Standard reference evaporation calculation for inland, south eastern Australia

Wayne S. Meyer

CSIRO Land and Water, Adelaide Laboratory
Technical Report 35/98, September 1999

CSIRO LAND and WATER

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ABSTRACT

Increasing emphasis on better matching crop water use and irrigation application, demands that procedures for estimating evaporation from crops be accurate and consistent. Revised international recommendations have been combined with measured values from the Griffith weighing lysimeters to provide a methodology for calculating reference evaporation (ET_o). Daily ET_o values are generally multiplied by crop specific coefficients to produce estimates of daily crop evaporation. A locally calibrated form of the Penman combination equation for ET_o estimation is developed which uses daily meteorological data. Where appropriate, preferred estimation procedures are given for factors needed in the equation followed by alternative procedures. The alternatives, together with an estimation of their effect on ET_o are given for reader information. Consistent with the most recent FAO recommendations, future developments in calculating reference evaporation will use a Penman-Monteith equation. However, readers are cautioned that should the standardised Penman-Monteith equation be used, considerable adjustment in crop coefficients will be needed.

Contents

1.0 Introduction and background

1.1 The need

Throughout Australia, and particularly in south eastern Australia, good quality water is rapidly being recognised as a finite and limiting resource. Irrigation uses large volumes of water. More than 90% of the total diverted in the Murray-Darling Basin is used for irrigation. Understandably, careful assessment of the efficient use of water in irrigation areas is being made with both distribution and application losses being scrutinised. There will be an increasing emphasis on matching the supply of irrigation water to crop water use caused by evaporation. Associated with this is a major trend towards more controlled forms of irrigation which allow more precision in the frequency and volume of water applied. Thus advice on estimated crop water use given for irrigation scheduling purposes must be accurate and consistent. Estimating reference evaporation provides fundamental information on the evaporative environment in which crops are growing and producing. Values of reference evaporation are combined with specific crop coefficients to give general estimates of daily crop water use.

1.2 Estimating daily reference evaporation

The development and subsequent adoption of methods to estimate evaporation and by association, crop evaporation, have been closely linked to developments in measurement techniques. Direct measurement of water evaporation from containers has taken many forms until, by force of common use, Class A pans have general adoption. The alternative to direct measurement, that is calculating evaporation from meteorological measurements, has been greatly influenced by the availability of data. Simple formulae using monthly or weekly temperature values only, have been superseded by formulae containing irradiance terms, and these in turn by "combination" formulae, containing both energy (primarily irradiance) and turbulent transfer (wind and vapour pressure) terms derived from daily and even hourly meteorological measurements.

The availability of automatically recording meteorological stations has greatly increased the availability of detailed measurements of weather data. However, those responsible for estimating daily evaporation are still faced with a bewildering array of calculation methods, many of which give small but significant differences in the final calculated value.

In an effort to provide standardised guidelines for estimating daily evaporation a group under the auspices of the Riverland Liaison Committee met to agree on a common approach. This approach would then enable consistency and improved accuracy of daily evaporation estimates to be available through various state agencies.

1.3 Rationalising an approach

Over time, it has become commonly recognised that estimates of daily crop evaporation, based on pan evaporation measurements, are a useful first approximation. However, experience has shown that maintenance and location of pans has a tremendous influence on measured values. With increased availability of real-time weather data, and more attention being given to maintenance of meteorological stations, more use is being made of calculated values of evaporation. Many people have used the Penman combination equation (Penman, 1948) and adopted the general methodology given in Doorenbos and Pruitt (1977). However, there has been no uniform adoption of a calculation procedure and this has resulted in different estimates of reference evaporation (ET_o) and variable reliability of these estimates. A major study by Jensen *et al*. (1990) clearly showed Penman-ET_o estimates to be the most reliable of all estimation methods although there were seven variations used. These variations reflect attempts to make the estimation procedure account for the last few percent of variation from observed crop ET values. A recent revision of the Doorenbos and Pruitt publication (Smith, 1992) although recommending the Penman-Monteith formula as the ideal approach, recognises that the simplest form of the Penman equation, especially with local calibration will provide most satisfactory estimates of ET_o.

This position was adopted in the present analysis. A straightforward use of the Penman equation, without external modifiers, but containing local calibration coefficients for the wind function developed from the best available measurements, should be used. The second requirement was for the calculation to be logical and consistent with international recommendations.

In the following sections a standard methodology for calculating reference evaporation is presented. The preferred equations are given first then some common alternatives are shown which are likely to be encountered when reading other material. In most cases these alternatives have been assessed and their effect on the final estimate of reference evaporation is indicated for reader information.

2.0 The aim

Provide a methodology for calculating reference evaporation (ET_o) based on daily meteorological data, using a locally calibrated form of the Penman combination formula.

