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Comparison and modification of methods for estimating evapotranspiration using diurnal groundwater level fluctuations in arid and semiarid regions

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SUMMARY

In arid and semiarid regions, vegetation growth largely depends on groundwater, and causes diurnal fluctuations of shallow groundwater levels. Diurnal groundwater level fluctuations have been widely used to estimate groundwater evapotranspiration (ET_G) in several methods. This study compared ET_G estimated by three commonly used methods. A groundwater flow model was created to generate synthetic diurnal groundwater level fluctuations caused by a given evapotranspiration. The model also calculates the change in groundwater storage and net groundwater inflow at locations of observation wells. The White method, the Hays method, and the Loheide method were applied to estimate ET_G with the model-generated diurnal groundwater levels. The comparison of the actual and estimated ET_G revealed the accuracy of each method and indentified the applicability of the methods. When the recovery limb of the groundwater level hydrograph is nonlinear, these existing methods underestimate daily ET_G. The Loheide method is comparatively better and can be improved by representing the rate of water table increase in the recovery limb of the hydrograph using an exponential equation. When the recovery limb of the groundwater level hydrograph is linear, all three methods can accurately estimate the daily ET_G. The modified White method can provide hourly ET_G estimates and is recommended for general use. In practical applications, the analysis of the shape of the water table recovery limb and the up and down gradient groundwater head differences can be used to identify the proper method for estimating ET_G.

1. Introduction

Arid and semiarid regions occupy approximately 30% of the land surface of the Earth (Dregne, 1991), including the majority of northern and southern Africa, the Middle East, western USA and the southern South America, most of Australia, large parts of central Asia, and parts of Europe (NOAA, 2010). Vegetation provides a natural protection against desertification and dust storms in these regions. Some vegetation is known as phreatophyte and is groundwater dependent (Bulter et al., 2007). Phreatophyte transpiration consumes groundwater and causes diurnal fluctuations of groundwater levels (Gribovszki et al., 2010). On the other hand, surface water is scarce, and groundwater is often the only reliable water resource for the social-economic development in arid regions (Scanlon et al., 2006). Irrigation water for crops is usually provided by the abstraction of groundwater. The over-exploitation of groundwater resources has caused decreasing groundwater levels and resulted in desertification in many parts of arid regions (Qi and Luo, 2006). The sustainable management of groundwater resources must consider both the water use by human activities and by nature. The starting point to develop a sustainable groundwater use plan is the assessment of groundwater balance. In arid environments, an important component of the groundwater balance is groundwater evapotranspiration (ET_G). For example, in the Ordos Plateau in northern China, ET_G accounts for over 60% of the natural groundwater discharge and other forms of discharge are baseflow (30%) and extraction (10%) (Yin et al., 2011). The shallower the water table and the drier the climate, the larger the ET_G and its contribution to groundwater discharge (Lubczynski, 2009).

ET_G is difficult to quantify directly due to its spatial and temporal variability (Mould et al., 2010). In the commonly used groundwater flow model, MODFLOW (McDonald and Harbaugh, 1988), ET_G is defined as a linear function of the water table depth. ET_G reaches the maximum when the water table is near the surface (Banta, 2000). It is considered to be zero when the water table is

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below a fixed depth, termed extinction depth. In real world, ET_G is driven by incoming solar radiation (Oke, 1978), therefore, has a strong diurnal signal caused by evapotranspiration of phreatophytic vegetation. In response to water loss, groundwater level fluctuates diurnally. The diurnal ET_G signal has also been observed in stream flows (Bulter et al., 2007; Chen, 2007; Gribovszki et al., 2010). Groundwater level fluctuations have long been recognized as valuable information for inferring daily groundwater evapotranspiration (Blaney et al., 1930; White, 1932; Troxell, 1936; Wicht, 1941). There are several practical advantages of using diurnal groundwater level fluctuations to estimate ET_G. With the application of high precision pressure transducers, diurnal groundwater level fluctuations can be recorded automatically. Monitoring of diurnal groundwater level fluctuations needs less intensive fieldwork and is much cheaper than monitoring of evaporation with pans or lysimeters (Lautz, 2008). There is a growing number of applications using diurnal groundwater level fluctuations to estimate ET_G (Gribovszki et al., 2010).

White (1932) first proposed the method of estimating ET_G using diurnal groundwater level fluctuations. The original White method has been modified and newly improved methods were developed, such as Hays (2003) and Loheide et al. (2005). These are the three commonly used methods (Lautz, 2008), there are, however, no systematic analysis and comparisons of the methods for estimating ET_G. The three methods use different algorithms to calculate groundwater inflow, and their estimated ET_G rates are different for a given hydrograph. No attempts have been made to compare the accuracy of these different estimates The objectives of this study are: (1) to compare the accuracy of the three methods for estimating ET_G, (2) to improve the accuracy of ET_G estimation methods and (3) to give the guidelines for proper method selection.

