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Review

A Review of Advances in Groundwater Evapotranspiration Research

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Abstract: Groundwater evapotranspiration (ET_g) is an important component of the hydrological cycle in water-scarce regions and is important for local ecosystems and agricultural irrigation management. However, accurate estimation of ET_g is not easy due to uncertainties in climatic conditions, vegetation parameters, and the hydrological parameters of the unsaturated zone and aquifers. The current methods for calculating ET_g mainly include the WTF method and the numerical groundwater model. The WTF method often requires data supplementation from the numerical unsaturated model to reduce uncertainty; in addition, it relies on point-monitoring data and cannot solve the spatial heterogeneity of ET_g. The ET_g calculation module of the numerical groundwater model is set up too simply and ignores the influence from the unsaturated zone and surface cover. Subsequent research breakthroughs should focus on the improvement of WTF calculation theory and the setting up of an aquifer water-table fluctuation monitoring network. The numerical groundwater model should couple the surface remote sensing data with the unsaturated zone model to improve the accuracy of ET_g calculation.

Keywords: groundwater evapotranspiration; water-table fluctuation method; unsaturated zone; aquifer; MODFLOW; HYDRUS

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1. Introduction

Water scarcity is already the biggest challenge for global agricultural development [1], with one third of the population in developing countries living in water-scarce areas and fifty-four percent of agricultural land also being located in water-scarce areas [2,3]. In water-scarce areas, groundwater evapotranspiration is an important part of the hydrological cycle; it is one of the main sources of regional evapotranspiration and the main consumer of groundwater in areas with a shallow water table [4,5]. Arid and semiarid regions occupy approximately 30% of the land surface of the Earth [6], including the majority of northern and southern Africa; the Middle East; western USA and southern South America; most of Australia; large parts of central Asia; and parts of Europe [7]. Vegetation provides natural protection against desertification and dust storms in these regions. Some vegetation, known as phreatophyte, is groundwater dependent [8]. Phreatophyte transpiration consumes groundwater and causes diurnal fluctuations of groundwater levels [9]. On the other hand, surface water is scarce, and groundwater is often the only reliable water resource for socio-economic development in arid regions [10]. 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 [11]. The sustainable management of groundwater resources must consider water use both from 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) [12]. Accurate estimation of groundwater evapotranspiration is essential for understanding hydrological cycle processes and sustainable groundwater resource use and management [13–15], and it is useful for natural ecosystem conservation and restoration [16]. The quantification of groundwater evapotranspiration is particularly important in areas dependent on groundwater ecosystems [17–20] and is important for water management of crops and the investigation of soil salinization processes [21–23].

Groundwater evapotranspiration (ET_g) can result in significant loss of groundwater storage. ET_g is generated when water moves from the unsaturated zone to replenish soil storage depleted by surface evapotranspiration and root water uptake. The shallow water table allows groundwater to be used to directly supply crop growth [24–29], so an accurate estimation of ET_g for farmland can help to improve irrigation management. Any efforts toward improving ET_g estimation methods are worthwhile for agricultural water management and land and water environmental protection [15].

However, accurate estimation of groundwater evapotranspiration remains a challenge because it is often influenced by uncertainties associated with climatic variables, vegetation parameters, geological variables, and hydrologic parameters [30,31].

Numerous studies have been conducted on evapotranspiration estimation models worldwide, and these studies have mainly included empirical statistical models, energy balance models, remote sensing models based on Penman's formula, complementary correlation models, and hydrological models. The evapotranspiration model (later called the Food and Agricultural Organization Penman–Monteith (FAO56-PM) model) proposed by Allen [32] is the most suitable model for estimating international evapotranspiration at present, and this model has been widely used worldwide. Evapotranspiration (ET) includes surface evaporation (E_s), evaporation of water from below the ground surface (E_{ss}), and transpiration of water by plants (T_{ss}). The latter two were together defined as subsurface evapotranspiration (ET_{ss}) [33], which includes groundwater evapotranspiration (ET_g) and unsaturated water evapotranspiration (ET_u) [34].

There are many computational methods applied to different scales to calculate evapotranspiration, including the method based on evapotranspiration lysimeter weighing [35–38], the method based on field water balance equations [39–42], micrometeorological methods [16], vorticity covariance methods [43–46], the Bowen ratio method [47], and regional-scale ET calculation models (for example, the TSEB model [48], the SEBAL model [49], the S-SEBI model [50], the SEBS model [51], and the LandSAF model [52]), as well as mapping of evapotranspiration using the internal scale method (METRIC) [53], the STSEB model [54], the GLEAM model [55], the MODIS-ET model [56], and the ETwatch product [57]. However, these methods are usually used more for the measurement and calculation of earth surface evapotranspiration and are not able to directly measure the value of groundwater evapotranspiration, mainly because the hydraulic connection between the earth surface and the aquifer is blocked by the unsaturated zone, and the uncertainty of the unsaturated zone increases the difficulty of calculating groundwater evapotranspiration. Therefore, the calculation of groundwater evapotranspiration often needs to consider the variation of moisture content in the unsaturated zone. In addition to this, the accurate estimation of ET_g needs to consider the variation of local atmospheric conditions [58] and groundwater table variation [59], and is also influenced by the spatial heterogeneity of land use [60]. Moreover, many factors such as lateral inflow at the recharge boundary, vertical recharge at the surface, and complex geological structure and soil composition in the unsaturated zone can affect ET_g [61–64]. In addition, intensive anthropogenic measures, such as water diversions and irrigation in water-scarce areas, also have a significant impact on groundwater evapotranspiration. External water transfers directly change the water storage capacity of inland lakes, and overall raise the groundwater table in the wetlands around the lakes and diversion channels. This change in groundwater is continuous, unlike the transfer of water for irrigation in agricultural areas. The shallower water table makes the water storage capacity and regulation capacity of the unsaturated zone weaker, the capillary zone is closer

