CHAPTER 4

EVAPORATION, EVAPOTRANSPIRATION AND SOIL MOISTURE

4.1 EVAPORATION, EVAPOTRANSPIRATION AND INTERCEPTION

4.1.1 General

Evaporation and transpiration are the primary abstractions of the hydrological cycle. These abstractions are small during a runoff event and can be neglected. The bulk of evaporation and transpiration takes place during the time between runoff events, which is usually long. Hence, these abstractions are the most important during this time interval. The combined effect of evaporation and transpiration is called evapotranspiration. Over large land areas in temperate zones, about two thirds of the annual precipitation is evapotranspired and the remaining one third runs off in streams and through the groundwater to the oceans. In arid regions, evapotranspiration may be even more significant, returning up to 90 per cent or more of the annual precipitation to the atmosphere. Evaporation also links hydrology to atmospheric science and, through transpiration, to agricultural sciences.

4.1.2 Definitions

Evaporation

The process by which water is changed from the liquid or solid state into the gaseous state through the transfer of heat energy is known as evaporation.

In the hydrological cycle evaporation is an important process, so much so that on a continental basis, approximately 70 to 75 per cent of the total annual precipitation is returned to the atmosphere by evaporation and transpiration. In hot climates, the loss of water by evaporation from rivers, canals and open-water storage equipment is a vital matter as evaporation takes a significant proportion of all water supplies. It is significant in the sense that most of the water withdrawn for beneficial uses ultimately returns to streams and aquifers and becomes available for reuse, while the loss of water due to evaporation is entirely lost from the usable supply. Even in humid areas, evaporation loss is significant although the cumulative precipitation tends to mask it so that it is ordinarily not recognized except during rainless periods. Storage reservoirs expose wide surfaces to evaporation and thus are a major source of water loss even though they may lessen natural evaporation by confining floods in deep storages instead of spreading over wide flood plains.

The factors controlling evaporation have been known for a long time, but evaluating them is difficult because of their interdependent effects. However, in general, evaporation is affected by temperature, wind, atmospheric pressure, humidity, water quality, water depth, soil type and nature, and shape of surface.

Transpiration

Transpiration is defined as a natural plant physiological process whereby water is taken from the soil moisture storage by roots and passes through the plant structure and is evaporated from cells in the leaf called stomata.

The amount of water held in storage by a plant is less than 1 per cent of that lost by it during the growing season. From the hydrological standpoint, therefore, plants are like pumps that remove water from the ground and raise it to the atmosphere.

It is difficult to make precise estimates of the water transpired because of the many variables responsible for the process. Available estimates should be used with due caution taking into consideration the conditions under which these estimates were obtained. Adequate relationships between climatic factors and transpiration are prerequisites if the data derived in one climatic region are supposed to have general utility.

Transpiration is affected by physiological and environmental factors. Stomata tend to open and close in response to environmental conditions such as light and dark, and heat and cold. Environmental factors that affect transpiration are essentially the same as for evaporation, but can be considered a bit differently. For practical purposes, vapour pressure gradient, temperature, solar radiation, wind and available soil moisture are the most important factors affecting transpiration.

Evapotranspiration

The term evapotranspiration (ET) is defined as the water vapour produced from the watershed as a result of the growth of plants in the watershed.

Evapotranspiration and consumptive use include both the transpiration by vegetation and evaporation from free surfaces, soil, snow, ice and vegetation. Here it will be important to give the difference between evapotranspiration and consumptive use. Consumptive use differs from evapotranspiration only in that it includes the water used to make plant tissues (Singh, 1994). In computing evapotranspiration both transpiration and soil evaporation are included. The actual evapotranspiration can be determined by the analysis of the concurrent record of rainfall and runoff from a watershed.

There is an important difference between evapotranspiration and free surface evaporation. Transpiration is associated with plant growth and hence evapotranspiration occurs only when the plant is growing, resulting thereby in diurnal and seasonal variations. Transpiration thus superimposes these variations on the normal annual free water-surface evaporation.

Potential evapotranspiration

The potential evapotranspiration (PET) is defined as the evapotranspiration that would result when there is always an adequate water supply available to a fully vegetated surface.

This term implies an ideal water supply to the plants. In case water supply to the plant is less than PET, the deficient would be drawn from the soil-moisture storage until about 50 per cent of the available supply is utilized. With further moisture deficiency, the actual evapotranspiration (AET) will become less than PET until the wilting point is reached, and when the evapotranspiration stops.

Interception

Interception is that portion of the precipitation that, while falling on the Earth's surface, may be stored or collected by vegetal cover and subsequently evaporated. The volume of water thus lost is called interception loss.

In studies of major storm events and floods the interception loss is generally neglected. However, it may be a very significant factor in water balance studies. Precipitation falling on vegetation may be retained on leaves or blades of grass, flow down the stem of plants and become stem flow or fall off the leaves to become part of the throughfall. The amount of water intercepted is a function of (a) the storm character, (b) the species, age and density of plants and trees and (c) the season of the year. Usually about 10 to 20 per cent of the precipitation falling during the growing season is intercepted and returned to the hydrological cycle through evaporation. Under very dense forest conditions, it may be even as high as 25 per cent of the total precipitation. In temperate regions, evaporation of water intercepted by the vegetation represents an important part of the evapotranspiration. There is a wide variety of techniques used to measure rain interception (water stored in the canopy), canopy-interception-storage capacity, time of leaf wetness, throughfall, canopy evapotranspiration, and interception evaporation (often, but less appropriately, called interception loss). Reviews of interception measurement and leaf wetness methods are given by, for example, Bouten and others (1991) and Lundberg (1993), whereas canopy-storage-capacity measurements are summarized by Klaassen and others (1998). Micrometeorological evaporation methods are described by, for example, Garratt (1984) and Sharma (1985).

4.1.3 Measurement of evaporation [HOMS C46]

For a general reference on measurement instruments, see the *Guide to Meteorological Instruments and Methods of Observation* (WMO-No. 8).

4.1.3.1 Direct methods

Reasonably accurate methods of measurement of evaporation and evapotranspiration are available from pans and small bodies of water and soil, but direct measurement of evaporation or evapotranspiration from large water or land surfaces is not possible at present. However, several indirect methods have been developed that give acceptable results. Evaporation pans and lysimeters are used in networks for this purpose, and are discussed in this chapter. For existing reservoirs and plots or small catchments, estimates can be made by water-budget, energy-budget, and aerodynamic approaches and other available methods. These latter techniques are discussed in this chapter only from the point of view of instruments and observational requirements. Computation of evaporation and evapotranspiration from water and land surfaces by the various indirect methods is also discussed separately in this chapter. Some of the direct methods are as follows.

Pan evaporation

For estimation of evaporation from open water bodies, evaporation records of pans are generally used. The pans could be either square or circular section, mounted entirely above the ground or sunk in the ground so that the water level is approximately that of the ground. They may be mounted on anchored floating platforms on lakes or other water bodies.

Three types of pans deserve special mention: the United States Class A pan (Figure I.4.1), the GGI-3000 pan (Figure I.4.2) and the 20-m^2 tank of the Russian Federation. The United States Class A pan has been recommended by WMO and the International Association of Hydrological Sciences as a reference instrument as its performance has been studied under a range of climatic conditions within wide limits of latitude and elevation. The GGI-3000 pan and 20-m^2 tank are used in the Russian Federation and some other countries with different climatic conditions, as they possess reliable operational qualities and an extremely stable relationship with the meteorological elements that influence evaporation. WMO sponsored comparative observations (WMO, 1976) of the Class A pan, the GGI-3000 pan and the 20-m^2 tank in several countries, which eventually led to some operational recommendations on the suitability of these pans in diverse climatic and physiographic conditions.

In addition to the pan, a number of other instruments, such as integrating anemographs or anemometers, non-recording precipitation gauges, thermometers or thermographs for pan water temperature, maximum and minimum thermometers or thermographs for air temperature or hygro-thermographs or psychrometers, are also needed.

When installing evaporation pans it is important to ensure that the site of the pan is reasonably level and free of obstruction. At sites where normal climate and soil do not permit the maintenance of a soil cover, the ground cover should be maintained as near as possible to the natural cover common in the area. Obstructions such as trees, buildings, shrubs or instrument shelters should not be closer than four times the height of the object above the pan. Under no circumstance should the pan or instrument shelter be placed on a concrete slab or pedestal, or over asphalt or gravel.

The instruments should be located on the evaporation station plot so as to prevent them from casting shadows over the pan. The minimum size of the plot should be 15 m x 20 m. The plot should be fenced to protect the instruments and to prevent animals from drinking the water. The fence should be constructed so that it does not affect the wind structure over the pan. At unoccupied sites, particularly in arid and tropical regions, it is often necessary to protect the pans from birds and small animals by using chemical repellants and a wire mesh. To estimate the error introduced by the wire-mesh screen on the wind field and thermal characteristics of the pan, readings from the protected pan should be compared with those of a standard pan at the nearest comparable occupied site.

The water level in the pan must be measured accurately before and after water is added.

This may be done in two ways:
(a) The water level may be determined by means of a hook gauge consisting of a movable scale and vernier fitted with a hook enclosed in a still-water chamber in the pan. An alternative arrangement is to use a float. A calibrated container is used to add or remove water at

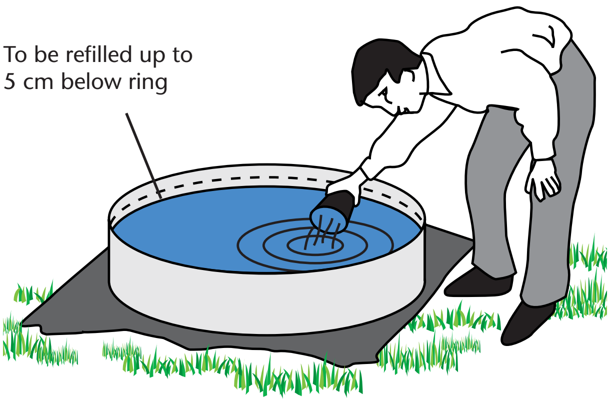

Figure I.4.1. United States Class A pan

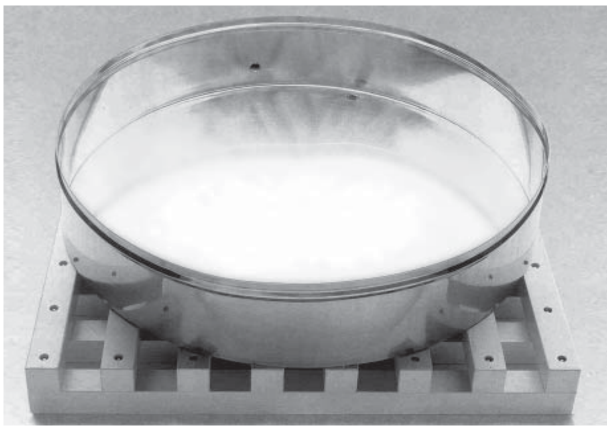

Figure I.4.2. GGI-3000 pan

each observation so as to maintain the water level to a pre-specified point;

(b) The water level may be determined by the following procedure:
   (i) A vessel of small diameter fitted with a valve is placed on top of a benchmark below the water surface in the pan;
   (ii) The valve is opened and the water level in the vessel is allowed to equalize with the water level in the pan;
   (iii) The valve is closed and the volume of water in the vessel is determined accurately in a measuring tube.

The height of the water level above the benchmark is determined from the volume of water in the vessel and the dimensions of the vessel.

Daily evaporation is computed as the difference in water level in the pan on successive days, corrected for any precipitation during the period. The amount of evaporation that has occurred between two observations of water level in the pan is determined by:

$$E = P \pm \Delta d \quad (4.1)$$

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.

Several types of automatic evaporation pans are in use. The water level in the pan is kept automatically constant by releasing water into the pan from a storage tank or by removing water from the pan in the case of precipitation. The amount of water added to or removed from the pan is recorded.

The major difficulty in using a Class A pan for the direct measurement of evaporation arises because of the use of coefficients to convert the measurements from a small tank to large bodies of open water. Fuzzy logic as suggested by Keskin and others (2004) can provide an alternative to the classical evaporation estimation.

Snow evaporation

Evaporimeters made of polyethylene or colourless plastic are used in many countries for measuring evaporation from, or condensation on, snow cover. Snow evaporimeters should have an area of at least 200 cm^2 and a depth of 10 cm.

A sample of snow is cut to fill the evaporimeter, the total weight is measured and the evaporimeter is set flush with the snow surface. Care should be taken that surface characteristics of the sample in the evaporimeter are similar to those of the snow cover in which it is placed. At the end of the measurement period, the evaporimeter is removed from the snow cover, the outside is wiped dry and a second measurement of weight is made. The difference between initial and final weights is converted to evaporation or condensation in centimetres. Measurements during periods of snowfall or blowing snow are not valid. During melt, the evaporimeters should be weighed and new samples should be cut at more frequent intervals as the snow cover will be settling, exposing the edge of the evaporimeter and altering air flow over the sample.

4.1.3.2 Indirect methods

Because of problems encountered in making direct measurements of evaporation from lakes and reservoirs, a number of indirect methods, such as the water-budget, the energy-budget, the aerodynamic approach or combination of these, are frequently used. The meteorological elements incorporated into these methods are solar and long-wave radiation, air and water-surface temperatures, atmospheric humidity or vapour pressure, and wind. Instruments and observational procedures for measuring these elements are described in the following subsections. The manner in which observations of the above elements are used in various indirect methods for estimating evaporation is described below in this chapter.

Solar radiation

Incident total solar (short-wave) radiation should be measured at a site near the reservoir with a pyranometer, and the output should be recorded continuously. Incoming short-wave radiation on a horizontal surface is measured with a pyranometer. Most modern types of pyranometers are based on multi-junction thermopiles and are covered by single or double glass domes that allow only radiation in the 0.3–3 µm range to reach the sensitive pyranometer surface (Figure I.4.3). Some types of pyranometer have the entire surface blackened with half the thermojunctions attached to it, with the other junctions located so that they sense the slowly varying reference temperature of a large, shielded brass block. Other types have a sensitive surface that consists of white and black painted surfaces, with thermojunctions attached to both.

