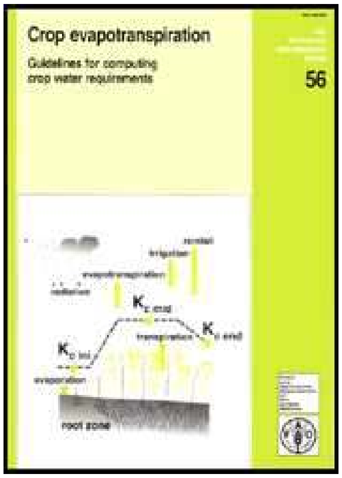
Crop evapotranspiration - Guidelines for computing crop water requirements - FAO Irrigation and drainage paper 56



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Evapotranspiration (ET)

The combination of two separate processes whereby water is lost on the one hand from the soil surface by evaporation and on the other hand from the crop by transpiration is referred to as evapotranspiration (ET).

Evaporation

Evaporation is the process whereby liquid water is converted to water vapour (vaporization) and removed from the evaporating surface (vapour removal). Water evaporates from a variety of surfaces, such as lakes, rivers, pavements, soils and wet vegetation.

Transpiration

Transpiration consists of the vaporization of liquid water contained in plant tissues and the vapour removal to the atmosphere. Crops predominately lose their water through stomata. These are small openings on the plant leaf through which gases and water vapour pass

Evapotranspiration (ET)

Evaporation and transpiration occur simultaneously and there is no easy way of distinguishing between the two processes. Apart from the water availability in the topsoil, the evaporation from a cropped soil is mainly determined by the fraction of the solar radiation reaching the soil surface. This fraction decreases over the growing period as the crop develops and the crop canopy shades more and more of the ground area. When the crop is small, water is predominately lost by soil evaporation, but once the crop is well developed and completely covers the soil, transpiration becomes the main process

Factors affecting evapotranspiration

Weather parameters, crop characteristics, management and environmental aspects are factors affecting evaporation and transpiration. The related ET concepts presented in Figure 3 are discussed in the section on evapotranspiration concepts.

Weather parameters

The principal weather parameters affecting evapotranspiration are radiation, air temperature, humidity and wind speed. The evaporation power of the atmosphere is expressed by the reference crop evapotranspiration (ET_o). The reference crop evapotranspiration represents the evapotranspiration from a standardized vegetated surface.

Crop factors

The crop type, variety and development stage should be considered when assessing the evapotranspiration from crops grown in large, well-managed fields. Differences in resistance to transpiration, crop height, crop roughness, reflection, ground cover and crop rooting characteristics result in different ET levels in different types of crops under identical environmental conditions.

Management and environmental conditions

Factors such as soil salinity, poor land fertility, limited application of fertilizers, the presence of hard or impenetrable soil horizons, the absence of control of diseases and pests and poor soil management may limit the crop development and reduce the evapotranspiration. Other factors to be considered when assessing ET are ground cover, plant density and the soil water content. The effect of soil water content on ET is conditioned primarily by the magnitude of the water deficit and the type of soil. On the other hand, too much water will result in waterlogging which might damage the root and limit root water uptake by inhibiting respiration.

Reference crop evapotranspiration (ET_o)

The evapotranspiration rate from a reference surface, not short of water, is called the reference crop evapotranspiration or reference evapotranspiration and is denoted as ET_o. The reference surface is a hypothetical grass reference crop with specific characteristics. The use of other denominations such as potential ET is strongly discouraged due to ambiguities in their definitions.

The only factors affecting ET_o are climatic parameters. Consequently, ET_o is a climatic parameter and can be computed from weather data. ET_o expresses the evaporating power of the atmosphere at a specific location and time of the year and does not consider the crop characteristics and soil factors.

Crop evapotranspiration under standard conditions (ET_c)

The crop evapotranspiration under standard conditions, denoted as ET_c, is the evapotranspiration from disease-free, well-fertilized crops, grown in large fields, under optimum soil water conditions, and achieving full production under the given climatic conditions.

❋ET_c will be between 1 to 9 mm/day from cool to warm average temperature.

The amount of water required to compensate the evapotranspiration loss from the cropped field is defined as crop water requirement. Although the values for crop evapotranspiration and crop water requirement are identical, crop water requirement refers to the amount of water that needs to be supplied, while crop evapotranspiration refers to the amount of water that is lost through evapotranspiration. The irrigation water requirement basically represents the difference between the crop water requirement and effective precipitation. The irrigation water requirement also includes additional water for leaching of salts and to compensate for non-uniformity of water application.

Energy balance

The energy arriving at the surface must equal the energy leaving the surface for the same time period.

All fluxes of energy should be considered when deriving an energy balance equation. The equation for an evaporating surface can be written as:

$R_n - G - \lambda ET - H = 0$ (1)

where R_n is the net radiation, H the sensible heat, G the soil heat flux and λ ET the latent heat flux.

