4 EVAPORATION AND TRANSPIRATION

Evaporation, the transfer of water from the basin surface to the atmosphere, is the main term facing rainfall input in the water balance equation. It is therefore an important factor in most hydrological studies, as the difference between rainfall and evaporation in its various forms largely determines the amount of water surplus in a river basin and thus the volume of river flow. It also affects the seasonal distribution of runoff; in many parts of the world the rainfall is highly seasonal, but in temperate areas where the rainfall varies less during the year, the temperature and thus the evaporation is seasonal. In terms of water resources studies, evaporation losses from storage reservoirs may be important, while irrigated crop water requirements are linked to evaporation.

However, evaporation from open water and bare soil, and the equivalent transpiration from grass or other short vegetation, and from woodland or forest, may be different from each other. Also, evaporation is not easy to measure directly, as the evaporation loss from a small pan cannot be accepted as equivalent to open water evaporation from a larger body of water. However, a number of indirect methods of estimating evaporation, which are based to a varying extent on the physical understanding of the process, are often used to estimate the different forms of evaporation.

MEASUREMENT OF DIFFERENT FORMS OF EVAPORATION

In physical terms evaporation is simply the conversion of water from a liquid form to water vapour; in this sense evaporation includes all the ways in which water at the surface of a river basin is converted into moisture in the air above the basin. However, it is generally referred to as evaporation from the surface of a water body or from bare soil, or as transpiration when it is routed through the leaves of grass or other vegetation such as woodland. A related process is evaporation from rainfall intercepted by vegetation and retained on the surface of the vegetation. There are variations in the rates of evaporation from these different surfaces, but all of them depend ultimately on an energy supply to provide the latent heat required; the evaporation rate also varies with the wind available to carry away the water vapour. These factors in turn depend in general terms on climate, so that evaporation is on the whole less variable than rainfall.

Evaporation from a large body of open water is perhaps the nearest to a standard example of the process, but is not easy to measure except as the residual from a water balance study of a lake. Measurements using evaporation pans, in which the loss of water over a known area is compared with measured rainfall, are not very satisfactory as the apparent evaporation loss varies with the dimension of the pan, and the instrument receives heat supply to the side of the pan. In arid climates the evaporating water surface is much moister than its surroundings, and thus evaporates at a higher rate than a larger body because of a so-called oasis effect. Various pans and tanks have been adopted as standard in different countries, with the United States "Class A" pan adopted as an international standard; however, empirical coefficients, which may vary with season, are required to adjust pan measurements to represent a large water body.

An early method of measuring evaporation was used in the Nile basin. Piche evaporimeters, in which a reservoir in a glass cylinder feeds a saturated disc of paper of standard dimension, were used throughout the Nile basin (Hurst, 1952) to measure the loss. This loss was then adjusted by a factor, which was taken as 0.50 in much of the basin, but which was found to vary towards 0.75 or 0.80 in more humid regions; this was clearly a rather artificial method of estimation.

Another standard definition of evaporation, known as potential transpiration, is the loss from a short grass surface which is amply supplied with moisture. Lysimeters have been designed to measure such an evaporation loss, either by continuous weighing or by adding sufficient water daily to induce drainage which is also measured. Such techniques are not easy to adapt to measuring evaporation losses from woodland, but experiments have been carried out which show these to be higher than from grass.

Although this brief discussion introduces topics relevant to later chapters, it illustrates some of the problems of direct measurement of evaporation.

ESTIMATION OF POTENTIAL TRANSPIRATION

Once potential transpiration from a moist grass surface had been proposed as a standard measure of evaporation, several approaches were developed to estimate this process from meteorological data on the assumption that the process was controlled by climate and that vegetation was simply responding to external conditions.

Thornthwaite

An empirical expression for potential transpiration was derived by Thornthwaite (1948) from records in the eastern United States of America. This formula is based on monthly mean temperatures and a heat index that is based on a complex sum of monthly temperatures throughout the year. An adjustment is made for the monthly hours of daylight corresponding to the latitude of the site. Because the relation between temperature and transpiration is derived from a specific region, this formula is not necessarily valid in other climates. A useful part of this approach is the Thornthwaite diagram (see Fig. 5.2), based on a comparison of monthly rainfall and potential transpiration; when potential transpiration exceeds rainfall, a period of deficit occurs and transpiration depends on soil moisture depletion, but when rainfall exceeds potential transpiration a period of surplus gives rise to soil moisture recharge and runoff.