3.0 Definition of reference evaporation[1]

Reference evaporation (ET_o) is defined as the rate of evaporation from a hypothetical crop with an assumed crop height (120 mm) and a fixed canopy resistance (70 s m^{-1})

1 The second half of this definition comes from one adopted in 1955 to describe potential evapotranspiration (Netherlands J. Agric. Sci. 4(1): p. 95, 1956).

and albedo (0.23) which would closely resemble evaporation from an extensive surface of green grass cover of uniform height, actively growing, completely shading the ground and not short of water (Smith, 1992[2]).

4.0 Preferred calculation methods

4.1 The Penman equation

Daily estimates of reference evaporation (ET_o mm day^{-1}) were made using the formula

$$ET_o = \left[\left(\frac{\Delta}{\Delta + \gamma} \right) (R_n - G) + \left(\frac{\gamma}{\Delta + \gamma} \right) f(U)\,(e_o - e_d) \right] / L \qquad (1)$$

Δ	:	Slope of the saturation vapour pressure-temperature curve at mean daily temperature [kPa $^oC^{-1}$]
γ	:	Psychrometric constant [kPa $^oC^{-1}$]
R_n	:	Net radiant energy [MJ m^{-2} day^{-1}]
G	:	Ground heat flux (positive when direction of flux is into the ground [MJ m^{-2} day^{-1}]
f(U)	:	Wind function of the form f(U) = a'+b'(U), where a' and b' are constants and U (km day^{-1}) is wind run [MJ m^{-2} kPa^{-1} day^{-1}]
e_o	:	Mean daily saturation vapour pressure at mean dry bulb temperature [kPa]
e_d	:	Actual mean daily vapour pressure [kPa], and
L	:	Latent heat of vaporisation of water (MJ kg^{-1}).

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[2] Note that this definition is particularly intended to encapsulate the needs of the Penman-Monteith formula but is equally applicable for the Penman equation. Note also that Smith (1992) uses "evapotranspiration" rather than evaporation. My use of "evaporation", rather than "evapotranspiration" follows the recommendation of Monteith.

4.2 Latent heat of vaporisation (L)

$$\mathbf{L = 2.50025 - 0.002365T_m} \qquad (2)$$

L : Latent heat of vaporisation [MJ kg^{-1}]

T_m : Mean daily air temperature [^oC]

Reference: Fritschen and Gay (1979), Eq 6.16

Harrison (1963)

4.3 Mean daily air temperature (T_m)

For situations where hourly mean values (T_{HM}) are recorded with automatic data logging equipment

$$\mathbf{T_m = \sum_0^{24} T_{HM} / 24} \qquad (3)$$

T_m : Mean daily air temperature [^oC]

T_{HM} : Mean hourly air temperature [^oC]

Note that daily temperatures refer to the period 0000 h to 2400 h.

At Griffith, T_{HM} is recorded as the ambient dry bulb temperature from a single reading on the hour. This is consistent with requirements for the Australian Bureau of Meteorology.

In situations where limited data are gathered

$$T_m = (T_{mx} + T_{mn}) / 2 \qquad \textbf{(4)}$$

T_{mx} : Maximum daily temperature [°C]
T_{mn} : Minimum daily temperature [°C]

Preferably, T_{mx} and T_{mn} values should be based on some averaged value determined over several minutes rather than instantaneous peak values. Again, the daily values refer to the period 0000 h to 2400 h.

Comparison of T_m from (3) and (4) for 10 years of Griffith data showed a mean difference of -0.127°C. Large differences occasionally occur when cold front activity causes rapid overnight temperature changes and the daily maximum is recorded in the early hours of the day (Shell, unpublished). The difference of -0.127°C caused a 1.2% difference in ET_o for winter/spring conditions and a -0.9% difference during summer conditions (Smith and Meyer, unpublished).

4.4 Slope of saturation vapour pressure - temperature curve (Δ)

$$\mathbf{\Delta = 0.611 * 17.27 * 237.3 \exp[17.27T_m / (237.3 + T_m)] / (237.3 + T_m)^2}$$

or

$$\mathbf{\Delta = 2504 \exp[17.27\, T_m / (237.3 + T_m)] / (237.3 + T_m)^2} \qquad \textbf{(5)}$$

Δ : slope of saturation vapour pressure, temperature curve [kPa °C^{-1}]
T_m : mean daily air temperature [°C]

Equation 5 comes from differentiating Eq 28 (Section 4.10) with respect to temperature. Another way of calculating Δ is to calculate two values of e_o from Eq 28 at $T_m \pm 0.1$, take the difference and divide by 0.2.

Equation 5 may also be seen in some literature as

$$\Delta = 4098 e_0 / (T_m + 237.3)^2 \qquad \textbf{(6)}$$

e_0 : saturation vapour pressure at T_m [kPa]

where e_0 is derived using Eq 28.

