5 Evapotranspiration

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5.1 Introduction

Evapotranspiration is important as a term in the hydrological cycle, e.g. in soil water and groundwater balances (Chapter 16), and in salinization (Chapter 15). In land drainage engineering, we therefore need to devote proper attention to its determination, particularly in arid and semi-arid areas. This applies not only to the various surveys and investigations that precede a drainage design, but also to the subsequent monitoring of the effects of drainage measures on parameters like watertable depth, soil salinity, and, ultimately, on crop yield.

In addition, agriculturists want to have information on the effects of a water supply on crop production. As there is often a direct relation between the ratio of actual to potential evapotranspiration and actual to potential crop yield, agriculturists want to know the specific water requirements of a crop, and whether these requirements are being met under the prevailing environmental conditions. Regular estimates of evapotranspiration may reveal water shortages and/or waterlogging, which can then lead to technical measures to improve irrigation and drainage, and, again ultimately, to an increase in crop yields.

This chapter, after explaining some basic concepts (in Section 5.2), provides brief information on how to measure actual evapotranspiration in the field and on how to estimate the evaporative demand of the atmosphere. Actual evapotranspiration can be measured with the soil water balance approach, or with micro-meteorological methods. These will be briefly discussed in Section 5.3. Actual evapotranspiration can also be estimated with computer models or remote-sensing techniques (Section 5.6.4).

A few empirical, temperature-based methods for estimating potential evapotranspiration are briefly discussed (Section 5.4). The theory of Penman's open water evaporation is treated fairly extensively in Section 5.5. This is followed by the recently accepted Penman-Monteith method of estimating the potential evapotranspiration from cropped surfaces, distinguishing between wet and dry crops, between full and partial soil cover, and between full and limited water supply (Section 5.6). How the preceding theory is applied in practice is explained in Section 5.7, with the use of a reference evapotranspiration and crop coefficients.

5.2 Concepts and Developments

In the past, many empirical equations have been derived to calculate potential evapotranspiration (i.e. evapotranspiration from cropped soils with an optimum water

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supply). Only two of these methods will be described: one based on air temperature and day length (Blaney and Criddle 1950), and another based on air temperature and solar radiation (Turc 1954; Jensen and Haise 1963).

These empirical correlation methods are often only valid for the local conditions under which they were derived; they are hardly transferable to other areas. Nowadays, therefore, the focus is mainly on physically-based approaches, which have a wider applicability.

For the process of evapotranspiration, three basic physical requirements in the soil-plant-atmosphere continuum must be met. There must be:

A: A continuous supply of water;

B: Energy available to change liquid water into vapour;

C: A vapour gradient to maintain a flux from the evaporating surface to the atmosphere.

The various methods of determining evapotranspiration are based on one or more of these requirements. For example, the soil water balance approach is based on A, the energy balance approach on B, and the combination method (energy balance plus heat and mass transfer) on parts of B and C.

Penman (1948) was the first to introduce the combination method. He estimated the evaporation from an open water surface, and then used that as a reference evaporation. Multiplied by a crop factor, this provided an estimate of the potential evapotranspiration from a cropped surface.

The combination method requires measured climatic data on temperature, humidity, solar radiation, and wind speed. Because even this combination method contains a number of empirical relationships, numerous modifications to adjust it to local conditions have been proposed by a host of researchers.

Analyzing a range of lysimeter data worldwide, Doorenbos and Pruitt (1977) proposed the FAO Modified Penman method, which has found worldwide application in irrigation and drainage projects. These authors adopted the same two-step approach as Penman to estimate crop water requirements (i.e. estimating a reference evapotranspiration, selecting crop coefficients per crop and per growth stage, and then multiplying the two to find the crop water requirements). They replaced Penman's open water evaporation by the evapotranspiration from a reference crop. The reference crop of Doorenbos and Pruitt was defined as 'an extended surface of an 8 to 15 cm tall green grass cover of uniform height, actively growing, completely shading the ground, and not short of water'. There was evidence, however, that the method sometimes over-predicted the crop water requirements.

Using similar physics as Penman did, Monteith (1965) derived an equation that describes the transpiration from a dry, extensive, horizontally-uniform vegetated surface, which is optimally supplied with water. In international literature, this equation is known as the Penman-Monteith equation. In The Netherlands, the name of Rijtema has been added, because this author independently derived a similar formula (Rijtema 1965).

