Comparison of seven models for estimation of evapotranspiration and groundwater recharge using lysimeter measurement data in Germany

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Abstract:
This study evaluates seven evapotranspiration models and their performance in water balance studies by using lysimeter measurement data at the Mönchengladbach hydrological and meteorological station in Germany. Of the seven evapotranspiration models evaluated, three models calculate actual evapotranspiration directly using the complementary relationship approach, i.e. the CRAE model of Morton, the advection–aridity (AA) model of Brutsaert and Stricker, and the GG model of Granger and Gray, and four models calculate first potential evapotranspiration and then actual evapotranspiration by considering the soil moisture condition. Two of the four potential evapotranspiration models belong to the temperature-based category, i.e. the Thornthwaite model and the Hargreaves model, and the other two belong to the radiation-based category, i.e. the Makkink model and the Priestley–Taylor model. The evapotranspiration calculated by the above seven models, together with precipitation, is used in the water balance model to calculate other water balance components. The results show that, for the calculation of actual evapotranspiration, the GG model and the Makkink model performed better than the other models; for the calculation of groundwater recharge using the water balance approach, the GG model and the AA models performed better; for the simulation of soil moisture content using the water balance approach, four models (GG, Thornthwaite, Makkink and Priestley–Taylor) out of the seven give equally good results. It can be concluded that the lysimeter-measured water balance components, i.e. actual evapotranspiration, groundwater recharge, soil moisture, etc., can be predicted by the GG model and the Makkink model with good accuracy. Copyright © 2005 John Wiley & Sons, Ltd.

KEY WORDS water balance; evapotranspiration; lysimeter; groundwater recharge

INTRODUCTION

Accurate estimations of actual evapotranspiration are required in hydrologic studies and water resources modelling of stationary and changing climate conditions. Three terms are usually used in describing evaporation, ET, in the literature: (1) The term free water evaporation \(ET_0\) is used for the amount of evaporation from open/free water surface, i.e. the water is returned to the atmosphere from lakes and reservoirs and, in some cases, from river channels in a river catchment (e.g. Peterson \textit{et al.}, 1995). (2) The term actual evapotranspiration \(ET_a\) describes all the processes by which liquid water at or near the land surface becomes atmospheric water vapour under natural condition (e.g. Morton, 1983). (3) The term potential evapotranspiration \(ET_p\) is water loss that will occur if at no time is there a deficiency of water in the soil for use of vegetation. (Thornthwaite, 1944). The concept of potential evapotranspiration was introduced to study the evaporative demand of the atmosphere independently of soil factors. For a give vegetation type,
the only factors affecting \( ET_p \) are climatic parameters. Consequently, \( ET_p \) is a climatic parameter and can be computed from weather data.

Most weather stations record rainfall, but few measure potential and/or actual evapotranspiration despite it being an important parameter in climatology. What happens to rainfall once it has reached the ground is of interest to ecologists, hydrologists and water engineers. The weighing lysimeter represents the best available technology for determining vegetative water use which gives additional information on soil water balance. However, it is physically and economically impossible to measure evapotranspiration in every area of interest. Since establishing and maintaining lysimeters for a long time period is costly, the detailed measurements associated with a lysimeter, i.e. evapotranspiration, groundwater recharge together with rainfall, are used to calibrate evapotranspiration models and evaluate water balance models. The information obtained could significantly enhance the reliability of evapotranspiration models and water balance models.

The main objectives of this study are: (1) to evaluate seven often-used evapotranspiration models and to examine their performance in water balance studies by using lysimeter measurement from a station in central Germany; (2) to compare different approaches for modelling soil water budget by using the lysimeter measurement.

Where detailed data about soil layers, depth to groundwater, and vegetation are not available, hydrologists have often resorted to simple bucket models and budgeting schemes to model near-surface hydrology. The soil-water budget is a simple accounting scheme used to predict soil-water storage, evapotranspiration, and water surplus. Surplus is the fraction of precipitation that exceeds evapotranspiration and is not stored in the soil. The simple model used here does not distinguish between surface and subsurface runoff, so surplus includes both. With this in mind, the basic equation for calculating surplus is:

\[
S = P - ET_a \pm \frac{\Delta W}{\Delta t}
\]  

where \( S \) is surplus (i.e. groundwater recharge, GWR, when no surface runoff is assumed), \( P \) is precipitation, \( ET_a \) is actual evapotranspiration, \( w \) is soil moisture, and \( t \) is time. A major source of uncertainty in evaluating Equation (1) is estimating actual evapotranspiration.

