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Estimation of evapotranspiration using diurnal groundwater level fluctuations: Comparison of different approaches with groundwater lysimeter data

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[1] In wetlands or riparian areas, water withdrawal by plants with access to groundwater or the capillary fringe often causes diurnal groundwater fluctuations. Various approaches use the characteristics of these fluctuations for estimation of daily groundwater evapotranspiration rates. The objective of this paper was to review the available methods, compare them with measured evapotranspiration and assess their recharge assumptions. For this purpose, we employed data of 85 rain-free days of a weighable groundwater lysimeter situated at a grassland site in the Spreewald wetland in north-east Germany. Measurements of hourly recharge and daily evapotranspiration rates were used to assess the different approaches. Our results showed that a maximum of 50% of the day to day variance of the daily evapotranspiration rates could be explained by the approaches based on groundwater fluctuations. Simple and more complex methods performed similarly. For some of the approaches, there were indications that erroneous assumptions compensated each other (e.g., when overestimated recharge counteracted underestimated storage change). We found that the usage of longer time spans resulted in improved estimates of the daily recharge rates and that the estimates were further enhanced by including two night averages. When derived from fitting estimates of recharge or evapotranspiration with according measurements the specific yield, needed to convert changes in water level to water volumes, differed considerably among the methods (from 0.022 to 0.064). Thus, the specific yield can be seen as "correction factor" that compensates for inadequate process descriptions.

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1. Introduction

[2] Sound water resource management in wetlands should be based on knowledge of prevailing processes and estimates of the water balance components. Evapotranspiration, in particular, is essential since often it is the main—sometimes almost the only [*Owen*, 1995] —water extracting constituent. A wide range of methods for estimating actual evapotranspiration or solely transpiration exist [*Drexler et al*., 2004], which typically differ in accuracy [*Allen et al*., 2011]. Often the equipment for estimating either is not only expensive but also only point measurements are provided. When remote sensing techniques are too costly, multiple sites for measurement may be necessary, especially to capture spatial heterogeneous areas, such as wetlands. Therefore, a cost-effective measurement approach is required.

[3] Diurnal groundwater fluctuations can be used for estimation of evapotranspiration at low cost. In environments, where plants have access to groundwater or the capillary fringe, e.g., riparian areas or wetlands with shallow groundwater levels, groundwater fluctuations can result from water extraction by phreatophytes. The characteristic diurnal pattern normally consists of a steep decline during the day and a recovery phase overnight. *Gribovszki et al*. [2010] showed in an elaborate review that this phenomenon has been frequently reported for decades.

[4] *White* [1932] used the diurnal pattern to set up a simple water balance for rain-free periods. The evapotranspiration is set equal to the sum of storage change plus groundwater recharge. The former is represented by the variation of the groundwater level, while the latter is derived from the water level behavior at night. Both quantities are multiplied by the specific yield (or drainable porosity), which serves as a conversion factor. Table 1 illustrates that the White method has been widely applied as well as modified. Entirely new approaches have been developed as well. Almost all are based on the same fundamental principle and encompass the three components: specific yield, recharge, and water level change. *Nachabe et al*. [2005] is the sole exception where instead of using groundwater

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Table 1. Review of Groundwater Fluctuation-Based Evapotranspiration Estimation Methods

Approach (and →Modifications)	Studies Applying the Approach
White method	*White* [1932], *Meyboom* [1964], *Tromble* [1977], *Farrington et al.* [1990], *Gerla* [1992], *Smithers et al.* [1995], *Owen* [1995], *Rosenberry and Winter* [1997], *Loheide et al.* [2005], *Butler et al.* [2007], *Shah et al.* [2007], *Lautz* [2008], *Martinet et al.* [2009], *Zhu et al.* [2011], and *Cheng et al.* [2013]
→ Correction factor for ET at night and plant density	*Gatewood* [1950]
→ No recharge considered	*Heikurainen* [1963] and *Schilling* [2007]
→ Determination of recharge not specified	*Lott and Hunt* [2001]
→ Recharge according to *Troxell* [1936]	*Tromble* [1977]
→ Alteration of periods used for recharge estimation (see Table 2)	*Rushton* [1996] and *Miller et al.* [2010]
→ Use of total soil moisture instead of groundwater levels (no specific yield required)	*Nachabe et al.* [2005]
→ Subtraction of groundwater trend of plot that was not affected by ET	*Bleby et al.* [1997]
→ Subtraction of regional groundwater fluctuations	*Engel et al.* [2005], *Nosetto et al.* [2007], and *Jobbagy et al.* [2011]
→ Using water table peaks of two subsequent days, recharge from night after ET	*Hays* [2003] and *Vincke and Thiry* [2008]
→ Subdaily ET estimates, intraday recharge by empirical and hydraulic approach	*Gribovszki et al.* [2008]
→ Subdaily ET estimates, intraday recharge related to detrended water table	*Loheide* [2008]
Dolan method (dual extrapolation of nighttime behavior)	*Dolan et al.* [1984]
Drawdown recharge method (using solely hydrograph extrema)	*Hays* [2003], *Hatler and Hart* [2009], and *Mould et al.* [2010]
Fourier-based method (using amplitude of sine function fitted to hydrograph)	*Soylu et al.* [2012]

level or specific yield they used changes in total soil moisture.

[5] For each method different assumptions are made, for example, how to select the time period to estimate recharge or how to decide which change in water level describes changes in water storage best. However, reliable data are often missing to verify the underlying theories. The final evapotranspiration estimates of newly proposed approaches were often only compared to the original White method [*Hays*, 2003; *Gribovszki et al.*, 2008], potential evapotranspiration [*Dolan et al.*, 1984; *Hays*, 2003; *Gribovszki et al.*, 2008; *Loheide*, 2008] and/or pan evaporation [*Dolan et al.*, 1984]. Verification with actual measured evapotranspiration was only performed for the methods of *White* [1932], appearing in several studies [e.g., *Gatewood*, 1950; *Farrington et al.*, 1990; *Engel et al.*, 2005], and of *Soylu et al.* [2012], who assessed their approach with eddy covariance measurements. Thus, most approaches are not verified within a study and furthermore, to our knowledge, a cross comparison of all methods together has not yet been performed.

[6] The objective of our study is therefore to evaluate the different evapotranspiration estimation approaches that use diurnal groundwater fluctuations, including their recharge assumptions. In addition, the role of the specific yield is examined. We employed data of a weighable groundwater lysimeter and compared them with the recharge and evapotranspiration estimates of the different methods. The water level in the lysimeter was permanently adjusted to a reference gauge, simulating the conditions of the surrounding area and providing measurements of evapotranspiration and all other water balance components with hourly resolution.

2. Methods for Estimating Evapotranspiration Based on Groundwater Fluctuations

[7] Besides the method by *White* [1932], we found five distinct approaches for evapotranspiration estimation based on groundwater fluctuations that were suitable for the available data and conditions. Estimates of recharge and evapotranspiration from six methods (complemented by modifications of the White method) were compared. All approaches are illustrated in Figure 1 and shortly described below. The original notations were kept to enable comparison with the corresponding sources. Since only two of the methods were named, the first authors name will be used in the following to distinguish between them. Moreover, wherever the term recharge will be used it will also encompass the possibility of being discharge under the corresponding conditions (e.g., after rainfall) without mentioning it separately.

2.1. White Method

[8] In his original work, *White* [1932] used groundwater fluctuations to calculate the volume of withdrawn water q [L/T]:

$$q = S_y(24r \pm s), \tag{1}$$

where S_y [L^3/L^3] is the specific yield, r [L/T] is the recharge rate calculated as the hourly rise of the water table between midnight and 4 A.M. (i.e., the slope of a fitted linear model), and s [L/T] is the net fall or rise of the water table in the 24 h period, i.e., from midnight to midnight (Figure 1). The quantity q is normally assumed to equate to the daily rate of evapotranspiration ET [L/T].

[9] The White method can be modified with regard to the period used for recharge estimation. For instance, *Miller et al.* [2010] derived r by employing the period from 10 P.M. at night to 7 A.M. of the next morning and averaging the calculated recharge rates of the previous night and the subsequent night. Since only the estimation of r is changed, equation (1) itself is not affected. Table 2 lists the modifications that were included in the investigation and the abbreviations used for them.