Penman

The Penman (1948) approach to the estimation of open water evaporation, or alternatively of potential transpiration from a grass surface which is never short of water, was more closely based on the physical processes involved. He combined the energy balance equation, according to which net solar energy is used both in evaporation and in heating the air, and also the aerodynamic equations in which evaporation is a function of saturation deficit and wind speed, and the corresponding equation in which heat flux is a function of temperature difference between the water surface and air, and of wind speed. By assuming the equivalence of the two wind speed expressions, Penman was able to eliminate the terms involving water surface temperature.

Plate 4.1 *Meteorological station at Esquel, Patagonia. From left to right: raingauge, evaporation pan, solarimeter, anemometer, raingauge, radiometer, anemometer mast, and Stevenson screen with recording raingauge behind.*

The expression obtained is $E_O = (\Delta H + \gamma E_A)/(\Delta + \gamma)$, where E_O is open water evaporation, H is the net incoming energy, E_A is evaporation at air temperature, Δ is the gradient of saturated vapour pressure against temperature, and γ is the hygrometric constant. This expression can be solved from standard meteorological observations (Plate 4.1): air temperature and humidity, wind speed and either incoming energy or sunshine hours. The value of H may be measured directly or estimated from sunshine hours; E_A, which is the smaller term, can be estimated from an aerodynamic equation using air temperature, humidity and wind speed. Tables were compiled to assist these calculations, and nowadays computer programs are also available. Because this approach is based primarily on physical reasoning with an element of empirical observation, it can be transposed from one climate to another and it has been used in many countries as a standard method. The detailed derivation is available in many textbooks, and is not repeated here.

Penman originally introduced empirical factors, based on observation in Britain, to transform open water estimates (E_O) to transpiration from grass (E_T), but he later demonstrated that the adjustment could be made by measuring the net energy over the different surface or by altering the albedo in the estimate. A more direct approach is the Penman-Monteith equation (Monteith, 1985).

ESTIMATION OF ACTUAL EVAPORATION

An empirical formula was developed by Turc (1954), expressing actual annual basin evaporation loss as a function of annual precipitation and mean air temperature. This approach was based on comparisons of rainfall and runoff from a number of basins in many countries. Although it has been popular as a useful means of estimating potential runoff from an area, it has largely been superseded by the more direct methods based on the Penman approach.

The simplest method of estimating actual evaporation is through comparison of rainfall and potential transpiration, when the periods of soil moisture use and recharge can be deduced. Some examples of the balance between rainfall and potential transpiration are illustrated in Fig. 4.1 (a)–(c), where examples of humid, arid and mixed climates are shown. The differences between these three situations are of fundamental importance to the approach to the hydrological water balance. In fact this simple comparison epitomises the hydrological character of an area, and largely determines the complexity of the analysis. This distinction will be stressed throughout subsequent chapters, but the humid climate is the simplest to deal with, while the arid climate is the most complex.

Climate examples: Sri Lanka, Botswana, India

The humid example is from Nuwara Eliya in Sri Lanka (Fig. 4.1(a)), where average rainfall exceeds potential evaporation in almost all months; in this case actual evaporation will always be close to the potential. In the arid case of Francistown in northeast Botswana (Fig. 4.1(b)), where potential evaporation exceeds average rainfall in all months, actual evaporation will be controlled by actual rainfall rather than potential transpiration. An intermediate case is that of Bhopal, in central India (Fig. 4.1(c)), where a short period of rainfall surplus is followed by a longer period of deficit. In this case the evaporation may be assumed to equal the potential rate during the wet season; at the beginning of the dry season the evaporation will approximate to the potential while the supply is maintained by soil moisture storage. Once the soil moisture storage has been reduced by plant water use from saturation to the so-called wilting point, the evaporation will greatly reduce or cease. The difference between the water which can be held against gravity, and that which remains at the wilting point, over the soil depth from the surface to the bottom of the rooting system, is known as the root constant (Penman, 1950). This depth of water is often assumed to be transpired at the potential rate before the evaporation is reduced.