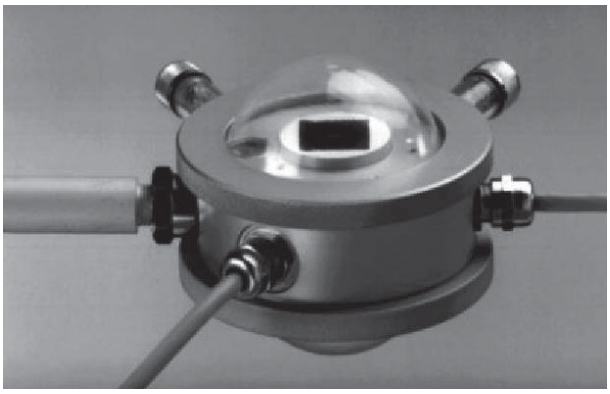

Figure I.4.3. Pyrradiometer (detail of the sensor)

Long-wave radiation

Long-wave radiation is measured indirectly with flat-plate radiometers. These instruments are not selective in response to different wavelengths and thus measure all wavelengths. The long-wave radiation is computed as the difference between the total radiation received from sun and sky as observed with a radiometer; the solar radiation is measured with a pyranometer at the same site.

One type of long-wave radiometer consists of a flat 5-cm^2 plate mounted horizontally in the exhaust of a small blower. The plate is a sandwich with a blackened aluminium upper surface and a polished aluminium lower surface. A thermopile measures the vertical temperature gradient across an insulating sheet that forms the centre layer of the sandwich. The thermopile voltage is proportional to the heat flow down through the plate, which in turn is proportional to the energy received at the blackened surface after deduction of the black-body radiation. To correct for the black-body radiation, a separate thermocouple measures the black-surface temperature. The function of the blower exhaust is to minimize the effects of wind on the calibration coefficient of the device.

Another type of instrument, a net pyrradiometer, measures the difference between total (short-wave and long-wave) incoming (downward) and outgoing (upward) radiation. The instrument consists of a horizontally mounted plate with two blackened surfaces. Half 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 the 0.3–100 µm band. These instruments are divided into two types: those that are ventilated and those that are shielded to reduce convective heat transfer from the sensing element. Instruments should be mounted at least 1 m above representative vegetation cover.

Air temperature

Air temperature should be measured 2 m above the water surface near the centre of the reservoir. For small reservoirs, the air temperature may not be greatly modified in its passage across the water surface, in which case satisfactory measurements can be made at an upwind shore site.

Although observations of air temperature at intervals of one, four or six hours may be satisfactory, continuous records are desirable, especially in connection with humidity measurements. Electrical thermographs, utilizing thermocouple thermometers, are suitable for recording on the multichannel recording potentiometers used for radiation measurements.

In measuring air temperature, thermometers must be shaded from the sun without restricting natural ventilation. Special radiation shields have been designed for thermocouple thermometers. Measurements of air temperature should be accurate to within ±0.3°C.

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, with voltmeter, with or without recorder, are used for the measurement of water temperature.

Particular applications will determine which thermometer is most suitable. For example, direct observations are best carried out with a mercury-in-glass thermometer, whereas continuous records may be obtained with resistance or thermocouple elements.

Thermographs, which produce a continuous record of temperature, usually comprise a mercury-in-steel sensing element immersed in the water, which is connected to a circular or cylindrical chart recorder with a Bourdon-tube transducer. Care should be taken in the installation of thermographs to ensure that measurements taken are representative of the water temperature (Herschy, 1971).

In the case of automatic stations where the measurement, which will usually include other variables, is recorded on a magnetic tape or transmitted over direct wire or radio-telemetry systems, the platinum-resistance or thermistor thermometers are used most frequently. As these have no moving parts, they are more reliable and offer greater accuracy and sensitivity of measurement. The sensing element is usually connected to a Wheatstone-bridge circuit and an electronic amplifier to produce an output signal that is suitable for recording or transmission.

In general, the precision required for the measurement of water temperature is ±0.1°C, except for special purposes where a greater accuracy may be required. However, in many circumstances precision of observation of ±0.5°C is adequate and there are many instances where statistical temperature data are quoted to the nearest 1°C. Thus, it is important to specify the operational requirement so that the most suitable thermometer is selected.

Humidity or vapour pressure of the air

Humidity measurements are made at the same location as air temperature. Psychrometers utilizing thermocouple thermometers are best suited for recording purposes. The thermocouple thermometers described in the preceding section on Air temperature, with an additional thermocouple thermometer to record wet-bulb temperatures, will give adequate results. Wet-bulb thermocouples require a wick and a reservoir that should be so arranged that the water will arrive at the wet-bulb temperature. Wet-bulb thermometers must be shielded from radiation and must, at the same time, maintain adequate ventilation to obtain a true wet-bulb temperature. A shield similar to the one used for air temperatures will provide adequate ventilation if wind speeds are greater than 0.5 ms^{-1}. In practice, the shield for the wet-bulb thermometer is placed just below the air temperature shield.

If measurements of dry- and wet-bulb temperatures are made to within ±0.3°C, the relative humidity should be within ±7 per cent for moderate temperatures. This is adequate for determining vapour pressure.

Wind

Wind speed should be measured near the centre of the lake or reservoir at a height of 2 m above the water surface. In practice, an anchored raft is used to support the instrumentation.

Any type of standard anemometer suitable for remote indication or recording should be adequate to determine the average daily wind speed. The three-cup rotor fan anemometers are most suited for remote recording. Accuracy of wind measurements by the three-cup or fan anemometers is usually within ±0.5 m s^{-1}, which is considered acceptable for evaporation measurements.

If a totalizing anemometer is used, provision must be made to read the counter at fixed intervals (preferably daily). If an electrical-contact anemometer is used, a recorder must be provided. This can be done by an electrical event marker on the margin of the temperature chart.

4.1.4 Measurement of evapotranspiration

Soil evaporimeters and lysimeters

Evapotranspiration can be estimated by the use of soil evaporimeters and lysimeters, by the water-budget or heat-budget methods, by the turbulent-diffusion method, or by various empirical formulae based on meteorological data. Use of soil evaporimeters and lysimeters allows direct measurement of evapotranspiration from different land surfaces and evaporation from the soil between cultivated plants. These instruments are simple and accurate if all requirements concerning their installation and observational techniques are fulfilled. Transpiration of vegetation is estimated as the difference between measured evapotranspiration and contemporaneously measured evaporation from the soil.

Soil evaporimeters and lysimeters are categorized according to their method of operation:
(a) Weight based, which use mechanical scales to account for changes in water content;
(b) Hydraulic based, which use the hydrostatic principle of weighing;
(c) Volumetric based, in which water content is kept constant and evapotranspiration is measured by the amount of water added or removed.

There is no single standard instrument for measuring evapotranspiration.

General requirements for the location of evaporation plots are as follows:
(a) The site selected for the plot should be typical of the surrounding area with respect to irrigation, soil characteristics (texture, layering, genetical type), slope and vegetative cover;

(b) The evaporation plot should be located beyond the zone of influence of individual buildings and trees. It should be situated at a distance not less than 100 to 150 m from the boundaries of the field and not more than 3 to 4 km from the meteorological station. Soil monoliths for inclusion in evaporimeters and lysimeters should be taken from within a radius of 50 m of the plot, and the soil and vegetative cover of the monolith should correspond to those of the plot.

4.1.5 Remote-sensing measurements of evaporation and evapotranspiration variables [HOMS D]

Remote-sensing observations combined with ancillary meteorological data have been used in obtaining indirect estimates of ET over a range of temporal and spatial scales (Schulz and Engman, 2000). Recently there has been a lot of progress in the remote-sensing of parameters, including:
(a) Incoming solar radiation;
(b) Surface albedo;
(c) Vegetative cover;
(d) Surface temperature;
(e) Surface soil moisture.

Remote-sensing of evaporation variables

Measurements of radiation and air temperature are usually made at the same locations, either at the centre of the lake or reservoir or at an upwind shore station. This permits recording several items in sequence on a single multichannel recorder. Integrating devices are sometimes used with strip-chart recorders. These devices present a visual readout of the average value of each item for the time period for which evaporation is to be computed (usually 10 days or two weeks).

Remote-sensing of several important parameters used to estimate evaporation is made by measuring the electromagnetic radiation in a particular waveband reflected or emitted from the Earth's surface. The incoming solar radiation can be estimated from satellite observations of cloud cover primarily from geosynchronous orbits using Multispectral Scanner (MSS) in the visible, near-infrared and thermal infra-red parts of EMS (Brakke and Kanemasu, 1981; Tarpley, 1979; Gautier and others, 1980). The surface albedo may be estimated for clear-sky conditions from measurements covering the entire visible and near-infra-red waveband (Jackson, 1985; Brest and Goward, 1987). The surface temperature may be estimated from MSS measurements at thermal IR wavelengths of the emitted radiant flux (Engman and Gurney, 1991).

However, there has been little progress in the direct remote-sensing of the atmospheric parameters that affect ET, such as:
(a) Near-surface air temperature;
(b) Near-surface water vapour gradients;
(c) Near-surface winds.

Furthermore, remote-sensing has a potentially important role because of its areal coverage in the spatial extrapolation process of ET.

Remote-sensing of evapotranspiration variables

Recently, researchers have begun using satellite data (for example, Bastiaanssen and others, 1998; Choudhury, 1997; Granger, 1997) to estimate regional actual evapotranspiration. Remote-sensing of several important parameters used to estimate ET is made by measuring the electromagnetic radiation in a particular waveband reflected or emitted from the Earth's surface. Estimates of incoming solar radiation, surface albedo and surface temperature may be done by the same satellite measurements described in 4.1.3. The soil moisture may be estimated using the measurement of microwave properties of the soil (microwave emission and reflection or backscatter from soil). However, there are uncertainties in such soil moisture estimates due to previously mentioned factors such as surface roughness and vegetative cover.

The most practical remote-sensing approach for the future will include repetitive observations at the visible, near and thermal infra-red, and microwave lengths. Components for determining the sensible heat flux will be measured by the EOS instruments. The latent heat flux cannot be measured directly but EOS instruments will provide some sampling capability. Furthermore, the future programme such as EOS should provide the necessary data for evaluating ET on local, regional and global scales.

4.2 ESTIMATING EVAPORATION FROM FREE SURFACES

4.2.1 General [HOMS I45]

Evaporation from water surfaces can be determined by various methods, such as:
(a) Water budget;
(b) Energy budget;
(c) Mass transfer methods;

(d) Combination methods;
(e) Empirical formulae.

Any of the methods described can be employed to determine evaporation. Usually, instrumentation for energy-budget and mass-transfer methods is quite expensive and the cost to maintain observations is substantial. For these reasons, the water-budget method and use of evaporation pans are more common. The pan method is the least expensive and will frequently provide good estimates of annual evaporation. Any approach selected is dependent, however, on the degree of accuracy required. As the ability to evaluate the parameters in the water budget and energy budget improves, so also will be resulting estimates of evaporation.

4.2.2 Water budget

The method is based on the continuity equation and can be utilized for the purpose of computing evaporation as:

$$E = I - O - \Delta S \qquad (4.2)$$

where E = evaporation, I = inflow, O = outflow and ΔS = change in storage.

By adding the suffixes s and g to the various components in equation 4.2 to denote vectors originating above and below ground surface respectively, the equation can be expressed as:

$$E_s = P + R_1 - R_2 - R_g - T_s - F - \Delta S_s \qquad (4.3)$$

where E_s = reservoir evaporation, P = precipitation, R_1 = surface runoff coming into the reservoir, R_2 = surface runoff going out of the reservoir, R_g = groundwater inflow, T_s = transpiration loss, F = infiltration (or seepage) and ΔS_s = change in storage.

If the net transfer of seepage $(R_g - F) = O_s$ and the transpiration term T_s equals zero, then equation 4.3 can be rewritten:

$$E_s = P + R_1 - R_2 + O_s - \Delta S_s \qquad (4.4)$$

All the terms are in volumetric units for a time period of interest that should be not less than a week. The water-budget method, although having the obvious advantage of being simple in theory, has the disadvantage in that the errors in the measurement of the parameters used in equation 4.4 are reflected directly in the computed amounts of evaporation. Therefore, it is not recommended that the method be applied to time periods of less than a month if the estimate of evaporation is expected to be within ±5 per cent of the actual amount.

Probably the most difficult term to evaluate is the seepage, F. This component can be estimated knowing the hydraulic conductivity of the lake bed and the hydraulic gradient. Nevertheless, it should be recognized that the water-budget method of determining evaporation will prove most successful when applied to relatively impervious lakes in which the seepage is negligible in comparison with the amount of evaporation.

To evaluate ΔS_s, an accurate area-capacity curve for the lake should be available. Even with these data, the bank storage component can introduce an error in the water budget. However, if the bank storage component is neglected, the water budget would not be useful on an annual cycle.

Although it is theoretically possible to use the water-budget method for the estimation of evaporation from any free surface, it is usually impractical to do so because of the effects of errors in measuring various parameters. Evaporation, estimated by this method, is residual and, therefore, may be subject to considerable error if it is small relative to other parameters.

In summary, the method is difficult and inaccurate under most conditions, particularly for short averaging time periods. Some of the most difficult parameters to measure are change in storage, seepage, groundwater flow and advected flows.

4.2.3 Energy budget

The energy-budget method illustrates an application of the continuity equation written in terms of energy. It has been employed to compute the evaporation from oceans and lakes, for example, at Elephant Butte Reservoir in New Mexico (Gunaji, 1968). The equation accounts for incoming and outgoing energy balanced by the amount of energy stored in the system. The accuracy of estimates of evaporation using the energy budget is highly dependent on the reliability and preciseness of measurement data. Under good conditions, average errors of perhaps 10 per cent for summer periods and 20 per cent for winter months can be expected.

The energy-budget equation for a lake may be written as (Viessman and others, 1989):

$$Q_0 = Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} + Q_v - Q_e - Q_h - Q_w \qquad (4.5)$$

where Q_0 = increase in stored energy by the water, Q_s = solar radiation incident at the water surface, Q_r = reflected solar radiation, Q_a = incoming long-wave radiation from the atmosphere, Q_{ar} = reflected long-wave radiation, Q_{bs} = long-wave radiation emitted by the water, Q_v = net energy advected (net energy content of incoming and outgoing water) into the water body, Q_e = energy used in evaporation, Q_h = energy conducted from water mass as sensible heat and Q_w = energy advected by evaporated water.

