3

EVAPORATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the process of evaporation and what controls its rate.
- A knowledge of the techniques for measuring evaporation directly.
- A knowledge of the techniques used to estimate evaporation.

Evaporation is the transferral of liquid water into a gaseous state and its diffusion into the atmosphere. In order for this to occur there must be liquid water present and available energy from the sun or atmosphere. The importance of evaporation within the hydrological cycle depends very much on the amount of water present and the available energy, two factors determined by a region's climate. During winter months in humid-temperate climates evaporation may be a minor component of the hydrological cycle as there is very little available energy to drive the evaporative process. This alters during summer when there is abundant available energy and evaporation has the potential to become a major part of the water balance. The potential may be limited by the availability of liquid water during the dry months. This can be seen in extremely hot, arid climates where there is often plenty of available energy to drive evaporation but very little water to be evaporated. As a consequence the actual amount of evaporation is small.

It is the presence or lack of water at the surface that provides the major semantic distinction in definitions of the evaporative process. **Open water evaporation** (often denoted as E_o) is the evaporation that occurs above a body of water such as a lake, stream or the oceans. Figure 1.6 shows that at the global scale this is the largest source of evaporation, in particular from the oceans. **Potential evaporation** (*PE*) is that which occurs over the land's surface, or would occur if the water supply were unrestricted. This occurs when a soil is wet and what evaporation is able to happen occurs without a lack of water supply. **Actual evaporation** (E_t) is that which actually occurs (i.e. if there is not much available water it will be less than potential). When conditions are very wet (e.g. during a rainfall event) E_t will equal PE, otherwise it will be less than PE. In hydrology we are most interested in E_o and E_t but normally require PE to calculate the E_t value.

All of these definitions have been concerned with 'evaporation over a surface'. In hydrology the surface is either water (river, lake, ponds, etc.) or the land. The evaporation above a land surface occurs in two ways – either as actual evaporation from the soil matrix or **transpiration** from plants. The combination of these two is often referred to as **evapotranspiration**, although the term *actual evaporation* is essentially the same (hence the *t* subscript in E_t). Transpiration from a plant occurs as part of photosynthesis and respiration. The rate of transpiration is controlled by the opening or closing of stomata in the leaf. Transpiration can be ascertained at the individual plant level by instruments measuring the flow of water up the trunk or stem of a plant. Different species of plants transpire at different rates but the fundamental controls are the available water in the soil, the plant's ability to transfer water from the soil to its leaves and the ability of the atmosphere to absorb the transpired water.

Evaporation is sometimes erroneously described as the only loss within the water balance equation. The water balance equation is a mathematical description of the hydrological cycle and by definition there are no losses and gains within this cycle. What is meant by 'loss' is that evaporation is lost from the earth's surface, where hydrologists are mostly concerned with the water being. To a meteorologist, concerned with the atmosphere, evaporation can be seen as a gain. Evaporation although not a loss, can be viewed as the opposite of precipitation, particularly in the case of dewfall, a form of precipitation. In this case the **dewfall** (or negative evaporation) is a gain to the earth's surface.

EVAPORATION AS A PROCESS

It has already been said that evaporation requires an energy source and an available water supply to transform liquid water into water vapour. There is one more precondition: that the atmosphere be dry enough to receive any water vapour produced. These are the three fundamental parts to an understanding of the evaporation process. This was first understood by Dalton (1766–1844), an English physicist who linked wind speed and the dryness of the air to the evaporation rate.

Available energy

The main source of energy for evaporation is from the sun. This is not necessarily in the form of direct radiation, it is often absorbed by a surface and then re-radiated at a different wavelength. The normal term used to describe the amount of energy received at a surface is **net radiation** (Q^*), measured using a net radiometer. Net radiation is a sum of all the different heat fluxes found at a surface and can be described by equation 3.1.

$$Q^* = Q_S \pm Q_L \pm Q_G \tag{3.1}$$

where Q_S is the sensible heat flux; Q_L is the latent heat flux and Q_G is the soil heat flux.

Sensible heat is that which can be sensed by instruments. This is most easily understood as the heat we feel as warmth. The sensible heat flux is the rate of flow of that sensible heat.

