International Conference "Water, Environment, Energy and Society" (WEES–2009)
New Delhi, 12–16 January 2009

Evapotranspiration in Hydrological Modeling

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ABSTRACT: Potential evaporation is one of the important inputs to hydrological models and actual evapotranspiration is one of the important output components of hydrological models. In general evapotranspiration provides the second largest quantity in a hydrological water balance. Accurate spatial and temporal predictions of evapotranspiration are required for hydrologic models. Research on the development and application of evaporation calculation methods at different complexity and climates has been carried out since the 19th century. As a result, a great deal of experience on various models and methods has been gained. This paper reviews these methods and discusses flexibility with different data availabilities. The applicability and limitations of the methods are addressed. The main purpose of the paper is to provide the readers with more or less a complete list of operational methods that are widely used by hydrologists and water resources engineers for calculating free water evaporation, potential/reference evapotranspiration, and regional actual evapotranspiration.

INTRODUCTION

Four terms are usually used in describing evaporation 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. (2) The term actual evapotranspiration, ET_a, describes all the processes by which liquid water at or near the land surface becomes atmospheric water vapor under natural conditions. (3) The term potential evapotranspiration, ET_p, was first introduced in the late 1940s and 50s by Penman (Penman, 1948, 1956; Brutsaert, 1982) and it is defined as "the amount of water transpired in a given time by a short green crop, completely shading the ground, of uniform height and with adequate water status in the soil profile". Note that in the definition of potential evapotranspiration, the evapotranspiration rate is not related to a specific crop, and there are many types of horticultural and agronomic crops that fit into the description of short green crops. The fourth term, reference evapotranspiration, ET_{ref}, was introduced by irrigation engineers and researchers in the late 1970s and the early 1980s (e.g., Allen *et al.*, 1998) to avoid ambiguities that existed in the definition of potential evapotranspiration. Reference evapotranspiration is defined as "the rate of evapotranspiration from a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 sec m^{-1} and an albedo of 0.23, closely resembling the evapotranspiration from an extensive surface of green grass of uniform height, actively growing, well-watered, and completely shading the ground". The method overcomes shortcomings of the previous Penman method and provides reference values of potential evapotranspiration for a uniform grass reference surface worldwide and there is no need for a local calibration. In the reference evapotranspiration definition, the grass is specifically defined as the reference crop and this crop is assumed to be free of water stress and diseases. By adopting a reference crop (grass), it has become easier and more practical to select consistent crop coefficients and to make reliable actual crop evapotranspiration (ET_a) estimates in new areas. As water is abundantly available at the potential/reference evapotranspiring surface, soil factors do not affect evapotranspiration rate, ET_p and ET_{ref}. The only factors affecting ET_p and ET_{ref} are climatic parameters. Consequently, ET_p and ET_{ref} are a climatic parameter and can be computed from weather data. It is not uncommon to use an equation for determination of evaporation from open water that was actually developed for determination of potential evapotranspiration from vegetated lands, and vice versa (see also Winter *et al.,* 1995) although they are not the same as defined above. In the rest of this paper, no further distinguish is made when discussion the methods for calculating potential evapotranspiration and reference evapotranspiration as is true in most hydrological literatures. Of the above four terms, ET_o or ET_p or ET_{ref} is used as an important input, together with precipitation, to hydrological models and ET_a is one of the important outputs of hydrological models, since only ET_a is a water balance component.

There exists a multitude of methods for estimation of free water evaporation and/or potential evapotranspiration, (e.g., Jensen *et al.,* 1990; Allen *et al.,* 1998; Xu and Singh, 1998, 2002); the techniques are based on one or more atmospheric variables, such as air temperature, solar or net total radiation and humidity, or some measurement related to these variables, like pan evaporation (ET_{pan}). Some of these methods are accurate and reliable; others provide only a rough approximation. Most of the methods were developed for use in specific studies and are most appropriate for use in climates similar to where they were developed (Penman, 1948; Jensen, 1973). The Penman-Monteith (P-M) approach was recommended by FAO (see Allen *et al.,* 1998) as a standard to calculate reference evapotranspiration wherever the required input data are available.

In a broad definition, the actual evapotranspiration is a combined process of both evaporation from soil and plant surfaces and transpiration through plant canopies. In practice, the estimation of actual evapotranspiration rate for a specific crop requires first calculating potential or reference evapotranspiration (ET_p or ET_{ref}) and then applying the proper crop coefficients (K_c) to estimate actual crop evapotranspiration (ET_a). In conceptual hydrological modelling, the procedure for calculating actual evapotranspiration is also first to estimate ET_p or ET_{ref} based on meteorological factors, then compute the amount of that potential that is utilized by the actual evapotranspiration processes, given the current status of the plant- and soil-moisture-related characteristics (Xu and Singh, 2004). Several other methods have been proposed in the literature for calculating actual evapotranspiration. Monteith (1963, 1965) introduced resistance terms into the well-known method of Penman (1948) and derived at an equation for evapotranspiration from surfaces with either optimal or limited water supply. This method is often referred to as Penman-Monteith method. Another approach is the complementary relationship proposed by Bouchet (1963). For areal estimation, this method is usually preferred because it requires only standard meteorological variables and does not require local parameter calibration. Different models have been derived using the complementary relationship concept, which include the Advection-Aridity (AA) model proposed by Brutsaert and Stricker (1979), the Complementary Relationship Areal Evapotranspiration (CRAE) model derived by Morton (1978, 1983), and the complementary relationship model proposed by Granger and Gray (1989) using the concept of relative evapotranspiration (the ratio of actual to potential evapotranspiration).