The latent heat flux (λ ET) representing the evapotranspiration fraction can be derived from the energy balance equation if all other components are known. Net radiation (R_n) and soil heat fluxes (G) can be measured or estimated from climatic parameters. Measurements of the sensible heat (H) are however complex and cannot be easily obtained. H requires accurate measurement of temperature gradients above the surface.

Penman-Monteith equation

The Penman-Monteith form of the combination equation is:

$$\lambda ET = \frac{\Delta(R_n - G) + \rho_a c_p \frac{(e_s - e_a)}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \quad (3)$$

where R_n is the net radiation, G is the soil heat flux, (e_s - e_a) represents the vapour pressure deficit of the air, ρ_a is the mean air density at constant pressure, c_p is the specific heat of the air, Δ represents the slope of the saturation vapour pressure temperature relationship, γ is the psychrometric constant, and r_s and r_a are the (bulk) surface and aerodynamic resistances.

Aerodynamic resistance (r_a)

The transfer of heat and water vapour from the evaporating surface into the air above the canopy is determined by the aerodynamic resistance:

$$r_a = \frac{\ln\left[\frac{z_m - d}{z_{om}}\right]\ln\left[\frac{z_h - d}{z_{oh}}\right]}{k^2 u_z} \quad (4)$$

where

r_a aerodynamic resistance [s m^{-1}],

z_m height of wind measurements [m],

z_h height of humidity measurements [m],

d zero plane displacement height [m],

z_{om} roughness length governing momentum transfer [m],

z_{oh} roughness length governing transfer of heat and vapour [m],

k von Karman's constant, 0.41 [-],

u_z wind speed at height z [m s^{-1}].

h crop height d = 2/3 h, $z_{om} = 0.123$ h

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

Assuming a constant crop height of 0.12 m and a standardized height for wind speed, temperature and humidity at 2 m ($z_m = z_h = 2$ m), the aerodynamic resistance r_a [s m^{-1}] for the grass reference surface becomes (Eq. 4):

$$r_a = \frac{\ln\left[\frac{2 - 2/3(0.12)}{0.123(0.12)}\right]\ln\left[\frac{2 - 2/3(0.12)}{(0.1)0.123(0.12)}\right]}{(0.41)^2 u_2} = \frac{208}{u_2}$$

(Bulk) surface resistance (r_s)

The 'bulk' surface resistance describes the resistance of vapour flow through the transpiring crop and evaporating soil surface. Where the vegetation does not completely cover the soil, the resistance factor should indeed include the effects of the evaporation from the soil surface. If the crop is not transpiring at a potential rate, the resistance depends also on the water status of the vegetation

$$r_s = \frac{r_l}{LAI_{active}} \quad (5)$$

where

r_s (bulk) surface resistance [s m^{-1}],

r_l bulk stomatal resistance of the well-illuminated leaf [s m^{-1}],

LAI_{active} active (sunlit) leaf area index [m^2 (leaf area) m^{-2} (soil surface)].

The LAI values for various crops differ widely but values of 3-5 are common for many mature crops. For a given crop, green LAI changes throughout the season and normally reaches its maximum before or at flowering (Figure 8). LAI further depends on the plant density and the crop variety.

A general equation for LAI_{active} is:

LAI_{active} = 0.5 LAI

which takes into consideration the fact that generally only the upper half of dense clipped grass is actively contributing to the surface heat and vapour transfer. For clipped grass a general equation for LAI is:

LAI = 24 h

where h is the crop height [m].

The stomatal resistance, r_l, of a single leaf has a value of about 100 s m^{-1} under well-watered conditions. By assuming a crop height of 0.12 m, the surface resistance, r_s [s m^{-1}], for the grass reference surface becomes (Eq. 5):

$$r_s = \frac{100}{0.5(24)(0.12)} \approx 70 \text{ s m}^{-1}$$

Reference surface

To obviate the need to define unique evaporation parameters for each crop and stage of growth, the concept of a reference surface was introduced. Evapotranspiration rates of the various crops are related to the evapotranspiration rate from the reference surface (ET_o) by means of crop coefficients

The FAO Expert Consultation on Revision of FAO Methodologies for Crop Water Requirements accepted the following unambiguous definition for the reference surface:

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

The reference surface closely resembles an extensive surface of green grass of uniform height, actively growing, completely shading the ground and with adequate water. The requirements that the grass surface should be extensive and uniform result from the assumption that all fluxes are one-dimensional upwards.

FAO Penman-Monteith equation

From the original Penman-Monteith equation (Equation 3) and the equations of the aerodynamic (Equation 4) and surface resistance (Equation 5), the FAO Penman-Monteith method to estimate ET_o can be derived (Box 6):

$$ET_o = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34 u_2)} \qquad (6)$$

where

ET_o reference evapotranspiration [mm day^{-1}],

R_n net radiation at the crop surface [MJ m^{-2} day^{-1}],

G soil heat flux density [MJ m^{-2} day^{-1}],

T mean daily air temperature at 2 m height [°C],

u_2 wind speed at 2 m height [m s^{-1}],

e_s saturation vapour pressure [kPa],

e_a actual vapour pressure [kPa],

$e_s - e_a$ saturation vapour pressure deficit [kPa],

Δ slope vapour pressure curve [kPa °C^{-1}],

γ psychrometric constant [kPa °C^{-1}].

