EVAPOTRANSPIRATION

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Evaporation and transpiration, that transfer water from land and vegetation to the atmosphere, are key components of the hydrologic cycle, especially at large space-time scales. These components are fundamental to agricultural irrigation, forest management, land use and land cover design, drought management, and water resources planning.

1 INTRODUCTION

Large amounts of water are evaporated from ponds and lakes, particularly in arid climates. Crops and trees transpire large portions of water that is applied through irrigation or is received through rainfall. Evaporation and transpiration together constitute what is called evapotranspiration (ET). The importance of ET depends upon the time scale and the type of hydrologic event. For an intense short storm event, abstractions due to evaporation and transpiration are usually small and often neglected in hydrologic balance computations. The bulk of evaporation and transpiration from a catchment takes place during the time between runoff events, which is usually long. ET links atmospheric, hydrologic, and agricultural processes.

At the catchment scale, ET represents the largest water flux after precipitation, but it is very challenging to measure ET at this scale due to the heterogeneity of the topography. In practice, ET is often calculated as the residual from precipitation and other water fluxes (e.g., runoff, change in soil water storage) and is subject to large errors at a short temporal scale and in watersheds with poorly defined surface and subsurface watershed boundaries.

1.1 Definitions

Evaporation is defined as the process by which water from water bodies (rivers, ponds, lakes) or soil is converted to vapor. It is commonly expressed in mm/hour.

Potential Evaporation is the quantity of water evaporated per unit time per unit area of an idealized extensive free water surface under existing atmospheric conditions. The time scale for measurement or estimation can be an hour, a day, a month, or a year. It is usually assumed uniform over the area under consideration and is expressed in mm, cm, or inches. Thus, it is commonly expressed in mm/day, cm/month, in/day.

Reference Evapotranspiration: evapotranspiration from a reference surface, not short of water, is called the reference evapotranspiration and is denoted as ET_o (UN Food and Agriculture Organization, www.fao.org). The reference surface is a hypothetical grass reference crop with specific characteristics. It is commonly expressed in mm/day or cm/month.

1.2 Evaporation Process

Since the conversion of liquid water into water vapor requires energy, the rate of evaporation depends on the availability of energy and also how easily water vapors can diffuse in the air. Hence, evaporation involves transfer of energy and hence the molecules that have energy to break all bonds with their neighbors and move into the air will evaporate. The heat required (in calories, or cal) to convert 1 gram of water at 1 atmosphere (atm) pressure to vapor is the latent heat of vaporization. Its value is 539 cal/gm at 1 atm pressure at 100°C which means that 539 cal of heat is required to evaporate 1 gram of water. The heat energy required to raise the temperature of water to its boiling point is less than that required to convert the liquid water to vapor. The water vapor is lighter than other atmospheric gases and hence it tends to rise above air.

When warmed, water molecules become more active and move from the liquid water to the atmosphere. As more molecules move into the atmosphere, the density of water molecules near the water surface increases. This leads to collision between molecules emerging from the water with those already in the air and it causes some water molecules to return to the water. Thus, there is a constant transfer of molecules to and from water, but the transfer from the water is dominant. Some molecules are carried away from the water body by wind. At some time, the air may become saturated and the vapor pressure of air may be equal to the vapor pressure of water. In this condition, evaporation becomes zero. Evaporation is high when the vapor pressure of air is low which means that the gradient between the vapor pressure of air and that of water is low. The difference in the vapor pressure gradient between the water surface and the air varies with the height above the water surface.

Heat energy, essential for evaporation, is derived generally from solar radiation. The relative humidity of air and the wind velocity across the water surface also influence the rate of evaporation. Near the equator, where the sea and winds are warm, the rate of evaporation is quite high. Tropical climates are typically humid, as is the case in the East Indies, the Amazon River basin, and in central Africa. Near the poles, evaporation is small since the cold winds cannot hold much moisture and weak sunshine contains little energy. Arctic climates are typically dry.

2.0 FACTORS AFFECTING EVAPORATION

The most important factors affecting evaporation are: (a) temperature, (b) solar radiation, (c) relative humidity, (d) vapor pressure difference (gradient), (e) wind speed, (f) atmospheric pressure, (g) water quality, and (h) water depth and soil type.

2.1 Temperature

Temperature of a place depends on its geographical location, altitude, season and distance from the sea or ocean. Temperature is influenced by the exchange of air masses and by cloudiness (which obstructs incoming radiation) and follows diurnal and seasonal cycles. Temperature falls with altitude at approximately 0.6^{o}C per 100 m rise in elevation for moist air and 0.9^{o}C per 100 m for dry air. Hence, evaporation is less at higher altitudes. The temperature variation near large water bodies, such as sea, is less due to the moderating influence and hence the annual and diurnal range is smaller. Generally, temperatures of nearby places are strongly correlated.

Temperature also influences the amount of dissolved gases and the rate of chemical and biological reactions and activities. As the temperature increases, the kinetic energy of water molecules also increases and this allows water molecules to escape from liquid water to the air faster than otherwise. Thus, the warmer the water is, the greater is the transfer of molecules from the water to the air and evaporation. Temperature is commonly reported in degree Centigrade; USA and some other countries follow the Fahrenheit unit.

Temperature is manually measured by a set of four thermometers which are usually installed in the Stevenson screen which provides ventilation and shade. The screen should face towards north. Four types of thermometers are used: dry bulb thermometer to measure ambient air temperature; wet bulb to measure temperature that is attained by a volume of air if cooled adiabatically to saturation by evaporation of water into it (this temperature is used to calculate relative humidity); maximum thermometers to measure the highest temperature reached since the last setting; and minimum thermometers to measure the lowest temperature reached since the last setting.

To measure water surface temperature, several types of thermometers, such as mercury-in-glass or mercury-in-steel (including maximum and minimum and reversing thermometer), platinum resistance or thermistor elements with electronic circuit and meter or recorder and thermocouple thermometers, are used. In general, the precision required for water temperature is ±0.1°C. However, precision of ±0.5°C is adequate in many circumstances.