The purposes of this paper are not to provide a deep discussion on the principle of evaporation theory, nor on measurement methods (to this end reader can refer to chapters 40 to 45 of the excellent book edited by Anderson and McDonnell, 2005). The sole purpose is to provide the readers with a more or less complete list of operational methods that are widely used by hydrologists and water resources engineers for calculating free water evaporation, potential/reference evapotranspiration, and regional actual evapotranspiration.

ESTIMATION OF FREE WATER EVAPORATION AND POTENTIAL/REFERENCE EVAPOTRANSPIRATION

The potential for evapotranspiration is usually defined as an atmospheric determined quantity. Therefore, techniques for estimating potential ET or ET_0 are based on one or more atmospheric variables, like solar or net radiation and air temperature and humidity, or some measurement related to these variables, like pan evaporation. Because climatic variables usually do not vary significantly over small areas, ET_p estimates can often be transferred some distance with minimal error. For most hydrologic applications, this is necessary because data are rarely available on the area where needed.

In the sections that follow many of the most commonly used techniques for estimating evaporation and potential evapotranspiration are described.

Climatological Methods

Air Temperature-Based Methods

In certain regions of the world, meteorological and climatological data may be quite limited. Models based almost solely on air temperature may be used in such cases to provide estimates of *ET*. The temperature methods are some of the earliest methods for estimating ET_p (Jensen *et al.*, 1990). If estimates are made for periods of several weeks or a month, reasonable approximations are possible. Most temperature-based equations take the form,

$$ET_P = cT^a \quad \text{... (1)}$$

or

$$ET_P = c_1 d_l T(c_2 - c_3 h) \quad \text{... (2)}$$

in which ET_P is potential evapotranspiration, T is air temperature, h is a humidity term, c_1, c_2, c_3 are constants, d_l is day-length. Many temperature-based equations have been developed and used. Some of the more common temperature-based models are described below. The following seven temperature-based equations each representing a special form of the equations (1) or (2) are discussed, namely: Thornthwaite (1948), Linacre (1977), Blaney-Criddle (1950), Hargreaves and Samni (1985), Kharrufa (1985), Hamon (1961), and Remanenko (1961) methods.

1. *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 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 Thornthwaite equation are discussed as follows:

Step 1: The annual value of the heat index I is calculated by summing monthly indices over a 12 months period. The monthly indices are obtained from the equation,

$$i = \left(\frac{T_a}{5}\right)^{1.51} \text{ and } I = \sum_{j=1}^{12} i_j \quad \text{... (3)}$$

in which I = annual heat index; i = monthly heat index for the month j, (which is zero when the mean monthly temperature is 0°C or less); T_a = mean monthly air temperature (degree Celsius); and j = number of months (1–12).

Step 2: The Thornthwaite general equation, Eq. 4a, calculates unadjusted monthly values of potential evapotranspiration, ET_P' (in mm), based on a standard month of 30 days, 12 hrs of sunlight/day,

$$ET_P' = C\left(\frac{10T_a}{I}\right)^a \quad \text{... (4a)}$$

in which C = 16 (a constant); and $a = 67.5 \times 10^{-8} I^3 - 77.1 \times 10^{-6} I^2 + .0179I + .492$.

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' is zero for temperature below zero degree Celsius.

Step 3: The unadjusted monthly evapotranspiration values ET_P' are adjusted depending on the number of days N in a month $(1 \le N \le 31)$ and the duration of average monthly or daily daylight d (in hr) which is a function of season and latitude,

$$ET_P = ET_P'\left(\frac{d}{12}\right)\left(\frac{N}{30}\right) \quad \text{... (4b)}$$

in which ET = adjusted monthly potential evapotranspiration (mm); d = duration of average monthly daylight (hr); and N = number of days in a given month, 1–31 (days).

Thornthwaite's equation has been widely criticized for its empirical nature but is widely used. Because Thornthwaite's method of estimating ET_p can be computed using only temperature, it has been one of the most misused empirical equations in arid and semi-arid irrigated areas where the requirement has not been maintained (Thornthwaite and Mather, 1955).