Recent comparative studies (e.g. those by Jensen et al. 1990, who analyzed various methods of estimating potential evapotranspiration) have shown the convincing performance of the Penman-Monteith approach under varying climatic conditions, thereby confirming the results of many individual studies reported over the past years.

An expert consultation on procedures to revise the prediction of crop water

Table 5.1 Meteorological and crop input data that are required for the various computation methods of potential evapotranspiration

Method	Rainfall	Air temperature	Solar radiation	Relative humidity	Wind speed	Aero-dynamic resistance	Basic canopy resistance
Blaney and Criddle (1950)		+					
Jensen and Haise (1963)		+	+				
Turc (1954)	+	+	+				
Penman (1948)		+	+	+	+	+	
Penman-Monteith (1965)		+	+	+	+	+	+

requirements was held in Rome (Smith 1990). There, it was agreed to recommend the Penman-Monteith approach as the currently best-performing combination equation. Potential and actual evapotranspiration estimates would, in principle, be possible with the Penman-Monteith equation, through the introduction of canopy and air resistances to water vapour diffusion.

This direct, or one-step, approach is increasingly being followed nowadays, especially in research environments. Nevertheless, since accepted canopy and air resistances may not yet be available for many crops, a two-step approach is still recommended under field conditions.

The reference crop evapotranspiration in the Penman-Monteith approach is defined as 'the evapotranspiration from a hypothetical crop fully covering the ground, and not short of water, with an assumed crop height of 12 cm, a fixed canopy resistance (70 s/m), and a canopy reflection coefficient of 0.23'. Details of the various parameters to be used in estimating this new reference evapotranspiration were worked out during the Rome meeting and are presented in Section 5.7.2.

The method selected to estimate potential evapotranspiration often depends on what meteorological data are available; the empirical approaches need fewer data than the physically-based methods. Table 5.1 indicates the meteorological input data that are needed for the computation methods discussed in this chapter.

5.3 Measuring Evapotranspiration

5.3.1 The Soil Water Balance Method

Both potential and actual evapotranspiration can be measured with the soil water balance method. The water balance of the soil accounts for the incoming and outgoing fluxes of a soil compartment. This compartment can be one-dimensional (e.g. the rootzone, or the soil profile to a greater depth). The soil water balance equation over a certain period (e.g. 7-10 days) can then be written as the change in water storage, ΔW. Defining ΔW as 'In – Out', we obtain, for a certain period of time

$$\Delta W = I + P - P_i + G - R - ET \tag{5.1}$$

where

I = irrigation (mm)
P = precipitation (mm)

P_i = intercepted precipitation (mm)
G = upward flow through the bottom (mm)
R = percolation through the bottom (mm)
ET = evapotranspiration (mm)

Re-arranging Equation 5.1 yields

$$ET = I + P - P_i + G - R - \Delta W \tag{5.2}$$

Because the soil water distribution over the profile is usually not uniform, ΔW in Equation 5.2 can be written as

$$\Delta W = \sum_{i=1}^{n} \Delta\theta_i D_i \tag{5.3}$$

where

n = number of soil layers (–)
$\Delta\theta_i$ = change in volumetric soil water content of layer i (–)
D_i = depth of the i-th soil layer (mm)

It is obvious that all errors in estimating the terms of Equation 5.2 will be reflected in the estimate of ET.

The problem with Equation 5.2 is that it is difficult to evaluate the quantity G – R properly. If there is no groundwater within reach of the bottom of the profile, this flow practically equals percolation, R. If a watertable influences the moisture conditions in the rootzone, however, capillary rise must also be considered.

For a proper evaluation of G – R (and the other terms of the water balance), one needs a lysimeter (Aboukhaled et al. 1982). A lysimeter is an isolated undisturbed column of soil, with or without a crop, in which one or more terms of the water balance can be assessed (Figure 5.1). There are two kinds of lysimeters: weighable and non-weighable. With a weighable lysimeter, ΔW can simply be determined by weighing. A reliable measurement of ET can only be obtained if the soil moisture conditions in the lysimeter are the same as those in the field. These conditions can be satisfied if the lysimeter is provided with a drainage system and a system to maintain the water potential of the soil at the bottom of the lysimeter at the same level as the water potential in the adjacent field.