In this study, different models/methods for calculating actual evapotranspiration are tested. These models can be classified into two groups. Group I models estimate the actual evapotranspiration directly using the complementary relationship approach, i.e. the CRAE model of Morton, the advection–aridity (AA) model of Brutsaert and Stricker, and the GG model proposed by Granger and Gray using the concept of relative evapotranspiration. Group II models use knowledge of potential evapotranspiration, available water-holding capacity of the soil, and a moisture extraction function. In this group of models, one aspect of the soil-water budget that involves significant uncertainty and ambiguity is estimating potential evapotranspiration that defines the upper limit of water loss, which is a function of energy budget, temperature, humidity and turbulence of the atmosphere. Various formulas exist. Physically based methods (e.g. Penman–Motheith) require a lot of data that may not be available, four representative empirical potential evapotranspiration models selected from two categories, namely the temperature-based approach (Thorntwaite and Hargreaves methods) and radiation-based approach (Makkink and Priestley–Taylor methods), are evaluated.

This study differs from those reported in the literature (e.g. Hobbins et al., 2001a,b; Xu and Li, 2003) in the following respects: (1) it verifies seven models for calculating actual evapotranspiration (three of them use a complementary relationship approach and four of them use a potential evapotranspiration approach) by using lysimeter data, which has not been done before, to our knowledge; (2) the performance of these evapotranspiration models in calculating soil water balance at different time scales is evaluated using the hydrological and meteorological data associated with a lysimeter. The remainder of the paper is organized as follows: the study region and data are described in the following section; the evapotranspiration models and the soil water balance calculation schemes are described in the third section. The results and discussions are given in the fourth section; followed by the summary and conclusions in the final section.
STUDY REGION AND DATA

The hydrological and meteorological station Mönchengladbach-Rheindahlen in Germany (6-43°E and 51-18°N) is run by the Stadtwerke Mönchengladbach GmbH. The station has a weighable lysimeter station, a completely automatic weather station, a shaft to measure the percolation water, and a data-processing unit. The water budget components and meteorological data have been collected since 1982 (Wellens and Schumacher, 1995). All measurements associated with lysimeter III at the station, which has a soil column 2 m deep, are used in this study; these include daily measurements on precipitation, air temperature (minimum, average and maximum), wind speed, relative humidity, solar radiation, evapotranspiration, groundwater recharge, change of lysimeter weight, etc. The vegetation cover of the study site is grass.

In order to have a brief idea about the climatologic and water balance conditions of the study site, monthly values for a number of water budget components from the lysimeter are plotted in Figure 1. Although the monthly precipitation is fairly variable, it is seen that the actual evapotranspiration shows a regular pattern that is mainly determined by seasonal variation of solar radiation and temperature. It is interesting to note that the peaks of the actual evapotranspiration often correspond well to relatively low precipitation, which is most likely associated with high global solar radiation and temperature. This indicates the importance of the storage term in the water budget, i.e. the soil water supply is usually not a limiting factor for evapotranspiration in the region, as it is rather limited by energy supply in temperate humid regions. The groundwater recharge that is measured by water flow at 2 m depth simulates the recharge to the groundwater. It has some similarity with precipitation, especially during cold periods when evapotranspiration is low and soil moisture content is high. Over a seasonal cycle, there is always a period (usually from May to October and sometimes longer) when groundwater recharge did not happen or was very limited (Krueger et al., 2001). This is the period when evapotranspiration demand is high. Owing to the role of the soil moisture played by the seasonal variation in evapotranspiration, the residual term representing precipitation minus actual evapotranspiration and groundwater recharge does not vanish on a monthly basis. However, it is a rather small term on an annual basis, as shown by the 1 year averages.

To check the consistency of the measurement, the residual term, i.e. precipitation minus evapotranspiration and groundwater recharge \((P - E - GWR)\) is plotted against the changes in the weight of the lysimeter in Figure 2. A nearly perfect 1:1 relation is found, which confirms the consistency of the measurement and the assumption that groundwater recharge represents the difference between precipitation and actual evapotranspiration.

METHODOLOGIES

In order to calculate the soil-water budget in Equation (1) an estimate of evapotranspiration and the soil’s ability to store water is required. In this study the actual evapotranspiration is estimated by two approaches. The first approach uses the complementary evapotranspiration models and the second approach considers the actual evapotranspiration as a fraction of potential evapotranspiration. The fraction is an increasing function of water content of the soil. The discussion of the evapotranspiration models used in this study is presented in detail below. As for the soil’s ability to store water, several terms are used by soil scientists to define the water storage capacities of soils under different conditions. The field capacity FC, defined as the water content of soil that has reached equilibrium with gravity after several days of drainage, is used in the study. The methods of the water balance calculation and the determination of the field capacity for the study site are discussed in the section ‘Budgeting soil-moisture storage to yield surplus’.

Description of the complementary evapotranspiration models

Utilizing an analysis based on an energy balance, Bouchet (1963) corrected the misconception that a larger potential evapotranspiration necessarily signified a larger actual evapotranspiration by demonstrating
that as a surface dried from initially moist conditions the potential evapotranspiration increased while the actual evapotranspiration was decreasing. The relationship that he derived is known as the complementary relationship between actual and potential evapotranspiration; it states that as the surface dries the decrease in actual evapotranspiration is accompanied by an equal, but opposite, change in the potential evapotranspiration. Thus, the potential evapotranspiration ranges from its value at saturation to twice this value. This relationship is described as

\[ \text{ET}_a + \text{ET}_p = 2\text{ET}_w \] 

(2)

where \( \text{ET}_w \) is the wet environment evapotranspiration.

