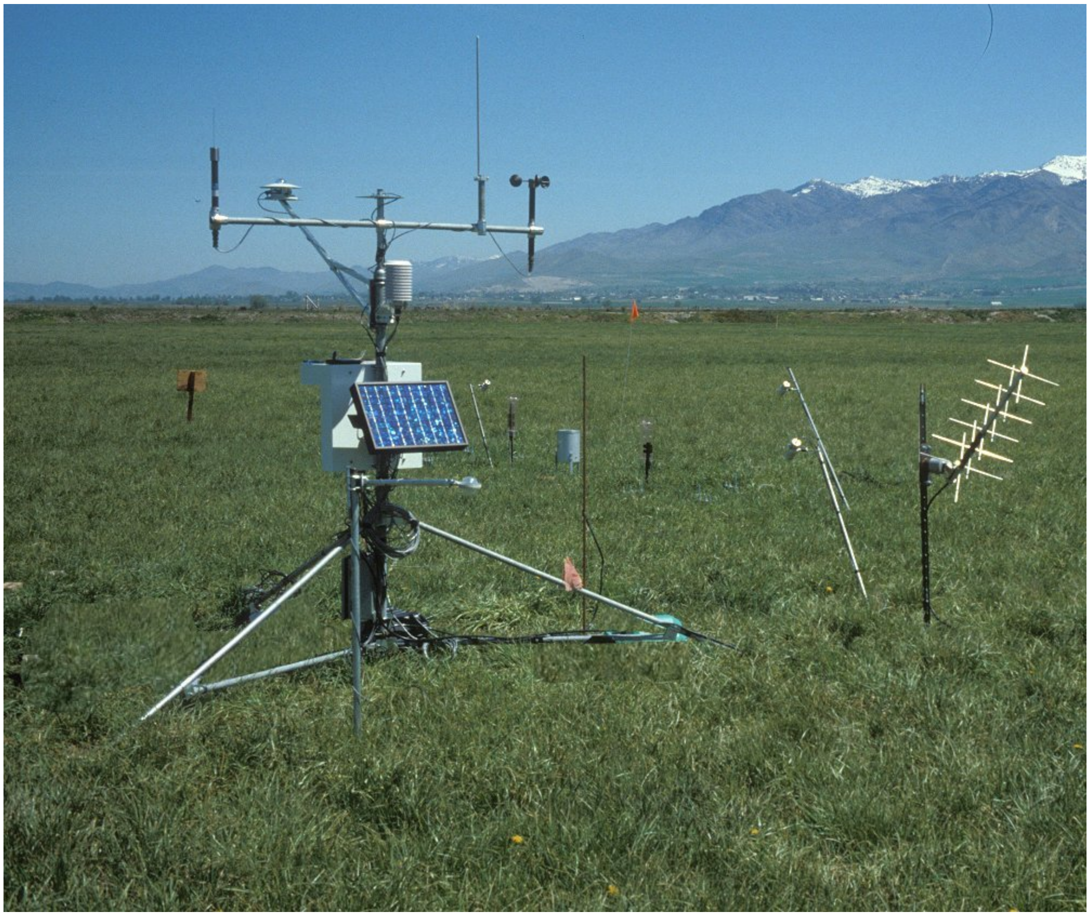
THE ASCE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION



Task Committee on Standardization of Reference Evapotranspiration

Environmental and Water Resources Institute
of
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THE ASCE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION

PREPARED BY

Task Committee on Standardization of Reference Evapotranspiration

of the

Environmental and Water Resources Institute

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THE IRRIGATION ASSOCIATION

Irrigation Association, 2004

ABSTRACT

This report describes the standardization of calculation of reference evapotranspiration (ET) as recommended by the Task Committee on Standardization of Reference Evapotranspiration of the Environmental and Water Resources Institute of the American Society of Civil Engineers. The purpose of the standardized reference ET equation and calculation procedures is to bring commonality to the calculation of reference ET and to provide a standardized basis for determining or transferring crop coefficients for agricultural and landscape use. The basis of the standardized reference ET equation is the ASCE Penman-Monteith (ASCE-PM) method of ASCE Manual 70. For the standardization, the ASCE-PM method is applied for two types of reference surfaces representing clipped grass (a short, smooth crop) and alfalfa (a taller, rougher agricultural crop), and the equation is simplified to a reduced form of the ASCE–PM. Standardized calculations for vapor pressure, net radiation and wind speed adjustment are recommended for application to hourly and daily calculation time steps. Guidelines on assessing weather data integrity and estimating values for missing data are provided.

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Task Committee on Standardization of Reference Evapotranspiration

PREFACE

The concept of reference evapotranspiration (ET) was developed in the 1970's as a practical and definable replacement for the term potential ET. Reference ET is a function of local weather, represents the ET from a defined vegetated surface, and serves as an evaporative index by which engineers, hydrologists, water managers and other technical professionals can predict ET for a range of vegetation and surface conditions by applying "crop" coefficients for agricultural or landscaped areas. During the past decade, for convenience and reproducibility, the reference surface has been expressed as a hypothetical surface having specific characteristics. In the context of this standardization, reference evapotranspiration is defined as the ET rate from a uniform surface of dense, actively growing vegetation having specified height and surface resistance, not short of soil water, and representing an expanse of at least 100 m of the same or similar vegetation. The EWRI Task Committee concluded that two standardized surfaces were needed to serve the needs of the agricultural and landscape communities and to provide for continuity with past reference ET usage. The ASCE Penman-Monteith (ASCE-PM) equation of ASCE Manual 70 is used to represent the standardized surface and is applied for two types of surfaces (short and tall)-- clipped, cool-season grass and alfalfa.

This recommended standardization follows commonly used procedures for calculating vapor pressure terms, net radiation, and soil heat flux. The standardization represents reference ET for each of the reference surfaces using a single equation having fixed constants and standardized computational procedures. The computational procedures are relatively simple to apply, are understandable, are supported by existing and historical data, are technically defensible, and are accepted by science and engineering communities. The Task Committee recognizes that the standardized reference equation, with fixed coefficients defining vegetation and surface conditions, may not correspond precisely with local measurements of ET from surfaces similar to the clipped, cool-season grass and full-cover alfalfa definitions. However, the Task Committee encourages the use of the standardized equation and procedure when possible to represent reference ET for the establishment of reproducible and universally transferable ET estimates, climatic description, and derived crop and landscape coefficients. The standardized equation has been investigated over a wide range of locations and climates across the United States and has the Task Committee's confidence for use as a standardized index of evapotranspirative demand.

Some of the computational procedures of the standardized reference method, for example, the computation of net radiation, may be updated by EWRI from time to time in the future, as developments and improvements in generalized computational techniques are made.

The development of this standardization report by EWRI was made at the request of, and has been endorsed by, the Irrigation Association.

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INTRODUCTION

In May 1999, The Irrigation Association (IA) requested the Evapotranspiration in Irrigation and Hydrology Committee – Environmental and Water Resources Institute (American Society of Civil Engineers) (ASCE-ET) to establish and define a benchmark reference evapotranspiration equation. The purpose of the benchmark equation is to standardize the calculation of reference evapotranspiration and to improve transferability of crop coefficients.

IA envisioned an equation that would be accepted by the U.S. scientific community, engineers, courts, policy makers, and end users. The equation would be applicable to agricultural and landscape irrigation and would facilitate the use and transfer of crop and landscape coefficients. In addition, IA requested guidelines for using the equation in regions where climatic data are limited and recommendations for

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incorporating existing crop and landscape coefficients and existing reference ET calculations.

An ASCE-ET Task Committee (TC) comprised of the authors of this report responded to the request by IA. Their initial response is included in Appendix A. Members of the TC jointly authored several papers (Allen, et al., 2000; Itenfisu, et al., 2000; Walter, et al., 2000) at the IA 4th National Irrigation Symposium in November 2000 that described issues, challenges and analyses conducted by the TC. This report provides detail on development of the ASCE Standardized equation, recommendations on use of the equation, and example calculations. In addition, this report provides guidelines for assessing the integrity of weather data used for estimating ET and methodologies that can be used where data are limited or missing.

DEFINITION OF THE EQUATION

Evapotranspiration (ET) represents the loss of water from the earth's surface through the combined processes of evaporation (from soil and plant surfaces) and plant transpiration (i.e., internal evaporation). Reference evapotranspiration (ET_{ref}) is the rate at which readily available soil water is vaporized from specified vegetated surfaces (Jensen et al., 1990). For convenience and reproducibility, the reference surface has recently been expressed as a hypothetical crop (vegetative) surface with specific characteristics (Smith et al., 1991, Allen et al., 1994a, Allen et al., 1998). In the context of this standardization report, reference evapotranspiration is defined as the ET rate from a uniform surface of dense, actively growing vegetation having specified height and surface resistance, not short of soil water, and representing an expanse of at least 100 m of the same or similar vegetation.

ASCE-ET recommends that the equation be referred to as the "Standardized Reference Evapotranspiration Equation" (ET_{sz}). ASCE-ET is of the opinion that use of the terms *standard* or *benchmark* may lead users to assume that the equation is intended for comparative purposes (i.e., a level to be measured against). Rather, the use of the term "standardized" is intended to infer that the computation procedures have been fixed, and not that the equation is a standard or a benchmark or that the equation has undergone the degree of review in the approval process necessary for standards adopted by ASCE, ASAE, American National Standards Institute, or the International Organization for Standardization.

ASCE-ET and IA-WM members concluded that two ET_{ref} surfaces with *standardized* computational procedures were needed. The two adopted ET_{ref} surfaces are (1) a short crop (similar to clipped grass) and (2) a tall crop (similar to full-cover alfalfa). Additionally, the TC recognized that an equation capable of calculating both hourly and daily ET_{ref} was needed.

RECOMMENDATION

ET_{ref} from each of the two surfaces is modeled using a single Standardized Reference Evapotranspiration equation with appropriate constants and standardized computational procedures. The surfaces/equation are defined as:

<u>Standardized Reference Evapotranspiration Equation, Short (ET_{os}):</u> Reference ET for a *short* crop with an approximate height of *0.12* m (similar to clipped, cool-season grass).

<u>Standardized Reference Evapotranspiration Equation, Tall (ET_{rs}):</u> Reference ET for a *tall* crop with an approximate height of *0.50* m (similar to full-cover alfalfa).

The two surfaces are similar to known full-cover crops of alfalfa and clipped, cool-season grass that have received widespread use as ET_{ref} across the United States. Each reference has unique advantages for specific applications and times of the year. As a part of the standardization, the ASCE Penman-Monteith (ASCE-PM) equation

(Appendix B and Jensen et al., 1990), and associated equations for calculating aerodynamic and bulk surface resistance have been combined and condensed into a single equation that is applicable to both surfaces.