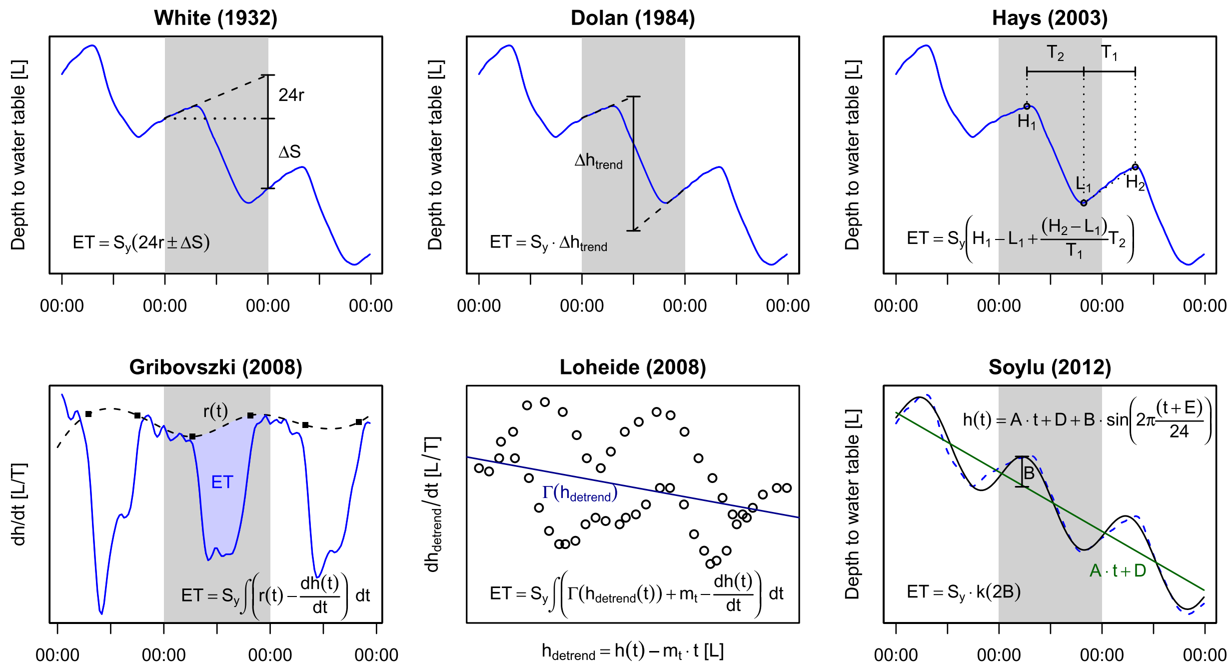

Figure 1. Illustration of the six methods for estimation of daily evapotranspiration *ET* based on groundwater fluctuation. The day under investigation is highlighted in gray. Note that the methods by Gribovszki and Loheide operate on a subdaily scale and have to be aggregated for obtaining daily values.

2.2. Dolan Method

[10] *Dolan et al.* [1984] manually extrapolated the nocturnal water table change forward to noon of the following day and backward to noon of the previous day (Figure 1). They hypothesized how the water level would behave, if no evapotranspiration would have occurred. The difference of the two noon water levels (multiplied with S_y) is attributed to *ET*. For our extrapolation, we used linear models based on hydrograph data between midnight and 4 A.M. of each night. When both fitted models pass through the according water levels at midnight, the method can also be expressed in form of equation (1) by setting r to the mean recharge of the two nights.

2.3. Hays Method (Drawdown Method)

[11] In his doctoral thesis, *Hays* [2003] proposed five different methods, using extrema of the hydrograph and the time spans between them. He recommend the Drawdown Method (Figure 1) for calculation of water use:

$$q = S_y\left[H_1 - L_1 + (H_2 - L_1)\frac{T_2}{T_1}\right], \tag{2}$$

where H_1 [L] is the maximum water level of the day under consideration, L_1 [L] the according minimum groundwater level, H_2 [L] the maximum of the following day, T_2 [L] the time interval between H_1 and L_1, and T_1 [L] the time span between L_1 and H_2. So the recharge rate $\frac{H_2 - L_1}{T_1}$ [L/T] is calculated in the subsequent night and evapotranspiration is just considered between H_1 and L_1.

[12] *Hays* [2003] also used the White method (equation (1)) with $s = H_2 - H_1$ and the recharge rate calculated from midnight to 4 A.M. in the following night.

2.4. Gribovszki Method

[13] Subdaily estimates of evapotranspiration can be calculated, when subdaily information on recharge rate is available. *Gribovszki et al.* [2008] estimated the rate by employing both a hydraulically derived and an empirical approach. Since the former needs hydraulic parameters that

Table 2. Variants of Daily Recharge Estimation Using the Slope of the Groundwater Hydrograph[a]

Basic Approach	Recharge Approach	Abbreviation	Start	End	Previous Night	Subsequent Night
White [1932]		White	0 A.M.	4 A.M.	x	
	Hays [2003]	White/Hays	0 A.M.	4 A.M.		x
	Loheide [2008]	White/Loheide	0 A.M.	6 A.M.	x	
	Rushton [1996]	White/Ruhston	6 P.M.*	6 A.M.	x	
	Miller et al. [2010]	White/Miller	10 P.M.*	7 A.M.	x	x
	Minimum RMSE	White/Minimum RMSE	6 P.M.*	6 A.M.	x	x
Dolan et al. [1984]		Dolan	0 A.M.	4 A.M.	x	x

[a]*Dolan et al.* [1984] originally did not use fixed time periods but interpolated the hydrograph manually. *Loheide* [2008] selected the period from midnight to 6 A.M. for recovery analysis but did not use the White method. "Minimum RMSE" was found to give the best recharge estimate concerning the root mean square error for our data (see section 5.2.2.). Times marked with an asterisk refer to the preceding day.

were not available at our study site, only the empirical method was used. Similar to *Troxell* [1936], the rate of the groundwater level change dh/dt [L/T] served as basis for deriving the transient recharge rate. For each day, maximum and minimum recharge rates were obtained by selecting the largest (positive) rate of the water level change and the mean of dh/dt between midnight and 6 A.M., respectively. These values were assigned to the time of the maximum and minimum groundwater level, respectively. Hence, for each day two points (knots) were defined. Based on the points of several days a spline interpolation was done, which describes the time-dependent recharge rate $r(t)$ [L/T]. Subdaily evapotranspiration (e.g., for $dt = 1h$) is then calculated as:

$$ET(t) = S_y\left(r(t) - \frac{dh}{dt}\right). \tag{3}$$

[14] Daily evapotranspiration can be calculated from the integral over time. As suggested by *Loheide* [2008], dh/dt at a specific time was determined by using the slope of a linear model fitted to three points (i.e., using the data at a specific time as well as one point before and after that time). We tested the inclusion of more data points, too. Also, a variant with a piecewise linear function (i.e., connecting the given knots linearly) as means of interpolation was applied (as suggested by Z. Gribovszki, personal communication, 2013).

2.5. Loheide Method

[15] The approach by *Loheide* [2008] differs from the Gribovszki method in terms of calculation of the recharge rate $r(t)$ in equation (3). It is assumed that the recovery rate is a function of the head difference between the observation well and the recovery source. The water level of the recovery source can be considered constant after detrending the hydrograph, suggesting that water levels of both locations change equally in the long term (i.e., their trend slopes m_t [L/T] are identical). Recovery is then only dependent on the measured groundwater table and can be calculated as:

$$r(t) = \Gamma(h_{detrend}) + m_t. \tag{4}$$

[16] The function $\Gamma(h_{detrend})$ [L/T] is a linear model describing the relationship between detrended groundwater level $h_{detrend}$ [L] and the change of detrended groundwater level over time $dh_{detrend}/dt$ (Figure 1), set up using data between midnight and 6 A.M. of two subsequent days. As for the Gribovszki method, $dh_{detrend}/dt$ was estimated as the slope of a fitted linear model.

