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Estimating actual, potential, reference crop and pan evaporation using standard meteorological data: a pragmatic synthesis

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Abstract

This guide to estimating daily and monthly actual, potential, reference crop and pan evaporation covers topics that are of interest to researchers, consulting hydrologists and practicing engineers. Topics include estimating actual evaporation from deep lakes and from farm dams and for catchment water balance studies, estimating potential evaporation as input to rainfall-runoff models, and reference crop evapotranspiration for small irrigation areas, and for irrigation within large irrigation districts. Inspiration for this guide arose in response to the authors' experiences in reviewing research papers and consulting reports where estimation of the actual evaporation component in catchment and water balance studies was often inadequately handled. Practical guides using consistent terminology that cover both theory and practice are not readily available. Here we provide such a guide, which is divided into three parts. The first part provides background theory and an outline of conceptual models of potential evaporation of Penman, Penman-Monteith and Priestley-Taylor, and discussions of reference crop evaporation and then Class-A pan evaporation. The last two sub-sections in this first part include techniques to estimate actual evaporation from (i) open-surface water and (ii) landscapes and catchments (Morton and the advection-aridity models). The second part addresses topics confronting a practicing hydrologist, e.g. estimating actual evaporation for deep lakes, shallow lakes and farm dams, lakes covered with vegetation, catchments, irrigation areas and bare soil. The third part addresses six related issues (i) hard-wired evaporation estimates, (ii) evaporation estimates without wind data, (iii) at-site meteorological data, (iv) dealing with evaporation in a climate change environment, (v) 24-h versus day-light hour estimation of meteorological variables, and (vi) uncertainty in evaporation estimates.

This paper is supported by supplementary material that includes 21 appendices enhancing the material in the text, worked examples of many procedures discussed in the paper, a program listing (Fortran 90) of Morton's WREVAP evaporation models along

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with tables of monthly Class-A pan coefficients for 68 locations across Australia and other information.

1 Introduction

Actual evaporation is a major component in the water balance of a catchment, reservoir or lake, irrigation region, and some groundwater systems. For example, across all continents evapotranspiration is 70 % of precipitation, and varies from over 90 % in Australia to approximately 60 % in Europe (Baumgarter and Reichel, 1975, Table 12). For major reservoirs in Australia, actual evaporation losses represent 20 % of reservoir yield (Hoy and Stephens, 1979, p. 1). Compared with precipitation and streamflow, the magnitude of actual evaporation over the long term is more difficult to estimate than either precipitation or streamflow.

This paper deals with estimating actual, potential, reference crop and pan evaporation at a daily and a monthly time-step using standard meteorological data. A major discussion of the use of remotely sensed data to estimate actual evaporation is outside the scope of this paper but readers interested in the topic are referred to Kalma et al. (2008) and Glenn et al. (2010) for relevant review material.

The inspiration for the paper, which is a considered summary of techniques rather than a review, arose because over recent years the authors have reviewed many research papers and consulting reports in which the estimation of the actual evaporation component in catchment and water balance studies was inadequately handled. Our examination of the literature yielded few documents covering both theory and practice that are readily available to a researcher, consulting hydrologist or practicing engineer. Chapter 7 *Evapotranspiration* in *Physical Hydrology* by Dingman (1992), Chapter 4 *Evaporation* (Shuttleworth, 1992) in the Handbook of Hydrology (Maidment, 1992) and, for irrigated areas, *FAO 56 Crop evapotranspiration: Guidelines for computing crop water requirements* (Allen et al., 1998) are helpful references. We refer heavily to these texts in this paper which is aimed at improving the practice of estimating actual and

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potential evaporation using standard daily or monthly meteorological data. This paper is not intended to be an introduction to evaporation processes. Dingman (1992) provides such an introduction. Readers, who wish to develop a strong theoretical background of evaporation processes, are referred to *Evaporation into the Atmosphere* by Brutsaert (1982), and to Shuttleworth (2007) for an historical perspective.

There are many practical tasks in which daily or monthly actual or potential evaporation needs to be estimated including for a deep lake or post-mining void, for a shallow lake or farm dam, for a catchment water balance study (in which actual evaporation may be land-cover specific or lumped depending on the style of analysis or modelling), as input to a rainfall-runoff model, or for a small irrigation area or for irrigated crops within a large irrigation district. Each of these tasks illustrates most of the practical issues that arise in estimating daily or monthly evaporation from meteorological data or from Class-A evaporation pan measurements. These tasks are used throughout the paper as a basis to highlight common issues facing practitioners.

Following this introduction, Sect. 2 describes the background theory and models under five headings: (i) potential evaporation, (ii) reference crop evapotranspiration, (iii) pan evaporation, (iv) open-surface water evaporation and (v) actual evaporation from landscapes and catchments. Practical issues in estimating actual evaporation from deep lakes, reservoirs and voids, from shallow lakes and farm dams, for catchment water balance studies, in rainfall-runoff modelling, from irrigation areas, from lakes covered by vegetation, bare soil, and groundwater are considered in Sect. 3. This section concludes with a guideline summary of preferred methods to estimate evaporation. Section 4 deals with several outstanding issues of interest to practitioners and, in the final section (Sect. 5), a concluding summary is provided. Readers should note there are 21 appendices in the supplementary material where more model details and worked examples are provided. (Appendices, tables and figures in the Supplement are indicted by an S before the caption number.)

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Definitions, time-step, units and input data

The definitions, time-steps, units and input data associated with estimating evaporation and used throughout the literature vary and, in some cases, can introduce difficulties for practitioners who wish to compare various approaches. Throughout this paper, consistent definitions, time-steps and units are adopted.

Evaporation is a collective term covering all processes in which water as liquid is transferred as water vapour to the atmosphere. The term includes evaporation of water from lakes and reservoirs, from bare soils, as well as from water intercepted by vegetative surfaces. Transpiration is the evaporation from within the leaves of a plant (Dingman, 1992, Sect. 7.5.1). This paper does not deal with sublimation from snow or ice.

Savenije (2004) argues that because actual evaporation of interception is a considerable proportion of total evaporation from vegetation, particularly in warm climates, the term evapotranspiration is misleading. This approach is consistent with Shuttleworth's (1992) chapter in which the term evapotranspiration is not used. However, we have retained the term "evapotranspiration" where we refer to literature in which the term is used, for example when discussing reference crop evapotranspiration.

Throughout the paper, unless otherwise stated, pan evaporation means a Class-A evaporation pan with a standard screen. A Class-A evaporation pan, which was developed in the United States and is used widely throughout the world, is a circular pan (1.2 m in diameter and 0.25 m deep) constructed of galvanised iron and is supported on a wooden frame 30 mm to 50 mm above the ground (WMO, 2006, Sect. 10.3.1). In Australia, a standard wire screen covers the water surface to prevent water consumption by animals and birds (Jovanovic et al., 2008, Sect. 2).

In this paper, the term "lake" includes lakes, reservoirs and voids (as a result of surface mining) and is defined, following Morton (1983b, p. 84), as a body of water so wide that the effects of upwind advections are negligible unless otherwise specified. Furthermore, Morton distinguishes between shallow and deep lakes, the former being

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one in which seasonal heat storage changes are insignificant. Deep lakes may also be considered shallow if one is interested only in annual or mean annual evaporation because at those time-steps seasonal heat storage changes are considered unimportant (Morton, 1983b, Sect. 2). However, for other procedures there is no clear distinction between shallow and deep lakes (see Table S5) and, therefore, we have identified them as shallow or deep in terms of the author's own description.

Because of the scope of evaporation topics across analyses and measurements, we deliberately restrict the content of the paper to techniques that can be applied at a daily and/or monthly time-step. Under each method we set out the time-step that is appropriate. Dealing with shorter time-steps, say one hour, is mainly a research issue and is beyond the scope of this paper.

In the literature, there is little consistency in the units for the input data, constants and variables. Here, except for several special cases, we use a consistent set of units and have adjusted the empirical constants accordingly. The adopted units are: evaporation in mm per unit time, pressure in kPa, wind speed in $\mathrm{m\,s^{-1}}$ averaged over the unit time, and radiation in $\mathrm{MJ\,m^{-2}}$ per unit time. Furthermore, we distinguish between measurements that are cumulated or averaged over 24 h, denoted as "daily" values, and those that are cumulated or measured during day-light hours, designated as "day-time" values (Van Niel et al., 2011).

Evaporation can be expressed as depth per unit time, e.g. $\mathrm{mm\,day^{-1}}$, or expressed as energy during a day and, noting that the latent heat of water is $2.45\,\mathrm{MJ\,kg^{-1}}$ (at $20\,^{\circ}\mathrm{C}$) it follows that $1\,\mathrm{mm\,day^{-1}}$ of evaporation equals $2.45\,\mathrm{MJ\,m^{-2}\,day^{-1}}$.

The evaporation models, discussed in this paper including Penman, Penman-Monteith, Priestley-Taylor, reference crop evaporation, PenPan, Morton and Advection-Aridity models, require a range of meteorological and other data as input. The data required are highlighted in Table 1 along with the time-step for analysis and the sections in the paper where the models are discussed. Availability of input data is discussed in Appendix S1. Appendices S2 and S3 list the equations for calculating the meteorological variables like saturation vapour pressure, and net radiation. Values of specific

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constants like the latent heat of vaporization, aerodynamic and surface resistances, and albedo values are listed in Tables S1, S2, and S3, respectively.

2 Background theory and models

The evaporation process over a vegetated landscape is linked by two fundamental equations – a water balance equation and an energy balance equation as follows:
Water balance

$$\bar{P} = \bar{E}_{\text{Act}} + \bar{Q} + \Delta S \quad (1a)$$

$$\bar{P} = \left(\bar{E}_{\text{Soil}} + \bar{E}_{\text{Trans}} + \bar{E}_{\text{Inter}}\right) + \bar{Q} + \Delta S \quad (1b)$$

Energy balance

$$\bar{R} = \bar{H} + \lambda \bar{E}_{\text{Act}} + \bar{G} \quad (2)$$

where, during a specified time period, e.g. one month, and over a given area, $\bar{P}$ is the mean rainfall (mm day^{-1}), $\bar{E}_{\text{Act}}$, $\bar{E}_{\text{Soil}}$, $\bar{E}_{\text{Trans}}$, and $\bar{E}_{\text{Inter}}$ are, respectively the mean actual evaporation (mm day^{-1}), the mean evaporation from the soil (mm day^{-1}), the mean transpiration (mm day^{-1}) and mean evaporation of intercepted precipitation (mm day^{-1}), $\bar{Q}$ is the mean runoff (mm day^{-1}), ΔS is the change in soil moisture storage (mm day^{-1}), $\bar{R}$ is the mean net radiation received at the soil/plant surfaces (MJ m^{-2} day^{-1}), $\bar{H}$ is the mean sensible heat flux (MJ m^{-2} day^{-1}), $\lambda \bar{E}_{\text{Act}}$ is the outgoing energy (MJ m^{-2} day^{-1}) as mean actual evaporation, $\bar{G}$ is the mean heat conduction into the soil (MJ m^{-2} day^{-1}), and λ is the latent heat of vaporisation (MJ m^{-2}). Models used to estimate evaporation are based on these two fundamental equations.

This section covers five types of models. Section 2.1 (Potential evaporation) discusses the conceptual basis for estimating potential evaporation which is followed by Sect. 2.2 (Reference crop evaporation) where estimating evaporation for reference crop conditions is considered. Section 2.3 (Pan evaporation) deals with the measurement

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and modelling of evaporation by a Class-A evaporation pan. Section 2.4 (Open-surface water evaporation) discusses actual evaporation from open-surface water of shallow lakes, deep lakes (reservoirs) and large voids. Finally, in Sect. 2.5 (Actual evaporation (from catchments)) actual evaporation from landscapes and catchments, where soil moisture limits soil evaporation and transpiration, is discussed.

2.1 Potential evaporation

In 1948, Thornthwaite (1948, p. 56) coined the term "potential evapotranspiration", the same year that Penman (1948) published his approach for modelling evaporation for a short green crop completely shading the ground. Penman (1956, p. 20) called this "potential transpiration" and since then there have been many definitions and redefinitions of the term potential evaporation or evapotranspiration.

In a detailed review, Granger (1989a, Table 1) (see also Granger, 1989b) examined the concept of potential evaporation and identified five definitions, but considered only three to be useful, which he labelled EP2, EP3 and EP5. They are related as:

$$\mathrm{EP5} \geq \mathrm{EP3} \geq \mathrm{EP2} \geq E_{\mathrm{Act}} \tag{3}$$

where E_{Act} is the actual evaporation rate. EP2, which is known as the equilibrium evaporation rate (see Sect. 2.1.4), is defined by only the available energy and represents the lower limit of actual evaporation from a wet surface. It is the first term in the Penman equation (Eq. 4). EP3 is equivalent to the Penman evaporation from a free-water surface and is dependent on available energy and atmospheric conditions. Granger (1989a, Table 1) denotes EP5 as "potential evaporation" that represents an upper limit of evaporation. It is defined by both the atmospheric conditions as well as the saturated vapour pressure at the actual *surface* temperature.

In the above context it is noted that Katerji and Rana (2011) argue that the concept of potential evapotranspiration is inadequate when applied to vegetated surfaces as evapotranspiration consists of two processes acting in opposite directions – evaporative demand, on the one hand, and canopy resistance which reduces the supply

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to the other. However, notwithstanding the previous comment, in this paper we have adopted the definition of Dingman (1992, Sect. 7.7.1) namely that "potential evapotranspiration ... is the rate at which evapotranspiration would occur from a large area completely and uniformly covered with growing vegetation which has access to an unlimited supply of soil water, and without advection or heating effects". Two other terms need to be defined – reference crop evapotranspiration and actual evaporation. Reference crop evapotranspiration or reference evapotranspiration is the evapotranspiration from a prescribed reference vegetated surface which is not short of water (Allen et al., 1998, p. 7) (see Sect. 2.2). The second term is actual evaporation (or actual evapotranspiration) which is defined as the quantity of water that is transferred as water vapour to the atmosphere from the evaporating surface (Wiesner, 1970, p. 5).

2.1.1 Penman

In 1948, Penman was the first to combine an aerodynamic approach for estimating potential evaporation with an energy equation based on net incoming radiation. This approach eliminates the surface temperature variable, which is not a standard meteorological measurement, resulting in the following equation, known as the Penman or Penman combination equation, to estimate potential evaporation (Penman, 1948, Eq. 16; see also Shuttleworth, 1992, Sect. 4.2.6; Dingman, 1992, Sect. 7.3.5):

$$E_{\text{Pen}} = \frac{\Delta}{\Delta + \gamma}\frac{R_{\text{n}}}{\lambda} + \frac{\gamma}{\Delta + \gamma}E_{\text{a}} \quad (4)$$

where E_{Pen} is the daily potential evaporation (mm day^{-1}) from a saturated surface, R_{n} is net daily radiation to the evaporating surface (MJ m^{-2} day^{-1}) where R_{n} is dependent on the evaporating surface albedo (Appendix S3), E_{a} (mm day^{-1}) is a function of the average daily wind speed (m s^{-1}), saturation vapour pressure (kPa) and average vapour pressure (kPa), Δ is the slope of the vapour pressure curve (kPa °C^{-1}) at air temperature, γ is the psychrometric constant (kPa °C^{-1}), and λ is the latent heat of

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vaporization (MJ kg^{-1}). The Penman equation assumes no heat exchange with the ground, no water-advected energy, and no change in heat storage (Dingman, 1992, Sect. 7.3.5). Penman (1956, p. 18) and Monteith (1981, 4 and 5 pp.) provide helpful discussions of the dependence of latent heat flux on surface temperature. Application of the Penman equation is discussed in Sect. 2.4.1 with further details provided in Appendix S4.