All the terms in equation 4.5 are in watt per square metre per day (W m^{-2}day). Heating brought about by chemical changes and biological processes is neglected, as it is the energy transfer that occurs at the water–ground interface. The transformation of kinetic energy into thermal energy is also excluded. These factors are usually very small, in a quantitative sense, when compared with other terms in the budget if large reservoirs are considered. As a result, their omission has little effect on the reliability of results.

Each of the various terms in the energy-budget equation is either measured directly or computed from known relationships. The procedure used in evaluating each term is described below.

The terms of equation 4.5 that can be measured are Q_s, Q_r and Q_a, and the net radiation balance is:

$$R_f = Q_s - Q_{sr} + Q_a - Q_{ar} - Q_{bs} \qquad (4.6)$$

All of the above values are expressed in W m^{-2}.

Detailed descriptions of the instruments and measuring techniques concerning the above-mentioned elements can be found in 4.1.3, 4.1.4 and 4.1.5, or in the *Guide to Meteorological Instruments and Methods of Observation* (WMO-No. 8).

Reflected long-wave radiation (Q_{ar}) may be taken as 3 per cent of the long-wave radiation received by the water surface.

Long-wave radiation emitted by the water (Q_{bs}) is computed according to the Stefan–Boltzmann law for black-body radiation, with an emissivity factor of 0.970 for water. The equation for computing radiation emitted by the water surface is:

$$Q_{bs} = 0.97\sigma\theta^4 \qquad (4.7)$$

where Q_{bs} is the radiation emitted by the water surface in W m^{-2}, σ is the Stefan-Boltzmann constant (5.67 x 10^{-8} *W* m^{-2} °K^{-4}), and θ is the temperature of the water surface in °K. For computing purposes, the average temperature of the water surface, as recorded near the centre of the reservoir, is determined for each period of study. The temperature is converted to °K, and the average radiation emitted by the water surface is computed for the period of study in W m^{-2}.

The thermal energy of the volume of water in the reservoir for a given date is computed from a temperature survey made on that date. These temperature measurements, which should be accurate to within 0.1°C, are usually made at biweekly or monthly intervals. The reservoir may be divided into several layers from the surface to the bottom. The volume of water for each of the layers is determined from the stage–volume relationship. All temperature observations made in a particular layer are averaged to obtain a mean temperature for that volume of water.

The summation of the products of volume and temperature (assuming a base temperature of 0°C) will give the total energy for that particular date. Density and specific heat are considered as unity for the range of temperatures that occur in the reservoir. In order to determine the energy utilized in evaporation, Q_e changes in energy storage resulting from advection of energy in the volumes of water entering or leaving the reservoir must be evaluated. Again, a base temperature of 0°C is usually chosen in computing the amount of energy in these volumes. Their temperatures are determined by observation or recordings (4.1.3) depending on the variation of temperature with the rate of flow. If the temperature of the water changes with the rate of flow, the mean temperature of the volume should be weighted according to the rate of flow. The temperatures of bank storage and net seepage are considered as being equal to the mean annual air temperature. This assumption is admittedly subject to error, but is not considered serious if the surface inflow is a large item in the water budget.

If precipitation is a significant item in the water budget, then the energy of this volume of water must be taken into account. The temperature of rainfall is assumed to be that of the wet bulb at the time of rainfall. In computing the energy for each of these volumes, centimetre-gram-second units are used, and density and specific heat are considered as unity for the range of temperatures that occur in these volumes. The product of temperature times volume will give the amount of energy for each volume in joules (net energy advected, Q_v). The difference between the computed energies of stored water for the thermal surveys made at the beginning and end of the period of study determines the change in energy storage (Q_0).

During winter months when ice cover is partial or complete, the energy budget only occasionally yields adequate results because it is difficult to measure reflected solar radiation, ice surface temperature and the areal extent of the ice cover. Daily evaporation estimates based on the energy budget are not feasible in most cases because reliable determination of changes in stored energy for such short periods is impractical. Periods of one week or longer are more likely to provide satisfactory measurements.

In using the energy-budget approach, it has been demonstrated that the required accuracy of measurement is not the same for all variables. For example, errors in measurement of incoming long-wave radiation as small as 2 per cent can introduce errors of 3–15 per cent in estimates of monthly evaporation, while errors of the order of 10 per cent in measurements of reflected solar energy may cause errors of only 1–5 per cent in calculated monthly evaporation. To permit the determination of evaporation by equation 4.5, it is common to use the following relation:

$$B = \frac{Q_h}{Q_e} \tag{4.8}$$

where B is known as Bowen's ratio (Bowen, 1926) and:

$$Q_w = \frac{c_p Q_e (T_e - T_b)}{L} \tag{4.9}$$

where c_p = the specific heat of water (cal/g°C) that is equal to 4186.8 J/kg°C, T_e = the temperature of evaporated water (°C) ; T_b= the temperature of an arbitrary datum usually taken as 0°C and L = the latent heat of vaporization (cal/g) that is equal to 2260 kJ/kg. Introducing these expressions in equation 4.5 and solving for Q_e, we obtain:

$$Q_e = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o + Q_v}{1 + B + c_p(T_e - T_b) / L} \tag{4.10}$$

To determine the depth of water evaporated per unit time, the following expression may be used:

$$E = \frac{Q_e}{\rho L} \tag{4.11}$$

where E = evaporation (m sec^{-1}) and ρ= the mass density of evaporated water (kg m^{-3}).

The energy-budget equation thus becomes:

$$E = \frac{Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} - Q_o + Q_v}{\rho \left\{ L(1 + B) + c_p(T_e - T_b) \right\}} \tag{4.12}$$

The Bowen ratio can be computed using:

$$B = 0.61 \frac{p(T_o - T_a)}{1000(e_o - e_a)} \tag{4.13}$$

where p = the atmospheric pressure (mb), T_o = the water-surface temperature (°C); T_a = the air temperature (°C), e_o = the saturation vapour pressure at the water-surface temperature (mb) and e_a = the vapour pressure of the air (mb).

This expression circumvents the problem of evaluating the sensible heat term, which does not lend itself to direct measurement.

Remote-sensing of several important parameters used to estimate evaporation is made by measuring the electromagnetic radiation in a particular waveband reflected or emitted from the Earth's surface as discussed earlier in 4.1.3.

Applicability of energy-budget approach

The points summarized below should be recognized first in order to apply the energy-budget approach for estimating the evaporation from free surfaces:

(a) The flow of heat from the bottom of the lake has not been accounted for. This, however, is important in the case of shallow lakes;
(b) Bowen's ratio is assumed to provide a sufficiently accurate estimate of Q_h;
(c) The approach neglects the effect due to radiative diffusivity, stability of the air and spray;
(d) The applicability of the approach hinges greatly on the ability to evaluate the advective energy components.

4.2.4 Mass-transfer method

The mass-transfer approach, as the name implies, is based on the determination of the mass of water vapour transferred from the water surface to the atmosphere. To better understand this, an insight into the physics of air movement is first discussed.

When air passes over land or water surfaces, the air thickness in the lower atmosphere may be divided into three layers: (a) the laminar layer near the surface; (b) the turbulent layer; and (c) the outer layer of frictional influence. The laminar layer, in which the air flow is laminar, is only of the order of a millimetre in thickness. In this layer the temperature, humidity and wind velocity vary almost linearly with height, and the transfer of heat, water vapour and momentum are essentially molecular processes. The overriding turbulent layer can be several metres in thickness depending on the level of turbulence. In this layer, temperature, humidity and wind velocity vary approximately linearly with the logarithm of height, and the transfer of heat, vapour and momentum through this layer are turbulent processes.

The mass-transfer approach is based on Dalton's aerodynamic law giving the relationship between evaporation and vapour pressure as:

$$E = k\,(e_s - e_a) \qquad (4.14)$$

where E = direct evaporation, k = a coefficient and depending on the wind velocity, atmospheric pressure and other factors, e_s and e_a = saturation vapour pressure corresponding to the water-surface temperature and the vapour pressure of the air, respectively. Mean daily temperature and relative humidity may be used in determining mean vapour pressure e_a and mean saturation deficit ($e_s - e_a$). Equation 4.14 was originally proposed by Harbeck and Meyers (1970).

4.2.5 Combination of aerodynamic and energy-balance methods

Perhaps the most widely used method for computing lake evaporation from meteorological factors is based on a combination of aerodynamic and energy-balance equations:

$$E_i = \frac{R_n \Delta + E_a \gamma}{\Delta + \gamma} \qquad (4.15)$$

where E_i is the estimated evaporation from a free-water surface, $\Delta = \dfrac{e_s - e_{sz}}{T_s - T_z}$ is the slope of the saturation vapour-pressure curve at any temperature θ_a, which is tabulated as γ/Δ versus T_z in Brutsaert (1982, Figure 10.2), R_n is the net radiation, γ is the constant in the wet and dry bulb psychrometer equation, and E_a is the same expressed in equation 4.14.

The psychrometer constant γ for °C is the same constant of the Bowen ratio, and its value at 1000-mb pressure is 0.61. The net radiation R_n (in MJ m^{-2} day) can be estimated by the following equation:

$$R_n = \left(0.25 + 0.5\frac{n}{N}\right) S_0 - \left(0.9\frac{n}{N} + 0.1\right) \left(0.34 - 0.14\sqrt{e_d}\right) \sigma T^4 \qquad (4.16)$$

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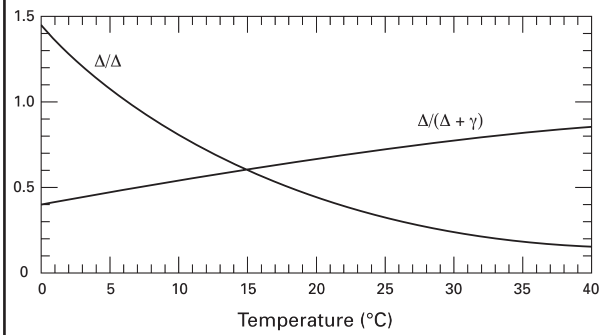


Temperature dependence of (γ/Δ) and $\Delta/(\Delta + \gamma)$ at 1000 mb
Adapted from Brutsaert (1982, Figure 10.2)

where n/N is the ratio of actual to possible hours of sunshine, S_0 is the extraterrestrial radiation (in MJ m^{-2} day), e_d is the actual vapour pressure of the air in mm of mercury, σ is the Stefan–Boltzmann constant, also expressed in equivalent evaporation in mm day^{-1}, and T is the mean air temperature (absolute) expressed in degrees Kelvin.

Although it may be necessary to use the above equation, it would be preferable to use measured values of solar and long-wave radiation.

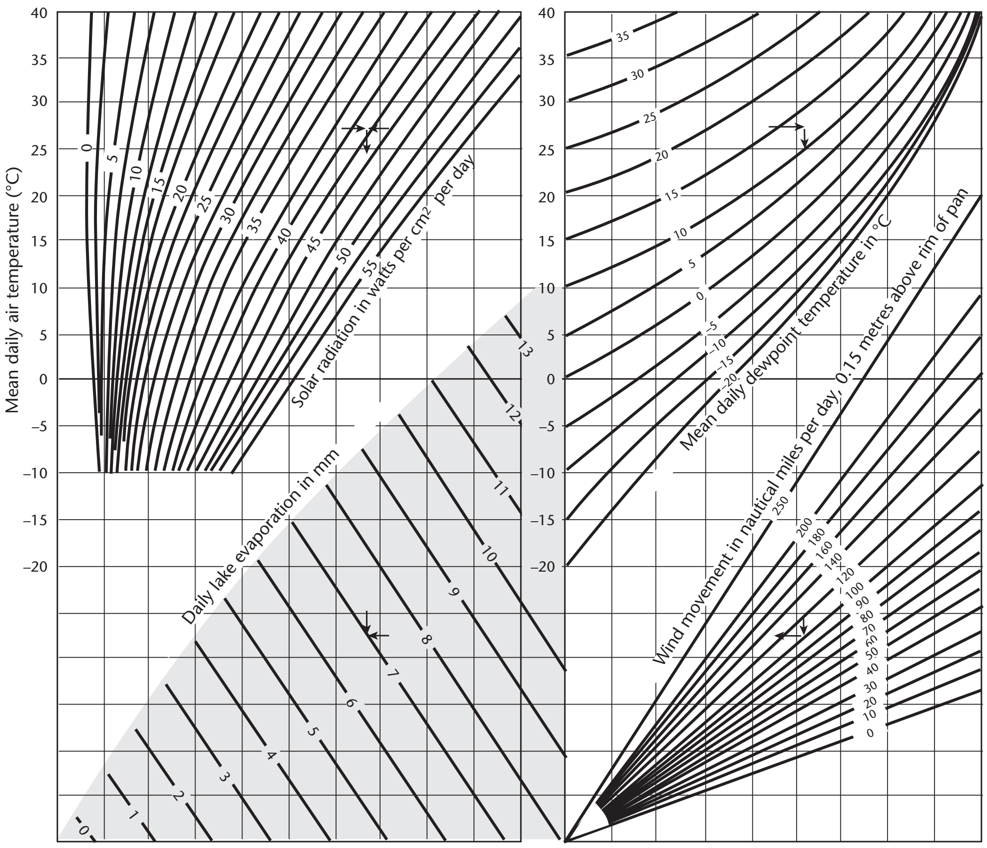
A similar approach was used by Kohler and others (1959) and a graphical presentation of the relationship is shown in Figure I.4.4. The meteorological observations of solar radiation, air temperature, dewpoint and wind movement at the anemometer height of a Class A pan are required for application of this technique. In the absence of solar-radiation observations, radiation may be estimated from the percentage of possible sunshine or cloud-cover data. Lake evaporation computed for short periods by this method would be applicable only to very shallow lakes with little or no advection of energy to the lake. For deep lakes and conditions of significant advection due to inflow and outflow, it is necessary to correct the computed lake evaporation for net advected energy and change in energy storage. These terms are described under the energy-budget method in 4.2.3. However, all of the advected energy and change in energy storage is not utilized for evaporation. The portion of this energy used for evaporation can be obtained from a relationship such as shown in Figure I.4.5. Observations of water-surface temperature and wind movement at 4 m above the water surface are required for application of this relationship. Reliable estimates of weekly or monthly lake evaporation can be obtained by this approach only if an evaluation is made of the energy-advection and storage factors.