In addition to the soil water balance method, there are various micro-meteorological methods to measure ET over periods of short duration. They are based on relationships concerning the energy balance, mass transfer, eddy correlation, or a combination of these. For an overview, see e.g. Jensen et al. (1990).

5.3.2 Estimating Interception

The amount of water that can adhere to the surface of the leaves of a crop depends on factors like intensity, amount and distribution of precipitation, evaporation flux, and the shape, stand, size, and nature of the leaves.

The amount of water intercepted by a crop can be measured by covering the ground below and around a number of individual plants with plastic sheets. The amounts

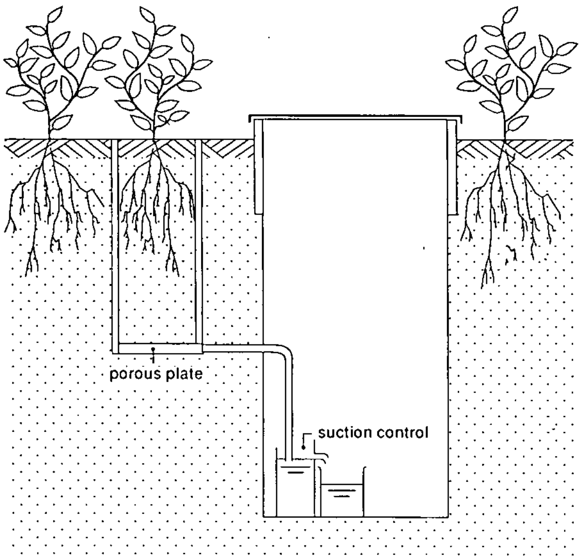

Figure 5.1 Example of a non-weighable lysimeter with suction control at the bottom

of water reaching these sheets (i.e. the throughfall) can be compared with measured rainfall to give the interception. Figure 5.2 illustrates measured interception for a small crop like grass (Rijtema 1965) and for a broad-leaved crop like red cabbage (Feddes 1971).

The scatter of the red cabbage data is largely due to variations in the different environmental factors. A smooth line was drawn through the points and, as is apparent from Figure 5.2, for a precipitation of less than 1 mm from one shower, 50 to 100% adhered to the leaves. With higher rainfall (> 5 mm), only 15% was intercepted by the leaves. Taking the scatter in the various data into account, we see that the curves for red cabbage and grass do not show significant differences.

Interception is especially important in periods of reduced evaporation. Interception

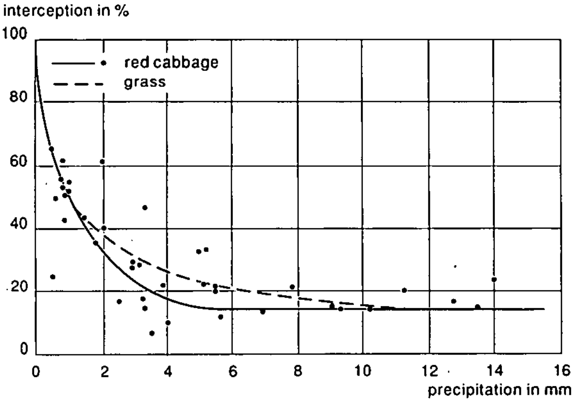

Figure 5.2 Relation between interception and rainfall depth for grass (after Rijtema 1965) and for red cabbage (after Feddes 1971)

increases the total evapotranspiration but, because part of the energy is used for the evaporation of the intercepted water (E_i), it reduces the transpiration of the crop. It should be noted that, when a relatively large error is made in estimating E_i, this leads to only a relatively small error in the final calculation of evapotranspiration.

Von Hoyningen-Hüne (1983) and Braden (1985) measured interception for various crops. On the basis of their data, a general equation can be given for the amount of water intercepted by the crop, P_i, (which is again considered to evaporate as E_i) as a function of precipitation amount, P, and leaf area index, I_l. It reads

$$P_i = a\, I_l \left(1 - \frac{1}{1 + \dfrac{bP}{a\, I_l}}\right) \qquad (5.4)$$

where

P_i = interception (mm)
a = a physical parameter, representing the crop-dependent saturation value (mm)
I_l = leaf area index (–)
b = degree of soil cover (–)
P = precipitation (mm)

5.3.3 Estimating the Evaporative Demand

Pan Evaporation

The evaporation from the free water surface of an open pan (Figure 5.3A) is widely used as an indicator of the evaporative demand of the atmosphere. Evaporation is given by the change in the water level inside the pan, after allowance is made for precipitation. Pan evaporation depends on the dimensions and exposure of the pan, the materials from which it has been constructed, and its colour, as well as on all the meteorological conditions.