Reference : Tetens (1930), Murray (1967).

An alternative expression with the temperature term converted to ^oK originated from Maas and Arkin (1978),

$$\Delta = 0.1[\exp\{21.255 - 5304 / (T_m + 273.1)\}] \\ *[5304 / (T_m + 273.1)^2] \qquad \textbf{(7)}$$

Comparison between (5) and (7) over a temperature range from -20°C to 100°C gave a mean difference between the two of about 10%. However, in the usual temperature range, from 0° to 30°C they were identical and at 50°C the difference was less than 2%. Thus either Eqs. 5 or 7 will be entirely adequate for calculation purposes, although Eq 5 is preferred because of its consistency with Eq 28.

4.5 Psychrometric constant (γ)

$$\gamma = 0.066 \qquad \textbf{(8)}$$

γ : psychrometric constant [kPa $^{o}C^{-1}$]

While γ is taken as a constant, it is more accurately defined as

$$\gamma = 0.00163\, P / L \tag{9}$$

P : atmospheric pressure [kPa]

Reference: Brunt (1952).

At P = 100 kPa, γ ranges from 0.0646 at $0^{o}C$ to 0.0672 at $40^{o}C$. The temperature value used to calculate L (Eq2) should be wet bulb temperature which more closely represents the evaporating surface temperature. However, given the general unavailability of mean P values and the limited range of γ, little error will be caused by treating γ as a constant (Eq 8).

4.6 Proportioning terms

$$\left(\frac{\Delta}{\Delta + \gamma}\right) \text{and} \left(\frac{\gamma}{\Delta + \gamma}\right)$$

Note that:

$$\left(\frac{\Delta}{\Delta + \gamma}\right) + \left(\frac{\gamma}{\Delta + \gamma}\right) = 1 \tag{10}$$

These proportioning terms give the relative importance of the energy (Rn-G) and aerodynamic [f(U) (e_o - e_d)] terms of Eq 1. They vary significantly during the year as mean daily temperature varies (see Table 1).

Table 1. Values of $5 / (5 + \gamma)$ as a function of temperature (Jensen et al., 1990, p91)

Temperature °C	$5 / (5 + \gamma)$*
0	0.403
10	0.553
20	0.683
30	0.782
40	0.852

* Note: Since γ is a function of atmospheric pressure, P (Eq 9), these values are valid for P = 100kPa. They will need adjustment for locations with significant elevation and thus lower values of P.

4.7 Net Radiant Energy (R_n)

$$\mathbf{R_n = (1 - \alpha) R_s - R_{Ln}} \qquad \textbf{(11)}$$

R_n : Net radiant energy [MJ m^{-2} day^{-1}]

α : Albedo

R_s : Solar irradiance [MJ m^{-2} day^{-1}]

R_{Ln} : Net longwave (thermal) radiant energy [MJ m^{-2} day^{-1}]

Albedo values for a green crop as recommended by Smith (1991)

$$\alpha = 0.23$$

Field measurements indicate this is accurate for summer growing crops (soybeans and maize) with closed canopies. However measurements made on irrigated wheat (Dunin, F.X., Pers. Comm.) gave a value of 0.184 ± 0.004. To be consistent with the ET_0 definition, a value of 0.23 should be used.

Solar irradiance or total global solar radiation should be measured routinely using a high quality, calibrated pyranometer (see Shell et al., 1997). Values should be reviewed to ensure that clear day R_s are within a few per cent of theoretical values determined from the latitude of the site.

4.7.1 Net longwave radiant energy (R_{Ln})

$$\mathbf{R_{Ln}} = \left[\mathbf{a} \frac{\mathbf{R_s}}{\mathbf{R_{so}}} + \mathbf{b} \right] \varepsilon' \sigma (\mathbf{T_m} + 273)^4 \qquad \textbf{(12)}$$

R_{so}	:	Daily clear day irradiance [MJ m^{-2} day^{-1}]
a	:	Empirical coefficient
b	:	Empirical coefficient
ε'	:	Net emissivity
σ	:	Stefan-Boltzmann constant [MJ m^{-2} day^{-1} K^{-4}]
T_m	:	Daily mean temperature [^{o}C]

4.7.2 Maximum daily clear day irradiance (R_{so})

Values of R_{so} can be obtained if long term data are available by fitting an envelope curve to maximum daily R_s values recorded at a nearby site. For Griffith the polynomial of this envelope curve is

$$R_{so} = 22.357 + 11.0947 \cos D - 2.3594 \sin D \qquad \textbf{(13)}$$

where

$$D = DOY / 365.25 * 2\pi \qquad \textbf{(14)}$$

and DOY = day number of the year

A comparative test between values produced by this function and observed data from Loxton, SA produced daily values agreeing within a few percent (Meissner, A., 1992, Pers. Comm.). This is not unexpected given the similarity of latitude, the flat topography, and the weather system similarity between Griffith and Loxton. However, this should not be taken to mean that average R_s values will be the same. For example, comparison between Griffith and Mildura measured R_s shows that annual average values were 16.2 MJm^{-2} and 18.6 MJm^{-2} respectively.

