainless period of the growing season.

The riparian groundwater hydrograph is a cumulative curve, the result of the dynamic interplay of replenishment as source term and groundwater evapotranspiration as a sink term (Troxell 1936, Gribovszki et al. 2008a). Since in dry periods stream baseflow predominantly originates from the saturated zone, the baseflow hydrograph has a similar shape and characteristics (disregarding of the short temporal shift observable in *Figure 1*) as the groundwater level curve.

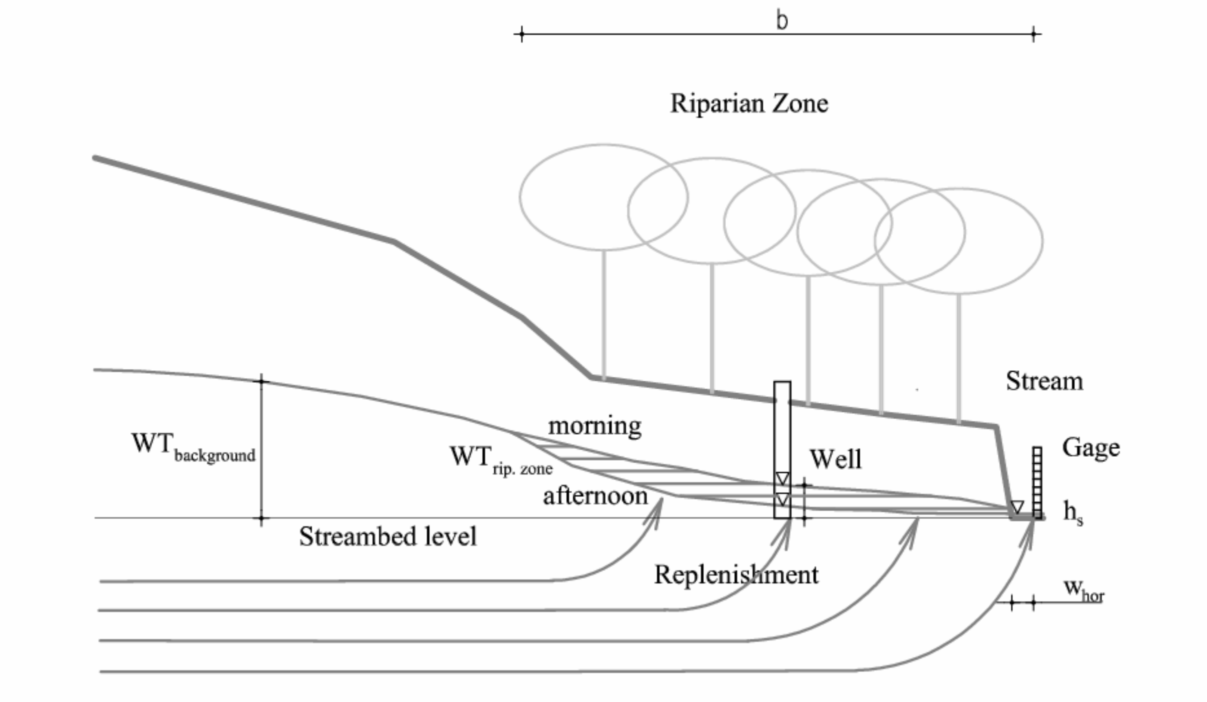

Figure 2. Schematic model of the water table (WT) diurnal change in the riparian zone

The riparian-zone groundwater ET estimation method of Gribovszki et al. (2008a), based on the diurnal fluctuations of the groundwater levels, can be reformulated for the baseflow signal through the employment of a linear transformation, similar to Loheide (2008). The physical interpretation of the linear transformation is that of a linear reservoir for the riparian zone.