The complementary relationship has formed the basis for the development of some evapotranspiration models (Brutsaert and Stricker, 1979; Morton, 1983; Granger and Gray, 1989), which differ in the calculation of \( \text{ET}_p \) and \( \text{ET}_w \). \( \text{ET}_a \) is usually calculated as a residual of Equation (2). For the sake of completeness, the
model equations are briefly summarized in what follows using the same notation as used by the original authors. For a more complete discussion, the reader is referred to the literature cited.

**AA model.** In the AA model (Brutsaert and Stricker, 1979), the $ET_p$ is calculated by combining information from the energy budget and water vapour transfer in the Penman (1948) equation:

$$ET_p = \frac{\Delta}{\Delta + \gamma} \frac{R_n}{\gamma} + \frac{\gamma}{\Delta + \gamma} E_a$$  \hspace{1cm} (3)

where $R_n$ is the net radiation near the surface, $\Delta$ is the slope of the saturation vapour pressure curve at the air temperature, $\gamma$ is the psychrometric constant, $\lambda$ is the latent heat, and $E_a$ is the drying power of the air, which in general can be written as

$$E_a = f(U_z)(e_s - e_a)$$  \hspace{1cm} (4)

where $f(U_z)$ is a function of the mean wind speed $U_z$ at a reference level $z$ above the ground, and $e_s$ and $e_a$ are the vapour pressure of the air and the saturation vapour pressure at the air temperature respectively. In this study, the empirical linear approximation for $f(U_z)$ originally suggested by Penman (1948) is used:

$$f(U_z) \approx f(U_2) = 0.0026(1 + 0.54U_2)$$  \hspace{1cm} (5)

which, for wind speed at 2 m elevation in metres per second and vapour pressure in pascals, yields $E_a$ in millimetres per day. This formulation of $f(U_2)$ was first proposed by Brutsaert and Stricker (1979) for use in the AA model operating at a temporal scale of a few days. Substituting Equation (4) and the wind function Equation (5) into Equation (3) yields the expression for $ET_p$ used by Brutsaert and Stricker (1979) in the original AA model:

$$ET_p^{AA} = \frac{\Delta}{\Delta + \gamma} \frac{R_n}{\gamma} + \frac{\gamma}{\Delta + \gamma} f(U_2)(e_s - e_a)$$  \hspace{1cm} (6)

The AA model calculates $ET_w$ (Brutsaert and Stricker, 1979) using the Priestley and Taylor (1972) partial equilibrium evapotranspiration equation:

$$ET_w^{AA} = \alpha \frac{\Delta}{\Delta + \gamma} \frac{R_n}{\lambda}$$  \hspace{1cm} (7)
where \( \alpha = 1.26 \). Different values for \( \alpha \) have been reported in the literature; the original value is tested in this study. Substitution of Equations (6) and (7) into Equation (2) results in the expression for \( E_{T_a} \) in the AA model:

\[
E_{T_a}^{AA} = (2\alpha - 1) \frac{\Delta R_n}{\Delta + \gamma} \frac{\gamma}{\Delta + \gamma} f(U_2)(e_s - e_a) \tag{8}
\]

**GG model.** Granger (1989) showed that an equation similar to Penman’s could also be derived following the approach of Bouchet’s (1963) complementary relationship. Granger and Gray (1989) derived a modified form of Penman’s equation for estimating the actual evapotranspiration from different non-saturated land covers:

\[
E_{T_a}^{GG} = \frac{\Delta G}{\Delta G + \gamma} \frac{R_n}{\lambda} + \frac{\gamma G}{\Delta G + \gamma} E_a \tag{9}
\]

where \( G \) is a dimensionless relative evapotranspiration parameter and the other symbols have the same meanings as in Equation (3). Granger and Gray (1989) showed that the relative evapotranspiration, the ratio of actual to potential evapotranspiration, \( G = E_{T_a}/E_{T_p} \) is a unique parameter for each set of atmospheric and surface conditions. Based on daily estimated values of actual evapotranspiration from water balance, Granger and Gray (1989) showed that there exists a unique relationship between \( G \) and a parameter that they called the relative drying power \( D \), given as

\[
D = \frac{E_a}{E_a + R_n} \tag{10}
\]

and

\[
G = \frac{1}{1 + 0.028 e^{0.345D}} \tag{11}
\]

Later, Granger (1998) modified Equation (11) to

\[
G = \frac{1}{0.793 + 0.20 e^{0.902D}} + 0.006D \tag{12}
\]