4.2.6 **Extrapolation from pan measurements** [HOMS C46]

The evaporation from pans exposed in or on the ground is influenced by the characteristics of the pan. Sunken pans are subject to undetected leaks, accumulation of debris on the water surface, and boundary conditions with the soil different from those of a large lake. Pans exposed above the ground are subject to heat exchange through the sides and to other effects that do not occur in lakes. Floating pans are subject to splash-in and splash-out, and are costly to install and operate.

Pans have much less heat storage than lakes and tend to experience a different annual cycle of evaporation, with pan-evaporation extremes occurring earlier in the season. Reliable estimates of annual lake evaporation can be obtained by multiplying the annual pan evaporation by the appropriate pan-to-lake coefficient. These estimates will be reliable only if it can be assumed that, on an annual basis, any energy advected to the lake is balanced by a change in heat storage. The pan-to-lake coefficient for a particular pan is determined by comparison with actual lake evaporation, if available, or more commonly by comparison with a pan large enough to simulate a lake (sunken pans 4 m or more in diameter). The coefficient for a specific pan is also dependent, to a degree, upon the climatic regime, that is, different for arid or humid conditions. For an evaporation pan to serve as a valid index to lake evaporation, the exposure of the pan should avoid the environmental effects of the lake. Such an exposure would be near the lake, but on the side toward the prevailing wind direction. An island exposure would not be satisfactory.

One method for determining the climatic variation of the pan coefficient is by field comparisons with large pans under the various conditions. This method is applied in the Commonwealth of Independent States with the GGI-3000 and 20-m^2 tanks. The pan-to-lake coefficients thus derived for



Note: The International Pyrheliometric Scale, which became effective in the United States on 1 July 1957, provides values that are 2.0 per cent less than those previously obtained. Therefore, for computations based on data subsequent to 1 July 1957, increase radiation values by 2 per cent

Figure I.4.4. Lake–evaporation relationship

the GGI-3000 range between 0.75 and 1.00. For estimates of monthly average evaporation, the coefficient for a floating GGI-3000 evaporation pan is estimated by the following equation:

$$\alpha = 0.8\ \frac{e_0 - e_{200}\beta}{e'_0 - e_{200}\beta} \tag{4.17}$$

where e_o is the average monthly vapour pressure, in hPa, estimated from the surface temperature of water body, e'_o is the average monthly vapour pressure, in hPa, estimated from surface-water temperature in the floating GGI-3000 pan, e_{200} is the average monthly vapour pressure at 200 cm above the water surface, in hPa, β is a correction factor for the area of a water body, and γ is a factor that depends on the distance *l* along the average direction of wind from the shore to the pan (fetch).

The ratio, β/γ, needs to be determined only for water bodies located in tundra, forest and forest-steppe zones and when the pan is located at a distance of up to 500 m from shore. In all other cases, this ratio is assumed to be equal to 1. For water bodies of approximately round or square shape, β is determined from the area of the water surface by using Table I.4.1.

Table I.4.1. Determination of β

Area of water body (km^2)	0.01	0.05	0.1	0.5	1.0	2.0	5.0
Correction factor β	1.03	1.08	1.11	1.18	1.21	1.23	1.26

For water bodies of irregular shape (long with islands and gulfs), the area used is that of an assumed circle with a diameter equal to an average distance, *l*, weighted with the frequency of wind direction in per cent from the eight points of the compass. The weighted distance can be computed by the equation:

$$\overline{l} = \frac{1}{100}\, i \sum_{i=1}^{i=8} l_i N_i \tag{4.18}$$

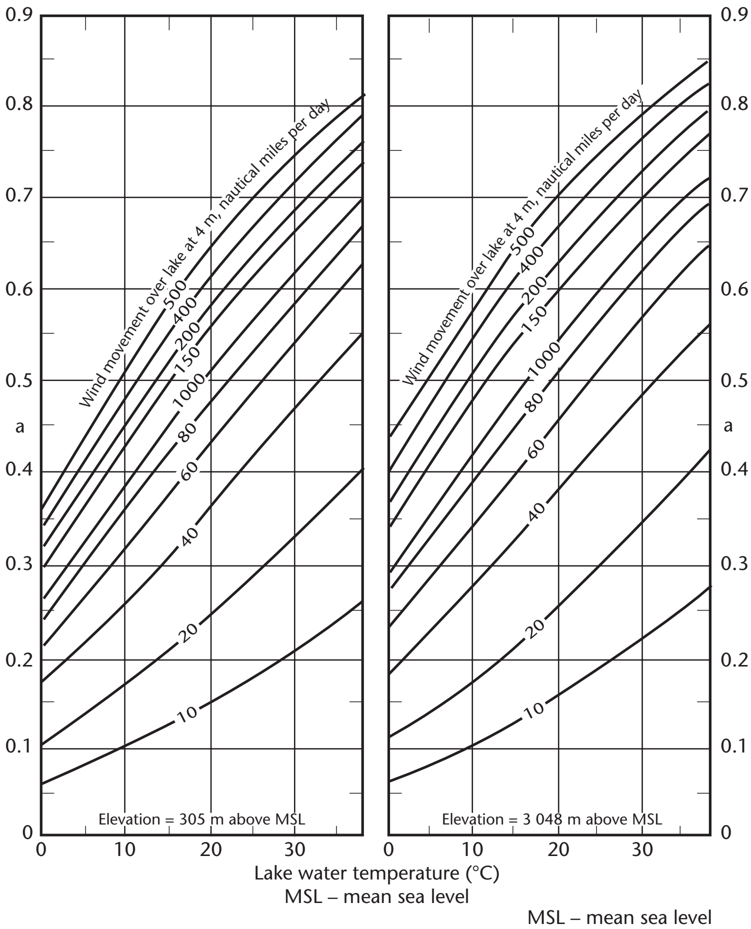


Figure I.4.5. Proportion of advected energy into a lake that is used for evaporation

where N_i is a frequency of wind direction from the eight points, in per cent; γ can be determined from Figure I.4.6.

Another method is the adjustment of the pan evaporation for heat gain or loss through the sides and bottom. An example of this method is the technique in estimating evaporation by using data from the Class A evaporation pan. In humid seasons and climates, the pan water temperature is higher than the air temperature, and the pan coefficient may be 0.80 or higher. In dry seasons and arid areas, the pan water temperature is less than air temperature, and the coefficient may be 0.60 or less. A coefficient of 0.70 is assumed to be applicable when water and air temperatures are equal. The relationships for estimating lake evaporation by adjusting Class A pan evaporation for heat gain or loss are shown in Figures I.4.7 and I.4.8. Owing to the important variation of wind with height, standard instrument heights are an essential requirement of the Class A station.

To obtain short-period estimates of lake evaporation with the pan method, it is also necessary to evaluate the net energy advection to the lake and change in energy storage as described in 4.2.3. It is useful to have pan evaporation near a lake or reservoir as a source of alternative data in the absence of other meteorological data and to help verify estimates made by the energy-budget and aerodynamic methods.

4.2.7 Empirical formulae

The energy-budget and mass transfer methods, though theoretically sound, require data which, for many studies, are not readily available. Moreover, in many cases even the economics of acquiring such data through instrumentation of the lake is also questionable. Thus, one has to make use of empirical formulae to obtain estimates of evaporation. Many empirical formulae to obtain estimates of evaporation have been developed (Mutreja, 1986) either on the basis of

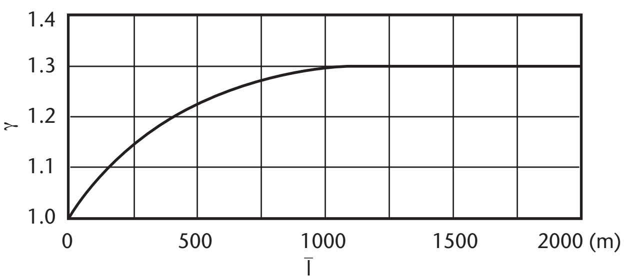


Figure I.4.6. Factor γ and $\bar{l}$ relationship

the energy-budget or mass transfer method. However, most of the equations are based on the simple aerodynamic equation given as equation 4.14.

A few of the more common of these empirical formulae used for estimating the evaporation from lake surfaces are given below:

Penman's formula, United Kingdom – small tank (Penman, 1948)

$$E(cm\ day^{-1}) = 0.89\ (1 + 0.15U_2)\ (e_s - e_a) \qquad (4.19)$$

where U_2 = the wind speed at 2 m above the water surface, e_s = saturation vapour pressure at water surface temperature and e_a = vapour pressure of the air at the specified height.

Marciano and Harbeck's formulae, United States (Marciano and Harbeck, 1954)

$$E(cm\ day^{-1}) = 0.0918U_8(e_s - e_8) \qquad (4.20)$$

$$E(cm\ day)^{-1} = 0.1156U_4(e_s - e_2) \qquad (4.21)$$

Kuzmin Formula, the then Union of Soviet Socialist Republics (Kuzmin, 1957) – reservoirs with surface >20–100 m

$$E(cm\ month^{-1}) = 15.24\ (1 + 0.13U_s)\ (e_s - e_a) \qquad (4.22)$$

United States Geological Survey (USGS), United States and Bureau of Reclamation's formula (USGS, 1977)

$$E(cm/year^{-1}) = 4.57T + 43.3 \qquad (4.23)$$

where T = mean annual temperature in °C

Shahtin Mamboub's formula, Egypt (Mutreja, 1986)

$$E(cm\ day^{-1}) = 0.35(e_s - e_a)\ (1 - 0.15U_2) \qquad (4.24)$$

where e_s = saturated vapour pressure at the water surface temperature (cm Hg^{-1}) and e_a = actual vapour pressure (cm Hg^{-1})

Unless specified in the above equations the wind speed (U) is in km x h^{-1} and vapour pressure is in cm of mercury. Further, the subscripts attached to the terms refer to the height in metres at which the measurements are to be taken. Also, the vapour pressure term e is frequently taken as the saturated vapour pressure at the mean air temperature during the interval of measurement.

The equations require surface temperature of the body of water, which is very difficult to measure. If this is substituted by the mean air temperature, then the effects of advected energy to the lake on evaporation are not considered. This may introduce considerable error in the computed amounts of evaporation, as small errors in temperature induce large errors in the computations. Furthermore, the measurements of the wind speed and vapour pressure should be measured at the height specified by the equation being used. Usually, it is difficult to adjust the data collected at different heights because neither an accurate wind law nor laws defining the variation in humidity with height are currently available.

The greatest appeal for the use of these empirical formulae lies in the fact that they are simple to use with the standard available meteorological data. Nevertheless, the limitations of these empirical formulae must be clearly understood.

4.3 EVAPOTRANSPIRATION FROM DRAINAGE BASINS [HOMS I50]

4.3.1 General

Evapotranspiration considers evaporation from natural surfaces whether the water source is in the soil, in plants, or in a combination of both. With respect to the cropped area, the consumptive use denotes the total evaporation from an area plus the water used by plant tissues, thus having the same meaning as evapotranspiration. The determination of evaporation and transpiration as separate elements for a drainage basin is unreliable. Moreover, their separate evaluation is not required for most studies.

Evapotranspiration is one of the most popular subjects of research in the field of hydrology and irrigation. Numerous procedures have been developed to estimate evapotranspiration. These

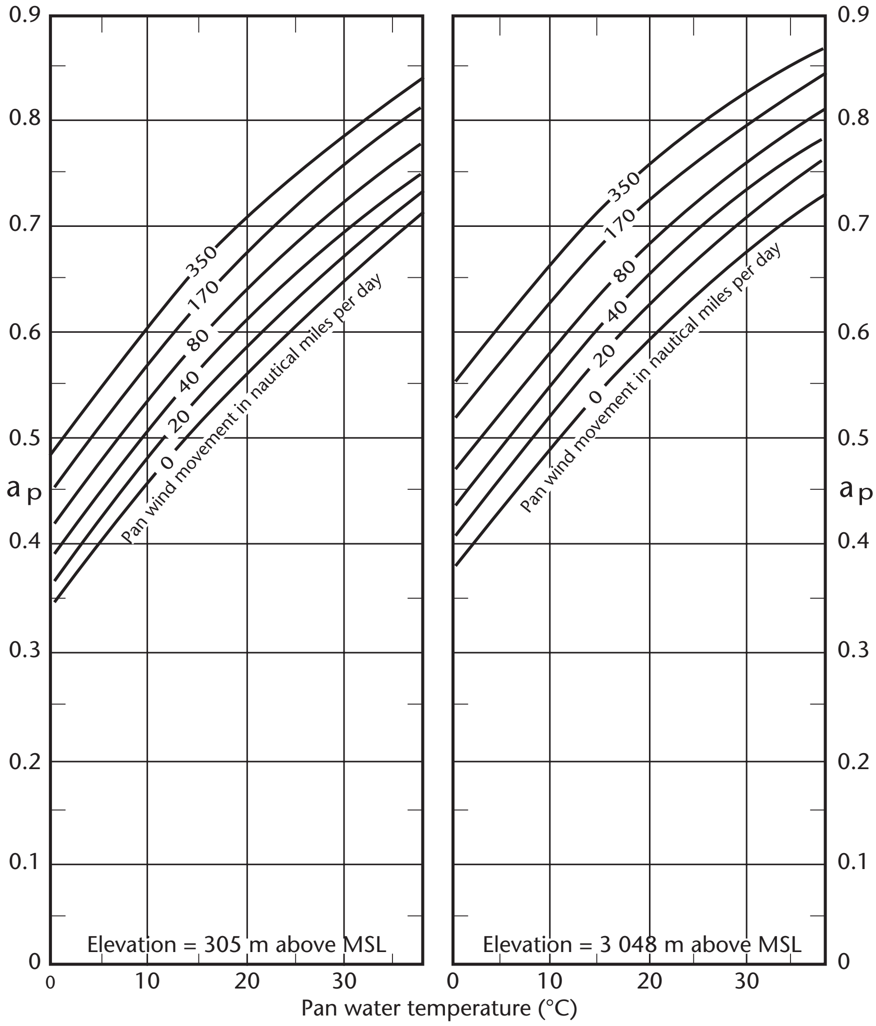


Figure I.4.7. Proportion of advected energy into a Class A pan that is used in evaporation