Accordingly, the streamflow-based ET-estimation method presented by Gribovszki et al. (2010b) employs the linear reservoir and water balance equations (written for the saturated zone), the latter as

$$\frac{\partial S_r}{\partial t} = S_y(t, WT)\frac{\partial WT}{\partial t} A_{rip} = Q_i - Q_o - ET_{gw} \cdot A_{rip} = Q_{net} - ET_{gw} \cdot A_{rip} \qquad (1a)$$

where dS_r / dt [L^3T^{-1}] is the time-rate of change in riparian groundwater storage (S_r), WT [L] the average groundwater level (above reference) in the riparian zone, S_y the specific yield, Q_i, the incoming discharge [L^3T^{-1}] to the riparian zone, and Q_o, the outgoing discharge from the riparian zone to the stream [L^3T^{-1}]. The net supply/replenishment rate is the difference of the incoming and outgoing discharges to and from the riparian zone, $Q_{net} = Q_i - Q_o$, [L^3T^{-1}]. ET_{gw}, is evapotranspiration (directly or indirectly) from the groundwater, $A_{rip} = l \cdot 2b$ [L^2] is the area of the riparian zone, with b denoting the average half-width of it, and l [L], the length of the stream valley (and not the stream), where the riparian vegetation (phreatophytes) is located.

In order to obtain the net supply rate (Q_{net}), let's write Eq. 1a for the late night/early morning hours when ET_{gw} is negligible

$$\frac{\partial S_r}{\partial t} = Q_i - Q_o = Q_{net} \tag{1b}$$

The linear storage equation for Q_0 is

$$Q_o = \frac{1}{T^*} S_r \tag{2}$$

where, T^* [T], is the average residence time of water within the riparian zone.

Eqs. (1b) and (2) yield

$$T^* \frac{\partial Q_o}{\partial t} = Q_i - Q_o \tag{3}$$

Q_o results from stream discharge measurements, while Q_i is obtained from Darcy's law (Gribovszki et al., 2010b), assuming a continuous flow system for the watershed (Tóth, 1963). Due to groundwater evapotranspiration, groundwater streamlines intersect the riparian zone, as illustrated in Fig. 2. Obviously, the density of the streamlines that intersect the riparian zone varies with the actual water demand of the riparian vegetation forming a considerable upward hydraulic gradient during rainless/drought periods of the growing season resulting in near-vertical streamlines directly below the root-zone.

Instead of using the real physical parameters (Gribovszki et al. 2010b), Eq. 3 can be written in a form that only a theoretical relationship (the most simple is linear) is assumed between the water balance components, e.g. $Q_i = f(Q_o)$, yielding regression equations. In this case one does not need a preliminary knowledge of the flow system, because the value of the resulting regression parameter indicates its type. Unfortunately, ET cannot be estimated directly in this way, however the reliability of a physically based method (Gribovszki et al. 2010b) under different conditions can still be tested. The assumed linear relationship between in and outflows, for a unit width of the riparian zone can thus be written as

$$q_i \approx a_1 + m_1 \cdot q_0 \tag{4}$$

where $q_i = Q_i/(2 \cdot l)$ and $q_o = Q_o/(2 \cdot l)$.

In this way Eq. 3 can be written in finite difference form as

$$T^* \frac{\Delta q_o}{\Delta t} = a_1 + m_1 \cdot q_0 - q_o = a_1 - (1 - m_1) \cdot q \tag{5}$$

Rearrangement for q_o yields

$$q_o = \frac{a_1}{1-m_1} - \frac{T^*}{1-m_1}\frac{\Delta q_o}{\Delta t} \tag{6}$$

Eq. 6 is of a linear (i.e., $y=a+m\cdot x$) form, where $y = q_0$, $x = \Delta q_0 / \Delta t$, $m = T^* \cdot (1-m_1)^{-1}$, and $a = a_1 \cdot (1-m_1)^{-1}$. The constant parameters (a and m) of Eq. 6 can be obtained from streamflow measurements in the late night/early dawn period of the day by fitting straight lines to the $\Delta q_0 / \Delta t$ and q_0 data pairs (Gribovszki 2010b). Fig. 3 illustrates the correlation of these data pairs (as a result of the numerical model).