**CRAE model.** Different forms of the CRAE model have been reported in the literature; in this study, the original form presented by Morton (1983) was used. To calculate \( E_{T_p} \) in the CRAE model, Morton (1983) decomposed the Penman equation into two separate parts describing the energy balance and vapour transfer process. A refinement was developed by using an ‘equilibrium temperature’ \( T_p \), which is defined as the temperature at which Morton’s (1983) energy budget method and mass transfer method for a moist surface and plants yields the same result for \( E_{T_p} \). The energy balance and vapour transfer equations can be expressed, respectively, as:

\[
E_{T_p}^{CRAE} = R_T - [\gamma f_T + 4\varepsilon\sigma(T_p + 273)^3](T_p - T) \tag{13}
\]

\[
E_{T_p}^{CRAE} = f_T(e_p - e_d) \tag{14}
\]

in which \( E_{T_p} \) is the potential evapotranspiration in the units of latent heat, \( T_p \) and \( T \) (°C) are the equilibrium temperature and air temperature respectively, \( R_T \) is the net radiation for soil–plant surfaces at the air temperature, \( \gamma \) is the psychrometric constant, \( \sigma \) is the Stefan–Boltzmann constant, \( \varepsilon \) is the surface emissivity, \( f_T \) is the vapour transfer coefficient, \( e_p \) is the saturation vapour pressure at \( T_p \), and \( e_d \) is the saturation vapour pressure at the dew-point temperature. The potential evapotranspiration estimate is obtained by using in Equation (13) the value of \( T_p \) obtained by an iterative process (Morton, 1983).

In calculating the wet-environment evapotranspiration, Morton (1983) modified the Priestley–Taylor equilibrium evapotranspiration Equation (7) to account for the temperature dependence of both the net radiation term and the slope of the saturated vapour pressure curve \( \Delta \). The Priestley–Taylor factor \( \alpha \) is replaced by a smaller factor \( b_2 = 1.20 \), while the addition of \( b_1 = 14 \text{ W m}^{-2} \) (or 0.49 mm day\(^{-1}\)) accounts for large-scale...
advection during seasons of low or negative net radiation and represents the minimum energy available for ET_w but becomes insignificant during periods of high net radiation:

$$ET_w^{CRAE} = b_1 + b_2 \frac{\Delta_p}{\Delta_p + \gamma} R_{T_p}$$

$$= b_1 + b_2 \frac{\Delta_p}{\Delta_p + \gamma} [R_n - 4\varepsilon \sigma T_p^3 (T_p - T_a)]$$

(15)

where $\Delta_p$ and $R_{T_p}$ are respectively the slope of the saturated vapor pressure curve and the net available energy adjusted to the equilibrium temperature $T_p$. Other symbols are as defined previously. Actual evapotranspiration is calculated as a residual of Equation (2), i.e.

$$ET_a^{CRAE} = 2ET_w^{CRAE} - ET_p^{CRAE}$$

(16)

**Potential evapotranspiration**

Earlier studies (Singh and Xu, 1997; Xu and Singh, 2000, 2001, 2002) have evaluated and compared various popular empirical potential evapotranspiration equations that belonged to three categories, viz. (1) mass-transfer-based methods, (2) radiation-based methods, and (3) temperature-based methods, and the representative methods from each category were evaluated by using the Penman–Monteith method as a standard. Determining whether there are significant differences between predicted runoff in a water-balance study using different methods is an important question, as some methods are more easily applied over different historical time periods and some methods have a more physical background. In this study, four representative empirical potential evapotranspiration equations selected from two categories, namely a temperature-based approach (Thornthwaite and Hargreaves methods) and a radiation-based approach (Makkink and Priestley–Taylor methods), are evaluated and their performance in water balance studies is compared with the complementary evapotranspiration models. The selection of the methods is based on their simplicity and popularity in the hydrological literature. A discussion of these methods now follows.

**Thornthwaite method.** A widely used method for estimating potential evapotranspiration was derived by Thornthwaite (1948), who correlated mean monthly temperature with evapotranspiration as determined from a water balance for valleys where sufficient moisture water was available to maintain active transpiration. In order to clarify the existing method, the computational steps of the Thornthwaite equation are now discussed.

**Step 1:** the annual value of the heat index $I$ is calculated by summing monthly indices over a 12 month period. The monthly indices are obtained using

$$i = \left( \frac{T_a}{5} \right)^{1.51}$$

(17a)

and

$$I = \sum_{j=1}^{12} i_j$$

(17b)

in which $i_j$ is the monthly heat index for the month $j$ (which is zero when the mean monthly temperature is 0 °C or less), $T_a$ (°C) is the mean monthly air temperature, and $j$ is the number of months (1–12).

**Step 2:** the Thornthwaite general equation, Equation (18a), calculates unadjusted monthly values of potential evapotranspiration $ET_p'$ (mm) based on a standard month of 30 days, with 12 h of sunlight per day.

$$ET_p' = C \left( \frac{10T_a}{I} \right)^a$$

(18a)

in which $C = 16$ (a constant) and $a = 67.5 \times 10^{-8}I^3 - 77.1 \times 10^{-6}I^2 + 0.0179I + 0.492$. 