The equation uses standard climatological records of solar radiation (sunshine), air temperature, humidity and wind speed. To ensure the integrity of computations, the weather measurements should be made at 2 m (or converted to that height) above an extensive surface of green grass, shading the ground and not short of water.

Atmospheric pressure (P)

$$P = 101.3\left(\frac{293 - 0.0065z}{293}\right)^{5.26} \quad (7)$$

where

P atmospheric pressure [kPa],

z elevation above sea level [m],

Psychrometric constant (γ)

The psychrometric constant, γ, is given by:

$$\gamma = \frac{c_p P}{\varepsilon \lambda} = 0.665 \times 10^{-3} P \quad (8)$$

where

γ psychrometric constant [kPa °C^{-1}],

P atmospheric pressure [kPa],

λ latent heat of vaporization, 2.45 [MJ kg^{-1}],

c_p specific heat at constant pressure, 1.013 10^{-3} [MJ kg^{-1} °C^{-1}],

ε ratio molecular weight of water vapour/dry air = 0.622.

Air temperature

For standardization, T_{mean} for 24-hour periods is defined as the mean of the daily maximum (T_{max}) and minimum temperatures (T_{min}) rather than as the average of hourly temperature measurements.

$$T_{mean} = \frac{T_{max} - T_{min}}{2} \quad (9)$$

It is dimensionless and is commonly given as a percentage.

K = °C + 273.16

Relative humidity

The relative humidity (RH) expresses the degree of saturation of the air as a ratio of the actual (e_a) to the saturation (e°(T)) vapour pressure at the same temperature (T):

$$RH = 100\frac{e_a}{e°(T)} \quad (10)$$

Mean saturation vapour pressure (e_s)

As saturation vapour pressure is related to air temperature, it can be calculated from the air temperature. The relationship is expressed by:

$$e°(T) = 0.6108 \exp\left[\frac{17.27T}{T + 237.3}\right] \quad (11)$$

where

e°(T) saturation vapour pressure at the air temperature T [kPa],

T air temperature [°C],

exp[..] 2.7183 (base of natural logarithm) raised to the power [..].

$$e_s = \frac{e°(T_{max}) + e°(T_{min})}{2} \quad (12)$$

Actual vapour pressure (e_a) derived from dewpoint temperature

$$e_a = e°(T_{dew}) = 0.6108 \exp\left[\frac{17.27\, T_{dew}}{T_{dew} + 237.3}\right] \quad (14)$$

dewpoint temperature (T_{dew}) [°C],

Actual vapour pressure (e_a) derived from psychrometric data

The relationship is expressed by the following equation:

$e_a = e° (T_{wet}) - \gamma_{psy} (T_{dry} - T_{wet})$ (15)

where

e_a actual vapour pressure [kPa],

a_{psy} = 0.000662 for ventilated (Asmann type) psychrometers, with an air movement of some 5 m/s,
0.000800 for natural ventilated psychrometers (about 1 m/s),
0.001200 for non-ventilated psychrometers installed indoors.

$e°(T_{wet})$ saturation vapour pressure at wet bulb temperature [kPa],

γ_{psy} psychrometric constant of the instrument [kPa °C^{-1}],

T_{dry}-T_{wet} wet bulb depression, with T_{dry} the dry bulb and T_{wet} the wet bulb temperature [°C].

Actual vapour pressure (e_a) derived from relative humidity data

• **For RH_{max} and RH_{min}:**

$$e_a = \frac{e°(T_{min}) \frac{RH_{max}}{100} + e°(T_{max}) \frac{RH_{min}}{100}}{2} \quad (17)$$

where

e_a actual vapour pressure [kPa],

$e°(T_{min})$ saturation vapour pressure at daily minimum temperature [kPa],

$e°(T_{max})$ saturation vapour pressure at daily maximum temperature [kPa],

RH_{max} maximum relative humidity [%],

RH_{min} minimum relative humidity [%].

For RH_{max}:

When using equipment where errors in estimating RH_{min} can be large, or when RH data integrity are in doubt, then one should use only RH_{max}:

$$e_a = e°(T_{min}) \frac{RH_{max}}{100} \quad (18)$$

• **For RH_{mean}:**

In the absence of RH_{max} and RH_{min}, another equation can be used to estimate e_a:

$$e_a = \frac{RH_{mean}}{100}\left[\frac{e°(T_{max}) + e°(Tmin)}{2}\right] \quad (19)$$

where RH_{mean} is the mean relative humidity, defined as the average between RH_{max} and RH_{min}.

Vapour pressure deficit (e_s - e_a)

The vapour pressure deficit is the difference between the saturation (e_s) and actual vapour pressure (e_a) for a given time period. For time periods such as a week, ten days or a month e_s is computed from Equation 12 using the T_{max} and T_{min} averaged over the time period and similarly the e_a is computed with one of the equations 4 to 19, using average measurements over the period.