Latent heat is the heat either absorbed or released during a phase change from ice to liquid water, or liquid water to water vapour. When water moves from liquid to gas this is a negative flux (i.e. energy is absorbed) whereas the opposite phase change (gas to liquid) produces a positive heat flux.

The **soil heat flux** is heat released from the soil having been previously stored within the soil. This is frequently ignored as it tends to zero over a 24-hour period and is a relatively minor contributor to net radiation.

Incoming solar radiation is filtered by the atmosphere so that not all the wavelengths of the electromagnetic spectrum are received at the earth's surface. Incoming radiation that reaches the surface is often referred to as short-wave radiation: visible light plus some bands of the infrared. This is not strictly true as clouds and water vapour in the atmosphere, plus trees and tall buildings above the surface, emit longer-wave radiation which also reaches the surface.

Outgoing radiation can be either reflected short-wave radiation or energy radiated back by the earth's surface. In the latter case this is normally in the infrared band and longer wavelengths and is referred to as long-wave radiation. This is a major source of energy for evaporation.

There are two other forms of available energy that under certain circumstances may be important sources in the evaporation process. The first is heat stored in buildings from an anthropogenic source (e.g. domestic heating). This energy source is often fuelled from organic sources and may be a significant addition to the heat budget in an urban environment, particularly in the winter months. The second additional source is **advective energy**. This is energy that originates from elsewhere (another region that may be hundreds or thousands of kilometres away) and has been transported to the evaporative surface (frequently in the form of latent heat) where it becomes available energy in the form of sensible heat. The best example of this is latent energy that arrives in cyclonic storm systems. In Chapter 1 it was explained that evaporating and condensing water is a major means of redistributing energy around the globe. The evaporation of water that contributes to cyclonic storms normally takes place over an ocean, whereas the condensation may occur a considerable distance away. At the time of evaporation, thermal energy (i.e. sensible heat) is transferred into latent energy that is then carried by the water vapour to the place of condensation where it is released as sensible heat once more. This 're-release' is often referred to as advective energy and may be a large energy source to drive further evaporation.

Water supply

Available water supply can be from water directly on the surface in a lake, river or pond. In this case it is open water evaporation (E_o). When the water is lying within soil the water supply becomes more complex. Soil water may evaporate directly, although it is normally only from the near surface. As the water is removed from the surface it sets up a soil moisture gradient that will draw water from deeper in the soil towards the surface, but it must overcome the force of gravity and the withholding force exerted by soil capillaries (see Chapter 4). In addition to this the water may be brought to the surface by plants using osmosis in their rooting system. The way that soil moisture controls the transformation from potential evaporation to actual evaporation is complex and will be discussed further later in this chapter.

The receiving atmosphere

Once the available water has been transformed into water vapour, using whatever energy source is available, it then must be absorbed into the atmosphere surrounding the surface. This process of *diffusion* requires that the atmosphere is not already saturated with water vapour and that there is enough buoyancy to move the water vapour away from the surface. These two elements can be assessed in terms of the **vapour pressure deficit** and atmospheric mixing.

Boyle's law tells us that the total amount of water vapour that may be held by a parcel of air is temperature and pressure dependent. The corollary of this is that for a certain temperature and air pressure it is possible to specify the maximum amount of water vapour that may be held by the parcel of air. We use this relationship to describe the **relative humidity** of the atmosphere (i.e. how close to fully saturated the atmosphere is). Another method of looking at the amount of water vapour in a parcel of air is to describe the *vapour pressure* and hence the *saturation vapour pressure*. The difference between the actual vapour pressure and the saturation vapour pressure is the *vapour pressure deficit* (vpd). The vpd is a measure of how much extra water vapour the atmosphere could hold assuming a constant temperature and pressure. The higher the vpd the more water can be absorbed from an evaporative surface.