The Standardized Reference Evapotranspiration Equation is intended to simplify and clarify the presentation and application of the method. As used in this report, the term ET_{sz} refers to both ET_{os} and ET_{rs}. Eq. 1 presents the form of the Standardized Reference Evapotranspiration Equation:

$$ET_{sz} = \frac{0.408\,\Delta(R_n - G) + \gamma \dfrac{C_n}{T+273} u_2\,(e_s - e_a)}{\Delta + \gamma(1 + C_d\, u_2)} \tag{1}$$

where:

ET_{sz} = standardized reference crop evapotranspiration for short (ET_{os}) or tall (ET_{rs}) surfaces (mm d^{-1} for daily time steps or mm h^{-1} for hourly time steps),

R_n = calculated net radiation at the crop surface (MJ m^{-2} d^{-1} for daily time steps or MJ m^{-2} h^{-1} for hourly time steps),

G = soil heat flux density at the soil surface (MJ m^{-2} d^{-1} for daily time steps or MJ m^{-2} h^{-1} for hourly time steps),

T = mean daily or hourly air temperature at 1.5 to 2.5-m height (°C),

u_2 = mean daily or hourly wind speed at 2-m height (m s^{-1}),

e_s = saturation vapor pressure at 1.5 to 2.5-m height (kPa), calculated for daily time steps as the average of saturation vapor pressure at maximum and minimum air temperature,

e_a = mean actual vapor pressure at 1.5 to 2.5-m height (kPa),

Δ = slope of the saturation vapor pressure-temperature curve (kPa °C^{-1}),

γ = psychrometric constant (kPa °C^{-1}),

C_n = numerator constant that changes with reference type and calculation time step (K mm s^3 Mg^{-1} d^{-1} or K mm s^3 Mg^{-1} h^{-1}) and

C_d = denominator constant that changes with reference type and calculation time step (s m^{-1}).

Units for the 0.408 coefficient are m^2 mm MJ^{-1}.

Table 1 provides values for C_n and C_d. The values for C_n consider the time step and aerodynamic roughness of the surface (i.e., reference type). The constant in the denominator, C_d, considers the time step, bulk surface resistance, and aerodynamic roughness of the surface (the latter two terms vary with reference type, time step and daytime/nighttime). C_n and C_d were derived by simplifying several terms within the ASCE-PM equation and rounding the result. Equations associated with calculation of required parameters in Eq. 1, the detailed derivation of the parameters in Table 1 and simplification of the terms listed in Table 2 are explained in more detail in Appendix B. Daytime is defined as occurring when the average net radiation, R_n, during an hourly period is positive.

Table 1. Values for C_n and C_d in Eq. 1

Calculation Time Step	Short Reference, ET_{os}		Tall Reference, ET_{rs}		Units for ET_{os}, ET_{rs}	Units for R_n, G
	C_n	C_d	C_n	C_d		
Daily	900	0.34	1600	0.38	mm d^{-1}	MJ m^{-2} d^{-1}
Hourly during daytime	37	0.24	66	0.25	mm h^{-1}	MJ m^{-2} h^{-1}
Hourly during nighttime	37	0.96	66	1.7	mm h^{-1}	MJ m^{-2} h^{-1}

Table 2. ASCE Penman-Monteith Terms Standardized for Application of the Standardized Reference Evapotranspiration Equation

Term	ET_{os}	ET_{rs}
Reference vegetation height, h	0.12 m	0.50 m
Height of air temperature and humidity measurements, z_h	1.5 – 2.5 m	1.5 – 2.5 m
Height corresponding to wind speed, z_w	2.0 m	2.0 m
Zero plane displacement height	0.08 m	0.08 m [a]
Latent heat of vaporization	2.45 MJ kg^{-1}	2.45 MJ kg^{-1}
Surface resistance, r_s, daily	70 s m^{-1}	45 s m^{-1}
Surface resistance, r_s, daytime	50 s m^{-1}	30 s m^{-1}
Surface resistance, r_s, nighttime	200 s m^{-1}	200 s m^{-1}
Value of R_n for predicting daytime	> 0	> 0
Value of R_n for predicting nighttime	≤ 0	≤ 0

[a] The zero plane displacement height for ET_{rs} assumes that the wind speed measurement is over clipped grass, even though the reference type is tall. This is done to accommodate a majority of weather stations that are located over

grass. See comments in Appendix B following Eq. B.14b. When wind speed is measured over a surface having vegetation taller than about 0.3 m, it is recommended that the "full" ASCE Penman-Monteith method (Eq. B.1) be employed, where the zero plane displacement can be varied. However, the standardized ET_{sz} equation can be used if wind speed are adjusted following guidelines in Appendix B.

USE OF THE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION

Based on an intensive review of reference evapotranspiration calculated for 49 sites throughout the United States (as described in the following section), ASCE-ET found the standardized reference evapotranspiration equation to be reliable and recommends its use for:

- Calculating reference evapotranspiration and, in turn, crop evapotranspiration (ET_c)
- Developing new crop coefficients
- Facilitating transfer of existing crop coefficients

CALCULATING STANDARDIZED REFERENCE CROP EVAPOTRANSPIRATION

This section describes data requirements, equations, and procedures necessary for calculating ET_{sz} on a daily and hourly time step. A daily time step has historically been commonly used in the calculation of ET_{ref}. Selection of the appropriate time step is a function of data availability, climate, the intended application, and user preference.

REQUIRED DATA FOR THE STANDARDIZED REFERENCE EQUATION

The calculation of ET_{sz} requires measurements or estimates for air temperature, humidity, solar radiation, and wind speed. These parameters are considered to be the minimum requirements to estimate ET_{os} and ET_{rs}. Examples of the calculation of ET_{sz} are provided in Appendix C. When humidity, solar radiation or wind speed measurements are not available, substitute values for daily and longer time periods may be estimated using procedures described in Appendix E.

The accuracy of any evapotranspiration calculation depends on the quality of the weather data, which requires good quality control and quality assurance procedures. When possible, weather data should be measured at stations that are located in open, well-watered, vegetated settings (preferably grass). Preferred locations have low growing, well-watered vegetation in the immediate and near vicinity of the weather station (~50 m) and mostly the same or other well-watered vegetation for a few hundred meters beyond that[2]. Suggestions for assessing and improving the integrity of collected weather data are described in Appendix D. Appendix D also provides

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[2] This recommendation is similar to those found in ASAE Engineering Practice EP505 (ASAE 2004).

guidelines for evaluating the weather station site and the possible impact upon the measured meteorological parameters. Suggestions for replacing missing data or data that are of poor quality are presented in Appendix E.

Appendix B provides background on the development of the standardized form of the ASCE equation. The full form of the ASCE-PM equation, which includes explicit terms for aerodynamic and surface resistance, is not required, nor is it recommended, for calculation of ET_{sz}. The full form of the ASCE-PM equation is recommended when ET is measured over grass or alfalfa vegetation having substantially different height than the 0.12 m height defined for the short reference (grass) or 0.50 m height defined for the tall reference (alfalfa). Values for vegetation height are fixed in the standardized equation.

CALCULATIONS REQUIRED FOR DAILY TIME-STEPS

The calculation process for ET_{sz} for daily time steps is presented in this section. Several of the calculations are identical to those required for hourly time steps. Some equations are repeated in the hourly calculation section so as to detail that calculation process completely.

Psychrometric and Atmospheric Variables[3]

Latent Heat of Vaporization (λ)

The value of the latent heat of vaporization, λ, varies only slightly over the ranges of air temperature that occur in agricultural or hydrologic systems. For ET_{sz}, a constant value of $\lambda = 2.45$ MJ kg^{-1} is recommended. The inverse of $\lambda = 2.45$ MJ kg^{-1} is approximately 0.408 kg MJ^{-1}. The density of water (ρ_w) is taken as 1.0 Mg m^{-3} so that the inverse ratio of $\lambda\ \rho_w$ times energy flux in MJ m^{-2} d^{-1} equals 1.0 mm d^{-1}.

Mean Air Temperature (T)

For the standardized method, the mean air temperature, T, for a daily time step is preferred as the mean of the daily maximum and daily minimum air temperatures rather than as the average of hourly temperature measurements to provide for consistency across all data sets.

$$T = \frac{T_{max} + T_{min}}{2} \qquad (2)$$

where:

- T = daily mean air temperature [°C]
- T_{max} = daily maximum air temperature [°C]
- T_{min} = daily minimum air temperature [°C]

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[3] Many of the equations presented here are the same as those reported in ASCE Manual 70 (Jensen et al., 1990) and in FAO-56 (Allen et al., 1998).

Atmospheric Pressure (P)

The mean atmospheric pressure at the weather site is predicted from site elevation using a simplified formulation of the Universal Gas Law[4]:

$$P = 101.3\left(\frac{293 - 0.0065\,z}{293}\right)^{5.26} \tag{3}$$

where:

P = mean atmospheric pressure at station elevation z [kPa], and
z = weather site elevation above mean sea level [m].

Psychrometric Constant (γ)

The standardized application using λ = 2.45 MJ kg^{-1} results in a value for the psychrometric constant, γ, that is proportional to the mean atmospheric pressure:

$$\gamma = 0.000665\ P \tag{4}$$

where P has units of kPa and γ has units of kPa $^{\circ}C^{-1}$.

Note: The variable γ is not the same variable as γ_{psy} used later in Eqs. 9 and 10 for converting psychrometric data (wet bulb and dry bulb temperature) to vapor pressure.

Slope of the Saturation Vapor Pressure-Temperature Curve (Δ)

The slope of the saturation vapor pressure-temperature curve[5], Δ, is computed as:

$$\Delta = \frac{2503\ \exp\left(\dfrac{17.27\,T}{T + 237.3}\right)}{(T + 237.3)^2} \tag{5}$$

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[4] Reference: Burman et al. (1987)

[5] References: Tetens (1930), Murray (1967)

where:

Δ = slope of the saturation vapor pressure-temperature curve [kPa $^{\circ}C^{-1}$], and

T = daily mean air temperature [°C].

Saturation Vapor Pressure (e_s)

The saturation vapor pressure[6] (e_s) represents the capacity of the air to hold water vapor.

For calculation of daily ET_{sz} , e_s is given by:

$$e_s = \frac{e^o(T_{max}) + e^o(T_{min})}{2} \tag{6}$$

where:

$e^o(T)$ = saturation vapor pressure function (Eq. 7) [kPa]

The function to calculate saturation vapor pressure is:

$$e^o(T) = 0.6108 \exp\left(\frac{17.27\,T}{T + 237.3}\right) \tag{7}$$

where vapor pressure is in units of kPa and temperature is in °C.

Actual Vapor Pressure (e_a)

Actual vapor pressure (e_a) is used to represent the water content (humidity) of the air at the weather site. The actual vapor pressure can be measured or it can be calculated from various humidity data, such as measured dew point temperature, wet-bulb and dry-bulb temperature, or relative humidity and air temperature data.

Preferred procedures for calculating e_a

When multiple types of humidity or psychrometric data are available for estimating e_a, the preferences listed in Table 3 are recommended for the calculation method. These recommendations are based on the likelihood that the data will have integrity and that estimates for e_a will be representative. The availability and quality of local data, as well as site conditions, may justify a different order of preference.

Table 3. Preferred method for calculating e_a for daily ET_{sz}

Method No.	Method	Preference Ranking	Equation(s)
1	e_a averaged over the daily period (based on hourly or more frequent measurements of humidity)[a,b]	1	7, 41
2	Measured or computed dew point temperature averaged over the daily period	1	8
3	Wet-bulb and dry-bulb temperature averaged over the daily period	2	7, 9, 10
4	Measured or computed dew point or measured wet-bulb and dry-bulb temperature at 7 or 8 am	2	8 or 7, 9, 10
5	Daily maximum and minimum relative humidity	2	7, 11
6	Daily maximum relative humidity	3	7, 12
7	Daily minimum relative humidity	3	7, 13
8	Daily minimum air temperature (see Appendix E)	4	--
9	Daily mean relative humidity	4	7, 14

[a] In many data sets, e_a may be expressed in terms of an equivalent dew point temperature.