Slope of saturation vapour pressure curve (Δ)

For the calculation of evapotranspiration, the slope of the relationship between saturation vapour pressure and temperature, Δ, is required. The slope of the curve (Figure 11) at a given temperature is given by.

$$\Delta = \frac{4098\left[0.6108\exp\left(\frac{17.27T}{T+237.3}\right)\right]}{(T+237.3)^2} \quad (13)$$

where

Δ slope of saturation vapour pressure curve at air temperature T [kPa °C^{-1}],

T air temperature [°C],

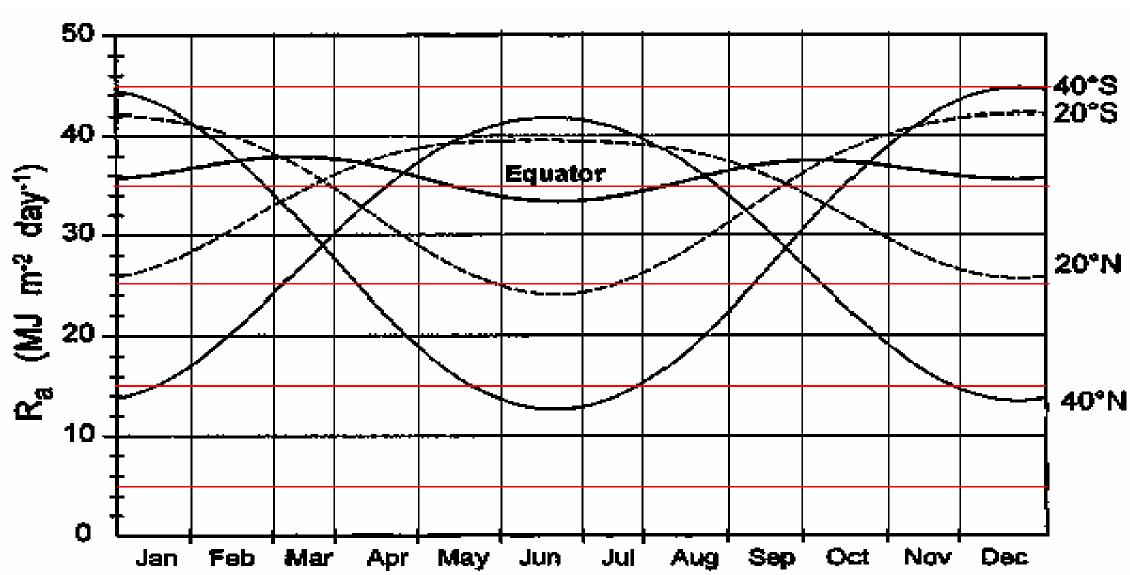
exp[..] 2.7183 (base of natural logarithm) raised to the power [..].

Values of slope Δ for different air temperatures are given in Annex 2 (Table 2.4). In the FAO Penman-Monteith equation, where Δ occurs in the numerator and denominator, the slope of the vapour pressure curve is calculated using mean air temperature.

Extraterrestrial radiation (R_a)

The radiation striking a surface perpendicular to the sun's rays at the top of the earth's atmosphere, called the solar constant, is about 0.082 MJ m^{-2} min^{-1}. The local intensity of radiation is, however, determined by the angle between the direction of the sun's rays and the normal to the surface of the atmosphere. This angle will change during the day and will be different at different latitudes and in different seasons. The solar radiation received at the top of the earth's atmosphere on a horizontal surface is called the extraterrestrial (solar) radiation, R_a.

FIGURE 13. Annual variation in extraterrestrial radiation (R_a) at the equator, 20 and 40° north and south



The extraterrestrial radiation, R_a, for each day of the year and for different latitudes can be estimated from the solar constant, the solar declination and the time of the year by:

$$R_a = \frac{24(60)}{\pi} G_{sc} d_r [\omega_s \sin(\varphi) \sin(\delta) + \cos(\varphi) \cos(\delta) \sin(\omega_s)] \quad (21)$$

where

R_a extraterrestrial radiation [MJ m^{-2} day^{-1}], G_{sc} solar constant = 0.0820 MJ m^{-2} min^{-1},

d_r inverse relative distance Earth-Sun (Equation 23), ω_s sunset hour angle (Equation 25 or 26) [rad],

φ latitude [rad] (Equation 22), δ solar decimation (Equation 24) [rad].

The inverse relative distance Earth-Sun, d_r, and the solar declination, δ, are given by:

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

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

where J is the number of the day in the year between 1 (1 January) and 365 or 366 (31 December).