This rather simple model is illustrated by Fig. 5.2 in which monthly average rainfall and potential transpiration are compared for the Betwa basin near Bhopal. During July, the first month where rainfall exceeds potential transpiration, the surplus water recharges the soil moisture store. After the soil moisture has been recharged, the water surplus is available for runoff, which occurs between late July and the end of September. From October, the transpiration is maintained at the potential rate by soil moisture storage until the storage, which is naturally equal to the recharge in July, is used up. There follows a period of water deficit, when transpiration is much reduced. This model, which is essentially the Thornthwaite diagram, enables the actual evaporation to be estimated from the potential rate and the value of the root constant. This topic will be discussed in greater detail once the concepts of soil moisture measurement and estimation have been introduced in Chapter 5.

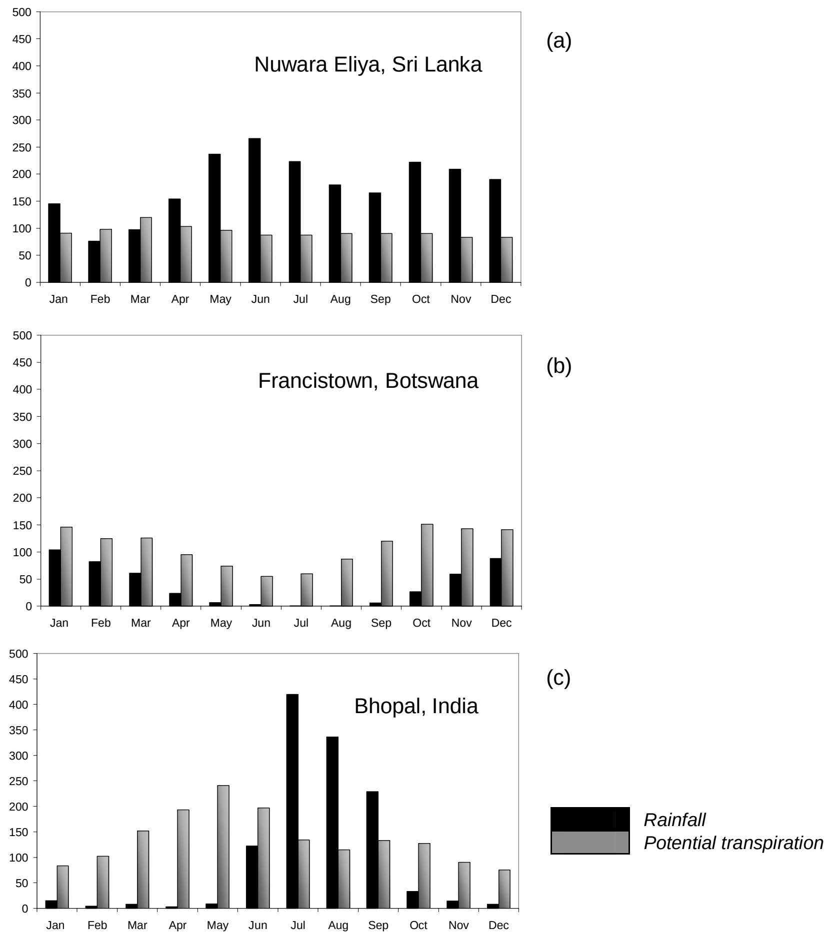

Fig. 4.1 *Comparison of monthly mean rainfall and potential transpiration (mm).*

MEASUREMENT OF OPEN WATER EVAPORATION

Botswana

In northeast Botswana there was an opportunity at Shashe reservoir near Francistown to measure open water evaporation fairly directly; the inflow was limited in most years to short periods during the summer months November–April, and spill from the reservoir was even more limited. The regime of the reservoir (Parks & Sutcliffe, 1987) is illustrated by Fig. 4.2, which shows the reservoir levels for 1984. The inflow to the reservoir was deduced from rises in reservoir level and from spill when the reservoir was full, to give a flow record which was not subject to the problems of measurement in these very ephemeral rivers. The evaporation from this open water body could be estimated from the water balance of the reservoir, taking into account inflow, spill, rainfall and abstractions, all of which were measured. Both inflow and evaporation could be estimated from a single water balance because inflow is concentrated in a small number of days.