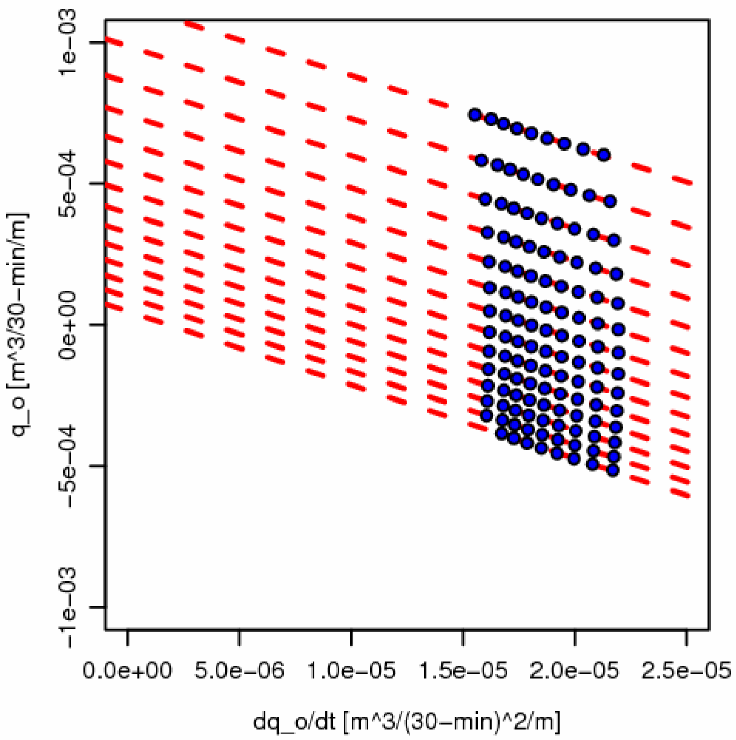

Figure 3. Numerical model data for a loamy aquifer. The dq_0 / dt vs q_0 values (dots) with the fitted first-order polynomials (one line per day) from the late-night hours

Now apply Eqs. 1a, 1b and 2 for the estimation of ET_{gw} ($ET_{gw} \cdot 2b = ET_{gw} \cdot A_{rip} / l$), which, using Eqs. 5 and 6, can be transformed into

$$ET_{gw} \cdot 2b = q_{net} - T^* \frac{\partial q_o}{\partial t} \approx a_1 - q_o(1-m_1) - T^* \frac{\Delta q_o}{\Delta t} \tag{7}$$

where $q_{net} = q_i - q_o$.

Figure 4 demonstrates the new ET estimation method applied with the synthetic outflow values of the numerical model.

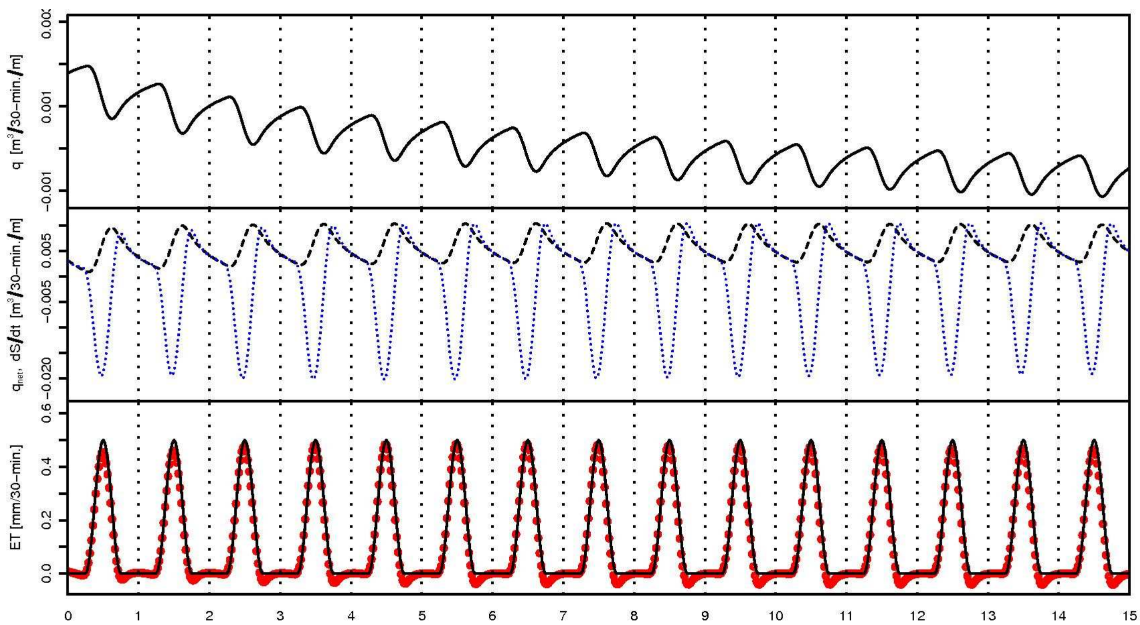

Figure 4. Components of the groundwater evapotranspiration estimation method based on diurnal fluctuations of streamflow (numerical model result without model spin up days in case of loamy aquifer). Top panel: q, measured streamflow; Middle panel: q_{net}, estimated net groundwater supply (broken line); dS_r / dt, estimated groundwater storage change of the riparian zone (punktuated). Bottom panel: ET_{gw}, the ET rates estimated by the new method (solid line); ET_s, prescribed model ET rates (points)