Atmospheric mixing is a general term meaning how well a parcel of air is able to diffuse into the atmosphere surrounding it. The best indicator of atmospheric mixing is the wind speed at different heights above an evaporating surface. If the wind speed is zero the parcel of air will not move away from the evaporative surface and will 'fill' with water vapour. As the wind speed increases, the parcel of air will be moved quickly on to be replaced by another, possibly drier, parcel ready to absorb more water vapour. If the evaporative surface is large (e.g. a lake) it is important that the parcel of air moves up into the atmosphere, rather than directly along at the same level, so that there is drier air replacing it. This occurs through turbulent diffusion of the air. There is a greater turbulence associated with air passing over a rough surface than a smooth one, something that will be returned to in the discussion of evaporation estimation.

One way of thinking about evaporation is in terms of a washing line. The best conditions for drying your washing outside are on a warm, dry and windy day. Under these circumstances the evaporation from your washing (the available water) is high due to the available energy being high (it is a warm day), and the receiving atmosphere mixes well (it is windy) and is able to absorb much water vapour (the air is dry). On a warm and still day, or a warm and humid day washing does not dry as well (i.e. the evaporation rate is low). Understanding evaporation in these terms allows us to think about what the evaporation rate might be for particular atmospheric conditions.

Evaporation above a vegetation canopy

Where there is a vegetation canopy the evaporation above this surface will be a mixture of transpiration, evaporation from the soil and evaporation from wet leaves *(canopy interception or interception loss or wet leaf evaporation)*. The relative importance of these three evaporation sources will depend on the degree of vegetation cover and the climate at the site. In tropical rain forests transpiration is the dominant water loss but where there is a seasonal soil water deficit the influence of canopy interception loss becomes more important. This is illustrated by the data in Table 3.1 which contrasts the water balance for two Pinus radiata forests at different locations in New Zealand (with different climates).

Transpiration by a plant leads to evaporation from leaves through small holes (stomata) in the leaf. This is sometimes referred to as dry leaf evaporation. The influence of stomata on the transpiration rate is an interesting plant physiological phenomenon. Some

Table 3.1 Estimated evaporation losses from two *Pinus radiata* sites in New Zealand

	Puruki (Central North Island, NZ) (% annual rainfall in brackets)		*Balmoral (Central South Island, NZ) (% annual rainfall in brackets)*	
Annual rainfall	1,405 mm		870 mm	
Annual interception loss	370 mm	(26)	220 mm	(25)
Annual transpiration	705 mm	(50)	255 mm	(29)
Annual soil evaporation	95 mm	(7)	210 mm	(24)
Remainder (runoff + percolation)	235 mm	(17)	185 mm	(21)

Source: Data adapted from Kelliher and Jackson (2001) plants are very effective at shutting stomata when under water stress, and therefore limit their water usage. The water stress occurs when the vapour pressure deficit is high and there is a high evaporative demand. In this situation the stomata within a leaf can be likened to a straw. When you suck hard on a soft straw it creates a pressure differential between the inside and outside and the sides collapse in; therefore you cannot draw air easily through the straw. Stomata can act in a similar manner so that when the evaporative demand is high (sucking water vapour through the stomata) the stomata close down and the transpiration rate decreases. Some plant species shut their stomata when under evaporative stress (e.g. conifers) while others continue transpiring at high rates when the evaporative demand is high (e.g. many pasture species). The ability of plants to shut their stomata can influence the overall water budget as their overall evaporation is low. This is illustrated in the case study later in this chapter on using a lysimeter to measure tussock evaporation.

It is the role of interception loss (wet leaf evaporation) that makes afforested areas greater users of water than pasture land (see Case Study on p. 42). This is because the transpiration rates are similar between pasture and forest but the interception loss is far greater from a forested area. There are two influences on the amount of interception loss from a particular site: canopy structure and meteorology.

