A Review of Methods to Estimate Irrigated Reference Crop Evapotranspiration across Australia

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CRC for Irrigation Futures/CSIRO Land and Water

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Description: Aerial view of cropping and irrigation practices in the Murrumbidgee Irrigation Area, Griffith, NSW. 1991.

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Executive Summary

The National Evapotranspiration Project being undertaken by the Cooperative Research Centre for Irrigation Futures (CRC IF) aims to identify an appropriate model for the calculation of reference crop evapotranspiration which can be adopted as an Australian standard. The purpose of this report is to review the measurement and calculation methods that could be used to provide daily estimates of reference crop evapotranspiration especially appropriate to irrigated crops. For context, we have set out the derivations of the most commonly used calculation methods and in so doing have highlighted the strengths and weaknesses of various approaches. The primary reason for doing this is to develop a systematic and quantitative assessment of the appropriateness of a standardised estimation of reference evapotranspiration (ET_o). Having a standardised form of ET_o available will then enable compilation of regional crop coefficients, which together, allow estimates of actual crop evapotranspiration (ET_c) to be made. The outcomes of this project will have application in the estimation of daily water requirements for agricultural crops grown in different climatic regions across Australia.

Table of Contents

Introduction

Evapotranspiration is the collective term used to describe evaporation of water from a plant-soil system. Water will evaporate into the surrounding air from the soil surface, from plant surfaces and from stomatal pore surfaces. Plants require water for structural purposes and physiological functioning but the majority of water is expended in the dissipation of radiant energy through transpiration and evaporation of water, to prevent overheating.

Around 67% of controlled water consumption in Australia is used for irrigation purposes. Improved efficiencies in the use of water in the agricultural sector has long been targeted by state and federal governments, however recent water shortages across the Murray-Darling basin is giving added impetus to the endeavour to reduce water extraction from the river systems.

Evapotranspiration measurement has an important role in the estimation of irrigated crop water requirement. Currently in Australia a number of methods are employed for calculating crop evapotranspiration (ET_C) and there is a pressing need to implement a national standard to facilitate estimation, comparison and management of crop water use. The present lack of consistency has resulted in an unnecessary level of confusion and is a major limitation to the credibility and confidence of calculated estimates of water use. The identification and adoption of a standard ET method would provide a benchmark for the planning and management of irrigation systems, and an accounting measure for political negotiations.

Identification of an appropriate model for calculating reference evapotranspiration across Australia is one of the aims of the “National ET project” being undertaken by the Cooperative Research Centre for Irrigation Futures (CRC IF). As well, crop coefficients for a series of Australian agro-climatic regions will be compiled to enable calculation of crop evapotranspiration. The outcomes of this project will have application in the estimation of daily water requirements for agricultural crops – particularly irrigated crops – grown in different climatic regions across Australia.

The aim of this report is to review the methods extant for estimating irrigated reference crop evapotranspiration. A derivation for each methodology is presented, and the assumptions and limitations are highlighted.

1 Overview of evapotranspiration from crops

Evaporation is an important component of the water cycle, where liquid water on the surface of the Earth vaporises into the atmosphere. This occurs from large water bodies such as oceans, lakes and rivers, as well as from plants and the soil. The term 'evapotranspiration' refers to the combined processes of transpiration and evaporation from vegetation and the surrounding soil.

Plant growth and productivity are directly related to the availability of water (Rosenberg *et al.*, 1983). Only about 1% of the water taken up by plants is actually involved in metabolic activity; most of the water is vaporised into the air, cooling the plant and preventing overheating. Since large quantities of energy are required to change phase from liquid to vapour (2.45MJ/kg for H_2O at 20°C), transpiration is a very effective means for the dissipation of heat.

Where the water available to crops is limited, growth can be impaired and yields reduced. In low rainfall areas, crops are often irrigated to achieve greater productivity. In such regions, the water available for irrigation is often limited, so it is important to be able to predict how much water is potentially required. Estimating the water requirements by estimating evapotranspiration provides quantification for planning and operational purposes. The rate of evapotranspiration depends on a number of interlinked factors, including but not limited to:

- the physiological and morphological nature of the crop;
- local and regional climate;
- soil properties;
- local topography; and
- regional land use.

The effects of many of these factors are difficult to measure. For this reason the estimation of evapotranspiration is a semi-empirical science, with predictions being made using a small subset of these variables.

When modelling evapotranspiration, it is important to understand the assumptions and limitations of the method being used. Use of a method should ideally be justified, as a method giving valid estimates in one region may not necessarily do the same in another, particularly if empirical fitting has been used.

This section describes the process of evapotranspiration and discusses the contributing factors. The aim is to highlight the complexities of the process and aid understanding of the predictive methods discussed in later sections.

Much of this section is derived from Rosenberg *et al.* (1983), and it is recommended that this text be consulted in the first instance for additional information.

1.1 The physics of evaporation

Brutsaert (1982) discusses the microscopic origin and thermodynamics of evaporation. Below is a brief qualitative introduction to the subject.

Evaporation is the process of a liquid changing into a gas. Consider an adiabatic system (one in which there is no heat exchange) containing *dry* air at a temperature *T*. Suppose a source of liquid water is added to the system which is in contact with the air. As air molecules impinge on the liquid water molecules, some of the liquid molecules will gain enough energy to vaporise. The rate of vaporisation will depend on the average speed and hence the temperature of the air molecules. As vaporisation occurs, the sensible heat of the air is being converted to latent heat, so the air will cool.

Some of the vapour molecules in the air will condense into the liquid, and this rate of condensation will increase as the water vapour pressure increases. Eventually the rate of condensation will equal the rate of evaporation, and a dynamic equilibrium will have been reached. At this point the air is said to be saturated, and the vapour pressure is called the saturation vapour pressure for the particular air temperature.

The net rate of evaporation depends on the temperature of the air and the water (for evaporation) and the vapour pressure in the air (for condensation).

If the temperature of the air is increased, the rate of evaporation will increase and a new dynamic equilibrium will eventually form with a higher saturation vapour pressure.

Hence for soil-plant surfaces, the rate of evaporation depends on the temperatures of the evaporating surface and the air, and the vapour pressure of the air.

1.2 Plant evaporation

Most of the water that evaporates from plant surfaces has passed through the plant. The water enters through the roots, passes through the vascular tissue to the leaves or other organs, and exits into the surrounding air primarily via stomates (but sometimes also through the cuticle). Evaporation of water that has passed through the plant is called transpiration. Since it is difficult to distinguish this water vapour from that caused by direct evaporation from soils, it is convenient to use the term evapotranspiration to describe the overall process.

The rate of water vapour loss from a crop and the underlying soil depends on: the physical and physiological properties of the crop; the relative heat flux to the crop and the soil; and the water content of the soil. The 'Leaf Area Index' (LAI) of a crop is used to define the effective surface area for water loss from the crop and the amount of shading of the ground surface below. LAI is defined as the ratio of the leaf area to the ground area; when the LAI is > 3 then transpiration dominates evapotranspiration, because most of the incoming radiant energy is captured by the crop canopy and the ground surface is almost fully shaded. For more sparse crop canopies (LAI is < 3) a greater portion of incoming radiant energy reaches the ground surface and, if this is often wet, then soil evaporation will be a greater proportion of the total evaporation.

The physiological nature of the plant and the availability of water in the soil play an important role in evapotranspiration. For maximum evapotranspiration to occur, the plant must behave passively, acting as a wick for the transport of water from soil to air. This does not happen in all circumstances; sometimes the rate at which the plant sources water from the roots, or transports it through the vascular tissue, can limit evapotranspiration. The stomates ultimately control the rate of transpiration, and their resistance is affected by environmental factors including leaf temperature, light, leaf water potential and possibly vapour pressure deficit, and by the physiology associated with CO_2 uptake.

The interconnected nature of water vapour loss (transpiration, λE) and CO_2 uptake (F_c) for photosynthesis through the stomates is described by the coupled expressions:

$$\lambda E = \left[\frac{\rho C_p}{\gamma} \left(e_{sat}(T_s) - e_o \right) \right] / r_{st} \qquad 1.1$$

And

$$F_c = (c_i - c_a) / r_c \qquad 1.2$$

where T_s, e_o and c_a are the temperature, vapour pressure and CO_2 concentration near the leaf surface, and c_i is the internal CO_2 concentration. The stomatal resistance to CO_2 transfer is r_c and water vapour is r_{st}.

Maintenance of stomatal aperture to achieve maximum CO_2 uptake and minimal water loss, all within an optimum growth temperature range, requires a very sensitive and responsive system. If it was possible to either measure or estimate the critical influences of stomatal function, effective over a whole canopy, it would be possible to anticipate transpiration rates.

However, this is not yet achievable and so an aggregate expression of resistances in the transpiration flow path are attempted.

The resistance of the plant and soil to evapotranspiration is sometimes described in terms of the 'surface resistance' and the 'aerodynamic resistance':

> "The surface resistance … describes the resistance of vapour flow through stomata openings, total leaf area and soil surface. The aerodynamic resistance … describes the resistance from the vegetation upward and involves friction from air flowing over vegetative surfaces." (Allen et al, 1998)

Evaporation can occur from water on the leaves which does not travel through the plant (i.e. intercepted water from rain or irrigation). Although this water still cools the plant, its release is not controlled by the plant. It is likely that the rate of evapotranspiration will increase when there is water lying on the leaves, and it could be higher even than free water evaporation because the leaf surface area can be greater than the ground surface area. Wallace (1994) notes that the evaporation rate can more than double when the leaves are wet, although this will very much depend on the type of crop and the weather conditions.

To summarise, there are three contributions to evapotranspiration from the soil-plant system:

1. transpiration through the stomates;
2. evaporation from the soil;
3. evaporation of water on the plant leaves.

1.3 Weather

Evapotranspiration is dependent on weather. Precipitation, solar radiation, temperature, humidity, and wind all contribute to the rate of evapotranspiration. The huge quantities of energy consumed in evapotranspiration are supplied almost entirely from two sources: radiant energy and the energy associated with air movement. Both sources of energy are traceable to solar radiation.

1.3.1 Radiant energy

Radiant energy in both direct and diffuse forms, includes short-wave solar radiation and long-wave thermal radiation. Short wave radiation arriving in the Earth's atmosphere can be reflected or absorbed by both clouds and the ground. These in turn emit long-wave radiation according to Stefan's Law. A fraction of solar radiation is also absorbed by atmospheric components such as water vapour, liquid water, dust and pollutants. The presence of cloud results in most radiant energy being diffuse rather than direct.

1.3.2 Temperature

The temperature of both the air and the evaporating surface has a major influence on the rate of evapotranspiration. Temperature is particularly important because of its bearing on other meteorological variables such as saturation vapour pressure and outgoing long-wave radiation – which are critical to the rate of evapotranspiration.

1.3.3 Humidity

In general, evapotranspiration increases in response to an increasing difference between the vapour pressure at the evaporating surface and the vapour pressure of the air. Hence, as the absolute humidity of the air decreases, or air temperature increases, there is an increase in evapotranspiration. The Dalton mass transfer equation, discussed in Section 4.1, is based on this theory.

1.3.4 Sensible heat advection

Advection is the "process of transport of an atmospheric property solely by the mass motion of the atmosphere …" (*Glossary of Meteorology*, quoted in Rosenberg *et al*., 1983). In the context of evapotranspiration, it can be considered as “the transport of energy or mass in the horizontal plane in the downwind direction” (Rosenberg *et al*., 1983).

Sensible heat (H) is an integral part of the surface energy balance as described by the equation:

$$R_n - G = H + \lambda E \qquad 1.3$$

Where R_n is net radiant energy, G is soil heat flux and λE is the latent heat associated with water vapour. Normally, the ground and vegetative surface is warmer than the immediately overlaying air and so sensible heat transfers from the surface into the atmosphere. When advective conditions occur, sensible heat, over and above that generated at the point of reference, moves in from a surrounding (upwind) area. Hence the energy available for evaporation is greater than that which is generated by the net radiant energy at the point of reference. This process is described as advection.

1.3.5 Wind

Wind plays an important role in the evapotranspiration process. Strong winds enhance turbulence, removing the water vapour from the plant canopy more quickly and mixing it into the surrounding drier air. In sub-humid and arid climates, wind can also transport sensible heat from dry surroundings into wet fields.

While wind primarily responds to atmospheric pressure differences, local turbulence can be strongly influenced by topographic features. Hence abrupt elevation changes and equivalent effects such as wind barriers can cause increased local turbulence and increased evapotranspiration.

1.4 Method and frequency of irrigation

Irrigation of crops generally aims to provide sufficient water to fully satisfy the evaporative demand of that crop. For a crop that is fully irrigated, it is expected that evapotranspiration will be greater if the ground surface and the crop canopy are frequently wetted. Thus the method of applying water (e.g. sprinkler or drip irrigation) will influence this.

For a crop that is less than fully irrigated in a limited rainfall environment, it is expected that evapotranspiration will decline below a potential, fully watered rate, especially as the crop experiences drying soil conditions and it undergoes the adaptive changes associated with water deficit stress. If a crop is “over-irrigated”, it is likely that the excess water will be lost through run-off and additional soil evaporation or drainage below the root zone.

1.5 Definition of reference evapotranspiration

Two alternative methodologies can be used to calculate evapotranspiration from irrigated crops. Either:

1. the physiological (stomatal) and boundary layer resistances of each crop are combined with meteorological data to calculate evapotranspiration on an individual basis; or
2. evapotranspiration is calculated for a single reference crop, and then simple crop coefficients are applied to extrapolate to other crops.