2.2 Solar Radiation

The instruments used to measure the incident solar short-wave radiation are called pyranometer. These pyranometers are based on multi-junction thermopiles; glass domes of these devices allow radiation in the 0.3–3 μm range only to reach the pyranometer surface. A net pyradiometer measures the difference between total (short-wave and long-wave) incoming (downward) and outgoing(upward) radiation (WMO 2008). It consists of a horizontally mounted plate with two blackened surfaces. Half of the junctions of a thermophile are attached to the upper surface and the others are attached to the lower surface, so that the thermopile output is proportional to net radiation in the0.3–100 μm band. WMO (2008) recommends that the instruments should be mounted at least 1 m above the representative vegetation cover.

2.3 Sunshine Duration

The sunshine duration is an indicator of radiation received at a place. The potential maximum sunshine duration is a function of latitude and longitude and season; the actual number of sunshine hours varies due to clouds, fog, etc. Sunshine duration is commonly measured by the Campbell Stokes sunshine recorder (Fig. 1) which consists of a glass sphere mounted on a section of a spherical bowl. The glass sphere focuses Sun's rays on a paper card which is mounted on the bowl and on which hours are indicated. When the sun is shining, its rays burn the card. The card is changed daily after sunset. The burnt traces on the card indicate the sunshine duration and data at the required time resolution is read from the card. Note that the sunshine recorder uses the movement of the sun instead to form the time base of the record. Different grooves in the bowl must be used in winter and summer and different card types be used for each season.

Fig. 1 Sunshine hour recorder.

6.2.4 Relative Humidity

Relative humidity (RH) in % can be defined as the ratio of the actual vapor pressure of air to the saturation vapor pressure at the same pressure and temperature. Due to availability of moisture, places close to sea have higher RH and a smaller daily variation than inland locations. The relative humidity does not vary rapidly with time.

RH is most commonly estimated by jointly measuring dry bulb and wet bulb temperatures and then calculating the dew point temperature and actual and saturated vapor pressures; it is also calculated from the wet bulb depression (difference between wet and dry bulb readings) using a set of tables. While the actual vapor pressure may vary little during the day, RH has a regular diurnal pattern with a minimum normally coinciding with the highest temperature (when the saturation vapor pressure is highest). RH also shows a seasonal variation.

Relative humidity may also be measured continuously by means of hygrograph where the sensor is human/horse hair whose length changes with relative humidity. Errors in the hygrograph may arise from bad calibration. Errors in RH may be detected by setting up upper and lower warning limits, depending upon the station and season. The maximum value for RH is 100%. The graph of daily series needs to be inspected to identify anomalous values.

6.2.5 Wind Speed

The wind speed at a place is controlled by the pressure gradient which depends on temperature, topographic features, and land use. The wind speed shows a wide variation with place and time. It is commonly measured by the use of a cup type anemometer (Fig. 2) – cups of this device rotate due to the pressure exerted by winds. The rotations of cups over a time interval are measured, displayed by a counter and/or recorded by a data logger. These rotations indicate the average wind speed over the chosen time interval. The usual practice is to measure the wind speed over a three-minute period and this is considered as the instantaneous wind speed at that time. Likewise, by noting the counter reading at the beginning and end of a period, average wind speed for that period can be determined.

Fig. 2 Cup type anemometer

The direction of wind is reported as one of the 16 points of the compass; it could be shown either as numerical or as an alphabetic code. Note that wind direction is not used to compute ET – it is not a variable in formulas. Wind direction may influence evaporation at a place if the wind is blowing from a place with significantly different humidity.

Under certain conditions, a thin film of saturated vapor exists over a water surface. If left undisturbed, this film acts as an insolating buffer between the water surface and the unsaturated air above and slows down the evaporation to a low rate. Usually, this condition is not seen in real life, because wind tends to remove the saturated film and expose the water surface to unsaturated air. This action creates a vapor-pressure gradient that provides conditions that are favorable for evaporation. The relation between a water surface and wind drag on that surface is complex. Evaporation is believed to increase with wind velocity until the vapor-pressure gradient reaches some nearly constant relation. At this point, a further increase in wind velocity will not increase evaporation appreciably. The effect of increased wind velocity on evaporation is believed to be related to the size of the water body. Wind removes water vapor from small bodies of water rather quickly, whereas on large bodies of water, the effect of wind requires a longer time to accomplish the same effect.

Due to the effect of other factors, such as relative humidity, vapor pressure, and wind speed, the pan evaporation is not directly related to the temperature.

6.2.6 Atmospheric Pressure

Since air is less dense at lower atmospheric pressure, the likelihood of collision of water molecules escaping in the air with air molecules reduces with atmospheric pressure. For this reason, evaporation is more when atmospheric pressure is low. Two changes take place as one moves to higher altitudes: atmospheric pressure decreases and temperatures decreases. Consequently, the relation between altitude and evaporation is not straightforward. It has been observed that evaporation from lakes and evaporation pans increase as elevation increases.

3.0 MEASUREMENT OF EVAPORATION

Evaporation pans are most widely used devicesto measure evaporation. An evaporation pan is a small shallow vessel filled with water which is placed in an open area and exposed to weather. Water level in the pan is carefully measured to determine how much water (in terms of depth) is lost by evaporation. This loss is related to the evaporation from a pond or lake.

Various agencies have brought out standardization in pans by fixing their size, material, and measurement procedures to bring standardization in measuring evaporation. Such efforts have resulted in, for example, the U.S. Weather Bureau Class A Pan, the U.S. Geological Survey (USGS) Floating Pan, and the GGI-3000 pan from Russia. Among these pans, the U.S. Weather Bureau Class A Pan is the most commonly used instrument in a large number of countries throughout the world. WMO has recommended the use of this pan and it is described here.

3.1 The U.S. Weather Bureau Class A Land Pan

The U.S. Weather Bureau Class A land panis 1.22 m in diameter and25.4 cm deep (Figure 6.6). White paint is applied to the inner base and sides of the pan. The pan is placed on a white painted level wooden stand above the ground. The pan is filled with water to a level at least 5.1 cm below the top and not more than 7.6 cm below the top of the pan. The fall in the water level in the pan (after accounting for rainfall) within a specified time is evaporation. After observing the water level in the pan, water is added to bring the level in the specified range. Many times, the pan is covered by a wire mesh to prevent birds and animals from drinking pan water.