2. *Linacre Method*

For the case of well-watered vegetation with an albedo of about 0.25, Linacre (1977) simplified Penman formula to give the following expression for the evaporate rate,

$$ET_P = \frac{500T_m/(100-A) + 15(T_a - T_d)}{(80 - T_a)} \quad \text{... (5)}$$

where ET_P = Linacre potential evapotranspiration in mm/d, $T_m = T + 0.006h$, h is the elevation (meters). A is the latitude (degrees) and T_d is the mean dew-point temperature. T_a, T_m and T_d are in °C. This formula requires only geographical data (A and h), the mean and the dew-point temperature.

3. *Blaney-Criddle Method*

The Blaney-Criddle (1950) procedure for estimating ET_P is well known in the western USA and has been used extensively elsewhere also (Singh, 1989). The usual form of the Blaney-Criddle equation converted to metric units is written as,

$$ET_P = kp(0.46T_a + 8.13) \quad \text{... (6)}$$

where ET_p is potential evapotranspiration from reference crop, in mm, for the period in which p is expressed. T_a is mean temperature in °C, p is percentage of total daytime hours for the used period (daily or monthly) out of total daytime hours of the year (365 × 12), and k is monthly consumptive use coefficient, depending on vegetation type, location and season. According to Blaney-Criddle, for the growing season (May to October) k varies from 0.5 for orange tree to 1.2 for dense natural vegetation.

4. *Kharrufa Method*

Kharrufa (1985) derived an equation through correlation of ET_P and T_a in the form of,

$$ET_P = 0.34\, pT_a^{\ 1.3} \qquad \text{... (7)}$$

where ET_P = Kharrufa potential evapotranspiration in mm/month, T_a and p have the same definitions as given in Eq. (6).

5. *Hargreaves Method*

Hargreaves and Samani (1982, 1985) proposed several improvements for the Hargreaves (1975) equation for estimating grass-related reference ET. Because solar radiation data frequently are not available, Hargreaves and Samani (1982, 1985) recommended estimating R_s from extraterrestrial radiation, R_A, and the difference between mean monthly maximum and minimum temperatures, TD in °C. The resulting form of the equation is,

$$ET_P = 0.0023\, R_A\, TD^{1/2}\, (T_a + 17.8) \qquad \text{... (8)}$$

The extra terrestrial radiation, R_A, is expressed in equivalent evaporation units. For a given latitude and day R_A is obtained from tables or may be calculated using a set of equations (see Jensen *et al.,* 1990, page 179). The only variable for a given location and time period is air temperature. Therefore, the Hargreaves method has become a temperature-based method.

6. *Hamon Method*

Hamon (1961) derived a potential evapotranspiration method based on the mean air temperature and is expressed as,

$$ET_P = 0.55\, D^2\, VP \qquad \text{... (9)}$$

where ET_P is potential evapotranspiration in inch/day, D is the hours of daylight for a given day in units of 12 hr, and VP is a saturated water vapour density term calculated by,

$$VP = \frac{4.95 e^{(0.062 T_a)}}{100} \qquad \text{... (10)}$$

where T_a is daily mean air temperature in °C.

7. *Remanenko Method*

Remanenko (1961) derived an evaporation equation based on the relationship using mean temperature and relative humidity,

$$ET_P = 0.0018\, (25 + T_a)^2\, (100 - RH) \qquad \text{... (11)}$$

where T_a is the mean air temperature in °C, RH is the mean monthly relative humidity, which is calculated by,

$$RH = \frac{e^{\circ}(T_d)}{e^{\circ}(T_a)} \qquad \text{... (12)}$$

in which $e°(T)$ is the vapour pressure calculated by (see Bosen, 1960),

$$e^{\circ}(T) = 33.8679 \left[(.00738T + .8072)^8 - .000019\left|1.8T + 48\right| + .001316 \right] \qquad \text{... (13)}$$

A comparative study of the above discussed temperature-based methods was done by Xu and Singh (2001).

Solar Radiation-Based Methods

The radiation-based approach has had wide application in estimation of potential evapotranspiration (ET_P) of land areas. Many empirical formulae have been derived based on this approach (Jensen *et al.,* 1990; Xu and Singh, 2000). Certain methods based on solar radiation also involve a temperature term.

Empirical radiation-based equations for estimating potential evaporation generally are based on the energy balance (Jensen *et al.,* 1990). Most radiation-based equations take the form,

$$\lambda ET_P = C_r(wR_s) \quad \text{or} \quad \lambda ET_P = C_r(wR_n) \qquad \text{... (14)}$$

where λ is the latent heat of vaporisation (in calories per gram), ET_P is the potential evapotranspiration (in mm per day), R_s is the total solar radiation (in calories per cm^2 per day), R_n is the net radiation (in calories per cm^2 per day), w is the temperature and altitude-dependent weighting factor, and C_r is a coefficient depending on the relative humidity and wind speed. Some well-known radiation based methods are presented in this section. For a more complete discussion, the reader is referred to the cited literature.