The $-m_1$ parameter value is the same as *b/(2w)* in Gribovszki et al. (2010b). So one can check if the $-m_1$ values calculated from the late night regression of the numerical discharge are equal or not with the *b/(2w)* constant values. If they are, one obtains the same ET rates as the prescribed ET time series values. If not, the model does not follow the pre-supposed flow system, therefore the method under the corresponding boundary conditions cannot be applied directly, only with some further modification.

The relationship between the replenishing and the outgoing groundwater discharge may be not linear. This nonlinearity can only be tested on a longer time series, so a 20-day interval was chosen as the time span of modelling. A 20-day time-period is long enough to exceed the model spin-up period significantly. So after omitting these spin up days, one is still left with a sufficient number of values to perform statistical analyses if needed.

3 THE NUMERICAL MODEL SETUP

To test the validity of the new ET estimation method, an adaptive, finite element 2D numerical model was employed for integrating the extended Richards equation (Lam et al., 1987, Szilágyi et al. 2008) vertically

$$\frac{\partial}{\partial x}\left(K(\Psi)\cdot\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K(\Psi)\cdot\frac{\partial h}{\partial y}\right)+s=m\cdot\chi\cdot\frac{\partial h}{\partial t} \tag{8}$$

where K [LT^{-1}] is the hydraulic conductivity (a function of the pressure head, Ψ); h [L] is the hydraulic head; m is the slope of the water retention curve which becomes the coefficient of volume change in the saturated zone; s is the sink term which represents ET [LT^{-1}] in this model; γ is the unit weight of water; and *x*, *y* [L] and t [T] are the horizontal, vertical and temporal coordinates.

First, this numerical model with geometrical and physical model parameters similar to that of Vadkan-valley (the test catchment in Gribovszki et al., 2010b) was used for generating an outflow flux (discharge) series. The numerical outflow values were then used to "back-calculate" the prescribed ET rates in the model.

Figure 5 describes the geometry of the model setup. Each simulation lasted for 20 days with a 30-minute time-step between outputs. The initial condition of the model was a horizontal groundwater level (1.5 m above the streambed) at t=0, equalling the water level in the adjacent stream. Stream-level then was dropped to a fixed elevation (i.e. 0.05 m) for t>0 to induce drawdown of the aquifer. The initial aquifer geometry is similar to Vadkan-valley, for which Gribovszki et al. (2010b) have already employed measured streamflow values for validation of the present method. The riparian aquifer width (*b*) is 20 m, the ground surface (*mf_rip*) has a gentle (1:20) slope, the impermeable layer is horizontal and found at *d* meter (*d* = 1 m) below the streambed. Prescribed horizontal extent (*wr*) of the watershed beyond the riparian zone is 50 m, with a valley-slope (*mf*) of 0.3. The *wr* value for the test catchment is about four-times greater (i.e., ~200 m), but in order to reduce the model run-time it was changed to 50 m. Sensitivity of the method to *wr* has been tested and was found that *wr* = 200 m gave practically the same ET estimates as *wr* = 50 m. Basic aquifer geometric parameters are listed in *Table 1*.

Table 1. The basic parameter-set of the model

b	=	20 m	{one side (half) width of the rip zone }
w_{hor}	=	1 m	{half of the stream width}
wr	=	50 m	{watershed boundary distance from rip. zone side}
d	=	1 m	{depth to the impermeable layer from stream bottom}
mf	=	0.3	{surface slope outside the rip. zone}
mf_rip	=	0.05	{surface slope of the rip. zone}
gyz	=	1 m	{depth of the root zone}
m	=	0	{slope of the impermeable layer}
h_0	=	1.5 m	{starting stream water level at t=0}
h_s	=	0.05 m	{stream water level at t>0}

Aquifer drainage rate was obtained by integrating the horizontal component of the Darcy-flux vectors at $x = b$ along the vertical stream bank and the vertical component of the Darcy-flux vectors where $b < x < (b+w)$ along the horizontal stream bed (w_{hor}). In order to obtain a correct value for w, the extent of the vertical seepage face with the corresponding seepage rate had to be considered and it had to be added to the half stream width (w_{hor}). The active vertical extent of the seepage face was determined by the ratio of the horizontal to vertical seepage rates and on average gave an additional 0.5 m (standard deviation ±0.10 m) to the horizontal half stream width.