2. Methods of estimation and comparison

2.1. Methods of estimation

White (1932) proposed a method of estimating ET_G from groundwater level hydrograph using the following equation:

$$ET_G = (24r \pm \Delta s) \times S_y \tag{1}$$

where S_y is the specific yield (1), r is the rate of water table rise between 00:00 and 04:00 (m h^{-1}) as shown in Fig. 1 and Δs is the net rise or fall of groundwater level during a 24-h period (m d^{-1}). The assumptions of the White method are: (1) evapotranspiration by plants causes diurnal water table fluctuations, (2) evapotranspiration is negligible relative to the groundwater inflow between 00:00 and 04:00, (3) the rate of groundwater inflow to the site is constant throughout the day (Loheide et al., 2005) and (4) specific

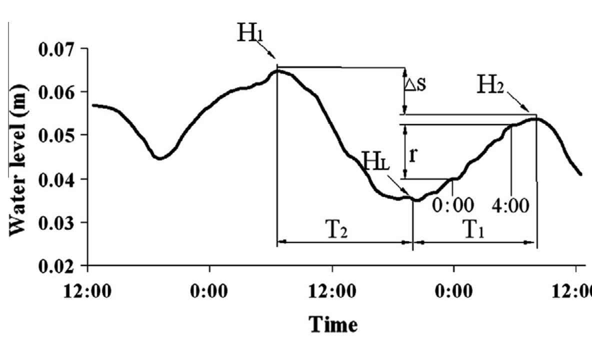

Fig. 1. Examples of diurnal groundwater level fluctuations, with variables used in Eqs. (1) and (2).

yield is constant. Most natural systems do not satisfy the third assumption, particularly in riparian zones where the White method is widely used. When ET_G occurs in riparian zones, groundwater flow direction has been observed to change (Rosenberry and Winter, 1997; Smerdon et al., 2005). As a result, groundwater inflow rate will increase due to the increasing pressure head difference between the constant head (rivers or lakes) and the water table where plants are located. Gribovszki et al. (2008) assessed the accuracy of the White method when the rate of groundwater inflow varies.

There are two main uncertainties associated with the use of the White method, i.e. specific yield of aquifers and the groundwater inflow. Improvements to the White method have been made to reduce uncertainties. Regarding specific yield, researchers suggested that the readily available specific yield should be used instead of the conventionally defined specific yield (Healy and Cook, 2002; Lautz, 2008; Loheide, 2008). The readily available specific yield is the amount of water that is released from the vadose zone during the period of the diurnal fluctuations. Loheide et al. (2005) proposed an equation for estimating the specific yield as a function of sediment texture, depth to water table, and elapsed time of drainage.

Calculation of groundwater inflow is another uncertainty. Due to nonlinearities associated with unsaturated flow, three-dimensional groundwater flow patterns, and transient effects, a mathematical form for calculating the groundwater inflow is difficult to obtain (Loheide, 2008). The groundwater inflow is usually determined from the rate of change in the water table during the night when evapotranspiration is assumed negligible. Different algorithms have been developed to calculate the groundwater inflow rate. In the White method, the groundwater inflow is calculated by $(24 \times r \times S_y)$ and r is the slope of the best fitted line to the hydrograph from 0:00 to 4:00 h.

Hays (2003) developed a new method to estimate ET_G, including a flexible time component for the ET_G period:

$$ET_G = \left[(H_1 - H_L) + \frac{H_2 - H_L}{T_1} T_2\right] \times S_y \tag{2}$$

As illustrated in Fig. 1, H_1 is the first peak of groundwater level in the morning (m), H_2 is the peak of the following day (m), H_L is the lowest groundwater level of the target day (m), T_2 is the hours of the water table decrease period, and T_1 is the hours of the water table rising period. The key assumptions of the Hays method are similar to the White method, except that it assumes ET_G occurs only during the water table decrease period. In the Hays method, groundwater inflow is calculated using the second term in Eq. (2).

Loheide (2008) modified the White method and estimated hourly evapotranspiration values by considering the influence of continuous groundwater flow in and out of the area of water table fluctuation. Loheide (2008) assumed that the rate of change in water table outside the area of water table fluctuation equals the overall rate of water table change at the observation location. The Loheide method first removes the trend from the groundwater level time series, $WT(t)$, using the following equation:

$$WT_{DT(t)} = WT_{(t)} - m_T \times t - b_T \tag{3}$$

where m_T is the trend slope (m t^{-1}), t is time (t), b_T is the intercept (m), and $WT_{DT(t)}$ is the detrended water table depth (m). $\Gamma[WT_{DT}]$ is defined as dWT_{DT}/dt, and can be estimated using water table data in the recovery period. Then the groundwater inflow rate, $r(t)$, can be estimated using:

$$r(t) = S_y(\Gamma(WT_{DT(t)}) + m_T) \tag{4}$$

ET_G can then be calculated using the following equation:

$$ET_{G(t)} = r(t) - S_y \frac{d(WT_{(t)})}{dt} \tag{5}$$

2.2. Methods of comparison

In this study, a groundwater flow model was used to generate diurnal groundwater level fluctuations by a given evapotranspiration flux. The White method, the Hays method and the Loheide method were applied to the model-generated diurnal groundwater level fluctuations to estimate ET_G. The estimated ET_G from the three methods were then compared with the actual ET_G of the groundwater flow model.

The groundwater flow model consists of a phreatophyte-dominated floodplain and an upland region as shown in Fig. 2, and is representative of small watersheds in arid regions. The width of the riparian zone is 100 m, the hillslope is 880 m long, and a river is 20 m wide located at the right boundary (Fig. 2). The elevation of the land surface decreases linearly from 55 m on the left water divide boundary to 46 m in the river valley. The elevation of the bottom of the aquifer is chosen as 0 m. The left and lower boundaries are no-flow boundaries and the river is simulated with a river package. The parameters of the river are: riverbed conductance: 400 $m^2\,d^{-1}$; river stage: 45 m; elevation of the riverbed bottom: 44 m. Horizontal hydraulic conductivity of the aquifer is 1.0 m h^{-1} and specific yield 0.25, corresponding to the properties of a medium-sand aquifer. Near the surface and the aquifer consists of fine-grained soils, specific yield could vary as the water table fluctuates (Duke, 1972; Loheide et al., 2005). In this study, the depth to water table is larger than 1 m and the aquifer material is coarse soil. A constant specific yield is assumed. The initial groundwater level linearly decreases from 50 m on the left boundary to 45 m at the river on the right. The model was run for a period of 480 h to obtain the initial condition for the transient simulation (Fig. 2).