to the surface, and thus groundwater evapotranspiration becomes stronger; in addition, the shallower water table allows more channels for groundwater to rise into the air, and the root systems of some deep-rooted plants can act directly on the unsaturated zone and even the aquifer. The branches and roots of plants establish the hydraulic connection between air and aquifer, and driven by the transpiration of plants, a large amount of groundwater enters the air directly without passing through the unsaturated zone. The shallower unsaturated zone channels and the newly added plant channels change the proportion of groundwater evapotranspiration (ET_g) in total evapotranspiration (ET_a). This combination of multiple factors makes ET_g difficult to calculate.

2. Methodology

2.1. Advances in Research on Groundwater Evapotranspiration

Historical studies of groundwater evapotranspiration by researchers date back to the 1920s. The early researchers found that groundwater is constantly supplied with evapotranspiration through "capillary rise," which is evident in groundwater when the water table is less than 3 m. Remson and Fox [65] proposed a method for estimating groundwater discharge by evapotranspiration from the water capillary rise of the water table. "Potential capillary water loss" is defined as a measure of the ability of the soil capillary gap to raise water from the groundwater to the earth surface, and they consider the depth of the water table as the most important factor affecting the magnitude of groundwater evapotranspiration. Gardner's [66] analysis showed that the evapotranspiration rate depends on the depth of the water table and the capillary flux. Subsequently, Gardner [66] and Willis [67] proposed the calculation of ET_g as a function of water table depth, and although their assumption of the constant moisture content in the unsaturated zone is inaccurate today, this idea has had a profound impact on subsequent studies of groundwater evapotranspiration. Schoeller [68] introduced the concept of ultimate evapotranspiration depth, and he theorized that groundwater evapotranspiration occurs only when the groundwater table depth is smaller than the ultimate evapotranspiration depth. The magnitude of the groundwater evapotranspiration value is determined by the groundwater table depth together with the potential evapotranspiration value, and this method is widely used because of its simple form. In the classic groundwater numerical model MODFLOW, the groundwater package EVT still follows this idea in the calculation of evapotranspiration. Doorenbos and Pruitt [69] proposed a method to calculate ET_g by quantifying the variation of soil water content in the root zone. They were the first to elaborate on the relationship between ET_g and soil moisture in the root zone of crops. This method has been used in soil water balance models for calculating the magnitude of ET_g. The calculation of soil water balance is a specific application of the water balance equation theory, which is also a common method for groundwater evapotranspiration calculations and is applicable to different scales [4,70–72].

Wang et al. [15] used the water balance equation to calculate the soil water content in the root zone under irrigation and rainfall conditions, and then proposed a new equation that integrates multiple influencing factors to estimate ET_g during the growing season based on the methods proposed by Doorenbos and Pruitt [69] and the Averianov equation [68]. Groundwater evapotranspiration was calculated using the following equation.

$$ET_g = K_c \times ET_0 \times \left(1 - \frac{H}{H\max}\right)^n \times \frac{\theta_{fc} - \theta}{\theta_{fc} - \theta_r} \quad (1)$$

where K_c is crop coefficient, ET_0 is reference crop evapotranspiration in mm·day^{-1}, H is the actual water table depth in m, H_{max} is the potential maximum depth in m, beyond which no ET_g occurs; n is the soil characteristics parameter, θ is the actual averaged soil water content in the root zone in cm^3·cm^{-3} (usually about 60 cm below the soil surface), θ_{fc} is the field capacity of the soil in the root zone in cm^3·cm^{-3}, θ_r is the soil water content close to

permanent wilting point in $cm^3 \cdot cm^{-3}$ (in this paper, a constant value 0.05 is used [73]). The method was tested and validated against the data from the lysimeter experiments.

Wang et al. [74] then used the following water balance equation to calculate ET_g based on ET_a obtained in a subsequent study.

$$ET_g = (P + I) + \Delta SWC - ET_a - P_c \quad (2)$$

where P and I are precipitation and irrigation in mm, respectively, ΔSWC is the variation in soil water storage up to 90 cm depth where most of the maize root system is concentrated [75], and P_c is the deep percolation to shallow groundwater. Since the daily ΔSWC was sometimes too small to accurately determine using Hydra probe measurements, the two-day average was used to calculate the water balance. In any event, it was assumed that either P_c or ET_g was zero. Lai et al. [76] considered quantifying the contribution of shallow groundwater to evapotranspiration (ET_g) as an important topic that has been extensively studied [77,78]. They used lysimeters to calculate groundwater evapotranspiration values for wheat fields in the lower Yellow River basin at different groundwater level conditions. He concluded that reasonable groundwater level control can help increase yields while reducing the risk of soil salinization and is important for sustainable management of the lower Yellow River basin.