1. *Turc Method*

Under general climatic conditions of western Europe, Turc (1961) computed ET_P in millimetres per day for 10 day periods as,

$$ET_P = 0.013\frac{T_a}{T_a+15}(R_s+50) \quad \text{for } RH \geq 50 \qquad \text{... (15)}$$

$$ET_P = 0.013\frac{T_a}{T_a+15}(R_s+50)\left(1+\frac{50-RH}{70}\right) \quad \text{for } RH < 50 \qquad \text{... (16)}$$

where T_a is the air temperature in °C, R_s is the total solar radiation in cal/cm^2/day, and RH is the relative humidity in percentage.

2. *Makkink Method*

Makkink (1957) estimated ET_P in millimetres per day over 10 day periods for grassed lands under cool climatic conditions of the Netherlands as,

$$ET_p = 0.61\frac{\Delta}{\Delta+\gamma}\frac{R_s}{58.5} - 0.012 \qquad \text{... (17)}$$

where Δ is the slope of saturation vapour pressure curve (in mb/°C), γ (in mb/°C) is the psychromatic constant. These quantities are calculated as (see also Singh, 1989),

$$\Delta = 33.8639\,[0.05904\,(0.00738\,T_a + 0.8072)^7 - 0.0000342] \qquad \text{... (18)}$$

$$\gamma(\text{mb/°C}) = \frac{c_p P}{0.622\lambda} \qquad \text{... (19)}$$

$$\lambda\,(\text{cal/g}) = 595 - 0.51T_a \qquad \text{... (20)}$$

$$P = 1013 - 0.1055\,EL \qquad \text{... (21)}$$

where EL is elevation (in metres), λ (in calories per gram) is latent heat, and P (in millibar) is atmospheric pressure. The specific heat of air c_p (in cal/g/°C) varies slightly with atmospheric pressure and humidity, ranging from 0.2397 to 0.260. An average value of 0.242 is reasonable.

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

$$ET_P = 0.7\frac{\Delta}{\Delta+\gamma}\frac{R_s}{\lambda} \qquad \text{... (22)}$$

where all the notations have the same meaning and units as in (17).

3. *Jensen-Haise Method*

Jensen and Haise (1963) evaluated 3000 observations of ET as determined by soil sampling procedures over a 35 years period, and developed the following relation,

$$\lambda ET_p = C_t(T_a - T_x)R_s \qquad \text{... (23)}$$

where λ and R_s have the same meaning and units as before, ET_p is in mm/day, C_T (temperature constant) = 0.025, and T_x = –3 when T_a is in degree Celsius. These coefficients were considered to be constant for a given area.

4. *Hargreaves Method*

Hargreaves (1975) and Hargreaves and Samani (1982, 1985) proposed several equations for calculating potential evapotranspiration, ET_p (in mm/day). One of the equations is written as,

$$\lambda ET_p = 0.0135(T_a + 17.8)R_s \qquad \text{... (24)}$$

All variables have the same meaning and units as before. The Hargreaves method was derived from eight years of cool season Alta fescue grass lysimeter data from Davis, California.

5. *Doorenbos and Pruitt Method*

Doorenbos and Pruitt (1977) presented a radiation method for estimating ET_p using solar radiation. The method is an adaptation of the Makkink (1957) method and was recommended over the Penman method when measured wind and humidity data were not available or could not be estimated with reasonable confidence,

$$ET_p = a\left[\frac{\Delta}{\Delta+\gamma}R_s\right] + b \qquad \text{... (25)}$$

where R_s is solar radiation in mm/day, b = –0.3 mm/day and a is an adjustment factor that varies with mean relative humidity and daytime wind speed. The adjustment factor a was presented in graphic and tabular forms, and can also be calculated from,

$$a = 1.066 - 0.13\times10^{-2}RH + 0.045U_d - 0.20\times10^{-3}RH\times U_d - 0.315\times10^{-4}RH^2 - 0.11\times10^{-2}U_d^2 \qquad \text{... (26)}$$

where RH is the mean relative humidity in percentage and U_d is the mean daytime wind speed in m/s.

6. *McGuinness and Bordne Method*

McGuinness and Bordne (1972) proposed a method for calculating potential evapotranspiration based on an analysis of a lysimeter data in Florida,

$$ET_p = \{(0.0082T_a - 0.19)(R_s/1500)\}2.54 \qquad \text{... (27)}$$

where ET_p is in cm/day for a monthly period, T_a is in degrees Fahrenheit, and R_s is in cal/cm^2/day.

7. *Abtew Method*

Abtew (1996) used a simple model that estimates ET_p from solar radiation as follows,

$$ET_p = K\frac{R_s}{\lambda} \quad \text{... (28)}$$

where ET_p is in mm/day, R_s is in $MJm^{-2}d^{-1}$, λ is in $MJ\ Kg^{-1}$, and K is a dimensionless coefficient.

8. *Priestley and 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 evaporation, ET_p. The aerodynamic component was deleted and the energy component was multiplied by a coefficient, α = 1.26, when the general surrounding areas were wet or under humid conditions.

$$ET_p = \alpha\frac{\Delta}{\Delta+\gamma}\frac{R_n}{\lambda} \quad \text{... (29)}$$

where R_n is the net radiation (cal $cm^{-2}d^{-1}$), and other notations have the same meaning and units as in equation (17).