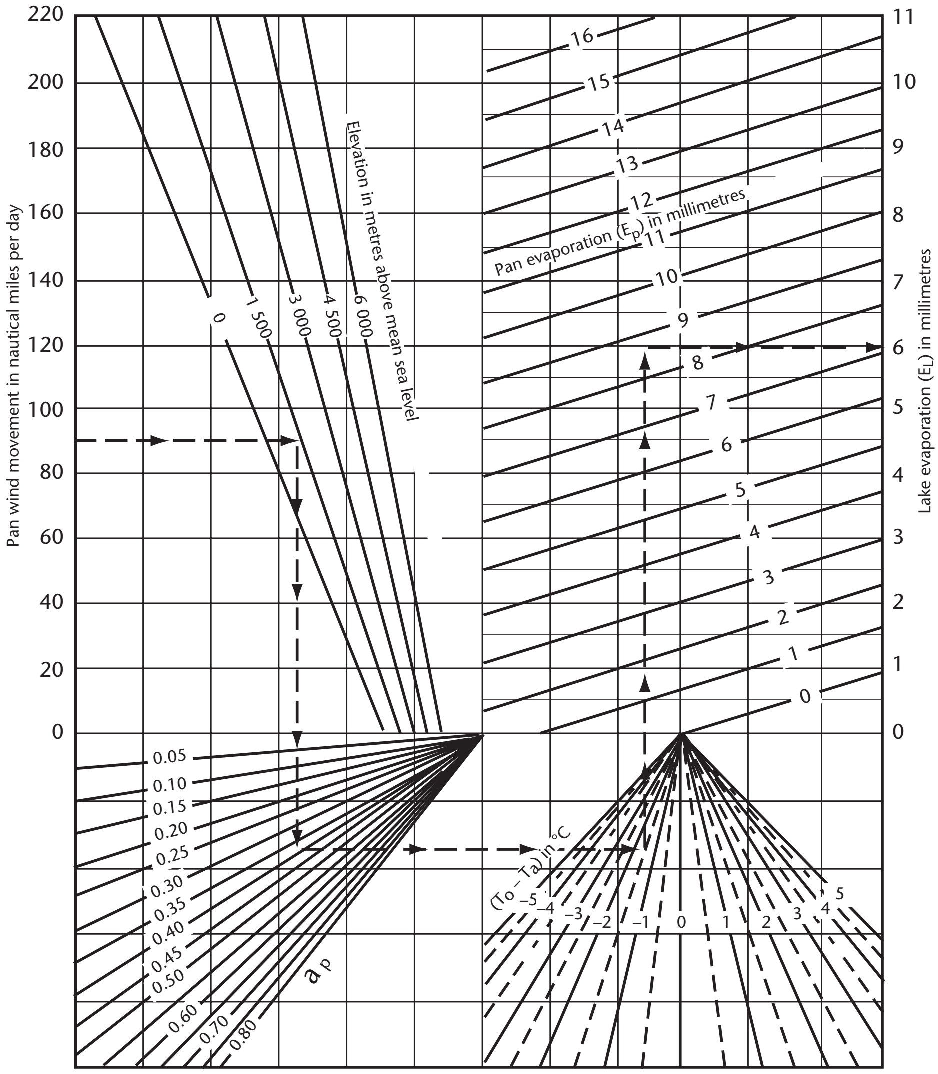


Figure I.4.8. Conversion of Class A pan evaporation into lake evaporation

fall in the categories of: (a) water balance methods, such as evapotranspirometers, hydraulic budget on field plots, and soil moisture depletion; (b) energy balance method; (c) mass-transfer methods, such as wind speed function, eddy flux and use of enclosures; (d) a combination of energy and mass-transfer methods, such as the Penman method; (e) prediction methods, such as the empirical equations and the indices applied to pan-evaporation data; and (f) methods for specific crops. These have been described in the *National Handbook of Recommended Methods for Water Data Acquisition* (USGS, 1977).

In the context of evapotranspiration, Thornthwaite and Holzman (1941) introduced the term "potential evapotranspiration" to define the evapotranspiration that will occur when the soil contains an adequate moisture supply at all times, that is, when moisture is not a limiting factor in evapotranspiration. The prediction methods estimate potential evapotranspiration. Most other methods apply to estimation of actual evapotranspiration under the condition of sufficient water at all times. The actual evapotranspiration from potential evapotranspiration is derived using a simple soil moisture function, $f(\phi)$ (Saxton and others, 1986):

$$\lambda E_{actual} = f(\phi)^{*} \lambda E \qquad (4.25)$$

where λE_{actual} is the actual evapotranspiration and the soil moisture function is a dimensionless variable estimated by a simple linear model. The soil moisture function is defined by the following:

$$f(\phi) = \text{M/Field capacity} \qquad (4.26)$$

where M is soil volumetric moisture at 20-cm depth (at rooting zone). Field capacity can be defined as the percentage of water remaining in a soil two or three days after it has been saturated and after free drainage has practically ceased. It has been shown (Brandes and Wilcox, 2000) that simple linear models of the evapotranspiration/soil moisture process are appropriate for hydrological modelling.

4.3.2 **Water-budget method**

The water-budget approach can be used to estimate evapotranspiration, ET, when precipitation, P, stream runoff, Q, deep seepage, Q_{ss}, and changes in storage, ΔS, can be measured or estimated. The equation is:

$$ET = P - Q - Q_{ss} \pm \Delta S \qquad (4.27)$$

The annual evapotranspiration from a basin for a water year can be estimated as the difference between precipitation and runoff if it can be established by hydrogeological studies that deep seepage is relatively insignificant. The date chosen for the beginning and ending of the water year should coincide with the dry season, when the amount of water in storage is relatively small and the change in storage from year to year is negligible.

If evapotranspiration is to be estimated for a shorter period, such as a week or a month, the amount of water storage in the ground and in the stream channel must be measured. This is feasible only on small basins, and application of the water-budget approach for such short periods is generally limited to experimental plots or catchments of a few acres.

For average annual evapotranspiration, the change in storage is usually negligible, and evapotranspiration can be estimated by the difference between average annual precipitation and average annual runoff.

The various terms of the above equation can be measured by conventional methods. The precipitation measurements can be made by a network of raingauges. For this purpose non-recording raingauges are adequate. The number of such raingauges would depend upon the expected variability of precipitation over the catchment. The streamflow measurements can be done by continuous measurement (Chapter 5). The change in water storage in the ground can be measured in two separate components, that is, the saturated and unsaturated components. For this purpose measurement of water table elevation in wells and measurement of soil moisture in the saturated zone are required. The elevation of the water table can be determined by measuring the distance from reference point to the water surface in wells at the end of each time period for which evapotranspiration is to be computed. The change in volume of water storage is equal to the average change in water elevation x the specific yield of the formation x the area of the catchment. Soil-moisture profiles from the saturation level (or to a point of constant soil moisture in arid regions) to the ground surface should be measured at the end of each computation period at a number of points over the catchment. The gain or loss of soil moisture during the period can then be computed. The amount of water that moves from or to the catchment as deep seepage cannot be measured directly. A hydrogeological study of the hydraulic characteristics of the underlying formations should indicate the relative magnitude of this flow, which must be considered when choosing the experimental area. This item should be small enough so that it can be neglected in water-budget studies.

4.3.3 **Energy-budget method**

This method (WMO, 1966) may be applied for the estimation of evapotranspiration when the difference between radiation balance and the heat flux into the soil is significant and exceeds the errors of measurement (4.2). This method is applied for estimation of evapotranspiration for periods of not less than 10 days. For shorter periods, the estimation of evapotranspiration by the energy-budget method is rather difficult.

Assuming that the surface energy balance equation is the primary boundary condition to be satisfied in computing ET, there are three techniques to solve the energy-balance equation. The first technique uses semi-empirical methods, the second employs analytical methods and the third utilizes numerical models.

The semi-empirical methods represent an effort to obtain a manageable model to estimate ET. These modern operational approaches are derived chiefly from Penman's original formulation, which is a combination of the diffusion and energy-balance approaches (Bailey, 1990). The Jackson model (Jackson and others, 1977) was later evaluated using empirical and theoretical results (Seguin and Itier, 1983). The energy-balance model is integrated over a 24-hour period and thus assumes that the soil heat flux is negligible. Furthermore, observations (Itier and Riou, 1982; Brunel, 1989) suggest that the daily ratio of sensible heat flux to the net radiation flux, R_n, can be approximated by that ratio estimated near midday under clear sky conditions. With some further approximations the energy-balance model can be recast as:

$$LE = R_n - B\,(T_s - T_a)_i + A \qquad (4.28)$$

where LE is the latent heat flux (evapotranspiration, ET), T_s is the surface temperature estimated remotely, say from a satellite-based thermal IR sensor, T_a is the near-surface air temperature obtained from a nearby weather station, the subscript i represents the "instantaneous" observation by the satellite over the area of interest, and A and B constants which vary with location (Caselles and Delegido, 1987). In practice, however, A and B vary with a wide range of both meteorological and surface factors (Bailey, 1990). This expression and derivatives of it have been tested and shown to produce reasonable estimates of daily ET (Brunel, 1989; Kerr and others, 1987; Nieuwenhuis and others, 1985; Rambal and others, 1985; Thunnissen and Nieuwenhuis, 1990; Riou and others, 1988). Although equation 4.28 is characterized by low demands for data provision and ease of operation, it is also characterized by limited spatial and temporal areas of application together with poor accuracy especially in the presence of cloud when using satellite thermal infra-red methods to obtain T_s (Bailey, 1990).

According to WMO, Germany is utilizing NOAA AVHRR data for input into numerical evaporation models in small-scale agricultural areas. Satellite data include vegetation, land-surface temperature gradients, soil moisture, diurnal temperature variations and solar irradiance. Extrapolation of the model results are to be tested (WMO, 1992*a*).

4.3.4 Aerodynamic approach

The application of this method (WMO, 1966) for the estimation of evapotranspiration is difficult because of the lack of reliable methods to determine the turbulent-exchange coefficient (4.2). Thus, it is seldom used. It is used only for approximate estimation of evaporation.

In some countries, evapotranspiration is estimated by empirical methods, the Penman method and the Thornthwaite formula. Penman's method is used in conditions of sufficient moisture, and the Thornthwaite formula (Thornthwaite and Holzman, 1941) is applied for regions with climatic conditions similar to those of the middle Atlantic coast of the United States on which this formula was based.

In the Commonwealth of Independent States, Konstantinov's method (Konstantinov, 1966) is applied for the estimation of evaporation based on observations of temperature and humidity of the air in a psychrometer shelter at 2 m above the ground. This method is mainly applicable for the computation of long-term mean monthly, seasonal or annual evapotranspiration.

4.3.5 Penman–Monteith method

The combination equation 4.14 represents the energy budget at the land surface and the transfer of water vapour and heat between the surface and the atmosphere. The Penman–Monteith method (Monteith, 1965) introduces aerodynamic and surface resistances. The former describes the effect of surface roughness on heat and mass transfer and the latter describes the resistance to the flow of water vapour between the evaporating surface and the air. Surface resistance for water surfaces is zero. In the case of vegetation, the surface resistance represents biological control of transpiration and is largely controlled by stomatal resistance. For drying soil, the surface resistance depends on soil moisture availability. This method may be used on an hourly or daily basis. However, its use is restricted because it requires sub-models for the surface resistance.

The Penman–Monteith model is expressed as:

$$\lambda E = (\Delta\Delta + C_p \rho D / r_{aa}) / (\Delta + \gamma + \gamma\, (r_{cs} / r_{aa})) \tag{4.29}$$

where r_{aa} is the aerodynamic resistance above the canopy, and r_{cs} is stomatal resistance of the canopy. For the Shuttleworth–Wallace model (Shuttleworth and Wallace, 1985), λE is separated into evaporation from the soil (λE_s) and transpiration from the canopy (λE_c), which are derived from the Penman–Monteith combination equations:

$$\lambda E_s = (\Delta\Delta_s + \rho c_p D_0 / r_{sa}) / (\Delta + \gamma (l + r_{ss} / r_{sa})) \tag{4.30}$$

$$\lambda E_c = (\Delta\Delta(-A_s + \rho c_p D_0 / r_{ca}) / (\Delta + \gamma (l + r_{cs} / r_{ca})) \tag{4.31}$$

where A_s is available soil energy, D_0 is vapour pressure deficit in the canopy, r_{sa} is the aerodynamic resistance between the substrate and canopy source height, r_{ca} is the boundary layer resistance of the vegetation, and r_{ss} is soil resistance. The aerodynamic resistance above the canopy (r_{aa}) and the aerodynamic resistance between the substrate and canopy source height (r_{sa}) are functions of leaf area index, eddy diffusivity decay constant, roughness length of the vegetation (function of vegetation height), zero plane displacement (function of vegetation height), a reference height above the canopy where meteorological measurements are available, wind speed, von Karman's constant, and roughness length of the substrate. D_0 is derived from the Ohm's law electrical analog for the vapour pressure and temperature difference between the canopy and the reference height above the canopy where fluxes out of the vegetation are measured. D_0 is a function of the measurable vapour pressure deficit at the reference height, D:

$$D_0 = D + (\Delta\Delta - r_{aa}\lambda E_c(\Delta + \gamma))/\rho c_p \qquad (4.32)$$

and D can thus be substituted for D_0 into the combination equations. The total evaporation from the crop, λE, for the Shuttleworth–Wallace model is the sum of the Penman–Monteith combination equations with D substituted for D_0:

$$\lambda E = C_c PMM_c + C_s PM_s \qquad (4.33)$$

where PM_c describes evaporation from the closed canopy, and PM_s describes evaporation from the bare substrate. The new Penman–Monteith equations have the form:

$$PM_c = \frac{(\Delta\Delta + (\rho\rho_p D - \Delta r_{ca} A_s) / (r_{aa} + r_{ca}))}{(\Delta + \gamma\,(1 + r_{cs} / (r_{aa} + r_{ca})))} \qquad (4.34)$$

$$PM_s = \frac{(\Delta\Delta + (\rho\rho_p D - \Delta r_{sa} A_s) / (r_{aa} + r_{sa}))}{(\Delta + \gamma\,(1 + r_{ss} / (r_{aa} + r_{sa})))} \qquad (4.35)$$

The coefficients C_c and C_s are resistance combination equations:

$$C_c = l/(l + R_c R_a/(R_s(R_c + R_a))) \qquad (4.36)$$

$$C_s = l/(l + R_s R_a/(R_p(R_s + R_a))) \qquad (4.37)$$

where

$$R_a = (\Delta + \gamma)r_{aa} \qquad (4.38)$$

$$R_s = (\Delta + \gamma)r_{sa} + \gamma r_{ss} \qquad (4.39)$$

$$R_c = (\Delta + \gamma)r_{ca} + \gamma r_{cs} \qquad (4.40)$$

4.3.6 Priestley–Taylor (radiation) method

The method of Priestley and Taylor (Priestley and Taylor, 1972) is based on the argument that, for large, wet areas, radiation controls of evaporation must dominate rather than advective controls. If the atmosphere remains saturated when in contact with the wet surface, then the latent-heat transfer (evaporation) may be expressed by:

$$\lambda E = \left(\frac{\varepsilon}{\varepsilon + 1}\right)(Q^* - G) \qquad (4.41)$$

where Q* is the available net radiation, G is the soil-heat flux, and ε equals $s\lambda/c_p$, with s equal to the slope of the saturation specific humidity curve, λ is the latent heat of vaporization, and c_p is the specific heat of water.