It is important to ensure that the pan is installed on a site which is leveled and free of obstructions. The ground cover should be as close as possible to that commonly found in the area. Any obstruction (tree, building, shrub or instrument shelter) should not be closer than four times the height of the obstruction. The pan should never be placed on an asphalt or concrete floor and no shadow should fall over it at any time. It should be preferably be installed in a plot whose preferred size is 15 m $\times$ 20 m. It is a good idea to fence the plot to protect the instruments and prevent animals from drinking the pan water but the fence should not affect the wind flow over the pan. Sometimes chemical repellants are used to prevent birds and animals from drinking the water but these should not pollute pan water.

On days without rain, at daily (or twice-daily) reading time, water is poured into the pan using a graduated cylinder to bring the level precisely to the top of the pointer gauge. The volume of water added is recorded and represents the depth of evaporation. On the days when rain has fallen since the last observation, the rainfall may exceed evaporation and water may have to be removed from the pan to bring the level to the hook level. The adjacent rain gauge is used to measure the rainfall. It there is a forecast of heavy rainfall on a given day, a measured amount of water may be removed from the pan in advance so as to avoid pan overflow.

The amount of evaporation (E in mm) between two successive observations is the difference between water levels in the pan. It is corrected for any precipitation during the period:

$$E = P \pm \Delta d \qquad (6.x)$$

where P is the depth of precipitation during the period between the two measurements, and Δd is the depth of water added (+) to or removed (–) from the pan.

A good practice is to install additional instruments along with a pan, such as an anemometer to measure wind movement over the pan, a precipitation gauge, and thermometers to measure temperature of pan water and surrounding air. The inner side of the pan is painted white. The water level in the pan changes due to evaporation and rainfall.

3.2 Estimation of Lake Evaporation from Pan Measurements

Evaporation data from pans are frequently used to estimate evaporation from water bodies, such as lakes and reservoirs and evapotranspiration from an area. But a pan is really small in size and its material is different than a natural body. Hence, its heat storage characteristics and air dynamics will be quite different than a large water body. Therefore, evaporation from a pan will be higher than a large open water body.

An estimate of lake or reservoir evaporation (E_R) is obtained by multiplying the pan evaporation by a coefficient which is called the pan coefficient:

$$E_R = K_P\, E_{pan} \tag{6.2}$$

where K_P is the pan coefficient; E_R is the depth of evaporation from the reservoir; and E_{pan} is the pan evaporation, both in mm/day. The pan coefficient modifies E_{pan} in accordance with the physical conditions of the lake/reservoir, since the pan has a much smaller heat capacity and is exposed to the climate from the sides and bottom. The value of pan coefficient depends on climate, geographical location, season, size, and depth of the water body. This coefficient generally varies from 0.6 to 0.85(annual scale) and from 0.3 to 1.7 on monthly scale, depending on the characteristics of water body. Lower values are typical of dry seasons and arid climates while higher values are appropriate for humid climates. In the absence of better estimates, a pan coefficient value of around 0.7 may be used. The pan coefficient for a particular pan can be determined by comparison with actual lake evaporation or with a large pan sunken to simulate a lake.

The advantages of using a pan are that: (a) it is quite cheap, simple and easy to install and operate, (b) the pan coefficient is stable, and (d) more data is available from using this pan. Pan measurements approximate lake evaporation reasonably close and have the advantage of less skilled labor. The disadvantages are that the pan coefficient varies in time and space.

Example 1: Compute daily evaporation from a Class A pan if the daily rainfall and the amount of water added to bring the water level in the pan to the fixed point are as follows:

Day	1	2	3	4	5	6
Rainfall (cm)	0	0.5	0.1	0	0	0.4
Water added (cm)	1.5	1.7	0.5	1.2	0.7	1.3

Solution: The daily pan evaporation equals the amount of water required to bring its water level to the fixed point plus the water contributed by rainfall. For each day, this amount is shown as follows, and varies from day to day, depending upon atmospheric conditions.

Day	1	2	3	4	5	6
Evaporation (cm)	1.5	2.2	0.6	1.2	0.7	1.7

Example 2: The surface area of a lake is 500 hectares. Compute daily evaporation for the data in Example 1. Assume the pan coefficient is 0.8.

Solution: The daily lake evaporation is obtained by multiplying the daily pan evaporation by the pan coefficient:

Day	1	2	3	4	5	6
Pan evaporation (cm)	1.5	2.2	0.6	1.2	0.7	1.7
Lake evaporation (cm)	1.2	1.76	0.48	0.96	0.56	1.36

The 6-day evaporation is the sum of daily values for 6 days = 6.32 cm. The 6-day loss of water = (6.32m/100) × 500 × 10,000 m^2 = 316,000 m^3. Since 1 m^3 = 1,000 liters, the 6-day loss = 316×10^6 liters.

Example 3: Compute the mean daily evaporation loss in hectare-meters for the month of July from a stream reach 100 km long and 50 m wide on average. The mean daily evaporation measured by a Class A pan for July is 0.6 cm. Assume the pan coefficient as 0.8.

Solution: Mean daily evaporation = 0.6 cm × 0.8 = 0.48 cm

This evaporation takes place from the entire stream reach whose area is 10,000 m × 50 m = 5 × 10^7 m^2, or 500 ha. Therefore,

$$\text{Daily Evaporation} = \frac{500\,ha \times 0.48\,m}{100} = 2.4\,ha - m$$

This loss of water due to evaporation must be considered in managing the water of this stream reach.

Example 4: It is desired to estimate the fall in water level due to evaporation from a lake on June 24, 2015. A Class A pan is located near the lake. On this day, the rainfall amount was 0.45 cm and the amount of water added to restore the water level to the value at the beginning of that day was 0.65 cm. Assume the pan coefficient as 0.78.

Solution Class A Evaporation = 0.44 cm + 0.65 cm = 1.1 cm
Evaporation from the lake = 1.1 cm x 0.78 = 0.858 cm
Thus, on June 24, the water level declined by 0.858 cm.

4.0 DETERMINATION OF EVAPORATION FROM WATER BODIES

Evaporation from water surfaces can be determined by (1) the water budget, (2) the energy budget, (3) mass transfer methods, (4) combination methods, and (5) evaporation formulas.