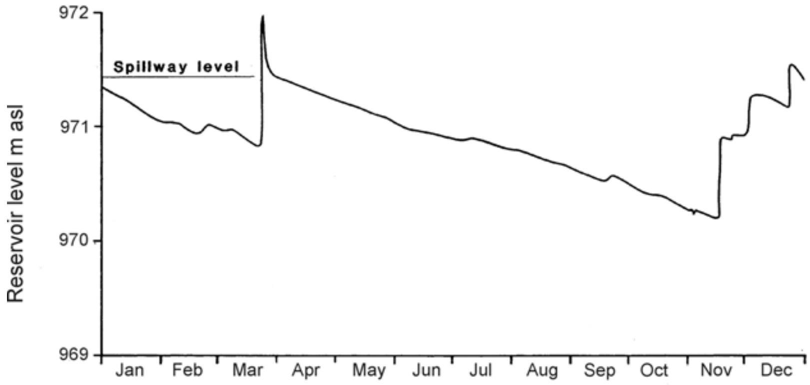

Fig. 4.2 *Daily levels of Shashe reservoir, northeast Botswana, 1984 (from Parks & Sutcliffe, 1987).*

After daily reservoir levels had been plotted to determine the periods of inflow, the periods of rise and the level changes were calculated. The inflow volumes were deduced from the level rises and the area–capacity curve. When levels exceeded the spillway level, they were converted to spill flows using an equation derived from model tests. The increases in reservoir volume and spills were tabulated separately as monthly totals. Daily rainfall totals were multiplied by current reservoir area to give rainfall volumes. The evaporation during periods of inflow and spill were estimated as a first iteration from daily rates estimated by the Penman method, and converted to volumes. The abstractions were derived from metered measurements, and the volumes during periods of inflow and spill were also deduced; because the abstractions were mainly for industrial use, they were fairly uniform.

The net evaporation was calculated from the initial and final daily reservoir levels, which were converted to area and volume. The net inflow, or total inflow less spill, was corrected for rainfall, evaporation and abstraction during the inflow period. The change of storage less corrected inflow then equals abstraction and net evaporation during the month. Seepage through the dam was known to be negligible. The net evaporation was converted to depth by dividing the volume by mean area, and the rainfall was added to give gross evaporation depth. These were tabulated and found to give an annual average of 1820 mm, which was similar to the Penman estimate. Over the 12-year period analysed, only 20 months had more than 10 days of inflow and spill, when the estimates might have been affected by measurement uncertainties. At the same time the total inflow volume was estimated for a 15-year period, with an annual average of 91 $m^3 \times 10^6$.

The results of this water balance analysis for 1984 are illustrated in Fig. 4.3 where monthly inflow volumes and evaporation depths are shown. They demonstrate the importance of reservoir evaporation in this semiarid example. Whereas net evaporation from a reservoir can almost be neglected in countries like Britain, where most reservoirs are built in upland areas where rainfall exceeds evaporation, the evaporation from reservoirs in Botswana is a most important component in the analysis and in fact is several times the net yield of the reservoir itself.

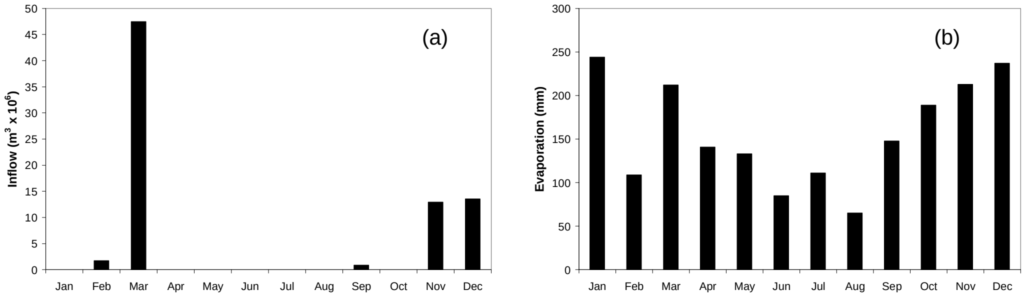

Fig. 4.3 *Water balance components for Shashe reservoir, 1984: (a) inflow ($m^3 \times 10^6$); (b) evaporation (mm).*