The Penman approach has spawned many other procedures (e.g. Priestley and Taylor, 1972; see Sect. 2.1.3) including the incorporation of resistance factors that extend the general method to vegetated surfaces. The Penman-Monteith formulation described in the following section is an example of the latter.

2.1.2 Penman-Monteith

The Penman-Monteith model, defined as Eq. (5), is usually adopted to estimate potential evaporation from a vegetated surface. Like Penman's equation, the Penman-Monteith depends on the unknown surface temperature of the evaporating surface (Monteith, 1965). Raupach (2001, p. 1154) provides a detailed discussion of the approaches to eliminate the surface temperature from the surface energy balance equations. The simplest solution results in the following well known Penman-Monteith equation (Allen et al., 1998, Eq. 3):

$$\mathrm{ET}_{\mathrm{PM}} = \frac{1}{\lambda} \frac{\Delta (R_\mathrm{n} - G) + \rho_\mathrm{a} c_\mathrm{a} \frac{(v_\mathrm{a}^* - v_\mathrm{a})}{r_\mathrm{a}}}{\Delta + \gamma \left(1 + \frac{r_\mathrm{s}}{r_\mathrm{a}}\right)} \tag{5}$$

where $\mathrm{ET_{PM}}$ is the Penman-Monteith potential evapotranspiration (mm day^{-1}), R_n is the net daily radiation at the vegetated surface (MJ m^{-2} day^{-1}), G is the soil heat flux (MJ m^{-2} day^{-1}), ρ_a is the mean air density at constant pressure (kg m^{-3}), c_a is the specific heat of the air (MJ kg^{-1} $°C^{-1}$), r_a is an "aerodynamic or atmospheric resistance" to water vapour transport (s m^{-1}) for neutral conditions of stability (Allen et al., 1998,

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p. 20), r_s is a "surface resistance" term ($\mathrm{s\,m^{-1}}$), $(v_a^* - v_a)$ is the vapour pressure deficit (kPa), λ is the latent heat of vaporization ($\mathrm{MJ\,kg^{-1}}$), Δ is the slope of the saturation vapour pressure curve ($\mathrm{kPa\,^{\circ}C^{-1}}$) at air temperature, and γ is the psychrometric constant ($\mathrm{kPa\,^{\circ}C^{-1}}$). Values of r_a and r_s are discussed in Appendix S5.

2.1.3 Priestley-Taylor

The Priestley-Taylor equation (Priestley and Taylor, 1972, Eq. 14) allows potential evaporation to be computed in terms of energy fluxes without an aerodynamic component as follows:

$$E_{PT} = \alpha_{PT}\left[\frac{\Delta}{\Delta+\gamma}\frac{R_n}{2.45} - \frac{G}{2.45}\right] \tag{6}$$

where E_{PT} is the Priestley-Taylor potential evaporation ($\mathrm{mm\,day^{-1}}$), R_n is the net daily radiation at the evaporating surface ($\mathrm{MJ\,m^{-2}\,day^{-1}}$), G is the soil flux into the ground ($\mathrm{MJ\,m^{-2}\,day^{-1}}$), Δ is the slope of the vapour pressure curve ($\mathrm{kPa\ ^{\circ}C^{-1}}$) at air temperature, and γ is the psychrometric constant ($\mathrm{kPa\,^{\circ}C^{-1}}$). α_{PT} is the Priestley-Taylor constant.

Based on field data, Priestley and Taylor (1972, Sect. 6) adopted $\alpha_{PT} = 1.26$ for "advection-free" saturated surfaces. Eichinger et al. (1996, p. 163) developed an analytical expression for α_{PT} and found that 1.26 was an appropriate value for wet surfaces. Lhomme (1997) developed a theoretical basis for the Priestley-Taylor coefficient of 1.26 for non-advective conditions. Based on field data in northern Spain, Castellvi et al. (2001) found that α_{PT} for Penman-Monteith reference crop rather than for water exhibited large seasonal (up to 27 %) and spatial ($\alpha_{PT} = 1.35$ to 1.67) variations. Improved performance was achieved by including adjustments for vapour pressure deficit and available energy. Pereira (2004), noting the analysis by Monteith (1965, p. 220) and Perrier (1975), considered the hypothesis $\alpha_{PT} = \Omega^{-1}$ where Ω is a decoupling coefficient and is a function of the aerodynamic and surface resistances, implying α_{PT} is not

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a constant. The decoupling coefficient is discussed in Appendix S5. Values of α_{PT} for a range of surfaces are listed in Table S8 and it is noted that α_{PT} values are dependent on the observation period, daily (24 h) or day-time. Priestley and Taylor (1972, Sect. 1) adopted a daily time-step for their analysis.

2.1.4 Equilibrium evaporation

Slatyer and McIlroy (1961) developed the concept of equilibrium evaporation (E_{EQ}) in which air passing over a saturated surface will gradually become saturated until an equilibrium rate of evaporation is attained. Edinger et al. (1968) defined equilibrium temperature as the surface temperature of the evaporating surface at which the net rate of heat exchange is zero. But because of the daily cycles in the meteorological conditions, equilibrium temperature is never achieved (Sweers, 1976, p. 377).

Stewart and Rouse (1976, Eq. 4) interpreted the Slatyer and McIlroy (1961) concept in terms of the Priestley and Taylor (1972) equation as

$$E_{EQ} = \frac{1}{\alpha_{PT}} E_{PT} \tag{7}$$

where E_{PT} and α_{PT} are defined in the previous section. McNaughton (1976) proposed a similar argument. However, based on lysimeter data Eichinger et al. (1996) question this concept of equilibrium evaporation and suggest that the Priestley-Taylor equation with $\alpha_{PT} = 1.26$ is more representative of equilibrium evaporation under wet surface conditions. In 2001, Raupach (2001) carried out a historical review and theoretical analysis of the concept of equilibrium evaporation. He concluded that for any closed evaporating system with steady energy supply, the system moves towards a quasi-steady state in which the Bowen Ratio (β) takes the equilibrium value of $\frac{1}{\varepsilon}$ where ε is the ratio of latent to sensible heat contents of saturated air in a closed system. Raupach (2001) also concluded that open systems cannot reach equilibrium.

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2.1.5 Other methods for estimating potential evaporation

There are many other potential evaporation equations proposed and evaluated during the past 100 or so years that could have been included in this paper. Some of these, e.g. Thornthwaite (Thornthwaite, 1948) and Makkink models (de Bruin, 1981, Eq. 5) are discussed in the Appendix S9.

2.2 Reference crop evaporation

Adopting the characteristics of a hypothetical reference crop (height = 0.12 m, surface resistance = 70 s m^{-1}, and albedo = 0.23; ASCE Standardization of Reference Evapotranspiration Task Committee, 2000; Allen et al., 1998, p. 15), the Penman-Monteith equation (Eq. 5 becomes Eq. 8), which is known as the FAO-56 Reference Crop or the Standardized Reference Evapotranspiration Equation, Short (ASCE, 2005, Table 1), as follows:

$$\mathrm{ET_{RC}} = \frac{0.408\Delta\left(R_n - G\right) + \gamma \frac{900}{T_a+273} u_2 \left(v_a^* - v_a\right)}{\Delta + \gamma\left(1 + 0.34u_2\right)} \quad (8)$$

where $\mathrm{ET_{RC}}$ is the daily reference crop evapotranspiration (mm day^{-1}), T_a is the mean daily air temperature (°C) at 2 m, and u_2 is the average daily wind speed (m s^{-1}) at 2 m. Other symbols are as defined previously. (A detailed explanation of the theory of reference crop evaporation is presented by McVicar et al., 2005, Sect. 2.) It should be noted that a second reference crop evapotranspiration equation has been developed for 0.5 m tall crop (ASCE, 2005, Table 1). Further details are included in Appendix S5.

The time-step recommended by Allen et al. (1998, Chapt. 4) for analysis using Eq. (8) is one day. Equations for other time-steps may be found in the same reference.

A detailed discussion of the variables is given in Appendix S5. G is a function of successive daily temperatures and, therefore, $\mathrm{ET_{PM}}$ and $\mathrm{ET_{RC}}$ are sensitive to G when there is a large difference between successive daily temperatures. An algorithm for estimating G is presented in Appendix S5. It should be noted that the Penman-Monteith

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equation assumes that the actual evaporation does not affect the overpassing air (Wang et al., 2001).

Other methods for estimating reference crop evaporation

There are other potential evaporation equations for estimating reference crop evaporation, e.g. FAO-24 Blaney and Criddle (Allen and Pruitt, 1986), Turc (1961), Hargreaves-Samani (Hargreaves and Samani, 1985), and the modified Hargreaves approach (Droogers and Allen, 2002). These are included in Appendix S9.

2.3 Pan evaporation

Evaporation data from a Class-A pan, when combined with an appropriate pan coefficient or with an adjustment for the energy exchange through the sides and bottom of the tank, can be considered to be open-water evaporation. Pan data can be used to estimate actual evaporation for situations that require free water evaporation as follows:

$$E_{\mathrm{fw},j} = K_j E_{\mathrm{Pan},j} \tag{9}$$

where $E_{\mathrm{fw},j}$ is an estimate of monthly (or daily) open-surface water evaporation (mm/unit time), j is the specific month (or day), K_j is the average monthly (or daily) Class-A pan coefficient, and $E_{\mathrm{Pan},j}$ is the monthly (or daily) observed Class-A pan value (mm/unit time). Usually, pan coefficients are estimated by comparing observed pan evaporations with estimated or measured open-surface water estimates, although Kohler et al. (1955) and Allen et al. (1998, p. 86) proposed empirically derived relationships. These are described in Appendix S16. Published pan coefficients are available for a range of regions and countries. Some of these are reported also in Appendix S16 and associated tables. In addition, monthly Class-A pan coefficients are provided for 68 locations across Australia (Appendix S16 and Table S6). In China, micro-pans (200 mm diameter, 100 mm high that are filled to 20 or 30 mm) are used to measure

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pan evaporation. Based on an analytical analysis of the pan energetics (McVicar et al., 2007b, p. 209), the pan coefficients for a Chinese micro-pan are lower than Class-A pan coefficients but with a seasonal range being similar to those of a Class-A pan.

Masoner et al. (2008) compared the evaporation rate from a floating evaporation pan (which estimated open-surface water evaporation – see Keijman and Koopmans, 1973; Ham and DeSutter, 1999) with the rate from a land-based Class-A pan. They concluded that the floating pan to land pan ratios were similar to Class-A pan coefficients used in the United States.

The disaggregation of an annual actual or potential evaporation estimate into monthly or especially daily values is not straightforward, assuming there is no concurrent at-site climate data which could be used to gain insight into how the annual value should be partitioned. One approach is to use monthly pan coefficients if available, as noted above. Another approach, that is available to Australian analysts, is to adopt average monthly values of point potential evapotranspiration for the given location (maps for each month are provided in Wang et al., 2001) and pro rata the values to sum to the annual values of E_{fw}. This suggestion is based on the recent analysis by Kirono et al. (2009, Fig. 3) who found that, for 28 locations around Australia, Morton's potential evapotranspiration $\mathrm{ET}_{\mathrm{Pot}}$ (see Sect. 2.5.2) correlated satisfactorily ($R^2 = 0.81$) with monthly Class-A pan evaporation values, although the Morton values over-estimated the pan values by approximately 11 %. Further discussion is provided in Sect. 3.1.3.

The PenPan model

There have been several variations of the Penman equation (Eq. 4) to model the evaporation from a Class-A evaporation pan. Linacre (1994) developed a physical model which he called the Penpan formula or equation. Rotstayn et al. (2006) coupled the radiative component of Linacre (1994) and the aerodynamic component of Thom et al. (1981) to develop the PenPan model (note the two capital Ps to differentiate it from Linacre's, 1994, contribution). Based on the PenPan model, Roderick et al. (2007, Fig. 1) and Johnson and Sharma (2010, Fig. 1) demonstrate separately that the model

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can successfully estimate monthly and annual Class-A pan evaporation at sites across Australia.

Following Rotstayn et al. (2006, Eq. 2) the PenPan equation is defined as:

$$E_{\mathrm{PenPan}} = \frac{\Delta}{\Delta + a_{\mathrm{p}}\gamma}\frac{R_{\mathrm{NPan}}}{\lambda} + \frac{a_{\mathrm{p}}\gamma}{\Delta + a_{\mathrm{p}}\gamma} f_{\mathrm{Pan}}(u)\left(v_{\mathrm{a}}^{*} - v_{\mathrm{a}}\right) \tag{10}$$

where E_{PenPan} is the modelled Class-A (unscreened) pan evaporation (mm day^{-1}), R_{NPan} is the net daily radiation at the pan (MJ m^{-2} day^{-1}), Δ is the slope of the vapour pressure curve (kPa °C^{-1}) at air temperature, γ is psychrometric constant (kPa °C^{-1}), and λ is the latent heat of vaporization (MJ kg^{-1}), a_{p} is a constant adopted as 2.4 (Rotstayn et al., 2006, p. 2), $v_{\mathrm{a}}^{*} - v_{\mathrm{a}}$ is vapour pressure deficit (kPa), and $f_{\mathrm{Pan}}(u)$ is defined as (Thom et al., 1981, Eq. 34):

$$f_{\mathrm{Pan}}(u) = 1.202 + 1.621 u_2 \tag{11}$$

where u_2 is the average daily wind speed at 2 m height (m s^{-1}). Details to estimate R_{NPan} and results of the application of the model to 68 Australian sites are given in Appendix S6.

2.4 Open-surface water evaporation

In this paper the terms open-water evaporation and free-water evaporation are used interchangeably and imply that water available to the evaporation surface is unlimited and that the heat and vapour fluxes have no impact on the over-passing air (Dingman, 1992, p. 276). We discuss two approaches to estimate open-water evaporation: Penman's combination equation and an aerodynamic approach.

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2.4.1 Penman equation

The Penman equation (Penman, 1948, Eq. 16) is widely and successfully used for estimating open-water evaporation as:

$$E_{\mathrm{PenOW}} = \frac{\Delta}{\Delta + \gamma}\frac{R_{\mathrm{nw}}}{\lambda} + \frac{\gamma}{\Delta + \gamma}E_{\mathrm{a}} \tag{12}$$

where E_{PenOW} is the daily open-surface water evaporation (mm day^{-1}), R_{nw} is the net daily radiation at the water surface (MJ m^{-2} day^{-1}), and other terms have been previously defined. In estimating the net radiation at the water surface, the albedo value for water should be used (Table S3). Details of the Penman calculations are presented in Appendix S4. Appendix S3 lists the equations required to compute net radiation with or without incoming solar radiation measurements. We note that of the 20 methods reviewed by Irmak et al. (2011) the method described in Appendix S3 (based on Allen et al., 1998, pp. 41 to 55) to estimate R_{nw} was one of the better performing procedures.