4.1 Water Budget

The water budget equation for estimating evaporation can be written (Horton, 1943b) as

$$E = I + P - O - O_s + \Delta S \tag{6.3}$$

where E is the evaporation, I is the inflow, P is the precipitation, O is the outflow, O_s is the seepage, and ΔS is the change in storage. Seepage, O_s, cannot be measured or evaluated directly and accurately, and the extent to which this quantity is accurate will affect the true value of evaporation. Inflow, outflow, precipitation, and change in storage can be measured reasonably accurately. The water budget method of determining long-term evaporation can be used as a standard for comparing other methods. This method is not perfect, but it is satisfactory for practical purposes.

Example 5: Estimate the evaporation for a month for a lake of 500 hectare surface area. The mean discharge from the lake is estimated to be 1.00 m^3/s. The monthly rainfall is about 10 cm. A stream flows with an average discharge of 2.00 m^3/s into the lake. The water level in the lake dropped about 5 cm in the month. The seepage losses are negligible.

Solution

Monthly inflow = I = 2 x 3600 x 24 x 30 m^3 = 51.84 x 10^5 m^3

Monthly outflow = O = 1 x 3600 x 24 x 30 m^3 = 25.92 x 10^5 m^3

Monthly rainfall = P = $\frac{10}{100} \times 500 \times 10{,}000$ m^3 = 5 x 10^5 m^3

Change in storage = ΔS = $\frac{5}{100} \times 500 \times 10{,}000$ m^3 = 2.5 x 10^5 m^3

Applying Equation 6.3 for estimating monthly evaporation,

$$E = (51.84 + 5.0 - 25.92 + 2.5) \times 10^5 \, m^3 = 33.42 \times 10^5 \, m^3$$

$$= \frac{33.42 \times 10^5}{500 \times 10^4} \times 100\, cm = 66.84\, cm$$

4.2 Energy Budget

Assuming unlimited availability of water, evaporation rate is chiefly governed by the amount of available radiant energy. The net amount of radiation available at the earth surface is the difference between the incoming and reflected solar (short wave) radiation (Fig. 3) plus the difference between the long wave incoming and outgoing radiations:

$$R_n = S_n + L_n \tag{6.4}$$

where R_n = the net radiation, S_n = the net short-wave (solar) radiation, and L_n is the net long-wave radiation, all in MJ/(m^2 day).

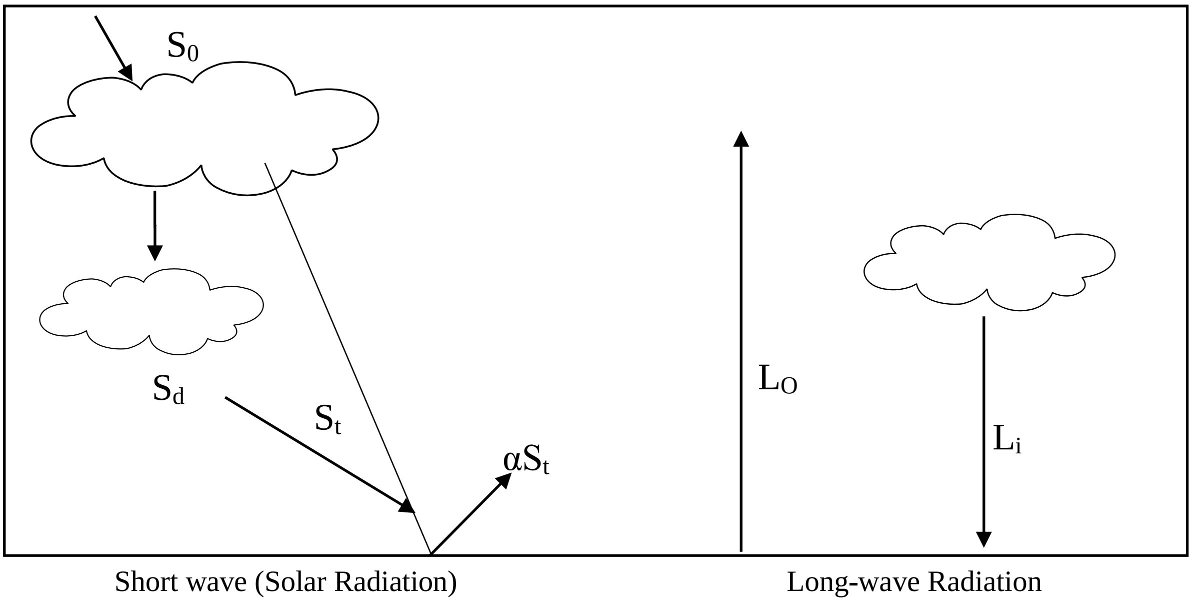

Fig. 3 Radiation balance at the ground. S_0 is the incoming short-wave radiation at the top of the atmosphere. After scattering and interception by the atmosphere and clouds, a part S_t comes to earth surface and a part of it is reflected. L_i is the incoming long-wave radiation; a part of it is reflected back.

The net short-wave radiation is equal to the incident short-wave radiation less the part due to reflection and is given by

$$S_n = S_t(1.0 - \alpha) \qquad (6.5)$$

where S_t is the incoming short wave (0.3to 3.0 μm) radiation and α is the reflection coefficient or albedo which depends on the land cover and the direction of solar rays. Taller vegetation reflects less radiation than shorter. In the absence of better information, Shuttleworth (1993) suggested that albedo can be assumed as0.08for open water, 0.11-0.16for tall trees, 0.20 – 0.26 for short crops and grass,, 0.10 (wet) – 0.35 (dry) for bare soil, and 0.20 (old) – 0.80 (fresh)for snow and ice. Incoming solar radiation can be estimated from the extraterrestrial radiation by

$$S_t = \left(a_s + b_s\frac{n}{N}\right)S_0 \qquad (6.6)$$

where $a_s$and b_s are the angstrom coefficients, n is the number of bright sunshine hours in a day of N hours, S_0 is the extraterrestrial radiation. Coefficients a_s and b_s can be estimated by regression analysis of data of S_0 and S_t. In the absence of locally calibrated data, one may adopta_s= 0.25 and b_s = 0.50 (Shuttleworth 1993).

A considerable amount of radiant energy is exchanged between the terrestrial surface and atmosphere in the form of long-wave (3 to 100 μm) radiation. The net energy received is the difference between the incoming and outgoing radiation.