Since the water flux path resistances from a crop cannot be measured directly, the second method is normally used. “To obviate the need to define unique evaporation parameters for each crop and stage of growth” the UN Food and Agricultural Organisation has chosen a single crop as a reference surface, which is defined as follows (Allen *et al.*, 1998):

> “A hypothetical reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s m^{-1} and an albedo of 0.23.”

“This reference closely resembles an extensive surface of green grass of uniform height, actively growing, completely shading the ground and with adequate water” (Allen *et al.*, 1998). The reference evapotranspiration, denoted ET_O, is calculated from meteorological data for this crop.

Reference evapotranspiration is a measure of the total quantity of water that the reference crop requires to avoid any water deficit and therefore would be expected to achieve maximum growth. The actual amount of water that is supplied to the crop would depend upon the efficiency of the irrigation system and the frequency of application.

In some literature, reference evapotranspiration is labelled “potential” evapotranspiration. Rosenberg *et al.* (1983) state that potential evapotranspiration should not exceed free water evapotranspiration (e.g. from an evaporation pan). Since this relationship does not always apply in sub-humid and arid climates, it is assumed in this report that evapotranspiration *can* exceed free water evaporation.

There have been many definitions of potential evapotranspiration. For the purposes of this report *potential evapotranspiration* refers to that maximum rate of evapotranspiration that a crop with full canopy cover (LAI>3) attains when it is well watered. In turn, reference evapotranspiration is that attained by the hypothetical crop as defined above or by some well defined closed canopy, well watered crop operating at potential evapotranspiration.

1.6 Crop coefficients and crop evapotranspiration

Evapotranspiration for a specific crop, ET_C, is estimated by multiplying the reference evapotranspiration, ET_O, by a crop coefficient, K_C. Hence:

$$ET_C = K_C ET_O \tag{1.4}$$

K_C is dependent on crop type, stage of growth, canopy configuration and regional climate. Throughout the growing cycle, K_C will change, increasing as the leaf area of the crop increases to a plateau when LAI > 3, then decreasing during senescence as the area of green leaf decreases.

For this report, K_C is thought to include both the effects of the crop configuration and physiology, as well as changes in soil evaporation due to surface wetting and drying. In subsequent reports, and as detailed by Allen *et al.* (1988), crop and soil effects are separated to obtain more accurate estimates of crop and soil evaporation.

2 Modelling evapotranspiration across Australia

Evapotranspiration is dependent on a large number of variables. To model the process it is necessary to simplify the problem by introducing a number of assumptions and abstractions.

The chosen model must satisfy a number of criteria:

1. The aims of the model must be fulfilled. In this project, the model must be able to produce daily estimations of *reference* evapotranspiration using available meteorological data. Appropriate regional crop coefficients must also be derived so that *crop* evapotranspiration can be estimated. The information that is produced must be in a format that is suitable to inform planners, water use operators and policy makers.
2. The accuracy of the model across Australia must be acceptable, for the purposes to which the output data will be used.
3. The availability and accuracy of the input data used in the model must be adequate. As much attention should be paid to the accuracy of the model data as to the accuracy of the model concepts.
4. Since the model may be used in political negotiations, the assumptions of the chosen methodology and the choice and quality of input data should be clear and transparent. Uncertainty analyses should be performed to calculate random and systematic errors in the final predictions.
5. Validation of the model against measurements must use existing experimental data. No new data will become available during the project.
6. If any empirical relationships are derived from measured data for use in the model, then their applicability across Australia must be justified.
7. Research, development and deployment of the model and data must be completed within the allocated budget.

The availability of input data and experimental measurements are discussed below. This is followed by a summary of the existing evapotranspiration models which need to be further investigated with respect to their ability to provide a relatively robust and reliable standard methodology for estimating evapotranspiration.

2.1 Input data for evapotranspiration calculations

Any evapotranspiration model will be limited by the quality, quantity and availability of input data.

Meteorological data are available from the Australian Bureau of Meteorology. Observations include:

- minimum daily temperature (°C) in the 24 hours to 9am;
- maximum daily temperature (°C) in the 24 hours to 9am;
- temperature readings at 9am and 3pm;
- precipitation (mm) in the 24 hours to 9am;
- “Class A” pan evaporation (mm) in the 24 hours to 9am;
- relative humidity readings at 9am and 3pm;
- fraction of sky obscured by cloud at 9pm and 3am;
- sunshine hours;
- maximum wind gust (direction, speed and time) in the 24 hours to midnight;

- wind speed at 9am and 3pm (averaged over 10 minutes prior);
- wind direction at 9am and 3pm (averaged over 10 minutes prior);
- atmospheric pressure reduced to mean sea level at 9am and 3pm.

The Bureau of Meteorology also calculates daily solar radiation (MJ/m^2) from satellite-derived data and cloud albedos. Satellites are calibrated against ground instruments. A change in satellite on 22 May 2003 resulted in the generation of slightly lower quality values than previously due to altered satellite characteristics. Estimated errors varied between ±2 MJ/m^2 under clear conditions in dry regions up to ±12 MJ/m^2 under heavy cloud in tropical regions. For clear day values in summer of around 30 MJ/m^2 these errors are very significant. On-going calibration is expected to reduce the bias (BoM, 2005).

There are a limited number of stations around Australia that measure all of the data above. In some locations, only a subset of the above data is measured. Over the years there have been some changes in methods of measurement. As well the number of measurements has increased. Hence the historic data is likely to be of lower quality than more recent data.

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2.2 Australian measurements of evapotranspiration

The success or failure of any model depends on the accuracy of its predictions, and the most common way of assessing this is by comparison with measurements. The model should be validated against a suitably wide range of experimental data. Since evapotranspiration is a locally variable process, it is important that any Australia-wide model be validated against measurements at a number of Australian sites to reduce the possibility of local bias in the model.

Unfortunately, there is not a single experimental method that has been performed at multiple sites across Australia; measurements at different sites have been performed by a number of teams using a variety of equipment. It is particularly important under these circumstances to assess each experiment for systematic and random errors.

During the 1980s and 1990s a team led by Meyer measured evapotranspiration from a number of crops at Griffith, NSW, using two carefully calibrated weighing lysimeters (Meyer, 1988, Meyer *et al.,* 1999). This study will be most useful for validation purposes.

McIlroy and Angus (1964) measured reference evapotranspiration at Aspendale, Victoria using 12 large and two small weighing lysimeters (two of these lysimeter balances were renovated by Meyer for the Griffith studies).

It would be preferable to validate the model against other experimental data in addition to that identified above, and ideally the validation should include Queensland and Western Australian data if at all possible. Limitations to the range and accuracy of experimental data will increase the random and systematic errors in the model predictions. It is anticipated that suitable data sets in addition to those identified above will be gathered to use for validation, testing and extrapolation purposes.

2.3 Evapotranspiration models

The key variables that impact on crop evapotranspiration have been briefly described in Section 1 . As has already been stated, defining and measuring all the contributing terms is problematic, consequently estimates are made using a subset of the contributing parameters.

Over the past century, a number of methodologies have been developed to calculate evapotranspiration. For the purposes of this project, only methodologies that use the meteorological input data described above are suitable. It is suggested that the following methodologies be considered:

1) Evaporation pan measurements
2) Penman combination equation
    a) Kohler-Parmele variation of the Penman equation
    b) Morton iterative variation of the Penman equation
3) Penman-Monteith equation
    a) FAO Penman-Monteith equation
4) Priestley-Taylor equation
    a) Meyer variation of the Priestley-Taylor equation
    b) Morton variation of the Priestley-Taylor equation

The evaporation pan method uses an empirical relationship to calculate reference evapotranspiration from evaporation pan measurements. The Penman and Penman-Monteith groups are based upon the Penman combination equation, and are collectively referred to as Penman-type equations in this report. The Priestley-Taylor-type equations are based on a variation of the energy balance expression in which the turbulent transfer term (for the water vapour and sensible heat components) is approximated with a constant.

These methodologies have been validated across a range of climates, so one might initially expect them to be suitable for Australia. However, the original Penman-based methodologies are primarily derived from data of humid climates where the energy contribution from advection is low. Meyer (1988) has compared some of these methods with measurements at Griffith, NSW, and notes that energy from advection is a significant contributor to the total latent heat flux. Similar behaviour is likely across Australia, since many irrigation zones are in sub-humid or arid areas. Hence the effect of advection, both regional and local, across the continent, which is likely to be both spatially and temporally heterogeneous, will require particular attention.

The primary purpose of locally calibrating the Penman combination equation and for further development of the Penman-Monteith equation is to more accurately account for local conditions including most advective influences. Testing local calibrations is the subject of subsequent reports; for this report the origins of the different methods is described.

All of the methodologies have a theoretical basis, but some also require empirical relationships to fit the predictions to measurements. The applicability across Australia of any empirical relationship would have to be justified.

Significant assumptions are made during the derivation of each methodology, and it is important that the reader understands these and can consider the implications for Australian calculations. In the following sections, a derivation of each method is reported to aid this process. The input data required by each methodology is also listed.

3 Evaporation pans

Evaporation pans have been used to measure evaporation for over a century. The pan is regularly filled to a specified height, and the water loss (equal to evaporation) noted. The Class-A pan is considered to be the standard international pan.

Jones (1992) suggests that the rate of evaporation depends on the:

- type of pan;
- type of pan environment;
- method of operating the pan;
- exchange of heat between pan and ground;
- solar radiation;
- air temperature;
- wind; and
- temperature of the water surface.

Pan evaporation rates are different to evapotranspiration rates. Allen *et al.* (1998) relate the reference evapotranspiration to the pan evaporation E_{pan} using an empirically derived pan coefficient K_p.

$$ET_O = K_p E_{pan} \qquad 3.1$$

Allen *el al.* (1998) provides a series of equations for K_P for different ground cover, fetch[1] and climatic conditions.

It should be noted that K_P can be greater than unity. For humid areas, maximum evapotranspiration usually does not exceed free water evaporation, however in subhumid and arid regions this is not the case. Rosenberg *et al.* (1983, p211) state that in the Great Plains region of the United State as well as in other arid regions, "well-watered crops that exert little canopy resistance [e.g. lucerne] can consume more energy and transpire more water than is evaporated from free water surfaces". The large number of pan measurements across Australia provides a valuable measure of evaporation (subject to the considerable errors associated with variable siting and operation) and in turn may provide an estimation of evapotranspiration. In particular, it might be possible to obtain an estimate of relative advection effects from across Australia by analysing pan measurements. This might provide, in the absence of any other comparable data, a means to obtain first order estimates of ET_O that account for local advective conditions.

Meyer *et al.* (1999) have compared reference evapotranspiration (as calculated using the Penman equation with Meyer wind coefficients) with Class A pan evaporation from Griffith during 1991 to 1993. Class A Pan values tended to be 7-8% higher than locally calibrated ET_O values for evaporation rates less than 10 mm day^{-1}. For greater evaporation values the pan tended to overestimate evapotranspiration rates by up to 30%.

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[1] Distance of the identified upwind surface type.

4 Formulae for evapotranspiration calculations

A number of equations are used in the derivation of the evapotranspiration formulae. The significant contribution of Penman (1948) was to combine the Dalton mass transfer and the energy balance equations. This section introduces these equations.

The energy balance equation relates net radiation to latent heat flux, sensible heat flux and ground heat flux. The calculation of these quantities is discussed. Finally, the saturation vapour pressure curve is introduced.

4.1 Dalton mass transfer equation

Modern study of evaporation began with Dalton in the late eighteenth century. Dalton "theorised that evaporation from a surface must be a consequence of the combined influence of the wind, atmospheric moisture content, and characteristics of the surface" (Rosenberg *et al.*, 1983). The Dalton aerodynamic equation for evaporation from a free water surface is:

$$E = f(u)\left(e_s - e_a\right) \qquad 4.1$$

where E is the evaporation in unit time (mm s^{-1}), $f(u)$ is a wind function, e_s is the vapour pressure at the evaporating surface (kPa), and e_a the vapour pressure of the atmosphere above (kPa) (Penman, 1948).

Empirical wind functions have been identified to fit the predictions to measurements. There are several reasons against using this equation to estimate evapotranspiration alone:

1. it is only appropriate for a free water surface;
2. the evaporating surface temperature – used to calculate e_s – is difficult to measure;
3. experimental work suggests that this approach is not adequate (Penman, 1948).

4.2 Energy balance equation

The energy consumed during evapotranspiration can be traced to solar radiation. Brutsaert (1982, p128) states the energy balance equation for vegetation or water bodies as:

$$\frac{\partial Q}{\partial t} = R_n - G - \lambda E - H + L_p F_p + A_h \qquad 4.2$$

Where
- Q is the energy stored in the soil-plant system,
- R_n is the net radiation,
- G is the soil heat flux,
- λ is the latent heat of vaporisation,
- H is the sensible heat flux,
- L_p is the thermal conversion factor for fixation of carbon dioxide,
- F_p is the flux of CO_2, and
- A_h is the energy advection into the soil-plant system from water flow.

Depending on the application, several terms can normally be discounted as being negligible (Brutsaert, 1982):

- On a daily basis, for thin layers of water or soil and for small canopies, the rate of change of stored energy term can be omitted. However, it is sometimes required in the case of tall vegetation, especially around sunrise and sunset.
- The energy advection term A_h represents the change in energy flux from precipitation or irrigation. It is likely to be negligible unless a whole lake or snow is being

considered – the magnitude of the term depends on the temperature difference between the incident water and the evaporating surface.

- Under favourable conditions, L_pF_p can be of the order of 5% of the net radiation, but it is usually closer to 1%. This term is normally neglected unless its determination is the main objective.

Hence in most situations, equation 4.2 can be written as:

$$R_n - G = \lambda E + H \quad 4.3$$

Equation 4.3 is the starting point for the majority of the methods described in this report.