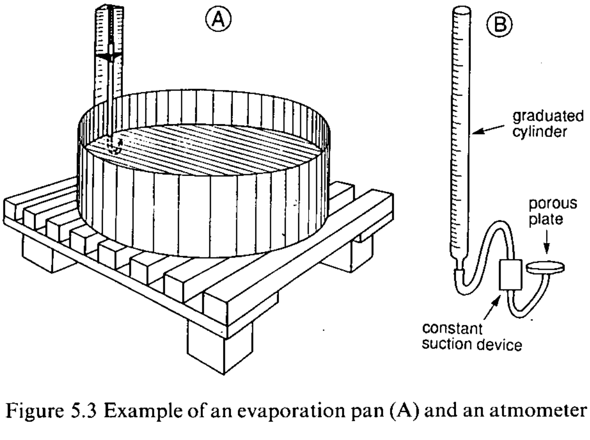

Figure 5.3 Example of an evaporation pan (A) and an atmometer (B)

The Class A pan of the U.S. Weather Bureau (122 cm in diameter and 25.4 cm high) is widely used as the standard pan (Doorenbos 1976). Because of the absorption of radiation through the pan wall and the transfer of sensible heat between the air and the pan wall, the above-ground pan receives an additional amount of energy, which results in higher evaporation rates than those calculated from meteorological data. Sunken pans might then be expected to give more reliable results, but heat exchange between the pan wall and the surrounding soil, and surface roughness effects, limit the accuracy of their results. Empirical correlations (e.g. pan factors) are required to convert measured pan evaporation rates into potential evapotranspiration rates of crops.

Atmometers

Atmometers are instruments with a porous surface connected to a supply of water in such a way that evaporation occurs from the porous surface (Figure 5.3B). A common atmometer is the Piche atmometer, made from a flat, horizontal disc of wetted blotting paper, with both sides exposed to the air. Another is the Bellani black-plate atmometer, which consists of a flat, black porous ceramic plate as the upper face of a non-porous hemisphere. Evaporation from an atmometer is affected by heat conduction through the water from the supply system. Furthermore, the transfer of sensible heat from the air is much greater with atmometers than with vegetation because the atmometer is usually placed at some height above the crop. Nevertheless, in many instances, satisfactory correlations have been found between the evaporation from an atmometer and the potential evapotranspiration from crops.

5.4 Empirical Estimating Methods

5.4.1 Air-Temperature and Radiation Methods

The formula by Turc (1954) reads

$$ET_p = \frac{P + 80}{\sqrt{1 + \left(\frac{P + 45}{L^{Tc}}\right)^2}} \qquad (5.5)$$

where

ET_p = 10-day potential evapotranspiration (mm)
P = 10-day precipitation (mm)
L^{Tc} = evaporative demand of the atmosphere, calculated as

$$L^{Tc} = \frac{(T_a + 2)\sqrt{R_s}}{11.1} \qquad (5.6)$$

in which

T_a = average air temperature at 2 m (°C)
R_s = incoming short-wave radiation (W/m^2)

The Jensen-Haise (1963) formula, with adjusted units, reads

$$ET_p = (0.025T_a + 0.08)\frac{R_s}{28.6} \tag{5.7}$$

where

ET_p = potential evapotranspiration rate (mm/d)
R_s = incoming short-wave radiation (W/m^2)
T_a = average air temperature at 2 m (°C)

Equations 5.5 and 5.7 generally underestimate ET_p during spring, and overestimate it during summer, because T_a is given too much weight and R_s too little.