The value of the exponent $a$ in the preceding equation varies from zero to 4.25 (e.g. Jain and Sinai, 1985), the annual heat index varies from zero to 160, and $ET_0^p$ is zero for temperatures below 0°C.

Step 3: the unadjusted monthly evapotranspiration values $ET_0^p$ are adjusted depending on the number of days $N$ in a month ($1 \leq N \leq 31$) and the duration of average monthly or daily daylight $d$ (h), which is a function of season and latitude:

$$ET_p = ET_0^p \frac{d \cdot N}{12 \cdot 30}$$

(18b)

in which $ET_p$ (mm) is the adjusted monthly potential evapotranspiration.

**Hargreaves method.** Hargreaves and Samani (1982, 1985) proposed several improvements for the Hargreaves (1975) equation for estimating grass-related reference $ET_p$ (mm day$^{-1}$); one of them has the form

$$ET_p = a R_a T^{1/2} (T_a + 17.8)$$

(19)

where $a = 0.0023$ is a constant, $T_D$ (°C) is the difference between maximum and minimum daily temperature, and $R_a$ is the extraterrestrial radiation expressed in equivalent evapotranspiration units. For a given latitude and day $R_a$ is obtained from tables or may be calculated using the equation recommended by the FAO (Allen et al., 1998). The only variables for a given location and time period are the daily mean, maximum and minimum air temperatures.

**Makkink method.** For estimating potential evapotranspiration $ET_p$ (mm day$^{-1}$) from grass, Makkink (1957) proposed the equation

$$ET_p = 0.61 \frac{\Delta}{\Delta + \gamma \lambda} R_s - 0.12$$

(20)

where $R_s$ (cal cm$^{-2}$ day$^{-1}$) is the total solar radiation, $\Delta$ (mbar °C$^{-1}$) is the slope of the saturation vapour pressure curve, $\gamma$ (in mbar °C$^{-1}$) is the psychrometric constant, $\lambda$ (cal g$^{-1}$) is the latent heat, and $P$ (mbar) is atmospheric pressure. These quantities are calculated using the FAO-98 recommended procedure (Allen et al., 1998) and units have to be converted to what are required in Equation (20).

On the basis of later investigation in the Netherlands and at Tastrup in Denmark, Hansen (1984) proposed the following form of the Makkink equation:

$$ET_p = 0.7 \frac{\Delta}{\Delta + \gamma \lambda} R_s$$

(21)

where all the symbols have the same meaning and units as in Equation (20). This equation will be used in this study, since the data used in determining the constant value are closer to the study region.

**Priestley–Taylor method.** Priestley and Taylor (1972) proposed a simplified version of the combination equation (Penman, 1948) for use when surface areas generally were wet, which is a condition required for potential evapotranspiration $ET_p$. The aerodynamic component was deleted and the energy component was multiplied by a coefficient, $\alpha = 1.26$, when the general surrounding areas were wet or under humid conditions.

$$ET_p = \alpha \frac{\Delta}{\Delta + \gamma \lambda} R_n$$

(22)

where $ET_p$ has units of millimetres per day, $R_n$ (cal cm$^{-2}$ day$^{-1}$) is the net radiation, and the other symbols have the same meanings and units as in Equation (20).
Budgeting soil-moisture storage to yield surplus

Soil-water budget calculations are commonly made using monthly or daily rainfall totals because of the way the data are recorded. The use of daily values is preferred over monthly values when possible, particularly in dry locations where the mean potential evapotranspiration for a given month may be higher than the mean precipitation, yet there is observed runoff. In this study, however, partly because the study region is humid and partly due to the fact that some potential evapotranspiration calculations are done on a monthly basis, the comparison is performed on a monthly basis.

The water balance calculation is performed in two different ways, depending on how the evapotranspiration is calculated.

Case one: when the actual evapotranspiration is calculated by using the three complementary evapotranspiration models, Equation (23) describes how soil moisture storage is computed.

\[
W(t) = W(t-1) + P(t) - ET_a(t)
\]  

(23)

In Equation (23), \(W(t)\) is the current soil moisture, \(W(t-1)\) is the soil moisture in the previous time step, \(P(t)\) is precipitation, and \(ET_a(t)\) is actual evapotranspiration. Several conditions apply when evaluating Equation (23):

- If \(W(t) < 0\) then \(W(t) = 0\) and Surplus = 0
- If \(W(t) > W^*\) then Surplus = \(W(t) - W^*\) and \(W(t)\) is set to \(W^*\)

where \(W^*\) is the water-holding capacity, i.e. field capacity FC in this study.