A comparative study of the above discussed methods is done by Xu and Singh (2000).

The Penman Combination Method

Penman (1948) was among the first to develop a method considering the factors of both energy supply and turbulent transport of water vapour away from an evaporating surface. The physical principles combine the two approaches, i.e. the mass-transfer and the energy balance. The basic equations are later modified and rearranged to use meteorological constants and measurements of variables made regularly at climatological stations. Following Shaw (1989), the Penman equation has the form of,

$$E_0 = ET_p = \frac{\Delta}{\Delta+\gamma}H + \frac{\gamma}{\Delta+\gamma}E_a \quad \text{... (30)}$$

where H is the available heat energy, which is the sum of energy for evaporation (latent heat flux) and energy for heating the air (sensible heat flux). E_a is the drying power of the air which is expressed as,

$$E_a = 0.35(0.5 + u_2/100)(e_a - e_d) \quad \text{... (31)}$$

where:

T_a mean air temperature for a week, 10 days or a month, °F or °C

e_d mean vapor pressure for the same period, mm of mercury

n bright sunshine over the same period, $h\ day^{-1}$

u_2 mean wind speed at 2 m above the surface, miles day^{-1}

Finally, a value of Δ is found from the curve of saturated vapor pressure against temperature corresponding to the air temperature, T_a.

Penman-Monteith Method

The Penman combination method (Eq. 30) was further developed by many researchers including the excellent work done by Monteith (1963, 1964). The resulted Penman-Monteith method has been recommended as the standard method for determining ET_{ref} by FAO (Allan *et al.,* 1998) (Eq. 32). The method has been selected because it is physically based and explicitly incorporates both physiological and aerodynamic parameter,

$$ET_{ref} = \frac{0.408\Delta(R_n - G) + \gamma\frac{900}{T_a+273}u_2(e_s - e_a)}{\Delta + \gamma(1+0.34u_2)} \quad \text{... (32)}$$

where: ET_{ref} = reference evapotranspiration [mm day^{-1}], R_n = net total radiation at the crop surface [MJ $m^{-2}\ day^{-1}$], G = soil heat flux density [MJ $m^{-2}\ day^{-1}$], T = mean daily air temperature at 2 m height [°C], u_2 = wind speed at 2 m height [ms^{-1}], e_s = saturation vapour pressure [kPa], e_a = actual vapour pressure [kPa], $e_s - e_a$ = saturation vapour pressure deficit [kPa], Δ = slope of the vapour pressure [kPa $°C^{-1}$], γ = psychrometric constant [kPa $°C^{-1}$].

Apart from the site location, the FAO Penman-Monteith equation requires air temperature, humidity, radiation and wind speed data for daily, weekly, ten-day or monthly calculations (Xu *et al.,* 2006). The procedure for using Eq. (32) for computing reference evapotranspiration has been given in Chapter 3 of the FAO paper 56 (Allen *et al.,* 1998).

Micrometeorological Methods

The Mass-Transfer-Based Methods

The mass-transfer method is one of the oldest methods (Dalton, 1802; Meyer, 1915; Penman, 1948) and is still an attractive method in estimating free water surface evaporation, ET_0, because of its simplicity and reasonable accuracy. The mass-transfer methods are based on the Dalton equation which for free water surface can be written as,

$$ET_0 = C(e_s - e_a) \quad \text{... (33)}$$

where ET_0 is free water-surface evaporation, e_s is the saturation vapor pressure at the temperature of the water surface, e_a is the actual vapor pressure in the air, and C is an empirically determined constant involving some function of windiness.

Therefore, equation (33) is expressed as,

$$ET_0 = f(u)(e_s - e_a) \quad \text{... (34)}$$

where $f(u)$ is the wind function. This function depends, among other factors, on the observational heights of the wind speed and vapor pressure measurements. Although the two heights need not be the same, the same experimental layout must be used for a particular value of the function. The mass-transfer method has had wide application in the estimation of lake evaporation and many empirical formulae have been derived based on this approach (Singh, 1989). Examples of empirical equations of this type are included in Table 1. An evaluation and comparison of mass-transfer methods was performed by Singh and Xu (1997a, b).

The wind speed (monthly mean) u is measured in miles per hour and vapor pressure e, in inches of Hg. The subscripts attached to u refer to height in meters at which the measurements are taken; no subscript refers to measurements near the ground or water surface.

Aerodynamic Method

Thornthwaite and Holzman (1942) were among the first modern micrometeorologists to apply the aerodynamic approach to measurement of ET_p. They proposed a relationship involving the gradients of specific humidity q and the logarithmic wind profile. Their expression, given here without derivation, is,

$$ET_p = \rho_a k^2 \frac{(q_2 - q_1)(U_2 - U_1)}{\ln(z_2 / z_1)^2} \quad \text{... (35)}$$

where ρ_a = density of moist air, k = von Karman's constant. Over a rough cropped surface $z - d$ is substituted for z. An error analysis of this method is given by Thompson and Pinker (1981).