2.2. Using the WTF Method to Calculate the ET_g

Among the methods for calculating groundwater evapotranspiration considering the influence of crops or plant roots, the groundwater table fluctuation (WTF) method is one of the most commonly used methods. In recent decades, there has been an increasing emphasis on using the WTF method to quantify ET_g. Calculating ET_g based on the WTF method is a relatively straightforward, simple and inexpensive method [9,79,80], Users can measure water loss due to evapotranspiration directly from groundwater table changes and therefore does not require additional measurements at the soil surface [73,81–83]. The use of the groundwater table fluctuation method assumes that groundwater table changes in a shallow aquifer are caused by evapotranspiration only [84], and this method has been widely used to estimate ET_g in riparian zones and wetlands [13,58,85–87].

The method of groundwater table fluctuation exploits the law of daily variation of the water table in riparian zones or wetlands. During the daytime, plant transpiration makes the groundwater table lower, and at night, the intensity of plant transpiration decreases and a significant rebound of the groundwater table level occurs. Based on the discovery of this pattern, White [81] proposed a classic method for calculating groundwater evapotranspiration, namely the White method. This method has long been used for groundwater evapotranspiration rates in arid and semi-arid areas [83,88–90]; it has been continuously improved by many researchers and has been applied to wetlands [8], riparian zones [79,91–93], prairies [94], and forests [95] in a variety of ecosystems.

2.2.1. White Method

The White method is the most classic method of the water-table fluctuation methods. White theorized that groundwater evapotranspiration is mainly composed of two parts—the recovery, and the storage variation in 24 h—based on the regular fluctuation of groundwater, which can be calculated by the following equation.

$$ET_g = S_y(24r \pm s) \quad (3)$$

where S_y is the specific yield of the shallow aquifer, r is the groundwater recharge rate from 00:00 a.m. to 04:00 a.m., and s is the 24-h shallow groundwater table variation. However, the White method is not universally applicable, and his application is based on four assumptions: (1) the day and night dynamics of groundwater table fluctuation are caused by plant evapotranspiration; (2) the evapotranspiration of plants from 0:00 a.m. to 4:00 a.m. at night is 0; (3) the groundwater recovery rate at night is a constant value; and (4) the specific

yield is a constant value and representative. Due to the simplicity and ease of operation of White method, this method is also the most widely used for evapotranspiration calculation at the site scale. However, some of the elements in the assumptions are subjective and limited, and researchers have made corresponding improvements in subsequent studies based on the shortcomings of the assumptions of the White method. Inspired by White method, many researchers have proposed other methods based on water-table fluctuation information, including the Dolan method [96], the Hays method [13], the Gribovszki method [85], the Loheide method [62], the Soylu method [86], etc. These methods are based on the White method and classed as approaches that improve upon the White method (Figure 1).

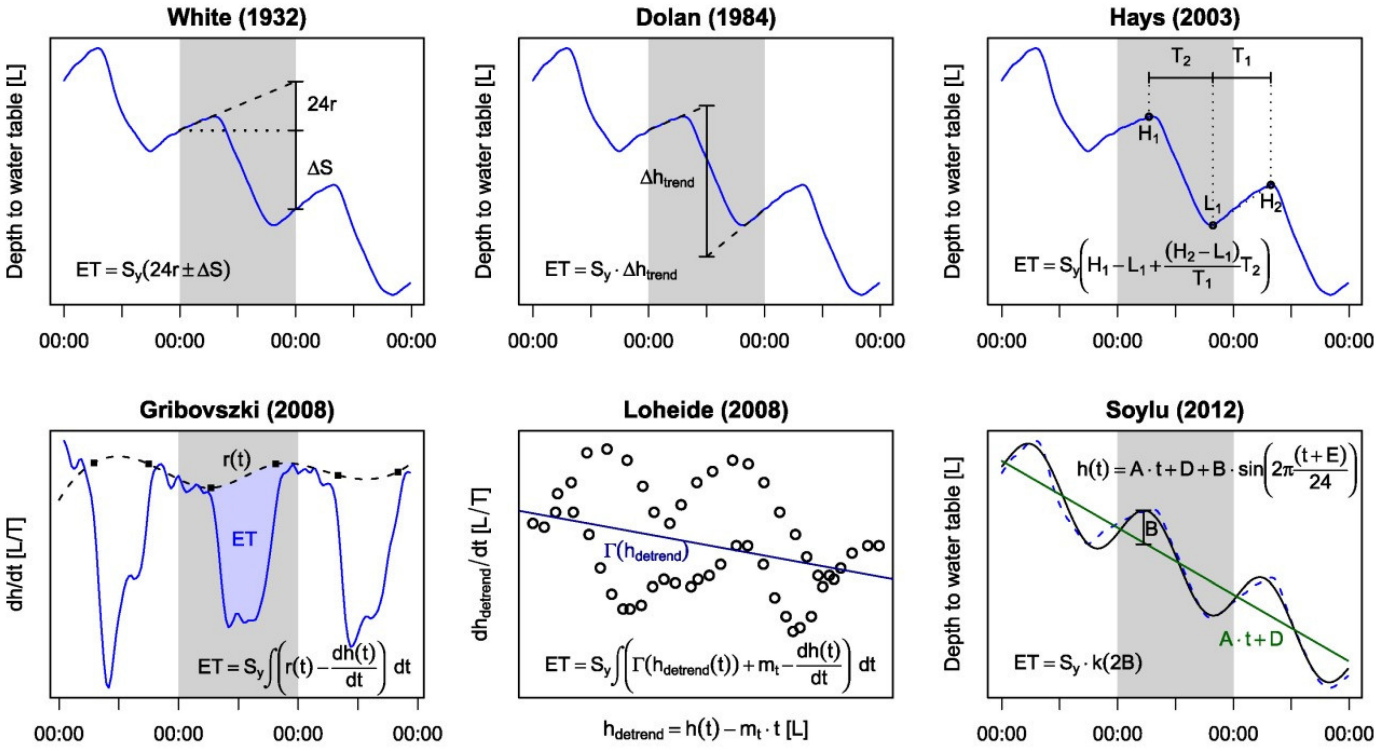

Figure 1. Computational schematic of the White method and its improved approaches [88].