For equilibrium evaporation, it is proposed that:

$$\lambda E = \alpha \left(\frac{\varepsilon}{\varepsilon + 1}\right)(Q^* - G) \qquad (4.42)$$

with α = 1.26, an empirical constant. This expression is used as an estimate of potential evaporation in the absence of local advection. It also gives good estimates for evaporation from well-watered but not wet vegetation in much smaller regions.

4.3.7 Complementary method

The complementary method, first suggested by Bouchet (1963), is increasingly used in hydrological applications for large areas because it essentially uses standard climatic data.

The method considers that potential evaporation is as much the effect of the actual evaporation as its cause. Heat and moisture released from the surface will modify the temperature and humidity of the air above it. It has been suggested that the increase in potential evaporation observed when an area dries out may be used as a measure of the actual evaporation rate.

If actual evaporation E is reduced below the potential rate E_{po} for an extensive wet region, then an amount of energy Q would be released, so that:

$$\lambda E_{po} - \lambda E = Q \qquad (4.43)$$

This energy change will affect temperature, humidity, turbulence and hence evaporation. If the area is big enough so that the change in energy does not result in changes in the transfer of energy between the modified air mass and that beyond, Q should equal the increase in λE_p, the potential evaporation for the drying region.

Hence:

$$\lambda E_p - \lambda E_{po} = Q \qquad (4.44)$$

Therefore:

$$E + E_p = 2\,E_{po} \qquad (4.45)$$

Most applications of the complementary relationship (Morton, 1982) have been concerned with finding appropriate expressions for E_p and E_{po}. These may be estimated with equation 4.15 and the

Priestley–Taylor method given in 4.3.6, respectively. The approach does not consider advection and assumes Q to remain constant. Also, the vertical exchange of energy, that is, with air masses brought in by large-scale weather systems, is not considered.

4.3.8 Crop coefficient and reference evapotranspiration method

In 1998, *Crop evapotranspiration – Guidelines for computing crop water requirements* (FAO-56 report), recommended a new standard for reference crop evapotranspiration using the Blaney–Criddle, Penman, radiation and pan evaporation methods. The FAO-56 approach (FAO, 1998; Allen 2000) first calculates a reference evapotranspiration (ET_o) for grass or an alfalfa reference crop and then multiplies this by an empirical crop coefficient (K_c) to produce an estimate of crop potential evapotranspiration (ET_c). The ET_c calculations used the dual crop coefficient approach that includes separate calculation of transpiration and evaporation occurring after precipitation and irrigation events.

The FAO-56 Penman–Monteith method computes reference evapotranspiration from net radiation at the crop surface, soil heat flux, air temperature, wind speed and saturation vapour pressure deficit. The crop coefficient is determined from a stress reduction coefficient (K_s), a basal crop coefficient (K_{cb}) and a soil water evaporation coefficient (K_e). The K_{cb} curve is divided into four growth stages: initial, development, mid-season and late season. Field capacity and wilting point estimates determine soil water supply for evapotranspiration. The downward drainage of the topsoil is included but no upward flow of water from a saturated water table was considered, possibly causing some overprediction of water stress between the known irrigations. Water stress in the FAO-56 procedure is accounted for by reducing the value of K_s.

4.3.9 Large aperture scintillometer

Estimation of actual evapotranspiration using the energy-balance method requires knowledge of the sensible heat flux. According to the Monin–Obukhov similarity theory, the sensible heat flux, H, is related to the structure parameter of temperature, C_T^2. A large aperture scintillometer is an instrument to collect path-average values of C_T^2 (de Bruin and others, 1995). The scintillometer directs a light source between a transmitter and receiver and the receiver records and analyses fluctuations in the turbulent intensity of the refractive index of the air. These fluctuations are due to changes in temperature and humidity caused by heat and moisture eddies along the path of the light. Additional data on temperature, pressure and humidity are necessary to compute the characteristic parameter of the refractive index. This can then be converted to sensible heat flux. An important feature of the scintillometer technique is that although the measurement is along the path of the light beam, because of the effects of wind, this is actually an estimate of H over an area. The method therefore forms an intermediate level between the field scale measurements and the large area remote-sensing estimates.

4.4 EVAPORATION REDUCTION

4.4.1 From free surfaces

Evaporation losses from a fully exposed water surface are essentially a function of the velocity and saturation deficit of the air blowing over the water surface, and the water temperature. Evaporation losses are held to a minimum by:

(a) Exposing the least possible water-surface area. This in turn means that streams and reservoirs should be kept deep instead of wide;
(b) Covering the water surface;
(c) Controlling aquatic growth;
(d) Creating afforestation around reservoirs that would act as windbreakers. However, this method has been found to be useful under limited conditions for small ponds;
(e) Storing water underground instead of creating a surface reservoir. To accomplish this there are physical and legal problems in preserving the water so stored from adverse withdrawal;
(f) Making increased use of underground water;
(g) Integrated operation of reservoirs;
(h) Treatment with chemical water evaporation retardents (WER).

The first seven methods mentioned above are direct and easily understandable methods. However, the last method needs some explanation. This method comprises dropping a fluid on the surface of the water so as to form a monomolecular film. The problem with the film, however, is that it becomes damaged by wind and dust, and is too rigid to enable repair of the film thus damaged. Chemicals such as hexadecanol and octadecanol, of course, can be used for the purpose (Gunaji, 1965).

Studies by the Bureau of Reclamation indicate that evaporation may be suppressed by as much as 64 per cent with hexadecanol films in 1.22-m diameter pans under controlled conditions. Actual reduction on large bodies of water would, of course, be significantly less than this because of problems of maintaining the films against wind and wave action. Evaporation reduction to the extent of 22 to 35 per cent has been observed on small lakes of roughly 100 ha in size with reductions of 9 to 14 per cent reported on larger lakes (La Mer, 1963).

In Australia, evaporation reduction to the extent of 30 to 50 per cent has been observed on medium lakes of roughly 100 ha in size. Although the use of the monomolecular film is still in the research stage, its relative case means that some measure of evaporation control can be obtained through this technique.

4.4.2 From soil surface

There are various methods of controlling evaporation losses from soil (Chow, 1964).

(a) *Dust mulch*: This is an age-old practice in cultivation of soil to keep it loose on the surface. It is based on the theory that loosening the surface will permit rapid drying and reduce contact between soil particles. Rapid drying will develop dry soil to act as a blanket to suppress evaporation. Reducing points of contact between soil particles will lessen capillary rise.
 It has been found that soil cultivation by tillage may be necessary only to kill weeds and keep the soil in a receptive condition to absorb water and deep tillage is futile as a means of overcoming drought or increasing yield. Experiments have also shown that mulching not only decreased the amount of water in the soil, but also caused loss of more moisture than in the bare, undisturbed soils. In tank and field trials it has also been found that mulching by thorough cultivation at weekly intervals failed to save soil moisture, but the surface shallow layer, by drying quickly, acted as a deterrent to further loss of moisture.
 Since these early investigations, the results of many others have been published. Many agricultural experiment stations have studied this problem, resulting in conclusions similar to those mentioned. Various experiments have also indicated that the soil mulch can reduce moisture loss only when the water table, perched or permanent, is within the capillary rise of the surface;

(b) *Paper mulch*: Covering the soil with paper to reduce evaporation was widely used in the late 1920s, but is now rarely done as it has been found that the effect of paper mulch is confined to limited surface of soil, which again is due to condensation of water beneath the paper;

(c) *Chemical alteration*: Experiments in the early 1950s indicated that chemical alteration of the soil moisture characteristics may decrease evaporation. The addition of polyelectrolytes to soils decreases the rate of evaporation and increases the water available to plants;

(d) *Pebble mulch*: In China this method has been used for partial control of evaporation in some dry areas.

4.5 SOIL MOISTURE MEASUREMENT [HOMS E55]

4.5.1 General

Below the surface of the Earth there exists a huge reservoir of freshwater. This subsurface water can be divided into soil moisture, vadose water, shallow groundwater and deep groundwater. The zones of soil moisture and vadose water are together known as the zone of aeration. The amount of water held as soil moisture at any time is an insignificant amount by comparison with the Earth's total available water, but it is crucial to plant life and food production and thus vital to life.

Soil moisture can be defined as the water held in the soil by molecular attraction. The forces acting to retain water in the soil are adhesive and cohesive forces. These forces act against the force of gravity and against evaporation and transpiration. Thus, the amount of moisture in the soil at any given time is determined by the strength and duration of the forces acting on the moisture, and the amount of moisture initially present.

Natural sources of soil water such as rainfall and snow melt are normally greatly reduced during drought. Slope shape, gradient and soil surface roughness will affect soil water content since surface or subsurface run-on from adjacent upslope sites can add to the soil moisture, while surface runoff can remove water from a site. Evaporation, evapotranspiration and deep percolation beyond rooting depth are other factors that deplete soil moisture.

Hence, soil water content must be defined in specific quantitative terms to accurately indicate the amount of water stored in the soil at any given time. At saturation, after heavy rainfall or snow melt, some water is free to percolate down through the soil profile. This excess water is referred to as gravitational water and can percolate below the rooting depth of some plants. Here it is important to define some terms in relation to soil moisture. Field capacity is defined as the amount of water remaining in the soil after percolation has occurred. Wilting point is defined as the soil water content at which the potential of plant roots to absorb water is balanced by the water potential of the soil. The amount of water between field capacity and wilting point is generally considered plant available water content although plants can also extract gravitational water while it is available.

The moisture content of the soil is a key component in making irrigation scheduling decisions. The root zone serves as a reservoir for soil moisture. During the rainy season the moisture content is high, but at harvest time the soil is commonly depleted of moisture. Thus the measurement of soil moisture is an important factor in preventing overirrigation resulting in wastage of water and leaching of fertilizers or under-irrigation, that result in water deficit.

Soil moisture is measured in two distinctly different methods: quantitatively and qualitatively, which is an indication of how tightly the water is held by the soil particles.

4.5.2 Quantitative methods

4.5.2.1 Gravimetric method (Oven dry and weigh)

The gravimetric method is one of the direct methods of measuring soil moisture. It involves collecting a soil sample (usually 60 cm^3), weighing the sample before and after drying it, and calculating its moisture content. The soil sample is considered to be dry when its weight remains constant at a temperature of 105°C. Many different types of sampling equipment, as well as special drying ovens and balances, have been developed and used for this method.

The gravimetric method is the most accurate method of measuring moisture content in the soil and serves as the standard for calibrating the equipment used in all other methods. However, it cannot be used to obtain a continuous record of soil moisture at any one location because of the necessity of removing the samples from the ground for laboratory work.

Sample collection

The procedure for collecting a sample for the gravimetric method depends on whether the soil-moisture determination is to be based on the dry mass of the sample or on its volume for dry-mass determination, but not for volumetric determination. The sample can be disturbed for dry-mass determination, but not for volumetric determination. Soil sampling is fraught with difficulties if the soil is very dry or very wet or if it contains stones or other material that preclude easy cutting by the sampling equipment.

The technique and equipment used for sample collection should be such that the samples do not lose or gain moisture or otherwise become altered or contaminated during sampling and transportation. When sampling through a wet layer into a dry layer, care must be taken to keep the sampling equipment as dry as possible and to prevent water from running down the hole into the drier material. If there is free water in the soil, the measured moisture content probably will be less than the correct value because some water will drip off as the sample is removed from the ground, or some may be squeezed out by compaction during sampling.

When dry, hard, fine-textured sediments are encountered it is difficult to drive the core barrels or to rotate the augers. When dry, coarse-textured sediments are sampled, the sample may slide out at the end of the core barrel or auger as it is withdrawn. Stony soils are very difficult to sample, especially volumetrically, because of the likelihood of hitting a stone with the cutting edges of the equipment and because representative samples must be large. Soils that contain a considerable amount of roots and other organic matter also present difficulty.

The amount of soil taken for the gravimetric moisture determination of a gravel soil needs to be substantially more than for non-gravel soils and depends proportionally on the size and content of the gravel. Moisture is determined as a percentage by mass (weight). If multiplied by bulk density, moisture as a percentage of volume is obtained.

In soil-moisture sampling, it is essential that all sampling operations, as well as the transfer of samples to cans, and the weighing of the moist samples be done as rapidly as possible to minimize moisture losses. Many difficulties in the use of sampling equipment may be avoided if the equipment is kept clean and free of moisture and rust.

Description of samplers

Auger samplers (Figure I.4.9)

The simplest equipment for soil-moisture sampling is the hand auger. Hand augers, with shaft extensions of aluminium pipe, have been used in sampling to depths as much as 17 m. One of the most useful types of hand augers consists of a cylinder 76 mm in diameter and 230 mm long, with a 1.4-m extension pipe on the top and two curved, cutting teeth on the bottom. Because the barrel is a solid cylinder, the sample is not as likely to become contaminated from the side of the test hole. Thus, a good, representative, but disturbed, sample is obtained by using this equipment. For ease in sampling at depths greater than 1.5 m, 0.9 m extensions of 19-mm aluminium pipe are added, as needed (Figure I.4.10).

To obtain a sample by the hand-auger method, the auger has to penetrate usually about 80 mm of the material in order to fill the cylinder barrel. The auger is then raised to the surface, and the barrel is struck with a rubber hammer to jar the sample loose.