2.6. Soylu Method (Fourier-Based Method)

[17] The Fourier-based method (Figure 1) takes the full diurnal cycle into account, but assumes the recovery rate, which is not explicitly determined, to be constant over the day. Following *Czikowsky and Fitzjarrald* [2004], *Soylu et al*. [2012] applied a sine function with a period of 24 h along a moving, multiday (i.e., 1–7 days) window of detrended groundwater level by fitting the water level time series to the function $h(t)$ [L]:

$$h(t) = A \cdot t + D + B \cdot sin\left(2\pi\frac{(t+E)}{24}\right), \tag{5}$$

where A [L/T] is the multiday trend, t [T] the time in hours, D [L] the mean bias, E [T] the diurnal signal phase, and B [L] the diurnal amplitude. Similar to the Hays approach, the daily evapotranspiration is estimated as function of the range between peak and trough of the day ($2B$):

$$ET = S_y \cdot k(2B). \tag{6}$$

[18] Since the amplitude B is affected by recharge and detrending, a scaling factor k is employed. It was found to be a function of the shape and the duration of the diurnal transpiration curve which depends largely on solar radiation. Hence, for the calculation of k, the sum of the solar radiation of the particular day is divided by the range of detrended solar radiation (i.e., differences of accumulated solar radiation and its 24 h moving average) of the same day. Alternatively, k can be calculated based on theoretical clear-sky radiation or can be set to the mean scaling factor of 1.9 [*Soylu et al*., 2012].

3. Data Set

3.1. Study Site

[19] The lysimeter is located in the Spreewald wetland, approximately 85 km southeast of Berlin, Germany (latitude 51°52′45″N, longitude 14°02′24″E). Measurements of the German Meteorological Service (Deutscher Wetterdienst (DWD)) at the weather station Lübben-Blumenfeld (51°55′N, 13°52′E) state an annual average temperature of 9.4°C and 570 mm of precipitation between 1981 and 2010. For the growing season, from April to October, values of 14.5°C and 355 mm were measured, respectively. The surrounding grassland site is subject to grazing during most of the vegetation period, with one additional cut taking place in June. Previous land use has led to degradation of the peat soils, resulting in a shallow peat layer over sand. The groundwater table is regulated via ditches and weirs that can be used for drainage or subirrigation depending on the current hydrological conditions.

3.2. Groundwater Lysimeter

[20] A weighable, monolithic groundwater lysimeter (2 m deep, 1 m^2 area) was used to determine the water balance of the grassland site (Figure 2). The water level in the lysimeter was adjusted via a compensation tank (diameter = 21 cm), acting as communicating vessel. Every hour the water level in the compensation tank was compared to the groundwater table measured at the nearby reference gauge and set to the latter if a threshold was exceeded. Net flux Q_n [L/T] to or from the lysimeter was derived from the water level change in the tank (by multiplication with the ratio of tank area to lysimeter area). A decreasing tank water table, for example, indicated an inflow from the tank to the lysimeter (i.e., recharge). A more in-depth description of the functioning can be found in *Bethge-Steffens et al*. [2004]. Storage change dS [L/T] was derived from mass change, recorded by three loading cells, and precipitation P [L/T] was derived from mass gain during rainfall periods, which were observed by a Hellmann rain gauge at

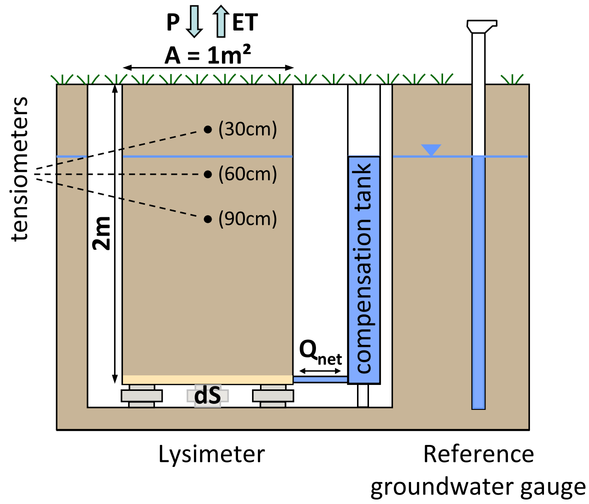

Figure 2. Schematic diagram of the groundwater lysimeter.

the site. Finally the evapotranspiration rate ET [L/T] was given by:

$$ET = P + Q_n + dS. \tag{7}$$

[21] All measurements, except the net flux ($\Delta t = 1h$), were conducted at 10 min intervals. For determination of the depth to groundwater in the lysimeter, the tensiometer (Tensio 151, UGT; accuracy: 0.3 kPa) in 60 cm depth was used. It was submerged nearly all the time, thus measuring the piezometric head. Throughout this study, we used hourly groundwater data, based on the 10 min measurements after applying a seven point moving average filter.

3.3. Period of Investigation and Data Preparation

[22] The evapotranspiration estimates were only calculated for rain-free conditions. Different studies also used days with (minor) precipitation, but distinct conditions prevailed (e.g., high interception storage [*Gribovszki et al.*, 2008] or high depth to groundwater in combination with deep rooting trees [*Miller et al.*, 2010]). For this study, rain-free periods longer than 120 h were selected from the vegetation periods (April to October) 2011 and 2012. After visualization of the corresponding hydrographs only those with pronounced diurnal fluctuations were chosen and truncated in order to start and end at midnight. In total 15 periods, ranging from 4 to 10 days, were selected. Days without properly logged recharge values, high negative hourly evapotranspiration values during daytime or groundwater data strongly deviant from the situation at the reference gauge were removed, since rigorous data filtering is a prerequisite for obtaining reliable estimates [*Mould et al.*, 2010]. In total, 85 days were left for investigation. Further, resulting from measurement errors of the water level changes in the compensation tank that were caused by remote data transfer (and subsequent voltage drop), the recharge values logged at 6 A.M. and at 7 A.M. were biased. Since a share of the recharge of the period from 7 A.M. to 8 A.M. was attributed to the period from 6 A.M. to 7 A.M., correction was done by replacing the individual values by the mean of the two periods.

4. Methodology for Evaluation of Evapotranspiration Estimates With Groundwater Lysimeter Data

4.1. Determination of Specific Yield

[23] The specific yield S_y [L^3/L^3] is of crucial importance whenever groundwater level changes are used for estimation of water budget components, e.g., storage change, groundwater recharge, or evapotranspiration. Since S_y is used as a factor in every approach under study (Figure 1), errors in specific yield estimation directly translate to the ET estimates [*Mould et al.*, 2010].

[24] The specific yield is defined as the volume of water extracted from the groundwater per unit area when the water table is lowered a unit distance [*Hillel*, 1998]. For deep groundwater conditions, a constant value for a defined porous medium is often assumed. This implies that the volume released is proportional to the water level change [*Nachabe*, 2002]. However, this does not apply for shallow water tables, when the soil moisture profile is truncated at the soil surface [*Childs*, 1960]. To account for the evolving depth dependence, the term apparent specific yield was used by *Duke* [1972].

[25] Further, a constant specific yield requires complete and instantaneous drainage at all points above the water table [*Duke*, 1972]. Yet as the water is released from the unsaturated zone, its transient nature [*Nachabe*, 2002], i.e., the time given for drainage, is of importance. Restriction of time, as in the case of evapotranspiration, can be expressed by the terms readily available yield [*Meyboom*, 1964] or transient specific yield [*Nachabe*, 2002] in contrast to the ultimate specific yield [*Duke*, 1972] that is reached after drainage has ceased. Moreover, S_y is affected by antecedent moisture conditions [*Loheide et al.*, 2005] and plant activity [*Logsdon et al.*, 2010]. As hysteresis can be an influencing factor as well [*Nachabe*, 2002], specific yield may be different for falling and rising water levels, which can be expressed by using the terms drainable and fillable porosity, respectively.

[26] In this article, the original term specific yield is kept, but defined (extending a definition by *Duke* [1972]) as the change in soil water volume per unit area per unit change of groundwater level during a specified time window under the specified conditions. A water budget approach [*Shah and Ross*, 2009; *Logsdon et al.*, 2010] was used to capture a maximum of influencing factors and processes. Referring to *Healy and Cook* [2002] the specific yield can be calculated as:

$$S_{y,wb} = dS\left(\frac{\Delta h}{\Delta t}\right)^{-1}, \tag{8}$$

where Δh [L] is the change in groundwater level, Δt [T] the time between the measurements, and dS [L/T] the time rate of water storage change in the lysimeter. The latter accounts for the water release from the entire monolith. Consequently, water extraction by plants from the unsaturated zone is included as well, although it may affect the groundwater table only to a certain degree [*Loheide et al.*, 2005]. Because the specific yield should be representative for the time scale of interest [*Butler et al*., 2007], i.e., for the daily period of evapotranspiration, the time span between 6 A.M. and 6 P.M. was chosen. This time window nearly covers the complete time frame in which evapotranspiration is taking place and furthermore, a period of 12 h was in accordance with *Gatewood* [1950] and *Loheide et al*. [2005].