2.2.2. Dolan Method

Dolan et al. [96] utilized continuous records of water-table elevation in the marsh soil, and related a drop in the water table to evapotranspiration loss. The observed rise or fall in the groundwater table at night represents the net inflow or outflow of water into or out of the marsh due to hydraulic forces alone. The rate of change of each night's groundwater table rise was extrapolated to noon the next day and back to noon the day before. The midday elevation represents the altitude of the groundwater table if no evapotranspiration had occurred during the 24 h period. The cycle is centered on each successive night. Thus, the difference between the elevation extrapolated from the previous night and the elevation extrapolated from the next night represents the water loss due to evapotranspiration on that day.

2.2.3. Hays Method

Hays [97] developed a new method for estimating ET_g. The biggest advantage of this method is the flexibility to determine the time of the water-table rise and fall according to the water-table waveform, instead of fixing it at a certain definite time period. Its calculation equation is:

$$ET_g = [(H_1 - H_L) + \frac{H_2 - H_L}{T_1} T_2] \times S_y \quad (4)$$

where H_1 is the maximum water table depth on the morning of the calculation day in m, H_2 is the maximum water table depth on the next day in m, H_L is the minimum water table depth on the day of the calculation in m, T_2 is the duration of the groundwater table decline period, and T_1 is the duration of the groundwater table rise period. The prerequisites for the use of the Hays method are similar to those of the White method.

2.2.4. Loheide Method

The Loheide method [62] is an improved method based on the White method. The main idea of the Loheide method is that it is first assumed that the trend of the water table change in the recharge source is consistent with the general trend of observed water table change, so that the measured groundwater table can be detrended. The detrending analysis of the water table included in this method not only improves the calculation accuracy of vegetation evapotranspiration, but also reduces the uncertainty in the calculation process. The change in groundwater storage near the water table observation wells is expressed as the change in water table with time (dWT/dt) and is controlled by the net inflow or outflow of groundwater in the vicinity (r(t)(L/T)) together with the ET_g.

$$S_y{}^* \frac{dWT}{dt} = r(t) - ET_g(t) \tag{5}$$

When ET_g is 0, Equation (3) can be simplified as

$$S_y{}^* \frac{dWT}{dt} = r(t) \tag{6}$$

It has been clarified that the daily recharge rate is a function of time, and Loheide [52] assumes a constant head in the recharge source area and that the recharge rate can be obtained from water table changes, so Equation (5) can be expressed as

$$r(WT) = S_y{}^* \frac{dWT}{dt} \tag{7}$$

The Loheide method considers the trend of the groundwater recharge zone water table to be included in the water table change at the observation point, so this trend is therefore removed from the water table change at the observation point.

$$WT_{DT}(t) = WT(t) - m_T \times t - b_T \tag{8}$$

where $WT_{DT}(t)$ is the detrended water table, WT(t) is the observed water table, m_T is the trend line slope, and b_T is the trend line intercept. dWT_{DT}/dt and $WT_{DT}(t)$ using the water table of ET_g as 0 in the early morning of the current day and the next day, established the functional relationship $\Gamma(WT_{DT})$, and thus the recharge rate function is obtained, which can be expressed as

$$r(t) = S_y^* \times [\Gamma(WT_{DT}(t)) + mT] \tag{9}$$

Finally, the available ET_g is expressed as:

$$ET_g(t) = r(t) - S_y^* \times \frac{dWT}{dt} \tag{10}$$

S_y* in the above equation is the complete specific yield instead of the traditional specific yield concept. The reason for the proposed concept of complete specific yield is because Loheide [62] theorized that the error of specific yield is an important factor in causing the error in the calculation of groundwater evapotranspiration. Therefore, he

proposed the concept of complete specific yield, and the equation of complete specific yield is as follows

$$S_y^* = S_{yu} - \frac{S_{yu}}{\left[1 + \alpha\left(\frac{z_i + z_f}{2}\right)^n\right]^{1-\frac{1}{n}}}, S_{yu} = \theta_s - \theta_r \quad (11)$$

where θ_s is the saturated soil water content, θ_r is the residual water content, and z_i and z_f are the initial and end values for the variation of groundwater table depth. The parameters α and n are in the van Genuchten model. The average values are taken when the groundwater table varies in the soil layer.

2.2.5. Yin Method

The Loheide method assumes that the rate of change of the detrended water table is linearly related to the detrended water table. In response to this assumption, Yin et al. [12] further refined the Loheide method in their study, and found that applying the exponential equation to fit the relationship curve between the rate of change of the detrended water table and the detrended water table, can yield more accurate calculation results. Inspired by the Loheide method, Yin et al. [12] proposed an improved calculation of hourly-scale evapotranspiration based on the White method, as Equation (12)

$$\mathrm{ET}_g = S_y \times (r + (H_{i-1} - H_i)) \quad (12)$$

where S_y is the specific yield, r is the water level recovery rate (LT^{-1}), i is the moment value (T), and H is the water table value at the i^{th} moment (T). Compared with the Loheide method, the calculation process of this method is simpler.