The sunset hour angle, ω_s, is given by:

ω_s = arccos [-tan (φ) tan (δ)] (25)

Extraterrestrial radiation for hourly or shorter periods (R_a)

For hourly or shorter periods the solar time angle at the beginning and end of the period should be considered when calculating R_a:

$$R_a = \frac{12(60)}{\pi} G_{sc} d_r [(\omega_2 - \omega_1) \sin(\varphi) \sin(\delta) + \cos(\varphi) \cos(\delta)(\sin(\omega_2) - \sin(\omega_1))] \quad (28)$$

where

R_a extraterrestrial radiation in the hour (or shorter) period [MJ m^{-2} $hour^{-1}$],

G_{sc} solar constant = 0.0820 MJ m^{-2} min^{-1},

d_r inverse relative distance Earth-Sun (Equation 23),

δ solar declination [rad] (Equation 24),

φ latitude [rad] (Equation 22),

ω_1 solar time angle at beginning of period [rad] (Equation 29),

ω_2 solar time angle at end of period [rad] (Equation 30).

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

$$\omega_1 = \omega - \frac{\pi t_1}{24} \quad (29)$$

$$\omega_2 = \omega + \frac{\pi t_1}{24} \quad (30)$$

where

ω solar time angle at midpoint of hourly or shorter period [rad],

t_1 length of the calculation period [hour]: i.e., 1 for hourly period or 0.5 for a 30-minute period.

The solar time angle at midpoint of the period is:

$$\omega = \frac{\pi}{12}[(t + 0.06667(L_z - L_m) + S_c) - 12] \quad (31)$$

where

t standard clock time at the midpoint of the period [hour]. For example for a period between 14.00 and 15.00 hours, t = 14.5,

L_z longitude of the centre of the local time zone [degrees west of Greenwich]. For example, L_z = 75, 90, 105 and 120° for the Eastern, Central, Rocky Mountain and Pacific time zones (United States) and L_z = 0° for Greenwich, 330° for Cairo (Egypt), and 255° for Bangkok (Thailand),

L_m longitude of the measurement site [degrees west of Greenwich],

S_c seasonal correction for solar time [hour].

Of course, $\omega < -\omega_s$ or $\omega > \omega_s$ from Equation 31 indicates that the sun is below the horizon so that, by definition, R_a is zero.

The seasonal correction for solar time is:

$S_c = 0.1645 \sin(2b) - 0.1255 \cos(b) - 0.025 \sin(b)$ (32)

$$b = \frac{2\pi(J - 81)}{364} \quad (33)$$

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

Daylight hours (N)

The daylight hours, N, are given by:

$$N = \frac{24}{\pi}\omega_s \quad (34)$$

where ω_s is the sunset hour angle in radians given by Equation 25 or 26. Mean values for N

Solar radiation (R_s)

If the solar radiation, R_s, is not measured, it can be calculated with the Angstrom formula which relates solar radiation to extraterrestrial radiation and relative sunshine duration:

$$R_s = \left(a_s + b_s \frac{n}{N}\right) R_a \quad (35)$$

where

R_s solar or shortwave radiation [MJ m^{-2} day^{-1}],

n actual duration of sunshine [hour],

N maximum possible duration of sunshine or daylight hours [hour],

n/N relative sunshine duration [-],

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

a_s regression constant, expressing the fraction of extraterrestrial radiation reaching the earth on overcast days (n = 0),

a_s+b_s fraction of extraterrestrial radiation reaching the earth on clear days (n = N).

R_s is expressed in the above equation in MJ m^{-2} day^{-1}. The corresponding equivalent evaporation in mm day^{-1} is obtained by multiplying R_s by 0.408 (Equation 20). Depending on atmospheric conditions (humidity, dust) and solar declination (latitude and month), the Angstrom values a_s and b_s will vary. Where no actual solar radiation data are available and no calibration has been carried out for improved a_s and b_s parameters, the values a_s = 0.25 and b_s = 0.50 are recommended.

Clear-sky solar radiation (R_{so})

The calculation of the clear-sky radiation, R_{so}, when n = N, is required for computing net longwave radiation.

• **For near sea level or when calibrated values for a_s and b_s are available:**

$R_{so} = (a_s+b_s)R_a$ (36)

where

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

a_s+b_s fraction of extraterrestrial radiation reaching the earth on clear-sky days (n = N).

Net solar or net shortwave radiation (R_{ns})

The net shortwave radiation resulting from the balance between incoming and reflected solar radiation is given by:

$R_{ns} = (1-\alpha)R_s$ (38)

where

R_{ns} net solar or shortwave radiation [MJ m^{-2} day^{-1}],

α albedo or canopy reflection coefficient, which is 0.23 for the hypothetical grass reference crop [dimensionless],

R_s the incoming solar radiation [MJ m^{-2} day^{-1}].

R_{ns} is expressed in the above equation in MJ m^{-2} day^{-1}.