The first term in Eq. (12) is the radiative component and the second term is the aerodynamic component. To estimate R_{nw}, the incoming solar radiation (R_{s}), measured at automatic weather stations or estimated from extraterrestrial radiation, is reduced by estimates of shortwave reflection, using the albedo for water, and net outgoing longwave radiation. E_{a} is known as the aerodynamic equation (Kohler and Parmele, 1967, p. 998) and represents the evaporative component due to turbulent transport of water vapour by an eddy diffusion process (Penman, 1948, Eq. 1) and is defined as:

$$E_{\mathrm{a}} = f(u)\left(v_{\mathrm{a}}^{*} - v_{\mathrm{a}}\right) \tag{13}$$

where $f(u)$ is a wind function typically of the form $f(u) = a + bu$, and $(v_{\mathrm{a}}^{*} - v_{\mathrm{a}})$ is the vapour pressure deficit (kPa).

There have been many studies dealing with Penman's wind function including Penman's (1948 and 1956) analyses (see Penman, 1956, Eqs. 8a and 8b, for a comparison of the two equations), Stigter (1980, pp. 322, 323), Fleming et al. (1989, Sect. 8.4),

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Linacre (1993, Appendix 1), Cohen et al. (2002, Sect. 4) and Valiantzas (2006). Based on Valiantzas's (2006, p. 695) summary of these studies, we recommend that the Penman (1956, Eq. 8b) wind function be adopted as the standard for evaporation from open water with $a = 1.313$ and $b = 1.381$ (wind speed is a daily average value in $\mathrm{m\,s^{-1}}$ and the vapour deficit in kPa). Typically, the wind function assumes wind speed is measured at 2 m above the ground surface but if not it should be adjusted using Eq. (S4.4). More details about alternative wind functions are provided in Appendix S4. It is noted here that because the wind function coefficients were empirically derived the Penman equation for a specific application is an empirical one.

In the Penman equation, it is assumed there is no change in heat storage nor heat exchange with the ground, and no advected energy and, hence, the actual evaporation does not affect the overpassing air (Dingman, 1992, p. 286). Data required to use the equation includes solar radiation, sunshine hours or cloudiness, wind speed, air temperature, and relative humidity (or dew point temperature). Although Penman (1948) carried out his computations of evaporation based on 6-day and monthly time-steps, most analysts have adopted a monthly time-step (e.g. Weeks, 1982; Fleming et al., 1989, Sect. 8.4; Chiew et al., 1995; Cohen et al., 2002; Harmsen et al., 2003) although several have used a daily or shorter time-step (e.g. Chiew et al., 1995; Sumner and Jacobs, 2005).

van Bavel (1966) amended the original Penman 1948 equation to take into account boundary layer resistance. The modified equation is considered in Appendix S4.

Linacre (1993, p. 239) discusses potential errors in the Penman equation and the accuracy of the estimates, and reports that lake evaporation estimates are much more sensitive to errors in net radiation and humidity than to errors in air temperature and wind.

2.4.2 Aerodynamic formula

The aerodynamic method is based on the Dalton-type approach (Dingman, 1992, Sect. 7.3.2), in which evaporation is the product of a wind function and the vapour

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pressure deficit between the evaporating surface and the overlying atmosphere. Following a review of 19 studies, McJannet et al. (2012, Eq. 11) proposed the following relationship to estimate open surface water evaporation.

$$E_{\text{Larea}} = (2.36 + 1.67u_2)\,A^{-0.05}\,(v_w^* - v_a) \quad (14)$$

where E_{Larea} is an estimate of open-surface water evaporation (mm day^{-1}) as a function of evaporating area, A, (m^2), u_2 is the wind speed (m s^{-1}) over land at 2 m height, v_w^* is the saturated vapour pressure (kPa) at the *water surface*, and v_a is the vapour pressure (kPa) at air temperature.

2.5 Actual evaporation (from catchments)

2.5.1 The Complementary Relationship

Figure 1 illustrates conceptually the Complementary Relationship (CR) of Bouchet (1963) which is the basis for estimating actual and potential evapotranspiration by the Morton (1983a) models (Complementary Relationship Areal Evapotranspiration – CRAE, Complementary Relationship Wet-surface Evaporation – CRWE and Complementary Relationship Lake Evaporation – CRLE) and by the Advection-Aridity (AA) model of Brusaert and Strickler (1979) with modifications by Hobbins et al. (2001a, b).

The Complementary Relationship (Morton, 1983a, Eq. 8) is expressed as:

$$\text{ET}_{\text{Act}} = 2\text{ET}_{\text{Wet}} - \text{ET}_{\text{Pot}} \quad (15)$$

ET_{Act} is the actual areal or regional evapotranspiration (mm per unit time) from an area large enough that the heat and vapour fluxes are controlled by the evaporating power of the lower atmosphere and unaffected by upwind transitions. ET_{Wet} is the potential evapotranspiration that would occur under steady state meteorological conditions in which the soil-plant surfaces are saturated and there is an abundant water supply, in other words, the wet-environment evapotranspiration (mm per unit time). According to

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Morton (1983a, p. 16) ET_{Wet} is equivalent to the conventional definition of potential evapotranspiration. ET_{Pot} is the potential evapotranspiration (mm per unit time) for an area so small that the heat and water vapour fluxes have no effect on the overpassing air, in other words, evaporation that would occur under the prevailing atmospheric conditions if only the available energy were the limiting factor.

In his 1983a paper, Morton argues that the CR cannot be verified directly, but based on a water balance study of four rivers in Malawi and another in Puerto Rico, he argued that the concept is plausible (Morton, 1983a, Figs. 7–9). Based on 192 observations in 25 catchments of actual and potential annual evapotranspiration, Hobbins and Ramírez (2004) and Ramírez et al. (2005) present independent evidence based on pan evaporation data and regional ET_{Act} in the US that the Complementary Relationship is at least approximately true. Using a mesoscale model over an irrigation area in south-eastern Turkey, Ozdogan et al. (2006) concluded that their results lend credibility to the CR hypothesis. However, research is underway into understanding whether the constant of proportionality ("2" in Eq. 15) varies and, if so, what is the nature of the asymmetry in the relationship (Ramírez et al., 2005; Szilagyi, 2007; Szilagyi and Jozsa, 2008). Some other references of relevance include Hobbins et al. (2001a), Yang et al. (2006), Kahler and Brutsaert (2006), Lhomme and Guilioni (2006), Yu et al. (2009), and Han et al. (2011).

2.5.2 Morton's models

F.I. Morton was at the forefront of evaporation analyses from about 1965 culminating in the mid-80s with the publication of the Program WREVAP (Morton et al., 1985). WREVAP, which is summarised in Table 2, combines three models namely CRAE (Morton, 1983a), CRWE (Morton, 1983b) and CRLE (Morton, 1986), typically at a monthly timestep. Details of the models are discussed briefly in this section, in Sects. 3.1.2 and 3.2, and in detail in Appendix S7.

Nash (1989, Abstract) concluded that Morton's analysis based mainly on the Complementary Relationship provides a valuable extension to Penman in that it allows one

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to estimate actual evapotranspiration under a limiting water supply. As air passes from a land environment to a lake environment it is modified and the complementary relationship takes this into account.

CRAE model

The CRAE model estimates the three components: potential evapotranspiration, wet-environment areal evapotranspiration and actual areal evapotranspiration. All are based on the Morton methodology.

Estimating potential evapotranspiration (ET_{Pot} in Fig. 1)

Because Morton's (1983a, p. 15) model does not require wind data, it has been used extensively in Australia (where historical wind data were unavailable until recently; see McVicar et al., 2008) to compute time series estimates of historical potential evaporation. Morton's approach is to solve the following energy-balance and vapour transfer equations, respectively for potential evaporation at the equilibrium temperature, which is the temperature of the evaporating surface:

$$ET_{Pot}^{MO} = \frac{1}{\lambda}\left\{R_n - \left[\gamma p f_v + 4\varepsilon_s\sigma\left(T_e + 273\right)^3\right]\left(T_e - T_a\right)\right\} \quad (16)$$

$$ET_{Pot}^{MO} = \frac{1}{\lambda}\left\{f_v\left(v_e^* - v_D^*\right)\right\} \quad (17)$$

where ET_{Pot}^{MO} is Morton's estimate of potential evaporation (mm day^{-1}), R_n is net radiation for soil-plant surfaces at air temperature (W m^{-2}), γ is the psychrometric constant (mbar °C^{-1}), p is the atmospheric pressure (mbar), f_v is the vapour transfer coefficient (W m^{-2} mbar^{-1}), ε_s is the surface emissivity, σ is the Stefan-Boltzmann constant (W m^{-2} K^{-4}), T_e and T_a are the equilibrium temperature (°C) and air temperature (°C), respectively, v_e^* is saturation vapour pressure (mbar) at T_e, v_D^* is the saturation vapour pressure (mbar) at dew point temperature, and λ is the latent heat of vaporisation

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(W day kg^{-1}). Solving for ET_{Pot} and T_e is an iterative process and guidelines are given in Appendix S7. A worked example is provided in Appendix S21.

Estimating wet-environment areal evapotranspiration (*ET$_{Wet}$ *in Fig. 1)

Morton (1983b, p. 79) notes that the wet-environment areal evapotranspiration is the same as the conventional definition of potential evapotranspiration. To estimate the wet-environment areal evapotranspiration, Morton (1983a, Eq. 14) added a term (b_1) to the Priestley-Taylor equation (discussed in Sect. 2.1.3) to account for atmospheric advection as follows:

$$ET_{Wet}^{MO} = \frac{1}{\lambda}\left\{ b_1 + b_2 \frac{R_{ne}}{\left(1 + \frac{\gamma p}{\Delta_e}\right)} \right\} \tag{18}$$

where ET_{Wet}^{MO} is the wet-environment areal evapotranspiration (mm day^{-1}), R_{ne} is the net radiation (W m^{-2}) for the soil-plant surface at the equilibrium temperature T_e (°C), γ is the psychrometric constant (mbar °C^{-1}), p is atmospheric pressure (mbar), Δ_e is slope of the saturation vapour pressure curve (mbar °C^{-1}) at T_e, b_1 (W m^{-2}) and b_2 are the empirical coefficients, and the other symbols are as defined previously. Details to estimate R_{ne} are given in Appendix S7.

Estimating (actual) areal evapotranspiration (*ET$_{Act}$ *in Fig. 1)

Morton (1983a) formulated the CRAE model to estimate actual areal evapotranspiration (ET_{Act}^{MO}) (mm day^{-1}) from the Complementary Relationship (Eq. 15) as follows:

$$ET_{Act}^{MO} = 2ET_{Wet}^{MO} - ET_{Pot}^{MO} \tag{19}$$

ET_{Pot}^{MO} and ET_{Wet}^{MO} are estimated from Eqs. (16), (17), and (18), respectively.

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In the Morton (1983a) paper (Fig. 13), Morton compared estimates of actual areal evapotranspiration with water-budget estimates for 143 river basins world-wide and found the monthly estimates to be realistic. Others have assessed various parts of the CRAE model. Based on a study of 120 minimally impacted basins in the US, Hobbins et al. (2001a, p. 1378) found that the CRAE model overestimated annual evapotranspiration by only 2.5 % of the mean annual precipitation with 90 % of values being within 5 % of the water balance closure estimate of actual evapotranspiration. Szilagyi (2001), inter alia, checked how well WREVAP (incorporating the CRAE program) estimated values of incident global radiation at 210 sites and estimates of pan evaporation at 19 stations with measured values. The respective correlations were 0.79 (Fig. 3 of Szilagyi, 2001) and 0.87 (Fig. 4 of Szilagyi, 2001).

For application of the CRAE model accurate estimates of air temperature and relative humidity are required from a representative land-based location (Morton, 1986, p. 378). For CRAE, Morton (1983a, p. 28) imposed a limit on the minimum time-step for analysis and advocated that the minimum be five days.

CRWE model

In Morton's (1983a) paper, he formulated and documented the CRAE model for land surfaces. In a second paper, Morton (1983b) converts CRAE to a complementary relationship lake evaporation which he designated as CRLE. However, in 1986 Morton (1986) introduced the complementary relationship wet-surface evaporation known as the CRWE model to estimate "lake-size wet surface evaporation" (Morton, 1986, p. 371), in other words, evaporation from shallow lakes. The evaporation from a shallow lake differs from the wet-environment areal evapotranspiration because the radiation absorption and vapour pressure characteristics between water and land surfaces are different (Morton, 1983b, p. 80) as documented in Table 2. It should also be noted that, for a lake, potential evaporation and actual evaporation will be equal but, for a land surface, actual evaporation will be less than potential evaporation, except when the surface is saturated (Morton, 1986, p. 81). Normally, land-based meteorological data

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would be used (Morton, 1983a, p. 70) but data measured over water has only a "... relatively minor effect ..." on the estimate of lake evaporation (Morton, 1983b, p. 96).

In the 1983b paper, Morton (1983b, Eq. 11) introduced an equation (Eq. 23 herein) to deal with estimating evaporation from small lakes, farm dams and ponds.

CRLE model

In the CRLE (and the CRWE) model, a lake is defined as a water body so wide that the effect of upwind advection is negligible. In the Morton context, a deep lake is considered shallow if one is interested only in annual or mean annual evaporation (Morton, 1983b, p. 84) and the CRWE formulation would be used.

Morton's (1983b, Sect. 3) paper provides, inter alia, a routing technique which takes into account the effect of depth, salinity and seasonal heat changes on monthly lake evaporation. This is only approximate as seasonal heat changes in a lake should be based on the vertical temperature profiles which rarely will be available. In 1986, Morton changed the form of the routing algorithm outlined in Morton (1983b, Sect. 3) to a classical linear storage routing model (Morton, 1986, p. 376). This is the one we have adopted in the Fortran 90 listing of WREVAP (Appendix S20) and in the WREVAP worked example (Appendix S21).

Morton (1979, 1983b) validated his approach for estimating lake evaporation against water budget estimates for ten major lakes in North America and East Africa. The average absolute percentage deviation between the model of lake evaporation and water budget estimates was 3.7 % of the water budget estimates (Morton, 1979, p. 72).

Morton (1986, p. 378) notes that, because the Complementary Relationship takes into account the differences in surrounding, for the CRLE model it matters little where the meteorological measurements are made in relation to the lake; they can be land-based or from a floating raft.

Because routing of solar and water-borne energy is incorporated in the CRLE model, a monthly time-step is adopted (Morton, 1983b, Sect. 9). Land-based meteorological data would normally be used (Morton, 1983b, p. 82) but as noted above data measured over water has only a minor effect on the estimate of lake evaporation (Morton, 1983b, p. 96; Morton, 1986, p. 378). Details of the application of Morton's procedures for estimating evaporation from a shallow lake, farm dam or deep lake are discussed in Appendix S7.

A worked example applying Program WREVAP using a monthly time-step is found in Appendix S21.

2.5.3 Advection-aridity and like models

Based on the Complementary Relationship (Eq. 15), Brutsaert and Strickler (1979, p. 445) proposed the original Advection-Aridity (AA) model in which they adopted the Penman equation (Eq. 4) for the potential evaporation (ET_{Pot}) and the Priestley-Taylor equation (Eq. 6) for the wet-environment evaporation (ET_{Wet}) to estimate actual evaporation as follows:

$$E_{Act}^{BS} = (2\alpha_{PT} - 1)\frac{\Delta}{\Delta + \gamma}\frac{R_n}{\lambda} - \frac{\gamma}{\Delta + \gamma} f(u_2)\left(v_a^* - v_a\right) \quad (20)$$

where E_{Act}^{BS} is the actual evaporation estimated by the Brutsaert and Strickler equation (mm day^{-1}), α_{PT} is the Priestley-Taylor coefficient, and the other symbols are as defined previously. In their analysis Brutsaert and Strickler (1979, Abstract) adopted a daily time-step.