[b] Some data logging systems may measure relative humidity (RH) and T, but calculate e_a or T_{dew} internally for output as averaged values over some time interval. See ASAE (2004) for further detail.

[6] Reference: Jensen et al. (1990) and Tetens (1930)

When humidity and psychrometric data are missing or are of questionable integrity, dew point temperature can be estimated from daily minimum air temperature as described in Appendix E. This estimation process should be verified locally. The assessment of weather data integrity is discussed in Appendix D.

e_a from measured dew point temperature

The dew point temperature (T_{dew}) is the temperature to which the air must cool to reach a state of saturation. For daily calculation time steps, average dew point temperature can be computed by averaging over hourly periods or, for purposes of estimating ET_{sz}, it can be determined by an early morning measurement (generally at 0700 or 0800 hours). The value for e_a is calculated by substituting T_{dew} into Eq. 7 resulting in:

$$e_a = e^o(T_{dew}) = 0.6108 \exp\left[\frac{17.27\ T_{dew}}{T_{dew} + 237.3}\right] \qquad (8)$$

e_a from psychrometric data

The actual vapor pressure can also be determined from the difference between the dry and wet bulb temperatures (i.e., the wet bulb depression) of the air:

$$e_a = e^o(T_{wet}) - \gamma_{psy}(T_{dry} - T_{wet}) \qquad (9)$$

where:

e_a = actual vapor pressure of the air [kPa],
$e°(T_{wet})$ = saturation vapor pressure at the wet bulb temperature [kPa] (Eq. 7),
γ_{psy} = psychrometric constant for the psychrometer [kPa °C^{-1}], and
T_{dry} - T_{wet} = wet bulb depression,
where
T_{dry} = dry bulb temperature and
T_{wet} = the wet bulb temperature [°C] (measured simultaneously).

The psychrometric constant for the psychrometer at the weather measurement site is given by:

$$\gamma_{psy} = a_{psy} P \tag{10}$$

where

a_{psy} = coefficient depending on the type of ventilation of the wet bulb [°C^{-1}], and

P = mean atmospheric pressure [kPa].

The coefficient a_{psy} depends primarily on the design of the psychrometer and on the rate of ventilation around the wet bulb. The following values are often used[7]:

a_{psy} = 0.000662 for ventilated (Asmann type) psychrometers having air movement between 2 and 10 m s^{-1} for $T_{wet} \geq 0$ and 0.000594 for $T_{wet} < 0$,

= 0.000800 for naturally ventilated psychrometers having air movement of about 1 m s^{-1}, and

= 0.001200 for non-ventilated psychrometers installed in glass or plastic greenhouses.

Generally, the wet-bulb and dry-bulb temperature data are measured once during the day.

<u>e_a from relative humidity data</u>

The actual vapor pressure of air can be calculated from relative humidity (RH) and the corresponding air temperature. When using RH data, it is essential that the RH and air temperature data are "paired," i.e., that they represent the same time of day or time period and that they are taken at the weather measurement site. For daily data, daily maximum relative humidity (RH_{max}) can be paired with T_{min}, which will both

[7] Allen et al., (1998).

occur, generally, during early morning. Daily minimum relative humidity (RH_{min}) is paired with T_{max}.

Depending on the availability of the RH data, the following equations apply, with preference of method listed in Table 3:

- Daily e_a from RH_{max} and RH_{min}.

$$e_a = \frac{e^o(T_{min})\dfrac{RH_{max}}{100} + e^o(T_{max})\dfrac{RH_{min}}{100}}{2} \tag{11}$$

where:

e_a = actual vapor pressure [kPa],
$e°(T_{min})$ = saturation vapor pressure at daily minimum temperature [kPa],
$e°(T_{max})$ = saturation vapor pressure at daily maximum temperature [kPa],
RH_{max} = daily maximum relative humidity [%], and
RH_{min} = daily minimum relative humidity [%].

When computing the average daily ET_{sz} during a week, a ten-day period or a month, RH_{max} and RH_{min} are obtained by averaging daily RH_{max} or RH_{min} values.

- Daily e_a from RH_{max}

Older styles of electronic relative humidity sensors, for example those manufactured before about 1990, often experienced difficulty in accurately measuring RH when at low levels. When using equipment where errors in estimating RH_{min} may be large, or when integrity of the RH data is doubtful, the actual vapor pressure can be computed from RH_{max}:

$$e_a = e^o(T_{min})\frac{RH_{max}}{100} \tag{12}$$

When accuracy of RH data is in doubt, error in RH_{max} causes smaller error in e_a than error in RH_{min}, due to the smaller value for the multiplier $e^o(T_{min})$ as compared to $e^o(T_{max})$. In addition, RH_{max} data are generally easier to assess for accuracy than is RH_{min}. The value of RH_{max} generally exceeds 90% and approaches 100% in well-watered settings such as within irrigation projects and in sub-humid and humid climates. This proximity to 100% serves as a first check on reasonableness, representativeness, and integrity of the data. Exceptions to this trend are where substantial advection of dry or warm air from dry regions outside the area occurs during nighttime, including, but not limited to,some desert areas of New Mexico, Arizona and California.

- Daily e_a from RH_{min}

Sometimes, only high quality estimates of daily RH_{min} are available and must be used to predict e_a:

$$e_a = e^o\left(T_{max}\right)\frac{RH_{min}}{100} \qquad (13)$$

However, estimates using Eq. 13 may be less desirable than estimates using Eq. 11 or 12, due to greater impact of error in RH_{min} on e_a, as discussed previously. In addition, it is more difficult to assess the integrity of RH_{min} data (see Appendix D):

- Daily e_a from RH_{mean}

In the absence of RH_{max} and RH_{min} data, but where daily RH_{mean} data are available, the actual vapor pressure may be estimated as:

$$e_a = \frac{RH_{mean}}{100} e^o\left(T_{mean}\right) \qquad (14)$$

where RH_{mean} is the mean daily relative humidity, generally defined as the average between RH_{max} and RH_{min} and T_{mean} is mean daily air temperature, defined in Eq. 2

Eq. 14 is less desirable than Eqs. 12 or 13 because the $e^o(T)$ relationship is nonlinear. Eq. 14 produces estimates of e_a that are closer to those by Eq. 11 and to daily average e_a computed from hourly values than is the use of alternative forms of Eq. 14, such as $e_a = RH_{mean}/100\ [e^o(T_{max})+e^o(T_{min})]/2$ described in Allen et al., (1998) or as $e_a = RH_{mean}/100\ [1/(50/e^o(T_{max}) + 50/e^o(T_{min}))]$ described in Smith et al., (1991). These latter two methods are not recommended in the standardized procedure.

Net Radiation (R_n)

Net radiation (R_n) is the net amount of radiant energy available at a vegetation or soil surface for evaporating water, heating the air, or heating the surface. R_n includes both short and long wave radiation components[8]:

$$R_n = R_{ns} - R_{nl} \tag{15}$$

where:

R_{ns} = net short-wave radiation, [MJ m^{-2} d^{-1}] (defined as being positive downwards and negative upwards),

R_{nl} = net outgoing long-wave radiation, [MJ m^{-2} d^{-1}] (defined as being positive upwards and negative downwards),

R_{ns} and R_{nl} are generally positive or zero in value.

Net radiation is difficult to measure because net radiometers are problematic to maintain and calibrate. There is good likelihood of systematic biases in R_n measurements. Therefore, R_n is often predicted from observed short wave (solar) radiation, vapor pressure, and air temperature. This prediction is routine and generally highly accurate. If R_n is measured, then care and attention must be given to the calibration of the radiometer, the surface over which it is located, maintenance of

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[8] Reference: Brutsaert (1982), Jensen et al., (1990), Wright (1982), Doorenbos and Pruitt (1975,1977), Allen et al., (1998).

the sensor domes, and level of the instrument. The condition of the vegetation surface is as important as the sensor. For purposes of calculating ET_{sz}, the measurement surface for R_n is generally assumed to be clipped grass or alfalfa at full cover.

Net Solar or Net Short-Wave Radiation (R_{ns})

Net short-wave radiation resulting from the balance between incoming and reflected solar radiation is given by:

$$R_{ns} = R_s - \alpha R_s = (1-\alpha) R_s \tag{16}$$

where:

R_{ns} = net solar or short-wave radiation [MJ m^{-2} d^{-1}],
α = albedo or canopy reflection coefficient, is fixed at 0.23 for the standardized short and tall reference surfaces [dimensionless], and
R_s = incoming solar radiation [MJ m^{-2} d^{-1}].

The calculation of ET_{sz} uses the constant value of 0.23 for albedo for daily and hourly periods. It is recognized that albedo varies somewhat with time of day and with time of season and latitude due to change in sun angle. However, because the solar intensity is less during these periods, the error introduced in fixing albedo at 0.23 is relatively small (Allen et al., 1994b). Users may elect to use a different prediction for albedo, however, it is essential to ascertain the validity and accuracy of an alternative method using good measurements of incoming and reflected solar radiation. Some types of pyranometers are invalid for measuring reflected radiation due to the difference in spectral response between the instrument and reflecting surface. Predictions of R_n made using an alternate method for albedo (i.e., other than 0.23) may not agree with those made using the ASCE standardized procedure.

Net Long-Wave Radiation (R_{nl})

There are several variations and coefficients developed for predicting the net long wave component of total net radiation. The standardized ASCE procedure is the same as that adopted by FAO-56 and is based on the Brunt (1932, 1952) approach for predicting net emissivity. If users intend to utilize a different approach for calculating R_{nl}, it is essential to ascertain the validity and accuracy of their R_n method using net radiometers in excellent condition and that are calibrated to some dependable and recognized standard. In all situations, users should compare measured R_n or R_n computed using an alternative method with R_n calculated using the standardized procedure. Substantial variation (more than 5 %) should give cause for concern and should indicate the need to reconcile or justify the differences.

R_{nl}, net long-wave radiation, is the difference between upward long-wave radiation from the standardized surface (R_{lu}) and downward long-wave radiation from the sky (R_{ld}), so that $R_{nl} = R_{lu} - R_{ld}$. The following calculation for daily R_{nl} follows the method of Brunt (1932, 1952) of using vapor pressure to predict net emissivity:

$$R_{nl} = \sigma \ f_{cd} \left(0.34 - 0.14\sqrt{e_a}\right)\left[\frac{T_{K\ max}^4 + T_{K\ min}^4}{2}\right] \tag{17}$$

where:

R_{nl}	=	net long-wave radiation [MJ m^{-2} d^{-1}],
σ	=	Stefan-Boltzmann constant [4.901 x 10^{-9} MJ K^{-4} m^{-2} d^{-1}],
f_{cd}	=	cloudiness function [dimensionless] (limited to $0.05 \le f_{cd} \le 1.0$),
e_a	=	actual vapor pressure [kPa],
$T_{K\ max}$	=	maximum absolute temperature during the 24-hour period [K] (K = °C + 273.16),
$T_{K\ min}$	=	minimum absolute temperature during the 24-hour period [K] (K = °C + 273.16).