Canopy structural factors include the storage capacity, the drainage characteristics of the canopy and the aerodynamic roughness of the canopy. The morphology of leaf and bark on a tree are important factors in controlling how quickly water drains towards the soil. If leaves are pointed upwards then there tends to be a rapid drainage of water towards the stem. Sometimes this appears as an evolutionary strategy by a plant in order to harvest as much water as possible (e.g. rhubarb and gunnera plants). Large broadleaved plants, such as oak (*Quercus*), tend to hold water well on their leaves while needled plants can hold less per leaf (although they normally have more leaves). Seasonal changes make a large difference within deciduous forests, with far greater interception losses when the trees have leaves than without. Table 3.2 illustrates the influence of plant morphology through the variation in interception found in different forest types and ages. The largest influence that a canopy has in the evaporation process is through the aerodynamic roughness of the top of the canopy. This means that as air passes over the canopy it creates a turbulent flow that is very effective at moving evaporated water away from the surface. The reason that forests have such high interception losses is because they have a lot of intercepting surfaces *and* they have a high aerodynamic roughness leading to high rates of diffusion of the evaporated water away from the leaf (Figure 3.1).

Meteorological factors affecting the amount of interception loss are the rainfall characteristics. The rate at which rainfall occurs (intensity) and storm duration are critical in controlling the interception loss. The longer water stays on the canopy the greater the amount of interception loss. Also important will be the frequency of rainfall. Does

Table 3.2 Interception measurements in differing forest types and ages

Tree type	*Age*	*Interception (mm)*	*% of annual precipitation*
Deciduous hardwoods	100	254	12
Pinus strobus (White Pine)	10	305	15
Pinus strobus	35	381	19
Pinus strobus	60	533.4	26

Source: From Hewlett & Nutter (1969)