Net longwave radiation (R_{nl})

The rate of longwave energy emission is proportional to the absolute temperature of the surface raised to the fourth power. This relation is expressed quantitatively by the Stefan-Boltzmann law. The net energy flux leaving the earth's surface is, however, less than that emitted and given by the Stefan-Boltzmann law due to the absorption and downward radiation from the sky. Water vapour, clouds, carbon dioxide and dust are absorbers and emitters of longwave radiation. Their concentrations should be known when assessing the net outgoing flux. As humidity and cloudiness play an important role, the Stefan-Boltzmann law is corrected by these two factors when estimating - the net outgoing flux of longwave radiation. It is thereby assumed that the concentrations of the other absorbers are constant:

$$R_{nl} = \sigma\left[\frac{T_{max}K^4 + T_{min}K^4}{2}\right]\left(0.34 - 0.14\sqrt{e_a}\right)\left(1.35\frac{R_s}{R_{so}} - 0.35\right) \quad (39)$$

where

R_{nl} net outgoing longwave radiation [MJ m^{-2} day^{-1}],

σ Stefan-Boltzmann constant [4.903 10^{-9} MJ K^{-4} m^{-2} day^{-1}],

$T_{max,\,K}$ maximum absolute temperature during the 24-hour period [K = °C + 273.16],

$T_{min,\,K}$ minimum absolute temperature during the 24-hour period [K = °C + 273.16],

e_a actual vapour pressure [kPa],

R_s/R_{so} relative shortwave radiation (limited to ≤ 1.0),

R_s measured or calculated. (Equation 35) solar radiation [MJ m^{-2} day^{-1}],

R_{so} calculated (Equation 36 or 37) clear-sky radiation [MJ m^{-2} day^{-1}].

An average of the maximum air temperature to the fourth power and the minimum air temperature to the fourth power is commonly used in the Stefan-Boltzmann equation for 24-hour time steps. The term $(0.34-0.14\sqrt{e_a})$ expresses the correction for air humidity, and will be smaller if the humidity increases. The effect of cloudiness is expressed by $(1.35\ R_s/R_{so} - 0.35)$. The term becomes smaller if the cloudiness increases and hence R_s decreases. The smaller the correction terms, the smaller the net outgoing flux of longwave radiation. Note that the R_s/R_{so} term in Equation 39 must be limited so that $R_s/R_{so} \leq 1.0$.

Where measurements of incoming and outgoing short and longwave radiation during bright sunny and overcast hours are available, calibration of the coefficients in Equation 39 can be carried out.

Net radiation **(*R_n*)**

The net radiation (R_n) is the difference between the incoming net shortwave radiation (R_{ns}) and the outgoing net longwave radiation (R_{nl}):

$R_n = R_{ns} - R_{nl}$ (40)

Soil heat flux **(*G*)**

Complex models are available to describe soil heat flux. Because soil heat flux is small compared to R_n, particularly when the surface is covered by vegetation and calculation time steps are 24 hours or longer, a simple calculation procedure is presented here for long time steps, based on the idea that the soil temperature follows air temperature

- **For day and ten-day periods:**

As the magnitude of the day or ten-day soil heat flux beneath the grass reference surface is relatively small, it may be ignored and thus:

$G_{day} \approx 0$ (42)

For hourly or shorter periods:

For hourly (or shorter) calculations, G beneath a dense cover of grass does not correlate well with air temperature. Hourly G can be approximated during daylight periods as:

$G_{hr} = 0.1\ R_n$ (45)

and during nighttime periods as:

$G_{hr} = 0.5\ R_n$ (46)

Where the soil is warming, the soil heat flux G is positive. The amount of energy required for this process is subtracted from R_n when estimating evapotranspiration.

Wind profile relationship

Wind speeds measured at different heights above the soil surface are different. Surface friction tends to slow down wind passing over it. Wind speed is slowest at the surface and increases with height. For this reason anemometers are placed at a chosen standard height, i.e., 10 m in meteorology and 2 or 3 m in agrometeorology. For the calculation of evapotranspiration, wind speed measured at 2 m above the surface is required. To adjust wind speed data obtained from instruments placed at elevations other than the standard height of 2m, a logarithmic wind speed profile may be used for measurements above a short grassed surface:

$$u_2 = u_z \frac{4.87}{\ln(67.8z - 5.42)} \quad (47)$$

where

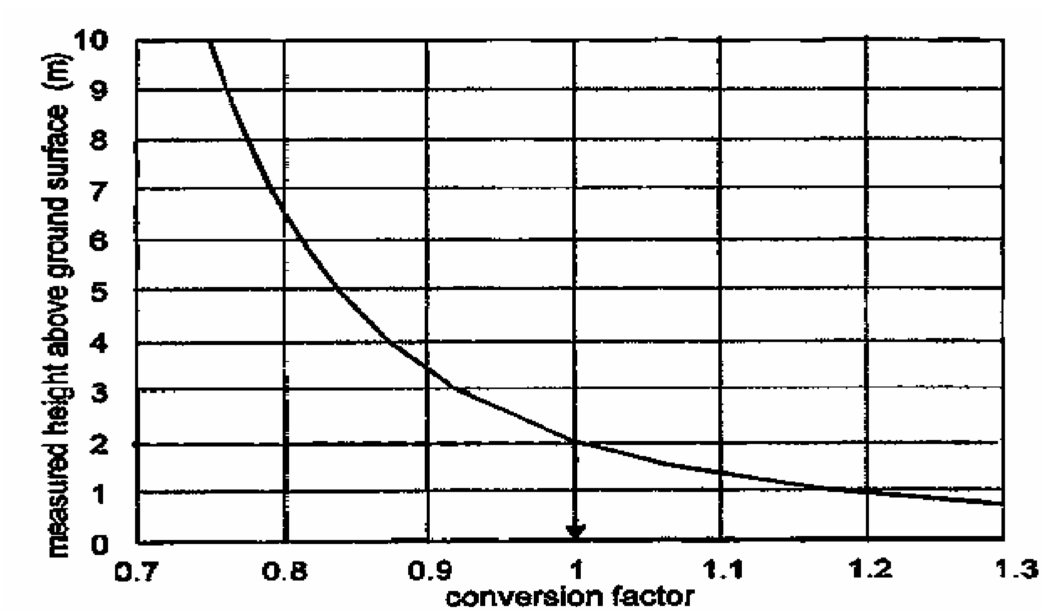
u_2 wind speed at 2 m above ground surface [m s^{-1}],

