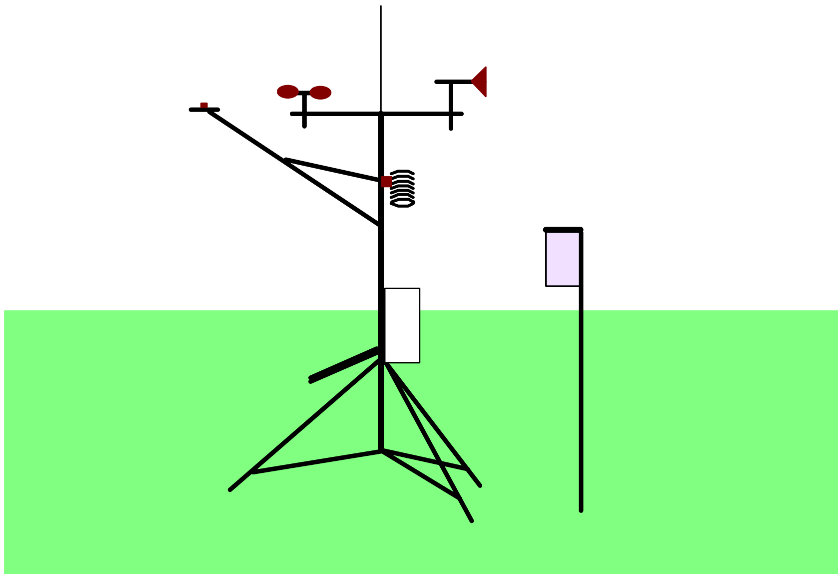
THE ASCE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION



Environmental and Water Resources Institute
of
the American Society of Civil Engineers

Standardization of Reference Evapotranspiration Task Committee

December 21, 2001
revised July 9, 2002
Draft

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Environmental and Water Resources Institute of the American Society of Civil Engineers, Standardization of Reference Evapotranspiration Task Committee[1]

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 that can be used 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 incorporating existing crop and landscape coefficients and existing reference ET calculations.

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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 additional 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:

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

Standardized Reference Evapotranspiration Equation, Tall (ET_{rs}): 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 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:

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

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, and
- C_d = denominator constant that changes with reference type and calculation time step.

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. 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.

USE OF THE STANDARDIZED REFERENCE EVAPOTRANSPIRATION EQUATION

Based on an intensive review of reference evapotranspiration calculated at 49 sites throughout the United States (described in the following section), the ASCE-ET found the standardized 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.

The accuracy of any evapotranspiration calculation depends on the quality of the weather data. When possible, weather data should be measured at stations that are located over grassed surfaces in open, well-watered settings and that have good quality control and quality assurance procedures in place. Suggestions for assessing and improving the integrity of collected weather data are described in Appendix D. Appendix D also provides 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 also provides background on the development of the standardized form of the PM 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[2]

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 recommended. The inverse of λ is approximately 0.408 kg MJ^{-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.

$$T = \frac{T_{max} + T_{min}}{2} \tag{2}$$

where:

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

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[3]:

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[2] 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).

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

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 $°C^{-1}$.

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[4], Δ, is computed as:

$$\Delta = \frac{2504\ \exp\left(\frac{17.27\,T}{T + 237.3}\right)}{(T + 237.3)^2} \tag{5}$$

where:

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

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

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

Saturation Vapor Pressure (e_s)

The saturation vapor pressure[5] (e_s) represents the capacity of the air to hold water vapor.

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

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

where:

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

The function to calculate saturation vapor pressure is:

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

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.

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

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	Preference Ranking	Equation(s)
e_a averaged over the 24-hour period (based on hourly or more frequent measurements of humidity)[a,b]	1	7, 41
Measured dew point temperature averaged over period	1	8
Wet-bulb and dry-bulb temperature averaged over period	2	7, 9, 10
Measured dew point temperature at 7 or 8 am	2	8
Daily maximum and minimum relative humidity	2	7, 11
Daily maximum relative humidity	3	7, 12
Daily minimum relative humidity	3	7, 13
Daily minimum air temperature (see Appendix E)	4	--
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.