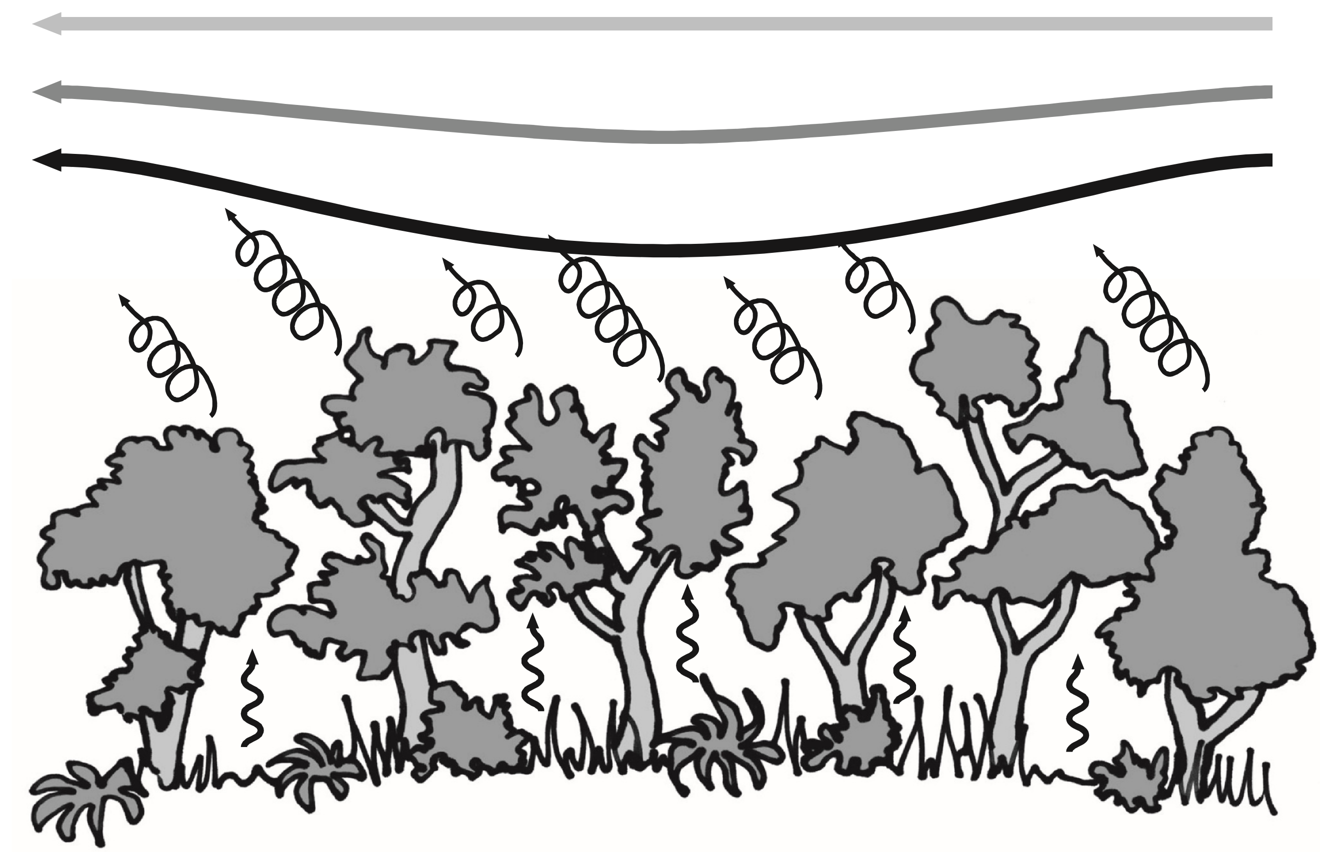

Figure 3.1 Factors influencing the high rates of interception loss from a forest canopy. The capacity of the leaves to intercept rainfall and the efficient mixing of water vapour with the drier air above leads to high evaporative losses (interception loss).

the canopy have time to dry out between rain events? If so, then the interception amount is likely to be higher. This is demonstrated in Figure 3.2 where the percentage of interception loss (interception ratio – broken line) is higher for small daily rainfall totals and the actual interception amount (solid line) reaches a maximum value of around 7mm even in the largest of daily rainfalls.

The amount of interception loss from an area is climate dependent. Calder (1990) used an amalgamation of different UK forest interception studies to show that there is a higher interception ratio (the interception loss divided by above-canopy rainfall) in drier than in wetter climates. The interception ratio ranges from 0.45 at 500 mm annual rainfall, to 0.27 at 2,700 mm annual rainfall. It is important to note that these interception ratio figures have considerable inter-annual variability.

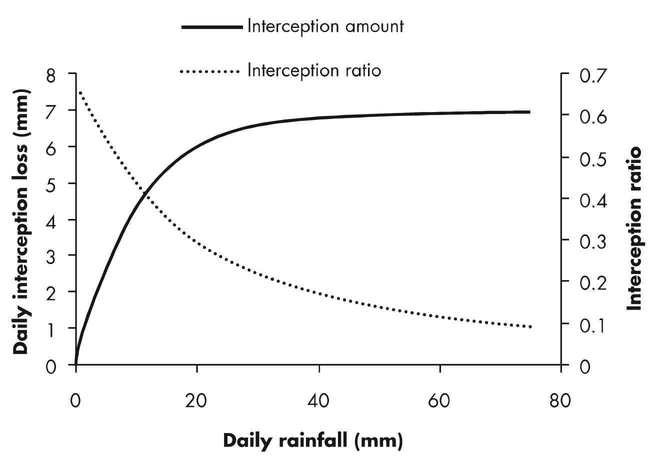

Figure 3.2 Empirical model of daily interception loss and the interception ratio for increasing daily rainfall. An interception ratio of 1.0 means all rainfall becomes interception loss.

Redrawn from Calder (1999)

Case study

FORESTS AND RAINFALL VS EVAPORATION

If you stand watching a forest during a warm summer shower it is common enough to see what appear to be clouds forming above the trees (see Plate 4). For many years it was believed that somehow trees attract rainfall and that cloud-forming was evidence of this phenomenon. As described by Pereira (1989) 'The worldwide evidence that hills and mountains usually have more rainfall and more natural forests than do adjacent lowlands has historically led to confusion of cause and effect'. This idea was taken further so that it became common practice to have forestry as a major land use in catchments that were being used to collect water for potable supply. In fact the cloud formation that is visible above a forest is a result of evaporation occurring from water sitting on the vegetation (intercepted rainfall). This 'wet leaf evaporation' can be perceived as a loss to the hydrologist as it does not reach the soil surface and contribute to possible streamflow. Throughout the latter half of the twentieth century there was considerable debate on how important wet leaf evaporation is.