In the absence of long-term R_s data for sites likely to be dissimilar to Griffith or Loxton (say outside 34.2^o S ± 2^o latitude[3]) the subroutine used by Stapper *et al.* (1986) to derive R_{so} values and given in Appendix 8.2 has been shown to be acceptably accurate (see Meyer *et al.* 1993).

4.7.3 Empirical coefficients (a and b)

In applying equation 12 we assume that on a perfectly clear day the term

$$\left[a\frac{R_s}{R_{so}}+b\right]=1$$

This then implies that when $R_s = R_{so}$, a + b = 1, and further that net longwave

[3] Comparison of calculated R_{so} values for latitudes 32 to 36^oS shows maximum (summer solstice) values differ by only + 0.6% while minimum (winter solstice) values differed by + 15.3%. Thus any errors during the summer irrigation season will be insignificant.

radiant energy (R_{Ln}) is described as a function of net emissivity and surface temperature as

$$\varepsilon'\sigma(T_m + 273)^4$$

Original values used by Jensen (1973, p. 26) were 1.22 and -0.18 for a and b respectively. Revised FAO recommendations (Smith, 1992) continue to use 1.35 and -0.35; the same as the first FAO values. However, there is also a recommendation to determine the most appropriate local coefficients for a range of seasons and conditions.

Accordingly, data from several seasons has been used to derive new local values (see Meyer *et al.* 1998).

$$\mathbf{a = 0.92} \qquad \mathbf{(15)}$$

$$\mathbf{b = 0.08} \qquad \mathbf{(16)}$$

In the situation where the value for clear day irradiance is calculated as a function of latitude only using extraterrestrial irradiance (R_{soa}), values of a and b in equation 15 will change. Using the 286 observations of R_s over four seasons, and the calculated values of R_{soa} from Marcel Fuchs formulae (Fuchs, M., Volcari Res. Inst., Israel. Pers. Comm.), new values were

$$a'' = 1.10$$
$$b'' = 0.18$$

It should be noted that the ratio of R_{so}/R_{soa} for Griffith varied from 0.67 in mid winter to 0.75 in summer. As a first approximation, this means that on perfectly clear days in summer, 25% of the above atmosphere irradiance does not reach the earth's surface at Griffith. In winter, up to 33% does not reach the ground.

It is possible to impose a quality check on irradiance values since observed surface values will be within a well defined range of extraterrestrial values at the same latitude. Based on the observations of R_s at Griffith,

$$0.23 < R_s/R_{soa} < 0.75$$

The lower limit on this ratio comes from the observation of the lowest mid winter R_s of 4 MJ m^{-2} and lowest mid summer R_s of 11 MJ m^{-2}. The corresponding R_{soa} values are 17 MJ m^{-2} and 47 MJ m^{-2}.

4.7.4 Net emissivity (ε')

$$\varepsilon' = \mathbf{c} + \mathbf{d}\,\sqrt{\mathbf{e_d}} \qquad \mathbf{(17)}$$

c	:	Empirical constant
d	:	Empirical constant
e_d	:	Vapour pressure at mean daily dew point temperature [kPa]

Reference: Jensen (1973)

Values for

$$\mathbf{c = 0.34} \qquad \mathbf{(18)}$$

$$\mathbf{d = -0.139} \qquad \mathbf{(19)}$$

Note that the value of c is different from that originally used (0.32) by Jensen (1973) and Meyer (1988) - this difference will induce a 3.8% decrease in ET_o estimates from original calculations. Ideally, locally derived values of c and d are required. It is anticipated that values can be obtained from independent radiant energy balance measurements currently being made.

e_d is calculated as in Eq (26).

4.7.5 The Stefan-Boltzmann constant (σ)

$$\sigma = 4.896 \times 10^{-9} \ [\mathrm{MJ\ m^{-2}\ day^{-1}\ K^{-4}}] \qquad \textbf{(20)}$$

4.7.6 Daily mean temperature (T_m)

Where T_m is derived as in Eq (3) it can be used directly in Eq (12). This is contrary to the FAO recommendations (Smith, 1992) where $(T_m + 273)^4$ is calculated from

$$\left[(T_{mx} + 273)^4 + (T_{mn} + 273)^4\right] / 2 \qquad \textbf{(21)}$$

The difference, however for $T_{mx} = 30^{o}C$, $T_{mn} = 10^{o}C$ and $T_m = 20^{o}C$ is 0.7%.