The energy-budget method to compute evaporation is the same as the water budget method but the energy budget deals with the conservation of energy. The energy available for evaporation is obtained by considering the incoming energy, the outgoing energy and the energy stored in the water body for a given time interval. Assuming that the temperature of water is constant, the change in heat stored is the change in the internal energy of water that is evaporated = $E_r l_v \rho_w$, where l_v is the latent heat of vaporization, and ρ_w is the density of water (= 996 kg/m^3). Hence, the energy balance for evaporation yields

$$E_r l_v \rho_w = R_n - H_s - G \qquad (6.7)$$

where H_s is the sensible heat flux and G is the heat conduction to the ground. When the sensible heat flux and the conduction to the ground are negligible, the evaporation rate E_r which is the rate when the entire incoming radiation is used to evaporate water can be computed as

$$E_r = \frac{R_n}{l_v \rho_w} \qquad (6.8)$$

Example 6: At a given location, the net radiation on May 01 was 180 W/m^2, air temperature was 25 ^{o}C, wind velocity at 2 m height was 2.5 m/s, and relative humidity was 70%. Use the energy method to determine the open water evaporation rate. Given ρ_w = 996 kg/m^3.

Solution: We first compute the latent heat of vaporization

$l_v = 2.501\times10^6 - 2370T = 2.501\times10^6 - 2370\times25 = 2442\times10^3$ J/kg = 2442 kJ/kg.

Hence,

$$E_r = \frac{180}{2442000 \times 996} = 7.4 \times 10^8\ m/s = 6.39\ mm/d.$$

Aerodynamic Method: This method computes evaporation by considering the ability of the atmosphere to transport water vapor away from the evaporating surface. Thus, evaporation is a function of the difference between the saturation vapor pressure at the surface at the ambient temperature and the vapor pressure at a height z_2 above the water surface. Applying the physical laws, the evaporation rate is given by

$$E_a = \frac{0.102u_2}{[\ln(z_2/z_0)]^2}(e_{as} - e_a) \qquad (6.9)$$

Here, u_2 is the wind velocity in m/s at height z_2 (cm), and z_0 is the roughness height of water surface. If temperature (T in °C) is known, the saturated vapor pressure e_{as} (Pa or N/m^2) can be computed by

$$e_{as} = 6.11 \times exp\left(\frac{17.27T}{237.3+T}\right) \qquad (6.10)$$

The vapor pressure e_a is given by

$$e_a = R_h \times e_{as} \qquad (6.11)$$

where R_h is the relative humidity.

Example 7: At a given location, the net radiation on May 01 was 180 W/m^2, air temperature was 25 °C, wind velocity at 2 m height was 2.5 m/s, and relative humidity was 70%. Use the aerodynamic method to determine the open water evaporation rate if the roughness height is 0.04 cm.

Solution: We first compute the saturated vapor pressure

e_{as} = 611×exp[17.27×25/(237.3+25)] = 3169 Pa.

So, *ea* = 0.70×3169 = 2218 Pa.

Hence,

$$E_a = \frac{0.102 \times 2.5}{[\ln(200/0.04)]^2}(3169 - 2218) \ = \ 3.34\, mm/day.$$

4.3 Combination Method

As the name suggests, the combination method combines the energy-budget and mass-transfer methods. The most popular combination method for computing evaporation from free water surfaces is that developed by Penman (1948). This method combines the fundamental physical principles and empirical concepts based on standard meteorological observations. The physical principles are the energy balance equation and the mass-transfer (or aerodynamic) equation.

The aerodynamic method can be used when the available energy is not limiting evaporation and the energy balance method can be used when the vapor transport is not a limiting factor. Normally both of these factors are not limiting and hence the combination method may be suitable. Here evaporation is calculated as:

$$E = \left(\frac{\Delta}{\Delta+\gamma}\right)E_r + \left(\frac{\gamma}{\Delta+\gamma}\right)E_a \qquad (6.12)$$

where the first term on the right side is due to vapor transfer and the second term is due to aerodynamic mass transfer, and γ is the psychrometric constant (= 66.8 Pa/°C). Notation Δ (kPa/°C) denotes the gradient of saturated vapor pressure e_{as} and is given by

$$\Delta = \frac{4098e_{as}}{(237.3+T)^2} \qquad (6.13)$$

where *T* is the temperature in °C**.**

Example 8: At a given location, the net radiation on May 01 was 180 W/m^2, air temperature was 25 °C, wind velocity at 2 m height was 2.5 m/s, and relative humidity was 70%. Use the aerodynamic method to determine open water evaporation rate if the roughness height is 0.04 cm.

Solution: For the given data, the saturated vapor pressure e_{as} was calculated in the earlier example as 3169 Pa. We now compute the gradient of e_{as}:

$$\Delta = \frac{4098e_{as}}{(237.3+T)^2} = \frac{4098\times3169}{(237.3+25)^2} = 188.75 \text{ Pa/°C}.$$

Hence, $E = \left(\frac{\Delta}{\Delta+\gamma}\right) E_r + \left(\frac{\gamma}{\Delta+\gamma}\right) E_a = \left(\frac{188.75}{188.75+66.8}\right) 6.39 + \left(\frac{66.8}{188.75+66.8}\right) 3.34$

$= 5.6$ mm /day.

5.0 TRANSPIRATION

Transpiration is the process by which plants use water for their metabolism and growth. Plants extract water from the soil through their roots and transpire it to atmosphere as vapor through stomata in their leaves. Stomata are small openings in plant leaves for exchange of moisture and gases with the atmosphere. Their number on a plant leave varies with the plant species and their function is vital to plant metabolism. Just to give an idea, the density of stomata can vary from about 8000 to 12500 per cm^2; the number on the lower leaf surface is much more than on the upper surface. Transpiration is bit similar to evaporation, except that here water escapes to the atmosphere not from the free water surface but through plant leaves. Transpiration is affected by plant physiological and environmental factors. The rate of movement of water through the plant is important, since nutrients and minerals are carried into the plant with water.

During daylight, the stomata actively transpire water vapor to the atmosphere but they close after sunset and transpiration ceases. Stomata allow carbon dioxide to enter the plant in the process of photosynthesis. Stomata respond to environmental conditions and their opening depends on factors, such as light and darkness, hot and cold weather, etc. Important plant physiological factors are: (a) leaf structure, (b) extent and properties of protective coverings, (c) density and properties of stomata, and (d) plant health. Leaves that are exposed to direct sun rays transpire more water than those shaded by other leaves. If soil moisture is adequate, the amount of moisture lost by transpiration increases with plant density. Healthy plants transpire at higher rates compared to those with diseases. Only about 1% of the water taken up by the roots is retained by plant tissues; most of the moisture passes from roots to atmosphere via leafs. Environmental factors that are the most important in affecting transpiration are vapor-pressure gradient, temperature, solar radiation, wind, and available soil moisture.