This example illustrates the fact that direct measurements of open water evaporation are complex, even when conditions are extremely favourable, as in this case of a reservoir with measured abstraction in an arid climate where periods of inflow are brief. Similar calculations can be made in other climates, but the result may depend on measurement of the differences between large numbers; the various components are each measured with given errors, but the differences between inflow and outflow components are estimated with greatly enhanced error where the components are of similar magnitude.

ESTIMATION OF WATER USE BY VEGETATION

The estimation of crop water use depends on the conversion of potential transpiration from grass to the potential transpiration from crops in various stages of growth and maturity, for which tables of crop coefficients (Doorenbos & Pruitt, 1977) are available. In general crops transpire at a rate below the potential when they are growing and not yet covering the ground surface, then at the potential rate when they are fully grown, and finally at a reduced rate when they reach maturity and are about to be harvested. The net irrigation requirement can be deduced from the gross requirement by allowing for rainfall during the periods of growth and maturity. Perennial crops, on the other hand, are likely to transpire at near the potential rate if water is available and thus their water needs will depend largely on climate.

One aspect of evaporation from vegetation is the effect of interception of rainfall by the leaves and stems of vegetation, especially woodland. Much research on this topic (e.g. Gash, 1979) has shown that the depth or fraction of rainfall intercepted depends not only on the type of vegetation but also on the time distribution of rainfall. For instance, the annual interception by broad-leaved woodland with rainfall distributed throughout the year, will exceed that from acacia-type vegetation in rainfall limited to intense storms in a short rainfall season. Much of the intercepted water will be evaporated from the leaf surface, and can be regarded as part of the water use of vegetation; the energy use may be increased by lateral flow or advection.

Water use in Bahrain

During a study of the water resources of Bahrain in 1967, the water balance of the island was found to depend on the balance between the groundwater inflow, the sole source of supply (which was derived from recharge in Saudi Arabia and flowed

Plate 4.2 *Date palm plantation in Bahrain.*

through the deep limestone aquifer under the sea), and groundwater abstraction for domestic and industrial use and for crop water use. It was necessary to make an estimate of water abstraction on the island which could be compared with the estimated inflow.

By 1967, the number of drilled wells had reached almost 800, and the measurement of even a sample number would have been a large task. The groundwater abstraction for urban and industrial water supply was measured, but the most important component was the abstraction from springs and wells for irrigation of date palm plantations. Because these irrigated areas were clearly shown on maps and air photography, the simplest way to estimate the abstraction was through the transpiration from these areas. It was possible to confirm the areas and estimate the densities of these plantations by field reconnaissance coupled with study of air photography. The water use was estimated from the product of the effective area and the transpiration rate. From the density of the date palms (Plate 4.2) the transpiration rate was assumed to correspond with the open water evaporation rate, which was estimated by the Penman method from meteorological records as 2580 mm per year; the annual rainfall of below 75 mm could be neglected in this case. Allowance was made for irrigation distribution losses, and leaching requirements to ensure drainage, at 25–50% of evaporation. This indirect estimate provided the major item in the water balance; with municipal and industrial use the annual abstraction totalled 122–147 $m^3 \times 10^6$. The use of this estimate is described in Chapter 6.

Abu Dhabi

In another case, in a rapid study of the water resources of Abu Dhabi, a major source of recharge to the main aquifer was runoff from the mountains of the Jebel Akhdar from which a number of streams emerged on to alluvial fans near the border with Oman. No measurements of these streams had been made, but it was possible after visiting the region to estimate their flows (see also Chapter 6) using the areas of perennial date palms and other crops irrigated near villages along the streams. Some of the agricultural areas relied on base flows diverted into channels by temporary dams on the main streams. Others relied on flood flows which recharged through fans into the alluvial aquifers; this subsurface flow was diverted from relatively narrow river valleys downstream, where flows were concentrated, by means of *aflaj* or underground tunnels similar to the *qanats* of Iran. The combined flows could be deduced from evaporation from the areas of perennial irrigation near these two types of villages, which were termed surface water and groundwater villages.