4.2. Examination of Groundwater Recharge Assumptions

[27] For estimation of evapotranspiration based on groundwater measurements, the recharge is required. Since complexity prevents the derivation of a general recharge function [*Loheide*, 2008], all six methods under investigation have their unique approaches. Commonly the nighttime behavior of the hydrograph is used, since it is assumed to be only a function of recovery and *ET* is considered to be negligible at that time. Some approaches [*White*, 1932; *Dolan et al*., 1984; *Hays*, 2003] use constant recharge rates and are therefore seeking average values that are representative for certain periods (e.g., 1 day for White's method, time between peak and trough of hydrograph for Hays' method). Other methods [*Gribovszki et al*., 2008; *Loheide*, 2008] assume changing rates due to varying gradients between recovery source and groundwater table.

[28] We evaluated the different recharge estimation approaches using hourly in/outflow data (Q_n) from the lysimeter. We assumed that the interactions between the compensation tank and the lysimeter (Figure 2) resemble the recharge conditions found at the field site, i.e., $r = Q_n$. Since the lysimeter groundwater level was adjusted to that of the reference gauge, the recharge could be measured without knowledge of the prevailing source(s) of recovery (e.g., upwelling groundwater or lateral flow from ditches).

[29] To reveal diurnal characteristics, we employed boxplots of each hour over the day. Then, the measured hourly recharge data was used for assessing the different methods. As the groundwater record can be perceived as accumulation of inflow and transpiration rates [*Troxell*, 1936], the measured recharge rates were also accumulated to use the estimation procedures in an unaltered form. The approaches by Soylu and Loheide were left out. The former does not account for recharge separately. The Loheide method bases its recharge estimation on a detrending procedure of the groundwater level, whose underlying assumptions cannot be applied to measured recharge data. However, the modifications of the White method were taken into account (Table 2).

[30] To find the best time period to estimate recharge, we tested all possible options with starting times from 6 P.M. to midnight of the preceding day, maximal ending time at 6 A.M. and a duration of at least 2 h. Additionally, for all these variants also the average of two nights (the previous and the subsequent night) were calculated. In total, only 76 days were included because some methods needed data for days prior or posterior to the selected day, which were not always available.

[31] Finally, recharge was estimated based on the measured groundwater level. For that purpose, the approach of *Loheide* [2008] was included and all 85 days were used. Due to the uncertainty of the specific yield derived from the water balance method ($S_{y,wb}$), we assessed the approaches without the incorporation of S_y. Instead, the specific yield $S_{y,rfit}$ which was necessary to fit estimates and measurements was derived and then compared to $S_{y,wb}$. It was given by

$$S_{y,rfit} = \frac{r_{lysimeter}}{r_{est}}, \tag{9}$$

where $r_{lysimeter}$ is the measured recharge and r_{est} is the normalized recharge estimate. The latter was given by dividing the recharge estimate by S_y. For the White method the first addend in equation (1), i.e., $S_y \cdot 24r$, corresponds to the estimated recharge. Hence, r_{est} is $24r$ for this approach. For all other methods, r_{est} can be derived accordingly.

4.3. Examination of Evapotranspiration Estimates

[32] Typically, the evapotranspiration estimates derived from groundwater level fluctuations are considered to account only for water extraction from the saturated zone, i.e., giving the share of transpiration that is sustained by groundwater. However, numerical simulations by *Loheide et al*. [2005] suggested that direct extraction from the capillary fringe and indirect extraction resulting from upward flow to the unsaturated zone in short distance above the capillary fringe are also represented. Hence, in wetlands with shallow water tables, the saturated and the unsaturated zone can be conceived as a single system [*Mould et al*., 2010]. Therefore, evapotranspiration measured in the groundwater lysimeter (which included all possible components of evapotranspiration) could be used as means of comparison under the prevailing conditions.

[33] The six different approaches (including various modifications of the White method) were applied for the selected 85 days. As in the case of recharge, the evapotranspiration estimates were also assessed using normalized values (ET_{GW}/S_y). So for the White method this quantity arises from dividing equation (1) by S_y and is given as $24r \pm s$. The relationship that evolves from plotting normalized estimates and measured evapotranspiration ($ET_{lysimeter}$) against each other represents the specific yield [analogous to *Engel et al*., 2005]. The relationship is linear when the specific yield is constant, and in this case the specific yield can be derived from the slope m of a fitted linear model. The slope was calculated as:

$$m = \frac{ET_{GW}/S_y}{ET_{lysimeter}}. \tag{10}$$

[34] The equation simplifies to $m = 1/S_y$ when the estimates fitted the observations ($ET_{GW} = ET_{lysimeter}$). This means that the inverse of the slope ($1/m$) is the specific yield that fitted observations and estimates ($S_{y,ETfit}$).

[35] Other kinds of relationship (i.e., nonlinear) would evolve if the specific yield was not constant (e.g., dependent on groundwater depth) or other assumptions were not valid (e.g., when a major share of evapotranspiration was not reflected by ET_{GW}). To test if specific yield was a function of the groundwater level, $S_{y,ETfit}$ was calculated for each day (similar to equation (9) the measured value was divided by the normalized estimate) and plotted against the mean depth to groundwater table of the corresponding day.

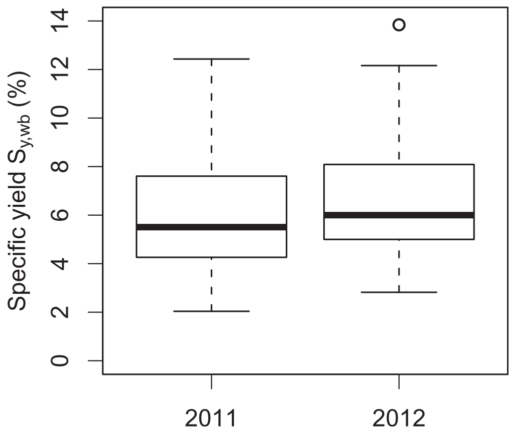

Figure 3. Specific yield $S_{y,wb}$ calculated by the water balance approach (equation (8)).

[36] Parameters had to be set for three of the evapotranspiration estimation approaches. For the methods by Loheide and Gribovszki, the slope of a fitted linear trend was used to estimate time rates of change of the (detrended) water table. We tested 3–13 data points for fitting the model. In case of Gribvoszki's approach, both piecewise linear and spline interpolation of recharge rates were assessed. We used Soylu's method with 1, 3, and 5 day windows and furthermore, we analyzed variants with calculated and standard scaling factors.

[37] Finally, for assessing potential errors involved in the different estimation methods, the root mean square error (RMSE) between evapotranspiration estimates and measured values was calculated. The specific yield gained from the water balance approach ($S_{y,wb}$) as well as the specific yield to fit evapotranspiration estimates and measurements ($S_{y,ETfit}$) were used for this purpose.

5. Results

5.1. Specific Yield Calculations

[38] An average specific yield $S_{y,wb}$ of 0.063 (median of 0.060) was obtained from the water balance approach (equation (8)) for 85 days of lysimeter data. Yet range and variation of the specific yield estimates were high. The first and third quartile were 0.044 and 0.079, respectively. No significant relationship between the specific yield and groundwater level was found. Therefore, dispersion is likely to be caused by a mixture of influencing factors. In Figure 3, only small differences between the 2 years can be seen, indicating that conditions were not transient. However, values of 2012 (mean: 0.067, median: 0.060) slightly exceeded those of 2011 (mean: 0.060, median: 0.055). In the following, a constant specific yield of 0.063 will serve as comparative value, however the value is highly uncertain.

5.2. Testing Groundwater Recovery Assumptions

5.2.1. Diurnal Characteristics of Recharge

[39] The measured recharge (i.e., the inflow to the lysimeter) showed notable variations (Figure 4). When comparing the hourly mean and the average rate (i.e., the uniformly distributed recharge) it can be seen that between 1 A.M. and 3 P.M. the hourly recharge is less than average. On the other hand, for the time span between 5 P.M. and midnight the inflow to the lysimeter was considerably higher than for the rest of the day. This coincides with the assumption that groundwater levels lowered by evapotranspiration cause higher gradients which should normally provoke higher recharge, even though the underlying processes are subject to nonlinearities [*Loheide*, 2008]. A clear diurnal variation was not found. This was probably caused by the limited resolution of the recharge measurements, especially since the amounts of hourly recharge were small.