The vapor-pressure gradient is the difference in vapor pressure between the space immediately inside the leaf and the outside air and it is a measure of the energy required to move the water from the leaf to the air. Expectedly, transpiration is less when the vapor-pressure gradient is less, e.g., during a rainfall event. However, plants transpire more when the surroundings are dry and warm, sun is shining, and plenty of soil moisture is available.

Transpiration is greatly affected by soil moisture condition. As the soil moisture depletes, the capillary forces holding soil moisture become stronger and it is hard for the plant roots to remove moisture. This is more true when the soil moisture is near the permanent wilting point. Of course, at the permanent wilting point transpiration stops. Wind usually increases transpiration by removing the moisture-laden air near the leaf and thereby increasing the vapor-pressure gradient. Gentle winds increase transpiration more than strong winds.

6.0 EVAPOTRANSPIRATION

Evapotranspiration (ET) and consumptive use include both the transpiration by vegetation, and evaporation from water surfaces, soil, snow, ice, and vegetation. For all practical purposes, the terms consumptive water use and evapotranspiration are synonymous. ET or consumptive uses converts water to a form (water vapor) which is not available for use again. This is in contrast to, say hydropower generation, where water is subsequently available for use again. ET is typically expressed in the units of depth (mm or cm) for a given period.

ET is an important component of the hydrologic cycle. Estimation of ET is necessary in many studies, such as catchment modeling, agricultural water management, determination of water balance, assessment of the impact of land use changes on the hydrologic response of a catchment, etc. In many watersheds, the return of moisture to the atmosphere through the process of ET is a large proportion of input precipitation.

Despite the widespread application and use of the concept of ET, there has been considerable ambiguity in the use of various terms, such as potential ET and reference crop ET. To overcome this, the Food and Agricultural Organization (FAO) of the United Nations brought out a report, commonly referred to as FAO-56 (Allen et al., 1998). Among other things, it introduced uniformity and standardization in the interpretation and use of various terms, such as potential ET and reference crop ET. FAO-56 discourages the use of the term potential ET due to ambiguities in its definition. Moreover, FAO recommended that a hypothetical reference surface "closely resembling an extensive surface of green grass of uniform height, actively growing, completely shading the ground and with adequate water" (Allen et al. 1998) be adopted as reference surface. In the FAO approach, the surface characteristics that influence ET are quantified in an unambiguous fashion.

The evapotranspiration rate from a reference surface, not short of water, is called the reference crop ET or reference evapotranspiration and is denoted as ET_0 (Allen et al. 1998). The reference surface is a hypothetical grass reference crop with specific characteristics. Further, crop ET under standard conditions (ET_c) refers to the evapotranspiration from excellently managed, disease-free, large, well-watered fields that achieve full production under the given climatic conditions. Further, due to suboptimal crop management and environmental constraints that affect crop growth and limit evapotranspiration (ET_c) under non-standard conditions generally requires a correction.

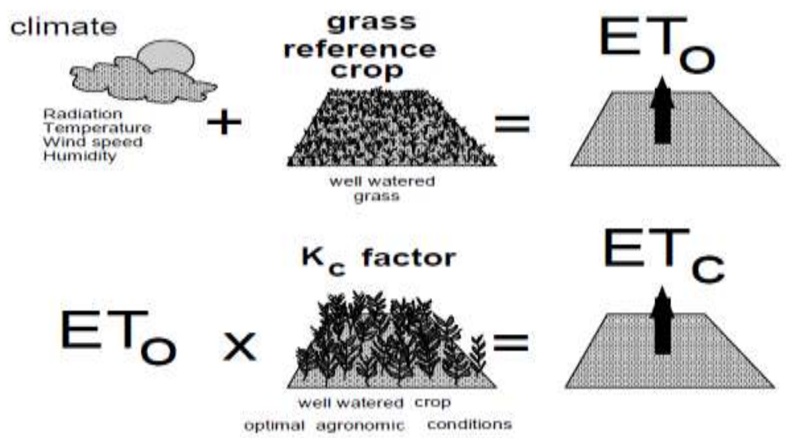

Fig. 4 Reference crop ET or $ET_0$and crop ET under standard conditions (ET_c). Source FAO-56

To estimate ET from a well-watered agricultural crop, reference evapotranspiration from a standard surface (ET_0) is first estimated. This value is multiplied by an empirical crop coefficient (Fig. 4) to obtain the ET from the crop (ET_c). The crop coefficient accounts for the difference between the standard surface and the crop. Reference ET is expressed in the units of depth/time, e.g., mm/day. It is a climatic parameter expressing the evaporative power of the atmosphere at a given space and time. Crop and soil features are not involved in computing it.

7.0 Estimation of Evapotranspiration

Evapotranspiration can be estimated by the water budget or heat-budget methods; many empirical formulae have been developed which are based on meteorological data. Food and Agricultural Organization (FAO) of the United Nations has adopted the Penman-Monteith (PM) equation as the standard technique to compute reference ET (Allen et al. 1998).

7.1 FAO-56 Penman-Monteith Method for Estimation of ET

Numerous reference ET equations have been developed and are being used, depending upon the availability of weather data. These equations range in sophistication from empirical solar radiation- or temperature-based equations to complex methods based on physical processes such as the combination method of Penman (1948). The combination approach links evaporation dynamics with the flux of net radiation and aerodynamic transport characteristics of a natural surface. Based on observations that latent heat transfer in plant stems is influenced not only by these abiotic factors, Monteith (1965) introduced a surface conductance term that accounted for the response of leaf stomata to its hydrologic environment. This modified form of the Penman equation is widely known as the Penman-Monteith (PM) equation.

The PM equation is physically based, because it attempts to incorporate the physiological and aerodynamic characteristics of the reference surface. While the use of the modified Penman method (Doorenbos and Pruit, 1977) was recommended by FAO, recent studies have suggested that this method overestimates ET (Sudheer et al., 2003). FAO has now recommended the use of the PM method to compute reference ET from a grass surface and has specified a grass reference ET equation (Allen et al. 1998). Studies have shown that the reference ET computed using the PM equation yields estimates that are close to observed reference ET values.