For a given point or a reference crop area and its associated atmosphere volume in contact and exchange with the evaporating surface, equation 4.3 describes the major energy components. In arid conditions an irrigated area surrounded by dry areas with sparse vegetation may be subject to additional sensible heat energy as lateral air movement (wind) moves in warmer, drier air. If the irrigated, evaporating area is relatively small, the air moving over may not come into complete equilibrium with the underlying evaporating surface. In this situation, there is energy available for heating and evaporation that is in addition to that from incident radiant energy, and hence R_n. The tendency is therefore to have evaporation rates (latent energy) that are higher than would result from R_n alone. This is then described as advective conditions. Thus, in arid and semi-arid areas, irrigated regions are often surrounded by dry areas and evapotranspiration rates exceeding those ascribed to R_n will generally be observed.

4.3 Calculation of the net radiation term

Solar irradiation entering the atmosphere can be absorbed or reflected, by clouds, the air or the ground. Some of the absorbed energy is radiated as longwave radiation. The net radiation R_n is calculated by considering the total radiation balance for the soil plant system (Meyer *et al.*, 1999):

$$R_n = R_{ns} + R_{nl} \quad 4.4$$

where R_{ns} is the net shortwave, and R_{nl} the net longwave, radiant energy.

4.3.1 Net shortwave radiation

R_{ns} is defined as the difference between incoming shortwave energy R_s and reflected shortwave radiation. The incoming shortwave energy can be measured on a horizontal surface using a pyranometer or solarimeter. The net shortwave radiation is then calculated using the equation:

$$R_{ns} = R_s - \alpha R_s \quad 4.5$$

where α is the albedo of the surface in question. The albedo is highly variable for different surfaces and for the angle of incidence or slope of the ground surface (Allen *et al.*, 1998). It can be as large as 0.95 for freshly fallen snow, or 0.05 for wet bare soil. Green vegetation cover tends to have an albedo of 0.18-0.25, and is generally assumed to have a value of 0.23. Tables of values for a variety of surfaces can be found in text books such as Rosenberg *et al.* (1983) and Oke (1987). Meyer *et al.* (1999) use a value of 0.184 for wheat, which is based on locally measured values.

4.3.2 Net longwave radiation

Meyer *et al.* (1999) use an empirical method to calculate the net longwave radiation:

$$R_{\ln} = R_{ld} - R_{lu} \quad 4.6$$

R_{ld} is the downwards longwave radiation emitted from the sky, clouds and aerosols in the atmosphere, and R_{lu} is the upwards longwave radiation emitted from the evaporating surface. R_{lu} is calculated using the equation:

$$R_{lu} = \varepsilon_{vs}\sigma(T_s + 273)^4 \qquad 4.7$$

Where ε_{vs} the effective surface emissivity, with a value for a green crop of 0.98,

σ the Stefan-Boltzmann constant (4.899 x 10^{-9} MJ m^{-2} day^{-1} K^{-4}), and

T_s the surface temperature (°C).

Since the evaporating surface temperature is difficult to measure, this term is often replaced with the air temperature measured close to the ground:

$$R_{lu} = \varepsilon_{vs}\sigma(T_a + 273)^4 \qquad 4.8$$

This assumption is discussed further in Section 6.

For a clear day (i.e. no clouds), the downward longwave radiation R_{ld} primarily comes from the first few hundred metres of atmosphere above the ground. The clear day downward longwave radiation is considered to be a function of the ground air temperature:

$$R_{ldo} = \varepsilon_a\sigma(T_a + 273)^4 \qquad 4.9$$

where ε_a is the atmospheric emissivity, which is principally affected by water vapour.

Substituting equations 4.8 and 4.9 into equation 4.6 gives the net longwave radiation for a clear day, R_{nlo}:

$$R_{nlo} = (\varepsilon_a - \varepsilon_{vs})\sigma(T_a + 273)^4 = \varepsilon'\sigma(T_a + 273)^4 \qquad 4.10$$

where ε' is the effective net emissivity. On clear days, ε_{vs} is always greater than ε_a, so ε' and hence R_{nlo} are always negative.

Modelling the effect of clouds is more difficult. Clouds act as effective black bodies with an emissivity of 1.0. One method is to express the net longwave radiation as a function of cloudiness and the clear day net longwave radiation:

$$R_{nl} = \left(a\frac{R_s}{R_{so}} + b\right)R_{nlo} \qquad 4.11$$

where R_s is the incoming shortwave energy described in the previous section, R_{so} is the total short wave energy that would be received on a totally clear day, and *a* and *b* are empirical constants. Substituting equation 4.10 into 4.11 gives:

$$R_{nl} = \left(a\frac{R_s}{R_{so}} + b\right)\varepsilon'\sigma(T_a + 273)^4 \qquad 4.12$$

Allen *et al.* (1998) suggest constant values of a=1.35 and b=-0.35. However, using regression analysis, Meyer *et al.* (1999) found that a=0.92 and b=0.08 were more suitable for Griffith.

Note that the sum of *a* and *b* should equal 1 so that equation 4.12 is valid on a clear day (when $R_s = R_{so}$).

4.3.3 Effective net emissivity

Since the atmospheric emissivity ε_a is principally affected by water vapour, an empirical relationship to derive ε' is used:

$$\varepsilon' = c + d\sqrt{e_a} \qquad 4.13$$

e_a is the mean daily water vapour pressure of the air, and c and d are constants with values of 0.34 and -0.139 respectively. Meyer *et al.* (1999) notes that daily values of ε' have a range from 0.15 to 0.22 for the winter/spring wheat season and the spring/summer soybean season at Griffith. ε' depends on the typical range of e_a through the growing season.

4.3.4 Calculation of the total shortwave radiation on a clear day

Meyer *et al.* (1999) considered formulae suggested by Stapper and Running to calculate R_{so}, but found that both of these overestimated R_{so} at Griffith, especially in summer. They proposed a new formula which was estimated from a fitted polynomial that is the effective upper envelope function for observed clear day solar irradiance values:

$$R_{so} = 22.357 + 11.0947\cos D - 2.3594\sin D \quad (4.14)$$

where $$D = 2\pi \frac{DOY}{365.25} \quad (4.15)$$

and DOY is the day number of the year. Equation 4.14 is optimised for Griffith, and requires alteration to calculate R_{so} at other latitudes.

Calculating clear sky radiation at ground level can be done using solar geometry. For all regions across Australian, it is recommended that the method set out in FAO 56 (Allen *et al.*, 1998) be used which requires only latitude and altitude to be specified.

4.3.5 Calculation of net longwave radiation using extraterrestrial radiation

The net long-wave radiation on a clear day, R_{nlo}, is measured at ground level. Equation 4.12 can be altered to use the extraterrestrial radiation, R_{soa}, instead (R_{soa} is the radiation incident on the upper atmosphere). Meyer *et al.* (1999) found that a=1.10 and b=0.18 for this situation, but noted that the ratio of R_{so}/R_{soa} varied from 0.67 in mid-winter to 0.75 in summer. Great care should be taken when using R_{soa}.

4.4 Calculation of the ground heat flux term

For soil-plant environments, the ground heat flux (G) is generally small compared to the net radiation, so is often omitted from energy balance calculations unless long term seasonal changes are being examined. Rosenberg *et al.* (1983) provide a general treatment of ground heat flux calculations. The approach used by Meyer (1988) is discussed here.

Jensen *et al.* (1971) estimates the ground heat flux using the equation (in metric units):

$$G = 0.377(T_a - T_{av}) \quad (4.16)$$

Where T_a is the average air temperature and T_{av} is the mean air temperature for the preceding 3 days. Meyer (1988) measured the ground flux at Griffith using three heat flux plates connected in series and buried 40 mm to 50 mm below the ground surface of two lysimeters. From regression analysis, it was suggested that the following equation be used for Griffith soils:

$$G = 0.12(T_a - T_{av}) \quad (4.17)$$

4.5 Saturation vapour pressure curve

The maximum vapour pressure content of the air depends on the air temperature. When the maximum has been reached for a particular temperature, the air is said to be saturated.

A graph of the saturation vapour pressure against temperature is shown in Figure 1. This is called the psychrometric chart. The gradient of the curve is denoted Δ and is a function of temperature T and saturated vapour pressure e^0. Allen *et al.* (1998) calculate Δ using:

$$\Delta = \frac{4098e^0}{(T + 237.3)^2} \quad (4.18)$$

The saturation vapour pressure is calculated using the equation:

$$e^0 = 0.6108\exp\left(\frac{17.27T}{T+237.3}\right) \qquad 4.19$$

For small temperature differences, the gradient can be approximated as:

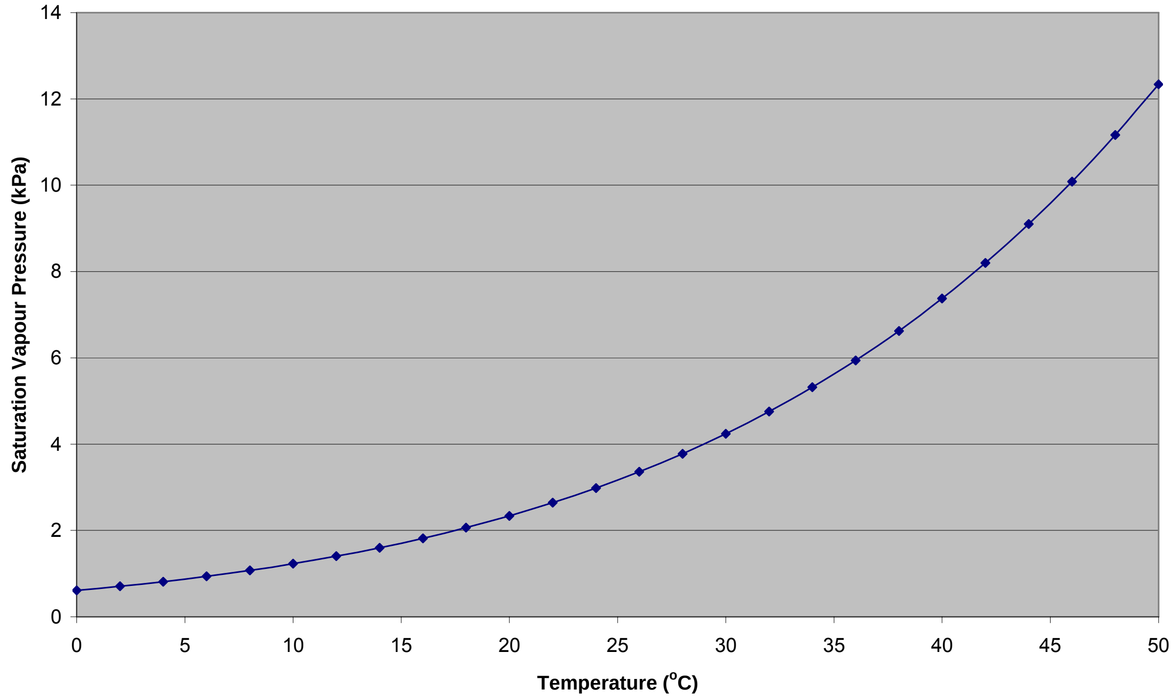
$$\Delta(T_1, T_2) = \frac{(e_1^0 - e_2^0)}{(T_1 - T_2)} \qquad 4.20$$

where T_1 and T_2 are the temperatures and e_1^0 and e_2^0 the respective saturation vapour pressures. For the evaporating surface, this equation is often written as:

$$\Delta(T_s, T_a) = \frac{(e_s^0 - e_a^0)}{(T_s - T_a)} \qquad 4.21$$

Although Δ should strictly be the gradient at the mid-point of T_s and T_a, it is often calculated at the air temperature T_a when the evaporating surface temperature is not known.

Figure 1: Saturation vapour pressure curve



5 Penman Combination Equation

Many of the methods that are discussed in this report are based on the Penman combination equation (Penman, 1948).

Prior to 1948, the Dalton mass transfer equation or the energy balance equation had been used to estimate evapotranspiration. Unfortunately, both equations require knowledge of the evaporating surface temperature of the soil-plant system, and this is difficult to measure. The Penman combination equation combines the two equations to remove the evaporating surface temperature term.

The Penman combination equation was developed to calculate evaporation from an open water source in a humid environment (the UK). The equation has subsequently been widely used to calculate evapotranspiration from well-watered crops. In a practical situation where the water supply is restricted, actual evapotranspiration will decline and the equation will no longer be valid.

The equation has been used for daily and even hourly calculations. Meyer (1988) found that hourly calculations at Griffith were not more accurate than daily predictions. Since some other studies have come to the opposite conclusion, it might be valuable to revisit the experimental measurements to search for possible limitations in the accuracy and adequacy of the measured data.

5.1 Derivation of the Penman combination equation

The glossary contains definitions of all of the terms used below.

Consider the simplified energy balance equation described in Section 4.2:

$$R_n - G = \lambda E + H \tag{4.3}$$

Rearranging this equation gives:

$$R_n - G = \lambda E\left(1 + \frac{H}{\lambda E}\right) = \lambda E(1 + \beta) \tag{5.1}$$

$$\lambda E = \frac{(R_n - G)}{(1 + \beta)} \tag{5.2}$$

where β is the ratio of sensible heat to latent heat, referred to as the Bowen ratio (see Appendix A).