The superscripts “4” in Eq. 17 indicate the need to raise the air temperature, expressed in Kelvin units, to the power of 4. For daily and monthly calculation timesteps, f_{cd} is calculated as[9]:

$$f_{cd} = 1.35 \frac{R_s}{R_{so}} - 0.35 \qquad (18)$$

where:

R_s/R_{so} = relative solar radiation (limited to $0.3 \le R_s/R_{so} \le 1.0$),
R_s = measured or calculated solar radiation [MJ m^{-2} d^{-1}], and
R_{so} = calculated clear-sky radiation [MJ m^{-2} d^{-1}].

The ratio R_s/R_{so} in Eq. 18 represents relative cloudiness and is limited to $0.3 < R_s/R_{so} \le 1.0$ so that f_{cd} has limits of $0.05 \le f_{cd} \le 1.0$.

Clear-Sky Solar Radiation (R_{so})

Clear-sky solar radiation (R_{so}) is used in the calculation of net radiation (R_n). Clear-sky solar radiation is defined as the amount of solar radiation (R_s) that would be received at the weather measurement site under conditions of clear-sky (i.e., cloud-free). The ratio of R_s to R_{so} in the equation for R_n is used to characterize the impact of cloud-cover on the downward emission of thermal radiation to the earth’s surface. Daily R_{so} is a function of the time of year and latitude. R_{so} is also impacted by station elevation (affecting atmospheric thickness and transmissivity), the amount of precipitable water in the atmosphere (affecting the absorption of some short-wave radiation), and the amount of dust or aerosols in the air.

Extraterrestrial radiation (R_a), as defined in Eq. 21, can be used as a means for determining a theoretical R_{so}envelope as illustrated in Figure 1. The envelope can be expressed in tabular form or as an equation. In this section, a simple procedure[10] is

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[9] Jensen et al., (1990); Allen et al., (1998)

[10] Reference: Allen (1996)

demonstrated for estimating R_{so} for purposes of predicting net radiation. A more involved procedure, useful for evaluating R_s data integrity, is described in Appendix D. The clear sky envelope can alternatively be developed using measured R_s from a period of one year or longer. The measured data should be confirmed for accuracy, including sensor calibration and maintenance (levelness and cleanliness). When measured R_s data are used to define an R_{so} envelope for a location, the resulting envelope should be compared with a theoretically derived envelope to confirm that there are no substantial differences in shape or magnitude. In general, the theoretically derived curve (Figure 1) is recommended.

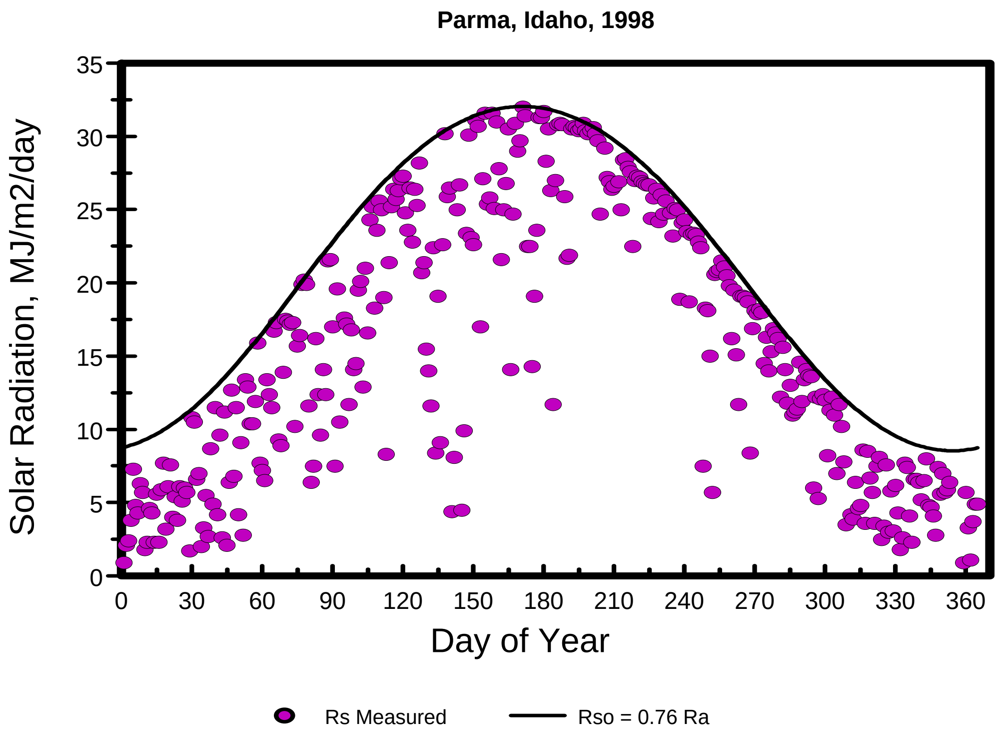

Figure 1. Daily R_s at Parma, Idaho during 1998 (elevation 703 m, Lat. 43.8^{o}) and R_{so} envelope from Eq. 19.

When a dependable, locally calibrated procedure for determining R_{SO} is not available, R_{so}, for purposes of calculating R_n, can be computed as:

$$R_{so} = \left(0.75 + 2 \times 10^{-5} z\right) R_a \tag{19}$$

where:

z = station elevation above sea level [m].

Eq. 19 predicts progressively higher levels of clear sky radiation with increasing elevation, and is the basis for the "0.76" factor for the R_{so} curve drawn in Figure 1. Elevation serves as a surrogate for total air mass and atmospheric transmissivity above the measurement site.

When dependable, locally calibrated values are available for applying the "Angstrom" formula (see Eq. A.44), the clear sky radiation can be computed as:

$$R_{so} = K_{ab} R_a \tag{20}$$

where:

R_{so} = clear-sky solar radiation [MJ m^{-2} d^{-1}],

R_a = extraterrestrial radiation [MJ m^{-2} d^{-1}],

K_{ab} = coefficient that can be derived from the a_s and b_s coefficients of the Angstrom formula, where $K_{ab} = a_s + b_s$, and where K_{ab} represents the fraction of extraterrestrial radiation reaching the earth on clear-sky days,

a_s = constant expressing the fraction of extraterrestrial radiation reaching the earth's surface on completely overcast days (see Eq. E.2 in Appendix E), and

b_s = constant expressing the additional fraction of extraterrestrial radiation reaching the earth's surface on a clear day (see Eq. E.2 in Appendix E).

Eqs. 19 or 20 are generally adequate for use in estimating R_{so} in Eq. 18 when predicting net radiation, R_n. More complex estimates for R_{so}, which include impacts of turbidity and water vapor on radiation absorption, can be used for assessing integrity of solar radiation data and are discussed in Appendix D. The difference in ET_{rs} or ET_{os} resulting from the use of Eq. 19 or 20, as opposed to the more complicated and accurate R_{so} equations D.1 – D.5 of Appendix D, will be generally less than a few percent over an annual period.

Extraterrestrial Radiation for 24-Hour Periods (R_a)[11]

Extraterrestrial radiation, R_a, defined as the short-wave solar radiation in the absence of an atmosphere, is a well-behaved function of the day of the year, time of day, and latitude. It is needed for calculating R_{so}, which is in turn used in calculating R_n. For daily (24-hour) periods, R_a can be estimated from the solar constant, the solar declination, and the day of the year:

$$R_a = \frac{24}{\pi} G_{sc}\, d_r \left[\omega_s \sin(\varphi)\sin(\delta) + \cos(\varphi)\cos(\delta)\sin(\omega_s)\right] \qquad (21)$$

where:

R_a	= extraterrestrial radiation [MJ m^{-2} d^{-1}],
G_{sc}	= solar constant [4.92 MJ m^{-2} h^{-1}],
d_r	= inverse relative distance factor (squared) for the earth-sun [unitless],
ω_s	= sunset hour angle [radians],
φ	= latitude [radians], and
δ	= solar declination [radians].

The latitude, φ, is positive for the Northern Hemisphere and negative for the Southern Hemisphere. The conversion from decimal degrees to radians is given by:

$$\text{Radians} = \frac{\pi}{180}\left(\text{decimal degrees}\right) \qquad (22)$$

and d_r and δ are calculated as:

$$d_r = 1 + 0.033\cos\left(\frac{2\pi}{365} J\right) \qquad (23)$$

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[11] Reference: Duffie and Beckman (1980).

$$\delta = 0.409 \sin\left(\frac{2\,\pi}{365} J - 1.39\right) \tag{24}$$

where:

J is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December). The constant 365 in Eqs. 23 and 24 is held at 365 even during a leap year. J can be calculated as[12]:

$$J = D_M - 32 + \mathrm{Int}\left(275\frac{M}{9}\right) + 2\ \mathrm{Int}\left(\frac{3}{M+1}\right) + \mathrm{Int}\left(\frac{M}{100} - \frac{\mathrm{Mod}(Y,4)}{4} + 0.975\right) \tag{25}$$

where:

D_M	= the day of the month (1-31),
M	= the number of the month (1-12), and
Y	= the number of the year (for example 1996 or 96).

The "Int" function in Eq. 25 finds the integer number of the argument in parentheses by rounding downward. The "Mod(Y,4)" function finds the modulus (remainder) of the quotient Y/4.

For monthly periods, the day of the year at the middle of the month (J_{month}) is approximately:

$$J_{month} = \mathrm{Int}(30.4\ M - 15) \tag{26}$$

The sunset hour angle, ω_s, is given by:

$$\omega_s = \arccos\left[-\tan(\varphi)\tan(\delta)\right] \tag{27}$$

The "arccos" function is the arc-cosine function and represents the inverse of the cosine. This function is not available in all computer languages, so that ω_s can alternatively be computed using the arc-tangent (inverse tangent) function:

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[12] Reference: Allen (2000).

$$\omega_s = \frac{\pi}{2} - \arctan\left[\frac{-\tan(\varphi)\tan(\delta)}{X^{0.5}}\right] \tag{28}$$

where:

$$X = 1 - [\tan(\varphi)]^2 [\tan(\delta)]^2 \tag{29}$$

$$\text{and } X = 0.00001 \text{ if } X \le 0$$

Soil Heat Flux Density (G)

Soil heat flux density is the thermal energy utilized to heat the soil. G is positive when the soil is warming and negative when the soil is cooling.

For Daily Periods

The magnitude of the daily, weekly or ten-day soil heat flux density, G, beneath a fully vegetated grass or alfalfa reference surface is relatively small in comparison with R_n. Therefore, it is ignored so that:

$$G_{day} = 0 \tag{30}$$

where:

G_{day} = daily soil heat flux density [MJ m^{-2} d^{-1}].

For Monthly Periods

Over a monthly period, G for the soil profile can be significant. Assuming a constant soil heat capacity of 2.0 MJ m^{-3} °C^{-1} and an effectively warmed soil depth of 2 m, G for monthly periods in MJ m^{-2} d^{-1} is estimated from the change in mean monthly air temperature as:

$$G_{month,i} = 0.07\,(T_{month,i+1} - T_{month,i-1}) \tag{31}$$

or, if $T_{month,i+1}$ is unknown:

$$G_{month,i} = 0.14\,(T_{month,i} - T_{month,i-1}) \tag{32}$$

where:

$T_{month,I}$ = mean air temperature of month i [°C],
$T_{month,i-1}$ = mean air temperature of previous month [°C], and
$T_{month,i+1}$ = mean air temperature of next month [°C].