Diurnal fluctuations of groundwater levels were simulated using MODFLOW-2000 (Harbaugh et al., 2000). The finite difference grid consisted of 25 rows, 50 columns and 1 layer, with a regular grid size of 20 × 20 m. Only a drying period was included in the study, as the White method is not applicable in wetting periods when water table rises (Loheide et al., 2005). Therefore, only ET_G was considered in the model. The maximum ET_G rate near the surface varies from 6:00 to 18:00 as shown in Fig. 3, corresponding to a maximum of 2 mm/h at noon and zero during the night. The magnitude of the maximum ET_G rate is comparable to some studies (Schilling, 2007; Lautz, 2008). The pattern of ET_G used in this study (Fig. 3) is probably the most common one according to the sap flow measurements of many species of trees or estimated ET_G pattern (Engel et al., 2005; Oguntunde, 2005; Bulter et al., 2007; Schilling, 2007; Gribovszki et al., 2008; Loheide, 2008). The extinction depth of ET_G was set to be 4 m, considering the presence of deep-rooted phreatophytic vegetation in arid environments (Naumburg et al., 2005). In the numerical model, ET package was used and the actual values of ET_G (mm h^{-1}) were calculated using the following equation (McDonald and Harbaugh, 1988).

$$ET_G = ET_{G\max} \times \left(\frac{D-(H_s-H)}{D}\right) \tag{6}$$

where $ET_{G\max}$ is the maximum ET_G when water table is near to the surface (mm h^{-1}), D is the extinction depth (m), H_s is the elevation of land surface (m), and H is the elevation of water table (m).

Four representative observation wells (Obs1 to Obs4) were located along the cross-section. Obs1 is located in the riparian zone at 10 m from the river. Obs2 is at the foot of the hill, Obs3 in the middle of the hill slope, and Obs4 near the water divide (Fig. 2). Under the given $ET_{G\max}$ shown in Fig. 3, the model was run on hourly stress periods for 10 days. The simulated groundwater level hydrographs at the four observation wells are shown in Fig. 4.

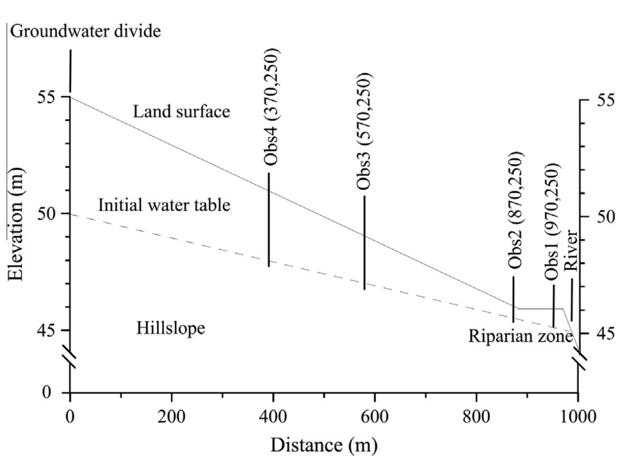

Fig. 2. Schematic hydrogeological cross-section and locations of the observation wells, and values in parentheses are the coordinates of the wells.

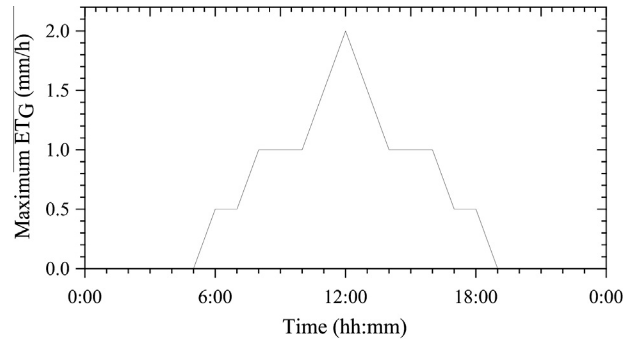

Fig. 3. Temporal pattern of $ET_{G\max}$ used in the numerical model.

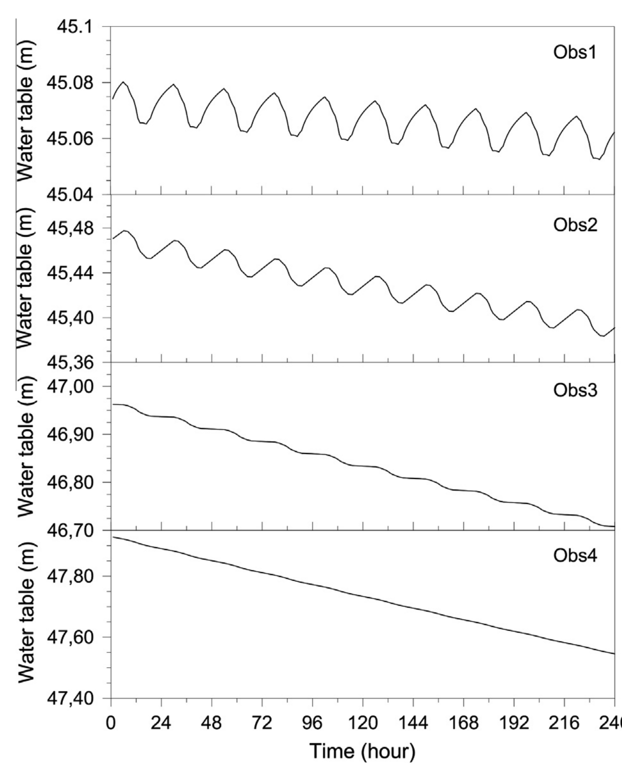

Fig. 4. Simulated groundwater level fluctuations at four observation wells.