If the temperature and vapour pressure gradients from the evaporating surface to the air are assumed to be linear, then equation A8 (Appendix A) can be substituted into equation 5.2:

$$\lambda E = \frac{(R_n - G)}{\left(1 + \gamma \frac{(T_s - T_a)}{(e_s - e_a)}\right)} \tag{5.3}$$

The air adjacent to the evaporating surface is assumed to be saturated, so e_s becomes e_s^0

$$\lambda E = \frac{(R_n - G)}{\left(1 + \gamma \frac{(T_s - T_a)}{(e_s^0 - e_a)}\right)} \tag{5.4}$$

In humid conditions, the evaporating surface temperature should be close to the air temperature. For these conditions, equation 4.21 can be rewritten as:

$$(T_s - T_a) = \frac{(e_s^0 - e_a^0)}{\Delta} \tag{5.5}$$

Substituting equation 5.5 into equation 5.3:

$$\lambda E = \frac{(R_n - G)}{\left(1 + \frac{\gamma}{\Delta}\frac{\left(e_s^0 - e_a^0\right)}{\left(e_s^0 - e_a\right)}\right)} \tag{5.6}$$

The evaporating surface temperature is required to calculate the saturated vapour pressure at the surface. Penman removed this requirement by taking the Dalton mass transfer equation (equation 4.1), and assuming saturation near the surface:

$$\lambda E = f(u)\left(e_s^0 - e_a\right) \tag{5.7}$$

He also defined a new mass transfer equation:

$$\lambda E_a = f(u)\left(e_a^0 - e_a\right) \tag{5.8}$$

Dividing equation 5.8 by equation 5.7 and rearranging:

$$\frac{\lambda E_a}{\lambda E} = \frac{\left(e_a^0 - e_a\right)}{\left(e_s^0 - e_a\right)} = \frac{\left(e_s^0 - e_a - e_s^0 + e_a^0\right)}{\left(e_s^0 - e_a\right)} = 1 - \frac{\left(e_s^0 - e_a^0\right)}{\left(e_s^0 - e_a\right)} \tag{5.9}$$

$$\frac{\left(e_s^0 - e_a^0\right)}{\left(e_s^0 - e_a\right)} = 1 - \frac{\lambda E_a}{\lambda E} \tag{5.10}$$

Substituting equation 5.10 into equation 5.6 gives:

$$\lambda E = \frac{(R_n - G)}{\left[1 + \frac{\gamma}{\Delta}\left(1 - \frac{\lambda E_a}{\lambda E}\right)\right]} \tag{5.11}$$

Rearranging this equation produces the **Penman** combination equation:

$$\lambda E = \frac{\Delta(R_n - G) + \gamma \lambda E_a}{\Delta + \gamma} \tag{5.12}$$

$$\lambda E = \frac{\Delta(R_n - G) + \gamma\ f(u)\left(e_a^0 - e_a\right)}{\Delta + \gamma} \tag{5.13}$$

The gradient Δ is usually calculated at the air temperature, making the assumption that the air and surface temperatures are similar. It would be more accurate to calculate Δ at the mid-point between the air and evaporating surface temperatures.

When calculating the net longwave radiation as described in Section 4.3.2, the calculation of emitted surface radiation is usually performed at the air temperature rather than the evaporating surface temperature.

5.2 Data requirements

To perform a reference evapotranspiration calculation using the Penman combination equation, the following data are required:

- mean air temperature;
- mean dewpoint (or dry and wet bulb temperature) or relative humidity;
- mean wind velocity at a standard height (for the wind function); and
- measurement of incident radiation or measurement of sunshine hours and cloud cover (from which incident solar radiation can be modelled).

In addition, crop coefficients are required to calculate the rate of crop evapotranspiration.

5.3 Wind function

The wind function, *f(u)*, is derived empirically to fit the predictions to measurements for the reference crop. Since the choice of wind function has significant impact on the predictions (Meyer, 1988), the types of wind functions that have been used are discussed separately in Section 10.

6 Kohler and Parmele Penman Variation

In Section 4.3.2 (equation 4.8), the longwave radiation from the soil-plant system is calculated using the air temperature instead of the evaporating surface temperature. Kohler and Parmele (1967) suggest a variation of the Penman equation which removes this approximation. This is the only approximation that can easily be removed while retaining an analytical solution.

The error from using the air temperature in the calculation of R_n is expected to be small. The main reason for presenting this derivation is to allow the reader to understand the Morton iterative equation, which takes a similar approach.

6.1 Derivation of the Kohler and Parmele equation

Suppose the net radiation is composed of two components:

$$R_n(T_s) = R_{ir} - R_{lu} \qquad 6.1$$

where R_{ir} is the difference between incident and reflected radiation (all-wave) to the evaporating surface, and R_{lu} the radiation emitted by the surface. From equation 4.7, R_{lu} is:

$$R_{lu} = \varepsilon_{vs}\sigma(T_s + 273)^4 \qquad 4.7$$

Substituting equation 4.7 into equation 6.1 gives:

$$R_n(T_s) = R_{ir} - \varepsilon_{vs}\sigma(T_s + 273)^4 \qquad 6.2$$

This equation can be rewritten and then expanded using the binomial expansion[1]:

$$R_n(T_s) = R_{ir} - \varepsilon_{vs}\sigma\big((T_a + 273) + (T_s - T_a)\big)^4 \qquad 6.3$$

$$R_n(T_s) = R_{ir} - \varepsilon_{vs}\sigma\big((T_a + 273)^4 + 4(T_a + 273)^3(T_s - T_a)\big) \qquad 6.4$$

Substituting equation 6.4 into the Penman equation 5.12 gives:

$$\lambda E = \frac{\Delta\Big(R_{ir} - \varepsilon_{vs}\sigma\big((T_a + 273)^4 + 4(T_a + 273)^3(T_s - T_a)\big) - G\Big) + \gamma\lambda E_a}{\Delta + \gamma} \qquad 6.5$$

$$\lambda E = \frac{\Delta(R_{ir} - G) - \Delta\varepsilon_{vs}\sigma(T_a + 273)^4 - 4\Delta\varepsilon_{vs}\sigma(T_a + 273)^3(T_s - T_a) + \gamma\lambda E_a}{\Delta + \gamma} \qquad 6.6$$

The evaporating surface temperature term can be removed by substituting equation 5.5 into equation 6.6:

$$\lambda E = \frac{\Delta(R_{ir} - G) - \Delta\varepsilon_{vs}\sigma(T_a + 273)^4 - 4\varepsilon_{vs}\sigma(T_a + 273)^3(e_s^0 - e_a^0) + \gamma\lambda E_a}{\Delta + \gamma} \qquad 6.7$$

Subtracting Penman's mass transfer equations 5.7 and 5.8 produces:

$$\lambda E - \lambda E_a = f(u)\big(e_s^0 - e_a - e_a^0 + e_a\big) = f(u)\big(e_s^0 - e_a^0\big) \qquad 6.8$$

Rearranging equation 6.8 in terms of the vapour pressure terms gives:

$$\big(e_s^0 - e_a^0\big) = \frac{\lambda E - \lambda E_a}{f(u)} \qquad 6.9$$

Substituting this into equation 6.7 and rearranging:

---

[1] Binomial expansion: $(a + x)^n \approx a^n + na^{n-1}(x - a)$, as long as $x \ll a$. Here, $a = (T_a + 273)$, $x = (T_s - T_a)$.

$$(\Delta + \gamma)\lambda E = \Delta\left(R_{ir} - G - \varepsilon_{vs}\sigma(T_a + 273)^4\right) - \frac{4\varepsilon_{vs}\sigma(T_a + 273)^3(\lambda E - \lambda E_a)}{f(u)} + \gamma\lambda E_a \qquad 6.10$$

Further rearranging produces the **Kohler and Parmele variation** of the Penman equation:

$$\lambda E\left(\Delta + \gamma + \frac{4\varepsilon_{vs}\sigma(T_a + 273)^3}{f(u)}\right) = \Delta\left(R_{ir} - G - \varepsilon_{vs}\sigma(T_a + 273)^4\right) + \lambda E_a\left(\gamma + \frac{4\varepsilon_{vs}\sigma(T_a + 273)^3}{f(u)}\right)$$

$$\lambda E = \frac{\Delta\left(R_{ir} - G - \varepsilon_{vs}\sigma(T_a + 273)^4\right) + \lambda E_a\left(\gamma + 4\varepsilon_{vs}\sigma(T_a + 273)^3 / f(u)\right)}{\Delta + \left(\gamma + 4\varepsilon_{vs}\sigma(T_a + 273)^3 / f(u)\right)} \qquad 6.11$$

The equation to calculate R_{ir} from the approximate R_n calculated in Section 4.3.2 is:

$$R_{ir} = R_n(T_a) + \varepsilon_{vs}\sigma(T_a + 273)^4 \qquad 6.12$$

Where $R_n(T_a)$ is calculated as in Section 4.3. Note that R_n in equations 6.1 to 6.4 is different to the approximate R_n that is calculated in Section 4.3.2; while the former calculates the longwave surface radiation using the evaporating surface temperature, the latter approximates by using the air temperature.

6.2 Data requirements

To perform a reference evapotranspiration calculation using the Kohler and Parmele variation of the Penman equation, the same data are required on a daily basis as for the Penman combination equation (refer Section 5.2). In addition, crop coefficients are required to calculate daily crop evapotranspiration.

7 Morton iterative Penman variation

Morton's complementary relationship areal evapotranspiration (CRAE) model was formulated to provide *monthly* estimates of *regional* evapotranspiration. The Penman equation is designed for use in humid climates where advection is low and the air and evaporating surface temperatures are close. In arid regions, this is often not the case. Morton (1983) suggests an iterative variation of the Penman equation to calculate an appropriate evaporating surface temperature.

There are two components of the CRAE model: potential evapotranspiration, E_{TP} and wet-environment areal evapotranspiration, E_{TW}. The complementary relationship is represented by (Morton, 1983):

$$E_T + E_{TP} = 2E_{TW} \quad 7.1$$

Or

$$E_T = 2E_{TW} - E_{TP} \quad 7.2$$

Potential evapotranspiration is found by solving the energy balance and vapour transfer equations:

$$E_{TP} = R_T - \lambda_s f_T\left(T_s - T_a\right) \quad 7.3$$

And

$$E_{TP} = f_T\left(e_s^0 - e_a^0\right) \quad 7.4$$

Where E_{TP} is the potential evapotranspiration, R_T is the net radiation, T_s and T_a are the equilibrium surface and air temperatures, respectively, f_T is the vapour transfer coefficient, e_s^0 is the saturation vapour pressure at T_s, e_a^o is the saturation vapour pressure at the dewpoint temperature and λ_s is a function of the atmospheric pressure and the equilibrium temperature (Granger and Gray, 1990).

One possible derivation for the energy balance equation starts with equation 6.4 of the Kohler and Parmele (1967) derivation:

$$R_n\left(T_s\right) = R_{ir} - \varepsilon_{vs}\sigma\left((T_a + 273)^4 + 4(T_a + 273)^3(T_s - T_a)\right) \quad 6.4$$

Substituting this into the rearranged version of the energy balance equation gives:

$$\lambda E = R_n\left(T_s\right) - G - H \quad \text{(rearranged)} \quad 4.3$$

$$\lambda E = R_{ir} - \varepsilon_{vs}\sigma\left((T_a + 273)^4 + 4(T_a + 273)^3(T_s - T_a)\right) - G - H \quad 7.5$$

This equation can be rearranged to group the net soil-plant irradiation terms:

$$\lambda E = \left(R_{ir} - \varepsilon_{vs}\sigma(T_a + 273)^4 - G\right) - 4\varepsilon_{vs}\sigma(T_a + 273)^3(T_s - T_a) - H \quad 7.6$$

Morton (1965, p78) states that the sensible heat transfer from the evaporating surface to the air has the form:

$$H = \gamma f(u)\left(T_s - T_a\right) \quad 7.7$$

Equation 7.7 is derived by substituting the Dalton aerodynamic equation (4.1) into the Bowen ratio (equation A8). Substituting equation 7.7 into equation 7.6 and rearranging:

$$\lambda E = \left(R_{ir} - \varepsilon_{vs}\sigma(T_a + 273)^4 - G\right) - 4\varepsilon_{vs}\sigma(T_a + 273)^3(T_s - T_a) - \gamma f(u)\left(T_s - T_a\right) \quad 7.8$$

$$\lambda E = \left(R_{ir} - \varepsilon_{vs}\sigma(T_a + 273)^4 - G\right) - \left(\gamma f(u) + 4\varepsilon_{vs}\sigma(T_a + 273)^3\right)\left(T_s - T_a\right) \quad 7.9$$

This can be rewritten as:

$$\lambda E = R_T - \lambda_a f(u)\left(T_s - T_a\right) \qquad 7.10$$

where $R_T = R_{ir} - \varepsilon_{vs}\sigma(T_a + 273)^4 - G$ and $\lambda_a = \gamma + 4\varepsilon\sigma(T_a + 273)^3 / f(u)$.

It should be noted that Morton (1983) uses temperature T_s instead of T_a in the equation for λ_a – the reasoning behind this has not been identified – and replaces *f(u)* with f_T, a vapour transfer coefficient independent of wind speed.

If saturation near the evaporating surface is assumed, then the Dalton aerodynamic equation (equation 2.1) becomes:

$$\lambda E = f(u)\left(e_s^0 - e_a\right) \qquad 7.11$$

Equations 7.10 and 7.11 can be combined to calculate the evaporating surface temperature. If T_s is the sum of a trial value (T_s') and a correction (δT_s), and if a similar relationship is used for the saturated vapour pressure at the surface (i.e. $e_s^{0'} = e_s^{0'} + \delta e_s^{0'}$), then equating equations 7.10 and 7.11 gives:

$$R_T - \lambda_a f(u)\left(T_s' + \delta T_s - T_a\right) = f(u)\left(e_s^{0'} + \delta e_s^0 - e_a\right) \qquad 7.12$$

Further assuming $\delta e_s^{0'} = \Delta_s'(T_s')\delta T_s$, where Δ_s' is the slope of the saturation vapour pressure curve at T_s', and rearranging, gives:

$$\delta T_s = \frac{R_T / f(u) + e_a - e_s^{0'} + \lambda_a\left(T_a - T_s'\right)}{\Delta_s'(T_s') + \lambda_a} \qquad 7.13$$

During the iterative process, R_T, *f(u)*, e_a, T_a and λ_a remain constant. T_s' is initially set equal to the air temperature, $e_s^{0'}$ is set equal to the saturation vapour pressure at the air temperature, and Δ_s' is set equal to the slope of the saturation vapour pressure-temperate curve at the air temperature. The iterative process is repeated until the absolute value of δT_s is less than 0.01°C; generally this should be satisfied within 4 iterations. The rate of evapotranspiration is then calculated by using the evaporating surface temperature T_s from the iteration process in equation 7.10.