As described in the Irrigation and Drainage Paper 56 (Allen et al. 1998), the FAO has adopted the Penman-Monteith (PM) equation (named here FAO56-PM) as the standard technique to compute reference ET. The FAO56-PM can be used for hourly or daily time steps. For hourly time steps, the equation is stated as (Allen et al. 1998):

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma \dfrac{37}{T_{hr} + 273} u_2 \left[e^0(T_{hr}) - e_a\right]}{\Delta + \gamma(1 + 0.34u_2)} \tag{6.14}$$

where ET_0 is the grass reference ET in mm/hour, R_n is the net radiation at the grass surface in MJ per m^2 per hour, G is the soil heat flux density in MJ per m^2 per hour, T is the mean hourly air temperature in °C, u_2 is the mean hourly wind speed at 2 m height in m/s, $e^0(T_{hr})$ is the saturation vapor pressure in kPa at air temperature T_{hr}, e_a is the actual hourly vapor pressure in kPa, Δ is the slope of vapor pressure versus temperature curve in kPa per °C, and γ is the psychrometric constant in kPa per °C. Allen et al. (1998) have described the procedure and steps for the application of the PM equation for various time step sizes.

An application of the FAO56-PM equation requires data of solar radiation, wind speed, air temperature, vapor pressure, and humidity. However, all these input variables may not be easily available for a given location. In developing countries in particular, difficulties are often faced in collecting accurate data of all necessary climatic variables and this can be a serious handicap in applying FAO56-PM equation. Among the inputs needed, temperature data are routinely measured and solar radiation can be estimated with sufficient accuracy. But the other variables are mostly measured only at a few locations.

7.2 Energy Balance (Bowen Ratio Energy Balance, BREB) Method

The diffusion processes that transport water vapor and sensible heat in the atmosphere are somewhat similar. Hence, it can be assumed that the aerodynamic resistance hindering the flow of water vapor relates the difference in vapor pressure between two elevations is equal to the resistance to the flow of sensible heat to the difference in temperature between the same heights. This fact was observed by Bowen (1926).

The energy balance for an area can be written as

$$\mathrm{R_n} - \mathrm{G} - \lambda E - H = 0$$

or

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

Dividing all terms by λE,

$$1+\frac{H}{\lambda E}=\frac{\mathrm{R_n-G}}{\lambda E} \qquad (6.15)$$

The Bowen ratio (β) is defined as the ratio of the sensible heat H to the latent heat λE, or$\beta = H/\lambda E$.It is directly related to the ratio of the differences in temperature and humidity measured at two different elevations:

$$\beta=\frac{H}{\lambda E}=\gamma\frac{\Delta \mathrm{T}}{\Delta \mathrm{e}} \qquad (6.16)$$

where γ is the psychrometric constant (=0.4g/kg K^{-1}). Hence, eq. (6.15) can be written as

$$1+\beta=\frac{\mathrm{R_n-G}}{\lambda E}$$

or

$$\lambda E=\frac{\mathrm{R_n-G}}{1+\beta} \qquad ((6.17)$$

If β is known, λE can be determined from the measured values of R_n and G. The Bowen ratio reflects the wetness of ground: the values of β below unity or λE exceeding H indicate ample moisture availability; the values of β over unity or H exceeding λE indicate drier surfaces. The Bowen ratio varies from 0.1 to 0.3 for humid conditions, 0.4 to 0.8 temperate forests and grasslands, 2 to 6 semi-arid regions with highly dry soils, and > 10 for deserts.

Using equation (6.16), it is possible to calculate the Bowen ratio from the measurement of temperature and humidity at two elevations. Also, by measuring the net radiation and soil heat flux and expressing them as water equivalent, one can compute the sum of these two fluxes by using the equation (6.17). This gives the following equation to compute evaporation.

$$E=\frac{\mathrm{R_n-G}}{1+\gamma(\Delta \mathrm{T}/\Delta \mathrm{e})} \qquad (6.18)$$

This method is also known as the Bowen ratio energy balance (BREB) method. Since diffusion of vapor and movement of sensible heat is similar, BREB method is considered more robust. A problem with this method is that the sign of H term changes in the morning and evening which means that this equation is not applicable at these times.

8.0 Reduced-set Methods for ET Estimation

While applying FAO56-PM in practice, time and cost (instruments, etc.) involved in the acquisition and processing of the requisite meteorological data are a hindrance. All the required meteorological variables are observed at limited number of stations in the world. Further, the number of stations where *reliable* data for these variables exist is even smaller. Concerns have been expressed about the correctness of the observed meteorological data since the instruments, particularly to measure solar radiation and relative humidity, require careful calibration and maintenance. Many weather stations are often inadequately irrigated, particularly during dry periods, and thus the relative humidity and air temperature data may not be representative. Wind data are often unavailable or lack reliability.

Due to these reasons, there is a need for methods to estimate ET_o that require variables that are commonly measured and give reliable results. This has been the motivation to develop practical methods of ET_o estimation. A number of such methods have been developed. According to the meteorological variables that play the dominant role in the method, the available methods can be grouped in three categories: a) temperature-based; b) radiation-based; and c) combination. The first two groups are also termed as the reduced-set methods, since they use a small set of variables. For comparing the methods, lysimeter measurements or the PMM are considered as the benchmark.

8.1 Temperature Based Equations

Temperature is probably the easiest to measure, widely available and reliable weather variable. Temperature-based methods assume that temperature is an indicator of the evaporative power of the atmosphere. While wind speed is important in arid climate, the number of sunshine hours is a more important variable in sub-humid and humid climates.