In a study of 120 minimally impacted basins in the United States, Hobbins et al. (2001a, Table 2) found that the Brutsaert and Strickler (1979) model underestimated actual annual evapotranspiration by 7.9 % of mean annual precipitation, and for the same basins, Morton's (1983a) CRAE model overestimated actual annual evapotranspiration by only 2.4 % of mean annual precipitation. Several modifications to the original AA model have been put forward. Hobbins et al. (2001b) reparameterized the wind function $f(u_2)$ on a monthly regional basis and recalibrated the Priestley-Taylor coefficient yielding small differences between computed evapotranspiration and water

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balance estimates. However, the regional nature of the wind function restricts the recalibrated model to the conterminous United States.

Alternatives to the Advection-Aridity model of Brutsaert and Strickler (1979) are the approach by Szilagyi (2007) amended by Szilagyi and Jozsa (2008), and the Granger model (Granger, 1989b; Granger and Gray, 1989), which is not based on the Complementary Relationship, and the Han et al. (2011) modification of the Granger model. Details are presented in Appendix S8.

3 Practical topics in estimating evaporation

To address the practical issue of estimating evaporation one needs to keep in mind the setting of the evaporating surface along with the availability of meteorological data. The setting is characterised by several features: the meteorological conditions in which the evaporation is taking place, the water available for evaporation, the energy stored within the evaporating body, the advected energy due to water inputs and outputs from the evaporating water body, and the atmospheric advected energy.

In this paper, water availability refers to the water that is available at the evaporating surface. This will not be limiting for lakes, yet will likely be limiting under certain irrigation practices and, certainly at times, will be limiting for catchments in arid, seasonal tropical and temperate zones. For a global assessment of water-limited landscapes at annual, seasonal and monthly time-steps see McVicar et al. (2012; Fig. 1 and associated material). Stored energy in deep bodies of water, where thermal or salinity stratification can occur, may affect evaporation rates and needs to be addressed as does the energy in water inputs to and outputs from the water body. How atmospheric advected energy is dealt with in an analysis depends on the size of the evaporating body and the procedure adopted to estimate evaporation. In this context we need to heed the advice of McVicar et al. (2007b, p. 197) that a regional surface evaporating at its potential rate would modify the atmospheric conditions and, therefore, change the rate of local potential evaporation. For a large lake or a large irrigation area dry incoming

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wind will affect the upwind fringe of the area but the bulk of the area will experience a moisture-laden environment. On the other hand, for a small lake or farm dam, a small irrigation area or an irrigation canal in a dry region, the associated atmosphere will be minimally affected by the water body and the prevailing upwind atmosphere will be the driving influence on the evaporation rate.

In the following discussion, we assume that: (i) at-site daily meteorological data from an automatic weather station (AWS) are available; or (ii) meteorological data measured manually at the site and at an appropriate time interval are available; or (iii) at-site daily pan evaporation data are available. At some AWSs, hard-wired Penman or Penman-Monteith evaporation estimates are also available. Methods to estimate evaporation where meteorological data are not available are discussed in Sect. 4.3.

When incorporating estimates of lake evaporation into a water balance analysis of a reservoir and its related catchment, it is important to note that double counting will occur if the inflows to the reservoir are based on the catchment area including the inundated area and then an adjustment is made to the water balance by adding rainfall to and subtracting lake evaporation from the inundated area. The correct adjustment is the difference between evaporation prior to inundation and lake evaporation (see McMahon and Adeloye, 2005, p. 97 for a fuller explanation of this potential error).

3.1 Deep lakes

This paper does not address the measurement of evaporation from lakes but rather the estimation of lake evaporation by modelling. A helpful review article on the measurement and the calculation of lake evaporation is by Finch and Calver (2008).

In dealing with deep lakes (including constructed storages, reservoirs and large voids), three issues need to be addressed. First, the heat storage of the water body affects the surface energy flux and, because the depth of mixing varies in space and time and is rarely known, it is difficult to estimate changes at a short time-step; typically, a monthly time-step is adopted. Second, the effects of water advected energy needs to be considered. If the inflows to a lake are equivalent to a large depth of the lake

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area and their average temperatures are significantly different, advected energy needs to be considered (Morton, 1979, p. 75). Third, increased salinity reduces evaporation and, therefore, changes in lake salinity need to be addressed. Next, we explore three procedures for estimating evaporation from deep lakes.

3.1.1 Penman model

To estimate evaporation from a deep lake, the Penman estimate of evaporation, E_{PenOW}, (Eq. 12) is a starting point. Water advected energy and heat storage are accounted for by the following equation recommended by Kohler and Parmele (1967, Eq. 12) and reported by Dingman (1992, Eq. 7–37) as:

$$E_{DL} = E_{PenOW} + \alpha_{KP}\left(A_w - \frac{\Delta Q}{\Delta t}\right) \quad (21)$$

where E_{DL} is the evaporation from the deep lake (mm day^{-1}), E_{PenOW} is the Penman or open-surface water evaporation (mm day^{-1}), α_{KP} is the proportion of the net addition of energy from water advection and storage used in evaporation during Δt, A_w is the net water advected energy during Δt (mm day^{-1}), and $\frac{\Delta Q}{\Delta t}$ is the change in stored energy expressed as a water depth equivalent (mm day^{-1}). The latter three terms are complex and are set out in Appendix S10 along with details of the procedure.

Vardavas and Fountoulakis (1996, Fig. 4), using the Penman model, estimated the monthly lake evaporation for four reservoirs in Australia and found the predictions agreed satisfactorily with mean monthly evaporation measurements. Change in heat storage is based on the monthly surface water temperatures. Thus:

$$E_{DL} = \frac{\Delta}{\Delta + \gamma}\left(\frac{R_n + \Delta H}{\lambda}\right) + \frac{\gamma}{\Delta + \gamma}E_a \quad (22)$$

where E_{DL} is the evaporation from the deep lake (mm day^{-1}), R_n is the net radiation at the water surface (MJ m^{-2} day^{-1}), E_a is the evaporation component (mm day^{-1}) due to

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wind, Δ is the slope of the vapour pressure curve (kPa °C^{-1}) at air temperature, γ is the psychrometric constant (kPa °C^{-1}), λ is the latent heat of vaporization (MJ kg^{-1}), and ΔH is the change in heat storage (MJ m^{-2} day^{-1}). We detail the Vardavas and Fountoulakis (1996) method in Appendix S10.

3.1.2 Morton evaporation

In Morton's WREVAP program, monthly evaporation from deep and shallow lakes can be estimated. As noted in Sect. 2.5.2, for annual evaporation estimates, there is no difference in magnitude between deep and shallow lake evaporation (see also Sacks et al., 1994, p. 331). In Morton's procedure, seasonal heat changes in deep lakes are incorporated through linear routing. Details are presented in Appendix S7. The data for Morton's WREVAP program are mean monthly air temperature, mean dew point temperature (or mean monthly relative humidity) and monthly sunshine hours as well as latitude, elevation and mean annual precipitation at the site. The broad computational steps are set out in Appendix S7 and details can be found in Appendix C of Morton (1983a). A Fortran 90 listing of a slightly modified version of the Morton WREVAP program is provided in Appendix S20 and a worked example is available in Appendix S21.

The Complementary Relationship Lake Evaporation of Morton (1983b, 1986) and Morton et al. (1985) may be used to estimate lake evaporation directly. Comparing CRLE lake evaporation estimates with water budgets for 17 lakes world-wide, Morton (1986, p. 385) found the annual estimates to be within 7 %. In a lake study in Brazil, dos Reis and Dias (1998, Abstract) found the CRLE model estimated lake evaporation to within 8 % of an estimate by the Bowen Ratio energy budget method. Furthermore, Jones et al. (2001) using a water balance incorporating CRLE evaporation for three deep volcanic lakes in western Victoria, Australia, satisfactorily modelled water levels in the closed lakes system over a period exceeding 100 yr.

Some further comments on Morton's CRLE model are given in Appendix S7.

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3.1.3 Pan evaporation for deep lakes

Dingman (1992, Sect. 7.3.6) implies that, through an application to Lake Hefner (US), Class-A pan evaporation data, appropriately adjusted for energy flux through the sides and the base of the pan, can be used to estimate daily evaporation from a deep lake. Based on the Lake Hefner study, Kohler notes that "annual lake evaporation can probably be estimated within 10 % to 15 % (on the average) by applying an annual coefficient to pan evaporation, provided lake depth and climatic regime are taken into account in selecting the coefficient" (Kohler, 1954).

In Australia, there was a detailed study of lake evaporation in the 1970s that resulted in two technical reports by Hoy and Stephens (1977, 1979). In these reports mean monthly pan coefficients were estimated for seven reservoirs across Australia and annual coefficients were provided for a further eight reservoirs. Values are listed in the Tables S11 and S12.

Garrett and Hoy (1978, Table III) modelled annual pan coefficients based on a simple numerical lake model incorporating energy and vapour fluxes. The results show that for the seven reservoirs examined, the annual pan coefficients change little with lake depth.

3.2 Shallow lakes, small lakes and farm dams

For large shallow lakes, less than a meter or so in depth, where advected energy and changes in seasonal stored energy can be ignored, the Penman equation with the 1956 wind function or Morton's CRLE model (Morton, 1983a,b) may be used to estimate lake evaporation. The upwind transition from the land environment to the large lake is also ignored (Morton, 1983b).

Stewart and Rouse (1976) recommended the Priestley-Taylor model for estimating daily evaporation from shallow lakes. Based on summer evaporation of a small lake in Ontario, Canada, the monthly lake evaporation was estimated to within ±10 % (Stewart and Rouse, 1976, p. 628). Galleo-Elvira et al. (2010) found that incorporating

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a seasonal advection component and heat storage into the Priestley-Taylor equation (Eq. 6) provided accurate estimates of monthly evaporation for a 0.24 ha water reservoir (maximum depth of 5 m) in semi-arid Southern Spain. Analytical details are given in Galleo-Elvira et al. (2010).

For shallow lakes, say less than 10 m, in which heat energy should be considered, Finch (2001) adopted the Keijman (1974) and de Bruin (1982) equilibrium temperature approach which he applied to a small reservoir at Kempton Park, UK. The procedure adopted by Finch (2001) is described in detail in Appendix S11.

Finch and Gash (2002) provide a finite difference approach to estimating shallow lake evaporation. They argue the predicted evaporation is in excellent agreement with measurements (Kempton Park, UK) and closer than Finch's (2001) equilibrium temperature method.

Using a similar approach to Finch (2001) but based on Penman-Monteith rather than Penman, McJannet et al. (2008) estimated evaporation for a range of water bodies (irrigation channel, shallow and deep lakes) explicitly incorporating the equilibrium temperature. The method is described in detail in Appendix S11 and a worked example is available in Appendix S19.

McJannet et al. (2012) developed a generalised wind function that included lake area (Eq. 14) to be incorporated in the aerodynamic approach (Eq. 13). The equation is of limited use as the equilibrium (surface water) temperature needs to be estimated.

For small lakes and farm (and aesthetic) dams the increased evaporation at the upwind transition from a land environment may need to be addressed. Morton (1983b, Eq. 11) recommends the following equation be used to adjust lake evaporation for the upwind advection effects:

$$E_{\mathrm{SLx}} = E_{\mathrm{L}} + (\mathrm{ET_P} - E_{\mathrm{L}})\frac{\ln\left(1+\frac{x}{C}\right)}{\frac{x}{C}} \quad (23)$$

where E_{SLx} is the average lake evaporation (mm day^{-1}) for a crosswind width of x m, E_{L} is lake evaporation (mm day^{-1}) large enough to be unaffected by the upwind transition,

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i.e. well downwind of the transition, ET_P is the potential evaporation (mm day^{-1}) of the land environment, and C is a constant equal to 13 m.

Morton (1986, p. 379) recommends that ET_P be estimated as the potential evaporation in the land environment as computed from CRWE and the lake evaporation E_L be computed from CRLE. ET_P could also be estimated using Penman-Monteith (Eq. 5) with appropriate parameters for the upwind landscape and the Penman open-water equation (Eq. 12) could be used to estimate E_L.

3.3 Catchment water balance

The traditional method to estimate annual actual evaporation for an unimpaired catchment is through a simple water balance:

$$\overline{ET}_{Act} = \bar{P} - \bar{Q} - \bar{G}_{DS} - \Delta S \quad (24)$$

where $\overline{ET}_{Act}$ is the mean annual actual catchment evaporation (mm yr^{-1}), $\bar{P}$ is the mean annual catchment precipitation (mm yr^{-1}), $\bar{Q}$ is the mean annual runoff (mm yr^{-1}), $\bar{G}_{DS}$ is the deep seepage (mm yr^{-1}), and ΔS is the change in soil moisture storage over the analysis period (mm yr^{-1}). At an annual time-step, ΔS is assumed zero. Often deep seepage is also assumed to be negligible. Based on an extensive review of the recharge literature in Australia, Petheram et al. (2002) developed several empirical relationships between recharge and precipitation. A more comprehensive and larger Australian data set was analysed by Crosbie et al (2010) who developed relationships between average annual recharge and mean annual rainfall for combinations of soil and vegetation types. A considerably larger global study, but only for semi-arid and arid regions, was conducted by Scanlon et al. (2006) who also developed several generalised relationships relating recharge to mean annual precipitation. These generalised relationships could be used if deep seepage was considered important and relevant data were available.

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An alternative and more direct method is to estimate actual monthly catchment evaporation either by Morton's CRAE model (Sect. 2.5.2) or one of the Advection-Aridity or like-models (discussed in Sect. 2.5.3). An interesting comparison of monthly estimates of catchment evaporation by the Morton and Penman methods was carried out by Doyle (1990) for the Shannon catchment in Ireland. In the Penman approach when water was not freely available, actual evaporation was estimated using a simple Thornthwaite soil moisture model. The study examined the strengths and weaknesses of both approaches and concluded that the Morton approach is a valuable alternative to the empiricism introduced through using the Thornthwaite algorithm to convert potential evaporation to actual evaporation, but Doyle also noted the strong degree of empiricism in accounting for advection in the Morton approach.

A very different approach to estimating mean annual actual evaporation is based on the Budyko formulation (Budyko, 1974), which is a balance between the energy and the water availability in a catchment. Equations that fall into this category include: Schreiber (Schreiber,1904), Ol'dekop (Ol'dekop, 1911), generalised Turc-Pike (Turc, 1954; Pike, 1964; Milly and Dunne, 2002), Budyko (Budyko, 1974), Fu (Fu, 1981; Zhang et al., 2004; Yang et al., 2008), Zhang 2-parameter model (Zhang et al., 2001), and a linear model (Potter and Zhang, 2009) and have the following simple form:

$$\bar{E}_{act} = \bar{P} f(\varnothing) \tag{25}$$

where $\bar{E}_{act}$ is mean annual actual catchment evaporation (mm yr^{-1}), $\bar{P}$ is mean annual catchment precipitation (mm yr^{-1}) and $\varnothing$ is the aridity index defined as $\bar{E}_{pot}/\bar{P}$ where $\bar{E}_{pot}$ is the mean annual catchment potential evapotranspiration (mm yr^{-1}). The available functions $f(\varnothing)$, based on the references above, are listed in Table 3. The Zhang function (Zhang et al., 2001, Eq. 8) allows long term estimates of actual evaporation for grasslands and forests to be estimated. Donohue et al. (2007, 2010b, 2011) and Zhang and Chiew (2011) found that the accuracy of estimates of long term actual evaporation can be improved by incorporating vegetation types and dynamics into the Budyko formulations.