When humidity and psychrometric data are missing or are of questionable integrity, dew point temperature can be predicted from daily minimum air temperature as described in Appendix E. This possibility 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:

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

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:

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

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[6]:

a_{psy} = 0.000662 for ventilated (Asmann type) psychrometers having air movement of about 5 m s^{-1},
= 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.

e_a from relative humidity data

The actual vapor pressure of air can also be calculated from relative humidity (RH) and the corresponding air temperature. When using RH data, it is critical 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}) should be paired with T_{min}, which will both occur, generally, during early morning. Daily minimum relative humidity (RH_{min}) is paired with T_{max}.

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

Depending on the availability of the RH data, the following equations apply:

- <u>Daily e_a from RH_{max} and RH_{min}</u>.

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

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.

- <u>Daily e_a from RH_{max}</u>

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}:

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

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 data. Exceptions to this trend are where substantial advection of dry or warm air from dry regions outside the area occurs during nighttime.

- Daily e_a from RH_{min}

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

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

However, these estimates may be less desirable than estimates using the above approaches, since often RH sensors, especially those manufactured prior to about 1990, tend to be less accurate at low RH values than at high RH values. 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, the actual vapor pressure may be estimated as:

$$e_a = \frac{RH_{mean}}{100}\left[\frac{e^o(T_{max}) + e^o(T_{min})}{2}\right] \tag{14}$$

where RH_{mean} is the mean daily relative humidity, generally defined as the average between RH_{max} and RH_{min}. Eq. 14 is less desirable than Eqs. 12 or 13 because the $e^o(T)$ relationship is highly nonlinear.

Net Radiation (R_n)

Net radiation (R_n) is the net amount of radiant energy available at the surface for evaporating water, heating the air, or heating the surface. R_n includes both short and long wave radiation components [7]:

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

where:

R_{ns}	=	net short-wave radiation, [MJ m^{-2} d^{-1}], defined as being positive downwards and negative upwards,
R_{nl}	=	net 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 one chooses to measure R_n, one must exercise care and attention 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 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:

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

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

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, 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 emissivity. If users intend to utilize a different approach for calculating R_n, they are strongly encouraged 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 upward long-wave radiation from the surface (R_{lu}) and downward long-wave radiation from the sky (R_{ld}):

$$R_{nl} = R_{lu} - R_{ld} \qquad (17)$$

The following calculation for <u>daily</u> net long-wave radiation follows the method of Brunt (1932, 1952) that uses vapor pressure from a weather station to predict net emissivity:

$$R_{nl} = \sigma\left[\frac{T_{K\ max}^4 + T_{K\ min}^4}{2}\right]\left(0.34 - 0.14\sqrt{e_a}\right)\left(1.35\frac{R_s}{R_{so}} - 0.35\right) \qquad (18)$$

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}],
$T_{K\ max}$ = 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),
e_a = actual vapor pressure [kPa],
R_s/R_{so} = relative solar radiation (limited to ≤ 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 superscripts "4" in Eq. 18 indicate the need to raise the air temperature, expressed in units of Kelvin, to the power of 4. The ratio R_s/R_{so} in Eq. 18 is used to indicate relative cloudiness and must be limited to $0.25 < R_s/R_{so} \leq 1.0$.

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. 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.

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[8] is 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.

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[8] Reference: Allen (1996)