2.2.6. Gribovszki Method

Gribovszki et al. [85] also concluded that the rate of water table recovery is not constant throughout the day and proposed the use of empirical interpolation methods to calculate a non-constant rate of water table recovery throughout the day. The main idea is to estimate the rate by using hydraulic derivation and empirical methods. The maximum (positive) rate of water table changes between midnight and 6 a.m. and the average of dh/dt are chosen to obtain the maximum and minimum recharge rates, respectively. These values were assigned to the times of maximum and minimum groundwater table rates of recovery, respectively. Therefore, two points (sections) were defined for each day. Based on the points for several days, a spline interpolation was performed to describe the recharge rate r(t) (L/T) over time. Subdaily evapotranspiration was then calculated (e.g., dt = 1 h).

$$\mathrm{ET(t)} = S_y\left(r(t) - \frac{dh}{dt}\right) \quad (13)$$

However, although the empirical interpolation method obtains the variation of water table recovery rate for each hour of the day, there may be some error in calculating the hour-by-hour water level recovery rate because the empirical interpolation method is based on only two pieces of daily data.

2.2.7. Soylu Method

Soylu et al. [86] proposed to use a new 'Fourier method' to calculate groundwater evapotranspiration. They found that the amplitude of the fluctuations of the water table after detrending can be used to calculate evapotranspiration directly by combining the water balance equation of the White method and the Fourier equation proposed by Czikowsky and Fitzjarrald [98]. The equations are as follows.

$$\mathrm{H(t)} = A \times t + D + B\sin\left(2\pi\frac{(t+E)}{24}\right) \quad (14)$$

$$\mathrm{ET}g = Sy \times k(2B) \quad (15)$$

where H is the water table depth (L), A is the water table depth change trend for multiple days (LT^{-1}), t is the time (T); D is the average deviation of water table depth change (L), B is the diurnal fluctuation amplitude (L), E is the day and night fluctuation phase (T), and k is the empirical correction factor for correcting the water table recovery and evapotranspiration components included in A × t.

2.2.8. Wang Method

Wang et al. [13] used the statistical method of day and night fluctuations of the water table to analyze the characteristics of de-trended groundwater table fluctuations, and then calculated the evapotranspiration at different time scales based on the relevant parameters. This method can effectively deal with continuously changing groundwater table fluctuation data.

$$\mathrm{ETg} = Sy \times \frac{\sigma}{\lambda} \tag{16}$$

where σ is the variance (L) of the variation of the de-trended water table, and λ is the evapotranspiration cycle correlation coefficient. It can be seen that the subsequent improvements in the water table fluctuation methods—both the Fourier method and the water table statistics method—are mostly improvements in the fluctuation characteristics about the detrended water table, and this series of improvement ideas is derived from the detrending theory proposed by Loheide.

2.2.9. Other Improvements and Applications

Wang et al. [99] concluded that one of the limitations of the White method is the large uncertainty in quantifying the daily groundwater recovery rate (r). Since ET_g is highly dependent on the shape and duration of the diurnal clear-sky solar radiation curve, using the groundwater recovery rate at short nighttime intervals to represent daily r may lead to large uncertainties in the ET_g estimates. They analyzed the dependence of the estimated daily r on sunset and sunrise times. and found that the estimated r is highly sensitive to the duration between sunset and sunrise and varies with the season. Instead of using fixed time spans (TS_s), they suggest using a more universal method for determining TS_s for estimating daily r. They tested this dynamic T of S method at a *Tamarix ramosissima*-dominated riparian site in northwestern China. The results proved that their improved method was better and less subjective than the traditional White method. Subsequently, many researchers have used these representative methods mentioned above in different study areas for comparison to see their applicability. Three water table fluctuation methods [13,69,74] were used by Su et al. [100] in 2017 to calculate daily ET_g in a riparian forest area in northwestern China. The purpose of comparing these three methods was to evaluate and compare their performance under various groundwater table conditions in the wild. The results indicated that the White method is applicable to the period of declining groundwater table. In addition, the selected time period may affect the estimation of ET_g. The Soylu and Hays methods performed well under various groundwater table conditions. Therefore, it appears that the Hays and Soylu methods are more suitable for long-term ET_g estimation in the wild. In addition, it was found that the percentage of water transpired by plants from groundwater varies during the growing season, and that riparian plants mainly use soil water in early growth and tend to use groundwater in late growth. Yin et al. [12] also compared three frequently used water table fluctuation methods: the White, Hayes, and Loheide methods. These three methods use the water table generated by the model to calculate the ET_g. The comparison of actual and estimated ET_g reveals the accuracy of each method and determines the applicability of the methods. When the recovery branch of the groundwater table process line is nonlinear, these methods underestimate daily ET_g. The Loheide method is relatively good, and all three methods can accurately estimate daily ET_g. The modified White method can provide hourly ET_g estimates and is recommended for general use. It was found that in practical applications, analysis of the shape of the groundwater table recovery branch and the difference in groundwater head between the upper and lower gradients can determine the appropriate estimation method

of ET_g. Fahle and Dietrich [101] compared the existing water table fluctuation method with their field measurements of evapotranspiration. They used 85 days of rain-free data from a weighable groundwater solution meter located in the wetland meadows of the Spreewald in northeastern Germany. It can be shown that some researchers have used multiple water table fluctuation methods in different study areas around the world and have continuously made targeted improvements. These case studies continue to promote the development and improvement of water table fluctuation methods.