Wind Profile Relationship

Wind speed varies with height above the ground surface. For the calculation of ET_{sz}, wind speed at 2 meters above the surface is required, therefore, wind measured at other heights must be adjusted. To adjust wind speed data to the 2-m height, Eq. 33 should be used for measurements taken above a short grass (or similar) surface, based on the full logarithmic wind speed profile equation B.14 given in Appendix B:

$$u_2 = u_z \frac{4.87}{\ln(67.8\,z_w - 5.42)} \tag{33}$$

where:

u_2 = wind speed at 2 m above ground surface [m s^{-1}],
u_z = measured wind speed at z_w m above ground surface [m s^{-1}], and
z_w = height of wind measurement above ground surface [m].

For wind measurements above surfaces other than clipped grass, the user should apply the full logarithmic equation B.14. A special application of Eq. B.14 is given in Appendix B for wind measured above alfalfa or similar vegetation having about 0.5 m height. It is noted that wind speed data collected at heights above 2 m are acceptable for use in the standardized equations following adjustment to 2 m, and may be preferred if vegetation adjacent to the station commonly exceeds 0.5 m. Measurement at a greater height, for example 3m, reduces the influence of the taller vegetation surrounding the weather measurement site.

CALCULATIONS REQUIRED FOR HOURLY TIME-STEPS

Many weather data networks collect and summarize hourly data that allow the user to calculate ET_{sz} for hourly periods. This capability is important where significant shifts in wind and humidity occur hourly. The calculation process for hourly time steps is analogous to that for daily calculations. The hourly equations can be used for shorter time periods, using fractional hours as the time parameter, but care must be taken to multiply the resultant ET rate in mm/h by the fractional hour. For example, if 30-minute data are used, one would input radiation in units of MJ m^{-2} h^{-1}. Then the output, in mm h^{-1}, would need to be multiplied by 0.5 h to arrive at the ET for the 30-minute period.

<u>Psychrometric and Atmospheric Variables</u>[13]

Latent Heat of Vaporization (λ)

The value of the latent heat of vaporization (λ), varies only slightly over the ranges of air temperature that occur in agricultural or hydrologic systems. For ET_{sz}, a constant value of λ = 2.45 MJ kg^{-1} is selected. The inverse of λ = 2.45 MJ kg^{-1} is approximately 0.408 kg MJ^{-1}. The density of water (ρ_w) is taken as 1.0 Mg m^{-3} so that the inverse ratio of $\lambda\ \rho_w$ times energy flux in MJ m^{-2} h^{-1} equals 1.0 mm h^{-1}.

Mean Air Temperature (T)

For hourly periods, the mean air temperature, T, represents an average over the period.

[13] Many of the equations presented here are the same as those reported in ASCE Manual 70 (Jensen et al., 1990) and used in FAO-56 (Allen et al., 1998).

Atmospheric Pressure (P)

The mean atmospheric pressure at the weather site is predicted from site elevation using a simplified formulation of the Universal Gas Law[14]:

$$P = 101.3\left(\frac{293 - 0.0065\,z}{293}\right)^{5.26} \tag{34}$$

where:

P = mean atmospheric pressure at station elevation z [kPa], and
z = weather site elevation above mean sea level [m].

Psychrometric Constant (γ)

The standardized application using λ = 2.45 MJ kg^{-1} results in a value for the psychrometric constant, γ, that is proportional to the mean atmospheric pressure:

$$\gamma = 0.000665\ P \tag{35}$$

where P has units of kPa and γ has units of kPa $^{\circ}$C^{-1}.

The variable γ is not the same variable as γ_{psy} used later in Eqs. 39 and 40 for converting psychrometric data (wet bulb and dry bulb temperature) to vapor pressure.

Slope of the Saturation Vapor Pressure-Temperature Curve (Δ)

The slope of the saturation vapor pressure-temperature curve[15], Δ, is computed as:

$$\Delta = \frac{2503\ \exp\left(\frac{17.27\ T}{T + 237.3}\right)}{(T + 237.3)^2} \tag{36}$$

where:

[14] Reference: Burman et al. (1987)

[15] References: Tetens (1930), Murray (1967)

Δ	= slope of the saturation vapor pressure-temperature curve [kPa °C^{-1}], and
T	= mean air temperature [°C].

Saturation Vapor Pressure (e_s)

The saturation vapor pressure[16], e_s, represents the capacity of the air to hold water vapor.

For calculation of hourly ET_{sz} , e_s is given by:

$$e_s = e^o(T) = 0.6108 \exp\left(\frac{17.27\,T}{T + 237.3}\right) \tag{37}$$

where vapor pressure is in units of kPa and T is mean air temperature during the hourly period in °C. $e^o(T)$ is the saturation vapor pressure function.

Actual Vapor Pressure (e_a)

Actual vapor pressure (e_a) is used to represent the water content (humidity) of the air at the weather site. The actual vapor pressure can be measured or it can be calculated from various humidity data, such as measured dew point temperature, wet-bulb and dry-bulb temperature, or relative humidity and air temperature data.

Preferred procedures for calculating e_a

When multiple types of humidity or psychrometric data are available for estimating e_a, the preferences listed in Table 4 are recommended for calculation method. These recommendations are based on the likelihood that the data will have integrity and that estimates for e_a will be representative of the reference ET environment. The availability and quality of local data may justify a different order of preference.

[16] Reference: Jensen et al. (1990) and Tetens (1930)

Table 4. Preferred method for calculating e_a for ET_{sz} for hourly periods

Method No.	Method	Preference Ranking	Equation(s)
1	e_a averaged over period[a,b]	1	--
2	Measured or calculated dew point temperature averaged over period	1	38
3	Average RH and T for the hour	1	37, 41
4	Wet-bulb and dry-bulb temperature	2	38, 39, 40
5	Daily minimum air temperature (see Appendix E)	3	--

[a] In many data sets, e_a may be expressed in terms of an equivalent dew point temperature.
[b] Some data logging systems may measure RH and T, but calculate e_a or T_{dew} internally for output as averaged values over some time interval.

When humidity and psychrometric data are missing or are of questionable integrity, dew point temperature can be estimated from daily minimum air temperature as described in Appendix E. This estimation procedure should be verified locally. The assessment of weather data integrity is discussed in Appendix D.

e_a from measured dew point temperature

The dew point temperature, T_{dew}, is the temperature to which the air must be cooled to reach a state of saturation. The value for e_a is calculated by substituting T_{dew} into Eq. 37 resulting in:

$$e_a = e^o(T_{dew}) = 0.6108 \exp\left[\frac{17.27\ T_{dew}}{T_{dew} + 237.3}\right] \tag{38}$$

e_a from psychrometric data

The actual vapor pressure can also be determined from the difference between the dry and wet bulb temperatures (i.e., the wet bulb depression) of the air:

$$e_a = e^o(T_{wet}) - \gamma_{psy}\left(T_{dry} - T_{wet}\right) \tag{39}$$

where:

e_a = actual vapor pressure of the air [kPa],
$e°(T_{wet})$ = saturation vapor pressure at the wet bulb temperature [kPa] (Eq. 37),
γ_{psy} = psychrometric constant for the psychrometer [kPa °C^{-1}], and
T_{dry}-T_{wet} = wet bulb depression, where T_{dry} is the dry bulb temperature and T_{wet} is the wet bulb temperature [°C] (measured simultaneously).

The psychrometric constant for the psychrometer at the weather measurement site is given by:

$$\gamma_{psy} = a_{psy} P \tag{40}$$

where:

a_{psy} = coefficient depending on the type of ventilation of the wet bulb [°C^{-1}], and
P = mean atmospheric pressure [kPa].

The coefficient a_{psy} depends primarily on the design of the psychrometer and on the rate of ventilation around the wet bulb. The following values are often used[17]:

a_{psy} = 0.000662 for ventilated (Asmann type) psychrometers having air movement between 2 and 10 m s^{-1} for $T_{wet} \geq 0$ and 0.000594 for $T_{wet} < 0$,
= 0.000800 for naturally ventilated psychrometers with air movement of about 1 m s^{-1}), and
= 0.001200 for non-ventilated psychrometers installed in glass or plastic greenhouses (List, 1984).

e_a from relative humidity data

The actual vapor pressure of air for hourly periods can be calculated from relative humidity (RH) and saturation vapor pressure at the corresponding air temperature (from Eq. 37):

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[17] Allen et al., (1998).

$$e_a = \frac{RH}{100} e^o(T) \qquad (41)$$

where:

RH = mean relative humidity for the hourly period, %, and
T = mean air temperature for the hourly period, ^oC.

Net Radiation (R_n)

Net radiation (R_n) is the net amount of radiant energy available at the vegetation or soil surface for evaporating water, heating the air, or heating the surface. R_n includes both short and long wave radiation components [18]:

$$R_n = R_{ns} - R_{nl} \qquad (42)$$

where:

R_{ns} = net shortwave radiation, [MJ m^{-2} h^{-1}] (defined as being positive downwards and negative upwards),
R_{nl} = net outgoing long-wave radiation, [MJ m^{-2} h^{-1}] (defined as being positive, upwards and negative downwards),

R_{ns} and R_{nl} are generally positive or zero in value.

Net radiation is difficult to measure because net radiometers are problematic to maintain and calibrate. There is good likelihood of systematic biases in R_n measurements. Therefore, R_n is often predicted from observed short wave (solar) radiation, vapor pressure, and air temperature. This prediction is routine and generally highly accurate. If R_n is measured, care and attention must be given to the calibration of the radiometer, the surface over which it is located, maintenance of the sensor domes, and level of the instrument. The condition of the vegetation surface is

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[18] Reference: Brutsaeart (1982), Jensen et al., (1990), Wright (1982), Doorenbos and Pruitt, (1975, 1977), Allen et al., (1998).

as important as the sensor. For purposes of calculating ET_{sz} , the measurement surface for R_n is generally assumed to be clipped grass or alfalfa at full cover.

Net Solar or Net Short-Wave Radiation (R_{ns})

Net short-wave radiation resulting from the balance between incoming and reflected solar radiation is given by:

$$R_{ns} = R_s - \alpha R_s = (1 - \alpha) R_s \tag{43}$$

where:

R_{ns}	= net solar or short-wave radiation [MJ m^{-2} h^{-1}],
α	= albedo or canopy reflection coefficient, is fixed at 0.23 for the standardized short and tall reference surfaces [dimensionless], and
R_s	= the incoming solar radiation [MJ m^{-2} h^{-1}].

The calculation of ET_{sz} uses the constant value of 0.23 for albedo for daily and hourly periods. It is recognized that albedo varies somewhat with time of day and with time of season and latitude due to change in sun angle. However, because the solar intensity is less during these periods, the error introduced in fixing albedo at 0.23 is relatively small (Allen et al., 1994b). Users may elect to use a different prediction for albedo, however, they are strongly encouraged to ascertain the validity and accuracy of an alternative method using good measurements of incoming and reflected solar radiation. Some types of pyranometers are invalid for measuring reflected radiation due to the difference in spectral response between the instrument and reflecting surface. Predictions of R_n made using an alternate method for albedo (i.e., other than 0.23) may not agree with those made using the ASCE standardized procedure.

Net Long-Wave Radiation (R_{nl})

There are several variations and coefficients developed for predicting the net long wave component of total net radiation. The standardized ASCE procedure is the same as that adopted by FAO-56 and is based on the Brunt (1932, 1952) approach for predicting net surface emissivity. If users intend to utilize a different approach for calculating R_n, it is essential to ascertain the validity and accuracy of their method using net radiometers in excellent condition and that are calibrated to some dependable and recognized standard. In all situations, users should compare measured R_n or R_n computed using an alternative method with R_n calculated using the standardized procedure. Substantial variation (more than 5 %) should give cause for concern and should indicate the need to reconcile or justify the differences.