The $b/2w$ values are listed in Tables 3 and 4 (with corresponding errors relative to the value of $-m_1$), calculated with both a constant 0.5-m long vertical seepage face and also with a seepage face whose length has been estimated by weighting the vertical and horizontal fluxes.

Figure 5 illustrates the setup of the model domain.

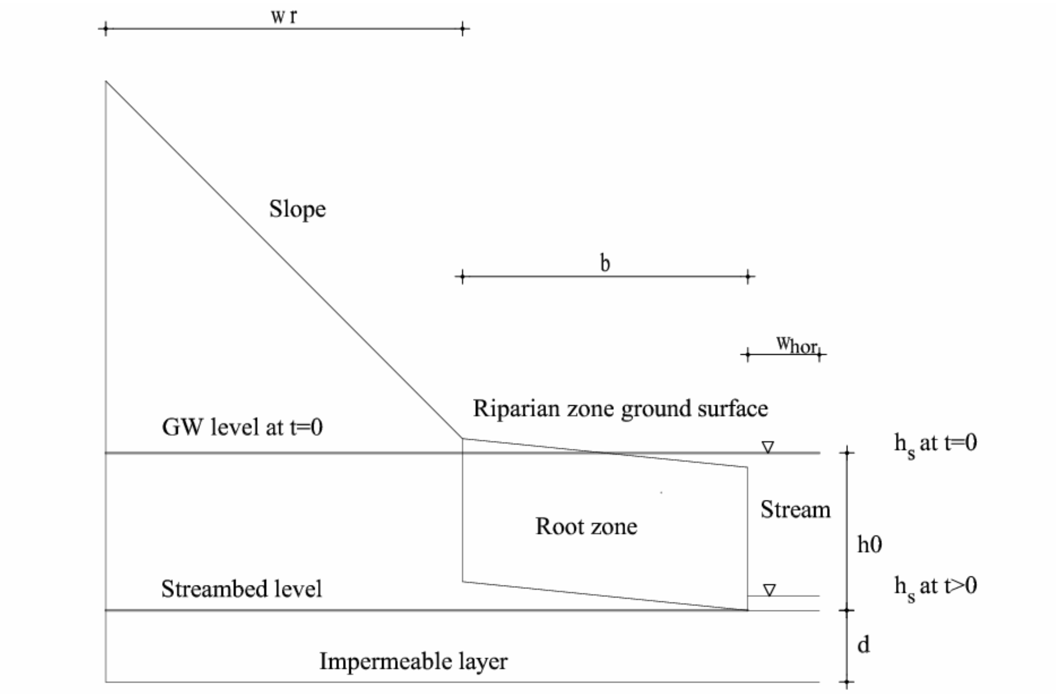

Figure 5. Schematic vertical cross-section of the aquifer employed in the numerical model. The left-side of the aquifer represents no-flow conditions (wr distance from the side of the riparian zone), and the stream (h_s is stream stage) is on the right-side of the riparian aquifer. The x_0, y_0 ($x=0$, $y=0$) coordinates of the model is set where the vertical riparian-zone boundary reaches the streambed level.

Prescribed aquifer material properties are listed in *Table 2.*

Table 2. Hydraulic properties of the model aquifer (after Campbell, 1974; as well as Clapp and Hornberger, 1978)