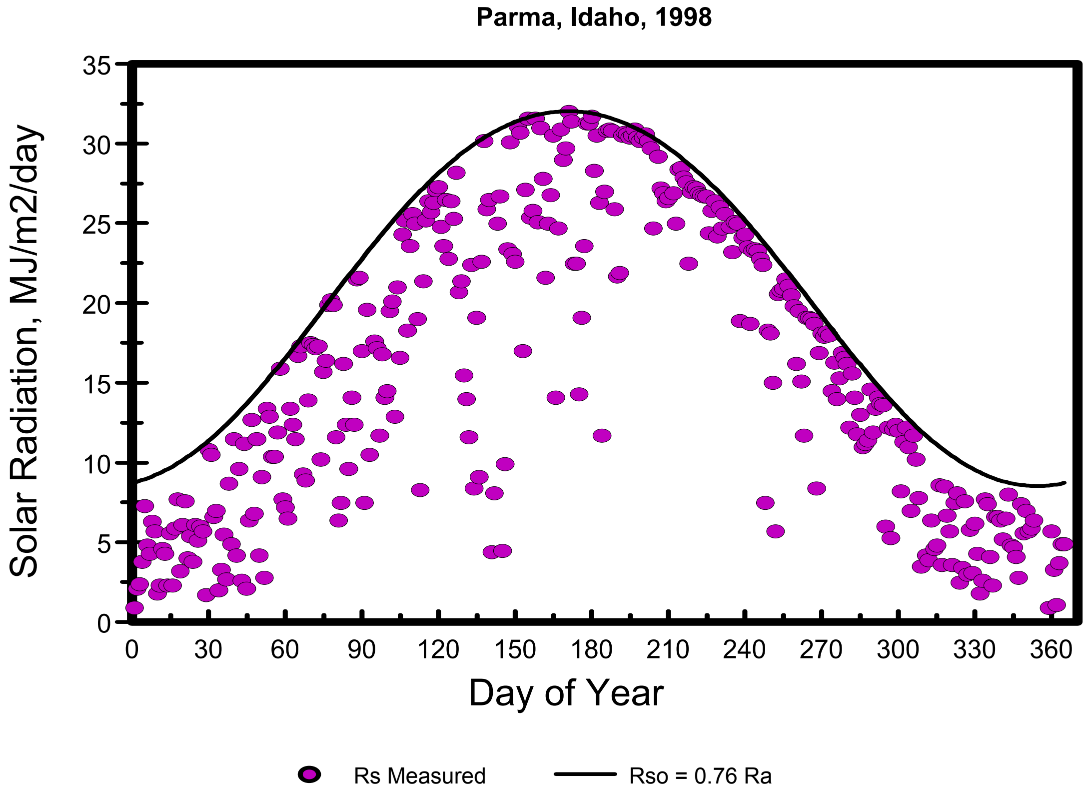

Figure 1. Daily R_s at Parma, Idaho during 1998 (elevation 703 m, Lat. 43.8°) and R_{so} envelope

When 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 was 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. Other 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 Eq. 19 or 20, as opposed to the more complicated R_{so} equation of Appendix D, will be generally less than 2% over an annual period. The R_{so} method in Appendix D is considered to produce a better estimate of R_{so}.

Extraterrestrial Radiation for 24-Hour Periods (R_a)[9]

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:

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

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

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:

$$\text{Radians} = \frac{\pi}{180}(\text{decimal degrees}) \tag{22}$$

and d_r and δ are calculated as:

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

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

where:

J is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December). J can be calculated as[10]:

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

where:

D_M	= the day of the month (1-31),

[10] Reference: Allen (2000).

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:

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

where:

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

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

<u>Soil Heat Flux Density (G)</u>

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 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 about 2.0 MJ m^{-3} $°C^{-1}$ and a soil depth of about 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 not measured at that height 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. 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.

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>[11]

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 recommended. The inverse of λ is approximately 0.408 kg MJ^{-1}.

Mean Air Temperature (T)

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

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[12]:

[11] 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).

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

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 $\lambda = 2.45$ 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 $°C^{-1}$.

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[13], Δ, is computed as:

$$\Delta = \frac{2504\ \exp\left(\frac{17.27\ T}{T + 237.3}\right)}{(T + 237.3)^2} \tag{36}$$

where:

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

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

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

Saturation Vapor Pressure (e_s)

The saturation vapor pressure[14], e_s, represents the capacity of the air to hold water vapor.

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

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

where vapor pressure is in units of kPa and T is mean air temperature during the hourly period in °C.

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. The availability and quality of local data may justify a different order of preference.