However, there are still some uncertainties regarding the White method and its improvements [102]. These uncertainties are mainly in the quantification of specific yield (S_y), the choice of recharge time, the high heterogeneity of surface vegetation, and the effect of rainfall or irrigation on the water table. The largest uncertainty of White method mainly lies in the estimation of the S_y. The S_y was defined as the volume of water released by gravity from a unit area of rock column extending from the water table to the ground surface as the water table decreases by one unit depth (head) [103]. Errors in the S_y estimates translate directly into ET_g estimates [62,69,70,104]; both variables are involved in the ET_g calculation in the White method, since storage variability and groundwater recharge variability need to be multiplied by the S_y. However, there are many difficulties in estimating S_y, as this parameter is not constant over time. The specific yield is highly variable in shallow aquifers, and its magnitude depends on factors such as soil texture, water table depth, and the state of drainage or recharge [42,62,70,85,103,105–107]. S_y is usually used only for groundwater drainage, but for rising water table conditions, the presence of air within the pore space is likely to reduce the value of S_y, i.e., the equivalent change in water table corresponds to a different change in water volume for the recharge and drainage states [42].

In addition, the White method for calculating the recharge rate (the underlying assumption for r) uses the average rate of groundwater table rise between 0:00 a.m. and 04:00 a.m. for a total of 4 h to equal the average groundwater recharge rate for the entire day (Table 1); therefore, this method is not applicable to variable groundwater fluctuations. The recharge rate is typically a function of the head difference between the observation well and the recharge source [62]. The transient recharge rate will vary with time. Replacing the average daily recharge rate with the average daily recharge rate determined over a range of time can cause errors in the estimation of ET_g [106]. Therefore, many researchers usually modify the White method according to the time period used for recharge estimation [62,94,108]. Some other methods even avoid using varying recharge rates [13,86].

Table 1. The time periods selected for water table recovery rate calculation in different studies.

Method	Experimental Period
White [81]	0:00~4:00
Dolan et al. [96]	0:00~4:00
Hays [97]	0:00~4:00
Loheide [62]	0:00~6:00
Yin et al. [12]	the previous day 21:00~5:00
Rushton [108]	the previous day 18:00~6:00
Miller et al. [94]	the previous day 22:00~7:00

In addition to the degree of water availability and the timing of recharge, the magnitude of ET_g is also influenced by surface vegetation cover and the growth state of vegetation [109–111]. Of course, vegetation effects are not limited to groundwater evapotranspiration ET_g; for total surface evapotranspiration ET, the contribution of plant transpiration (T) to evapotranspiration (T/ET) is estimated to range from 38 to 77% at the global scale, with an average of 64 ± 13% [112–114]. Thus, terrestrial vegetation is an important force in the global water cycle [115]. In arid areas or areas with a low leaf area index (LAI), the average contribution of T to ET can reach 70% to 95% [113,116,117]. In arid or semi-arid zones, plants can only extract water from deeper soils or directly from the shallow water table for transpiration during the growing season due to the low water content of the topsoil.

Multiple indications suggest a strong correlation between the distribution of plant species and the depth of the groundwater table [118,119]. Vegetation has a strong adaptive capacity, especially in water-scarce ecosystems, to make full use of water in the unsaturated zone and aquifers through root growth [120,121]. Nepstad et al. [122] suggest that the semi-enclosed forests of the Brazilian Amazon rely on deep root systems to maintain a green canopy during the dry season. Evergreen forests can meet evapotranspiration requirements during droughts of up to 5 months by absorbing water from the soil at depths greater than 8 m. Maraux and Lafolie [123] found, in a maize-sorghum field in Nicaragua, that upward infiltration of water fluxes into the root zone reached 2 mm per day during drought, while actual evapotranspiration ranged from 2 to 4 mm per day. Kleidon and Heimann [124] similarly found that water uptake from deep soil or groundwater plays an important role in dry season transpiration in Amazonia. More recently, Saleska et al. [125] used Moderate Resolution Imaging Spectroradiometer (MODIS) satellite data to find that the greenness of Amazonian forests increased even during the 2005 drought, and concluded that trees were able to use deep roots to access groundwater to survive extreme drought periods. Considering the influence of crop growth processes on ET_g, Liu and Luo [126] combined two different methods, proposed by Doorenbos and Pruitt [59] and Schmid et al. [127], to calculate the values of groundwater evapotranspiration in areas with burial depths of less than 1.5 m for rainy periods (presence of rainfall or irrigation) and non-rainy periods, respectively. This method used a negative linear relationship between the water content of the unsaturated zone (root zone) and ET_g. However, when the groundwater depth is greater than 1.5 m, the variation in ET_g and the effect of irrigation can cause the relationship to deviate from the linear relationship. Under field conditions, the calculation of ET_g is influenced by multiple factors. Even during non-rainy periods (when there is no rainfall or irrigation) ET_g is still affected by multiple factors such as soil properties, crop water requirements, available soil water content, and groundwater table depth [128]. During rainy periods, a mixed upward and downward water potential gradient is formed in the soil profile [126], and downward fluxes caused by rainfall or irrigation may lead to a gradual development of local downward water potential gradients toward the bottom of the root zone. Some researchers, such as Yuan et al. [129], observed a significant positive correlation between evapotranspiration ET_g and potential evapotranspiration (PET). In addition, the results of Carlson Mazur [58] recognized a significant positive correlation between the two. Some studies have also reported that ET_g and potential PET have a weak positive correlation. According to some previous calculations, the R^2 range of ET_g and PET was 0.02–0.43 [130]. Lautz [79] also reported a similar correlation between groundwater evapotranspiration (ET_g) and potential evapotranspiration (PET) in semi-arid riverfront areas. Zhang et al. [131] reported that the strength of the correlation between ET_g and PET was significantly different in different research regions or conditions. The above complexity of the relationship between PET and ET_g is mainly due to their different controlling factors. Compared with PET, which is affected by climate conditions, there are many influencing factors for groundwater evapotranspiration. including plant species in the study area, root depth, local meteorological conditions, groundwater level variation patterns, lateral recharge and discharge of groundwater, and so on. Under the combined effect of these factors, groundwater evapotranspiration exhibits a large spatial and temporal heterogeneity.