u_z measured wind speed at z m above ground surface [m s^{-1}],

z height of measurement above ground surface [m].

The corresponding multipliers or conversion factors are given in Annex 2 (Table 2.9) and are plotted in Figure 16.

FIGURE 16. Conversion factor to convert wind speed measured at a certain height above ground level to wind speed at the standard height (2 m)



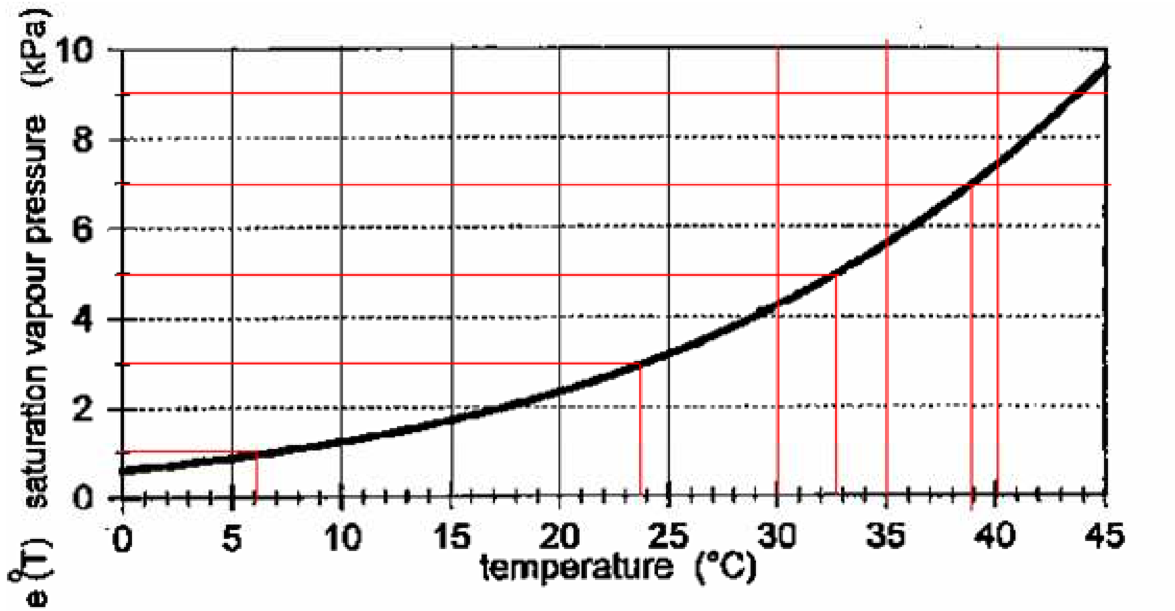


FIGURE 11. Saturation vapour pressure shown as a function of temperature: e°(T) curve

FIGURE 12. Variation of the relative humidity over 24 hours for a constant actual vapour pressure of 2.4 kPa