Hydraulic property	Value/equation	Explanation
φ [–]	Sand=0.395, Loamy sand=0.410, Sandy loam=0.435, Silt loam=0.485, Loam=0.451, Sandy clay loam=0.42, Silty clay loam=0.477, Clay loam=0.476, Sandy clay=0.426, Silty clay=0.492, Clay=0.482	φ – total porosity
K_s [cm/min]	Sand=15.21, Loamy sand=13.51, Sandy loam=3.00, Silt loam=0.622, Loam=0.600, Sandy clay loam=0.544, Silty clay loam=0.147, Clay loam=0.212, Sandy clay=0.187, Silty clay=0.089, Clay=0.111	*Ks* – saturated hydraulic conductivity
Ψ_{ae} [kPa]	Sand=1.21, Loamy sand=0.90, Sandy loam=2.18, Silt loam=7.86, Loam=4.78, Sandy clay loam=2.99, Silty clay loam=3.56, Clay loam=6.30, Sandy clay=1.53, Silty clay=4.90, Clay=4.05	Ψ_{ae} – air-entry pressure
$\Psi(\Theta)$	$\lvert\Psi_{ae}\rvert\cdot(\varphi/\Theta)^b$	Θ – volumetric water content [–]
$K(\Theta)$	$K_s(\Theta/\varphi)^{2b+3}$ *b*: Sand=4.05, Loamy sand=4.38, Sandy loam=4.90, Silt loam=5.30, Loam=5.39, Sandy clay loam=7.12, Silty clay loam=7.75, Clay loam=8.52, Sandy clay=10.4, Silty clay=10.4, Clay=11.4	*b* – pore size distribution index

The 1-m thick root zone considered in the model starts at the surface and lies parallel with the sloping ground surface. The water consumption of riparian vegetation is represented by the term *s* (a sink for groundwater and soil moisture, respectively, depending whether the root zone is saturated or not) in (8), restricted to the root zone only. Diurnal water use fluctuation is described by (Szilágyi et al., 2008)

$$s = -c \cdot \sin^2\left(\frac{\pi \cdot t}{12}\right), ...0 \leq \text{mod}\left(\frac{t}{24}\right) < 12 \tag{9a}$$

$$s = 0, ...12 \leq \text{mod}\left(\frac{t}{24}\right) < 24 \tag{9b}$$

where *c* (=2.4 10^{-3}) is a constant and *mod* is the modulus of division with time measured in hours. Eqs. 9a and 9b ensure that for one-half of the day water uptake by the vegetation from the root zone is zero, and follows a sine-like curve for the other half, taking only positive values.

The basic prescribed value (used in most model validation scenario) of *c* yields a daily mean ET of 6 mm, which is close to the growing season daily mean ET value reported by Gribovszki et al. (2010b). The fixed value of the stream level for $t > 0$ in the model allowed for possible induced recharge during the drawdown.

4 RESULTS

The effect of geometry and ET magnitude

First, sensitivity of the new method to the geometry of the riparian zone and to the magnitude of the prescribed evapotranspiration rate for a loamy aquifer (similar to that of Vadkan Valley) was tested. Table 3 displays the results of the sensitivity test. The first column lists the parameters whose values were modified while keeping the rest of the parameters unchanged and listed in Table 1.

Table 3. ET-estimation sensitivity to geometry and prescribed evapotranspiration magnitude

Parameter	$b/2w/{-m_1}$ if *w* used as a constant	$b/2w/{-m_1}$ if *w* calculated from num. model	$-m_1$	R (Corr. coef.)	Error (%) (*w* as a constant)	Error (%) (*w* as a result of num. model)
w_hor=1 m, *d*=1 m	1.31	1.31	5.08	0.991	–	–
w_hor=0 m, *d*=0 m	1.14	1.14	10.95	0.979	–13.0	–13.0
w_hor=2 m , *d*=1 m	0.78	0.66	5.12	0.991	–40.5	–49.4
w_hor=1, *d*=2	1.69	1.93	3.94	0.997	28.9	47.3
mf=0.01	1.22	1.17	5.47	0.992	–7.1	–10.7
mf=0.1	1.27	1.13	5.27	0.991	–3.6	–13.7
mf=1	1.36	1.46	4.89	0.99	3.9	11.5
mf_rip=0.025	1.45	1.35	4.61	0.989	10.2	3.1
mf_rip=0.1	1.10	0.85	6.04	0.992	–15.9	–35.1
b=40 m	1.02	1.00	13.06	0.993	–22.2	–23.8
b=10 m	2.30	2.44	1.45	0.991	75.2	86.3
ET=2 mm	1.48	1.50	4.49	0.992	13.1	14.5
ET=10 mm	1.11	1.13	5.99	0.99	–15.2	–13.7

Error means the estimated ET value’s deviation from the prescribed ET rate.

The following conclusions can be drawn from *Table 3*.

1) The increase (over 4m) of the horizontal streambed width ($2{\cdot}w_{hor}$) strongly influences the ET estimation. The new method is stable in the range of a 0–2 m width, so it can only be used reliably for small headwater catchments.