4.8 Ground heat flux (G)

There is a diurnal and an annual pattern with respect to G. While the soil at the surface is heating up, due principally to solar irradiance, there is a gradient into the soil and a positive value for G. At night, the soil and air cool and the gradient reverses as the upper layers lose energy by radiation and convection and become colder. The net gain or loss on any day is quite small.

However when seasons change quite sharply it is worthwhile accounting for short term trends. It is also useful to have the concept before us if we consider applying the energy balance over much shorter periods, say hourly.

$$\mathbf{G = (T_m - T_{av}) \times 0.12} \qquad \mathbf{(22)}$$

G : Ground heat flux [MJ m^{-2} day^{-1}]

T_m : Mean daily temperature [^oC]

T_{av} : Average daily temperature for the previous 3 days [^oC]

$$T^i_{av} = \left[T^{i-1}_m + T^{i-2}_m + T^{i-3}_m \right] / 3 \qquad \mathbf{(23)}$$

The factor (0.12) in Eq 22 is derived from field measurements over several seasons on lysimeters at Griffith and should be general for the red brown earth and transitional red brown earths common in the Riverina region.

4.9 Wind function, f(U)

This term is effectively a conductance term giving expression to the transport of water vapour away from a well watered crop canopy in response to a water vapour pressure gradient. Given the importance of turbulence in the transport of water vapour away from the crop canopy, f(U) is calculated as a function of wind run as

$$\mathbf{f(U) = a' + b'(U)} \qquad \mathbf{(24)}$$

U: Daily total wind run measured at 2 m height [km day^{-1}]

Many values of a' and b' from various derivations have been proposed. Derived values (a' = 6.623, b' = 0.0662) from Doorenbos and Pruitt (1977) were developed from a wide range of measured data. However, the calculation methods used by Doorenbos and Pruitt were slightly different from those used here. This can affect the size of the values (see Meyer *et al.* (1987) for a discussion of the need for consistency of calculation method between the VPD and f(U) terms). There is general agreement that local calibration is desirable.

Back calculation using evaporation rates measured by lysimeters cropped with wheat and then soybeans, and the estimation procedures described here resulted in

$$\mathbf{f(U) = 17.8636 + 0.0440U} \qquad \mathbf{(25)}$$

Reference: Meyer *et al.* (1998).

4.10 Water vapour pressure, e_o and e_d

The term (e_o - e_d) in Eq 1 is generally called the vapour pressure deficit or saturation deficit term. The actual mean daily vapour pressure (e_d) is only slowly variable over any given day and major changes only occur during a change in air mass such as the passage of a cold front or introduction of a maritime air mass.

e_d is calculated at dewpoint temperature which is the best way of approximating the actual daily vapour pressure.

$$\mathbf{e_d = 0.611 \exp [17.27\, T_{dew} / (T_{dew} + 237.3)]} \qquad \mathbf{(26)}$$

e_d : actual daily vapour pressure [kPa]
T_{dew} : mean daily dew point temperature [^oC]

T_{dew} is best derived from

$$T_{dew} = \sum_{O}^{23} T_{Hdew} / 24 \qquad \textbf{(27)}$$

T_{Hdew} : Mean hourly dew point temperature [°C]

Values of T_{Hdew} are best obtained from wet bulb and dry bulb temperature sensors with aspiration, i.e. a well maintained psychrometer is the ideal. Air flow over the sensors should be >3 m/s. As a second option, which is quite satisfactory, non aspirated wet and dry bulb sensors within a Stevenson screen can be monitored. At Griffith, single values of T_{wet} and T_{dry} are recorded on the hour. These values are stored and then processed using the Goff-Gratch equation (see Shell et al., 1997). This procedure is used to calculate values consistent with the tables given in Chapter 6 of Fritschen and Gay (1979). The dew point temperature, T_{Hdew}, is the temperature at which saturation will occur if moist air is cooled at constant pressure. This value is calculated by iteratively solving the Goff-Gratch equation using the Newton-Rhapson method. This requires the Goff-Gratch equation to be differentiated. Details of this method are given in Shell et al. (1997) which describes guidelines for obtaining good quality data from automatic, low cost weather stations.

The potential daily vapour pressure e_o is calculated at the daily mean temperature T_m as

$$\mathbf{e_o = 0.611 \exp [17.27\, T_m / (T_m + 237.3)]} \qquad \textbf{(28)}$$

e_o : Mean daily saturation vapour pressure at daily mean temperature [kPa]

T_m : Mean daily air temperature [°C]

Note that in Smith (1992)

$$e_o = \left(e_{oT_{mx}} + e_{oT_{mn}}\right)/2 \qquad \textbf{(29)}$$

with ($e_{o\,Tmx}$ + $e_{o\,Tmn}$) calculated with Eq 28 but with daily maximum and minimum temperature respectively replacing T_m. The value of e_o calculated using Eq 28 was about 12% higher than that from Eq 29 when compared over one year of observed data from Griffith.