The water table fluctuation (WTF) method, as a typical point-scale ET_g calculation method, is not a good choice in solving problems in terms of spatial heterogeneity. Due to some restrictive nature of its own application conditions and influenced by the uncertainty of the envelope, many researchers use numerical models as an auxiliary method to the water table fluctuation method to solve the problems of envelope uncertainty and spatial heterogeneity. Using an arid desert environment in northwestern China as his study area, Wang et al. [13] found that groundwater evapotranspiration is an important factor controlling hydrological processes in arid riparian zones, while the accuracy of estimating groundwater evapotranspiration values is influenced by the groundwater flow rate in the

aquifer, the redistribution of water within the riparian aquifer during river flow [132], and the specific yield [84]. He also used HYDRUS as a complement to understand the different seasonal variations in groundwater evapotranspiration values, but the point-scale HYDRUS+WTF approach still cannot address the uncertainties associated with time-varying lateral flow velocities, spatial variations in groundwater dynamic patterns, and specific yields in the riparian corridor, and he argued that monitoring networks rather than individual monitoring of point settings is necessary. Jia et al. [133] also used HYDRUS as a complement to the WTF method and found that in areas with shallow groundwater depths (<1 m), the groundwater replenishment of ET-induced depleted soils during nighttime is significant, and the use of the traditional White method would seriously underestimate groundwater evapotranspiration values because this factor is neglected; he therefore used HYDRUS to correct the omission of White method and used an improved method to estimate ET_g. Diouf et al. [134] also used HYDRUS and they found that in urban or shallow depth groundwater areas with agricultural irrigation, groundwater table fluctuation methods may be influenced by urban water use or agricultural irrigation, so in order to analyze the applicability of the WTF method to this area, he found that the WTF method achieved better accuracy in the evapotranspiration values calculated in the dry season using the water table simulated by HYDUS; however, the accuracy was not guaranteed for different surfaces. The accuracy could be guaranteed only under the vegetation type. Therefore, the feasibility of using the simulation results of the numerical model as the source data for the WTF method to calculate the evapotranspiration needs to be further demonstrated. By studying historical data on water content and pressure level fluctuations on soil profiles. Zhao et al. [135] found that in a study area in northwest China, direct groundwater recharge does not occur from mid-June to mid-September when evapotranspiration is high, and soil water content changes in the upper unsaturated zone are mainly controlled by atmospheric conditions; however, in deeper parts, they are controlled by pressure surface fluctuations. Accordingly, a one-dimensional model (HYDRUS-1D) with variable head for the lower boundary condition (BC) was developed to explain the response of groundwater flow to changes in atmospheric and groundwater conditions, and the lower BC model with variable head reproduced the observed variation in soil moisture content, with a much smaller total evapotranspiration value obtained for the variable head below BC compared with the fixed head corresponding to the mean water table depth, which is similar to Zhu et al.'s model results [136].

2.3. Using the Numerical Model to Calculate the Groundwater Evapotranspiration

Unlike HYDRUS, another commonly used groundwater flow numerical model, MODFLOW [137], is often used without the need to incorporate the WTF method because it has a well-established ET_g calculation package. The calculation of ET_g is based on the fact that potential evapotranspiration shows a linear variation depending on the depth of the water table, with ET_g reaching a maximum when the water table is near the surface [138]. When the water table is below a fixed depth (ultimate evapotranspiration depth), ET_g is zero. In addition, MODFLOW includes the segmented linear evapotranspiration (ETS) package, which considers the segmented linear relationship between the depth of groundwater burial and the ET_g rate [138], and the ETS package more accurately represents the numerical variation of ET_g at different burial depth stages [139]. In the ETS package, the relation of ET rate to head is conceptualized as a segmented line (a piecewise linear function) in the variable interval. The segments that determine the shape of the function in the variable interval are defined by intermediate points where adjacent segments join. The ends of the segments at the top and bottom of the variable interval are defined by the ET surface and the extinction depth. For the simplest case, where a single ET segment is used (equivalent to the EVT package in MODFLOW2005), the ET rate is calculated as:

$$\mathrm{RET_{nb}} = EVTR_{\mathrm{nb}},\ h_n > SURF_{nb} \tag{17}$$

$$RET_{\mathrm{nb}} = EVTR_{\mathrm{nb}} \frac{h_n - (SURF_{nb} - EXDP_{nb})}{EXDP_{nb}},\ (SURF_{nb} - EXDP_{nb}) \leq h_{\mathrm{n}} \leq SURF_{nb} \quad (18)$$