3. Results of comparison

3.1. Comparison of the estimated ET_G for Obs1

The simulated hydrographs are shown in Fig. 4 and they are insensitive to the spatial discretization. The average difference of the simulated water table is about 1 mm when the spatial discretization was reduced to 2 × 2 m. Using the simulated hydrograph at Obs1 (Fig. 4), ET_G was calculated by the three methods. The groundwater inflow was calculated using the recovery period of 0:00 to 4:00 for the White method, the whole recovery period for the Hays method and 0:00 to 5:00 for the Loheide method. The average of the actual daily ET_G was calculated by Eq. (6) to be 12.54 mm/d. The average daily ET_G estimated by the White method, the Hays method and the Loheide method were 4.76 mm/d, 6.88 mm/d and 8.33 mm/d, respectively. It is clear that all three methods underestimate the actual daily ET_G. The White method underestimates the average daily ET_G by 62.0%, the Hays method underestimates by 45.1%, the Loheide method underestimates by 33.6%. The hourly ET_G can be calculated using the Loheide method. As shown in Fig. 5, the pattern of hourly ET_G estimated by the Loheide method mimics the actual hourly ET_G. But the averaged peak value of ET_G of 1.4 mm/h is smaller than the actual values of 1.9 mm/h.

The White method inherently introduces a significant source of error by assuming a constant net groundwater inflow for the 24-h period. As noted by Troxell (1936), net groundwater inflow in general is not constant over time and the variable groundwater inflow has been observed in the literature (Bulter et al., 2007; Lautz, 2008; Loheide, 2008; Gribovszki et al., 2008). In this study, Zonebudget (Harbaugh, 1990) was used to calculate the inflow and outflow of the model cell where the observation well is located. Flow into the observation cell is positive and out of the observation cell is negative. The sum of flows from surrounding cells to the observation cell is the net groundwater inflow as shown in Fig. 6a. It is clear that net groundwater inflow is not a constant, but also shows diurnal fluctuations.

The net groundwater inflow starts to increase immediately when ET_G occurs, and peaks around 13:00 that is about 1 h after ET_G reaches its maximum (Fig. 6a). During the night from 18:00 to 06:00, groundwater inflow decreases. In the White method, groundwater inflow from 00:00 to 04:00 is used to calculate the daily groundwater inflow, therefore, groundwater inflow is underestimated because during this period the inflow is small (Gribovszki et al., 2008). As shown in Fig. 6b, the daily groundwater inflow calculated by the White method was 4.40 mm/d, much lower than 12.18 mm/d calculated from the groundwater model.

Hays (2003) applied the hourly rate of water table rise between the minimum water level of the target day and maximum of the following day for the ET_G calculation. The groundwater inflow for the period of ET_G occurrence was obtained using the formula $S_y(H_2 - H_L)T_2/T_1$. As shown in Fig. 6b, the groundwater inflow calculated by the Hays method is 5.98 mm/d that is higher than the White method, but is less than half of that from the groundwater flow model.

The Loheide method calculates higher groundwater inflow (8.37 mm/d), but much less than that from the groundwater model (Fig. 7). The Loheide method calculates the recovery slope in finer time steps (hourly), it captures better the temporal variations in the recovery limb of the hydrograph.

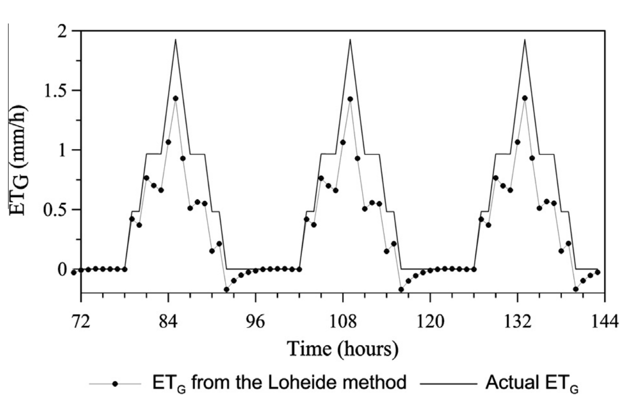

Fig. 5. Hourly ET_G estimated by the Loheide method and the actual hourly ET_G.

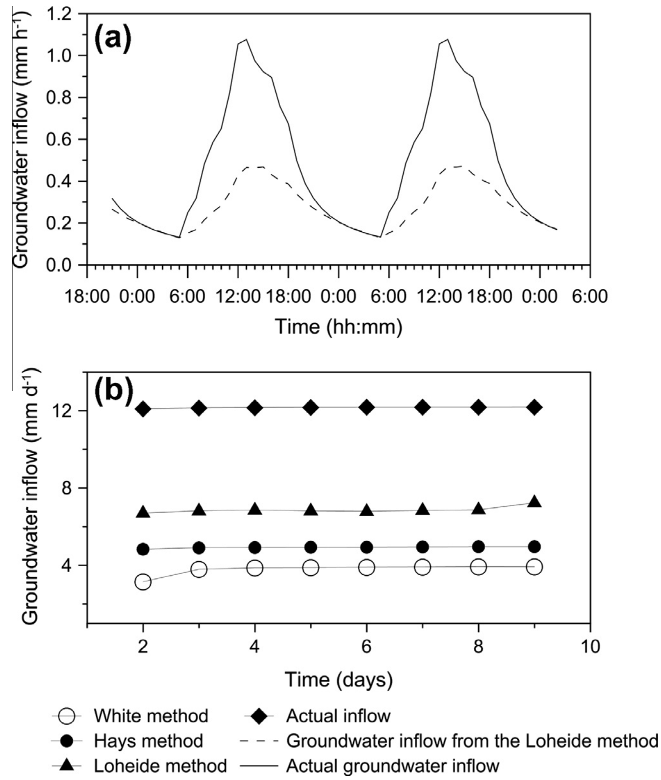

Fig. 6. Variation of groundwater inflow with time calculated by the Loheide method against actual groundwater inflow calculated by the model (a); and actual groundwater inflow and groundwater inflow calculated by the White method, the Hay method, and the Loheide method (b).