In the situation where wet and dry bulb temperatures are being recorded automatically on the hour, it is straightforward to calculate an hourly vapour pressure deficit VPD_H as

$$VPD_H = e_{Hd} - \left[e_{Hw} - \gamma\left(T_{Hd} - T_{Hw}\right)\right] \qquad \textbf{(30)}$$

e_{Hd} : Saturation vapour pressure at hourly dry bulb temperature [kPa]

e_{Hw} : Saturation vapour pressure at hourly wet bulb temperature [kPa]

T_{Hd} : Hourly dry bulb temperature [^oC]

T_{Hw} : Hourly wet bulb temperature [^oC]

γ : Psychrometric constant [kPa $^oC^{-1}$]

Both e_{Hd} and e_{Hw} can be calculated using the Tetens equation (Eq 28) with the appropriate temperature. While a constant value for γ (0.066 kPa $^oC^{-1}$) is used for aspirated wet and dry bulb thermometers, corrections (see Eq 9) for atmospheric pressure P, and the latent heat of vapourisation from the wet bulb temperature are possible for greater precision. From Fritschen and Gay (1979, p. 133)

$$\gamma = 0.00066\,(1 + 0.00115 T_{Hw})(T_{Hd} - T_{Hw})P \qquad \textbf{(31)}$$

Note that for non-aspirated wet and dry bulb thermometers the value of γ in Eq 30 should be 0.0799 (kPa $^{o}C^{-1}$) for conditions when $T_{Hw} > 0^{o}C$ (see Unwin, 1980 p. 73).

From hourly values of VPD, a daily mean can be found as

$$(e_o - e_d) = \sum_0^{23} VPD_H \;/ 24 \qquad \textbf{(32)}$$

When ET_o was calculated over a summer crop season using the procedure above (Eq 32), and compared to that using Eqs 26, 27 and 28, there was a 2.8% increase in seasonal total ET_o.

In the case of manually observed data it is possible to average T_{dew} observations at 09.00 and 15.00 hours and at as many other observations as are made. In the absence of multiple readings of dew point a single 09.00 hours observation is a reasonable approximation as it occurs after the regional boundary layer has been re-established following an overnight thermal stratification, dew fall, and/or frost deposition.

A fourth alternative is to calculate from hourly measures of relative humidity. The method as set out in the FAO recommendations (Smith, 1991) is

$$\mathbf{RH_m = (RH_{mx} + RH_{mn}) / 2} \qquad \textbf{(33)}$$

RH_m : Mean daily relative humidity

RH_{mx} : Maximum daily relative humidity [%]

RH_{mn} : Minimum daily relative humidity [%]

$$e_d = RH_m / \left[\frac{50}{e_{oTmn}} + \frac{50}{e_{oTmx}} \right] \qquad \textbf{(34)}$$

Values of $e_{o\ Tmn}$ and $e_{o\ Tmx}$ are calculated as in Eq (28) with the appropriate temperature, T_{mn} and T_{mx} respectively.

As a fifth alternative, but one which is not encouraged to be used, e_d can be roughly approximated by using T_{mn} as a substitute for T_{dew} in Eq (26).

5.0 Future developments - the Penman-Monteith equation

An expert group under the auspices of FAO met in 1990 to review the Doorenbos and Pruitt (1977) recommendations. This group subsequently recommended that calculation of ET_o be done using the Penman-Monteith equation in future. The advantage of this combination method approach is that crop evapotranspiration is directly integrated into the formula through crop and air resistance factors. This would then eliminate the need for a 2 step process to estimate actual crop evaporation viz. ET_o and a crop coefficient K_c. However, there is no way of measuring the resistance values and the quantification of them for specific crop and water availability responses is still difficult and considerable development is still needed.

To overcome the need for individual crop resistance factors in the first instance, and to maintain consistency with the FAO 24 report (Doorenbos and Pruitt, 1977) on reference crop evaporation, Smith (1992) recommended a parameterised form of the Penman-Monteith approach. This has the following form

$$ET_o = \frac{0.408\,\Delta\,(R_n - G) + \gamma \dfrac{900}{(T_m + 273)}\,U_2\,(e_o - e_d)}{\Delta + \gamma\,(1 + 0.34\,U_2)} \qquad \textbf{(35)}$$

ET_o	:	Reference crop evaporation [mm day^{-1}]
R_n	:	Net radiant energy [MJ m^{-2} day^{-1}]
G	:	Ground heat flux [MJ m^{-2} day^{-1}]
T_m	:	Mean daily air temperature [^oC]
U_2	:	Mean daily wind speed at 2 m height [m s^{-1}]
e_o	:	Mean daily saturation vapour pressure at mean dry bulb temperature [kPa]
e_d	:	Actual mean daily vapour pressure [kPa]
γ	:	Psychrometric constant [kPa$^oC^{-1}$]
Δ	:	Slope of saturation vapour pressure-temperature curve at T_m.