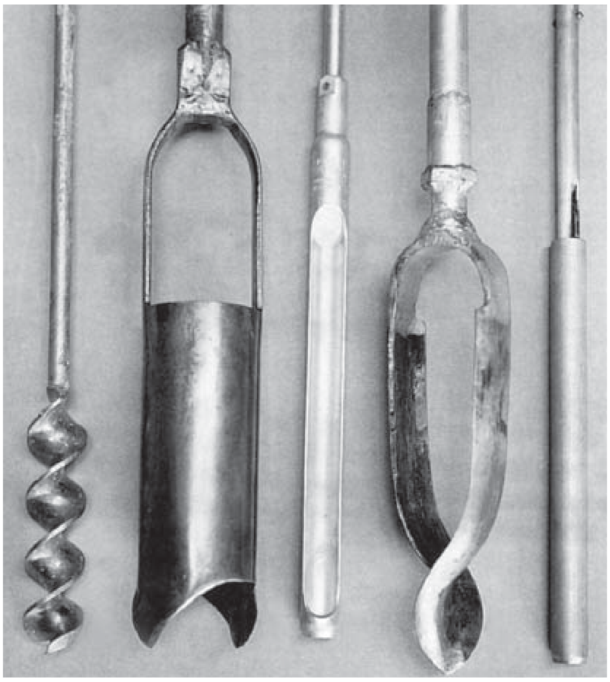


Figure I.4.9. Soil augers and tubes (left to right: screw or worm auger; barrel auger; sampling tube; Dutch mud auger; peat sampler)
(Source: http://soils.usda.gov/technical/manual/print_version/complete.html)

Tube or core-barrel samplers (Figure I.4.9)

A soil-sampling tube, core barrel or drive sampler offers an advantage in soil-moisture sampling as volumetric samples can be obtained for calculating moisture content by volume. Core samplers provide uncontaminated samples if the equipment is kept clean. Oil should never be used on the samplers, and they should be kept free of dirt, rust and moisture. A two-person crew is normally recommended for deep sampling, and depths of 20 m may be sampled (Figure I.4.11). A volume of soil core of at least 100 cm^3 is recommended.

The open-drive sampler consists of a core barrel of 50 mm inside diameter and 100 mm long, with extension tubes of 25 mm in diameter and 1.5 m long for sampling at depth. Brass cylinder liners, 50-mm in length, are used to retain the undisturbed core samples. The samples are removed from the core barrel by pushing a plunger. A light drill rod or 15-mm pipe may be used as extensions.

A simple and economical sampler for obtaining volumetric core samples from shallow depths consists of a thin-walled brass tube 50 mm in diameter and 150 mm long mounted on the end of a 90-cm T-handle of 19-mm pipe. After samplers are removed from the hole, they are pushed out of the core barrel by the central plunger. Since the inside diameter of the core barrel is known, volumetric samples may be obtained easily by cutting off predetermined lengths of the core as it is removed from the sampler.

Laboratory procedure

First, the wet soil samples are weighed individually in their transport containers. The containers are then opened and placed in a drying oven that is capable of maintaining a temperature of 105°C ±0.5. For samples that contain peat or significant

Figure I.4.10. Soil sampling kit
(Source: http://www.colparmer.com/catalog/product_view.asp?sku=9902640)

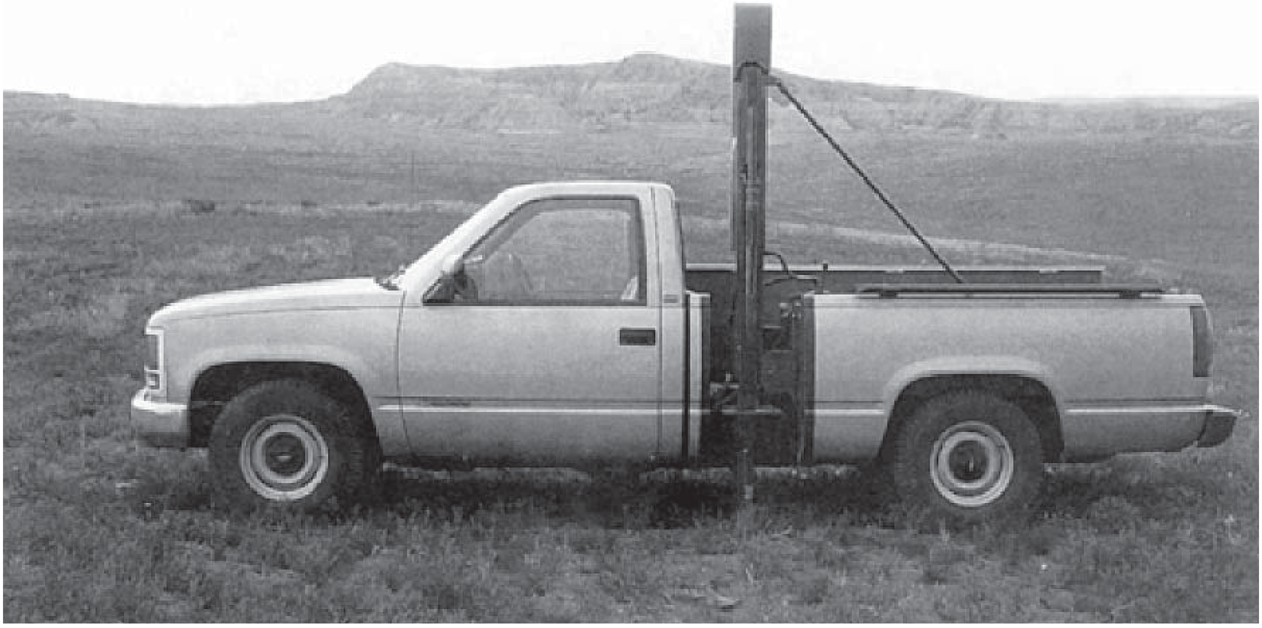


Figure I.4.11. A hydraulically operated sampling tube mounted on a small lorry. The open-faced tube is in place. Hydraulic controls are at the right.

amounts of gypsum, the oven temperature should be 50°C ±0.5, which will then require a longer time for the sample to reach a dry state.

After drying, the samples are reweighed in their containers. The difference in the wet and dry weights for a sample is the measure of its original water content. Other drying processes that are faster than the standard oven may be used, for example, alcohol roasting, infra-red lamps and microwave ovens.

If the samples contain gravel and stones, the above procedure can be modified if the weights or volumes of the gravel and/or stones can be determined separately.

The advantages and disadvantages of the method are given below.

Advantages: This technique is relatively inexpensive, simple and highly accurate.

Disadvantages: This technique is time-consuming, labour-intensive and difficult in rocky soils.

4.5.2.2 **Neutron scatter method** [HOMS C58]

The neutron method indicates the amount of water per unit volume of soil. The soil volume measured by this method is bulb-shaped and has a radius of 1 to 4 m, according to the moisture content and the activity of the source.

This method is based on the principle of measuring the slowing of neutrons emitted into the soil from a fast-neutron source (Greacen, 1981). The energy loss is much greater in neutron collisions with atoms of low atomic weight and is proportional to the number of such atoms present in the soil. The effect of such collisions is to change a fast neutron to a slow neutron. Hydrogen, which is the principal element of low atomic weight found in the soil, is largely contained in the molecules of the water in the soil. The number of slow neutrons detected by a counter tube after emission of fast neutrons from a radioactive source tube is electronically indicated on a scale.

Instruments

A typical set of equipment consists of a portable battery-powered or spring-wound timer that has a time-accounting range of 0.5 to 5 minutes and weighs approximately 16 kg, and a moisture probe containing a 100-mCi fast-neutron source of americium-241 and finely ground beryllium (half-life, 458 years). The probe has a length of about 400 mm, a diameter of about 40 mm and a weight of 20 kg when complete with a lead and paraffin shield that is 150 mm in diameter and 100 mm long (Figure I.4.12). These probes have been used with up to 60 m of cable.

The source and detector are lowered into the soil through a hole cased with aluminium tubing, and readings can be taken at any depth except close to the surface. The inside diameter of the tube should be only slightly larger than the diameter of the probe. The tube should be installed by augering the soil inside the tube, if possible, to ensure close contact between the outside surface of the tube and the soil.

Similar gauges have been developed to make measurements in the surface layers of the soil. In this case, the equipment is placed on the ground surface and gives the moisture content of a hemispherical volume of 15- to 40-cm radius.

Access tubes

The installation of access tubes must be performed carefully to prevent soil compaction and to ensure soil contact around the outside of the tubes, that is, no voids in the soil should be created outside the tubes during their installation. Access tubes may be installed:
(a) By inserting the tubes into prepared holes of the same or slightly smaller diameter (the holes can be prepared by using either a hand-powered or motorized auger);
(b) By driving the tubes into the soil with a hammer and then removing the soil from inside the tubes with an auger.

The bottom ends of the tubes should be sealed to prevent infiltration of groundwater. The top ends of the tubes should be sealed with a cap or a stopper when not in use.

Calibration

The probe should be calibrated by gravimetric sampling (4.5.2.1) of the type of soil that is to be tested and in the size and type of casing into which the probe is to be lowered. Sufficient samples should be taken around the test hole to define the soil moisture profile. It is difficult to obtain a good calibration in heterogeneous soil or when soil moisture is changing rapidly with depth. An approximate calibration can also be carried out in the laboratory by using a container filled with soil material. The type and size of casing and the method of installation of the access tube have a considerable effect on the readings, and new calibration curves should be obtained for each type of installation.

Measurements and accuracy

The access tubes must be kept free of excess moisture or erroneous readings will result.

After lowering the probe to the proper depth in the access tube, the number of counts over a known time period is determined. The average count is converted to soil moisture content by using the calibration curve. The accuracy of a determination depends primarily on:
(a) The validity of the calibration curve;
(b) The number of counts per determination.

Because of the randomness of the emission and the impact of neutrons, random count errors can occur. Timing errors may be kept to a minimum by using a standard-count timing cycle of two minutes.

Salt concentrations in the range ordinarily found in soil moisture do not materially affect data obtained by the neutron method, but at salt concentrations at the level of seawater, the effect is appreciable. There is some evidence of a temperature effect.

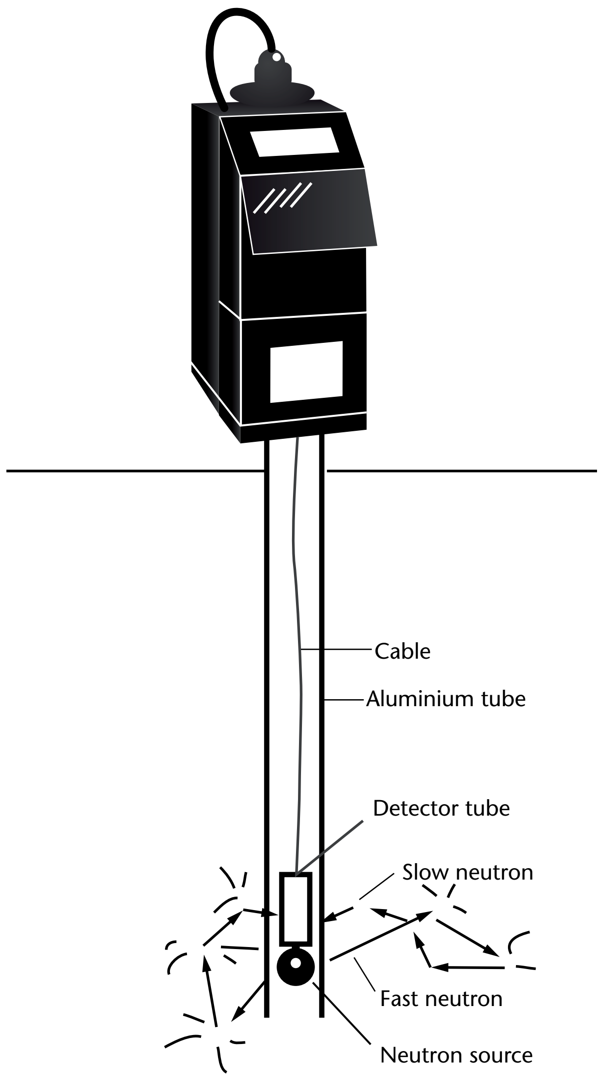


Figure I.4.12. Neutron probe

Readings close to the surface are affected by the position of the probe with respect to the air-soil interface. Proximity of the interface causes lower counts than would be indicated for the same moisture content at a greater depth.

When the error sources are minimized, the accuracy of an individual determination can reach 0.5 to 1 per cent. For repeated determinations over time, such as might be performed in a water-balance study, the changes in water content of soil can be even more accurate because of the elimination of systematic errors.

The advantages and disadvantages of the method and the availability of instruments for its use are summarized below (Prichard, 2003):

Advantages: The neutron probe allows a rapid, accurate, repeatable measurement of soil moisture content to be made at several depths and locations.

Disadvantages: The use of radioactive material requiring a licensed and extensively trained operator, the high equipment cost and extensive calibration required for each site.

Readily available instruments: Neutron probes are available commercially.

4.5.2.3 **Dielectric methods** [HOMS C60]

The dielectric constant methods seek to measure the capacity of a non-conductor (soil) to transmit high-frequency electromagnetic waves or pulses. The resultant values are related through calibration to soil moisture content.

The basis for use of these instruments is that dry soil has dielectric values of about 2 to 5 and that of water is 80 when measured between 30 MHz and 1 GHz.

Two approaches have been developed for measuring the dielectric constant of the soil water media and estimating the soil volumetric water content:
(a) Time domain reflectrometry (TDR);
(b) Frequency domain reflectrometry (FDR).

Neither TDR nor FDR use a radioactive source, thereby reducing the cost of licensing, training and monitoring when compared with the use of the neutron probe.

Time domain reflectrometry

The TDR device propagates a high-frequency transverse electromagnetic wave along a cable attached to a parallel conducting probe inserted into the soil. The signal is reflected from one probe to the other, then back to the meter, which measures the time between sending the pulse and receiving the reflected wave. By knowing the cable length and waveguide length, the propagation velocity can be computed. The faster the propagation velocity, the lower the dielectric constant and thus lower soil moisture.

Waveguides are usually a pair of stainless steel rods, which are inserted into the soil a few centimetres apart. The measurement is the average volumetric water content along the length of the waveguide if so calibrated. Waveguides are installed from the surface to a maximum depth of usually 45–60 cm. Pairs of rods can be permanently installed to provide water content at different depths. If deeper measurements are needed, a pit is usually dug after which the waveguides are inserted into the undisturbed pit wall. The soil disruption can change water movement and water extraction patterns, resulting in erroneous data.