5.2.2. Recharge Estimates Based on Measured Recharge

[40] Most methods were able to estimate daily recharge reasonably well when we applied hourly measured recharge data. As indicated in Figure 5, from the established methods Gribovszki's (employing linear interpolation) as well as Miller's and Rushton's adaptations of the White method resulted in the highest coefficients of determination R^2 (0.88, 0.84, and 0.84, respectively). With respect to the root mean square error (RMSE) the approaches of Miller, Gribovszki, and Dolan performed best (0.48, 0.54, and 0.54

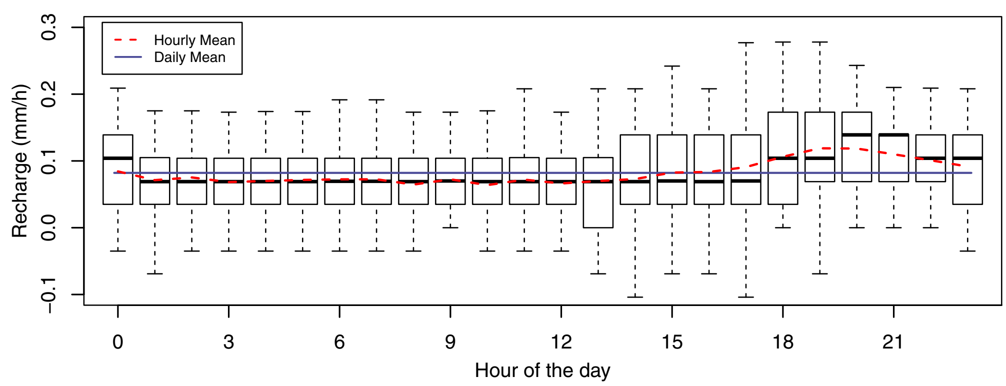

Figure 4. Measured flow to/from the lysimeter. Outliers (i.e., values lying outside of 1.5 times the interquartile range) are not shown.

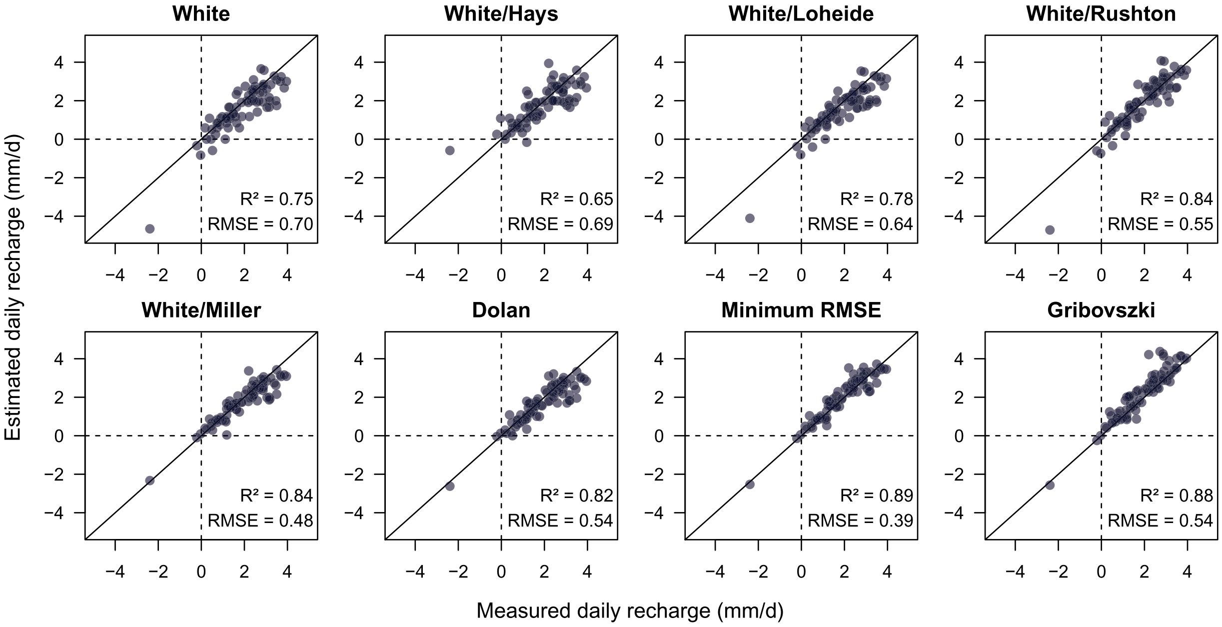

Figure 5. Comparison of daily recharge against estimates calculated based on measured hourly recharge data. The one-to-one line is shown in each plot. Root mean square errors (RMSE) are in millimetre per day. All correlations, expressed by the coefficient of determination R^2, are significant ($p < 0.05$).

mm/d, respectively). In summary, methods that considered gradient-dependent recharge (Gribovszki), two nights average (Miller and Dolan) or a very long time span (Rushton) showed the best results. In contrast, deviations between estimated and measured recharge were highest only when a short period of one night was used.

[41] These findings were confirmed when all possible recharge periods were tested for their performance. Superior results were achieved when the mean of two nights was used—both in regard to RMSE and R^2. The lowest RMSE was achieved using two night periods from 6 P.M. to 6 A.M. (data shown in Figure 5 as "Minimum RMSE"). The same time span was proposed by *Rushton* [1996], yet only using one night's data. The highest R^2 (0.92) was achieved when employing the two night average of recharge data between 6 P.M. and 1 A.M., but due to the high RMSE (0.72 mm/d) this time window is not considered hereafter.

[42] For Hays' method (not shown, RMSE = 0.54 mm/d, $R^2 = 0.47$), the recharge is measured only for the time between maximum and minimum of the groundwater hydrograph, which was on average 13 h. Since the calculated recharge only reflects this part of the day, the RMSE could not be compared directly to the other methods, however, the correlation was the lowest of all methods.

5.2.3. Recharge Estimates Based on Groundwater Data

[43] Normalized daily recharge estimates (r_{est}) and values of $S_{y,rfit}$ (i.e., required specific yield to fit estimates and measurements) differed considerably among the approaches when applied to groundwater data. Median values of both entities (Table 3) characterized the recharge patterns better than the mean values. In some cases the mean values, especially of $S_{y,rfit}$, were heavily influenced by larger values, which resulted from r_{est} values, appearing as denominator in equation (9), close to zero.

[44] Median values of recharge normalized by the specific yield ranged from 37.8 mm/d (Loheide's modification of the White method) to 77.5 mm/d (Loheide method). The approach by Hays showed the lowest value, yet again it only represents the share of the day for which evapotranspiration was assumed to occur. Also, median values of $S_{y,rfit}$ differed up to 70%, ranging from 0.022 to 0.039. These values were much lower than the specific yield estimated by the water balance approach ($S_{y,wb} = 0.063$). Assuming that $S_{y,wb}$ was the appropriate specific yield, all methods would have substantially overestimated the recharge. For example, Gribovszki's approach would have estimated a mean recharge of 4.9 mm/d, although only 2.0 mm/d were observed. An often assumed relation between specific yield

Table 3. Results of Recharge Estimation Based on Groundwater Data[a]

Method	Median of Rest $\left(\frac{mm}{d}\right)$	Median of $S_{y,rfit}$ $\left(\frac{m^3}{m^3}\right)$
White	49.7	0.033
White/Hays	61.0	0.03
White/Loheide	37.8	0.039
White/Rushton	47.2	0.033
White/Miller	44.7	0.035
White/Minimum RMSE	56.6	0.032
Dolan	54.9	0.032
Hays	32.3	0.025
Gribovszki	76.4	0.025
Loheide	77.5	0.022

[a]Median values of normalized recharge estimates (r_{est}) and specific yield required to fit estimates with measurements ($S_{y,rfit}$).