One of the first pieces of field research to promote the idea of canopy interception being important was undertaken at Stocks Reservoir, Lancashire, UK. Law (1956) studied the water balance of an area covered with conifers (Sitka spruce) and compared this to a similar area covered with grassland. The water balance was evaluated for areas isolated by impermeable barriers with evaporation left as the residual (i.e. rainfall and runoff were measured and soil moisture assumed constant by looking at yearly values). Law found that the evaporation from the forested area was far greater than that for the pasture and he speculated that this was caused by wet leaf evaporation – in particular that the wet leaf evaporation was far greater from the forested area as there was a greater storage capacity for the intercepted water. Furthermore, Law went on to calculate the amount of water 'lost' to reservoirs through wet leaf vegetation and suggested a compensation payment from the forestry owners to water suppliers.

Conventional hydrological theory at the time suggested that wet leaf evaporation was not an important part of the hydrological cycle because it compensated for the reduction in transpiration that occurred at the same time (e.g. Leyton and Carlisle, 1959; Penman, 1963). In essence it was believed that the evapotranspiration rate stayed constant whether the canopy was wet or dry.

Following the work of Law, considerable research effort was directed towards discovering whether the wet leaf/dry leaf explanation was responsible for discrepancies in the water balance between grassland and forest catchments. Rutter (1967) and Stewart (1977) found that wet leaf evaporation in forests may be up to three or four times that from dry leaf. In contrast to this, other work has shown that on grassland, wet leaf evaporation is approximately equal to dry leaf (McMillan and Burgy, 1960; McIlroy and Angus, 1964). In addition, transpiration rates for pasture have been found to be similar to that of forested areas. When all this evidence is added up it confirms Law's work that forested areas 'lose' more rainfall through evaporation of intercepted water than grassland areas.

However there is still a question over whether the increased wet leaf evaporation may lead to a higher regional rainfall; a form of water recycling. Bands *et al*. (1987) write that: 'Forests are associated with high rainfall, cool slopes or moist areas. There is some evidence that, on a continental scale, forests may form part of a hydrological feedback loop with evaporation contributing to further rainfall'. Most researchers conclude that in general there is little, if any, evidence that forests can increase rainfall. However Calder (1999: 24, 26) concludes, 'Although the effects of forests on rainfall are likely to be relatively small, they cannot be totally dismissed from a water resources perspective . . . Further research is required to determine the magnitude of the effect, particularly at the regional scale.'

MEASUREMENT OF EVAPORATION

In the previous chapter there has been much emphasis on the difficulties of measuring precipitation due to its inherent variability. All these difficulties also apply to the measurement of evaporation, but they pale into insignificance when you consider that now we are dealing with measuring the rate at which a gas (water vapour) moves away from a surface. Concentrations of gases in the atmosphere are difficult to measure, and certainly there is no gauge that we can use to measure total amounts in the same way that we can for precipitation.

In each of the process chapters in this book there is an attempt to distinguish between measurement and estimation techniques. In the case of evaporation this distinction becomes extremely blurred. In reality almost all the techniques used to find an evaporation rate are estimates, but some are closer to true measurement than others. In this section each technique will include a sub-section on how close to 'true measurement' it is.

Direct micro-meteorological measurement

There are three main methods used to measure evaporation directly: the eddy fluctuation (or correlation), aerodynamic profile, and **Bowen ratio** methods. These are all micro-meteorological measurement techniques and details on them can be found elsewhere (e.g. Oke, 1987). An important point to remember about them all is that they are attempting to measure how much water is being evaporated above a surface, a very difficult task.

The eddy fluctuation method measures the water vapour above a surface in conjunction with a vertical wind speed and temperature profiles. These have to be measured at extremely short timescales (e.g. microseconds) to account for eddies in vertical wind motion. Consequently, extremely detailed micrometeorological instrumentation is required with all instruments having a rapid response time. In recent years this has become possible with hot wire **anemometers** and extremely fine thermistor heads for thermometers. One difficulty is that you are necessarily measuring over a very small surface area and it may be difficult to scale up to something of interest to catchment-scale hydrology.

The aerodynamic profile (or turbulent transfer) method is based on a detailed knowledge of the energy balance over a surface. The fundamental idea is that by calculating the amount of energy available for evaporation the actual evaporation rate can be determined. The measurements required are changes in temperature and humidity giving vertical humidity gradients. To use this method it must be assumed that the atmosphere is neutral and stable, two conditions that are not always applicable.