TDR units are relatively expensive. However, once properly calibrated and installed, the TDR technique is highly accurate. Since surface measurements can be made easily and in multiple sites, it works well for shallow rooted crops.

Frequency domain reflectrometry

This approach uses radio frequency waves to measure soil capacitance. The soil acts as the dielectric completing a capacitance circuit, which is part of a feedback loop of a high-frequency transistor oscillator. The frequency varies between instrument manufacturers but is generally about 150 MHz. The soil capacitance is related to the dielectric constant by the geometry of the electric field established around the electrodes. The dielectric constant is then related to the volumetric water content as discussed in the TDR method. Two distinct types of instruments use the FDR techniques – an access tube method and a hand-held push probe.

Access tube type

An access tube of PVC material similar to one being used in the neutron probe and the electrodes is lowered into the access well and measurements are taken at various depths. It is necessary to ensure a very close fit between the walls of the access tube and the soil to ensure reliable values as air gaps affect the travel of the signal in the soil. Calibration to soil volumetric water content is required (especially in clayey soils and those with high bulk densities) to ensure accurate values. If properly calibrated and installed, the accuracy of the probe can be good.

Many of the advantages of the neutron probe are available with this system, including rapid measurements at the same locations and depths over time.

Another variant of this technology is the use of a permanent installation, which reads multiple depths. These are used in conjunction with electronics to make frequent readings and transmit results to a central data-collection device.

Hand-push probe

The other type of capacitance device is a hand-push probe, which allows rapid, easy, near-surface readings. These probes provide a qualitative measurement of soil water content on a scale from 1 to 100 with high readings indicating higher soil moisture content. Probe use in drier soils and those containing stones or hard pans is difficult. Deeper measurements are possible using a soil auger to gain access to deeper parts of the root zone. The probe is best used in shallow-rooted crops.

Advantages: The advantages of the TDR and FDR equipment is that they are relatively accurate (±1–2 per cent); can provide direct readouts of volumetric, available plant soil moisture percentages or continuous readings if used with a data logger; do not require calibration; and are relatively unaffected by salts in the soil. TDR is more accurate and less affected by salts while FDR can detect "bound" water in fine particle soils, which is still available to plants. Thus, the TDR instrument would be preferable for extensive acreage of salt-affected soils. However, if dealing with primarily fine-textured, non-saline soils, the FDR instrument would be preferable. In general, these instruments are accurate, reasonably priced, easy to use and very suitable for large areas.

Disadvantages: Owing to the cost of the instruments, these methods are more expensive than others. Readings can be affected if good contact is not made with the soil, and prongs can be damaged in hard or rocky soils. TDR has complex electronics and is the most expensive, whereas FDR is more susceptible to soil salinity errors. Data logger readings are in the form of graphs requiring interpretation.

4.5.2.4 Gamma-ray attenuation

The intensity of a gamma ray that passes through a soil section undergoes an exponential decrease that principally depends on the apparent density of the soil, the water contained in the soil and the coefficients of attenuation of the soil and of the water, which are constants. The method consists of concurrently lowering a gamma-ray source (generally Caesium 137) and a gamma-ray detector (scintillator-photomultiplier) down a pair of parallel access tubes that have been installed in the soil. At each measurement level, the signal can be translated into the apparent wet density of the soil or, if the apparent dry bulk density of the soil is known, the signal can be converted into a measure of the volumetric soil-moisture content.

The measuring equipment permits tracking of the evolution of wet density profiles and of the volumetric soil-moisture at several tens of centimetres of depth below the soil surface if the dry density does not vary with time.

The method has the advantage of having a high spatial resolution (it measures over a slice of soil 20 to 50 mm in thickness with the access tubes separated by about 3 m). However, the measurements are not specific to water alone. The apparent variations in dry density can confound the measurements of soil moisture.

Some complex equipment has two energy sources with different intensities of gamma rays, which permit the joint study of the variations in both apparent density and soil moisture. Such equipment is used primarily in laboratories and not under field conditions.

4.5.3 Qualitative methods

4.5.3.1 Tensiometric method [HOMS C62]

The components of a tensiometer include the porous cup, the connecting tube and/or the body tube and the pressure sensor. The porous cup is made of a porous, rigid material, usually ceramic. The pores of the cup wall are small enough to prevent the passage of air. A semi-rigid connecting tube and/or a rigid body tube are used to connect the tensiometer cup to the pressure sensor. The system is filled with water and the water in the point or cup comes into equilibrium with the moisture in the surrounding soil. Water flows out of the point as the soil dries and creates greater tension, or flows back into the point as the soil becomes wetter thereby decreasing the tension. These changes in pressure or tension are indicated on the measuring device. Multiple tensiometres located at several depths permit the computation of a soil-moisture profile.

Tensiometers provide data on soil-water potential (pressure components). If a tensiometer is used for moisture determinations, a calibration curve is needed. The calibration curve may be a part of the soil-moisture retention curve, but it is recommended that field data from the gravimetric method (4.5.2.1) and tensiometer readings be used for the calibration. Even so, the moisture data are only approximate, because of the hysteresis between the wetting and drying branches of the soil-moisture retention curve. The range of use is restricted to 0 to 0.8 bars (0 to 8 m of negative hydraulic head). Therefore, the method is suitable only for wet regions.

The pressure measuring device is usually a Bourdon-tube vacuum gauge or a mercury manometer. The tensiometer may also be attached to an electrical pressure transducer to maintain a continuous record of tension changes. Because the system is under a partial vacuum during unsaturated soil conditions, it is necessary that all parts or joints be impermeable to air. For field use, Bourdon vacuum gauges are more convenient than mercury manometers, but they have a lower accuracy. Electrical pressure transducers are both convenient and precise.

The tensiometer response time is much faster with pressure transducers that have small volume displacements than with other pressure sensors. The disadvantage of the cost can be offset by using only one electrical pressure transducer connected to several tensiometers via a scanning device. Another solution consists of using a measuring apparatus that briefly samples the pressure in the tensiometer by means of a needle. This needle perforates a special bulb on the tensiometer tube only during the moment of the measurement. A single needle apparatus can be used to sample numerous tensiometers placed in the soil. However, unlike the system described above, this type of tensiometer cannot be used to record changes of pressure potential.

Tensiometers should be filled with de-aerated water. Then it would be possible to remove air trapped inside the system by using a vacuum pump. Tensiometers are generally inserted vertically into the soil in pre-augered holes of the same diameter as the porous cup. The centre of the porous cup is located at the depth where pressure measurement is required. Tensiometers are affected by temperature fluctuations that induce thermal expansion or contraction of the different parts of the system and that influence the pressure readings. In the field, protection from solar radiation is recommended for tensiometers that are above ground to minimize this influence. Similarly, tensiometers used in the winter should be protected against frost damage to the water tube and the pressure sensor. Tensiometers need to be purged periodically to remove accumulated air from the system.

A tensiometer reading indicates the pressure in the porous cup minus the pressure difference caused by the water column between pressure sensor and porous cup. Therefore, the pressure potential of the soil water at the depth of the cup is the pressure sensor reading plus that of this water column. If the pressure is expressed in terms of suction, that is, atmospheric pressure minus gauge pressure, then the pressure potential of the soil equals the sensor reading minus the pressure difference caused by the water column in the tube. Corrected pressure potential of the soil can be generated directly with pressure transducer systems.

It is difficult to state the precision of a tensiometer measurement of soil-water pressure potential. The accuracy of a measurement is influenced by temperature, the accuracy of the pressure sensor and the quantity of air accumulated within the system. Moreover, the response time of tensiometers can cause erroneous measurements if the soil-water potential is changing quite rapidly in time. In this case, equilibrium between the soil water and the tensiometer water cannot be obtained. Recent studies have shown that semi-permeable plastic points provide much faster response than ceramic points (Klute, 1986).

The tensiometer is probably the easiest to install and the most rapidly read of all soil-moisture measuring equipment. However, tensiometers are not suitable for installation at depths greater than 3 m. At normal atmospheric pressures, the method is limited to a range of pressure potential down to about –85 kPa. Tensiometers require frequent servicing to obtain reliable measurements under field conditions.

Advantages: Tensiometers are not affected by the amount of salts dissolved in the soil water. They measure soil water tension with a reasonable accuracy in the wet range.

Disadvantages: Tensiometers only operate between saturation and about –85 kPa. Thus they are not suited for measurement in dry soils.

4.5.3.2 **Porous blocks/electrical resistance blocks** [HOMS C60]

Porous blocks are made of gypsum, glass/gypsum matrix, ceramic, nylon and fibreglass. They are buried at the depth of measurement desired. Over time, the blocks come to equilibrium with the moisture content in the surrounding soil. Therefore, the subsequent measurement is related to soil water tension.

In the case of electrical resistance blocks, two electrodes are buried inside the block with a cable extending to the surface. The electrical resistance is measured between the two electrodes using a meter attached to the cable. Higher resistance readings mean lower block water content and higher soil water tension.

Porous blocks require the same careful installation as tensiometers and good soil contact is important. Maintenance requirement is small and is much less than for tensiometers. Gypsum blocks are proven to break down in alkaline soils and will eventually dissolve, necessitating an abandonment or replacement. Soils high in soluble salts may cause erroneous readings, as salts influence soil conductivity and resistance. Gypsum blocks are best suited for fine-textured soils, as they are not generally sensitive below 1 000 hPa For most sandy soils, this would be outside the level of available water.

A newer type of gypsum block consists of a fine granular matrix with gypsum compressed into a block containing electrodes. The outside surface of the matrix is incised in a synthetic membrane and is placed in a perforated PVC or stainless steel protective cover. The construction materials enhance water movement to and from the block, making it more responsive to soil water tensions in the 300–2 000 hPa range. This makes them more adaptable to a wider range of soil textures.

Thermal dissipation blocks: These are made of a porous, ceramic material. Embedded inside a porous block is a small heater and temperature sensor attached by cable to a surface meter. A measurement is made by applying voltage to an internal heater and measuring the rate at which heat is conducted away from the heater (heat dissipation). The rate of heat dissipation is related to moisture content.

Thermal dissipation sensors are sensitive to soil water across a wide range of soil water contents; however, to yield water content they must be individually calibrated. These blocks are considerably more expensive than electrical resistance blocks.

Advantages: The method is quick, repeatable and relatively inexpensive.

Disadvantages: The blocks do not work well in coarse-textured, high shrink-swell, or saline soils. Accuracy is rather poor unless blocks are individually calibrated for the soil being monitored using a pressure plate extractor or gravimetric method. Blocks should be replaced every one to three years. Major consideration is that the sensitivity of the blocks is poor in dry soil conditions. The blocks need to be soaked in water for several hours before they are installed in the field.

4.5.4 **Remote-sensing** [HOMS D]

The remote-sensing technique is the most recent tool being used to estimate soil moisture properties at or near the surface. This information may be used to infer soil moisture profiles down to several metres. Remote-sensing of soil moisture can be accomplished using visible, infra-red (near and thermal), microwave and gamma data (Engman and Gurney, 1991; Schultz and Engman, 2000). However, the most promising techniques are based on the passive and active microwave data. The visible and near-infra-red techniques, which are based on the measurement of reflected solar radiation, are not particularly viable because there are too many noise elements that confuse the interpretation of the data. The thermal infra-red techniques are based on the relationship between the diurnal temperature cycle and soil moisture, which depend upon soil type and is largely limited to bare soil conditions. A main problem associated with thermal infra-red techniques is cloud interference. Microwave techniques for measuring soil moisture include both passive and active microwave approaches; each has distinct advantages. Microwave techniques are based on a large contrast between dielectric properties of liquid water and dry soil. The variation of natural terrestrial gamma radiation can be used to measure soil moisture because gamma radiation is strongly attenuated by water. It appears that operational remote-sensing of soil moisture will involve more than one sensor. Furthermore, both active microwave and thermal infra-red applications need much additional research before they can be used to extract soil moisture information.

The reflection from bare soil, in the visible and near-infra-red parts of the electromagnetic spectrum, can only be used under limited conditions to estimate soil moisture. The accuracy of this method is poor and absolute values of soil moisture cannot be obtained. More spectral bands and a much higher geometrical resolution in the (VIS/NIR) infra-red visible/near range are needed for soil moisture and agricultural purposes, than that available from

Landsat, SPOT and the NOAA satellites. Soil moisture has been estimated by using precipitation indices; operational applications have been developed by FAO using geostationary imagery over intertropical regions (WMO, 1993). With the advent of the International Geosphere–Biosphere Programme (IGBP) the need for high-resolution data is increasing.

Thermal infra-red techniques have been successfully used to measure the few surface centimetres of soil moisture. A limitation to the thermal approach is that it cannot effectively be applied to surfaces with vegetation cover.

Attempts have been made to evaluate the soil moisture through observation of the Apparent Thermal Inertia using both AVHRR data from Landsat and SPOT and geostationary images; applications have been more of pilot projects rather than operational (WMO, 1993).

Microwave techniques have shown a lot of potential for measuring soil moisture but still need varying amounts of research to make them operational. In order to progress to operational soil moisture monitoring by remote-sensing techniques, multifrequency and multipolarization satellite data will be required; such data are needed to quantify different surfaces and thus reduce the amount of ground truth required.

Only in the microwave region is there a direct physical relationship between soil moisture and the reflection or emission of radiation. A unique advantage of using the microwave region is that at long wavelengths the soil moisture measurements can be made through clouds. It has also been illustrated that the synergistic use of optical and microwave data in agrometeorological applications is advantageous. The passive microwave region has been exploited the most so far. At present, microwave radiometers capable of measuring soil moisture are available only on aircraft. These are being used in both research and a few operational applications.

Soil moisture information at a depth of several metres can be obtained from short pulse radar (wavelengths of 5–10 cm) techniques. In the Russian Federation, this aircraft-based method is used for soil moisture measurements in forested areas and for detecting zones of saturation down to a depth of 5–10 m. The use of gamma radiation is potentially the most accurate of the remote-sensing methods developed for soil moisture measurement. The attenuation of gamma radiation can be used to determine changes in soil moisture in the top 20–30 cm of the ground. This technique requires that some field measurements of soil moisture be made during the measurement flight, because it does not give the absolute values of soil moisture. (WMO, 1992*b*).

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