3.2. Comparison of the estimated ET_G for Obs2, Obs3 and Obs4

Modeled groundwater level hydrographs at Obs2, Obs3 and Obs4 are shown in Fig. 4. The magnitudes of diurnal fluctuations decrease with the increase of the water table depth towards the water divide. No recovery of groundwater levels during the night were observed in Obs3 and Obs4, only the slope of the water table decrease becomes smaller than that during the daytime. All three methods were applied to the simulated hydrograph of Obs2 to estimate ET_G using all three methods, while only the White method and Loheide method were used to calculate ET_G for Obs3 and Obs4 (Table 1) because the Hays method is not applicable in the case of the continuous decline of groundwater levels. Eq. (6) was used to calculate actual ET_G (Table 1). The average values of ET_G estimates by the White method, the Hays method and the Loheide method were close to the average of actual ET_G value for Obs2. The ET_G estimates from the White method and the Hays method are 10.70 mm/d and 10.79 mm/d respectively, about 3% higher than the actual value. While the Loheide method estimates about 5% higher than the actual value. For Obs3, the White method and

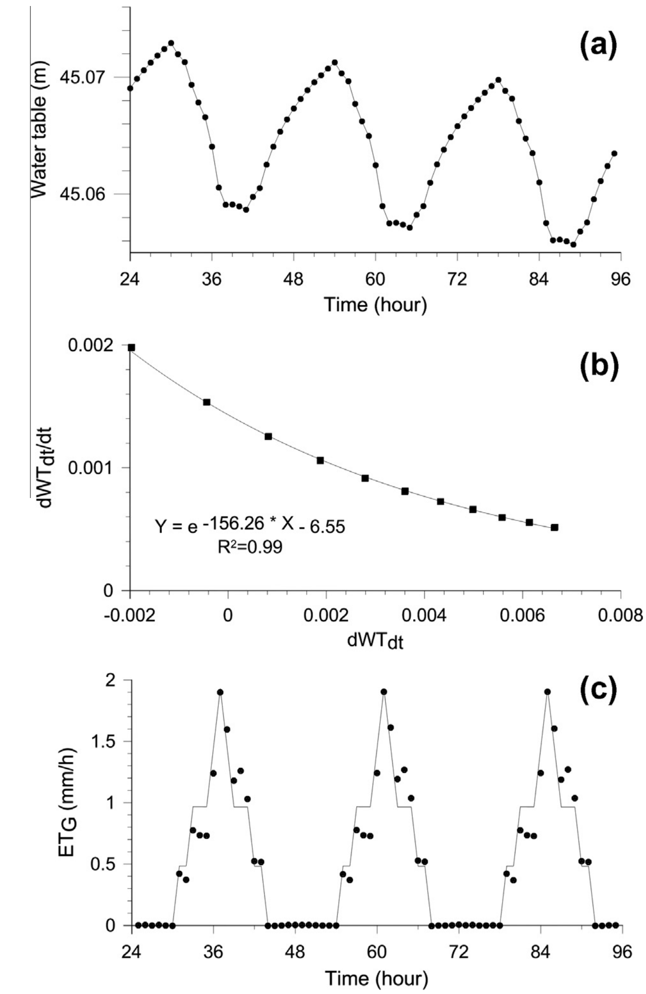

Fig. 7. Groundwater levels (a); relation between detrended water level and time rate of change of detrended water level (b); and the estimated hourly ET_G (dots) and actual values (solid line) and (c).

Table 1
Averaged actual ET_G and ET_G in mm/d estimated for Obs2, Obs3 and Obs4.

Observation wells	White	Hays	Loheide	Actual
Obs2	10.73	10.79	10.92	10.40
Obs3	5.05		5.02	5.03
Obs4	1.33		1.27	1.32

the Loheide method provide an estimation of 5.05 mm/d and 5.02 mm/d respectively, and they are almost identical to the actual ET_G (5.03 mm/d). For Obs4, the White method gives an estimation of 1.33 mm/d, slightly better than the Loheide method (1.27 mm/d).

The results show that all three methods can provide accurate daily ET_G estimates. The main reason is that the slope of groundwater level recovery is nearly constant in the hillslope. The slope during the period of 0:00 to 4:00 is 1.40×10^{-3} cm h^{-1} and the slope for the whole recovery period is 1.41×10^{-3} cm h^{-1} for the second day at Obs2. As a result, the basic assumption of the constant groundwater inflow in the White method is satisfied. In addition, the difference between the White method and the Hays method becomes minimal (Table 1) due to the relative constant slope. The reason for the Loheide method to overestimate ET_G is that the assumption of the linear relation between groundwater inflow and water table depth may not be met.