The Bowen ratio method is similar to the aerodynamic profile method but does not assume as much about the atmospheric conditions. The Bowen ratio is the ratio of sensible heat to latent heat and requires detailed measurement of net radiation, soil heat flux, temperature and humidity gradient above a surface. These measurements need to be averaged over a 30-minute period to allow the inherent assumptions to apply.

All of these micro-meteorological approaches to measuring evaporation use sophisticated instruments that are difficult to leave in the open for long periods of time. In addition to this they are restricted in their spatial scope (i.e. they only measure over a small area). With these difficulties it is not surprising that they tend to be used at the very small scale, mostly to calibrate estimation techniques (see pp. 46–52). They are accurate in the assessment of an evaporation rate, hence their use as a standard for the calibration of estimation techniques. The real problem for hydrology is that it is not a robust method that can be relied on for long periods of time.

Indirect measurement (water balance techniques)

Evaporation pans

The most common method for the measurement of evaporation is using an **evaporation pan** (see Figure 3.3). This is a large pan of water with a water depth measuring instrument or weighing device underneath that allows you to record how much water is lost through evaporation over a time period. This technique is actually a manipulation of the water balance equation, hence the terminology used here of a water balance technique. An evaporation pan is constructed from impervious material and the water level is maintained below the top so that no seepage or leakage occurs. This eliminates runoff (Q term) from the water balance. Therefore it can be assumed that any change in storage is related to either evaporative loss or precipitation gain. This means that the water balance equation can be rearranged as shown in equation 3.2.

$$E = \Delta S - P \tag{3.2}$$

If there is a precipitation gauge immediately adjacent to the evaporation pan then the P term can be accounted for, leaving only the change in storage (ΔS) to be measured as either a weight loss or a drop in water depth. At a standard meteorological station the evaporation is measured daily as the change in water depth. For a finer temporal resolution (e.g. hourly) there are load cell instruments available which measure and record the weight at regular intervals.

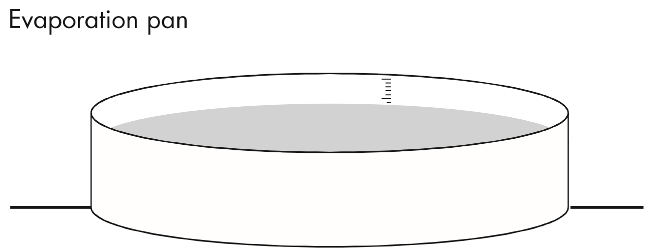

Figure 3.3 An evaporation pan. This sits above the surface (to lessen rain splash) and has either an instrument to record water depth or a continuous weighing device, to measure changes in volume.

An evaporation pan is filled with water, hence you are measuring E_o, the open water evaporation. Although this is useful, there are severe problems with using this value as an indicator of actual evaporation (E_t) in a catchment. The first problem is that E_o will normally be considerably higher than E_t because the majority of evaporation in a catchment will be occurring over a land surface where the available water is contained within soil and may be limited. This will lead to a large overestimation of the actual evaporation. This factor is well known and consequently evaporation pans are rarely used in catchment water balance studies, although they are useful for estimating water losses from lakes and reservoirs.

There are also problems with evaporation pans that make them problematic even for open water evaporation estimates. A standard evaporation pan, called a Class A evaporation pan, is 1,207 mm in diameter and 254 mm deep. The size of the pan makes them prone to the 'edge effect'. As warm air blows across a body of water it absorbs any water vapour evaporated from the surface. Numerous studies have shown that the evaporation rate is far higher near the edge of the water than towards the centre where the air is able to absorb less water vapour (this also applies to land surfaces). The small size of an evaporation pan means that the whole pan is effectively an 'edge' and will have a higher evaporation rate than a much larger body of water. A second, smaller, problem with evaporation pans is that the sides, and the water inside, will absorb radiation and warm up quicker than in a much larger lake, providing an extra energy source and greater evaporation rate.