Table 1: Some Mass-transfer-based Evaporation Equations for Estimation of Evaporation (Singh and Xu, 1997a)

Sl. No.	*Author*	*Equation*	*Remarks*
1.	Dalton (1802)	ET_0 (in./mo) = $a(e_s - e_a)$	a = 15 for small, shallow water, and a = 11 for large deep water
2.	Fitzgerald (1886)	ET_0 (in./mo) = $(.4 + .199u)(e_s - e_a)$	
3.	Meyer (1915)	ET_0 (in./mo) = $11(1+.1u)(e_s - e_a)$	e_a is measured at 30 ft above the surface
4.	Horton (1917)	ET_0 (in./mo) = $.4[(2-\exp(-2u))(e_s - e_a)]$	
5.	Rohwer (1931)	ET_0 (in./da) = $.77(1.465 - .0186p_b)\cdot(.44 + .118u)(e_s - e_a)$	p_b = barometric pressure in in. of *Hg*.
6.	Penman (1948)	ET_0 (in./da) = $.35(1+.24u_2)(e_s - e_a)$	
7.	Harbeck *et al.* (1954)	ET_0 (in./da) = $.0578u_8(e_s - e_a)$ ET_0 (in./da) = $.0728u_4(e_s - e_a)$	
8.	Kuzmin (1957)	ET_0 (in./mo) = $6.0(1+.21u_8)(e_s - e_a)$	
9.	Harbeck *et al.* (1958)	ET_0 (in./da) = $.001813u(e_s - e_a)(1-.03(T_a - T_w))$	T_a = average air temperature °C +1.9°C; T_w = average water surface temperature °C.
10.	Konstantinov (1968)	ET_0 (in./da) = $.024(t_w - t_2)/u_1 + .166u_1)(e_s - e_a)$	
11.	Remanenko (1961)	ET_0 (cm/mo) = $.0018(T_a + 25)^2(100 - hn)$	hn = relative humidity
12.	Sverdrup (1946)	$ET_0(in./h) = \dfrac{.623\rho K_0^2(u_8 - u_2)(e_2 - e_8)}{p[\ln(800/200)]^2}$	K_0 = von Karman's const ρ = density of air p = atmospheric pressure
13.	Thornthwaite & Holzman (1939)	$ET_0(in./h) = \dfrac{.623\rho K_0^2(u_8 - u_2)(e_2 - e_8)}{p[\ln(800/200)]^2}$	

Following Thornthwaite and Holzman's work, many others (e.g., Pasquil, 1950; Pruitt, 1963; Dyer, 1974) have proposed stability-corrected aerodynamic methods for estimating the flux of vapor. Aerodynamic methods require stringently accurate observations of wind speed and specific humidity or vapour pressure at a number of heights above the surface, as well as temperature measurement to permit stability corrections to be made. Because of its origins in classical fluid dynamics theory, aerodynamic methods have been popular with scientists. However, the methods have not reached a degree of development that makes them applicable for routine use, for example, in hydrological modeling.

Bowen Ratio-energy Balance Method

Bowen (1926) introduced a relationship between latent heat flux, λE and sensible heat flux, H known as the Bowen ratio β. This is defined by,

$$\beta = \frac{H}{\lambda E} = \frac{PC_p}{\lambda}\left(\frac{M_a}{M_w}\right)\left(\frac{K_h}{K_w}\right)\frac{\partial T/\partial z}{\partial e/\partial z} = \gamma\frac{K_h}{K_w}\frac{\partial T/\partial z}{\partial e/\partial z} \quad \text{... (36)}$$

where M_w and M_a are the molecular weights of water vapor and air, K_h and K_w are the turbulent exchange coefficients for sensible heat and water vapor. Other notations are previously defined.

This relationship is generally simplified by assuming that the turbulent exchange coefficient for heat transport K_h = the exchange coefficient for water vapor transport K_w and that $(\partial T/\partial z)/(\partial e/\partial z) \approx \Delta T/\Delta e$ where $\Delta T = T_2 - T_1$, and $\Delta e = e_2 - e_1$. Equation (36) then becomes,

$$\beta \approx \gamma\frac{\Delta T}{\Delta e} \quad \text{... (37)}$$

a simplified form of energy balance equation at the earth's surface can be written as,

$$R_n + S + \lambda E + H = 0 \quad \text{... (38)}$$

From (36), $H = \beta\lambda E$. Substitution into (38) and solution for λE yields,

$$\lambda E = -\frac{R_n + S}{1+\beta} = -\left[\frac{R_n + S}{1+\gamma\Delta T/\Delta e}\right] \quad \text{... (39)}$$

Equation (39) is the so called "Bowen Ratio-Energy Balance" (BREB) method of estimating λE.