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3.4 Daily and monthly rainfall-runoff modelling

Most rainfall-runoff models at a daily or monthly time-step (e.g. Sacramento, Burnash et al., 1973; Système Hydrologique Européen – SHE, Abbott et al., 1986a, b; AWBM, Boughton, 2004; SIMHYD, Chiew et al., 2002) require as input an estimate of potential evaporation in order to compute actual evaporation. In these models typically:

$$\mathrm{ET_{Act}} = f\left(\mathrm{SM}, \mathrm{ET_{PET}}\right) \tag{26}$$

where $\mathrm{ET_{Act}}$ is the estimated actual daily evaporation (mm day^{-1}), SM is a proxy soil moisture level for the given day (mm), and $\mathrm{ET_{PET}}$ is the daily potential evaporation (mm day^{-1}). In arid catchments and for much of the time in temperate catchments, actual evaporation will be limited by soil moisture availability with potential evaporation becoming more important in wet catchments where soil moisture is not limiting. Generally, one of four approaches has been used to estimate potential evaporation in rainfall-runoff modelling: Penman-Monteith and variations (Beven, 1979; Watson et al., 1999), Priestley-Taylor and variations (Raupach et al., 2001; Zhang et al., 2001), Morton's procedure (Chiew et al., 1993; Siriwardena et al., 2006), and pan evaporation (Zhao, 1992; Lidén and Harlin, 2000; Abulohom et al., 2001; McVicar et al., 2007a; Welsh, 2008; Zhang et al., 2008). These approaches are discussed in detail in Appendix S13.

In detailed water balances, interception and, therefore, interception evaporation are key processes. Two important interception models are the Rutter (Rutter et al., 1971, 1975) and the Gash (Gash, 1979) models. The Rutter model incorporates Penman (1956, Eq. 8b) equation to estimate potential evaporation while the Penman-Monteith equation is the evaporation model used in the Gash model. Details are provided in Appendix S14. Readers are referred to a recent and comprehensive review by Muzylo et al. (2009), who addressed the theoretical basis of 15 interception models including their evaporation components, identified inadequacies and research questions, and noted there were few comparative studies about uncertainty in field measurements and model parameters.

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3.5 Irrigation areas

Internationally, the FAO-56 Reference Crop equation (Eq. 8) which is the Penman-Monteith equation for specific reference conditions, is the accepted procedure to estimate reference crop evaporation (ET_{RC}). It is assumed that both water advected energy and heat storage effects can be ignored (Dingman, 1992, p. 299). Reference crop evapotranspiration is usually different to the actual evapotranspiration of a specific crop under normal growing conditions. To estimate crop evapotranspiration under standard conditions (disease-free, well-fertilised crop, grown in large fields, under optimum soil water conditions and achieving full yield) a crop coefficient (K_c) is applied to ET_{RC}. Values of K_c are a function of crop characteristics and soil moisture conditions. Because of the large number of crops and potential conditions, readers are referred to the details in Allen et al. (1998, Chapters 6 and 7).

The FAO-56 Reference Crop method (Allen et al., 1998, Chapt. 4) (Sect. 2.2) for computing reference crop evaporation is a two-step procedure and, according to Shuttleworth and Wallace (2009), humid conditions are a prerequisite for its applicability (Shuttleworth and Wallace, 2009, p. 1905). In irrigation regions like Australia that are arid and windy, Shuttleworth and Wallace (2009) recommend the FAO-56 method be replaced by a one-step method known as the Matt-Shuttleworth procedure in which the crop coefficients are replaced by their equivalent surface resistances. Some more details are set out in Appendix S5.

For small irrigation areas in dry regions, atmospheric advection may need to be taken into account for the same reason as discussed for a small lake in Sect. 3.2. The significance of this situation, which is known as the "oasis effect", is illustrated in Fig. 2 (adapted from Allen et al., 1998, Fig. 46). As observed in the figure, the effect can be important. For the climate and vegetation conditions examined by Allen et al. (1998) for a 100 m wide area of irrigation in dry surroundings, the crop coefficient of K_c would be increased by a little more than 30 %, and for a 300 m wide area, the increase is

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20 %. However, as cautioned by Allen et al. (1998, p. 202) care needs to be exercised in adopting these sorts of adjustments.

Estimating actual crop evapotranspiration under non-optimum soil-water conditions is not straightforward. Details are set out in Allen et al. (1998, Chapt. 8). Sumner and Jacobs (2005, Sect. 7) found that Penman-Monteith and Priestley-Taylor models could reproduce actual evapotranspiration from a non-irrigated crop but both models required local calibration.

3.6 Evaporation from lakes covered by vegetation

There is an extensive body of literature addressing the question of evaporation from lakes covered by vegetation. Abtew and Obeysekera (1995, Table 1) summarise 19 experiments which, overall, show that the transpiration of macrophytes is greater than open-surface water evaporation. However, most experiments were not in situ experiments. On the other hand, Mohamed et al. (2008, Table 2) lists the results of 11 in situ studies (mainly eddy correlation or Bowen Ratio procedures) in which wetland evaporation is, overall, less than open-surface water. These issues are discussed in Appendix S12 and a comparison is provided (Table S7) of equations to estimate evaporation from lakes covered by vegetation.

3.7 Bare soil evaporation

Numerous writers (e.g. Ritchie, 1972; Boesten and Stroosnijder, 1986; Katul and Parlange, 1992; Kondo and Saigusa, 1992; Yunusa et al. 1994; Daaman and Simmonds, 1996; Qiu et al., 1998; Snyder et al. 2000; Mutziger et al., 2005; Konukcu, 2007) have discussed bare soil evaporation. Most methods require field data in addition to the meteorological data to estimate evaporation from an initially wet surface. Ritchie (1972, p. 1205) proposed a two-stage model following Philip (1957) to estimating bare soil evaporation: Stage-1 evaporation, which is atmosphere-controlled (that is, the soil has adequate moisture so that the moisture can move to the surface at a rate that does

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not impede evaporation), and Stage-2 evaporation, which is soil-moisture controlled. It is noted that McVicar et al. (2012, p. 183) observed that Stage-1 evaporation is more appropriately described as energy-limited evaporation and Stage-2 as water-limited evaporation. Salvucci (1997) developed this approach further. Details are provided in Appendix S15.

3.8 Groundwater evaporation

Luo et al. (2009) noted that a significant amount of groundwater evaporates from irrigated crops and phreatophytes as a result of shallow groundwater tables. After reviewing the literature, they field-tested four widely used groundwater relationships (linear, linear segment, power and exponential) which relate evapotranspiration to the depth to the groundwater table and the maximum evapotranspiration. The authors concluded that so long as appropriate parameters are chosen, the four functions can be used to describe the relationship between evapotranspiration from groundwater and water table depth. Readers are referred to the Luo et al. (2009) paper for details.

3.9 Guidelines in estimating monthly evaporation

This section provides a brief justification of Table 4 which is a succinct summary of the preferred options for estimating monthly evaporation for the set of practical topics discussed in Sect. 3. In the table we have adopted four levels of guidelines: *preferred*, *acceptable*, *not preferred or insufficient field testing*, and *not recommended*. Each assessment in Table 4 is based on the information summarised in the paper and in the supplementary materials along with the authors' personal experiences in applying or reviewing evaporation estimation procedures. Analysts using the table as a basis for choosing a specific procedure should be aware of the inadequacies of the procedures which are discussed in the relevant sections and in the supplementary material. Some comments on several of the assessments follow.

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For deep lakes, the Morton (1986) method is the preferred approach because it has a theoretical background and the evaporation estimates have been widely tested. The Kohler and Parmele (1967) and the Vardavas and Fountoulakis (1996) approaches are acceptable as both take into account heat storage effects. However, testing of these two models has not been as extensive as testing of the Morton (1986) procedure. Pan coefficients are required to apply evaporation pan data to estimating deep lake evaporation. These are available for selected reservoirs (Hoy and Stephens, 1977, 1979), but there appears to be little consistency in their monthly values. For this reason in Table 4, we adopt the "not preferred" guideline for pan coefficients.

For shallow lakes, less than 2 m depth, Penman (1956) is the preferred approach whereas for deeper lakes Morton's (1983a) CRWE model is preferred. Both models are based on theoretical analysis and have been subject to extensive field tests. Based on theoretical analysis, equilibrium temperature methods of Finch (2001) and McJannet et al. (2008) are acceptable along with the finite difference procedure of Finch and Gash (2002). We acknowledge that pan evaporation data are widely used to estimate shallow lake evaporation, but we have adopted the "not preferred" guideline on the basis that reliable local pan coefficients are often not available.

To estimate the actual monthly evaporation component for catchment water balance studies, Morton (1983a) CRAE model is acceptable. It is not a preferred method because the parameters f_Z, b_1 and b_2 were required to be calibrated for the Australian landscape (see Supplement Appendix S7). Both the Brutsaert and Strickler (1979) and Szilagyi and Jozsa (2008) models have theoretical backgrounds and have been tested mainly against Morton (1983a). Both procedures can generate negative values of actual evaporation and are not preferred.

To estimate crop water requirements in humid regions, FAO-56 Reference Crop (Allen et al., 1998) is widely adopted and preferred. However, for specific crops in windy semi-arid regions, the Matt-Shuttleworth model (Shuttleworth and Wallace, 2009) is acceptable. Again, we do not advise the evaporation pan approach because reliable local pan coefficients are not always readily available. The FAO-56 Reference Crop potential

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evapotranspiration is converted to a specific crop water requirement through the application of a crop coefficient.

There are no preferred procedures for lakes with vegetation and bare soil evaporation. The major issue here is that there is little evidence in the literature of adequate testing of the methodologies.

Rainfall-runoff modelling requires potential evaporation as input and the selection of an adequate potential evaporation model is more important in energy-limited catchments, where soil moisture is readily available, than in water-limited catchments. From a literature review, we regard several models as acceptable for this application – Penman-Monteith (Monteith, 1965), Priestly-Taylor (Priestly and Taylor, 1972), Morton (Morton, 1983a), and evaporation pan. These models are acceptable provided they are used as input during calibration of the rainfall-runoff model. To this list we add Penman (Penman, 1948, 1956), although its use in rainfall-runoff modelling has been generally restricted to estimating interception evaporation.

In practice, analysts must take several issues into consideration in the selection of the most appropriate option to estimate monthly actual or potential evaporation. The guidelines presented in Table 4 are predominantly based on the strength of the theoretical basis of the method and the results of testing. These are important characteristics that should influence the selection of an appropriate method. However, amongst other things, analysts must also consider the availability of the input data and the effort required to generate the monthly evaporation estimates. A summary of the data required by each method is given in Table 1 and Sect. 4.3 discusses approaches to estimate evaporation without at-site data. To our knowledge, there are no studies that have compared the relative accuracy of these methods when the data inputs are based on spatial interpolation or modelling. The effort required to generate monthly evaporation estimates varies between the methods available and, in some situations, it may be appropriate for an analyst to adopt a simpler, but less preferred method.

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4 Outstanding issues

Within the context of this paper we have identified six issues that require discussion: (i) hard-wired potential evaporation estimates at AWSs; (ii) estimating evaporation without wind data; (iii) estimating evaporation without at-site data; (iv) dealing with a climate change environment: increasing annual air temperature but decreasing pan evaporation rates; (v) daily meteorological data average over 24 h or day-light hours only; and (vi) finally, uncertainty in evaporation estimates.

4.1 Hard-wired evaporation estimates

Some commercially available AWSs, in addition to providing values of the standard climate variables, output an estimate of Penman evaporation or Penman-Monteith evaporation. For practitioners, this will probably be the data of choice rather than recomputing Penman or Penman-Monteith evaporation estimates from basic principles. However, users need to understand the methodology adopted and check the values of the parameters and functions (e.g. albedo, wind function, r_a and r_s) used in the AWS evaporation computation.

4.2 Estimating potential evaporation without wind data

Many countries do not have access to historical wind data to compute potential evaporation. In rainfall-runoff modelling in which potential evaporation estimates are required, several researchers (Jayasuria et al., 1988, 1989; Chiew and McMahon, 1990) overcame this situation by using Morton's algorithms (Morton, 1983a, b) (Sect. 2.5.2) which do not require wind information. In developing a water balance model for three volcanic crater lakes in western Victoria, Jones et al. (1993) successfully applied Morton's CRLE model (Morton et al., 1985) to estimate actual lake evaporation using air and dew point temperatures and sunshine hours or global radiation. Valiantzas (2006) developed two empirical equations to provide approximate estimates of Penman's open-surface water

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evaporation and reference crop evaporation without wind data. Details are set out in Appendices S4 and S5. Another approach to estimating potential evaporation without wind data is the modified Hargreaves procedure described in Appendix S9.

4.3 Estimating potential evaporation without at-site data

Where at-site meteorological or pan evaporation data are unavailable, it is recommended that evaporation estimates be based on data from a nearby weather station that is considered to have similar climate and surrounding vegetation conditions to the site in question. This would mean that both stations would have similar elevation and would be exposed to similar climatic features.

In many parts of the world an alternative approach is to use outputs from spatial interpolation and from spatial modelling (Sheffield et al., 2006; McVicar et al., 2007b; Vicente-Serrano et al., 2007; Thomas, 2008; Donohue et al., 2010a; Weedon et al., 2011). However, sometimes this cannot be achieved as proximally located meteorological stations do not exist. If seeking an estimate of evaporation for a large area (e.g. a catchment or an administrative region) then using gridded output is required. Details for Australia are provided in Appendix S1.

Errors in lake evaporation estimates introduced by transposed data were studied using an energy budget by Rosenberry et al. (1993) for a lake in Minnesota, United States from 1982 to 1986. Their key conclusions are:

1. Replacing raft-based air temperature or humidity data with those from a land-based near-shore site affected computed estimates of annual evaporation between +3.7 % and −3.6 % (averaged −1.2 %).

2. Neglecting heat transfer from the bottom sediments to the water resulted in an increase in lake evaporation of +7 %.

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3. Substituting lake shortwave solar radiation, air temperature and atmospheric vapour pressure with values from a site 110 km away resulted in errors of +6 % to +8 %.

4.4 Dealing with a climate change environment: increasing annual temperature but decreasing pan evaporation

Based on analysis of regions, across seven countries, with more than 10 pan evaporation stations, Roderick et al. (2009a, Table 1) reported negative trends in pan evaporation measures over the last 30 to 50 yr. Recently, McVicar et al. (2012; Table 5) showed that declining evaporative demand, as measured by pan evaporation rates, was globally widespread. In their review of 55 studies reporting pan evaporation trends, the average trend was 3.19 mm yr^{-2}. Reductions over the past 40 yr have also been observed in Australia (Roderick and Farquhar, 2004; Kirono and Jones, 2007; Jovanovic et al., 2008) and in China (Liu et al., 2004; Cong et al., 2009).