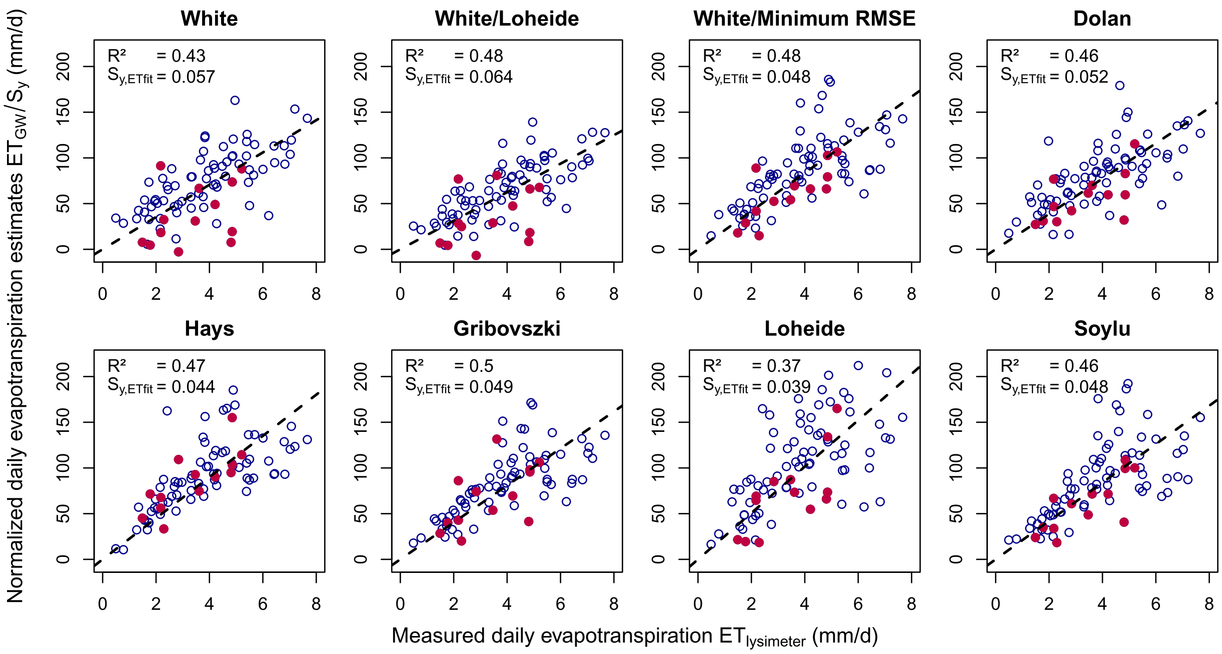

Figure 6. Comparison of normalized daily evapotranspiration estimates with values measured by the lysimeter for different methods. Filled points highlight days which followed days with rain. The given specific yield ($S_{y,ETfit}$) is the value that fits estimates and measured ET data. All correlations, expressed by the coefficient of determination R^2, are significant ($p < 0.05$).

and groundwater level was not the cause for these deviations, since no significant (linear) correlation ($p < 0.05$) between $S_{y,rfit}$ and groundwater level and no consistent depth profile (data not shown) could be found.

5.3. Evapotranspiration Estimates Based on Groundwater Fluctuation

5.3.1. Comparison of Normalized Evapotranspiration Estimates With Lysimeter Measurements

[45] Differences between most of the evapotranspiration estimation approaches were rather small with respect to the coefficient of determination R^2 (Figure 6). However, the specific yield that fits estimates and measurements differed notably as do the deviation patterns.

[46] The best results were achieved using the method of Gribovszki. Other approaches performed equally well. Only the Loheide method for subdaily evapotranspiration estimation and the original White method performed poorly. In both cases the underlying recharge assumptions were suspected to cause the poor results. For the Loheide approach, the supposed relationship between detrended water table depth and time rate of change of detrended water table was very weak in our case. The mean correlation of the two quantities was very small ($R^2 = 0.14$). This probably caused the poor evapotranspiration estimates, since the equations of Gribovszki, whose method performed best, and of Loheide just differ in terms of $r(t)$ (see equation (3) and (4)). For the White method, the period used for recharge calculation was the crucial point, as the performance could be improved notably by its modification. The two best variants (White/Loheide and White/Minimum RMSE) are shown in Figure 6. Also, the adaptation of Miller improved the performance ($R^2 = 0.46$, $S_{y,ETfit} = 0.055$) while the modifications of Rushton and Hays decreased the performance ($R^2 = 0.42$, $S_{y,ETfit} = 0.052$, and $R^2 = 0.39$, $S_{y,ETfit} = 0.051$, respectively).

[47] The performance of the tested approaches depended on the evapotranspiration rate (Figure 6). The estimates of the methods of Hays, Soylu, White/Minimum RMSE, and Gribovszki scattered when measured evapotranspiration was large (i.e., higher than 4 mm/d), while they performed substantially better on days with low evapotranspiration rates. In contrast, the White method (including the modification according to Loheide) and the Dolan method were able to reproduce estimates of evapotranspiration at higher values, although deviations occurred at lower ET rates.

[48] Concerning days after precipitation, i.e., conditions where at least parts of the day can be subject to drainage, the White approach (as well as its modification by Loheide) clearly underestimated evapotranspiration (Figure 6). The negative recharge rate (i.e., discharge of groundwater prevailed) derived from the beginning of the day was subject to change and not representative for the day (see also Figure 5). Hence, most of the day's groundwater lowering was attributed to (overestimated) discharge, leading to underestimation of evapotranspiration. All other approaches performed considerably well after days with rain. Only Hays' Drawdown Method tended to overestimate evapotranspiration sometimes, as it was not designed to reproduce discharge conditions. Drainage leads to a drop of groundwater level after evapotranspiration has already ceased. This causes problems when hydrograph extrema are used, since the low of the current day and the high of the next day are reached around midnight and are almost identical. So the recharge rate $(H_2 - L_1)/T_1$ (equation (2)) will yield (negative) values close to zero, generating recharge underestimation. Furthermore, the recharge can be affected easily by noise since the time T_1 between the two extrema is very

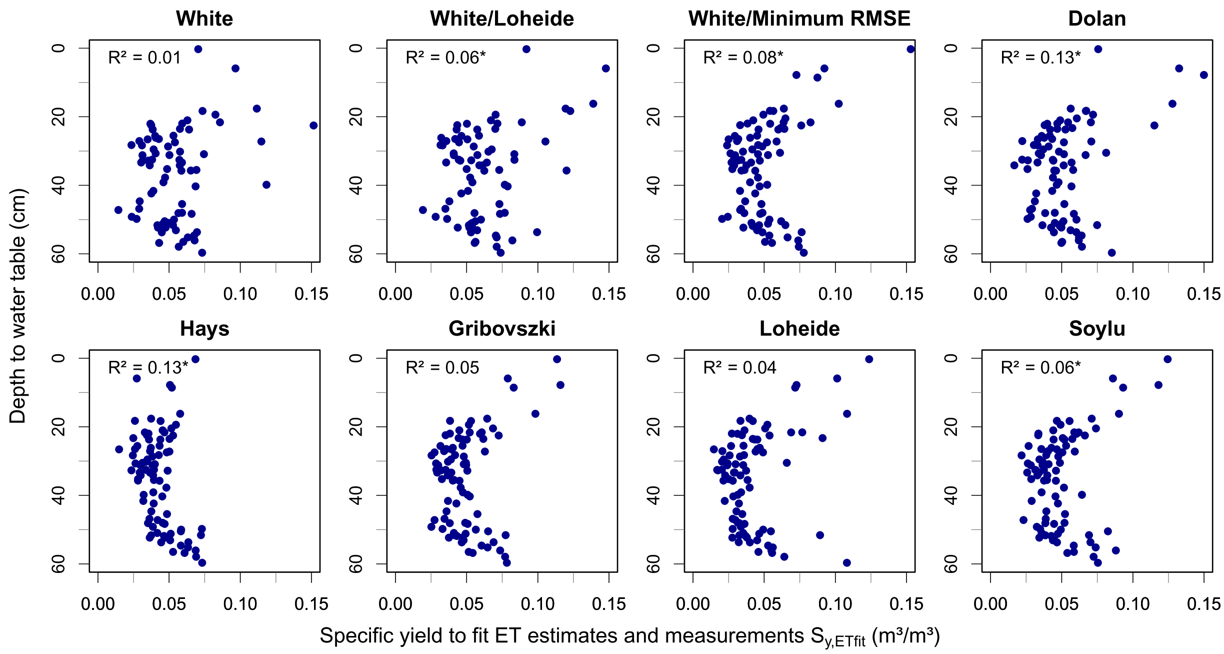

Figure 7. Comparison of $S_{y,ETfit}$ calculated for each day under investigation to mean groundwater depth of the corresponding day. Some outliers are not shown as limits of the axes were kept constant. All correlations, expressed by the coefficient of determination R^2, marked with an asterisk are significant ($p < 0.05$).

small. Despite this apparent source of error the Hays method estimated the evapotranspiration fairly well for most rain influenced days of our data set. Probably the incorrect discharge estimation was compensated by other effects.