2) If the riparian aquifer is deeper than 2 m below the streambed, considerable (30–50%) errors can be detected, so the method can only be applied for fully or near-fully incised streams.

3a) The steepness of background watershed surface (*mf*) influences the evaporation estimate only slightly. Even a slope of 100% does not lead to a significant error in the ET estimates.

3b) Changing the slope of the riparian zone surface (*mf_rip*) does not cause any significant change, but if the slope is over 10%, the error becomes significant. A flat valley bottom assumption, however, is relevant in most natural cases.

4) Even a two-fold increase in the riparian zone width (from 20 to 40 m) does not induce significant errors in the results of the method, but on the other hand its decrease (narrowing) influences the evaporation estimate more significantly. The errors become substantial only under a width of 10 m, but this situation is rare even on headwater catchments. It is so because a narrow vegetation stripe cannot induce significant upward gradients in a considerably wide zone, as required by the method.

5) Applying an ET range of 2 to 10 mm/d on rainless days of the growing season as lower and upper envelope values, the method gives +13–15% and –14–15% errors, respectively. Small ET rates were slightly overestimated, while large ET values were slightly underestimated by the method.

The effect of soil physical parameters

The prescribed ET values are recovered only for the sandy-loam aquifer, because it corresponds best to the theoretical groundwater flow system *(Table 4)*. Only in this case will the *b/(2w)* ratio be about equal with the $-m_1$ constant in Eqs. 4–7.

Table 4. Results of the soil parameter sensitivity tests

1	Soil texture	*k* (m/d)	*b/2w/–m1* if *w* used as a constant (1.5m)	*b/2w/–m1* if *w* calculated from num. model	*–m1*	*R* (Corr. coef.)	Error (%) (*w*=1.5m as a constant)	Error (%) (*w* as a result of num. model)
2	Sand	15.21	2.89	3.09	2.31	0.976	188.6	209.0
3	Loamy sand	13.51	2.04	2.26	3.27	0.943	103.9	126.0
4	Sandy loam	3.00	2.26	2.36	2.95	0.984	126.0	136.0
5	Silt loam	0.62	2.19	1.97	3.04	0.968	119.3	97.0
6	Loam	0.60	1.31	1.31	5.08	0.991	31.2	31.0
7	Sandy clay loam	0.54	1.02	0.96	6.54	0.999	1.9	–4.0
8	Silty clay loam	0.15	0.55	0.52	12.2	0.999	–45.4	–48.0
9	Clay loam	0.21	0.81	0.81	8.25	0.998	–19.2	–19.0
10	Sandy clay	0.19	0.38	0.41	17.55	0.993	–62.0	–59.0
11	Silty clay	0.09	0.55	0.81	12.14	0.997	–45.1	–19.0
12	Clay	0.11	0.60	0.576	11.12	0.996	–40.0	–42.4

Error means the estimated ET value’s deviation from the prescribed ET rate.

For low hydraulic conductivity aquifers (e.g. clay, silt), root water uptake induces a considerably depressed water table under the riparian zone. The streamlines will be perpendicular to the surface of the depression cone and not to the bottom of the root zone. In this case $b/2w < -m_1$, so the method underestimates the prescribed ET rate. Interestingly, the error of the estimation as a function of the physical soil type is not linear, but roughly oscillates around an average value of –50%.

For coarse aquifer type (loam, loamy sand, sand) the flow lines do not become vertical below the root zone because the decreasing day-time storage can be readily replaced via the high conductivities. Thus the water demand of the vegetation is almost immediately met via the main horizontal gradients. In this case $b/2w > -m_1$ so the method overestimates the ET rate. It is also remarkable that the error does not change linearly with the change of the hydraulic conductivity, but rather (except for sand and loam types), it oscillates around +100–130%.

In summary, it can be stated that a strong linear relationship exists in all model settings between the prescribed and estimated ET values (the model spin up period excluded). The correction-factor values valid for the linear relationships changed only minutely during the 20-day modelling period (Fig. 4).

Since the model results depend strongly on the soil physical characteristics, the method requires a correction. Fortunately, the correlation between prescribed and estimated ET rates is very high (generally R = 0.98–1.00) and linear, accordingly the required correction by introducing a constant multiplier as a function of the soil type, can be achieved and remains the task of the future to be tested on natural catchments.

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