The Pan Method

Measured evaporation from a shallow pan of water is one of the oldest and common methods for estimating ET_0. It is an indirect integration of the principal atmospheric variables related to ET_0. Pans are inexpensive, relatively easy to maintain and simple to operate in regions where frost is not a significant factor. Using pan evaporation for estimating reference evapotranspiration is another common method, especially in American and Asian countries (Golubev *et al.,* 2001; Liu *et al.,* 2004). The evaporation rate from pans filled with water is easily obtained. In the absence of rain, the amount of water evaporated during a period (mm/day) corresponds with the decrease in water depth in that period. Pans provide a measurement of the integrated effect of radiation, wind, temperature and humidity on the evaporation from an open water surface. Although the pan responds in a similar fashion to the same climatic factors affecting crop transpiration, several factors produce differences in loss of water from a water surface and from a cropped surface (e.g., Allen *et al.,* 1998): "Reflection of solar radiation from water in the shallow pan might be different from the grass reference surface. Storage of heat within the pan can be appreciable and may cause significant evaporation during the night while most crops transpire only during the daytime. There are also differences in turbulence, temperature and humidity of the air immediately above the respective surfaces. Heat transfer through the sides of the pan occurs and affects the energy balance". Notwithstanding the difference between pan-evaporation and evapotranspiration of cropped surfaces, the pan has proved its practical value and has been widely used to estimate reference evapotranspiration by applying empirical coefficients to relate ET_{pan} to ET_{ref} for periods of 10 days or longer (Allen *et al.,* 1998). In humid regions, pans may give realistic estimates of potential evapotranspiration, ET_p. However, care must be taken in relating evaporation from pans to ET in arid climates (Rosenberg *et al.,* 1983, page 262). Given some standardization of pan shape, environmental setting, and operation, good correlations have been developed between pan evaporation, ET_{pan}, and potential evaporation, ET_p, by a simple relation,

$$ET_p = C_{ET}E_{pan} \quad \text{... (40)}$$

where C_{ET} is a coefficient.

Pan-to-ET coefficients (C_{ET}) are necessary because evaporation for a pan is generally more than for a well-wetted vegetated surface, or even a pond, due to the pan's excessive exposure and lower reflectance of solar radiation (Allen *et al.,* 1998). The values of C_{ET} vary normally from 0.5 to 1.0. The actual value depends, among other factors, on the type of pan, the location of the measurement, and the season. Although specific coefficient values for application to any given situation or pan may have to be found by calibration, mean monthly values are usually shown in a table or graphically shown in a map for some major meteorological stations or regions.

ESTIMATION OF ACTUAL EVAPOTRANSPIRATION

A large number of methods have been developed in recent years for actual evapotranspiration, ET_a predictions each has its own requirements and emphases. The available methods range from quite simple to very complex. There are a few formulas which are more common than others. In this section will we discuss some commonly used methods in hydrological modeling studies.

Actual Evapotranspiration Calculated From Potential Evapotranspiration

In hydrological modeling studies, many investigators have found it necessary to derive "actual" evapotranspiration as a function of potential evapotranspiration and the dryness of the soil (Palmer, 1965; Saxton and McGuinness, 1982; Dyck, 1983). As the model storage ratio (actual soil moisture storage divided by the maximum storage) is representative of the 'wetness' of the soil, it would be conceptually acceptable to extract moisture at the potential rate when the storage was full, that is at field capacity, and reduce the extraction to zero when the storage was empty (when the soil moisture deficit had reached its maximum). However, the nature of the function that estimates actual evapotranspiration for conditions between these limits is not known. A number of functions operating between the limits of potential rate and zero have been tried by a number of modellers. A general form of such equations can be shown as,

$$ET_a = ET_p \cdot f(SMT / SMC) \qquad \text{... (41)}$$

where SMT is the actual soil moisture storage and SMC is the soil moisture storage at field capacity. Dyck (1983) provided a summary of some moisture extraction functions used by different investigators. Mintz and Walker (1993) also illustrated several moisture extraction functions. Many researchers agree on the general pattern of the soil's behavior that moisture is extracted from the soil at the potential rate until some critical moisture content is reached when evapotranspiration is no longer controlled by meteorological conditions. Below this critical moisture content, there is a linear decline in soil moisture extraction until the wilting point is reached. This type of behavior is illustrated by Shuttleworth (1993) and Dingman, (1994). Shuttleworth (1993) notes that the critical moisture content divided by the field capacity is typically between 0.5 and 0.8. This type of moisture extraction function is also used in the HBV model (e.g. Bergström, 1995) where actual evapotranspiration is computed as,

$$ET_a = ET_p \frac{SMT}{LP \cdot FC} \qquad \text{... (42)}$$

where FC is the field capacity and LP is a parameter ranging 0.5 to 0.8.