These reductions imply that there has been a decline in evaporative demand as measured by pan evaporation (Petersen et al., 1995) which is in contrast to the increased air temperatures that have been observed during the same period (Hansen et al., 2010). Roderick et al. (2009b, Sect. 2.3) suggest that the decline in evaporative demand is due to increased cloudiness and reduced wind speeds and, for the Indian region, Chattopadhyay and Hulme (1997) suggested that relative humidity was also a factor. After an extensive literature review, Fu et al. (2009) concluded that more investigations are required to understand fully global evaporation trends. McVicar et al. (2012; Table 7) demonstrated that broad generalisations pointing to one variable controlling evaporation trends is not possible. All variables influencing the evaporative process (wind speed, atmospheric humidity, radiation environment and air temperature) need to be taken into account. It is interesting to note that Jung et al. (2010, p. 951) argue that global annual actual evapotranspiration increased, on average, by 7.1 mm yr^{-1} $decade^{-1}$ from 1982 to 1997, after which the increase ceased. They suggested that the switch is due mainly to lower soil moisture in the Southern Hemisphere

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during the past decade. Further understanding of the area-average evaporation and pan evaporation is offered by Shuttleworth et al. (2009), who concluded from their study, that there are two influences on pan evaporation operating at different spatial scales and in opposite directions. The study confirmed that changes in pan evaporation are associated with: (i) large-scale changes in wind speed, with surface radiation having a secondary impact and (ii) the landscape coupling between surface and the atmospheric boundary layer through surface radiation, wind speed and vapour pressure deficit (Shuttleworth et al., 2009, p. 1244). However, the above explanations are further complicated by analyses of Brutsaert and Parlange (1998), Kahler and Brutsaert (2006) and Pettijohn and Salvucci (2009) and summarised by Roderick et al. (2009b). Roderick et al. (2009b) examined a generalised Complementary Relationship incorporating pan evaporation and suggested that in water-limited environments declines in pan evaporation may be interpreted as evidence of increasing terrestrial evaporation if rainfall increases while in energy-limited environments terrestrial evaporation is decreasing.

As pointed out by Roderick et al. (2009b), to apply the reductions in pan evaporation to the terrestrial environment is not straightforward because of the importance of supply and demand of water through rainfall and evaporation and because of the operation of the Complementary Relationship (Sect. 2.5.1) (see also the comment by Brutsaert and Parlange, 1998). Roderick et al. (2009b, Sect. 4) described the issue as follows. "In energy-limited conditions, declining pan evaporation generally implies declining actual evapotranspiration. If precipitation were constant then one would also expect increasing runoff and/or soil moisture. In water-limited conditions, the interpretation is not so straightforward because actual evapotranspiration is then controlled by the supply and not demand. In such circumstances, one has to inspect how the supply (i.e. precipitation) has changed before coming to a conclusion about how actual evapotranspiration and other components of the terrestrial water balance have changed ...". Recently, McVicar et al. (2012; Fig. 1) in a global review of terrestrial wind speed trends mapped the areas that are climatologically water-limited and energy-limited.

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For rainfall-runoff modelling, several researchers (Oudin et al., 2005; Kay and Davies, 2008) have observed that a model calibrated with potential evaporation inputs based on air temperature perform at least as well as with inputs from the more data intensive Penman-Monteith model in predominately water-limited environments. However, when considering climatic changes, the recent evidence is compelling (Roderick et al., 2009a; McVicar et al., 2012) that in a climate-changing environment all relevant and interacting climate variables should be taken into account wherever possible (McVicar et al., 2007b). Using air temperature as the only forcing variable for estimating potential evaporation will lead potentially to an incorrect outcome particularly in energy-limited environments which are important head-waters for many major river systems across the globe (see McVicar et al., 2012 for discussion). In this context it is worth noting that by not considering one of the key variables (radiation, air temperature, relative humidity or wind) in an evaporation equation, it is implicitly assumed that variable is non-trending. This can be a very poor assumption as highlighted in the global wind review by McVicar et al. (2012), following the original wind "stilling" paper by Roderick et al. (2007).

Finally, in the context of a changing climate, Donohue et al. (2010a) compare potential evaporation computed by Penman (Sect. 2.1.1), Priestley-Taylor (Sect. 2.1.3), Morton point (Sect. 2.5.2), Morton areal (Sect. 2.5.2), and Thornthwaite (1948). Donohue et al. (2010a) concluded that the Penman model produced the most reasonable estimation of the dynamics of potential evaporation (Donohue et al., 2010a, p. 196). Their finding echoed several previous papers (e.g. Chen et al., 2005; Garcia et al., 2004; McKenney and Rosenberg, 1993; Shenbin et al., 2006) and is confirmed by an extensive review paper by McVicar et al. (2012). In this context it is noted that estimates of the Morton point potential evapotranspiration by Donohue et al. (2010a) were very high. R. Donohue (personal communication, 2012) advised that "the reason Morton point potential values were so high in Donohue et al. (2010b) was because, in their modelling of net radiation, they explicitly accounted for actual land-cover dynamics. This procedure differs from Morton's (1983) methodology, developed over 25 yr

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ago, when remotely sensed data were not routinely available, and thus Donohue et al. (2010b) is in contradiction to Morton's (1983) methodology."

4.5 Daily (24 h) or day-time (day-light hour)

An issue that arose during this project relates to whether or not daily meteorological data used in evaporation equations should be averaged over a 24-h daily period or averaged during daylight hours when evaporation is mainly taking place. Most authors are silent about this as they are using standard meteorological daily data provided by the relevant agency. Furthermore, most procedures incorporate empirically derived coefficients which were estimated using the standard meteorological data. In view of this, except where we specifically have noted in the text and in the appendices that the input climate data are averaged over day-light hours, in all analyses standard meteorological data should be used. Although the definitions may vary slightly from jurisdiction to jurisdiction, the issue is far too large to be further considered here. This is an important question that needs addressing. Stigter (1980, p. 328) and Van Niel et al. (2011) provide a starting point for such a discussion.

4.6 Uncertainty in evaporation estimates and model performance

In the previous sections we describe several models for estimating actual and potential evaporation. These models vary in complexity and in data requirements. In selecting an appropriate model, analysts should consider the uncertainty in alternative methods.

Winter (1981) provides a useful starting point. He examined the uncertainties in the components of the water balance of lakes. Regarding evaporation, he concluded that closing the surface energy balance was considered the most accurate method – annual estimates $< \pm 10\,\%$. Errors of 15 to 20 % in Dalton-type equations were assessed in terms of the mass transfer coefficient. Errors in monthly Class-A pan data were reported to be up to 30 %. In addition, several studies reported large variations in pan

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to lake coefficients (for error analyses see Hounam, 1973; Ficke, 1972; Ficke et al., 1977).

Nichols et al. (2004) also provide a detailed error analysis based on a semi-arid region in New Mexico, US using a standard error propagation method. The conditions adopted in the uncertainty analysis using a daily time-step included: air temperature ±0.1 %, relative humidity ±3 %, vapour deficit, ±4 %, wind speed ±5 %, net radiation ±15 %, γ ±0.1 %, and Δ ±0.5 % from which the following uncertainties were computed: Penman (1948 equation) ±13 %, Priestley-Taylor ±18 %, and Penman-Monteith ±10 %. McJannet et al. (2008, Table 6.1), in a review of open-surface water evaporation estimates in the Murray-Darling Basin, Australia, assessed through sensitivity analysis errors in actual evaporation due to meteorological and other inputs as follows: temperature (input ±1.5 °C) ±3 %, solar radiation (input ±10 %) ±6 %, vapour pressure (input ±0.15 kPa) ±3 %, wind speed (input ±50 %) ±7 %, elevation (input ±50 %) ±1 %, latitude (input ±2°) ±1 %, water depth (input ±1 m) ±1 %, and water area (±20 %) ±20 %.

Fisher et al. (2011) compared three models – Thornthwaite, Priestley-Taylor and Penman-Monteith – at 10 sites in the Americas and one in South Africa. The potential evapotranspiration estimates varied by more than 25 % across the sites, the PM model generally gave the highest PET estimates and Thornthwaite 20–30 % lower than PT or PM. At the global and continental scales, the three models gave similar averaged PET estimates.

To provide a more detailed guide to relative differences in the estimates of evaporation based on the models discussed in this paper, we review and summarise 27 references in Table 5 where cross-comparisons are carried out. (A consolidated list of relative differences is presented in Table 6 and a consolidated list of uncertainty estimates is available in Table 7.) For each case in Table 5 we have provided, where possible, an estimate of the mean annual evaporation by the specific procedure as a ratio of one of three base methods: (i) estimates based on a water balance, eddy correlation or Bowen Ratio study; (ii) estimates based on lysimeter measurements of evaporation; or (iii) estimates compared with another procedure; we have used Penman-Monteith, FAO-56 Reference Crop, Priestley-Taylor and Hargreaves-Samani methods. An estimate of the uncertainty for each analysis is also summarised in Tables 7. Although space precludes a detailed discussion of the errors here, Fig. 3 provides a summary of the relative differences between the procedures where the results from at least two studies were available for comparison. A detailed discussion of the results presented in Tables 5, 6 and 7 and Fig. 3 is provided in Supplement Appendix S17.

Rather than undertake a direct comparison of the potential evapotranspiration estimates, Oudin et al. (2005) compared the efficiency of rainfall-runoff models when 27 different potential evapotranspiration models were used. Four lumped rainfall-runoff models were examined for a sample of 308 catchments from Australia, France and the United States. These catchments were mainly water-limited where potential evapotranspiration is less important to model performance. The study found little improvement in the efficiency of the rainfall-runoff models when the more complex and data intensive models were used. The models based on air temperature and radiation provided the best results (Oudin et al., 2005).

The majority of the literature has focused on providing a relative accuracy through ranking of the various models. Lowe et al. (2009) adopted a different approach and present a framework to quantify the uncertainties associated with estimates of reservoir evaporation generated using the pan coefficient method. The uncertainty in each model input was assessed (including rainfall and Class-A evaporation measurements, bird guard adjustment factor, pan coefficients and spatial transposition) and combined using Monte Carlo simulations. They applied the framework to three reservoirs in South-East Australia. The largest contributor to the overall uncertainty was the estimation of Class-A evaporation at locations without monitoring, followed by uncertainty in annual pan coefficients. The overall uncertainty in reservoir evaporation was found to be as large as ±40 % at three study sites (Lowe et al., 2009, p. 272). Factors affecting measurement errors in evaporation are discussed in detail in Allen et al. (2011). These can be combined with a methodology like that presented in Lowe et al. (2009) to assess uncertainty in evaporation estimates.

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5 Concluding summary

This is not a review paper, but rather a considered summary of techniques that are readily available to the researcher, consulting hydrologist and practicing engineer to estimate both actual and potential evaporation. There are three key procedures that are used to estimate potential evaporation: Penman, Penman-Monteith and Priestley-Taylor. To estimate reference crop evaporation, FAO-56 Reference Crop equation, which is a Penman-Monteith equation for a 0.12 m high hypothetical crop in which the surface resistance is $70\,\mathrm{s\,m^{-1}}$, is used world-wide. It is applicable to humid conditions. If reliable pan coefficients are available, Class-A evaporation pans provide useful data for a range of studies and the PenPan model, which models very satisfactorily Class-A pan evaporation, is a useful tool to the hydrologist. The Penman equation estimates actual evaporation from shallow open-surface water in which the heat and vapour fluxes have no impact on the over-passing air. There are two wind functions (Penman, 1948, 1956, Eqs. 8a and 8b) which have been widely used that form part of the aerodynamic term in the Penman equation. We prefer the Penman (1956, Eq. 8b) wind function for most studies. There are a range of techniques available to estimate monthly actual evaporation of a catchment including Morton's procedure, the aridity-advection model of Brutsaert-Strickler, and the models of Szilagyi-Jozsa and Granger-Gray. The Budyko-like equations may be used to estimate annual actual evaporation. However, analysts need to be aware that changes in land surface conditions due to vegetation and lateral inflow may occur; these are best modelled using remote sensed data as inputs, an issue that is not explored here.

Turning to other practical topics, we observed that the Penman or Penman-Monteith models, incorporating a seasonal heat storage component and a water advection component, and the Morton CRLE model can be used to estimate evaporation from deep lakes and large voids. For shallow lakes or deep lakes, where only a mean annual evaporation estimate is required, the Morton model can be applied. Both the Penman and the Penman-Monteith equations modified to take into account heat storage effects

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are also acceptable procedures. For catchment water balance studies, in addition to the traditional simple water balance approach, the Morton model CRAE can be used. Our review of the literature suggests that any one of a number of the techniques can be used to estimate potential evaporation in rainfall-runoff modelling where the model parameters are calibrated. It has been customary in recent years to apply the FAO-56 Reference Crop method to estimate crop water requirements. It is noted for semi-arid windy regions that a more suitable method is the Matt-Shuttleworth model. Other practical topics, that are considered, include evaporation from lakes covered by vegetation, bare soil evaporation and groundwater evaporation.

There are six additional issues addressed. We noted that care needs to be exercised in using hard-wired evaporation estimates from commercially available automatic weather stations. Where wind data are not available we observed that Morton's procedure can be used and Valiantzas (2006) developed an empirical equation to simulate Penman without wind data. Where at-site meteorological data or Class-A pan data are not available at or nearby the target site, outputs from spatial interpolation and spatial modelling offer an approach. The paradox of increasing annual temperature but decreasing evaporative power observed in many parts of the world is briefly addressed. We observe that in the context of a changing climate the four key variables for estimating evaporative demand (radiation, air temperature, relative humidity and wind) should be taken into account. We note that, except for several exceptions recorded in the paper and supplementary appendices, standard meteorological data averaged (or estimated as an average) over a 24-h day rather than during daylight hours should be used in analysis. The last issue to be addressed is uncertainty in evaporation estimates. The main focus here was a literature review in which measures of uncertainty were collated, allowing the relative accuracies of most potential and actual evaporation procedures to be assessed.

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Table 1. Data required to compute evaporation using key models described in the paper.

Models	Penman	Penman-Monteith	Priestley-Taylor	FAO 56 Ref. Crop	PenPan	Morton CRAE	Morton CRWE	Morton CRLE	Advection-Aridity
Sub-section discussed	2.1.1	2.1.2	2.1.3	2.2	2.3.1	2.5.2	2.5.2	2.5.2	2.5.3
Time-step (D = daily, M = Monthly)	D or M	D	D or M	D	M	M (or D)	M (or D)	M	D
Sunshine hours or solar radiation	yes	yes	yes	yes	yes	yes	yes	yes	yes
Maximum air temperature	yes	yes		yes	yes	yes	yes	yes	yes
Minimum air temperature	yes	yes		yes	yes	yes	yes	yes	yes
Relative humidity	yes	yes		yes	yes	yes	yes	yes	yes
Wind speed	yes	yes		yes	yes				yes
Latitude						yes	yes	yes	
Elevation						yes	yes	yes	
Mean annual rainfall						yes	yes	yes	
Salinity of lake								yes	
Average depth of lake								yes	

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Table 2. Morton's models (α is albedo, ε_s is surface emissivity, and b_0, b_1, b_2 and f_Z are defined in Appendix S7).