5.4.2 Air-Temperature and Day-Length Method

The formula of Blaney-Criddle (1950) was developed for the western part of the U.S.A. (i.e. for a climate of the Mediterranean type). It reads

$$ET_p = k\,p\,(0.457T_{am} + 8.13)(0.031T_{aa} + 0.24) \tag{5.8}$$

where

ET_p = monthly potential evapotranspiration (mm)
k = crop coefficient (–)
p = monthly percentage of annual daylight hours (–)
T_{am} = monthly average air temperature (°C)
T_{aa} = annual average air temperature (°C)

The last term, with T_{aa}, was added to adapt the equation to climates other than the Mediterranean type. The method yields good results for Mediterranean-type climates, but in tropical areas with high cloudiness the outcome is too high. The reason for this is that, besides air temperature, solar radiation plays an important role in evaporation. For more details, see Doorenbos and Pruitt (1977).

More commonly used nowadays are the more physically-oriented approaches (i.e. the Penman and Penman-Monteith equations), which give a much better explanation of the evaporation process.

5.5 Evaporation from Open Water: the Penman Method

The Penman method (1948), applied to open water, can be briefly described by the energy balance at the earth's surface, which equates all incoming and outgoing energy fluxes (Figure 5.4). It reads

$$R_n = H + \lambda E + G \tag{5.9}$$

where

R_n = energy flux density of net incoming radiation (W/m^2)
H = flux density of sensible heat into the air (W/m^2)
λE = flux density of latent heat into the air (W/m^2)
G = heat flux density into the water body (W/m^2)

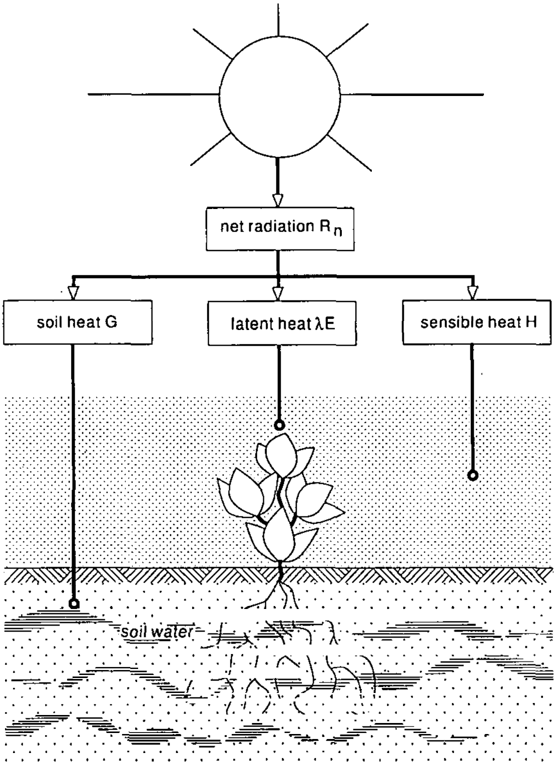

Figure 5.4 Illustration of the variables involved in the energy balance at the soil surface

The coefficient λ in λE is the latent heat of vaporization of water, and E is the vapour flux density in kg/m^2 s. Note that the evapo(transpi)ration in Equation 5.1 is expressed in mm water depth (e.g. over a period of one day). To convert the above λE in W/m^2 into an equivalent evapo(transpi)ration in units of mm/d, λE should be multiplied by a factor 0.0353. This factor equals the number of seconds in a day (86 400), divided by the value of λ (2.45×10^6 J/kg at 20°C), whereby a density of water of 1000 kg/m^3 is assumed.

Supposing that R_n and G can be measured, one can calculate E if the ratio H/λE (which is called the Bowen ratio) is known. This ratio can be derived from the transport equations of heat and water vapour in air.

The situation depicted in Figure 5.4 and described by Equation 5.9 shows that radiation energy, $R_n - G$, is transformed into sensible heat, H, and water vapour, λE, which are transported to the air in accordance with

$$H = c_1 \frac{(T_s - T_a)}{r_a} \tag{5.10}$$

$$\lambda E = c_2 \frac{(e_s - e_d)}{r_a} \tag{5.11}$$

where

c_1, c_2 = constants
T_s = temperature at the evaporating surface (°C)
T_a = air temperature at a certain height above the surface (°C)
e_s = saturated vapour pressure at the evaporating surface (kPa)
e_d = prevailing vapour pressure at the same height as T_a (kPa)
r_a = aerodynamic diffusion resistance, assumed to be the same for heat and water vapour (s/m)