Case two: when potential evapotranspiration \(ET_p\) is used in the calculation, the procedure is as follows:

- If \(ET_a(t) > P(t)\), then soil water will be depleted to compensate the supply. At the same time, we have \(ET_a(t) < ET_p(t)\) and Surplus = 0. Under this condition, \(ET_a(t)\) is assumed to be proportional to \(W(t)/W^*\).
- When \(ET_p(t) = P(t)\), then \(ET_p(t) = ET_a(t)\), Surplus = 0.
- If \(ET_a(t) < P(t)\), then \(ET_a(t) = ET_p(t)\). \(W(t)\) is first estimated with Surplus = 0. When \(W(t) > FC\), \(WT(t) = W(t) - FC\); when \(W(t) ≤ FC\), Surplus = 0.

If the initial soil moisture is unknown, which is typically the case, then a balancing routine is used to force the net change in soil moisture from the beginning to the end of a specified balancing period \(N\) time steps) to zero. To do this, the initial soil moisture is set to the water-holding capacity and budget calculations are made up to the time period \(N + 1\). \(W(1)\), the initial soil moisture at time 1, is then set equal to the soil moisture at time \(N + 1\) \((W(N + 1))\) and the budget is recomputed until the difference \(W(1) - W(N + 1)\) is less than a specified tolerance.

One of the parameters that has to be specified is the FC of soil. FC represents the amount of water remaining in a soil after the soil layer has been saturated and the free (drainable) water has been allowed to drain away (a few days). The maximum water content of the soil at six different layers of lysimeter III listed by Schumacher and Wellens (1993) was used. The averaged volumetric saturated water content is 0·81 m$^3$ m$^{-3}$ for the 2 m deep soil in lysimeter III, which gives a field capacity of 810 mm. For the first metre this parameter was calculated to be 356 mm.

RESULTS AND DISCUSSION

Calculation of actual evapotranspiration

The three complementary evapotranspiration models are first applied to data at the Mönchengladbach station to calculate actual evapotranspiration. The calculations are made on a daily basis for the period of
1983 to 1994. Four potential evapotranspiration models, as discussed earlier, are used to calculate the monthly potential evapotranspiration, which is then used to calculate actual evapotranspiration following the procedure outlined in Case two. The comparisons of the methods are first made on an annual basis. The observed yearly evapotranspiration and the relative errors of the calculated evapotranspiration to the observed values are presented in Table I. It is seen that on an annual basis the AA model gives the smallest error compared with the observed value, followed by the Priestley–Taylor, GG and Makkink models. The above four models have a mean annual error of less than 5%. The largest annual errors are found for the CRAE and Hargreaves models, with mean annual errors of 11-5% and 11-3% respectively (i.e. an overestimation of about 60 mm year\(^{-1}\)). In order to see the ability of the models in calculating monthly and seasonal evapotranspiration values, the mean monthly values are compared in Figure 3. Two features can be seen from Figure 3: (1) For the complementary evapotranspiration models, the GG model follows the seasonal variation better than the other two, whereas for the potential evapotranspiration models the Makkink model follows the seasonal variation better than the others. (2) Almost all the models overestimate monthly evapotranspiration for summer months and some of them underestimate monthly evapotranspiration for winter months. For example, although the AA model gives a best fit for annual value, it overestimates summer evapotranspiration by about 20 mm. This phenomenon can, of course, be improved by a local calibration of the parameter values involved in each model (e.g. Xu and Singh, 2005). Recalibration of the parameter values is not done in this study because our main purpose is to examine the applicability and accuracy of the models in the study region with their original parameter values. The correlations between monthly values computed from the seven evapotranspiration models and the observed values are analysed using the simple linear regression equation

\[
Y = mX + c
\]  

where \(Y\) represents \(ET_a\) computed by evapotranspiration models, \(X\) is the observed \(ET_a\), and \(m\) and \(c\) are constants representing the slope and the intercept of the regression equation respectively. The resulting values of the regression equation parameters are presented in Table II. For illustrative purposes, the plots are shown in Figure 4 for the complementary group models only. It is seen that the GG model of the complementary model group and the Makkink model of the potential evapotranspiration model group have the highest coefficient

<table>
<thead>
<tr>
<th>Year</th>
<th>(ET_{obs}) (mm)</th>
<th>AA</th>
<th>CRAE</th>
<th>GG</th>
<th>Thornthwaite</th>
<th>Hargreaves</th>
<th>Makkink</th>
<th>Priestley–Taylor</th>
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<td>31.6</td>
<td>6.9</td>
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<td>27.8</td>
<td>16.9</td>
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<td>453.3</td>
<td>9.6</td>
<td>25.0</td>
<td>11.9</td>
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<td>11.3</td>
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</table>

\(\text{Error (\%)} = \frac{ET_{cal} - ET_{obs}}{ET_{obs}} \times 100.\)
of determination values, with $R^2 = 0.92$ and 0.91 respectively. The Thornthwaite method has the lowest value, with $R^2 = 0.86$. Concerning the slope and intercept of the regression equations, all four potential evapotranspiration models give good results compared with the complementary evapotranspiration models. The worst case is the AA model, which has largest intercept value and the worst slope value; this is due to its mismatch of monthly variation, as shown in Figure 3.