There are several drawbacks to using simple soil moisture extraction functions. Mintz and Walker (1993), cite field studies that show $f(SMT/SMC)$ may vary not only for a given soil wetness but may also vary with leaf-area index. In addition, it is difficult to determine the spatial variation of the water-holding capacity. A new and surely better approach to determine the relationship between plant transpiration and potential evapotranspiration is to correlate $f(SMT/SMC)$ with satellite-derived indices of vegetation activity so that $f(SMT/SMC)$ will reflect the plant growth stage and spatial vegetation patterns. Some examples for $f(SMT/SMC)$ and other functions are given in Table 2.

Thornthwaite Water Balance Approach

Thornthwaite-Mather Soil-water budget calculations approach (Thornthwaite and Mather, 1955; Xu and Chen, 2005) is commonly made using monthly totals because of its easy availability. The use of daily values is preferred over monthly values when possible particularly in dry locations where the mean potential evaporation for a given month may be higher than the mean precipitation, yet there is observed run-off.

The simiplified water balance calculation procedure is as follows:

- If $ET_p(t) > P(t)$, 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/FC.
- When $ET_p(t) = P(t)$, then $ET_a(t) = ET_p(t)$, Surplus = 0.
- If $ET_p(t) < P(t)$, then $ET_a(t) = ET_p(t)$. $W(t)$ is first estimated with Surplus = 0. When $W(t) > FC$, Surplus = $W(t) - FC$; when $W(t) <= FC$, Surplus = 0.

Table 2: Some Examples for Function $f(SMT / SMC)$ (updated from Dyck, 1983)

Reference	*f(SMT/SMC) =*
DAILY VALUES	
Minhas *et al.* (1974)	$\dfrac{1-\exp(-\gamma SMT)}{1-2\exp(-\gamma SMC)+\exp(-\gamma SMT)}$
Norero (1969)	$\left[1+(SMT/SMC)^{b \cdot k}\right]^{-1}$ $k = 2.69\exp(-0.09PET)^{-0.62}$
Baier & Robertson (1966)	$\sum_{j=1}^{n} k_j(SMT_{j,i-1}/SMC_j)Z_j$
Bergström (1995)	$\dfrac{SMT}{LP}$ (in the following equations $RAT = SMT/SMC$)
Roberts (1978)	$(RAT)^{0.5}$ $RAT^2/((RAT^2)+(1-RAT)^2)$ $2\times RAT^2 \times (1/(1+RAT)^{RAT})$ $2\times RAT\times(1/(1+RAT)^{RAT})$ $RAT^{1/2}+(RAT^{1/2}-RAT)$ RAT^2 RAT
5-DAY VALUES	
Renger *et al.* (1974) see Dyck, 1983	$0.2+2.0\times RAT-1.2\times RAT^2$
MONTHLY VALUES	
Budyko & Zubenok (1961)	RAT
Xu *et al.* (1996)	$1-a_1^{((SMT_{i-1}+P_i)/ET)}$

Where: SMT = actual soil moisture; SMC = soil moisture at field capacity; $SMT_{j,i-1}$ = actual soil moisture in the j–th zone at the end of the previous day ($i-1$); Z_j = fraction of available soil moisture at which $ET < PET$ and plant stress sets in; K_j = fraction of soil moisture extraction at that zone; γ = free parameter; b = soil specific constant; M = vegetation canopy density.

If the initial soil moisture is unknown, which is typically the case, 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 ($N + 1$). The initial soil moisture at time 1 ($w(1)$) is then set equal to the soil moisture at time $N + 1$ ($w(N + 1)$) and the budget is re-computed 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 field capacity of soil, FC. Field capacity 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).

Complementary Relationship Methods

Another approach is the complementary relationship proposed by Bouchet (1963). Utilizing an analysis based on 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, i.e. evaporative capacity increased, while the actual evapotranspiration decreased as the available water decreased. The relationship that he derived has come to be 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; the potential evapotranspiration thus ranges from its value at saturation to twice this value. This relationship is described as,

$$ET_a = 2ET_w - ET_p \quad \text{... (43)}$$

where ET_a, ET_p and ET_w are actual, potential and wet environment evapotranspiration, respectively. For areal estimation, this method is usually preferred because it requires only standard meteorological variables and does not require local parameter calibration. Different models have been derived using the complementary relationship concept, the methods differ in the way the ET_p and ET_w are calculated. The methods include the Advection-Aridity (AA) model proposed by Brutsaert and Stricker (1979), the Complementary Relationship Areal Evapotranspiration (CRAE) model derived by Morton (1978, 1983), and the complementary relationship model proposed by Granger and Gray (1989) using the concept of relative evapotranspiration (the ratio of actual to potential evapotranspiration). Although the above three models are derived using the complementary relationship concept, the assumptions and derived model forms are different. Besides the above cited references, there are a number of studies on evaluating the validity of the complementary relationship model (e.g., Doyle, 1990; Lemeur and Zhang, 1990; Chiew and McMahon, 1991; Granger and Gray, 1990; Hobbins *et al.*, 2001a,b; Xu and Li, 2003; Xu and Chen, 2005; Xu and Singh, 2005). For a more complete discussion, the reader is referred to the cited literature.

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