7.1 Data requirements

To perform a reference evapotranspiration calculation using the Morton iterative variation of the Penman equation, the following daily data are required:

- mean air temperature;
- mean dewpoint (or dry and wet bulb temperature) or relative humidity;
- mean wind velocity at a standard height (for the wind function); and
- measurement of incident radiation or measurement of sunshine hours and cloud cover (from which incident solar radiation can be modelled).

8 Penman-Monteith equation

The Penman combination equation estimates the rate of evapotranspiration from a reference grass crop only. The equation is designed to use purely meteorological data, and as such it does not include a treatment of the physiological behaviour of the plant. Following his landmark paper in 1948, Penman attempted to alter the equation to include the physiological resistances to evapotranspiration. The research culminated in work by Monteith (1965), who derived the Penman-Monteith equation. The philosophy of the approach is to replace crop coefficients with physiological and aerodynamic resistances that better represent the water flux pathways of evapotranspiration.

Monteith worked at the same research establishment in the UK as Penman, and advective effects there would have been small. In more arid climates such as Australia, the air temperature is likely to be higher than the evaporating surface temperature, and a significant proportion of latent heat is likely to come from sensible heat transfer. The effect of advection in this situation must be considered carefully.

8.1 Derivation of the Penman-Monteith equation

Monteith (1965), Monteith & Unsworth (1990), and Wallace (1994) provide the most comprehensive reviews of the Penman-Monteith equation.

The first step is to calculate the evaporation from a wet surface. The resistances are then altered to represent a plant leaf or crop canopy.

8.1.1 Calculation of evaporation from a wet surface

The sensible heat flux to the air is estimated in Appendix A to be:

$$H = \rho_a c_p K_h \frac{\partial \theta}{\partial z} \qquad \text{A1}$$

By using the assumption in Appendix A that the potential temperature can be replaced by the actual temperature, and by integrating the transfer coefficient between the surface and the air, Equation A1 becomes:

$$H \int \frac{1}{K_h} dz = \rho_a c_p \int_{T_a}^{T_s} dT \qquad 8.1$$

$$H = \frac{\rho_a c_p (T_s - T_a)}{\int \frac{1}{K_h} dz} = \frac{\rho_a c_p (T_s - T_a)}{r_h} \qquad 8.2$$

where r_h is the resistance to sensible heat transfer.

The latent heat flux to the air is estimated in Appendix A to be:

$$\lambda E = \rho_a \frac{\lambda (M_w / M_a)}{P} K_w \frac{\partial e_a}{\partial z} \qquad \text{A4}$$

Substituting γ from Equation A6 into Equation A4, assuming the air at the evaporating surface is saturated, and again integrating the transfer coefficient between the surface and the air:

$$\lambda E \int \frac{1}{K_w} dz = \frac{\rho_a c_p}{\gamma} \int_{e_a}^{e_s^0} de \qquad 8.3$$

$$\lambda E = \frac{\frac{\rho_a c_p}{\gamma}\left(e_s^0 - e_a\right)}{\int \frac{1}{K_w} dz} = \frac{\rho_a c_p}{\gamma\, r_v}\left(e_s^0 - e_a\right) \qquad 8.4$$

where r_v is the resistance to latent heat transfer.

In humid conditions, where the evaporating surface temperature is close to the air temperature, equation 5.5 is applicable:

$$\left(T_s - T_a\right) = \frac{\left(e_s^0 - e_a^0\right)}{\Delta} \qquad 5.5$$

Substituting this into equation 8.2 and rearranging gives:

$$H = \frac{\rho_a c_p}{r_h}\frac{\left(e_s^0 - e_a^0\right)}{\Delta} \qquad 8.5$$

$$e_s^0 = \frac{\Delta r_h H}{\rho_a c_p} + e_a^0 \qquad 8.6$$

Equation 8.6 can then be substituted into equation 8.4 to remove the evaporating surface vapour pressure term:

$$\lambda E = \frac{\rho_a c_p}{\gamma\, r_v}\left(\frac{\Delta r_h H}{\rho_a c_p} + e_a^0 - e_a\right)$$

$$\lambda E = \frac{\Delta r_h H + \rho_a c_p\left(e_a^0 - e_a\right)}{\gamma\, r_v} \qquad 8.7$$

The energy balance equation (4.3) can be rearranged and substituted into equation 8.7:

$$H = R_n - G - \lambda E \qquad \text{(rearranged)} \quad 4.3$$

$$\lambda E = \frac{\Delta r_h\left(R_n - G - \lambda E\right) + \rho_a c_p\left(e_a^0 - e_a\right)}{\gamma\, r_v} \qquad 8.8$$

Rearranging gives the equation for evaporation from a wet surface:

$$\lambda E\left(1 + \frac{\Delta r_h}{\gamma r_v}\right) = \frac{\Delta r_h\left(R_n - G\right) + \rho_a c_p\left(e_a^0 - e_a\right)}{\gamma\, r_v}$$

$$\lambda E = \frac{\Delta\left(R_n - G\right) + \frac{\rho_a c_p}{r_h}\left(e_a^0 - e_a\right)}{\Delta + \gamma\frac{r_v}{r_h}} \qquad 8.9$$

Note that if $r_v \approx r_h$, then equation 8.9 simplifies to the Penman combination equation, with:

$$\gamma\; f(u) = \frac{\rho_a c_p}{r_h} \qquad 8.10$$

8.1.2 Calculation of evapotranspiration from a plant canopy

For plant canopies, the resistance to vapour transfer from the canopy to the air can be separated into:

- an aerodynamic resistance, r_a, for the flux path from the vegetation to the air above;

- a bulk surface resistance, r_s, which is mainly affected by the resistance to water vapour transfer from the wet mesophyll cell surfaces through the stomatal opening into the boundary air layer surrounding the leaf. It also includes the resistance to water vapour transfer from the moist or wet ground or leaf surfaces.

The total resistance to water vapour transfer is given by the equation:

$$r_v = r_a + r_s \qquad 8.11$$

If the aerodynamic resistances for sensible and latent heat are assumed to be equal, then:

$$r_a = r_h = r_l \qquad 8.12$$

Substituting equations 8.11 and 8.12 into equation 8.9 gives:

$$\lambda E = \frac{\Delta(R_n - G) + \frac{\rho_a c_p}{r_a}\left(e_a^0 - e_a\right)}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \qquad 8.13$$

This is the **Penman-Monteith** equation.

With this form of evapotranspiration equation it is possible, provided appropriate values of r_s and r_a can be specified, to describe evapotranspiration from any crop, whether well watered or not. Water stress is modelled by increasing the bulk surface resistance r_s.

8.2 Data requirements

To perform a crop evapotranspiration calculation using the Penman-Monteith equation, the following daily data are required:

- mean air temperature;
- mean dewpoint (or dry and wet bulb temperature) or relative humidity;
- measurement of incident radiation or measurement of sunshine hours and cloud cover (from which incident solar radiation can be modelled);
- aerodynamic and bulk surface resistance terms for the crop being analysed; and
- mean wind velocity at a standard height (if required for the aerodynamic resistance term).

Unfortunately, it is not possible to directly measure the resistance terms, so use of the Penman-Monteith equation is mainly restricted to research studies in which the resistance terms are derived as functions of canopy characteristics and wind profile.

9 FAO56 Penman-Monteith equation

A consultation of experts and researchers was organised by the FAO in May 1990 to review the FAO methodologies on crop water requirements and to advise on the revision and update of procedures (Allen, 1998).

The panel of experts recommended the adoption of the Penman-Monteith equation as the new FAO standard for reference evapotranspiration calculations, and advised on procedures for calculation of the various parameters. The reference crop for evapotranspiration calculations was defined as described in Section 1.5.

The FAO Penman-Monteith equation is derived from the Penman-Monteith equation by calculating appropriate values for the aerodynamic and surface resistances for the reference crop. It can be considered as a practical use of the Penman-Monteith equation for a single crop, and therefore has the same assumptions and limitations (e.g. the lack of treatment of advection).

Since crop coefficients are then applied to find the crop evapotranspiration, the philosophy of the FAO approach is more in line with that of the Penman equation than the Penman-Monteith approach.

9.1 Derivation of the FAO Penman-Monteith equation

The derivation below is split into three parts to aid understanding of the assumptions in the resistance calculations. The aerodynamic and bulk surface resistances are calculated first, and then these are substituted into the Penman-Monteith equation to produce the FAO equation. All of these calculations are documented in Allen (1998).

9.1.1 Calculation of the aerodynamic resistance term

The aerodynamic resistance in neutral atmospheric conditions, r_a, can be determined from the equation (Allen, 1998):

$$r_a = \frac{\ln\left(\frac{z_m - d}{z_{om}}\right)\ln\left(\frac{z_h - d}{z_{oh}}\right)}{k^2 u_z} \qquad 9.1$$

Where z_m the height of the wind measurement;

z_h the height of the humidity measurement;

d the zero plane displacement height;

z_{om} is the roughness length governing momentum transfer;

z_{oh} is the roughness length governing heat and vapour transfer;

k is von Karman's constant (0.41); and

u_z is the wind speed at height z.

The zero plane displacement height is the height within the canopy where the distribution of shearing stress over the canopy is aerodynamically equivalent to the imposition of the entire stress at height d (Monteith, 1990). Allen (1998) provides the derivation of the empirical equations which have been developed for estimating d, z_{om} and z_{oh}. For a wide range of crops, it can be estimated from the crop height h by the equation:

$$d = \frac{2}{3}h \qquad 9.2$$

The roughness length governing momentum transfer, z_{om}, can be estimated using:

$$z_{om} = 0.123h \qquad 9.3$$

The roughness length governing transfer of heat and vapour, z_{oh}, can be estimated using:

$$z_{oh} = 0.1 z_{om} \quad 9.3$$

For the reference crop with a constant crop height of 0.12 m and a standardised measurement height at 2 m ($z_m = z_h = 2$ m), the aerodynamic resistance becomes:

$$r_a = \frac{\ln\left(\frac{2 - \frac{2}{3}(0.12)}{0.123(0.12)}\right)\ln\left(\frac{2 - \frac{2}{3}(0.12)}{(0.1)0.123(0.12)}\right)}{0.41^2 u_2}$$

$$r_a = \frac{208}{u_2} \quad 9.4$$

where u_2 is the wind speed (m s^{-1}) measured at a height of 2m.

9.1.2 Calculation of the surface resistance term

The surface resistance, r_s, can be calculated as a function of the leaf area index (LAI):

$$r_s = \frac{r_l}{LAI_{active}} \quad 9.5$$

where r_l is the bulk stomatal resistance of the well-illuminated leaf, and LAI_{active} is the sunlit LAI. Allen (1998) provides a general equation for the sunlit LAI:

$$LAI_{active} = 0.5 LAI \quad 9.6$$

For the reference grass crop, the LAI is found from the equation:

$$LAI = 24h \quad 9.7$$

where h is the crop height (0.12 m). The bulk stomatal resistance of a single leaf is assumed to be about 100 s m^{-1}, so the bulk stomatal resistance for the reference crop is:

$$r_s = \frac{100}{0.5(24)(0.12)} \quad 9.8$$

Hence the bulk surface resistance is assumed to be constant for the reference crop at 70 s/m. The ratio of surface resistance to aerodynamic resistance for the reference crop is:

$$\frac{r_s}{r_a} = \frac{70}{208} u_2 = 0.34 u_2 \quad 9.9$$

9.1.3 FAO Penman-Monteith equation

Rearranging equation 4.23 gives:

$$c_p = \frac{\gamma \lambda M_w}{P M_a} \quad 9.10$$

The ideal gas law can be used to calculate the air density ρ_a

$$\rho_a = \frac{P}{T_{Kv} R} \quad 9.11$$

where R is the specific gas constant (0.287 kJ kg^{-1} K^{-1}), and T_{Kv} is the virtual temperature which is given by:

$$T_{Kv} = 1.01(T_a + 273) \quad 9.12$$

Multiplying equation 9.10 by equation 9.11, substituting in equation 9.12 and dividing by r_a:

$$\frac{\rho_a c_p}{r_a} = \frac{\gamma \lambda M_w}{1.01(T+273) M_a R(208)} u_2$$ with units MJ m^{-2} K^{-1} s^{-1} 9.13

Altering the units to a daily calculation and substituting in M_w / M_a = 0.622 gives:

$$\frac{\rho_a c_p}{r_a} = 86400 \frac{\gamma (0.622) \lambda}{1.01(T+273)(0.287)(208)} u_2$$

$$\frac{\rho_a c_p}{r_a} \approx \gamma \frac{900\lambda}{T+273} u_2$$ with units MJ m^{-2} K^{-1} day^{-1} 9.14

The Penman-Monteith equation is derived in Section 8.1:

$$\lambda E = \frac{\Delta(R_n - G) + \frac{\rho_a c_p}{r_a}\left(e_a^0 - e_a\right)}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{8.13}$$

Substituting equations 9.14 and 9.9 into equation 8.13 gives:

$$\lambda E = \frac{\Delta(R_n - G) + \gamma \frac{900\lambda}{T+273} u_2 \left(e_a^0 - e_a\right)}{\Delta + \gamma(1 + 0.34 u_2)} \tag{9.15}$$

The latent heat of vaporisation, λ, is a weak function of temperature. If λ is assumed to be constant with a value of 2.45 MJ kg^{-1} (valid for a temperature of 20°C), then:

$$E = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 \left(e_a^0 - e_a\right)}{\Delta + \gamma(1 + 0.34 u_2)} \tag{9.16}$$

This is the **FAO Penman-Monteith** equation.