Table II. Regression relationships between observed evapotranspiration (independent variable) and estimated evapotranspiration (dependent variable)

<table>
<thead>
<tr>
<th>Methods</th>
<th>Slope $M$</th>
<th>Intercept $C$</th>
<th>$R^2$</th>
</tr>
</thead>
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<tr>
<td><strong>Complementary methods</strong></td>
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<tr>
<td>AA model</td>
<td>1.30</td>
<td>-13.49</td>
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<tr>
<td>GG model</td>
<td>1.13</td>
<td>-6.93</td>
<td>0.92</td>
</tr>
<tr>
<td>CRAE model</td>
<td>1.10</td>
<td>0.63</td>
<td>0.90</td>
</tr>
<tr>
<td><strong>Potential evapotranspiration methods</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thornthwaite</td>
<td>1.04</td>
<td>0.92</td>
<td>0.86</td>
</tr>
<tr>
<td>Hargreaves</td>
<td>1.06</td>
<td>2.20</td>
<td>0.89</td>
</tr>
<tr>
<td>Makkink</td>
<td>0.95</td>
<td>3.70</td>
<td>0.91</td>
</tr>
<tr>
<td>Priestley–Taylor</td>
<td>1.13</td>
<td>-5.20</td>
<td>0.89</td>
</tr>
</tbody>
</table>

Figure 3. Observed and computed mean monthly evapotranspiration using (a) the complementary relationship models and (b) the potential evapotranspiration models.
Simulation of groundwater recharge

Following the procedure earlier, the monthly soil water surplus, i.e. groundwater recharge when no surface runoff is assumed, is calculated. Figure 5 shows the yearly values of the observed and estimated groundwater recharge from the seven models. It is seen that two of the three complementary relationship models, i.e. the GG model and AA model, result in the best estimate of annual groundwater recharge. Among the potential evapotranspiration methods, the Makkink and the Priestley–Taylor methods give better estimations than the temperature-based methods, i.e. the Thornthwaite and Hargreaves methods. Regarding the mean monthly and mean annual estimates (Table III), two features can be seen: (1) For the mean annual estimates, the best results are found for the GG and AA models, with mean annual errors of 4.4% and 5.7% respectively. The worst cases are found for the Hargreaves and CRAE models, which underestimate the annual groundwater recharge by about 30%. (2) For the monthly values, all methods underestimate the groundwater recharge to some extent for the months from April to August, which is due to the overestimation of actual evapotranspiration for those months, as is shown in Figure 3.

Simulation of soil moisture content

By rewriting the water balance Equation (1), the soil moisture content can be estimated using the observed precipitation and calculated or observed groundwater recharge and evapotranspiration. The monthly soil moisture content calculated using the seven models and the observed soil moisture content are compared in

![Figure 4. Correlation between observed and computed monthly evapotranspiration using the complementary relationship models](image-url)
Table III. Comparison of the observed and calculated mean monthly and mean annual groundwater recharge values

<table>
<thead>
<tr>
<th>Month</th>
<th>Observed</th>
<th>AA</th>
<th>CREA</th>
<th>GG</th>
<th>Thornthwaite</th>
<th>Hargreaves</th>
<th>Makkink</th>
<th>Priestley–Taylor</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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</tr>
<tr>
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<td>2.7</td>
<td>4.9</td>
<td>3.1</td>
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<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>8</td>
<td>0.4</td>
<td>0.0</td>
<td>0.0</td>
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<td>0.0</td>
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<td>22.9</td>
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<td>193.1</td>
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<tr>
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<td>-21.5</td>
<td>-30.9</td>
<td>-16.7</td>
<td>-11.0</td>
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</tbody>
</table>

* Error (%) = \( \frac{GWR_{cal} - GWR_{obs}}{GWR_{obs}} \times 100. *

Figure 5. Observed and computed annual groundwater recharge using (a) the complementary evapotranspiration models and (b) the potential evapotranspiration models
Figure 6. In this figure, the calculated soil moisture content means that it is calculated with Equation (1) using calculated evapotranspiration and groundwater recharge, whereas the observed soil moisture content means that it is calculated with the same equation but using observed groundwater recharge and evapotranspiration. It is seen from Figure 6 that most models simulate monthly soil moisture content quite well. In order to have a numeric comparison, the mean monthly relative error between the calculated soil moisture content \( W(t)_{\text{cal}} \) and the observed soil moisture content \( W(t)_{\text{obs}} \) is calculated, and the results are presented in Table IV. Figure 6 and Table IV reveal that: (1) All seven models simulate the mean annual soil moisture content well. Six out of the seven models have a maximum error of less than 5%. (2) For the mean monthly estimations, four (GG, Thornthwaite, Makkink and Priestley–Taylor) out of the seven models give a mean annual error of less than 10%. The worst case for simulating monthly soil moisture content is the CRAE model, followed by the AA and Hargreaves models. It should be kept in mind that the comparison is made without a recalibration of the model parameter values.