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

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

Method	Preference Ranking	Equation(s)
e_a averaged over period[a,b]	1	--
Measured dew point temperature averaged over period	1	38
Average RH and T for the hour	1	37, 41
Wet-bulb and dry-bulb temperature	2	38, 39, 40
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 predicted from daily minimum air temperature as described in Appendix E. 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[15]:

a_{psy} = 0.000662 for ventilated (Asmann type) psychrometers, with air movement of approximately 5 m s^{-1},
= 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).

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

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

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

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

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 surface for evaporating water, heating the air, or heating the surface. R_n includes both short and long wave radiation components [16]:

$$R_n = R_{ns} - R_{nl} \tag{42}$$

where:

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

R_{nl} = net 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 one chooses to measure R_n, one must exercise care and attention 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

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

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:

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

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, they are strongly encouraged 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 surface (R_{lu}) and long-wave radiation radiated downward from the atmosphere (R_{ld}):

$$R_{nl} = R_{lu} - R_{ld} \tag{44}$$

The following calculation for net long-wave radiation follows the method of Brunt (1932, 1952) that uses near surface vapor pressure to predict net surface emissivity:

$$R_{nl} = \sigma T_{K\,hr}^{4} \left(0.34 - 0.14\sqrt{e_a}\right)\left(1.35\frac{R_s}{R_{so}} - 0.35\right) \tag{45}$$

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}],
$T_{K\,hr}$	= mean absolute temperature during the hourly period [K] (K = °C + 273.16),
e_a	= actual vapor pressure [kPa],
R_s/R_{so}	= relative short-wave radiation (limited to ≤ 1.0), (see discussion under R_{so})
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 superscript "4" in Eq. 45 indicates the need to raise the air temperature, expressed in units of Kelvin, to the power of 4. The ratio R_s/R_{so} is used to indicate relative cloudiness and must be limited to $0.25 < R_s/R_{so} < 1.0$. The following section describes how to determine values for R_s/R_{so} for nighttime conditions.

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 hourly R_{so} it is recommended that R_{so}, for purposes of calculating R_n, be calculated using the following simple approach:

$$R_{so} = \left(0.75 + 2 \times 10^{-5}\, z\right) R_a \tag{46}$$

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 46 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 46 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. These 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. 46 will generally be less than 3% across a day, and less than 2% over an annual period.

Figure 2 illustrates a comparison of measured hourly solar radiation with R_{so} computed using Eq. 46 and using the more complicated 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. 46. Measured R_s exceeded the R_{so} curves for the 1100 reading on June 19 because of reflection from clouds adjacent to the weather site. In general, the solar radiation data appear to be of excellent quality and calibration.

R_s/R_{so} for Hourly Periods

The ratio R_s/R_{so} is used to represent cloud cover in the calculation of net radiation. When calculating R_{nl} during the nighttime, use the ratio R_s/R_{so} for a time period occurring 2-3 hours before sunset. The hourly period that is 2 to 3 hours before sunset can be identified during computation of R_a as the period where the solar time angle ω is within the range $(\omega_s - 0.79) \leq \omega \leq (\omega_s - 0.52)$, where ω and sunset hour angle ω_s are calculated in the following section. Another approach is to use both the preceding late afternoon and subsequent early morning data to approximate cloud cover during the nighttime hours (Dong et al., 1992).

As a more approximate alternative, one can assume R_s/R_{so} = 0.4 to 0.6 during nighttime periods in humid and sub-humid climates and R_s/R_{so} = 0.7 to 0.8 in arid and semiarid climates. A value of R_s/R_{so} = 0.3 presumes total cloud cover.