5.3.2. Relation Between Groundwater Level and Specific Yield Needed to Fit Estimates and Measurements

[49] The values of the specific yield needed to fit evapotranspiration estimates and measurements ($S_{y,ETfit}$) showed a significant ($p < 0.05$) linear correlation to groundwater depth for five approaches (Figure 7). This was in contrast to recharge, where the values of $S_{y,rfit}$ were not correlated to the water level at all. Yet even for $S_{y,ETfit}$ the obtained (linear) correlation was low (R^2 was 0.14 at maximum). However, patterns in Figure 7 suggested a nonlinear relation, as for most of the approaches a crescent-like shape with high values of $S_{y,ETfit}$ at high and low groundwater levels was observed. However, this shape did not necessarily reflect characteristics of the soil layer. Rather, it appeared that high specific yield values for groundwater levels close to the surface resulted from underestimation of evapotranspiration under drainage or low recharge conditions (as discussed in the preceding subsection). Consequently the method by Hays, which estimated *ET* under drainage conditions well, did not show this pattern. Nevertheless, a trend for larger values of $S_{y,ETfit}$ was observed when groundwater levels dropped below 50 cm.

5.3.3. Effect of Parameter Modification

[50] For the methods of Gribovszki and Loheide, extensive smoothing proved to enhance the performance. Best results concerning R^2 were achieved when nine points for determining the slopes of $dh_{detrend}/dt$ (Loheide) and dh/dt (both approaches) were used. In other words, 4 h before and after the respective time were used. The coefficient of determination between estimated and measured evapotranspiration increased from 0.4 to 0.5 (Gribovszki) and 0.33 to 0.37 (Loheide) when we implemented nine points instead of three points. No matter how many points for slope determination were used, Gribovszki's approach always performed better regarding R^2 when linear (in contrast to spline) interpolation was applied. Yet the best results required considerable less smoothing when we used splines (five points for slope determination, $R^2 = 0.45$).

[51] Fitting the sine function to just 1 day of data provided the best solution for the Fourier-based method by *Soylu et al.* [2012]. Accompanied with the superior fit of the sine function to the hydrograph (R^2 were 0.98, 0.82, and 0.72 for windows of 1, 3, and 5 days, respectively), 1 day windows also showed the best results concerning the correlation between ET estimates and measurements (R^2 of 0.46, in contrast to 0.22 and 0.23 for 3 and 5 days). Furthermore, the usage of measured solar radiation for derivation of the scaling factor (mean value was 1.97) proved to be useful since performance deteriorated from 0.46 to 0.39 when a constant factor of 1.9 was applied.

5.3.4. Comparison of Nonnormalized Evapotranspiration Estimates With Lysimeter Measurements

[52] When we used $S_{y,wb}$, the methods by White/Loheide, White/Miller, and White showed the smallest deviations between estimates and measurements (Table 4). This was expected since their $S_{y,ETfit}$ values (0.064, 0.055, and 0.057, respectively) were close to the average $S_{y,wb}$ (0.063). Conversely, methods with a low $S_{y,ETfit}$ (e.g., the methods of Loheide, Hays, and Soylu) performed worse.

Table 4. Root Mean Square Errors (RMSE) of Evapotranspiration Estimates[a]

Approach	$S_{y,ETfit}$ $\left(\frac{m^3}{m^3}\right)$	RMSE for $S_{y,wb}$ $\left(\frac{mm}{d}\right)$	RMSE for $S_{y,ETfit}$ $\left(\frac{mm}{d}\right)$
White	0.057	1.83	1.59
White/Hays	0.051	2.14	1.51
White/Loheide	0.064	1.50	1.53
White/Rushton	0.052	2.21	1.70
White/Miller	0.055	1.76	1.45
White/Minimum RMSE	0.048	2.25	1.39
Dolan	0.052	1.93	1.40
Hays	0.044	2.56	1.33
Gribovszki	0.049	2.04	1.32
Loheide	0.039	3.58	1.61
Soylu	0.048	2.39	1.49

[a]RMSE are calculated using the mean specific yield derived from the water balance approach ($S_{y,wb} = 0.063$) and from the linear model which fitted normalized ET estimates and measurement best ($S_{y,ETfit}$).

However, the specific yield gained from the water balance approach was uncertain. The $S_{y,ETfit}$ values were almost as likely to represent the conditions in the lysimeter, as nearly all values of $S_{y,ETfit}$ fell in the range of first quartile (0.044) and median (0.06) of $S_{y,wb}$.

[53] When $S_{y,ETfit}$ of the corresponding approaches were applied, differences regarding RMSE declined. The lowest RMSE values were achieved for the methods of Gribovszki and Hays. Still, even the lowest RMSE exceeded 1.3 mm/d. So compared to the mean evapotranspiration of the 85 days (3.8 mm/d), the error produced by the different estimation techniques was relatively high. In addition the $S_{y,ETfit}$ gained from the slope of the fitted linear trend did not necessarily secure a minimum RMSE (e.g., the approach of White/Loheide behaved slightly better with a specific yield of 0.063). This was probably caused by forcing the linear model to pass through the point of origin.

6. Discussion

[54] Our overall comparison between groundwater fluctuation estimates of evapotranspiration against groundwater lysimeter measurements show that up to 50% of the daily variance could be explained with a RMSE of at least 1.32 mm/d. These results are in the range of former studies that compared evapotranspiration estimates from groundwater level fluctuations with measured evapotranspiration. High correlations were reported by *Farrington et al*. [1990], $R^2 = 0.81$ (White method versus ventilated chamber method, only 6 days), and *Engel et al*. [2005], $R^2 = 0.78$ (modified White method versus sap flow measurements). In contrast, previous comparisons to eddy covariance measurements showed mixed results. *Martinet et al*. [2009] investigated four sites and the average correlation for the White method spread from no correlation up to $R^2 = 0.5$ (7 years of data). The results of *Soylu et al*. [2012] were, for some of their measurements, even worse. However, their approach (R^2 of 0.4 and 0.62, *RMSE* of 0.64 and 1.1 mm/d) performed better than the White method (R^2 of 0.2 and 0.28, *RMSE* of 0.85 and 1.77 mm/d).

[55] For our data set, most of the methods under investigation (including modifications) performed almost equally well. Even more sophisticated approaches [e.g., *Soylu et al*., 2012] did not notably improve the results. Moreover, the approaches we used did not necessarily perform well for the right reasons. For example, the assumption that nighttime evapotranspiration does not occur, needed for the method of *Hays* [2003], is invalid. Mean evapotranspiration during the time period not covered by the Hays approach (i.e., before maximum and after minimum groundwater level) added up to 10.7% of the daily value. Yet overestimated recharge (on average 27% when the approach was applied to measured hourly recharge) compensated for omitted nighttime evapotranspiration to a certain degree. On the other hand, two approaches performed considerably worse than the others: the methods by *Loheide* [2008] and by *White* [1932], the latter being the most commonly used approach. The Loheide method assumes a recovery source that is constant or following the general trend of the water table. In our study, we did not find this to be the case. *Loheide* [2008] suggested that, if the assumption is not met, a slope other than the trend from the observed record should be removed, but this is beyond the scope of this investigation. In contrast, we found that evapotranspiration estimates of the White method could be improved by adjusting the procedure to determine the daily recharge.

[56] Originally, *White* [1932] chose the period between midnight and 4 A.M. for estimation of the recharge rate. It is often claimed that the method considers recharge to be constant throughout the day. In fact, *White* [1932] pointed out that he was considering an average recharge rate for the day. He argued that the supposed time period is suitable, as the water level is approximately at its mean value of the day, i.e., corresponding to the mean gradient between water level and recovery source. However, the hourly recharge data (Figure 4) indicated that during this period recharge is (slightly) lower than the daily mean, hence leading to underestimation when this period is used. This underestimation of recharge is mostly due to the considerably higher recharge values in the late afternoon and early night after cessation of evapotranspiration (i.e., when groundwater levels are at a minimum). Better results for the estimation of the average recharge rate were achieved when longer time spans were used. Calculations revealed that usage of a two night average, as suggested by *Loheide et al*. [2005], always outperformed recharge estimation based only on the night before evapotranspiration. For measured recharge, the best results were achieved when using the average of the trends between 6 P.M. to 6 A.M. of two nights.