MEASUREMENT OF WETLAND EVAPORATION

Large wetlands, with a mixture of open water and swamp vegetation either floating or anchored in saturated soil, provide a surface with some similarity to open water; there is no complication caused by the transpiration decreasing as a result of soil moisture deficit. However, it is not easy to measure evaporation and transpiration directly.

The Sudd, the area of swamps of the Bahr el Jebel in the southern Sudan, where about half the inflows are lost to the White Nile outflow due to evaporation, provides an example of a large wetland in an area of moderate rainfall; the existence of the wetland depends on the inflow fed by Lake Victoria and the other East African lakes.

Direct measurement

At a time when the estimation of evaporation from meteorological data had not been developed, an attempt was made to measure evaporation (Hurst & Phillips, 1938) near the centre of the swamps by establishing a tank 10 m square and 2.5 m deep. The tank was sunk in the swamp and planted with papyrus; the water in the tank was kept at the same level as the swamp, and the loss of water was measured regularly and adjusted for rainfall. It was noted that losses were higher on rainy days, which was attributed to interception. The observations from 1926 to 1932 gave an annual mean of 1318 mm, but the annual totals decreased over this period. It was noted that the papyrus within the tank was less luxuriant than that outside, and it was found difficult to maintain typical growth.

Moreover, the evaporation losses deduced from the average area of flooding, compared with inflows, rainfall and outflows, were insufficient to explain the losses in the Sudd. Hurst & Phillips (1938) suggested that evaporation must have been underestimated by about 30% in order to explain the balance. In later studies, Migahid (1948, 1952) carried out a botanical investigation of the area and improved the evaporation tanks, with growth in the tank as good as that outside. Measurements in 1947–1948 gave annual results approaching 2400 mm, which were similar to losses from a tank full of water in an open lagoon. The interception by vegetation was measured as 45% of rainfall. Penman (1963) commented that the diurnal weather cycle, with relative humidity decreasing to a minimum during the day as the wind speed reaches a maximum, would lead to extreme evaporation and to transpiration from papyrus similar to evaporation from an open lagoon.

Indirect estimation

These experiments confirm that the direct measurement of wetland losses is difficult. In many ways it is easier to establish evaporation rates as a residual of the water balance. In an investigation of the Sudd to deduce the potential effect of the Jonglei Canal on areas of flooding, Sutcliffe & Parks (1982, 1987) developed a simple hydrological model. Using monthly river inflows and outflows for the period 1905–1980, rainfall measurements (average 870 mm) and swamp evaporation taken as open water evaporation (2150 mm) estimated by the Penman method, they estimated monthly areas of flooding. They took account of infiltration into newly flooded areas, and used a linear relation between areas and volumes of flooding to convert cumulative differences between inflow and outflow volumes to flooded areas. The series of flooded areas covering the period of flow records were compared with a number of estimates of these areas, taken from air photography and satellite imagery or from vegetation maps compiled at different dates. These comparisons (see also Chapter 8) showed that the modelling was realistic and therefore that the Penman open water evaporation was a reasonable estimate of evaporation from the wetland.

ESTIMATION OF NET RAINFALL

In the case of open water evaporation from a reservoir or evaporation from a wetland, the net rainfall can be deduced simply from the difference between rainfall and evaporation; in fact annual rainfall was less than evaporation in the case of the Sudd, so the net rainfall was negative. When the actual evaporation reduces below the potential transpiration rate following the development of a moisture deficit, the derivation of actual evaporation from the potential rate must take account of the soil moisture state. Further discussion of actual rather than potential evaporation follows the discussion of soil moisture storage in the next chapter.

CONCLUSION

Evaporation takes a variety of routes, directly from open water or soil, or from intercepted water, or alternatively as transpiration through vegetation, but in general the process is controlled by climate and may be estimated from meteorological records. The process is reasonably conservative in time and space, and fewer measurements or estimates are required than for rainfall. It is possible, but not usually straightforward, to measure evaporation directly in some cases, but it is often deduced as the residual of the water balance.