Apart from a change in units of wind run to mean daily wind speed, the terms in (35) are the same as those in (1).

This parameterised form was derived for a grass reference crop of 120 mm height, and assuming a constant canopy resistance.

A comparison of the standardised Penman-Monteith approach (Eq 35) with the locally calibrated Penman method for reference evaporation conditions was made. This showed that the Penman-Monteith method gave daily values which are 30% less than ET_o values. Direct adoption of Eq 35 would require considerable increases in the crop coefficients to give accurate crop evaporation values.

However it is expected that refinements can be made to the factors in the Penman-Monteith approach and that this will ultimately develop into the standard methodology. The implementation of the Penman-Monteith approach will be relatively straightforward if users understand and adopt the locally calibrated Penman method given in this paper.

6.0 Acknowledgements

Development and testing of local coefficients has been greatly assisted by David Smith. Robert White and Graeme Shell have also helped in testing. I am indebted to Mr Mick Fleming and Dr Richard Allan for advice and suggestions to help conform with accepted world standards.

7.0 References

Brunt, D. (1952). Physical and dynamical meteorology, 2nd ed. University Press, Cambridge. 428 pp.

Doorenbos, J. and Pruitt, W.O. (1977). Guidelines for predicting crop water requirements. FAO Irrigation and Drainage Paper 24, 2nd ed. Rome 156 pp.

Fritschen, L.J. and Gay, L.W. (1979). Environmental instrumentation. Springer Advanced Texts in Life Sciences, D.E. Reickle, Ed. Spring Verlag, New York, 216 pp.

Harrison, L.P. (1963). Fundamental concepts and definitions relating to humidity. In: Humidity and Moisture. Vol. 3. Wexler, A. (ed). Reinhold Publishing Company, New York.

Jensen, M.E. (1973). Consumptive use of water and irrigation water requirements. ASCE. New York 215 pp.

Jensen, M.E., Burman, R.D. and Allen, R.G. (1990). Evapotranspiration and irrigation water requirements. ASCE Manual No. 70.

Maas, S.J. and Arkin, G.F. (1978). User's guide to SORGF. A dynamic grain sorghum growth model with feedback capacity. (Blackland Research Center: Temple, Texas).

Meyer, W.S., Dunin, F.X., Smith, R.C.G., Shell, G.S.G. and White, N.S. (1987). Characterising water use by irrigated wheat at Griffith, New South Wales. Aust. J. Soil Res. 25: 49-215.

Meyer, W.S. (1988). Development of management strategies for minimising salinization due to irrigation: quantifying components of the water balance under irrigated crops. AWRAC Research Project 84/162 AWRAC, DPIE Research Project Final Report. 54 pp.

Meyer, W.S., Smith, D. and Shell, G. (1998). Estimating E from weather data - An addendum to AWRAC research project 84/162. CSIRO Land and Water, Technical Memorandum.

Murray F.W. (1967). On the computation of saturation vapor pressure. J. Appl. Meteor. 6: 203-204.

Penman, H.L. (1948). Natural evaporation for open water, bare soil and grass. Proc. R. Soc. London. A193, 120-146.

Shell, G.S., Meyer, W.S. and Smith, D.J. (1997). Guidelines on installation and maintenance of low cost automatic weather stations with particular reference to the measurement of wet bulb temperature in arid climates and the calculation of dew point temperature. CSIRO Land and Water, Griffith, Oct. 1997, Technical Report 28/97. 30 pp.

Smith, M. (1992). Report on the expert consultation on revision of FAO methodologies for crop water requirements. Land and Water Development Division, Food and

Agriculture Organisation of the United Nations, Rome 54 pp.

Stapper, M., Keegan, G.M. and Smith, R.C.G. (1986). SIRAGCROP computer-based crop management : irrigation scheduling service. Irrigation 86. Proceedings. Darling Downs Soil and Water Studies Centre and Irrigation Association of Australia. Sept 1986 137-153.

Tetens, O. (1930). Uber einige meteorologische Begriffe. Z Geophys. 6: 297-309.

Unwin, D.M. (1980). Microclimate measurement for ecologists. Biological Techniques Series. Academic Press. London. 97pp.