9.1.4 Comparison of the FAO Penman-Monteith and Penman equations

The FAO Penman-Monteith equation is similar to the Penman equation:

$$\lambda E = \frac{\Delta(R_n - G) + \gamma\, f(u)\left(e_a^0 - e_a\right)}{\Delta + \gamma} \tag{5.13}$$

The major difference is that in the FAO equation, both the radiant energy term (R_n – G) and the turbulent transfer term are influenced by wind. Also embedded in the FAO equation is an aerodynamic resistance term; the Penman equation simply uses a wind function surrogate for this term.

9.1.5 Calculation of evapotranspiration from wet leaves

Jensen *et al.* (1989, p44) propose a formula to calculate evapotranspiration for wet leaves that is loosely based on the Penman-Monteith equation:

$$\lambda E = \frac{\Delta(R_n - G) + \frac{\rho_a c_p}{r_a}\left(e_a^0 - e_a\right)}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{8.13}$$

$$\lambda E = \frac{\Delta(R_n - G)}{\Delta + \gamma} + \frac{\rho_a c_p}{\Delta + \gamma} \frac{k^2 u_z}{\left[\ln\left(\frac{z-d}{z_{om}}\right)\right]\left[\ln\left(\frac{z-d}{z_{ov}}\right)\right]} \frac{K_h}{K_m}\left(e_a^0 - e_a\right) \tag{9.17}$$

Since there is water on the leaves, transpiration will be negligible and r_s has been ignored.

The aerodynamic resistance term, r_a, is taken from equation 9.1. K_h and K_m are the eddy transfer coefficients for heat and for momentum, respectively. Note that when $K_h = K_m$, the equation is the Penman-Monteith equation.

9.2 Data requirements

To perform a reference evapotranspiration calculation using the FAO Penman-Monteith equation, the same data are required as for the Penman combination equation:

- mean air temperature;
- mean dewpoint (or dry and wet bulb temperature) or relative humidity;
- measurement of incident radiation or measurement of sunshine hours and cloud cover (from which incident solar radiation can be modelled); and
- mean wind velocity at a standard height.

In addition, crop coefficients are required to calculate the crop evapotranspiration.

10 Wind functions for Penman-type equations

All of the methods described in the previous sections require some form of wind or vapour transfer function. Aside from the notable exception of the FAO Penman-Monteith equation, the function is empirically derived to minimise the discrepancies between predictions and measurements.

Three types of wind function are discussed in this section:

1. the basic Penman wind function, and the coefficients proposed by Meyer for use in inland south eastern Australia;
2. the FAO derivation of aerodynamic resistance as a function of wind;
3. Morton's vapour transfer coefficient (strictly, this is not a wind function as it is independent of wind).

Each method is discussed in the following sections.

10.1 Penman type of wind function

Penman (1948) suggests a wind function with the form:

$$f(u) = a + bu \quad (10.1)$$

where *a* and *b* are constants and *u* is the wind run (usually daily values, km/day).

A number of studies have derived values of *a* and *b* from analysis of measurements. Table 1 lists several suggested values, and also shows the ratio of the two constants for a wind run of 200km/day.

Table 1. Constants *a* and *b* derived by a number of studies, to be used in the equation *f(u) = a +bu*. All of the constants have been converted into units km/day and kPa.

Study	Surface	*a*	*b*	*a* / (*b**200km/day)
Rohwer (from Penman, 1948)	Sunken pans	3.00	0.013	1.14
Penman (1948)	Sunken pans	2.63	0.016	0.82
Doorenbos & Pruitt (1977)	Grass	2.70	0.027	0.50
Stigter (from Rosenberg, 1983)	Grass	3.70	0.023	0.80
Meyer (1988)	Wheat, soybean	17.9	0.044	2.03
Meyer *et al.* (1999)	Wheat, soybean	6.24	0.038	0.82

All of the constants are reasonably consistent with the notable exception of Meyer. While Rohwer and Penman performed their own experiments, Doorenbos & Pruitt and Stigter interpreted a number of studies to find appropriate values. Meyer derived the constants from lysimeter measurements of wheat and soybean at Griffith, NSW.

It is notable that the Meyer constants are higher than the others, particularly in the case of the "*a*" term. Part of the increase might be due to Meyer fitting to crops that behave differently to a grass reference crop. A second, more important, issue is the method of calculation of the vapour pressure deficit (VPD) value $(e_0 - e)$. For both the 1988 and 1999 Meyer coefficients, the mean daily saturation vapour pressure, e_0, is calculated at the mean daily dry bulb temperature. However the 1988 Meyer coefficients relate to an *e* value derived using daily mean dew point temperature while for the 1999 Meyer coefficients the actual daily vapour pressure *e*, was mostly derived using minimum daily relative humidity. These two methods result in quite different estimates of the daily VPD and hence the *a* and *b* coefficients in the wind function will differ. Clearly, the derivation method for daily VPD can have a significant affect on the wind function constants and it is important to know the basis of the VPD derivation when applying the wind function.

A third consideration, relating to the higher constant value seen in the Meyer (1988) and Meyer *et al.* (1999) wind functions, is the difference in climatic conditions between those for which the Penman (and many of the subsequent equations) has been derived and the actual conditions. The Penman equation was largely developed with humid conditions where it is assumed that the evaporating surface temperature is near ambient temperature. This assumption is less often applicable in the hot dry climes of inland Australia. Application of the Penman type equation in low humidity, more advective regions is liable to introduce more error than in more humid and less advective regions.

Doorenbos & Pruitt (1977) caution that the wind function depends on the temporal distribution of wind and relative humidity at the site, and note that using their equation produced a systematic error of 30% in either direction at some experimental sites.

10.2 FAO 56 wind function

The Penman-Monteith equation uses aerodynamic and surface resistance terms instead of an empirical wind function. If the bulk surface resistance is assumed to be negligible (e.g. the leaves are wet), then the Penman-Monteith equation can be simplified to the Penman equation, with the wind function calculated using the equation:

$$f(u) = \frac{\rho_a c_p}{\gamma\, r_a} \qquad 10.2$$

The FAO Penman-Monteith equation derives the aerodynamic resistance term as a function of wind speed (see Section 9.1), so implicitly includes a wind function. The FAO aerodynamic resistance term can be substituted into equation 10.2 to calculate *f(u)*. If the elevation is close to sea level, then:

$$f(u) = 0.085u \qquad 10.3$$

10.3 Morton vapour transfer coefficient

Morton (1983) refers to the wind function as the vapour transfer coefficient f_T. For reasons outlined below, Morton assumes the vapour transfer coefficient is independent of wind speed.

To account for long term atmospheric processes, especially stability, Morton (1983) assumes the form of the vapour transfer coefficient (f_T) to be:

$$f_T = \left(\frac{P_s}{P}\right)^{0.5} \frac{f_Z}{\zeta} \qquad 10.4$$

Where: f_Z is a constant; ζ is a dimensionless stability factor; and P and P_s are the atmospheric pressure and sea-level atmospheric pressure respectively.

The pressure term represents the effect of altitude on evapotranspiration.

The stability factor, ζ takes into account the decrease in the vapour transfer coefficient that occurs when the temperature of the evaporating surface is much below the air temperature. This is usual during winter in extratropical latitudes and during other seasons in very dry environments. The relationship is given as:

$$\frac{1}{\zeta} = 0.28\left(1 + \frac{e_a}{e_a^0}\right) + \frac{\Delta R_{TC}}{\gamma\left(P/P_s\right)^{0.5} b_0 f_Z (e_a^0 - e_a)}$$
$$\frac{1}{\zeta} \le 1 \qquad 10.5$$

b_0 = 1.00 for areal evapotranspiration; R_{TC} = R_T but with $R_{TC} \geq 0$.

The second term of equation 10.5 assumes that the ratio of radiation to vapour transfer in the Penman equation provides a good index of the effects of atmospheric stability on the vapour transfer coefficient. The first term is arbitrarily chosen to reduce non-linearity and scatter in the datasets used by Morton.

Morton assumes that the vapour transfer coefficient f_T is independent of the wind speed because:

1. f_T increases as both surface roughness and wind speed rises, and wind speeds tend to be lower in rough areas than in smooth areas;
2. f_T increases as the instability of the atmosphere rises, and this effect is more pronounced at low wind speeds than at high wind speeds;
3. the use of climatological observations of wind speed can lead to significant error because of local variations in exposure and instrument height.

Hence Morton argues that the use of routinely observed wind speeds in estimating evapotranspiration does not significantly reduce error and may in fact increase it.

Morton derives the constant f_Z by trial and error to meet the following criterion:

$$b_1 \overline{\frac{\Delta_s(T_s) + \gamma}{\Delta_s(T_s) R_{nsur}}} + b_2 = 1.32 \qquad 10.6$$

This equation is discussed in Section 11.2. For now, it is enough to note that Wang *et al.* (2001) have used the Morton method to produce evapotranspiration maps across Australia. They recalibrated the coefficients so that equation 10.6 equalled the commonly used 1.26 (from Priestley & Taylor (1972)) instead of Morton's 1.32, and found:

- f_Z = 29.2 W m^{-2} mbar^{-1}, equivalent to 25.2 MJ m^{-2} kPa^{-1} day^{-1}
- b_1 = 13.4 W m^{-2}, equivalent to 1.2 MJ m^{-2} day^{-1}
- b_2 = 1.13

Since the vapour transfer coefficient is designed for long-term atmospheric conditions, it might not be appropriate to use it for evapotranspiration calculations of periods of less than a month (refer Granger and Gray, 1990).

10.3.1 Calculating the atmospheric pressure term

From Allen *et al.* (1998), atmospheric pressure is calculated using the equation:

$$P = 101.3\left(\frac{293 - 0.0065z}{293}\right)^{5.26} \qquad 10.7$$

Table 2 summarises pressure at different heights. Between 0m and 500m above sea level, the change in pressure is only 6%, which is equivalent to a 2.4% change in equation 9.4. Hence for irrigated areas below 500m (e.g. Riverland, South Australia), it is reasonable to neglect the pressure term if elevation data is not available.

Table 2. Average atmospheric pressure as a function of height above sea level

Height above sea level *z* (m)	**Pressure *p* (kPa)**
0	101.3
100	100.1
200	99.0
300	97.8
400	96.7
500	95.5

10.4 Wind functions, vapour transfer and advection

In arid climates such as Australia, irrigation will generally occur during the dry part of the year. It is a reasonable expectation that most irrigated areas in southern Australia will be influenced by regional and local advective conditions. Northern Australian, winter dry conditions may result in occasional advection over irrigated areas while humid summer conditions close to eastern and northern coastal areas are less likely to have advective situations.

Our expectation is therefore that many irrigated areas, made up of either ribbon development along inland rivers, or as patchworks of irrigated and non-irrigated (dry) paddocks will be subject to advective conditions. In these situations, if the vapour pressure deficit is significantly large and the irrigated crop canopy and aerodynamic resistances are low, then there will be a downwards directed sensible heat flux because the effective canopy surface is cooler than the air immediately above. In this situation, none of the Penman or Penman-Monteith equations include this advective forcing of evapotranspiration. The problem in these situations is that there is a net horizontal flux divergence and the calculation of the resistances may not be correct for this non-equilibrium boundary layer. Local calibration of the Penman equation wind function may partly account for this effect but the calibration may only have limited geographic applicability. For the Penman-Monteith equation it will be important to test general applicability, especially in situations where advection is thought to be significant.

11 Priestley-Taylor equation

All of the methods described previously require knowledge of the mean humidity or dew point, and in most cases the wind speed. This data is not always available. Priestley and Taylor (1972) derived an equation from first principles that described the theoretical limit for evaporation from an extensive, well watered region. In essence it describes the limit of the Penman-Monteith equation when $r_a \rightarrow \infty$ and $r_s \rightarrow 0$.

The Priestley-Taylor equation contains an empirical coefficient *α,* which accounts for water vapour entrainment as air moves over an extensive well watered, evaporating region:

$$\lambda E = \alpha \frac{\Delta(R_n - G)}{\Delta + \gamma} \qquad 11.1$$

A value α = 1.26 was derived from averages of data for both water and wet land surfaces. 1.28 has been used successfully for crops in humid areas (Meyer, 1988). Jury and Tanner (1975) used α = 1.57. As would be expected, the value of *α* needs to increase to account for evapotranspiration in less humid conditions.

This equation is most suitable for estimating areal evapotranspiration from a wet environment, where the effect of local advection is minimal. For example, with α = 1.26, the equation models conversion of around 90% of the net radiation into latent heat and 10% into sensible heat. It is less useful in arid irrigated areas, where the wind speed and the vapour deficit are more likely to be significant variables.

Meyer (1988) and Morton (1983) have suggested variations to equation 11.1, and these are discussed below.

11.1 Meyer variation

Meyer (1988) uses maximum air temperature as a conditioning parameter for α. The empiricism was developed for wheat and soybeans grown at Griffith, NSW.

The ratio $\lambda E/R_n$ was calculated for all days with LAI>3. On days when this ratio was less than or equal to 1, it was likely that there was no sensible heat from advection. The mean maximum air temperature on these days was 19.6°C.