3.3. Modification of methods

Although the Loheide method produced a better estimate of ET_G at Obs1, the estimate still differs from the actual value (Fig. 5). As suggested by Loheide (2008), it is preferable to use the longest period to determine $\Gamma[WT_{DT}]$. But determining the time when ET_G stops is difficult. The time when the water table begins to increase may not be the time ET_G stops. At Obs1, water table starts to increase at 16:00 when ET_G is still occurring (Fig. 3). Groundwater inflow contributes to both the storage increase and ET_G when the water table is rising and ET_G is occurring at the same time. Only when ET_G stops, the rise of water table becomes faster. Therefore, the rate of water table change during recovery periods may be used as an indicator of the cease of ET_G. Fig. 7a shows that the fastest increase occurred around 18:00 that corresponded to the time when ET_G ended (Fig. 3). As a result, $\Gamma[\mathrm{WT_{DT}}]$ can be determined using the water table recovery data from 18:00 to 5:00. The best fitting curve was an exponential function for $\Gamma[WT_{DT}]$ (Fig. 7b), rather than a linear equation. Therefore, the Loheide method was modified by using the fitted exponential function to calculate ET_G. The average of daily ET_G calculated from the modified Loheide method is 12.31 mm/d, only about 2% lower than the average of actual daily ET_G. The hourly ET_G estimates using the modified Loheide method are much closer to the actual data (Fig. 7c) than those from the original Loheide method (Fig. 5).

The Loheide method is the only method that provides hourly estimates of ET_G, but its estimates of ET_G at Obs2 to Obs4 are less accurate. The White method provides accurate daily estimates of ET_G. A modified White method is proposed to calculate hourly ET_G:

$$ET_G = S_y \times (r + (H_{i-1} - H_i)) \quad (7)$$

where ET_G and S_y are the same as defined in the original White method; r is the slope of the recovery period of the previous day; i is the time step; H_i and H_{i-1} are the water tables in the ith and $(i-1)$th time steps, respectively. Fig. 8 shows the hourly ET_G calculated from the modified White method for Obs3 (Eq. (7)). The estimated hourly ET_G closely approximate the actual ET_G values with a 0.03% difference.

4. Case study

4.1. Estimate of ET_G in the Last Chance Watershed

The above methods were applied to a case study in the riparian zone of the Last Chance Watershed, California. USA. The detailed site descriptions and data collection can be found in Bulter et al. (2007). Data from the Doyle crossing well was used to calculate ET_G for the period of 7–11 July 2004 and can be found in Fig. 11

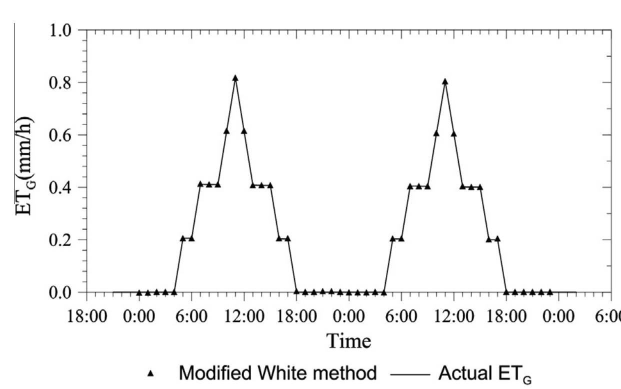

Fig. 8. Comparison of the actual ET_G and ET_G estimated by the modified White method.

of Bulter et al. (2007). Hourly water table data on 7 July are shown in Fig. 9a. In the study, S_y is assigned a value of 0.02. The White method, the Hays method, the original and modified Loheide methods were used to calculate ET_G. Fig. 9a shows that the fastest increase of groundwater level occurred around 22:00, indicating the recovery period begins in 22:00. Therefore, $\Gamma[WT_{DT}]$ was determined using the water table recovery data from 22:00 to 8:00 using an exponential function in the modified Loheide method and using a linear function in the original Loheide method.

The results are shown in Fig. 9b. Average values of daily ET_G were estimated to be 3.09, 3.41 mm/d, respective by the White-method and the Hays-method. The modified and original Loheide methods gave the average daily ET_G of 4.62 mm/d and 5.64 mm/d, respectively. The results show that the White-method calculates the lowest ET_G value, and the modified Loheide-method gives the highest value, and the Hays-method yields the value in between. The results are consistent with the previous modeling study.

4.2. Estimate of ET_G in the Hailiutu River Catchment

The study area is located in the Hailiutu River Catchment and detailed site descriptions can be found in Yang et al. (2011). The water table was measured in a well located in a hillslope, about 8 km away from the river. The hourly data was collected for five days using a Kellor DCX-22A data logger (Kellor, Winterthur, Switzerland) and are shown in Fig. 8c. Regular diurnal fluctuations with a magnitude of ~2 cm were recorded (Fig. 9c). A pumping test gave a specific yield of 0.12. ET_G was calculated using the three methods and the results are shown in Fig. 9d. The ET_G estimated by the White method, the Hays method and the Loheide method are similar and the averaged values are 5.24 mm/d, 4.73 mm/d and 4.92 mm/d, respectively. The conclusion is consistent with the modeling study described earlier in this study.