$$\mathrm{RET_{nb}} = 0,\ h_{\mathrm{n}} < (SURF_{nb} - EXDP_{nb}) \quad (19)$$

where RET_{nb} is the rate of loss per unit surface area of water table due to ET, in units of volume of water per unit area per unit time (LT^{-1}); hn is the head, or water-table elevation in cell n from which the ET occurs (L); $EVTR_{\mathrm{nb}}$ is the maximum possible value of RET (LT^{-1}); $SURF_{\mathrm{nb}}$ is the ET surface elevation, or the water table elevation at which this maximum value of ET loss occurs (L); and $EXDP_{\mathrm{nb}}$ is the cutoff or extinction depth (L), such that when the distance between h_{n} and $SURF_{\mathrm{nb}}$ exceeds $EXDP_{\mathrm{nb}}$, ET ceases.

Although the ETS package is an improvement in computational accuracy over the traditional EVT package, the calculations of the ETS package ignore other influences on $\mathrm{ET_g}$, such as plant canopy cover, leaf area, community composition, and water content in the air pocket.

Pozdniakov et al. [140] argues that groundwater vapor emission has an important influence on the water balance in river valley zones, and he uses MODFLOW with the RIV and EVT packages to calculate $\mathrm{ET_g}$. Based on the MODFLOW evapotranspiration package (EVT), Baird and Maddock [141] developed the Riparian Evapotranspiration for MODFLOW (RIP-ET) package to simulate riparian and wetland $\mathrm{ET_g}$ using a nonlinear $\mathrm{ET_g}$ curve that accounts for reduced $\mathrm{ET_g}$ rates due to hypoxic conditions. The spatial and temporal variability of riparian $\mathrm{ET_g}$ is controlled by climate, vegetation structure, water content of the unsaturated zone, and groundwater table depth (DTWT). To address the spatial heterogeneity of riparian zone vegetation $\mathrm{ET_g}$, Ajami et al. [142] implemented a GIS tool in conjunction with RIPGIS-NET, which creates data input files in the MODFLOW-2000 or RIP-ET packages and visualizes MODFLOW results. Combining RIP-ET in MODFLOW with the GIS functionality of RIPGIS-NET can be used to calculate $\mathrm{ET_g}$ at different scales. This relationship was later modified in the evapotranspiration (ETS1) package to a segmented linear relationship, to include the nonlinearity between DTWT and $\mathrm{ET_g}$ rates. El-Zehairy et al. [143] calculated groundwater evapotranspiration $\mathrm{ET_g}$ and unsaturated zone evapotranspiration $\mathrm{ET_{uz}}$ for reservoir areas with large stage fluctuations in the water table by using MODFLOW with the addition of the packages SFR7, UZF1, and LAK7. Sergey et al. [144] analyzed the role of groundwater evapotranspiration in the water balance by using the MODFLOW-2005 hydrogeological model with the STR package; the annual variation of recharge was obtained with the codes from Surfbal and HYDRUS. The program code SurfBal is a surface water balance model, which was developed to simulate the processes of precipitation and heat energy transformations on the land surface in order to generate the upper boundary meteorological conditions for HYDRUS. Hou et al. [145] used the remote sensing evapotranspiration data ETwatch with the UZF1 package and MODFLOW to calculate the region-scale $\mathrm{ET_g}$ of the shallow depth groundwater zone in the eastern North China Plain.

Some other numerical models have also become common in recent years to calculate groundwater evaporation $\mathrm{ET_g}$. Researchers conducting climate simulations have also started to calculate groundwater evapotranspiration [146–157]. Other researchers have combined numerical simulations with experiments to estimate groundwater evapotranspiration from agricultural fields [69,158]. Blin et al. [159] used evapotranspiration (ET) from the Earth Engine Evapotranspiration Flux (EEFlux) tool as calibration data, then used Parameter Estimation Software (PEST) as a tool to calibrate $\mathrm{ET_g}$ from MODFLOW for an undeveloped basin located in the arid Chilean highlands. Liu et al. [160] proposed an alternative approach to estimate ET through the lower boundary of the root zone (<1.0 m depth). Askri et al. [161] developed the hydrological model OASIS-MOD to study the effects of irrigation management on groundwater table fluctuations and soil salinity. The model involves evapotranspiration processes in the unsaturated zone, crop transpiration, and groundwater evapotranspiration processes.

3. Conclusions

Based on the above discussion, it is evident that accurate estimation of groundwater evapotranspiration is still a challenging task at present. The progress and breakthroughs in the technical methods related to this problem are mainly focused on two directions; the first is based on the theory of the WTF method, for which further improvement of the key parameters is required (such as specific yield S_y and groundwater recovery rate, etc.) There are various methods for improving the accuracy of the parameters, such as using numerical models or measuring in sample fields with the help of various instruments. It should be noted that the improvement of accuracy should not be limited only to the methods of obtaining parameters, and the discovery of new computational theories should not be ignored. The second type of method to calculate ET_g mainly relies on the numerical groundwater model. The distributed groundwater model can better solve the problem of spatial heterogeneity, and remote sensing technology can provide more accurate hydrological and meteorological parameters for the model, but the uncertainty of the unsaturated zone blocks the combination of remote sensing technology and the groundwater model, so the coupled model of surface remote sensing data and the unsaturated zone model and groundwater model, should subsequently be constructed for calculating the groundwater evapotranspiration at a regional scale.

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