**Verification of the calculation and observation results**

According to the water balance concept (Equation (1)), the residual of precipitation minus evapotranspiration and groundwater recharge should be equal or close to zero over 1 year or a longer period. The annual and long-term residuals \( (P - E - \text{GWR}) \) are presented in Table V. A careful examination of Table V shows that: (1) For the annual residuals calculated by the seven models, fluctuations of up to 150 mm are obtained. This is mainly due to the fact that the calculation is based on a calendar year rather than a hydrological year. For example, in 1989 most models give a relatively large negative residual followed by a relatively large positive residual in 1990. This means that a water transfer from 1989 to 1990 can be expected. This problem becomes less important as the calculation period becomes longer. For the 12-year sum of residuals, five out of the seven models give zero or nearly zero values. For the AA and CRAE models, residuals of \(-85 \) mm and \(-54 \) mm respectively are obtained, after the twelfth year of calculation. It is anticipated that the residuals approach zero as the calculation period increases. (2) For the residuals calculated from the observed data, an unbalanced condition is obtained on an annual basis, with a general decreasing trend beginning in 1988. The reason for this systematic error in the measured data is to be investigated in a future study.

### Table IV. Comparison of mean monthly observed and calculated soil moisture content

<table>
<thead>
<tr>
<th>Month</th>
<th>( W_{\text{obs}} ) (mm)</th>
<th>( W_{\text{cal}} ) (%)</th>
</tr>
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<tr>
<td></td>
<td>AA</td>
<td>CRAE</td>
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* Error (%) = \[ \frac{W(t)_{\text{cal}} - W(t)_{\text{obs}}}{W(t)_{\text{obs}}} \times 100. 

Figure 6. Observed and computed monthly soil moisture content using the seven evapotranspiration models
Table V. Observed and computed annual residuals (RL = P − E − GWR) using the seven evapotranspiration models

<table>
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<tr>
<th>Year</th>
<th>Precipitation (mm)</th>
<th>RL_{obs} (mm)</th>
<th>Model RL (mm)</th>
<th>AA</th>
<th>CRAE</th>
<th>GG</th>
<th>Thornthwaite</th>
<th>Hargreaves</th>
<th>Makkink</th>
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<td>−51</td>
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SUMMARY AND CONCLUSIONS

The hydrological and meteorological data associated with lysimeter III at Mönchengladbach, Germany, have been used to verify seven models of water balance calculations. The water balance study is performed in three steps. First, actual evapotranspiration is computed using seven models that fall into two different approaches, namely the complementary relationship approach (three models) and the potential evapotranspiration approach (four models). Of the four potential evapotranspiration models, two are temperature-based methods (i.e. the Thornthwaite and Hargreaves methods) and other two are radiation-based methods (i.e. the Makkink and Priestley–Taylor methods). Second, groundwater recharge, expressed as soil water surplus in the water balance model, is calculated using measured precipitation and calculated or observed evapotranspiration. Third, the annual residual, defined as precipitation minus evapotranspiration and groundwater recharge, is computed using measured data and computed results from the above-mentioned methods. The following conclusions can be drawn from the study.

Where the complementary relationship methods are concerned, for calculating the actual evapotranspiration in the study region the GG-model gives the best estimates compared with the measured values of the lysimeter, followed by the AA-model. The CRAE model gives the worst results. When the potential evapotranspiration methods are used in calculating actual evapotranspiration, the radiation-based methods (Makkink and Priestley–Taylor) performed better than the temperature-based methods (Thornthwaite and Hargreaves). The best models from each approach performed equally well in calculating actual evapotranspiration, both on annual and monthly bases.

For calculating groundwater recharge using the water balance equation with measured precipitation and calculated evapotranspiration from the above methods, the GG-model and the AA model of the complementary relationship approach performed better than the other models, with mean annual errors of −5.7% and −4.4% respectively.

For calculating soil moisture content using the water balance equation with measured precipitation and calculated evapotranspiration from the above methods, four out of the seven models (GG, Thornthwaite, Makkink and Priestley–Taylor) give equally good results, with a maximum mean monthly error of less than 10%. The worst case for simulating monthly soil moisture content is the CRAE model, followed by the AA and Hargreaves models.
A careful examination of the residual (defined as precipitation minus evapotranspiration and groundwater recharge) reveals that five out of the seven models give zero or nearly zero values for the 12-year calculation period. It is anticipated that, as the calculation period becomes longer, the other two models (AA and CRAE) will also give a residual approaching zero. The study also shows that the residuals calculated from the observed data do not approach zero after the 12-year calculation period; a general decreasing trend beginning at 1988 is obtained, which is to be investigated in a future study. It should be kept in mind that the comparison is made without a recalibration of the model parameter values.

ACKNOWLEDGEMENTS

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