8.2 The Hargreaves- Samani Method

Hargreaves (1975) found that ET can be adequately computed by using the average temperature and solar radiation data. Hargreaves (1975) proposed the following equation to compute ET_o:

$$ET_o = 0.0135(T + 17.78)R_s \qquad (6.19)$$

where R_s is the solar radiation is terms of water evaporation (mm/day), and T is temperature in ºC. Hargreaves and Samani (1982) showed that ET can be estimated by knowing the difference between the maximum (T_{max}) and the minimum (T_{min}) daily temperatures. Clear skies allow solar radiation to reach the ground easily and T_{max} is high while long wave radiation is able to escape easily in night and T_{min} is low. On the other hand, clouds obstruct incoming solar radiation as well as outgoing long wave radiation. Hence, T_{max} is lower and T_{min} is higher. Based on this, Hargreaves and Samani (1982) proposed the following equation to estimate solar radiation using the temperature difference (ΔT):

$$\frac{R_s}{R_a} = K_T(T_{max} - T_{min})^{0.5} \qquad (6.20)$$

where R_a is the extra-terrestrial radiation in mm/day which can be obtained from published tables or be calculated (Allen et al. 1998). For interior regions where land mass dominates, K_T can be taken as 0.162; a value 0.190 was suggested for coastal regions where air is influenced by sea. R_s/R_a ranges from 0.75 on a clear day to 0.25 on a day with dense clouds. Based on equations (5) and (6), Hargreaves and Samani (1985) developed an equation requiring only temperature, day of year and latitude to calculate *ETo (*HS equation):

$$ET_o = 0.0135K_T(T + 17.78) \times (T_{max} - T_{min})^{0.5}R_a \qquad (6.21)$$

Equation (5) was originally calibrated for semi-arid conditions and does not explicitly account for relative humidity. Hence, it may overestimate *ETo* in humid regions. Wind removes saturated air from the boundary layer and thus increases ET. Itensifu et al. (2003) found that the HS equation does not work well in high winds and high vapor pressure deficit situations. Wind speed is a major factor affecting the performance of the HS equation and this equation should be re-calibrated when it is applied in areas with very high or low wind speeds. The HS equation tends to overestimate *ETo* when mean daily *ETo* is relatively low and underestimate when *ETo* is relatively high and this is a common issue with most of the reduced set methods.

8.3 The Thornthwaite Method

Thornthwaite (1948) proposed a method by using the mean air temperature and number of hours of daylight to estimate *ETo* for short vegetation with adequate water supply. Monthly *ETo* can be estimated by the following Thornthwaite equation:

$$ET_o = ET_{osc} \times \frac{N}{12} \times \frac{dm}{30} \qquad (6.22)$$

where N is the maximum number of sunny hours as a function of month and latitude and *dm* is the number of days per month. *ETosc* is the gross ET (without corrections) which can be calculated as:

$$ET_{osc} = 16 \times \left(\frac{10T_a}{I}\right)a \qquad (6.23)$$

where T_a is the mean daily temperature (°C), a is an exponent as a function of the annual index:

a = 0.49239 + 1792 × 10^{-5} I - 771 × 10^{-7} I^2 + 675×10^{-9} I^3; and I is the annual heat index obtained from monthly heat indices:

$$I = \sum_{m=1}^{12} \left(\frac{T_m}{5}\right)^{1.514}$$

Under dry and arid conditions, the Thornthwaite equation strongly underestimates *ETo* (Garcia et al. 2004), because it does not consider the saturation deficit of air. At high altitudes also, it underestimates the effect of radiation because the equation is calibrated for temperate low altitude climates. Generally, the Thornthwaite method underestimates ET in humid areas.

Thornthwaite (1948) derived an equation for used in the case of limited water conditions. This equation can be written as

$$\mathrm{ET} = cT^a \quad (6.24)$$

where ET is in cm, c is a coefficient, T is the mean monthly temperature in ^oC, and a is an exponent. Both a and c depend on the location, and a can be estimated as

$$a = (67.5 \text{ x } 10^{-8})I^3 - (77.1 \text{ x } 10^{-6})I^2 + 0.0179I + 0.492 \quad (6.25)$$

where I is the heat index expressed as

$$I = \sum_{j=1}^{12} (T_j / 5)^{1.51} \quad (6.26)$$

where T_j. is the mean temperature of the jth month. The heat index is an integral element of Thornthwaite's classification of climates. The value of ET is modified by a factor to account for the number of daylight hours and the number of days in a month. Thornthwaite has tabulated values of this factor corresponding to various degrees and months of the year.

Assuming each month has 30 days and each day has 12 hours of sunshine, Equation 6.24 reduces to

$$\mathrm{ET} = 1.62\ (10T/I)^a \quad (6.27)$$

Taking its logarithm,

$$\log \mathrm{ET} = \log 1.62 + a(\log 10 + \log T - \log I) \quad (6.43)$$

Obviously, ET = 1.62 when I = 10T for 10T = log I. It has been shown by Thornthwaite that all lines obeying this equation have a common point of convergence at T = 26.5°C and E = 1.35 cm.

8.4 Blaney-Criddle method

The Blaney-Criddle (1950) method computes ET based on temperature data and coefficients that are computed from the data of humidity, sunshine and wind:

$$ET_o = \alpha + \beta[p(0.46T + 8.13)] \quad (6.29)$$

where p is the mean annual percentage of daytime hours. The value of α can be calculated using the daily *RHmin* and *n/N* as follows:

$$\alpha = 0.043RH_{min} - \left(\frac{n}{N}\right) - 1.41 \quad (6.30)$$

$$\left(\frac{n}{N}\right) = 2\left(\frac{R_s}{R_a}\right) - 0.5 \quad (6.31)$$

Value of β is obtained by calibration.

The Blaney and Criddle method has been used extensively. The underlying assumption of this procedure is that the heating of air and evaporation share the heat budget in a fixed proportion. As a result, ET varies directly with the sum of the products of mean monthly air temperature and monthly percentage of daytime hours with an actively growing crop with sufficient soil moisture. Estimates of ET by the Blaney–Criddle method are quite consistent.

9.0 Summary

Both temperature and radiation can be used to calculate daily *ETo* values with satisfactory accuracy. It is best to use these equations for areas having a climate similar to the one for which the original equations were developed. Most of the equations can be used with some confidence for areas with moderate conditions of humidity and wind speed. Globally, the Turc equation has produced good results for humid or semi-humid areas, while the Thornthwaite equation underestimates *ETo*. The Priestley-Taylor equation should not be used in the winter months in locations with high latitude, such as northern Europe. Both the Hargreaves and the reduced-set Penman-Monteith methods can be effectively used with only temperature measurements, although the results can be improved if wind speed is taken into consideration.

The use of the reduced-set equations can be very helpful in irrigation water management since the errors in applying these methods can be much smaller than those arising if data from a distant weather station are used.

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