Program	WREVAP (Morton 1983a, 1983b, 1986)			
Environment	Land environment		Shallow lake	Deep lake
Radiation input (if not using Morton, 1983a method)	α = 0.10–0.30, depending on vegetation ε_s = 0.92		α = 0.05 ε_s = 0.97	α = 0.05 ε_s = 0.97
Models	CRAE		CRWE	CRLE
Data	Latitude, elevation, mean annual precipitation, and daily temperature, humidity and sunshine hours		As for CRAE plus lake salinity (Morton, 1986, Sect. 4, item 2)	As for CRAE plus lake salinity and average depth (Morton, 1986, Sect. 4)
Component models and variable values	ET_{Pot}^{MO} Potential evapotranspiration Morton (1983a) b_0 = 1.0 (page 64) f_Z = 28 W m^{-2} mbar^{-1} (page 25)	For Australia (Chiew and Leahy, 2003, Sect. 2.3) b_0 = 1.0 f_Z = 29.2 W m^{-2} mbar^{-1}	E_{Pot}* Potential evaporation (in the land environment) or pan-size wet surface evaporation Morton (1983a, p. 26) b_0 = 1.12 f_Z = 25 W m^{-2} mbar^{-1}	
	ET_{Wet}^{MO} Wet environment areal evapotranspiration Morton (1983a, p. 25) b_1 = 14 W m^{-2} b_2 = 1.2	For Australia (Chiew and Leahy, 2003, Sect. 2.3) b_1 = 13.4 W m^{-2} b_2 = 1.13	ET_{Wet}^{MO} Shallow lake evaporation Morton (1983a, p. 26) b_1 = 13 W m^{-2} b_2 = 1.12 R_{ne} (net radiation at T_e °C)	ET_{Wet}^{MO} Deep lake evaporation b_1 = 13 W m^{-2} b_2 = 1.12 R_{ne} (net radiation at T_e °C) with seasonal adjustment of solar and water borne inputs
Outcome	Actual areal evapotranspiration $ET_{Act}^{MO} = 2ET_{Wet}^{MO} - ET_{Pot}^{MO}$		E_{SL} Shallow lake evaporation	E_{DL} Deep lake evaporation

* According to Morton (1986, p. 379, item 4) in the context of estimating lake evaporation, E_{Pot} has no "... real world meaning ..." because the estimates are sensitive to both the lake energy environment and the land temperature and humidity environment which are significantly out of phase. This is not so with lake evaporation as the model accounts for the impact of overpassing air.

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Table 3. Functional relationships for the Budyko-like relationships (Ø is the aridity index, *e* is the Turc-Pike parameter, *f* is the Fu parameter, *w* is the plant available water coefficient, and *c* is a parameter in the linear model).

Model	Model details	Reference
Schreiber	$F(\varnothing) = [1 - \exp(-\varnothing)]$	Schreiber (1904)
Ol'dekop	$F(\varnothing) = \varnothing \tanh\left(\varnothing^{-1}\right)$	Ol'dekop (1911)
Generalised Turc-Pike	$F(\varnothing) = \left(1 + \varnothing^{-e}\right)^{\frac{-1}{e}}$ For the Turc-Pike model, $e = 2$	Milly and Dunne (2002); Turc (1954); Pike (1964)
Budyko	$F(\varnothing) = \left\{\varnothing[1 - \exp(-\varnothing)]\tanh\left(\varnothing^{-1}\right)\right\}^{0.5}$	Budyko (1974)
Fu-Zhang	$F(\varnothing) = 1 + \varnothing - \left[1 + (\varnothing)^{f}\right]^{f^{-1}}$	Fu (1981); Zhang et al. (2004)
Zhang 2-parameter model	$F(\varnothing) = (1 + w\varnothing)\left(1 + w\varnothing + \varnothing^{-1}\right)^{-1}$	Zhang et al. (2001)
Linear model	$F(\varnothing) = c\varnothing$	Potter and Zhang (2009)

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Table 4. Practical application in estimating monthly evaporation. (This summary is based on models described in the paper and supplementary material that have an appropriate theoretical background, and a range of field testing. We have not included empirically-based techniques that are discussed in Supplement Sect. S9.)

Model, (Reference), Section	Application, (ET_{type}), Section: Deep lakes (ET_{act}), 3.1	Shallow lakes (ET_{act}), 3.2	Catchment water balance (ET_{act}), 3.3	Estimating crop requirements (ET_{act}), 3.5	Lakes with vegetation (ET_{act}), 3.6	Bare soil evaporation (ET_{act}), 3.7	Rainfall-runoff modelling (ET_{pot}), 3.4
Penman 1956, (Penman, 1956), 2.4.1	×	♣♣♣ < 2 m*	×	×	×	×	♣♣
Penman plus Kohler and Parmele, (Kohler and Parmele, 1967), 3.1.1	♣♣	×	×	×	×	×	×
Penman plus Vardavas-Fountoulakis, (Vardavas and Fountoulakis, 1996), 3.1.1	♣♣	×	×	×	×	×	×
Penman based on equilibrium temperature, (Finch 2001), 3.2	×	♣♣	×	×	×	×	×
Penman-Monteith, (Monteith, 1965), 2.1.2	×	×	×	♣	×	×	♣♣
FAO-56 Ref Crop, (Allen et al., 1998), 2.2	×	×	×	♣♣♣ (humid)	×	×	×
Matt-Shuttleworth, (Shuttleworth and Wallace, 2009), 3.5	×	×	×	♣♣♣ (windy, semi-arid)	×	×	×
Weighted Penman-Monteith, (Wessel and Rouse, 1994), 3.6	×	×	×	×	♣	×	×
Penman-Monteith based on equilibrium temperature, (McJannet et al., 2008), 3.2	♣♣	♣♣	×	×	×	×	×
Priestley-Taylor, (Priestley and Taylor, 1972), 2.1.3	×	×	×	×	×	×	♣♣
Morton, (Morton 1983a, 1986), 2.5.2	♣♣♣	♣♣♣	♣♣	×	×	×	♣♣
Advection-Aridity, (Brutsaert and Strickler, 1979), 2.5.3	×	×	♣	×	×	×	×
Szilagyi-Jozsa, (Szilagyi and Jozsa, 2008), 2.5.3	×	×	♣	×	×	×	×
Granger-Gray, (Granger, 1989b; Granger and Gray, 1989), 2.5.3	×	×	♣	×	×	×	×
Budyko-like models, (Budyko, 1974; Potter and Zhang, 2009), 3.3	×	×	♣♣ (annual)	×	×	×	×
Lake finite-difference model, (Finch and Gash, 2002),3.2	♣	♣♣	×	×	×	×	×
Salvucci for bare soil, (Salvucci, 1997), 3.7	×	×	×	×	×	♣	×
Class-A pan evaporation or PenPan, (Rotstayn et al., 2006), 2.3	♣	♣	×	♣	×	×	♣♣

♣♣♣ preferred; ♣♣ acceptable; ♣ not preferred or insufficient field testing; × not recommended.
* Based on Monteith (1981, p. 9) and others (see Appendix S11), we suggest that Penman (1956) be not used for lakes greater than 2 m in depth.

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Table 5. Bias and uncertainty in published estimates of actual and/or potential evaporation. (P48 or P56: Penman 1948- or 1956-wind function; PM: Penman-Monteith; FAO56 RC: FAO-56 Reference Crop; SW: Shuttleworth-Wallace; Mo: Morton CRAE; BS: Brutsaert-Strickler; GG: Granger-Gray; PT: Priestley-Taylor; modH: modified Hargreaves; Dalton-type: equations with a structure similar to Dalton; Th 1948 or 1955; Ma; Makkink equation; BC: Blaney and Criddle FAO-24 Reference Crop; Tu: Turc (1961) equation; HS: Hargreaves-Samani).

Ref	P48 or P56	PM	FAO56 RC	SW	Mo	BS or GG	PT	modH	Dalton-type	Th 1948 or 1955	Ma	BC	Tu	HS
1	Keijman and Koopmans (1973), Table 1: Comparison with lake water balance in Holland over 32 days. Dalton coefficient based on Lake Mead data. Bold values are ratios of average daily actual evaporation to water balance estimates.													
	1948 **1.00**								**1.51**					
2	Gunston and Batchelor (1982), Figs. 1, 2, 3: Comparing monthly Priestley–Taylor with Penman. R^2 is correlation coefficient squared, b is slope of regression between the two estimates for 30 world-wide data sets.													
	wet months $R^2 = 0.76$, $b = 0.91$, intermediate months $R^2 = 0.81$, $b = 0.85$, dry months $R^2 = 0.23$, $b = 0.34$													
3	Jensen et al. (1990), Table 7.20: Comparison with lysimeter measurements adjusted to reference crop values. 11 sites world-wide. Results based on average of monthly values at all locations. Bold values are the percentage differences from lysimeter measurements. Values in parenthesis are weighted standard errors of estimate (mm day^{-1}).													
	Arid locations													
	1956 **0.98** (0.70)		**0.99** (0.49)				**0.73** (1.89)			1955 **0.63** (2.40)		**1.00** (0.76)	**0.74** (1.88)	**0.91** (1.17)
	Humid locations													
	1956 **1.14** (0.60)		**1.04** (0.32)				**0.97** (0.68)			1955 **0.96** (0.86)		**1.16** (0.79)	**1.05** (0.56)	**1.25** (0.79)
4	Stannard (1993), Table 3: Comparison with eddy correlation measurements in sparsely vegetated semiarid rangelands over 58 days during four-year period. Parameters representing surface control of upward vapour flux were estimated by calibration Bold values are the ratios of median daily model actual daily evaporation to the eddy correlation daily evaporation estimates.													
		1.14		**1.12**			**1.05**							
5	Amatya et al. (1995), Tables 5 and 6: Estimates compared with PM Ref Crop at three sites in North Carolina. Daily, monthly, annual analysis. Bold values are the ratios of average daily estimates with respect to PM reference crop estimates. Values in parenthesis are the average root mean square daily estimates (mm day^{-1}) of regression estimates of PM values at the three sites.													
			1.00^{+}				**0.91** (0.80)			1948 **0.84** (1.40)	**0.86** (0.83)		**1.00** (0.84)	**1.14** (1.15)
6	Abtew and Obeysekera (1995), Figs. 6, 8 and 9: Based on evaporation from a lysimeter containing cattails (*Typha domingensis*) located in South Florida. Values are slopes of regression with intercept between modelled actual evaporation and lysimeter data. The Penman wind function was calibrated for the site. Values in parentheses are the standard errors of estimate (mm day^{-1}).													
	1.01 (0.57)	0.75 (0.38)					0.70 (0.53)							

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Table 5. Continued.

Ref	P48 or P56	PM	FAO56 RC	SW	Mo	BS or GG	PT	modH	Dalton-type	Th 1948 or 1955	Ma	BC	Tu	HS
7	Federer et al. (1996) Fig. 1: Analysis for seven locations across U.S., and based on one year of data. *Reference surface potential evaporation (1948 Penman wind function and albedo = 0.25) as italics*. Bold values are the ratios of potential evaporation with respect to Penman.													
	1948 *1.00*									1948 **0.84**			**0.89**	
	Surface-dependent potential evaporation as italics. Bold values are ratios of the potential evaporation with respect to Penman-Monteith for cultivation, grassland, conifers and broadleaf vegetation.													
	Grassland	*1.0*		**1.24**			**0.89**							
	Conifer	*1.0*		**1.15**			**1.18**							
	Broadleaf	*1.0*		**1.23**			**1.45**							
	Cultivation	*1.0*		**1.05**			**1.02**							
8	Souch et al. (1998), Table 3: For a wetland (undisturbed, disturbed and wet, disturbed and dry condition) in Indiana, computed ETs for the three conditions were compared with eddy correlation values. Bold values are the ratios of computed to observed values only for dry conditions.													
	1948 **1.11**	**1.01**					**1.10**							
9	Xu and Singh (2000), Table III: Data recorded over five years at a climate station in Switzerland. Bold values are the ratios compared with Priestley-Taylor values.													
							1.0				**0.88**		**0.93**	**1.02**
10	Xu and Singh (2001), Table II: Data recorded over five years at two climate stations in Canada for 12 and 15 yr. Bold values are the ratios compared with Hargreaves-Samani values.													
										0.95		**1.22**		*1.0*
11	Abtew (2001), Fig. 3, Table 6: Based on detailed analysis of five years of data for Lake Okeechobee, United States. The Penman wind function was calibrated for a nearby site. Results are compared with water budget estimates. Calibrated value of α_{PT} = 0.18 in PT. Bold values are ratios of mean estimates to mean water balance values.													
	1.05*						**1.03**						**0.88**	
12	Xu and Singh (2002), Fig. 2: Data recorded over five years for a grassland site in Switzerland. Estimates compared with PM Ref Crop. Values are slopes of regression (intercept not zero) between method and PM Reference Crop estimate.													
			1.00				0.68				0.95	1.00		0.89
13	Rosenberry et al. (2004), Figs. 2 and 4: Based on five years of data for a small wetland in North Dakota. Estimates compared with Bowen Ratio energy balance values. Bold values are ratios of mean estimates to energy balance estimates. Values in parentheses are the standard deviations of the mean differences (mm day^{-1}).													
	1956 **1.02** (0.4)					BS **0.98** (0.7)	**1.02** (0.3)				**0.93** (0.4)			
14	Sumner and Jacobs (2005): 30-min modelled values of evaporation for 19 months from a non-irrigated pasture in Florida U.S. were compared with eddy correlation measurements. Values are the standard errors (mm day^{-1}) (in parentheses) and R^2. Model coefficients were calibrated.													
		(1.48) 0.81					(1.08) 0.88							
15	Lu et al. (2005), Figs. 5–8: Based on monthly estimates of PET for four catchments (0.25 to 1036 km^2, 23 to 30 yr) in US. Bold values are the ratios of PET estimates compared to the Priestley-Taylor estimates.													
							1.00			**0.79**	**0.92**		**1.02**	**1.20**
16	Xu and Chen (2005) Table 1: Comparisons of actual evapotranspiration based on 12 yr of annual lysimeter data of grass in Germany. Bold values are the mean estimates as ratios of observed values. Values in parenthesis are the average standard deviation of annual errors in percent.													
					CRAE **1.12** (8.95)	BS **1.00** (6.22) GG **0.98** (6.75)	**1.01** (11.5)			**1.07** (13.1)	**1.05** (10.5)			**1.12** (12.9)

Table 5. Continued.