[57] Our findings on recharge estimation did not translate directly to the estimation of evapotranspiration. We found for the White method, that better recharge calculation procedures (e.g., the modification by *Rushton* [1996]) did not necessarily improve evapotranspiration estimates. The modification to the White method by *Loheide* [2008] provided better estimates of evapotranspiration than the modification by *Miller et al*. [2010], although the latter performed better in estimating the recharge. However, the evaluation of the approaches concerning the estimation of recharge was done using measured hourly recharge. Deviations may arise when the approaches are applied to groundwater levels. Even if only a small amount of evapotranspiration was measured at nighttime, the groundwater table could still be affected by hydraulic redistribution (toward equilibrium) in the unsaturated zone [*Miller et al*.,

2010] and water uptake by plants replenishing their storage [*Loheide*, 2008], among others, which would mask the recharge signal and influence estimated recharge. Nevertheless, the best recharge procedure we found also represented the best modification of the White method.

[58] We found the specific yield is of crucial importance to estimating evapotranspiration based on diurnal groundwater fluctuations, as pointed out before by others [e.g., *Bleby et al*., 1997; *Lautz*, 2008; *Martinet et al*., 2009], because errors in the specific yield estimates directly translate to errors in the evaporation estimates [*Loheide*, 2008]. In contrast to theoretically derivable relations [e.g., *Duke*, 1972; *Nachabe*, 2002], the specific yield $S_{y,wb}$ obtained by application of a water balance approach varied notably. *Logsdon et al*. [2010] also observed considerable variation for a similar approach and showed that evapotranspiration demands and soil-water dynamics influenced the specific yield. A certain error was introduced in our case by using water storage change of the lysimeter since it incorporated water losses in the unsaturated zone that did not affect the groundwater table directly. To prevent the propagation of the uncertainty inherent to $S_{y,wb}$, normalized variants (i.e., excluding the specific yield) of recharge and evapotranspiration were used for the model evaluation.

[59] The specific yield which best fitted estimates and measurements for recharge ($S_{y,rfit}$) and evapotranspiration ($S_{y,ETfit}$), varied among the different approaches in a considerable range. However, the variance of $S_{y,ETfit}$ values in *Soylu et al*. [2012] was even more pronounced. Our values of $S_{y,ETfit}$ agree with those of the water balance approach, as they ranged from the first quartile to the median value of $S_{y,wb}$. The values of $S_{y,rfit}$ were even lower. The differences between $S_{y,ETfit}$ and $S_{y,rfit}$ values can originate from the dependence of the specific yield on the direction of water level change, for which it can be divided in drainable and fillable porosity. But since $S_{y,ETfit}$ and $S_{y,rfit}$ differ enormously (the values of the latter are between 50% and 95% higher), it rather seems that recharge is overestimated and then compensated by the assumptions concerning storage change.

[60] Nearly all calculated variants of the specific yield showed no dependency on depth to groundwater. A nonmonotonic relationship was found in an experiment conducted in winter of 2009 (data not shown), where the groundwater level in the lysimeter was stepwise raised and afterward lowered with the surface of the lysimeter covered. However, these steady state conditions do not reflect the behavior under transient conditions. For the values calculated by the water balance approach ($S_{y,wb}$), no depth dependence was visible. Most likely the dependency on depth to groundwater was masked by the multitude of influencing factors and processes (e.g., changing evapotranspiration demands, varying conditions in the vadose zone). For some methods, a rather weak dependence of $S_{y,ETfit}$ on the groundwater level was seen (Figure 7). However, it appears that high $S_{y,ETfit}$ values near the surface compensate for difficulties of the different approaches under conditions with drainage or very low recharge. Higher values for deeper groundwater levels could result from vadose zone depletion or high recharge, as a significant correlation ($r = -0.61$, $p < 0.05$) between recharge and groundwater depth was found. In contrast, *Soylu et al*. [2012] found an inverse exponential relationship and attributed the higher values for shallow water levels to enhanced water availability in the vadose zone or higher contribution from lateral flow. Altogether, the use of a depth-independent specific yield throughout the study might have introduced a minor error in evapotranspiration estimates.

[61] The comparison of the evapotranspiration rates is subject to limitations, since the rates measured by the lysimeter include more components than the rates estimated from groundwater fluctuations. Above all, the latter do not include interception evaporation. The impact should be small, since only a few selected days followed days with rain and rain events often ended hours ago. In general, the evapotranspiration estimates of those days may be biased as water redistribution in the vadose zone can influence the groundwater fluctuations. Other components only considered in the lysimeter are dew formation and evaporation, canceling out their overall impact, and only resulting in minor shifts should dew formation differ between days. Evapotranspiration coming from the unsaturated zone is not covered by fluctuation methods per se and cannot be measured separately in the lysimeter. But, since only daily evapotranspiration values were considered (i.e., there was time for redistribution after transpiration ceased) and the vadose zone was close to saturation due to high groundwater levels, the effect should be small. In other environments a closer examination of the hydraulic connection between saturated and unsaturated zone may be needed.

[62] Despite trying to reproduce natural conditions, the lysimeter is an artificial environment. We assume that the in- or outflow to the lysimeter equals the groundwater recharge or discharge at the field site, but small deviations may emerge. For example, the extent of groundwater fluctuation in the lysimeter was slightly higher than that of the reference gauge. This was probably caused by an oasis effect, as the lysimeter was elevated compared to the encompassing field, or other related effects [see *Allen et al*., 2011]. Higher evapotranspiration rates caused lower water levels in the lysimeter compared to those of the reference gauge, yet, as the former is adjusted to the latter, inflows can be slightly exaggerated. This can generate deviations in the recharge patterns which might affect their estimation based on the groundwater hydrograph. Still, the lysimeter reflects realistic behavior and the expected errors of the water balance components are small. In particular, lysimeters typically provide the most accurate evapotranspiration measurements [see *Allen et al*., 2011]. This is of primary importance since the measured evapotranspiration served as means of comparison. Hence, every error would have had an unpredictable impact on the conducted method evaluation.

[63] Nevertheless, it must be pointed out that the assessment of the different methods is only valid for the specific conditions of this study, i.e., a grassland site with shallow groundwater table. Results from different environments (e.g., riparian areas with trees) could differ notably, for instance, *Martinet et al*. [2009] claimed that the White method performs better at sites with coarse texture and deeper groundwater table. The available data set influenced the obtained results, as well. Some of the best-performing approaches yielded poor results for days with high evapotranspiration, which were underrepresented due to wet

conditions in the summers of 2011 and 2012. A different composition of days could have affected the outcome of the study. Though, the results contradict the thesis by *Zhu et al*. [2011] who stated that evapotranspiration estimates are more error-prone during times of low fluctuations.

[64] In total, in spite of the imprecise evapotranspiration estimates and the problems concerning the determination of the specific yield, groundwater fluctuation-based methods can still be used for comparison of evapotranspiration at different locations. The correlation between two stations is not affected by the specific yield, at least when the latter is assumed to be constant, but can reveal differences regarding vegetation and recharge conditions [*Lautz*, 2008]. Another promising option is the application of the approach by *Nachabe et al*. [2005]. They interpreted the White method by application of total soil moisture instead of groundwater levels, solving two of the main problems. First, evapotranspiration from the unsaturated zone is taken into account and second, no determination of the specific yield is necessary. However, requirements concerning the equipment are higher and restricted the application for our study site, since a prohibitive number of moisture probes would be needed to capture the soil moisture variability between soil surface and groundwater table with an appropriate resolution.

7. Conclusion

[65] The aim of this paper was to evaluate different approaches for estimation of evapotranspiration based on diurnal groundwater fluctuations, also with regard to their assumptions concerning the recharge and the specific yield. For the 85 rain-free days under investigation the method proposed by *Gribovszki et al*. [2008] obtained the highest correlation with evapotranspiration measured at a groundwater lysimeter. In general, the complexity of an approach was no indicator for effectiveness. For example, the White method, which is simple but the most frequently applied approach, performed comparable to the best methods when the period for recharge estimation (by default between midnight and 4 A.M.) was modified appropriately. In our case, the average of the periods between 6 P.M. and 6 A.M. of the nights before and after the day under consideration gave the best results. Specific yield values showed high variation, both for a water balance approach and when calculated to fit evapotranspiration or recharge measurements and estimates. The differences of the latter among the methods implied that the specific yield can function as a "correction factor," which offsets deficiencies in the underlying assumptions. In essence, for our conditions the accuracy of the evapotranspiration estimates was limited, but the errors involved could not generally be assigned to the specific yield since low correlation between the evapotranspiration estimates and measurements indicated an inadequate process description of the approaches.

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