When the concept of the similarity of transport of heat and water vapour is applied, the Bowen ratio yields

$$\frac{H}{\lambda E} = \frac{c_1 (T_s - T_a)}{c_2 (e_s - e_d)} \tag{5.12}$$

where

$c_1/c_2 = \gamma$ = psychrometric constant (kPa/°C)

The problem is that generally the surface temperature, T_s, is unknown. Penman therefore introduced the additional equation

$$e_s - e_a = \Delta (T_s - T_a) \tag{5.13}$$

where the proportionally constant Δ (kPa/°C) is the first derivative of the function $e_s(T)$, known as the saturated vapour pressure curve (Figure 5.5). Note that e_a in Equation 5.13 is the saturated vapour pressure at temperature T_a. Re-arranging gives

$$\Delta = \frac{e_s - e_a}{T_s - T_a} \approx \frac{de_a}{dT_a} \tag{5.14}$$

The slope Δ in Figure 5.5 can be determined at temperature T_a, provided that $(T_s - T_a)$ is small.

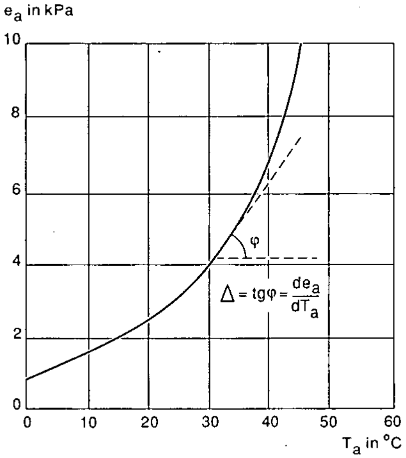

Figure 5.5 Saturated water vapour pressure e_a as a function of air temperature T_a.

From Equation 5.13, it follows that $T_s - T_a = (e_s - e_a)/\Delta$. Substitution into Equation 5.12 yields

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\frac{e_s - e_d}{e_s - e_a} \tag{5.15}$$

If $(e_s - e_a)$ is replaced by $(e_s - e_d - e_a + e_d)$, Equation 5.15 can be written as

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\left[1 - \frac{e_a - e_d}{e_s - e_d}\right] \tag{5.16}$$

Under isothermal conditions (i.e. if no heat is added to or removed from the system), we can assume that $T_s \approx T_a$. This implies that $e_s \approx e_a$. If we then introduce this assumption into Equation 5.11, the isothermal evaporation, λE_a, reads as

$$\lambda E_a = c_2 \frac{e_a - e_d}{r_a} \tag{5.17}$$

Dividing Equation 5.17 by Equation 5.11 yields

$$\frac{E_a}{E} = \frac{e_a - e_d}{e_s - e_d} \tag{5.18}$$

The ratio on the right also appeared in Equation 5.16, which can now be written as

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\left(1 - \frac{E_a}{E}\right) \tag{5.19}$$

From Equation 5.9, it follows that $H = R_n - \lambda E - G$. After some rearrangement, and writing E_o (subscript o denoting open water) for E, substitution into Equation 5.19 yields the formula of Penman (1948)

$$E_o = \frac{\Delta(R_n - G)/\lambda + \gamma E_a}{\Delta + \gamma} \tag{5.20}$$

where

E_o = open water evaporation rate (kg/m^2 s)
Δ = proportionality constant de_a/dT_a (kPa/°C)
R_n = net radiation (W/m^2)
λ = latent heat of vaporization (J/kg)
γ = psychrometric constant (kPa/°C)
E_a = isothermal evaporation rate (kg/m^2 s)

The term $\frac{\Delta}{\Delta + \gamma}(R_n - G)/\lambda$ is the evaporation equivalent of the net flux density of radiant energy to the surface, also called the radiation term. The term $\frac{\Delta}{\Delta + \gamma} E_a$ is the corresponding aerodynamic term. Equation 5.20 clearly shows the combination of the two processes in one formula.

For open water, the heat flux density into the water, G, is often ignored, especially for longer periods. Also note that the resulting E_o in kg/m^2 s should be multiplied by 86 400 to give the equivalent evaporation rate E_o in mm/d.

As was mentioned in Section 5.2, E_o has been used as a kind of reference evaporation