R_{nl} is the difference between long-wave radiation radiated upward from the standardized surface (R_{lu}) and long-wave radiation radiated downward from the atmosphere (R_{ld}), so that $R_{nl} = R_{lu} - R_{ld}$. The following calculation for R_{nl} is the method introduced by Brunt (1932, 1952) that uses near surface vapor pressure to predict net surface emissivity:

$$R_{nl} = \sigma\, f_{cd} \left(0.34 - 0.14\sqrt{e_a}\right) T_{K\,hr}^4 \qquad (44)$$

where

R_{nl} = net outgoing long-wave radiation [MJ m^{-2} h^{-1}],
σ = Stefan-Boltzmann constant [2.042 x 10^{-10} MJ K^{-4} m^{-2} h^{-1}],
f_{cd} = cloudiness function [dimensionless] (limited to $0.05 \le f_{cd} \le 1.0$),
e_a = actual vapor pressure [kPa],
$T_{K\,hr}$ = mean absolute temperature during the hourly period [K] (K = °C + 273.16).

The superscript "4" in Eq. 44 indicates the need to raise the air temperature, expressed in Kelvin units, to the power of 4. For periods <u>during daytime</u> when the sun is more than about 15° above the horizon (see procedures below), f_{cd} is calculated as:

$$f_{cd} = 1.35\frac{R_s}{R_{so}} - 0.35 \qquad (45)$$

where:

R_s/R_{so} = relative solar radiation (limited to $0.3 \le R_s/R_{so} \le 1.0$),
R_s = measured or calculated solar radiation [MJ m^{-2} h^{-1}], and
R_{so} = calculated clear-sky radiation [MJ m^{-2} h^{-1}].

The ratio R_s/R_{so} in Eq. 45 represents relative cloudiness and is limited to $0.3 < R_s/R_{so} \le 1.0$ so that f_{cd} has limits of $0.05 \le f_{cd} \le 1.0$.

During nighttime, R_{so}, by definition, equals 0, and Eq. 45 is undefined. Furthermore, even small out of level of a pyranometer or imperfect correction for cosine error of the instrument (required for accurate measurement at low sun angle) can cause substantial deviation in the value for R_s/R_{so} when the sun is near the horizon (i.e., when R_{so} is small). Therefore, f_{cd} during periods of low sun angle and nighttime is defined using f_{cd} from a prior period having sufficient sun angle.

When sun angle above the horizon (β)[19] at the midpoint of the hourly or shorter time period is less than 0.3 radians (~17°), then:

$$f_{cd} = f_{cd\ \beta>0.3} \tag{46}$$

where:

$f_{cd\ \beta>0.3}$ = cloudiness function for the time period prior to when sun angle β (in the afternoon or evening) falls below 0.3 radians [dimensionless].

Note that if the time period is shorter than one hour, f_{cd} from several periods can be averaged into $f_{cd\ \beta>0.3}$ to obtain a representative average value. In mountain valleys where the sun may set near or above 0.3 radians (~17°), the user should increase the sun angle at which $f_{cd\ \beta>0.3}$ is computed and imposed. For example, for a location where mountain peaks are 20° above the horizon, a period should be selected for computing $f_{cd\ \beta>0.3}$ where the sun angle at the end of the time period is 25 to 30°

[19] The sun angle β is defined as the angle of a line from the measurement site to the center of the sun's disk relative to a line from the measurement site to directly below the sun and tangent to the earth's surface. This definition assumes a flat surface.

above the horizon. The same adjustment is necessary when deciding when to resume computation of f_{cd} during morning hours when mountains lie to the east.

Only one value for $f_{cd\ \beta>0.3}$ is calculated per day for use during dusk, nighttime and dawn periods. That value for $f_{cd\ \beta>0.3}$ is hen applied to the time period when β at the midpoint of the period first falls below 0.3 radians (~15°) and to all subsequent periods until after sunrise when β again exceeds 0.3 radians. Computation of β is given in Eq. 62 in the following section.

Equations 45 – 46 will not apply at latitudes and times of the year when there are no hourly (or shorter) periods having sun angle of 0.3 radians or greater. These situations occur for latitudes at 50° for about one month per year (in winter), for latitudes at 60° for about 5 months per year, and for latitudes at 70° for about 7 months per year. At extreme latitudes, some fall and winter months have little or no daylight. Under these conditions, the application can average $f_{cd\ \beta>0.3}$ from fewer time periods or, in the absence of any daylight, can assume a ratio of R_s/R_{so} ranging from 0.3 for complete cloud cover to 1.0 for no cloud cover. Under these extreme conditions, the prediction of R_n during nighttime and low sun angle is only approximate.

The application of Eq. 46 presumes that cloudiness during periods of low sun angle and nighttime is similar to that during late afternoon or early evening. This is generally a reasonable assumption and is commensurate with the relative simplicity and moderate accuracy of Eq. 45. Some applications may wish to split the nighttime period into two halves, with the first half using $f_{cd\ \beta>0.3}$ computed from late afternoon or early evening and the second half using $f_{cd\ \beta>0.3}$ computed from R_s measured during the following morning (for the period when β is first > 0.3 radians). However, this additional computation requires looking ahead within a data set and will generally not add accuracy to the computations, since the timing of any shift in cloudiness during nighttime is unknown and due to the general, approximate accuracy of the f_{cd} function (Eq. 45).

Clear-sky solar radiation

Clear-sky solar radiation, R_{so}, is used in the calculation of net radiation, R_n. Clear-sky solar radiation is defined as the amount of solar radiation, R_s, that would be received at the weather measurement site under conditions of clear-sky (i.e., cloud-free). The ratio of R_s to R_{so} in the equation for R_n is used to characterize the impact of cloud-cover on the downward emission of thermal radiation to the earth's surface. The value for R_{so} is a function of the time of year and latitude, and, in addition, the time of day for hourly calculation periods. These parameters affect the potential incoming solar radiation from the sun. Clear-sky solar radiation is also impacted by the station elevation (affecting atmospheric thickness and transmissivity), the amount of precipitable water in the atmosphere (affecting the absorption of some short wave radiation), and the amount of dust or aerosols in the air.

A daily R_{so} "envelope" was developed earlier in Figure 1 and compared to measured R_s. For purposes of calculating R_n, hourly R_{so} can be calculated using the following simple approach:

$$R_{so} = \left(0.75 + 2 \times 10^{-5} z\right) R_a \tag{47}$$

where:

R_{so}	= clear-sky solar radiation [MJ m^{-2} h^{-1}],
z	= station elevation above sea level [m], and
R_a	= extraterrestrial radiation [MJ m^{-2} h^{-1}].

Equation 47 predicts progressively higher levels of clear sky radiation with increasing elevation. Elevation serves as a surrogate for total air mass and atmospheric transmissivity above the measurement site. Equation 47 is generally adequate for use in predicting the ratio R_s/R_{so} when calculating net radiation, R_n. Other more complex estimates for R_{so}, which include turbidity and water vapor effects as well as impact of sun angle are discussed in Appendix D. Those equations are recommended for assessing integrity of solar radiation data and may provide improved estimates for R_{so} for calculating R_n. The impact on ET_{SZ} of using the equations in Appendix D for R_{so} rather than Eq. 47 will generally be less than a few percent across a day and over an annual period.

Figure 2 illustrates a comparison of measured hourly solar radiation with R_{so} computed using Eq. 47 and using the more complicated, but generally more accurate, method presented as Eq. D.1-D.5 of Appendix D. Data from two days in late June are plotted. June 19 had some morning and mid-day cloudiness. June 20 was a cloud-free day. The R_s data from June 20 compare relatively well with both R_{so} methods throughout the day. The measured data plot slightly higher than either R_{so} estimate at mid-day, with the more complicated R_{so} method from Appendix D having better agreement than Eq. 48. Measured R_s exceeded the R_{so} curves for the 1100 reading on June 19 because of reflection from clouds near the weather site. In general, the solar radiation data appear to be of excellent quality and calibration.

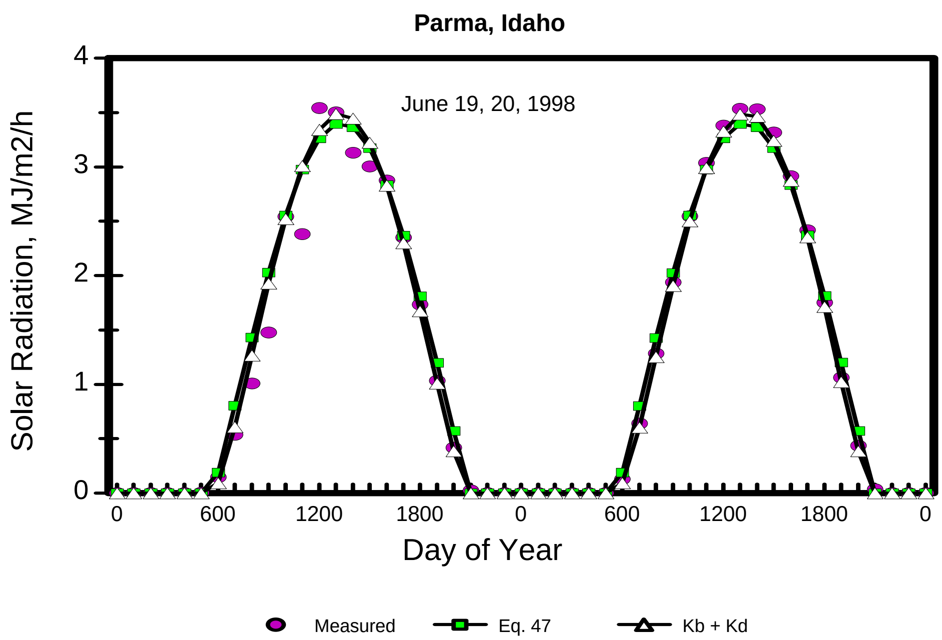

Figure 2. Measured and calculated hourly R_{so} for two days at Parma, Idaho during 1998 using Eq. 47 and using the more accurate $K_B + K_D$ method of Appendix D.

Extraterrestrial radiation for hourly periods (R_a)[20]

Extraterrestrial radiation, R_a, defined as the short-wave solar radiation in the absence of an atmosphere, is a well-behaved function of the day of the year, time of day, latitude, and longitude. For hourly time periods, the solar time angle at the beginning and end of the period serve as integration endpoints for calculating R_a:

$$R_a = \frac{12}{\pi} G_{sc}\, d_r \left[(\omega_2 - \omega_1)\sin(\varphi)\sin(\delta) + \cos(\varphi)\cos(\delta)\left(\sin(\omega_2) - \sin(\omega_1)\right)\right] \qquad (48)$$

where

R_a	= extraterrestrial radiation during the hour (or shorter) period [MJ m^{-2} $hour^{-1}$],
G_{sc}	= solar constant = 4.92 MJ m^{-2} h^{-1},
d_r	= inverse relative distance factor (squared) for the earth-sun [unitless],
δ	= solar declination [radians],
φ	= latitude [radians],
ω_1	= solar time angle at beginning of period [radians], and
ω_2	= solar time angle at end of period [radians].