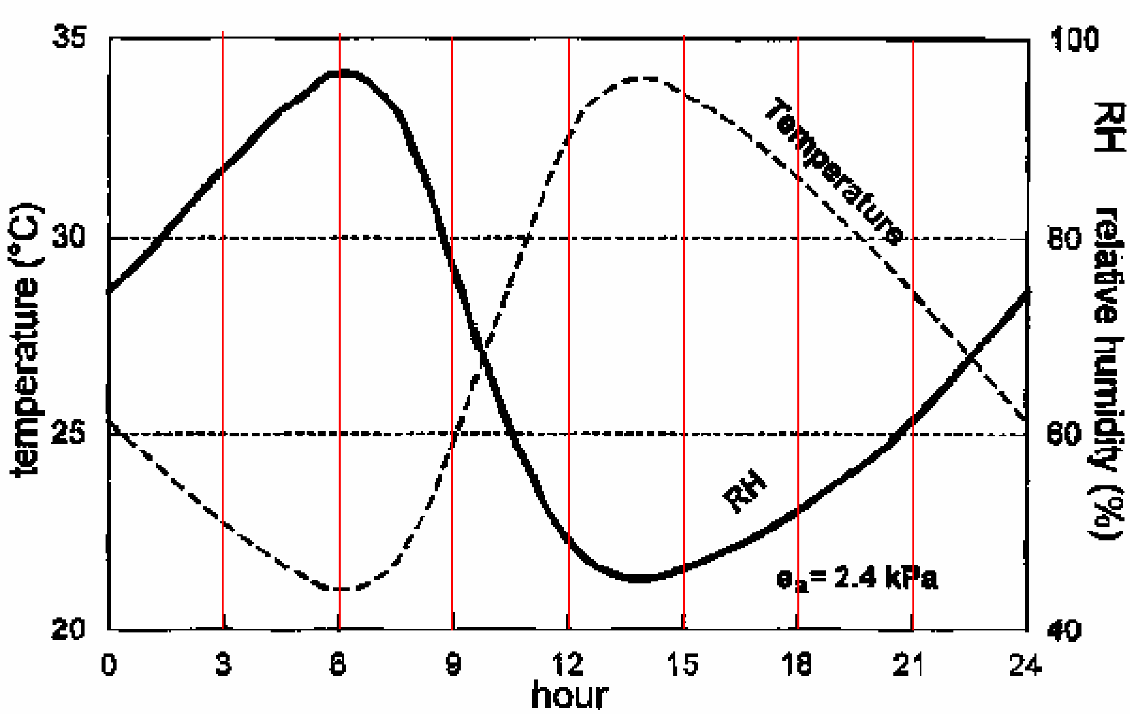


FIGURE 14. Annual variation of the daylight hours (N) at the equator, 20 and 40° north and south

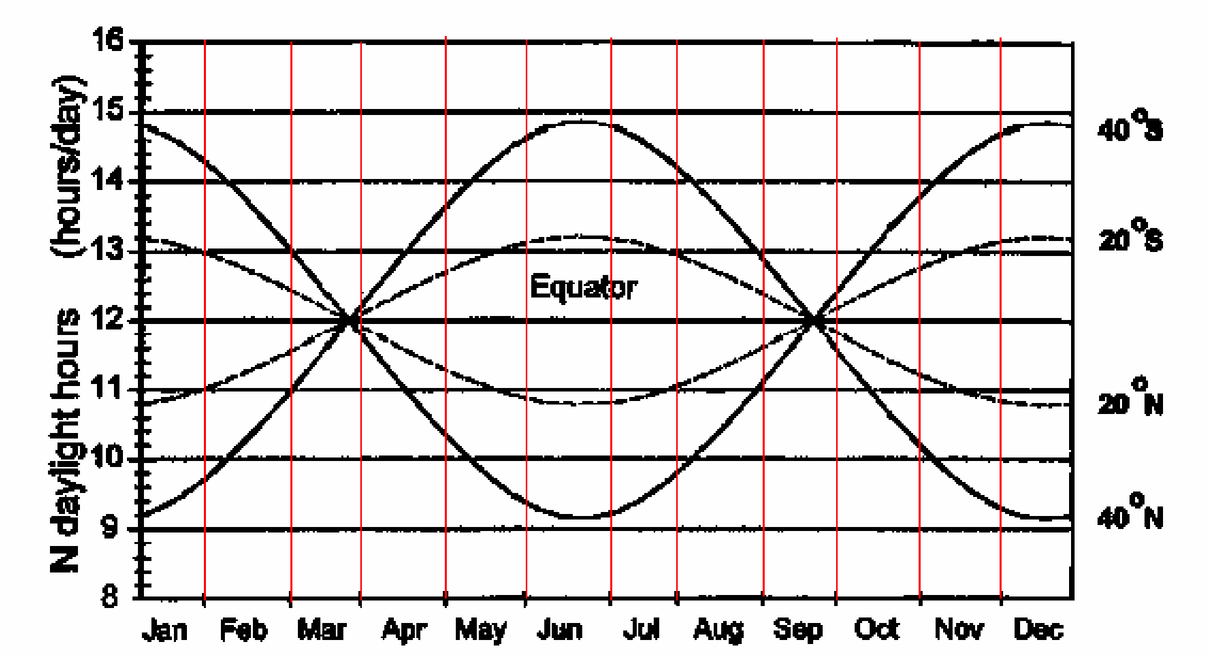


TABLE 1. Conversion factors for Evapotranspiration

	depth	volume per unit area		energy per unit area *
	$mm\ day^{-1}$	$m^3\ ha^{-1}\ day^{-1}$	$l\ s^{-1}\ ha^{-1}$	$MJ\ m^{-2}\ day^{-1}$
1 $mm\ day^{-1}$	1	10	0.116	2.45
1 $m^3\ ha^{-1}\ day^{-1}$	0.1	1	0.012	0.245
1 $l\ s^{-1}\ ha^{-1}$	8.640	86.40	1	21.17
1 $MJ\ m^{-2}\ day^{-1}$	0.408	4.082	0.047	1

* For water with a density of 1000 kg m^{-3} and at 20°C.