APPENDIX 8.1 Symbol List

Symbol	Explanation	Units
ET_o	Reference evaporation	mm d^{-1}
Δ	Slope of saturation vapour pressure-temperature curve	kPa $^oC^{-1}$
γ	Psychrometric constant	kPa $^oC^{-1}$
R_n	Net radiant energy	MJ m^{-2} day^{-1}
G	Ground heat flux	MJ m^{-2} day^{-1}
f(U)	Wind function	MJ m^{-2} kPa^{-1} day^{-1}
a’	Empirical coefficient in wind function term	
b’	Empirical coefficient in wind function term	
U	Daily wind run at 2 m	km day^{-1}
U_2	Mean daily wind speed at 2 m	m s^{-1}
e_o	Mean daily saturation vapour pressure	kPa
e_d	Actual daily vapour pressure	kPa
L	Latent heat of vaporisation	MJ kg^{-1}
T_m	Mean daily air temperature	^oC
T_{HM}	Mean hourly air temperature	^oC
T_{mx}	Maximum daily temperature	^oC
T_{mn}	Minimum daily temperature	^oC
P	Atmospheric pressure	kPa
R_s	Solar irradiance	MJ m^{-2} day^{-1}
R_{Ln}	Net longwave radiant energy	MJ m^{-2} day^{-1}
α	Albedo	
R_{so}	Daily clear day irradiance	MJ m^{-2} day^{-1}
a	Empirical coefficient in R_{LN} equation	
b	Empirical coefficient in R_{LN} equation	
D	Day number of a calendar year	
c	Empirical coefficient in ε’ equation	
d	Empirical coefficient in ε’ equation	

T_{wet}	Mean hourly wet bulb temperature	^oC
T_{dry}	Mean hourly dry bulb temperature	^oC
VPD_H	Hourly vapour pressure deficit	kPa
e_{Hd}	Saturation vapour pressure at hourly dry bulb temperature	kPa
e_{Hw}	Saturation vapour pressure at hourly wet bulb temperature	kPa
T_{Hd}	Hourly dry bulb temperature	^oC
T_{Hw}	Hourly wet bulb temperature	^oC
$e_{o\ Tmn}$	Saturation vapour pressur at T_{mn}	kPa
$e_{o\ Tmx}$	Saturation vapour pressure at T_{mx}	kPa
ε'	Net emissivity	
σ	Stefan-Boltzmann constant	MJ m^{-2} day^{-1} K^{-4}
T_{av}	Average daily temperature for previous 3 days	^oC
T_{dew}	Mean daily dew point temperature	^oC
T_{Hdew}	Mean hourly dew point temperature	^oC
RH_m	Mean daily relative humidity	%
RH_{mx}	Maximum daily relative humidity	%
RH_{mn}	Minimum daily relative humidity	%
K_c	Crop coefficient	

APPENDIX 8.2

Fortran subroutine for calculating daily clear day irradiance R_{so} (MJ m^{-2} day^{-1}) at any site in the southern hemisphere (From SIRAGCROP; Stapper, 1986). Inputs needed are latitude, day number of the year, and measured solar irradiance (MJ m^{-2}) for that day. Output (RADMX) is R_{so} for the day in question.

```
!         LATITUDE = Latitude, e.g. Griffith: 34.3°S
!         DAYLN = Daylength
          JD = Day number of the year
          RAD = Solar irradiance (measured)
! ****************************DAYLENGTH************************************************
          ANGLE = 90.+50./60.
          PI = 3.141593
          RD = JD*2.*PI/365.
          DEC1=0.397-22.98*COS(RD)+3.631*SIN(RD)-0.388*COS(2*RD)
          DECLIN=DEC1+0.039*SIN(2*RD)-0.16*COS(3*RD)
          DECLIN= -DECLIN                     !!Southern hemis: negative
          CF=2.*PI/360.
          COCO=COS(LATITUDE*CF)*COS(DECLIN*CF)
          SISI=SIN(LATITUDE*CF)*SIN(DECLIN*CF)
          CO1=AMAX1(-1.,(COS(ANGLE*CF)-SISI)/COCO)
          CO2=SQURT(1.-(CO1**2))
          HAS= -ATAN(CO1/CO2)+PI/2.
          DAYLN=2.*HAS*24./(2.*PI)

! *******************SOLAR IRRADIANCE AT TOP OF ATMOSPHERE******************
          RADIVC=1.001-0.03349*SIN(JD-94.)*2.*PI/365.
          RADTOP=37.23*RADIVC*(HAS*SISI+COCO*SIN(HAS))

! ********************IRRADIANCE ON A CLEAR DAY*************************************
          AA=1.799E-06*EXP(0.0273*LATITUDE)
          BB=6.52E-04*EXP(0.0273*LATITUDE)
          CC=0.87*EXP(-0.0064*LATITUDE)
          FRAC= -AA*(JD-182)*(JD-182)+BB*(IABS(JD-182))+CC
          RADMX=FRAC*RADTOP
          IF(RADMX.LT.RAD) RADMX=RAD

          RETURN
          END
```