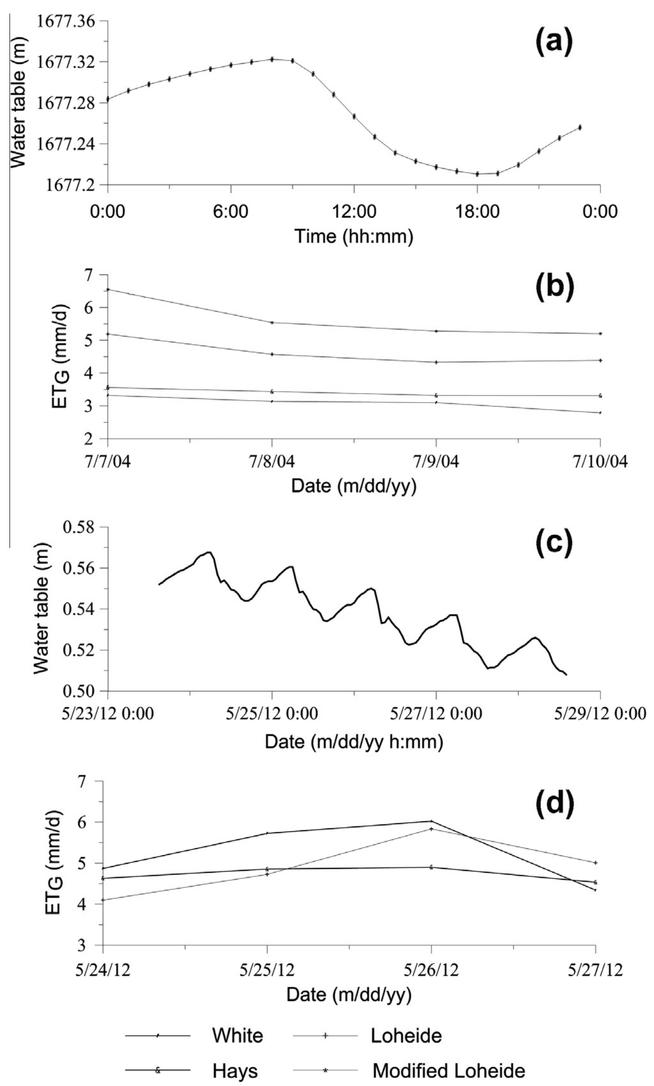

Fig. 9. Water table fluctuations on 7 July 2004 in the Doyle crossing well (a); ET_G estimation in the riparian zone (b); water table fluctuations of a well in the Hailiutu River catchment (c); ET_G estimation in the hillslope zone (d).

5. Discussions

ET_G has two contributing sources, i.e. the groundwater storage and net groundwater inflow. Therefore, the uncertainty in estimating ET_G may originate from the uncertainties in groundwater storage and/or net groundwater inflow. Fig. 10 shows the hourly contributions of groundwater storage and net inflow to ET_G at the observation wells. The positive values of net inflow indicate flow into is more than flow out of the target cell and the positive values of groundwater storage change indicate the depletion of groundwater storage.

For the well Obs1, both the change of groundwater storage and net groundwater inflow show the similar diurnal variations as the ET_G (Fig. 10a). During night, the net groundwater inflow supplies water to the target cell and the water table recovers. During daytime, the net groundwater inflow is positive, but the water table declines (positive values of groundwater storage change). This indicates ET_G consumes both groundwater inflow and groundwater storage. The groundwater inflow supplies more water to ET_G than the groundwater storage since the decrease of water table by evapotranspiration induces more groundwater inflow. The groundwater inflow contributes 57% of ET_G from 6:00 to 12:00 h, and accounts for 91% of ET_G from 12:00 to 18:00 h. The underestimation of ET_G by the three methods is due to the inaccuracy in calculating the contribution from groundwater inflow as discussed before. The modified Loheide method improves the accuracy of calculating groundwater inflow, and therefore, better results can be obtained.

For wells Obs2, Obs3 and Obs4, the change of groundwater storage shows similar diurnal variations as the ET_G, while the net groundwater inflow is approximately constant over time (Fig. 10b–d). Only at Obs2 near the foot of the hill, the groundwater inflow contributes significantly to the ET_G during the day and groundwater storage recovery during the night (Fig. 10b). Therefore, the water table at Obs2 exhibits diurnal fluctuations (Fig. 4). At the well Obs3 in the middle of the hill, the net groundwater inflow is almost zero, groundwater evapotranspiration consumes solely groundwater storage, and therefore, groundwater levels in Obs3 does not recover (Fig. 4). At the well Obs4 near the water divide, the net groundwater inflow becomes negative (Fig. 10d), the groundwater storage is consumed not only by evapotranspiration, but also by the lateral flow to down gradient, and groundwater levels continue to decline (Fig. 4). All three methods can estimate accurately the change of groundwater storage and the constant groundwater inflow. Therefore, they can provide similar accurate estimates of ET_G (Table 1). Moreover, the modified White method can provide hourly ET_G estimates.

In real systems, groundwater level variations are more complex because of multiple influences of many factors, such as heterogeneous hydraulic conductivity, topography, vegetation cover, and human activities. As a rule of thumb for selecting a proper method to estimate ET_G from groundwater level hydrograph, patterns of diurnal groundwater level variations should be analyzed. Fig. 11a shows the groundwater head differences between the observation