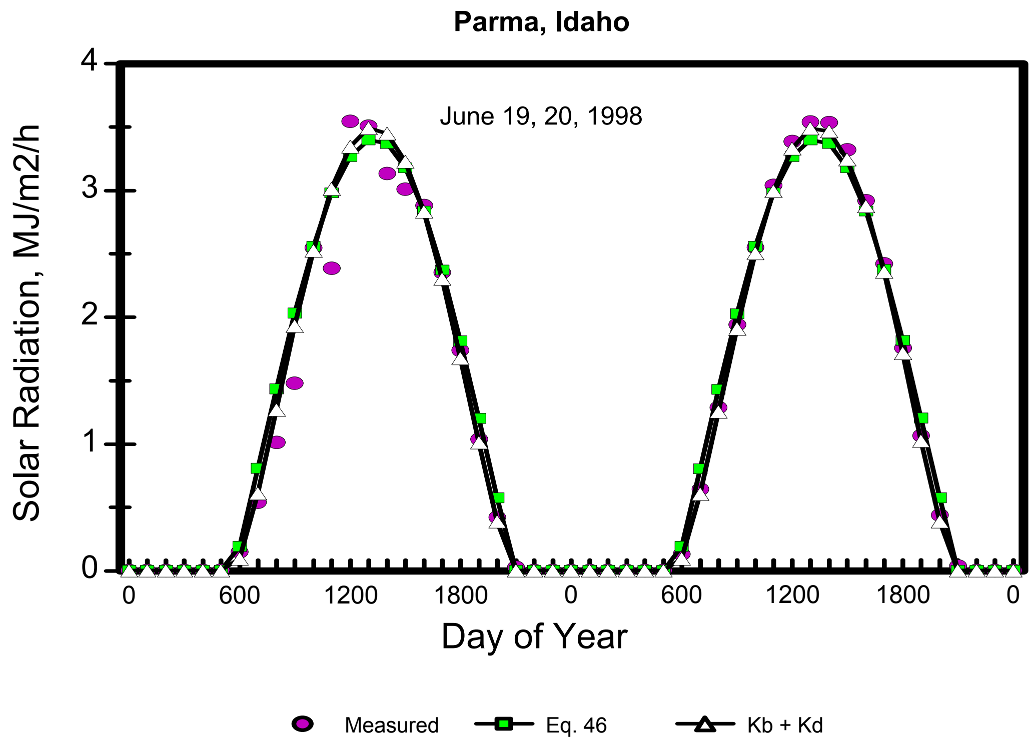

Figure 2. Measured and calculated hourly R_{so} for two days at Parma, Idaho during 1998 using Eq. 46 and using the $K_B + K_D$ method in Appendix D.

Extraterrestrial radiation for hourly periods (R_a)[17]

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:

$$R_a = \frac{12}{\pi} G_{sc}\, d_r \left[(\omega_2 - \omega_1)\sin(\varphi)\sin(\delta) + \cos(\varphi)\cos(\delta)(\sin(\omega_2) - \sin(\omega_1))\right] \tag{47}$$

where

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

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:

$$\text{Radians} = \frac{\pi}{180}(\text{decimal degrees}) \tag{48}$$

and d_r and δ are calculated as:

$$d_r = 1 + 0.033 \cos\left(\frac{2\pi}{365} J\right) \tag{49}$$

$$\delta = 0.409 \sin\left(\frac{2\pi}{365} J - 1.39\right) \tag{50}$$

where J is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December). J can be calculated as[18]:

$$J = D_M - 32 + \text{Int}\left(275\frac{M}{9}\right) + 2\ \text{Int}\left(\frac{3}{M+1}\right) + \text{Int}\left(\frac{M}{100} - \frac{\text{Mod}(Y,4)}{4} + 0.975\right) \tag{51}$$

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).

[18] Reference: Allen (2000)

The "Int" function in Eq. 51 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:

$$\omega_1 = \omega - \frac{\pi\ t_1}{24} \tag{52}$$

$$\omega_2 = \omega + \frac{\pi\ t_1}{24} \tag{53}$$

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 midpoint of the period is:

$$\omega = \frac{\pi}{12}\left[(t + 0.06667(L_z - L_m) + S_c) - 12\right] \tag{54}$$

where:

- t = standard clock time at the midpoint of the period [hour]. For example for a period between 1400 and 1500 hours, t = 14.5,
- 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 measurement site [expressed as positive degrees west of Greenwich, England], and
- S_c = seasonal correction for solar time [hour].

