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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 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.

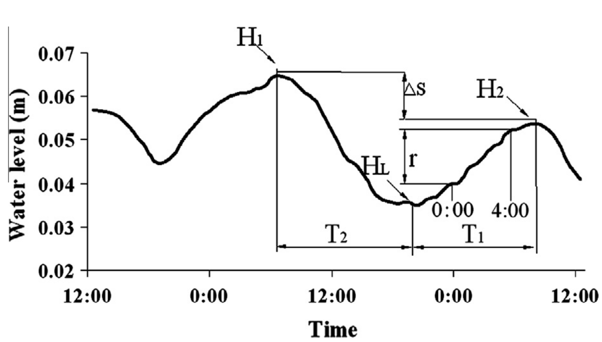

Fig. 1. Examples of diurnal groundwater level fluctuations, with variables used in Eqs. (1) and (2).

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:

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