Ref	P48 or P56	PM	FAO56 RC	SW	Mo	BS or GG	PT	modH	Dalton-type	Th 1948 or 1955	Ma	BC	Tu	HS
17	Nandagiri and Kovoor (2006), Table 6: Monthly and daily analysis for four climate stations in India. Bold values are the average percentage differences from FAO-56 Ref. Crop of the four sites and the average monthly values. Values in parenthesis are the average standard errors of estimate (mm day^{-1}) of regression with PM values of the four sites.													
17a	Arid		*1.00*				**1.02** (0.95)					**1.08** (0.95)	**0.87** (1.4)	**1.27** (1.01)
17b	Semi-arid		*1.00*				**1.09** (1.16)					**1.44** (0.59)	**1.17** (1.05)	**0.87** (0.67)
17c	Sub-humid		*1.00*				**1.11** (0.56)					**1.01** (0.57)	**1.20** (0.74)	**0.87** (0.38)
17d	Humid		*1.00*				**0.89** (0.64)					**0.98** (0.48)	**1.53** (0.20)	**0.88** (0.62)
18	Rosenberry et al. (2007), Fig. 4: Based on 37 months of data for small lake in North Hampshire. Estimates are compared with Bowen Ratio energy budget measurements. Bold values are the ratios of mean estimates to energy balances. Values in parentheses are the standard deviations of the differences between the values and the Bowen Ratio estimates (mm day^{-1}).													
	1956 **1.09** (0.2)					BS **1.09** (0.4)	**1.09** (0.2)				**0.98** (0.5)	**1.42** (0.8)		
19	Schneider et al. (2007), Table 2: Comparison is based on two years of actual crop estimates produced by SWAT model incorporating a specific evaporation model (without calibration) with eddy flux measurements for a region in semi-arid northern China. Values are ratio of model to eddy flux observations.													
			0.83				**0.85**				**0.95**			
20	Weiß and Menzel (2008), Table 2: Based on 23 sites in Jordon River region. PT is compared with PM. Bold values are the ratios of average annual ET estimated by PT and PM. The values in parenthesis are the RMSEs expressed as a percentage of average annual PM ETs.													
			1.00				**1.17** (17.0%)							**1.40** (43.9%)
21	Alexandris et al. (2008), Tables 1 and 2: Based on analysis of two summers of data at one grassland site in Serbia. Bold values are the ratios of the mean between model and PM Reference Crop estimates. Values in parenthesis are the RMSEs (mm day^{-1}).													
			1.00				**1.05** (0.20)				**0.79** (0.60)		**0.95** (0.23)	**1.13** (1.01)
22	Shi et al. (2008), Table 2: Model estimates for three growing seasons of temperate mixed forest in Changbai Mountains in northeastern China compared with eddy covariance observations. Bold values are the ratios of daily model estimates compared with eddy covariance values.													
		KP **0.94**^ TD **1.36**					**1.12**							
23	Ali et al. (2008), Table VI: Model estimates compared with Bowen Ratio energy balances over four years for a small lake in semi-arid region of India. Bold values are the ratios of the model estimates to Bowen Ratio measurements.													
							1.00		**1.06**					
24	Yao (2009) Fig. 7, Table 4: Based on a small lake in Ontario, Canada. Methods compared with energy budget. 23 yr of data. Bold values are the slopes of regression for zero intercept. Values in parentheses are the RMSEs and the Nash-Sutcliffe efficiencies.													
	1948 **1.06** (4.1, 0.92)						**1.10** (4.7, 0.89)				**0.92** (5.6, 0.85)			
25	Trajkovic and Kolakovic (2009), Table 3. Monthly values are compared with FAO-56 Reference Crop for seven climate stations in Croatia and Serbia. Bold values are the ratios of average modelled ET to PM values. Values in parenthesis are the RMSDs (mm month^{-1}).													
			1.00				**1.01** (9.1)			**0.89** (15.5)			**0.95** (8.8)	**1.23** (19.2)

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Table 5. Continued.

Ref	P48 or P56	PM	FAO56 RC	SW	Mo	BS or GG	PT	modH	Dalton-type	Th 1948 or 1955	Ma	BC	Tu	HS
26	Douglas et al. (2009), Table 5 and Fig. 5: Based on 18 sites in Florida covering forest, grassland, citrus, wetlands and lakes. Observed daily ET was estimated from a range of energy budget techniques including eddy covariance and Bowen Ratio. Here, model estimates for forests, grass/pastures and lakes are compared with daily measured values over periods from 507 to 968 days where Bowen Ratios > 1. Bold values are the ratios of model estimates to observed estimates. Values in parenthesis are the RMSEs (mm day^{-1}).													
26a	Forest	**0.97**** (1.82)					**1.37** (2.32)						**1.37** (1.75)	
26b	Grass	**0.72** (0.83)					**1.29** (1.13)						**1.42** (1.23)	
26c	Lakes	**0.99** (1.05)					**0.94** (1.25)						**0.80** (1.43)	
27	Elsawwaf et al. (2010): Based on 10 yr of data for Lake Nasser, Egypt. Monte Carlo analysis was used to assess uncertainty in lake evaporation measurements. Values in parenthesis are the uncertainty estimates as percentage of mean evaporation rates.													
	1956 (12.7)						(13.3)		(15.3)	(14.1)				

+ *1.00* in italics indicates model adopted for comparison.
\# Energy budget method.
* Wind function coefficients were calibrated using data from a cattail marsh (Abtew and Obeysekera, 1995).
ˆ KP Aerodynamic and canopy resistances defined by Katerji and Perrier (1983) and calibrated as discussed by Shi et al. (2008).
TD Aerodynamic and canopy resistances defined by Todorovic (1999).
** Adjusted from 0.47 to 0.97 based on at-site data in Douglas et al. (2009), Fig. 3.

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Estimating actual, potential, reference crop and pan evaporation

T. A. McMahon et al.

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Table 6. Consolidated list of biases expressed as ratios of model estimations of actual and/or potential evaporation to field measurements, lysimeter observations or comparison with evaporation equations. (P48: Penman, 1948; P56: Penman 1956; PM: Penman-Monteith; FAO56 RC: FAO-56 Reference Crop; SW: Shuttleworth-Wallace; BS, GG: Brutsaert-Strickler or Granger-Gray, respectively; PT: Priestley-Taylor; Dalton: Dalton-type model; Th: Thornthwaite; Ma: Makkink; BC: Blaney-Criddle; Tu: Turc; HS: Hargreaves-Samani.)

Ref#	Surface	Location/climate	P48	P56	PM	FAO56 RC	SW	BS, GG	PT	Dalton	Th	Ma	BC	Tu	HS
Comparisons with water balance, eddy correlation or Bowen Ratio															
1	Lake	Holland/temperate	1.00							1.51					
11	Lake	Florida/sub-tropical		1.05*					1.03					0.88	
18	Lake	North Dakota/cold		1.09				1.09 BS	1.09			0.98	1.42		
23	Lake	India/semi-arid							1.00	1.06					
24	Lake	Canada/cold	1.06						1.10			0.92			
26c	Lake	Florida/sub-tropical			0.99				0.94					0.80	
		Count	2	2					5	2		2		2	
		Average	1.03	1.07					1.03	1.29		0.95		0.84	
Comparisons with eddy correlation or Bowen Ratio															
8	Dry wetland	Indiana/cold	1.11		1.01				1.10						
13	Wetland	North Dakota/cold		1.02				0.98 BS	1.02			0.93			
22	Forest	NE China/cold			1.15$				1.12						
26a	Forest	Florida/sub-tropical			0.97				1.37					1.37	
4	Rangeland	Colorado/semi-arid			1.14		1.12		1.05						
19	Rangeland	China/semi-arid				0.83			0.85			0.95			
26b	Grassland	Florida/sub-tropical			0.72				1.29					1.42	
		Count			5				7			2		2	
		Average			1.00				1.11			0.94		1.40	
Comparisons with lysimeter measurements															
3a	Grass	World-wide/arid		0.98		0.99			0.73		0.63		1.00	0.74	0.91
3b	Grass	World-wide/humid		1.14		1.04			0.97		0.96		1.16	1.05	1.25
6	Wetland	South Florida/humid		1.01	0.75				0.70						
16	Grass	Germany/ temperate/cold						1.00 BS 0.98 GG	1.01		1.07	1.05			1.12
		Count		3	3				4		3		2	2	3
		Average		1.04	0.93				0.85		0.89		1.08	0.90	1.09

Table 6. Continued.

Ref#	Surface	Location/climate	P48	P56	PM	FAO56 RC	SW	BS, GG	PT	Dalton	Th	Ma	BC	Tu	HS
Comparisons with Penman-Monteith (average values as ratio of PM values = *1.00*)															
5	Ref. Crop	North Carolina/temperate				*1.00*			0.91		0.84	0.86		1.00	1.14
21	Grassland	Serbia/temperate/cold				*1.00*			1.05			0.79		0.95	1.13
7a	Grassland	USA/cold to semi arid			*1.00*		1.24		0.89						
7b	Conifer	USA/cold to semi arid			*1.00*		1.15		1.18						
7c	Broadleaf	USA/cold to semi arid			*1.00*		1.23		1.45						
7d	Cultivation	USA/cold to semi arid			*1.00*		1.05		1.02						
17a	Climate stn.	India/arid				*1.00*			1.02				1.08	0.87	1.27
17b	Climate stn.	India/semi-arid				*1.00*			1.09				1.44	1.17	0.87
17c	Climate stn.	India/semi-humid				*1.00*			1.11				1.01	1.2	0.87
17d	Climate stn.	India/humid				*1.00*			0.89				0.98	1.53	0.88
20	Irrigation to desert	Jordon /arid				*1.00*			1.17						1.40
25	Climate stn.	Croatia, Serbia/humid				*1.00*			1.01	0.89				0.95	1.23
		Count					4		12		2	2	4	7	7
		Average				*1.00*	1.17		1.07		0.87	0.83	1.13	1.10	1.06
Comparisons with Priestley-Taylor (average values as ratio of PT values = *1.00*)															
9	Climate stn.	Switzerland/cold							*1.00*			0.88		0.93	1.02
15	Forest	USA/Humid							*1.00*		0.79	0.92		1.02	1.20
		Count										2		2	2
		Average							*1.00*			0.90		0.98	1.11
Comparison with Hargreaves-Samani (average values as ratio of H-S value = *1.00*)															
10	Climate stn.	Canada/cold									0.95		1.22		*1.00*

Numbers refer to references listed in Table 5. * Indicate which of the three models the results refer to. $ Average of KP and TD values in item 22 of Table 6.

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Table 7. Consolidated list of uncertainty estimates as RMSE or SEE expressed as ratio of the equivalent values estimated for the Priestley-Taylor equation. (P56: Penman 1956; PT: Priestley-Taylor; Ma: Makkink; PM: Penman-Monteith; BC: Blaney-Criddle; HS: Hargreaves-Samani; Tu: Turc; Th: Thornthwaite.)

Ref.*	P56	PT	Ma	PM	BC	HS	Tu	Th
RMSE (mm day^{-1})								
#5		*1.00*	1.04			1.44	1.05	1.75
#21		*1.00*	3.00			5.05	1.15	
#25		*1.00*				2.10	0.97	1.70
#26a		*1.00*		0.78			0.75	
#26b		*1.00*		0.73			1.09	
#26c		*1.00*		0.84			1.14	
Median		1.00	2.02	0.78		2.10	1.07	1.73
SEE (mm day^{-1})								
#3A	0.37	*1.00*		0.26	0.40	0.62	0.99	1.27
#3H	0.88	*1.00*		0.47		1.16	0.82	1.26
#6	1.08	*1.00*		0.72				
#13	1.33	*1.00*	1.33					
#17a		*1.00*			1.00	1.06	1.47	
#17b		*1.00*			0.51	0.58	0.91	
#17c		*1.00*			1.02	0.68	1.32	
#17d		*1.00*			0.75	0.97	0.31	
Median	0.98	1.00	1.33	0.47	0.75	0.83	0.95	1.27

* Numbers refer to references listed in Table 5.

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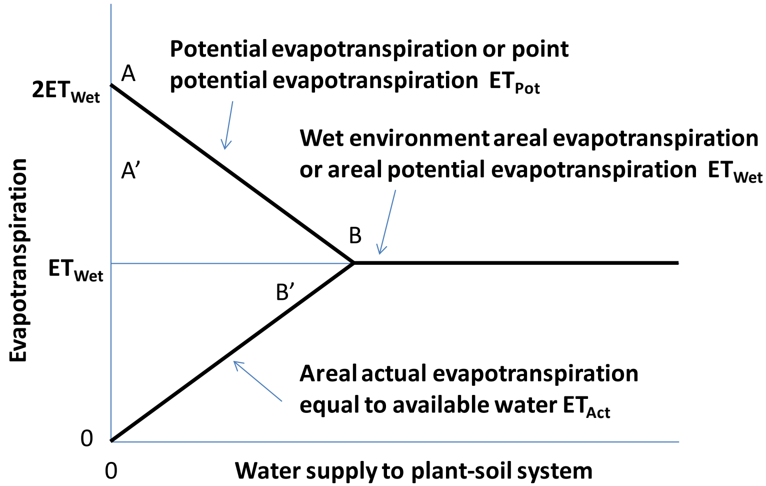

Fig. 1. Theoretical form of the Complementary Relationship.

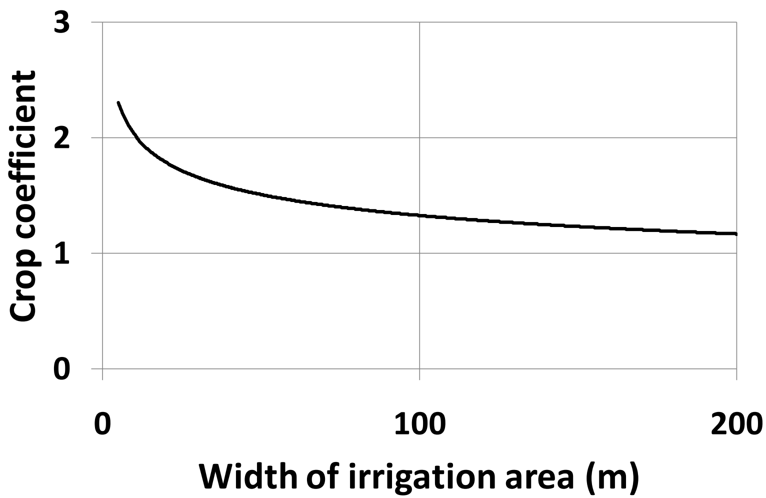

Fig. 2. Effect of an "oasis" environment on irrigation water requirement (adapted from Allen et al., 1998).

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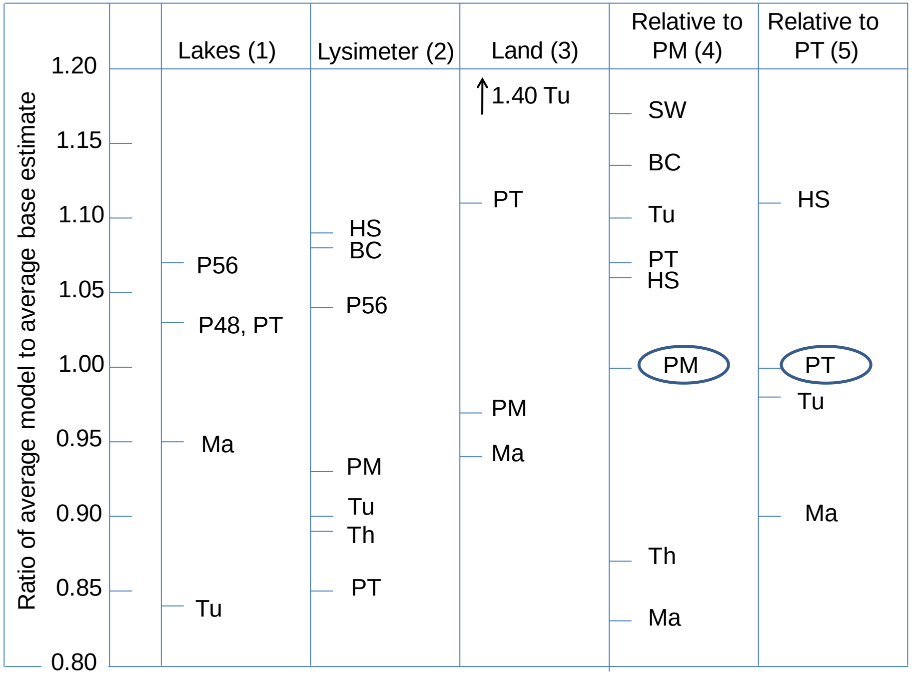

Fig. 3. Pictorial comparison of published evaporation estimates from Table 6. Values are average ratios of the nominated procedures to base evaporation. For the lakes, the base evaporation estimation was by water balance, eddy correlation or Bowen Ratio, and for lysimeter results the base was estimated for lysimeters containing grass. Land estimates were based on eddy correlation or Bowen Ratio. For the two columns to the right, the values were compared directly with Penman-Monteith or Priestley-Taylor, both set arbitrarily to a ratio of 1.00. Symbols are defined at the head of Table 6.

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