Values of $\omega < -\omega_s$ or $\omega > \omega_s$ from Eq. 54, where ω_s is the sunset hour angle and $-\omega_s$ is the sunrise hour angle (noon has $\omega = 0$), indicates that the sun is below the horizon so that, by definition, R_a

is zero. When the values for ω_1 and ω_2 span the value for $-\omega_s$ or ω_s, this indicates that sunrise or sunset occurs within the hourly (or shorter) period. In this case, the integration limits for applying Eq. 47 can be correctly set using the following conditionals:

$$\begin{gathered}\text{If } \omega_1 < -\omega_s \text{ then } \omega_1 = -\omega_s \\ \text{If } \omega_2 < -\omega_s \text{ then } \omega_2 = -\omega_s \\ \text{If } \omega_1 > \omega_s \text{ then } \omega_1 = \omega_s \\ \text{If } \omega_2 > \omega_s \text{ then } \omega_2 = \omega_s \\ \text{If } \omega_1 > \omega_2 \text{ then } \omega_1 = \omega_2\end{gathered} \qquad (55)$$

The above conditionals insure numerical stability of the application of Eq. 47 as well as correctly computing any theoretical quantities of solar radiation early and late in the day. The user should recognize that Eqs. 47-55 presume an infinitely flat ground surface and locate 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:

$$S_c = 0.1645 \sin(2\,b) - 0.1255 \cos(b) - 0.025 \sin(b) \qquad (56)$$

$$b = \frac{2\pi(J-81)}{364} \qquad (57)$$

where J is the number of the day in the year.

The sunset hour angle, ω_s, is given by:

$$\omega_s = \arccos\left[-\tan(\varphi)\tan(\delta)\right] \qquad (58)$$

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:

$$\omega_s = \frac{\pi}{2} - \arctan\left[\frac{-\tan(\varphi)\tan(\delta)}{X^{0.5}}\right] \tag{59}$$

where:

$$X = 1 - [\tan(\varphi)]^2 [\tan(\delta)]^2 \tag{60}$$

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

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 will 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.

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 does correlate well with net radiation 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[19].

For the standardized short reference ET_{os} :

$$G_{hr\,daytime} = 0.1\ R_n \tag{61a}$$

$$G_{hr\,nighttime} = 0.5\ R_n \tag{61b}$$

For the standardized tall reference ET_{rs} :

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

$$G_{hr\,daytime} = 0.04\ R_n \tag{62a}$$

$$G_{hr\,nighttime} = 0.2\ R_n \tag{62b}$$

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}. For standardization, nighttime is defined as when measured or calculated hourly net radiation R_n is < 0 (i.e., negative).

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 not measured at that height must be adjusted. To adjust wind speed data to the 2-m height, Eq. 63 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.

$$u_2 = u_z \frac{4.87}{\ln(67.8\ z_w - 5.42)} \tag{63}$$

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. 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.

DEFINITION AND APPLICATION OF CROP COEFFICIENTS

Calculation of crop evapotranspiration ($ET_{c)}$ requires the selection of the correct 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 to be calculated as shown in Eq. 64 as.

$$ET_c = K_{co} * ET_{os} \quad \underline{\text{or}} \quad ET_c = K_{cr} * ET_{rs} \tag{64}$$

TRANSFER AND CONVERSION OF CROP COEFFICIENTS

Crop coefficients (K_c) and landscape coefficients available in the literature are referenced to either clipped 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. An alfalfa reference crop is defined 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 equation or 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 predict 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 are 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).

Some grass and alfalfa based crop coefficients are “mean” crop coefficients (e.g., Wright, 1979). 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 errors.

The publications referenced in the above paragraphs contain descriptions on determination and application of crop coefficients during growing periods. This information will not be repeated 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 is generally less than reference ET because plants may become dormant during non-growing periods and therefore may have substantially increased surface resistance. Besides increased 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 vegetation surfaces. 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 a type of evaporative index. However, the user must be aware that 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 ($mm\ m^{-1}\ s\ d^{-1}\ ^oC\ kPa^{-1}$ or $mm\ m^{-1}\ s\ h^{-1}\ ^oC\ kPa^{-1}$)
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} $^oC^{-1}$)
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)
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})
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)
ϕ	"phi" = angle of the sun above the horizon (radians)
ρ_a	"rho" = air density (Kg 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)