The latitude, φ, expressed in radians is positive for the Northern Hemisphere and negative for the Southern Hemisphere. The conversion from decimal degrees to radians is given by:

$$\text{Radians} = \frac{\pi}{180}\left(\text{decimal degrees}\right) \qquad (49)$$

and d_r and δ are calculated as:

$$d_r = 1 + 0.033 \cos\left(\frac{2\pi}{365} J\right) \qquad (50)$$

$$\delta = 0.409 \sin\left(\frac{2\pi}{365} J - 1.39\right) \qquad (51)$$

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[20] Reference: Duffie and Beckman (1980).

where J is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December). The constant 365 in Eqs. 50 and 51 is held at 365 even during a leap year. J can be calculated as[21]:

$$J = D_M - 32 + \mathrm{Int}\left(275\frac{M}{9}\right) + 2\ \mathrm{Int}\left(\frac{3}{M+1}\right) + \mathrm{Int}\left(\frac{M}{100} - \frac{\mathrm{Mod}(Y,4)}{4} + 0.975\right) \quad (52)$$

where:

D_M	= the day of the month (1-31),
M	= the number of the month (1-12), and
Y	= the number of the year (for example 1996 or 96).

The "Int" function in Eq. 52 finds the integer number of the argument in parentheses by rounding downward. The "Mod(Y,4)" function finds the modulus (remainder) of the quotient Y/4.

The solar time angles at the beginning and end of each period are given by:

$$\omega_1 = \omega - \frac{\pi\ t_1}{24} \quad (53)$$

$$\omega_2 = \omega + \frac{\pi\ t_1}{24} \quad (54)$$

where:

ω	= solar time angle at the midpoint of the period [radians], and
t_1	= length of the calculation period [hour]: i.e., 1 for hourly periods or 0.5 for 30-minute periods.

The solar time angle at the midpoint of the period is:

$$\omega = \frac{\pi}{12}\left[\left(t + 0.06667(L_z - L_m) + S_c\right) - 12\right] \quad (55)$$

where:

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[21] Reference: Allen (2000)

t = standard clock time at the midpoint of the period [hour] (after correcting time for any daylight savings shift). For example for a period between 1400 and 1500 hours, t = 14.5 hours,

L_z = longitude of the center of the local time zone [expressed as positive degrees west of Greenwich, England]. In the United States, L_z = 75, 90, 105 and 120° for the Eastern, Central, Rocky Mountain and Pacific time zones, respectively, and L_z = 0° for Greenwich, 345° for Paris (France), and 255° for Bangkok (Thailand),

L_m = longitude of the solar radiation measurement site [expressed as positive degrees west of Greenwich, England], and

S_c = seasonal correction for solar time [hour].

Because ω_s is the sunset hour angle and $-\omega_s$ is the sunrise hour angle (noon has $\omega = 0$), values of $\omega < -\omega_s$ or $\omega > \omega_s$ from Eq. 55 indicate that the sun is below the horizon, so that, by definition, R_a and R_{so} are zero and their calculation has no meaning. When the values for ω_1 and ω_2 span the value for $-\omega_s$ or for ω_s, this indicates that sunrise or sunset occurs within the hourly (or shorter) period. In this case, the integration limits for applying Eq. 48 should be correctly set using the following conditionals:

$$\begin{gathered}
\text{If } \omega_1 < -\omega_s \text{ then } \omega_1 = -\omega_s \\
\text{If } \omega_2 < -\omega_s \text{ then } \omega_2 = -\omega_s \\
\text{If } \omega_1 > \omega_s \text{ then } \omega_1 = \omega_s \\
\text{If } \omega_2 > \omega_s \text{ then } \omega_2 = \omega_s \\
\text{If } \omega_1 > \omega_2 \text{ then } \omega_1 = \omega_2
\end{gathered} \tag{56}$$

The above conditionals can be applied for all timesteps to insure numerical stability of the application of Eq. 48 as well as correctly computing the theoretical quantity of solar radiation early and late in the day. The user should recognize that Eqs. 48-56 and 62 presume an extensive, flat ground surface and are based on a vector to the center of the sun's disk. The calculations do not account for diffuse radiation occurring shortly before sunrise and shortly after sunset. Where there are hills or mountains, the hour angle when the sun first appears or disappears may increase for sunrise or decrease for sunset.

The seasonal correction for solar time is:

$$S_c = 0.1645 \sin(2\,b) - 0.1255 \cos(b) - 0.025 \sin(b) \tag{57}$$

$$b = \frac{2\pi(J-81)}{364} \tag{58}$$

where J is the number of the day in the year and b has units of radians.

The sunset hour angle, ω_s, is given by:

$$\omega_s = \arccos\left[-\tan(\varphi)\tan(\delta)\right] \tag{59}$$

The “arccos” function is the arc-cosine function and represents the inverse of the cosine. This function is not available in all computer languages, so that ω_s can alternatively be computed using the arc tangent (inverse tangent) function:

$$\omega_s = \frac{\pi}{2} - \arctan\left[\frac{-\tan(\varphi)\tan(\delta)}{X^{0.5}}\right] \tag{60}$$

where:

$$X = 1 - [\tan(\varphi)]^2\,[\tan(\delta)]^2 \tag{61}$$

$$\text{and } X = 0.00001 \text{ if } X \le 0$$

The user should confirm accurate setting of the datalogger clock. If clock times are in error by more than 5-10 minutes, estimates of extraterrestrial and clear sky radiation may be significantly impacted. This can lead to errors in estimating R_n on an hourly or shorter basis, especially early and late in the day. A shift in “phase” between measured R_s and R_{so} predicted from R_a according to the data logger clock can indicate error in the reported time. More discussion is given in Appendix D.

The angle of the sun above the horizon, β, at the midpoint of the hourly or shorter time period is computed as:

$$\beta = \arcsin\left[\sin(\varphi)\sin(\delta) + \cos(\varphi)\cos(\delta)\cos(\omega)\right] \quad (62)$$

where

- β = angle of the sun above the horizon at midpoint of the period [radians],
- φ = latitude [radians],
- δ = solar declination [radians],
- ω = solar time angle at the midpoint of the period [radians] (from Eq. 55).

The "arcsin" function is the arc-sine function and represents the inverse of the sine. This function is not available in all computer languages, so that β can alternatively be computed using the arc tangent (inverse tangent) function:

$$\beta = \arctan\left[\frac{Y}{\left(1 - Y^2\right)^{0.5}}\right] \quad (63)$$

where:

$$Y = \sin(\varphi)\sin(\delta) + \cos(\varphi)\cos(\delta)\cos(\omega) \quad (64)$$

and all other parameters are defined following Eq. 62.

Soil Heat Flux Density (G)

Soil heat flux density is the thermal energy that is utilized to heat the soil. G is positive when the soil is warming and negative when the soil is cooling. For hourly calculation periods, G beneath a dense cover of grass or alfalfa does not correlate well with air temperature, but can be significant. Hourly G generally correlates well with net radiation and amount of vegetative cover and can be approximated as a fraction of

R_n. The following equations are based on Eq. B.13 of Appendix B for fixed vegetation height and leaf area index[22].

For the standardized short reference ET_{os} :

$$G_{hr\,daytime} = 0.1\ R_n \quad (65a)$$

$$G_{hr\,nighttime} = 0.5\ R_n \quad (65b)$$

where G and R_n have the same measurement units (MJ m^{-2} h^{-1} for hourly or shorter time periods). For the standardized tall reference ET_{rs} :

$$G_{hr\,daytime} = 0.04\ R_n \quad (66a)$$

$$G_{hr\,nighttime} = 0.2\ R_n \quad (66b)$$

For standardization, nighttime is defined as when measured or calculated hourly net radiation R_n is < 0 (i.e., negative). When the soil is warming, the soil heat flux density, G, has a positive value. The amount of energy consumed by G is subtracted from R_n when estimating ET_{os} or ET_{rs}.

Wind Profile Relationship

Wind speed varies with height above the ground surface. For the calculation of ET_{sz}, wind speed at 2 meters above the surface is required, therefore, wind measured at other heights must be adjusted. To adjust wind speed data to the 2-m height, Eq. 68 should be used for measurements above a short grass (or similar) surface, based on the full logarithmic wind speed profile equation B.14 given in Appendix B.

$$u_2 = u_z \frac{4.87}{\ln(67.8\ z_w - 5.42)} \quad (67)$$

where

u_2 = wind speed at 2 m above ground surface [m s^{-1}],
u_z = measured wind speed at z_w m above ground surface [m s^{-1}], and

[22] Leaf area index (LAI) is defined as the area (one-sided) of leaves per unit area of ground surface. Units are dimensionless (i.e., m^2 m^{-2})

z_w = height of wind measurement above ground surface [m].

For wind measurements above surfaces other than clipped grass, the user should apply the full logarithmic equation B.14. A special application of Eq. B.14 is given in Appendix B for wind measured above alfalfa or similar vegetation having about 0.5 m height. It is noted that wind speed data collected at heights above 2 m are acceptable for use in the standardized equations following adjustment to 2 m, and may be preferred if vegetation adjacent to the station commonly exceeds 0.5 m. Measurement at a greater height, for example 3 m, reduces the influence of the taller vegetation.

Negative Values Computed for ET_{sz}

Values calculated for reference ET for nighttime hours occasionally take on negative values. In practice, the user may wish to set negative values to zero before summing over the 24-hour period. However, in some situations, negative hourly computed ET_{os} or ET_{rs} may indicate some condensation of vapor during periods of early morning dew and should therefore be registered as negative during the summing of 24-hour ET. In other situations, negative hourly ET_{os} or ET_{rs} during nighttime reflect the uncertainties in some parameter estimates including R_n and assumptions implicit to the combination equation. The impact of negative hourly values on ET summed over daily periods is usually less than a few percent. In general, it may be appropriate to retain the negative values.

DEFINITION AND APPLICATION OF CROP COEFFICIENTS

Calculation of crop evapotranspiration ($ET_{c)}$ requires the selection of the appropriate crop coefficient (K_c) for use with the standardized reference evapotranspiration (ET_{os} or ET_{rs}). It is recommended that the abbreviation for crop coefficients developed for use with ET_{os} be denoted as K_{co} and the abbreviation for crop coefficients developed for use with ET_{rs} be denoted as K_{cr}. ET_c is calculated as:

$$ET_c = K_{co} * ET_{os} \quad \underline{\text{or}} \quad ET_c = K_{cr} * ET_{rs} \tag{68}$$

TRANSFER AND CONVERSION OF CROP COEFFICIENTS

Crop coefficients (K_c) and landscape coefficients available in the literature are referenced to either clipped, cool season grass or full-cover alfalfa. Without appropriate adjustment, crop coefficients for the two references are not interchangeable. For this standardization effort, a grass reference crop is defined as an extensive, uniform surface of dense, actively growing, cool-season grass with a height of 0.12 m, and not short of soil water. A full-cover alfalfa reference crop is defined here as an extensive, uniform surface of dense, actively growing alfalfa with a height of 0.50 m, and not short of soil water.

Grass-based crop coefficients should be used with ET_{os}, and alfalfa-based coefficients should be used with ET_{rs}. If a calculated or measured reference other than ET_{os} or ET_{rs} was used to develop the crop coefficients, it must be established that the reference equation or reference measurements provide values that are equivalent to ET_{os} or ET_{rs} (see Appendix A for comparisons between selected methods). It is important to establish the differences between ET equations since some equations developed to estimate grass or alfalfa reference ET may not agree exactly with ET_{os} or ET_{rs} during all time periods or under all climatic conditions.