Days when the maximum air temperature (T_{amax}) exceeded 20°C were assumed to have significant local or regional advection, and α was increased as a function of maximum temperature to account for this. The following equations were developed using regression analysis for days where T_{amax} > 20°C:

Wheat	$\alpha = 1.28 + 0.15(T_{a\max} - 20)$	R^2 = 0.06	
Soybeans	$\alpha = 1.28 + 0.09(T_{a\max} - 20)$	R^2 = 0.09	
Combined	$\alpha = 1.28 + 0.11(T_{a\max} - 20)$	R^2 = 0.07	11.2

On days where T_{amax} < 20°C, α = 1.28.

Although the relationship between α and maximum air temperature (T_{amax}) is small, application of this modifier in situations of local or regional advection is helpful. In these situations, the vapour pressure deficit would be expected to be high with this being closely related to temperature. This adjustment therefore involves using one of the variables already employed, thus retaining a real advantage of this equation – the requirement for relatively few data parameters.

11.2 Morton variation

Morton (1983) suggests a variation to estimate monthly areal evapotranspiration in a wet environment, as opposed to the estimation of point evapotranspiration for which the Penman equation is used.

First, he suggests that α = 1.26 is an underestimate for land surfaces because they are rougher and more heterogenous than water surfaces. He suggests that an average of the daily values for land surfaces, α = 1.32, would be more appropriate.

Further, Morton raises two conceptual problems with equation 11.1:

1. "the slope of the saturation vapour pressure curve and the net radiation are temperature-dependent, and both the air and land surface temperatures vary with changes in the availability of water for areal evapotranspiration, and;
2. that it does not take into account the effects of the advection of heat and water vapour associated with large-scale weather systems which can become significant during seasons of low net radiation when potential temperature and specific humidity inversions persist down to the surface for long periods of time."

These concerns are accounted by altering the Priestley-Taylor equation to produce:

$$\lambda E = b_1 + b_2 \frac{\Delta_s (R_{nsur} - G)}{\Delta_s + \gamma} \qquad 11.3$$

where b_1 and b_2 are constants, Δ_s is the slope of the saturation vapour pressure curve at evaporating surface temperature T_s, and R_{nsur} is the net radiation at temperature T_s. Note that Δ_s and R_{nsur} are calculated at a different temperature to Δ and R_n in equation 11.1. Morton calculates the surface temperature using his iterative variation of the Penman equation described in Section 7, and that equation uses his vapour transfer coefficient described in Section 10.3.

The constants b_1 and b_2 are found by trial and error using the equation:

$$b_1 \overline{\frac{\Delta_s + \gamma}{\Delta_s (R_{nsur} - G)}} + b_2 = 1.32 \qquad 11.4$$

where 1.32 is the constant α. Morton proposed using f_z = 28 W m^{-2} mbar^{-1}, b_1 = 14 W m^{-2} and b_2 = 1.20.

Wang *et al.* (1998) have used this method to produce monthly areal wet-environment evapotranspiration maps for Australia. They recalibrated the constants and set them to f_z = 29.2 W m^{-2} mbar^{-1}, b_1 = 13.4 W m^{-2} and b_2 = 1.13, to give a value of α = 1.26 instead of α = 1.32. In terms of the units used in this report, the Wang constants are:

f_z = 25.2 MJ m^{-2} kPa^{-1} day^{-1}

b_1 = 1.2 MJ m^{-2} day^{-1}

b_2 = 1.13

The Morton variation requires a prior calculation of the Morton iterative variation of the Penman equation to calculate R_{ns} and Δ_s. To avoid this, a relationship similar to equation 11.3 could be used:

$$\lambda E = b_1 + b_2 \frac{\Delta (R_n - G)}{\Delta + \gamma} \qquad 11.5$$

Values for b_1 and b_2 would be derived empirically.

11.3 Data requirements

To perform a reference evapotranspiration calculation using any of the Priestley-Taylor variations, the following data are required:

- mean air temperature; and
- measurement of incident radiation or measurement of sunshine hours and cloud cover (from which incident solar radiation can be modelled).

In addition, the maximum air temperature is required for the Meyer variation.

The Morton variation requires the prior calculation of Morton iterative version of the Penman equation, and so requires the same data as the Penman combination equation.

12 Framework for crop evapotranspiration calculations across Australia

A number of methods have been introduced in the previous sections. It is useful to review some of the findings and limitations, with the aim of identifying an appropriate framework for crop evapotranspiration calculations across Australia.

12.1 Review of the input data

All of the methodologies require suitable meteorological data. While there is sufficient data available across Australia to perform the Penman-type calculations, the quality of the data is difficult to assess. The historical data will certainly be of variable quality.

The meteorological data would have to be carefully converted for use in the Penman-type equations:

- the mean air vapour pressure will have to be assessed from the measured 09:00 vapour pressure, possibly by considering regional climatic behaviour;
- daily temperature data will have to be produced from the 09:00 to 09:00 data.

Although the SILO system will be used to produce daily reference evapotranspiration predictions, it might be necessary to average the predictions over a longer period to remove random errors caused by using uneven measuring times. Great care will be necessary to identify and quantify any systematic errors that have been introduced from the data.

The crop coefficient data is specific to the method used to estimate evapotranspiration, and changes throughout the growing season. Planting crops at different times can also affect the crop coefficient. Hence one might expect significant uncertainty in the accuracy of any available crop coefficient.

12.2 Review of the methodologies

Evaporation pan measurements are highly sensitive to local site conditions, and its use in calculating reference evapotranspiration is both highly empirical and regionally variable. Since evaporation pan data implicitly incorporate the effects of regional and local advection, they might represent the only Australia-wide quantifiable source of data for this effect. With this in mind it may be justifiable to collate the regional relationships between pan evaporation and reference evapotranspiration as a way of calibrating crop coefficients in the first instance.

The three variations of the Penman combination equation have broadly the same assumptions and limitations. While the Kohler-Parmele and the Morton iterative approach may produce slightly more accurate estimations than the original equation, it is possible that any improvement would be dwarfed by uncertainties in the meteorological and crop coefficient input data. The more complicated calculations would be more difficult to verify and validate on a day-to-day basis, in particular for the Morton iterative method, and so use of these might be difficult to justify.

The empirical wind function derived by Meyer *et al.* (1999) is different to those derived by other wider studies. It is believed this reflects the correction involved when applying the Penman type equations, which have been formulated for a surface (grass) crop in more humid, less advective environments, to taller crops in less humid and more advective environments. This highlights the need to be aware of the assumptions on which the various evapotranspiration equations are based and that, given these assumptions, the equations may be unsuited to many Australian regions (particularly northern Australia and more humid environments) without regional calibration.

The vapour transfer coefficient developed by Morton is designed to be used over much longer periods (i.e. periods of at least a month) than planned in this project. However, it still might be useful in this project, particularly if suitable wind run data are not available.

The FAO Penman-Monteith equation uses a more generally derived relationship between evapotranspiration and wind, so is more likely to be applicable across Australia. The applicability of this equation for different Australian conditions should be tested with observed data, and compared with values obtained by the locally calibrated Penman equation.

The Priestley-Taylor-type equations are likely to be less accurate than the Penman-type equations, and will probably only be considered if suitable vapour pressure and wind run data are not available. In this case, the Meyer variation of the Priestley-Taylor equation is likely to be most suitable because it implicitly includes an additional energy flux contribution from advection.

12.3 Spatial and temporal considerations

As in most scientific applications, it is likely that the choice of model will be heavily influenced by spatial and temporal considerations.

The meteorological input data is interpolated to a 5km grid resolution, and it is unlikely that suitable data will be available at a higher resolution. Hence the choice of spatial resolution appears to be straightforward. Heat from advection is likely to be the most problematic heterogenous spatial variable, and treatment of this will have to be considered carefully.

The temporal considerations present a more difficult challenge. Meyer (1988) calculated reference evapotranspiration on a daily basis using only meteorological data, and daily estimates have been proposed for the SILO database. However, there are three longer term effects that must be considered for crop evapotranspiration calculations which cannot be derived from daily averaged meteorological data:

1. The rate of evapotranspiration can increase following irrigation or rainfall, possibly due to intercepted water lying in and on the canopy. Since it is extremely difficult to differentiate evaporation and transpiration in measurements, this phenomenon represents a systematic error in the prediction of evapotranspiration from meteorological data. It could be accounted for by increasing the crop coefficient for the time it takes for the free water to evaporate. Analysis of the Meyer data suggests that the fluctuations due to this phenomenon can be removed by averaging the measurements and predictions over 7 days. Such averaging should be performed with care, however, as longer term changes in climate or crop behaviour might be hidden by such an approach.
2. Seasonal climatic changes can cause advection changes. At any particular time, the heat flux from advection will depend on the upwind conditions and hence the wind direction. The total monthly heat flux from advection will depend on the prevailing winds during that month. It is not feasible to directly model advection as a function of upwind conditions, so it will probably be necessary to consider average advection over a longer period. The only realistic method of modelling advection on a daily basis would be to use evaporation pan data to calibrate and modify the reference evapotranspiration.
3. The behaviour of the crop changes during the crop growing season. Normally this is accounted for by changing the crop coefficient during the growing season. Unfortunately, the crop coefficient can be a function of the time of planting due to changing climate. The type of soil can also affect the behaviour of the crop (Meyer, 1988). Accurately modelling these variables could be difficult, so appropriate error assessment will be necessary.

We recognise that the accurate estimation of crop ET on a daily basis is difficult. This report highlights the importance of understanding the limits and uncertainty associated with data and the approximations made in the ET estimation equations. Nonetheless there is still a need to provide daily ET_o for use by water managers and irrigators for water balance estimates. In addition, estimates produced by this project are targeted at policy makers, irrigation scheme planners and academics. The policy makers and irrigation scheme planners are likely to want evapotranspiration data that is averaged on at least a monthly basis, and it would be useful if statistical routines could be incorporated into the SILO system to produce this data.

Another modification would be to add crop coefficients to the SILO system so that crop evapotranspiration (ET_c) can be estimated. This would need a set of look up tables, developed on a regional basis for a range of crops using a degree day growth scale. It may be more appropriate to have such tables in a separate but linked site, which would allow regional updating as the information base improves.

12.4 Proposal for validation of reference ET and estimating matching crop coefficients

Five important modelling issues have emerged in this report:

1. the availability and quality of input data;
2. the availability of suitable experimental data;
3. the modelling of heterogenous spatial advection across Australia;
4. the collation and standardisation of crop coefficients;
5. the integration of several variables with diverse temporal scales.

Below is a brief outline of the proposed approach to institute a “National ET Protocol” which will include the adoption of a national standard for ET estimation and the creation of a listing of irrigated crop coefficients.

This proposed approach assumes that the “patched point” weather datasets available on the SILO website will be the basis for ET calculations. While there are some localities in Australia where more comprehensive (and possibly more accurate) datasets are available, the SILO facility has the advantage of being a widely accessible resource providing a standard dataset for localities across Australia.

The intended outcome of this project is that, for any given Australian irrigation locality, daily reference evapotranspiration (ET_o) can be calculated using SILO data and a standard ET method. Crop coefficients (K_c), which are dependent on crop type and region, can then be applied to obtain daily crop evapotranspiration estimates (ET_c).

12.4.1 Validating and extending reference ET

The following process could be adopted to produce the most appropriate model:

1) Calculate reference ET from SILO patched point datasets for selected sites where independent weather data is also available. Identify any additional errors introduced by limitations in SILO weather data.
2) Compare ET_o from SILO data with ET_o from independent weather data – use FAO 56 and Penman-Meyer calculation methods.
3) Compare estimated ET_o with measured ET for well watered, full canopy crops from contrasting regional environments. It is known that the sources of measured data will be very limited.
4) Compare ET_o, measured ET and pan evaporation form the same sites and from this assess whether pan evaporation could be used to adjust ET_o for local advective conditions.

12.4.2 Estimating matching crop coefficients

1) Identify reasonable regional groupings on the basis of agro-climatic regions. These are likely to be those associated with tropical, summer rainfall and winter rainfall regions. In addition, urban settings associated with more maritime climatic conditions should be considered.

2) Compile and collate as wide a range of existing crop coefficients (K_c) as possible. It will be important to identify the base or reference ET against which the K_c values were established.

3) Convert the different K_c values to the standardised ET_o estimate in SILO. We expect that many existing K_c values have been developed using E_{pan} as the base and so the relation between ET_o and E_{pan} for the appropriate region will be important.

13 Conclusions

Modelling evapotranspiration is a difficult task, particularly across a country as large and diverse as Australia. The difficulty is further increased by the availability of input data and accurate measurements.

A number of methodologies have been reviewed that could be used to calculate reference evapotranspiration. A full annotated derivation has been presented in each case to aid understanding of the assumptions and limitations. It is likely that most of the methodologies would be suitable for the purpose of crop evapotranspiration calculations across Australia.

The main challenges likely to be faced by the project are:

- the availability and quality of meteorological data;
- the availability of suitable experimental data;
- the incorporation of spatially heterogenous advection into the model;
- the production of suitable crop coefficients;
- the choice of temporal scales;
- the assessment of systematic and random uncertainties in the model;
- the development of appropriate reporting tools to summarise the predictions.

It is recommended that the project concentrate effort on these areas rather than the choice of evapotranspiration methodology. The framework described in Section 12.4 can be used to identify the most suitable modelling approach.

The success or failure of the recommended approach, and the project itself, will depend on the accuracy of the estimation with respect to measurements. To reduce the risk of failure or cost overrun, all parts of the model should be tested against a wide range of experimental results prior to the deployment of the model on SILO.