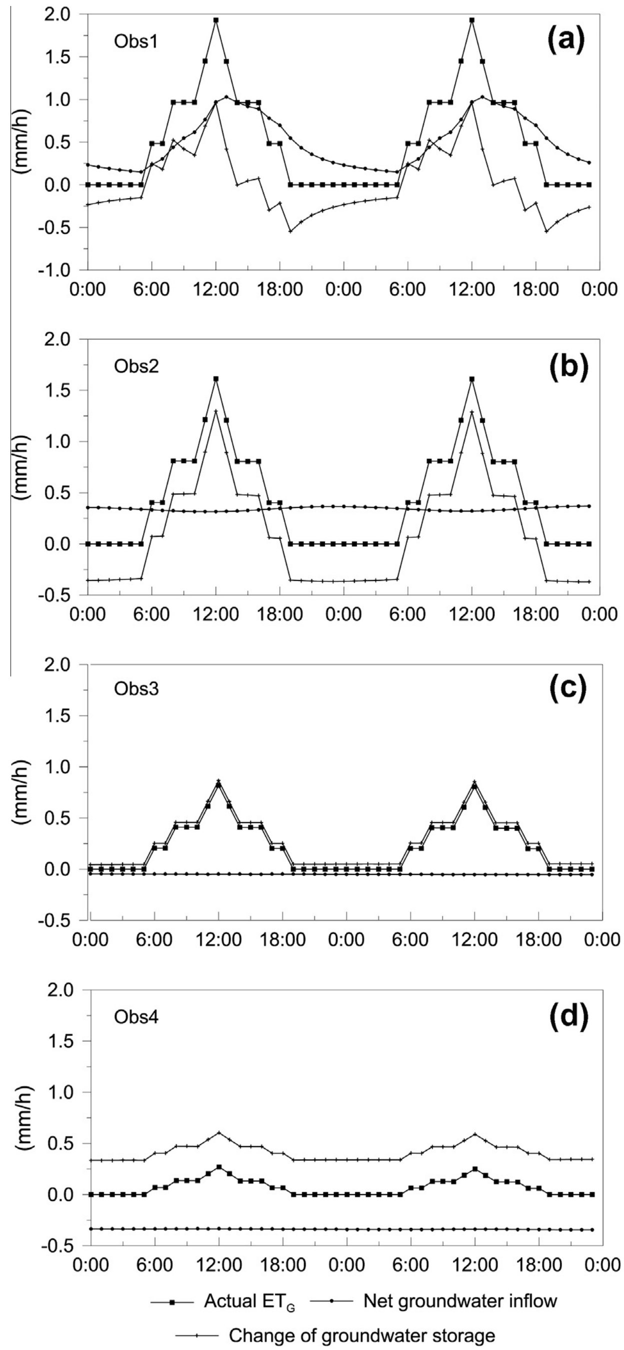

Fig. 10. Contribution of groundwater storage and inflow to ET_G in observation wells.

well (Obs1) and the up-gradient well and down-gradient well. The head differences show diurnal variations, and the head differences in the up- and down-gradient areas differ with larger head differences for the up-gradient area. Moreover, the two curves are not parallel which suggests that the rates of head differences for the two areas vary. This will lead to a varying slope of groundwater level recovery. In this case, the Loheide or modified Loheide methods should be selected. Fig. 11b shows that the groundwater head differences between the observation well (Obs2) and the up-gradient well and down-gradient well in the hillslope. The up- and down-gradient differences are parallel, resulting in a constant net groundwater inflow. This will result in a constant slope of groundwater level recovery. In this case, the White or modified White

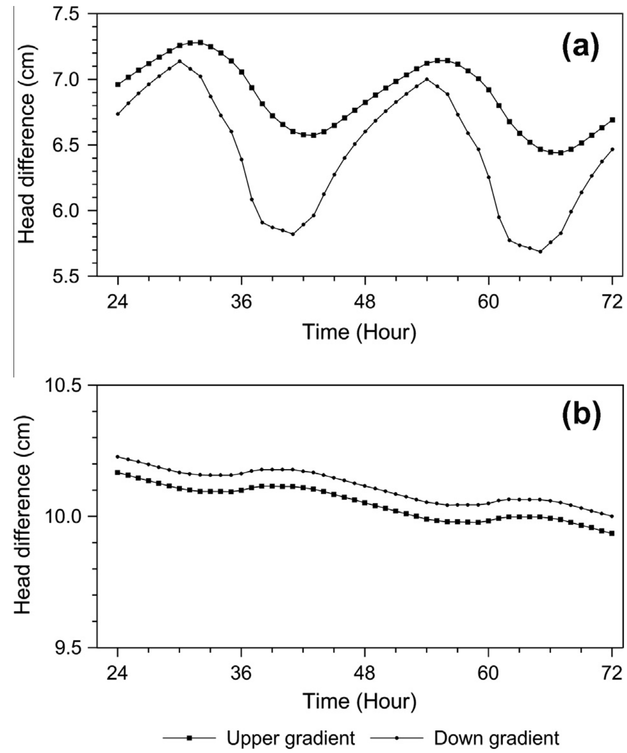

Fig. 11. Up- and down-gradient head differences at observation wells Obs1 (a) and Obs2 (b).

methods would be better choices because of their simplicity in the calculation.

6. Conclusions

This study compared methods of estimating ET_G using hourly groundwater level hydrographs. The hourly groundwater level hydrographs were generated by a groundwater flow model with diurnal evapotranspiration. Four representative groundwater observation wells were selected: one in the riparian zone near to the river, one at the foot of the hill, one in the middle of the hill slope, and one near the water divide. The model simulated hourly groundwater levels for 10 days and calculated the actual ET_G, net groundwater inflow and change of groundwater storage at these four observation wells. The White method, the Hays method and the Loheide method were applied to the four hydrographs to estimate ET_G.

All three methods underestimated daily ET_G at Obs1 where net groundwater inflow contributes more water to evapotranspiration than the groundwater storage. This is because these methods do not account for varying net groundwater flow. The Loheide method is comparatively better with a highest ET_G value. When an exponential equation is used to calculate the rate of water table recovery, the Loheide method can approximate hourly ET_G more accurately.

All three methods estimated accurately daily ET_G at the other three wells as groundwater storage contributes more water to evapotranspiration than net groundwater flow. The net groundwater inflow is also almost constant. All methods can calculate accurately the change of groundwater storage and the constant inflow. The modified White method can also calculate hourly ET_G and is preferably to be used.

The analysis of the pattern of the recovery limb of the groundwater level hydrograph and up and down gradient groundwater level differences can help select a suitable method to estimate ET_G. When the recovery limb of the groundwater level hydrograph is linear and the up and down gradient groundwater level differences are parallel and constant, the White or modified White methods can be used. Otherwise, the Loheide or modified Loheide methods are recommended.

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