K_c values that can be used with ET_{os} without adjustment include those reported in FAO-56 (Allen et al., 1998) and ASCE Manual 70 (Jensen et al., 1990, Table 6.8). Coefficients that can be used as is with ET_{os} for most practical applications are those reported by FAO-24 (Doorenbos and Pruitt, 1977) and SCS NEH Part 623 Chapter 2 (Martin and Gilley, 1993). Coefficients based on the CIMIS Penman equation (Snyder and Pruitt, 1992) should not require adjustment for use with ET_{os}. K_c values that can be used as is with ET_{rs} for most practical applications are those reported by Wright (1982) and ASCE Manual 70 (Jensen et al., 1990, Tables 6.6 and 6.9). There is a tendency for relatively minor overestimation of ET using K_c from Wright (1982) with ET_{rs} in spring and fall. Thus, the K_c values by Wright (1981, 1982) have been converted for direct use with ET_{rs}, with ET_{rs} calculated on a 24-h time step (Allen and Wright, 2002).

Some grass and alfalfa based crop coefficients are "mean" crop coefficients (e.g., Wright, 1979; 1981; Doorenbos and Pruitt, 1977). Mean crop coefficients incorporate the effects of irrigation, rainfall, and soil type at the development site. Users of these mean crop coefficients are cautioned that differences in irrigation frequency, rainfall patterns, and/or soil drying characteristics between the development site and the study site could cause error in the ET_c estimate.

The publications referenced in the above paragraphs contain descriptions on determination and application of crop coefficients during growing periods. This information will not be presented here. The following section discusses the application of ET_{sz} and K_c during non-growing periods.

CALCULATION OF REFERENCE EVAPOTRANSPIRATION DURING NON-GROWING PERIODS

During cold periods in many regions, freezing temperatures preclude vegetation from remaining green and actively growing. These periods are referred to as non-growing periods. Evapotranspiration from non-active vegetation during non-growing periods is generally less than reference ET because plants may be dormant and therefore may have substantially increased surface resistance. Besides surface resistance, albedo or reflectance of dormant or dead vegetation is generally greater than that of green vegetation. Both of these characteristics reduce the potential rate of ET from plant residue. This may make it difficult to assess the validity of reference ET equations under these conditions.

While it is recognized that the reference ET equations do not represent measurable quantities during non-growing periods, the ET_{sz} equation can still be useful as an evaporative index. However, the user must be aware that conditions for the reference surfaces for ET_{os} and ET_{rs} may not exist during non-growing periods. Under many non-growing conditions, it is possible to incorporate the differences between dormant or dead vegetation ET and ET_{sz} into the K_c value. However there are other factors to be considered. For example, the soil heat flux estimates may be uncertain, low sun angles and snow cover will influence albedo, and short day lengths will affect the calculation of net radiation and ET_{sz} for daily time steps.

In this document the Task Committee will not recommend a methodology for the application of reference evapotranspiration during non-growing seasons. Two other ASCE Task Committees are investigating evaporative losses during non-growing seasons and are developing estimation methodologies.

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GLOSSARY OF TERMS

FOR THE

ASCE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION

C_d	denominator constant that changes with reference type and calculation time step (s m^{-1})
C_n	numerator constant that changes with reference type and calculation time step (K mm s^3 Mg^{-1} d^{-1} or K mm s^3 Mg^{-1} h^{-1})
D_M	day of the month (1-31)
ET	Evapotranspiration (mm d^{-1} or mm h^{-1})
ET_c	Crop evapotranspiration
ET_{os}	Reference ET for a *short* crop with an approximate height of *0.12* m (similar to clipped grass) (mm d^{-1} or mm h^{-1})
ET_{ref}	Reference Evapotranspiration (mm d^{-1} or mm h^{-1})
ET_{rs}	Reference ET for a *tall* crop with an approximate height of *0.50* m (similar to full-cover alfalfa) (mm d^{-1} or mm h^{-1})
ET_{sz}	Standardized Reference Evapotranspiration Equation
G	soil heat flux density at the soil surface (MJ m^{-2} d^{-1} for daily time steps or MJ m^{-2} h^{-1} for hourly time steps)
G_{day}	daily soil heat flux density (MJ m^{-2} d^{-1})
$G_{hr\ daytime}$	hourly soil heat flux density during daytime (MJ m^{-2} h^{-1})
$G_{hr\ nighttime}$	hourly soil heat flux density during nighttime (MJ m^{-2} h^{-1})
G_{month}	monthly soil heat flux density (MJ m^{-2} d^{-1})
G_{sc}	solar constant (4.92 MJ m^{-2} h^{-1})
J	day of the year (1 – 365)
J_{month}	month of the year (1 –12)
K_{ab}	coefficient derived from the a_s and b_s coefficients of the Angstrom formula (unitless)
K_B	the clearness index for direct beam radiation (unitless)
K_c	crop coefficient
K_{co}	crop coefficient for use with ET_{os}
K_{cr}	crop coefficient for use with ET_{rs}
K_D	the transmissivity index for diffuse radiation (unitless)
K_G	coefficient used to calculate hourly soil heat flux (unitless)
K_t	atmospheric turbidity coefficient (unitless)
K_{time}	units conversion, equal to 86,400 s d^{-1} for ET in mm d^{-1} and equal to 3600 s h^{-1} for ET in mm h^{-1}
K_o	average difference between T_{min} and mean daily T_{dew} (^oC)

LAI	leaf area index = area (one-sided) of leaves per unit area of ground surface (m^2 m^{-2})
LAI_{active}	active (sunlit) leaf area index, m^2 (leaf area) m^{-2} (soil surface)
L_m	longitude of the measurement site (expressed as positive degrees west of Greenwich, England)
L_z	longitude of the center of the local time zone (expressed as positive degrees west of Greenwich, England)
M	number of the month (1-12)
N	maximum duration of sunshine or daylight hours (h)
P	atmospheric pressure at station elevation z (kPa)
P_o	atmospheric pressure at sea level = 101.3 (kPa)
R	specific gas constant = 287 (J kg^{-1} K^{-1})
R_a	extraterrestrial radiation (MJ m^{-2} d^{-1}) or (MJ m^{-2} h^{-1})
RH	relative humidity (%)
RH_{max}	daily maximum relative humidity (%)
RH_{mean}	mean daily relative humidity
RH_{min}	daily minimum relative humidity (%)
R_{lu}	long-wave radiation emitted from the surface
R_{ld}	long-wave radiation emitted from the atmosphere
R_n	net radiation at the crop surface (MJ m^{-2} d^{-1} or MJ m^{-2} h^{-1})
R_{nl}	net long-wave radiation (MJ m^{-2} d^{-1} or MJ m^{-2} h^{-1}), defined as being positive upwards and negative downwards
R_{ns}	net short-wave radiation (MJ m^{-2} d^{-1} or MJ m^{-2} h^{-1}), defined as being positive downwards and negative upwards
R_s	measured or calculated solar radiation (MJ m^{-2} d^{-1}) or (MJ m^{-2} h^{-1})
R_{so}	clear-sky radiation (MJ m^{-2} d^{-1}) or (MJ m^{-2} h^{-1})
S_c	seasonal correction for solar time (h)
T	mean daily or hourly air temperature at 1.5 to 2.5-m height (°C)
T_{dew}	dew point temperature (°C)
T_{dry}	dry bulb temperature (°C)
T_{hr}	mean hourly air temperature (°C)
T_K	mean absolute temperature (K)
$T_{K\ hr}$	mean absolute temperature during the hour (K)
T_{Ko}	reference temperature at elevation z_o (K)
$T_{K\ max}$	maximum absolute temperature during the 24-hour period (K)
$T_{K\ min}$	minimum absolute temperature during the 24-hour period (K)
T_{Kv}	mean virtual temperature for period (K)
T_{hr}	mean hourly air temperature (°C)
T_{max}	daily maximum air temperature (°C)
T_{mean}	mean air temperature for the time period of calculation (°C)
T_{min}	daily minimum air temperature (°C)
T_{month}	monthly mean air temperature (°C)
T_{wet}	wet bulb temperature (°C)
W	precipitable water in the atmosphere (mm)

Y	number of the year (for example 1996 or 96)
a_{psy}	coefficient depending on the type of ventilation of the wet bulb of a psychrometer (°C^{-1})
a_s	coefficient of the Angstrom formula (unitless)
b_s	coefficient of the Angstrom formula (unitless)
c_p	specific heat of the air, (MJ kg^{-1} °C^{-1})
d	zero plane displacement height, (m)
daytime	hourly or shorter period when $R_n \geq 0$
d_r	inverse relative distance earth-sun (unitless)
e_a	mean actual vapor pressure at 1.5 to 2.5-m height (kPa)
$e^o(T)$	saturation vapor pressure function (kPa)
e_s	saturation vapor pressure at 1.5 to 2.5-m height (kPa)
f_{cd}	cloudiness function (unitless)
$f_{cd\ \beta>0.3}$	cloudiness function for the time period prior to when sun angle β (in the afternoon or evening) falls below 0.3 radians (unitless)
g	gravitational acceleration = 9.807 (m s^{-2})
h	reference vegetation height (m)
k	von Karman's constant, 0.41, (dimensionless)
k_{Rs}	adjustment coefficient for predicting R_s from air temperature (°$C^{-0.5}$)
n	recorded duration of sunshine during a day (h)
nighttime	hourly or shorter period when $R_n < 0$
r_a	aerodynamic resistance (s m^{-1})
r_l	bulk stomatal resistance of a well-illuminated leaf (s m^{-1})
r_s	surface resistance (s m^{-1})
t	standard clock time at the midpoint of the period
t_l	length of the calculation period (h)
u_2	mean daily or hourly wind speed at 2-m height (m s^{-1})
u_z	wind speed at height z (m s^{-1})
z	weather site elevation above mean sea level (m)
z_h	height of air temperature and humidity measurements (m)
z_o	elevation at reference level (i.e., sea level) (m)
z_{om}	roughness length governing momentum transfer (m)
z_{oh}	roughness length for transfer of heat and vapor (m)
z_w	height corresponding to wind speed (m)
α	"alpha" = albedo or canopy reflection coefficient (unitless)
α_l	constant lapse rate moist air = 0.0065 (K m^{-1})
β	"beta" = angle of the sun above the horizon (radians)
γ	"gamma" = psychrometric constant (kPa °C^{-1})
γ_{psy}	psychrometric constant for the psychrometer (kPa °C^{-1})
Δ	"delta" = slope of the saturation vapor pressure-temperature curve (kPa °C^{-1})
δ	"delta" = solar declination (radians)

ε	"epsilon" = ratio of the molecular weight of water vapor to dry air (unitless) (ε = 0.622)
λ	"lambda" = latent heat of vaporization (MJ/kg)
φ	"phi" = latitude (radians)
ρ_a	"rho" = air density (Kg m^{-3})
ρ_w	water density (Mg m^{-3}) (taken as 1.0 Mg m^{-3})
σ	"sigma" = Stefan-Boltzmann constant (4.901 10^{-9} MJ K^{-4} m^{-2} d^{-1})
ω	"omega" solar time angle (radians), solar noon = 0.
ω_s	sunset hour angle (radians)
ω_1	solar time angle at beginning of hourly or shorter period (radians)
ω_2	solar time angle at end of hourly or shorter period (radians)