Glossary

ET_O, E	mm day^{-1}	Reference evapotranspiration
ET_C	mm day^{-1}	Crop evapotranspiration
K_C		Crop coefficient multiplier
E_{pan}	mm day^{-1}	Class 'A' pan evaporation
K_p		Class 'A' pan coefficient
u, u_2	m s^{-1}	Wind speed[1]
FET	m	Fetch – length of field upwind of the evaporation pan
RH_{mean}	%	Mean relative humidity
λ	MJ kg^{-1}	Latent heat of vapourisation
$f(u)$	MJ m^{-2} kPa^{-1} day^{-1}	Empirical wind function
e_a	kPa	Actual vapour pressure of air (at the dew point temperature)
e_s	kPa	Actual vapour pressure of air near the evaporating surface (at the dew point temperature)
e_a^0	kPa	Saturation vapour pressure of air (at temperature T_a)
e_s^0	kPa	Saturation vapour pressure of air near the evaporating surface (at temperature T_s)
Q	MJ m^{-2}	Energy stored in the soil-plant system
t	day^{-1}	Time
R_n	MJ m^{-2} day^{-1}	Net radiant energy
G	MJ m^{-2} day^{-1}	Ground heat flux
H	MJ m^{-2} day^{-1}	Sensible heat (enthalpy)
L_p	MJ kg^{-1}	Thermal conversion factor for fixation of carbon dioxide
F_p	kg m^{-2} day^{-1}	Specific flux of CO_2
A_h	MJ m^{-2} day^{-1}	Energy advection into the soil-plant system from precipitation or irrigation
R_{ns}	MJ m^{-2} day^{-1}	Net shortwave radiation
R_{nl}	MJ m^{-2} day^{-1}	Net longwave radiation
R_s	MJ m^{-2} day^{-1}	Incoming shortwave radiation measured near the ground
α		Effective albedo of the surface
R_{ld}	MJ m^{-2} day^{-1}	Downwards longwave radiation from the sky, clouds and aerosols
R_{lu}	MJ m^{-2} day^{-1}	Upwards longwave radiation from the soil-plant interface
ε_{vs}		Effective emissivity of the surface
σ	MJ m^{-2} day^{-1} °C^{-4}	Stefan-Boltzmann constant (4.896 x 10^{-9})
T_s	°C	Evaporating surface temperature

[1] The FAO 56 equation is derived for the wind speed measured at 2m above ground for a crop of height 12cm, and is denoted u_2. There is a conversion equation provided to interpolate to wind speeds measured at other heights.

T_a	°C	Air temperature
R_{ldo}	MJ m^{-2} day^{-1}	Clear day downward longwave radiation
ε_a		Effective emissivity of the atmosphere
ε'		Effective net emissivity ($\varepsilon' = \varepsilon_a - \varepsilon_{vs}$)
R_{nlo}	MJ m^{-2} day^{-1}	Effective longwave radiation on a clear day
a, b, c ,d		Constants in various formulae
DOY		Day of the year
R_{soa}	MJ m^{-2} day^{-1}	Extraterrestrial radiation incident on the upper atmosphere
T_{av}	°C	Mean air temperature over the preceding 3 days
ρ_a	kg m^{-3}	Air density
c_p	MJ kg^{-1} $°C^{-1}$	Specific heat of air at constant pressure
K_h	m s^{-1}	Turbulent exchange coefficient for sensible heat
θ	°C	Potential temperature
z	m	Spatial direction perpendicular to the surface
M_w	amu	Molecular weight of water vapour
M_a	amu	Molecular weight of air
P	kPa	Atmospheric pressure
K_w	m s^{-1}	Turbulent exchange coefficient for water vapour
P_s	kPa	Sea level or standard pressure (≈100 kPa)
R	kJ kg^{-1} $°C^{-1}$	Specific gas constant for air (R = 0.287 kJ kg^{-1} $°C^{-1}$ for dry air)
γ	kPa $°C^{-1}$	Psychrometric constant
β		Bowen ratio
Δ	kPa $°C^{-1}$	Slope of saturation vapour pressure curve, calculated at the air temperature[1]
E_a	MJ m^{-2} day^{-1}	Arbitrary mass transfer function defined by Penman
R_{ir}	MJ m^{-2} day^{-1}	Difference between the incident and reflected radiation to the surface
R_T	MJ m^{-2} day^{-1}	Net radiation term in the Morton iterative variation of the Penman equation
λ_a		Arbitrary constant defined in the Morton iterative variation of the Penman equation
r_h	s m^{-1}	Resistance to sensible heat flux
r_v	s m^{-1}	Total resistance to latent heat flux
r_a	s m^{-1}	Aerodynamic (or canopy) resistance
r_s	s m^{-1}	Surface resistance
r_l	s m^{-1}	Aerodynamic resistance to latent heat flux
z_m	m	Height of the wind measurements
z_h	m	Height of the humidity measurements

[1] Different methods are used to calculate Δ in the Meyer *et al.* (1998) and Allen *et al.* (1998) reports, although these give almost identical results in the temperature range of interest.

z_{om}	m	Roughness length governing momentum transfer
z_{oh}	m	Roughness length governing heat and vapour transfer
d	m	Zero plane displacement height, also used in the calculation of atmospheric emissivity as a constant
k	Dimensionless	von Karman's constant
h	m	Crop height
LAI	Dimensionless	Leaf area index
LAI_{active}	Dimensionless	Sunlit leaf area index
r_l	s m^{-1}	Bulk stomatal resistance of a well-illuminated leaf
T_{Kv}	°C	Virtual temperature
f_T	MJ m^{-2} kPa^{-1} day^{-1}	Morton's vapour transfer coefficient
f_z	MJ m^{-2} kPa^{-1} day^{-1}	Constant used by Morton
ζ		Dimensionless stability constant used by Morton
b_0		Constant (from Morton, b_0 = 0 for areal evapotranspiration)
R_{TC}	MJ m^{-2} day^{-1}	Net radiation term in the stability constant calculation
Δ_s	kPa °C^{-1}	Slope of saturation vapour pressure curve calculated at the evaporating surface temperature
R_{nsur}	MJ m^{-2} day^{-1}	Net radiation term in the Morton variation of the Priestley-Taylor equation, calculated at evaporating surface temperature T_s
b_1, b_2		Constants in the Morton variation of the Priestley-Taylor equation

References

Allen, RG, Pereira, LS, Raes, D, Smith, M 1998, 'Crop evapotranspiration: Guidelines for computing crop water requirements', I*rrigation and Drainage Paper 56,* Food and Agricultural Organisation of the United Nations.

Bowen, IS 1926, 'The Ratio of Heat Losses by Conduction and by Evaporation from any Water Surface', *Physical Review*, vol. 27, no. 6, pp. 779–789.

Brutsaert, W 1982, *Evaporation into the atmosphere: theory, history and applications,* Dordrecht, Holland.

BoM 2005, 'Daily solar radiation model description', Bureau of Meteorology internet site, www.bom.gov.au/sat/solradinfo.shtml, viewed May 2005.

Doorenbos, J & Pruitt, WO 1975, 'Guidelines for predicting crop water requirements', I*rrigation and Drainage Paper 24,* Food and Agricultural Organisation of the United Nations.

Fairweather, H 2005, Private communication.

Granger, RJ and Gray, DM 1990, 'Examination of Morton's CRAE model for estimating daily evaporation from field-sized areas', *Journal of Hydrology*, vol. 120, pp. 309-325.

Jensen, ME, Wright, JL, Pratt, BJ 1973, 'Estimating soil moisture depletion from climate, crop and soil data', *Transactions of the American Society of Agricultural Engineering*, vol. 14, pp. 954–959.

Jones, FE 1992, *Evaporation of water: with emphasis on applications and measurements*, 1st edn, Lewis Publishers, Chelsea, Michigan.

Jury, WA & Tanner, CB 1975, 'Advection modification of the Priestley and Taylor evapotranspiration formulae', *Agronomy Journal*, vol. 67, pp. 840–842.

Kohler, MA & Parmele, LH 1967, 'Generalized Estimates of Free-Water Evaporation', *Water Resources Research*, vol. 3, no. 4, pp. 997–1005.

McIlroy, IC & Angus, DE 1964, 'Grass, Water and Soil Evaporation at Aspendale', *Agricultural Meteorology*, vol. 1, pp. 201–224.

Meyer, WS 1988, 'Development of Management Strategies for Minimizing Salination due to Irrigation: Quantifying Components of the Water Balance under Irrigated Crops', *AWRAC Research Project 84/162*.

Meyer, WS, Smith, DJ, Shell, G 1999, 'Estimating reference evaporation and crop evapotranspiration from weather data and crop coefficients', *Technical Report 34/98*, CSIRO Land and Water.

Meyer, WS 1999, 'Standard Reference Evaporation Calculation for Inland, South Eastern Australia', *Technical Report 35/98*, CSIRO Land and Water.

Monteith, JL 1965, 'Evaporation and Environment', *19th Symposia of the Society for Experimental Biology*, vol. 19, pp 205–234.

Monteith, JL & Unsworth, MH 1990, *Principles of Environmental Physics*, Second Edition, Edward Arnold, London.

Morton, FI 1965, 'Potential Evaporation and River Basin Evaporation', *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers*, vol. 91, no. HY6, pp. 67–97.

Morton, FI 1983, 'Operational Estimates of Areal Evapotranspiration and their Significance to the Science and Practice of Hydrology', *Journal of Hydrology*, vol. 66, pp. 1–76.

Oke, TR 1987, *Boundary Layer Climates*, Second Edition, Methuen, London.

Penman, HL 1948, ‘Natural evaporation from open water, bare soil and grass’, *Procedures of the Royal Society of London, Series A*, vol. 193, pp. 120–145.

Priestley, CHB & Taylor, RJ 1972, ‘On the Assessment of Surface Heat Flux and Evaporation Using Large-Scale Parameters’, *Monthly Weather Review*, vol. 100, no. 2, 81–92.

Rosenberg, NJ, Blad, BL, Verma, SB 1983, *Microclimate: The Biological Environment*, 2nd edn, John Wiley & Sons, New York.

Tanner, CB 1957, ‘Factors affecting evaporation from plants and soils’, *Journal of Soil Water Conservation*, vol. 12, pp. 221–227.

Wallace, JS 1994, ‘Calculating evaporation: resistance to factors’, *Agricultural and Forest Meteorology*, vol. 73, pp. 353–366.

Wang, QJ, McConachy, FLN, Chiew, FHS, James, R, deHoedt, GC, Wright, WJ 2001, ’Climatic atlas of Australia: Maps of Evapotranspiration’, *Bureau of Meteorology*.

Appendix A: The Bowen ratio

The transfer of water vapour and sensible heat to the air can be modelled as being caused by small eddies within the vegetation canopy. From aerodynamic theory, the sensible heat and water vapour fluxes can be estimated using the equations (Rosenberg *et al.*, 1983, p142):

$$H = \rho_a c_p K_h \frac{\partial \theta}{\partial z} \tag{A1}$$

$$E = \rho_a \frac{(M_w / M_a)}{P} K_w \frac{\partial e_a}{\partial z} \tag{A2}$$

where ρ_a is the air density,

c_p is the specific heat at constant pressure,

P is the atmospheric pressure,

K_h is the turbulent exchange coefficient for sensible heat,

K_w is the turbulent exchange coefficients for water vapour,

$\partial\theta/\partial z$ is the vertical gradient of potential temperature,

$\partial e_a/\partial z$ is the vertical gradient of vapour pressure,

M_w is the molecular weight of water vapour, and

M_a is the molecular weight of air.

The potential temperature, θ, is the temperature that a dry air parcel would have if transported adiabatically from its ambient temperature and pressure to a pressure of 1000 mbar. It is calculated using the equation:

$$\theta = T \left(\frac{P_s}{P} \right)^{R/c_p} \tag{A3}$$

where P_s is the standard pressure (1000mbar, close to the sea level atmospheric pressure) and R is the ideal gas constant. If the evaporating surface is close to sea level, then P_s/P is close to 1 and the potential temperature, θ can be replaced with the temperature, T.

The latent heat flux is calculated by multiplying equation 4.19 by the latent heat of vaporisation (λ):

$$\lambda E = \rho_a \frac{\lambda (M_w / M_a)}{P} K_w \frac{\partial e_a}{\partial z} \tag{A4}$$

The ratio of the sensible heat flux to the latent heat flux is called the Bowen ratio, after Bowen (1926):

$$\beta = \frac{H}{\lambda E} = \frac{c_p P M_a}{\lambda M_w} \frac{K_h}{K_w} \frac{\partial T / \partial z}{\partial e_a / \partial z} \tag{A5}$$

The first fraction is called the psychrometric constant (γ):

$$\gamma = \frac{c_p P M_a}{\lambda M_w} \tag{A6}$$

γ is actually not a constant, as it depends on the atmospheric pressure P and is weakly related to temperature through the latent heat of vaporisation.

If stable atmospheric conditions are assumed (generally when the evaporating surface temperature is lower than that of the air immediately above), then $K_h/K_w \approx 1$. Hence equation 4.22 can be rewritten:

$$\beta = \frac{H}{\lambda E} = \gamma \frac{\partial T / \partial z}{\partial e_a / \partial z} \qquad \text{A7}$$

If the temperature and vapour pressure gradients from the evaporating surface to the air are assumed to be linear, then equation 4.24 can be further simplified to:

$$\beta = \frac{H}{\lambda E} = \gamma \frac{(T_s - T_a)}{(e_s - e_a)} \qquad \text{A8}$$

Unfortunately, atmospheric conditions near the evaporating surface are not always stable as the result of the surface temperature being greater that the immediately connected air, and in this situation the Bowen ratio assumptions do not apply. Also, the approximation that $\partial T/T_z = (T_s - T_a)$ is likely to apply only over small distances which makes the accurate measurement of tiny differences very demanding.