To overcome the edge effect, empirical (i.e. derived from measurement) coefficients can be used which link the evaporation pan estimates to larger water body estimates. Doorenbos and Pruitt (1975) give estimates for these coefficients that require extra information on upwind fetch distance, wind run and relative humidity at the pan (Goudie *et al.*, 1994). Grismer *et al.* (2002) provide empirical relationships linking pan evaporation measurements to potential evapotranspiration, i.e. from a vegetated surface not open water evaporation.

Lysimeters

A **lysimeter** takes the same approach to measurement as the evaporation pan, the fundamental difference being that a lysimeter is filled with soil and vegetation as opposed to water (see Figure 3.4). This difference is important, as E_t rather than E_o is being indirectly measured. A lysimeter can also be made to blend in with the surrounding land cover, lessening the edge effect described for an evaporation pan.

There are many versions of lysimeters in use, but all use some variation of the water balance equation to estimate what the evaporation loss has been. One major difference from an evaporation pan is that a lysimeter allows percolation through the bottom, although the amount is measured. Percolation is necessary so that the lysimeter mimics as closely as possible the soil surrounding it; without any it would fill up with water. In the same manner as an evaporation pan it is necessary to measure the precipitation input immediately adjacent to the lysimeter. Assuming that the only runoff (Q) is through percolation, the water balance equation for a lysimeter i shown in equation 3.3.

$$E = \Delta S - P - Q \tag{3.3}$$

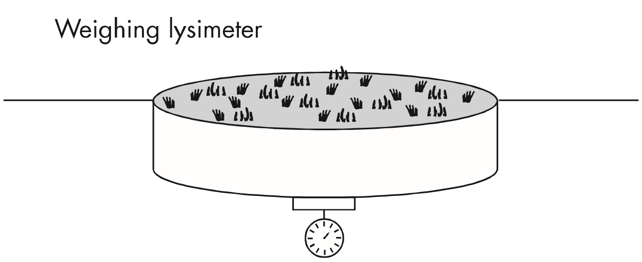

Figure 3.4 A weighing lysimeter sitting flush with the surface. The cylinder is filled with soil and vegetation similar to the surroundings.

A lysimeter faces similar problems to a rain gauge in that it is attempting to measure the evaporation that would be lost from a surface if the lysimeter were not there. The difference from a rain gauge is that what is contained in the lysimeter should closely match the surrounding plants and soil. Although it is never possible to recreate the soil and plants within a lysimeter perfectly, a close approximation can be made and this represents the best efforts possible to measure evaporation. Although lysimeters potentially suffer from the same edge effect as evaporation pans, the ability to match the surrounding vegetation means there is much less of an edge effect.

A *weighing lysimeter* has a weighing device underneath that allows any change in storage to be monitored. This can be an extremely sophisticated device (e.g. Campbell and Murray, 1990; Yang *et al.*, 2000), where percolation is measured continuously using the same mechanism for a tipping-bucket rain gauge, weight changes are recorded continuously using a hydraulic pressure gauge, and precipitation is measured simultaneously. A variation on this is to have a series of small weighing lysimeters (such as small buckets) that can be removed and weighed individually every day to provide a record of weight loss. At the same time as weighing, the amount of percolation needs to be recorded. This is a very cheap way of estimating evaporation loss for a study using low technology.

Without any instrument to weigh the lysimeter (this is sometimes referred to as a *percolation gauge*) it must be assumed that the change in soil moisture over a period is zero and therefore evaporation equals rainfall minus runoff. This may be a reasonable assumption over a long time period such as a year where the soil storage will be approximately the same between two winters. An example of this type of lysimeter was the work of Law who investigated the effect that trees had on the water balance at Stocks Reservoir in Lancashire, UK (Law, 1956; see Case Study on p. 42).

A well-planned and executed lysimeter study probably provides the best information on evaporation that a hydrologist could find. However, it must be remembered that it is not evaporation that is being measured in a lysimeter – it is almost everything else in the water balance equation, with an assumption being made that whatever is left must be caused by evaporation. One result of this is that any errors in measurement of precipitation and/or percolation will transfer and possibly magnify into errors of evaporation measurement.

ESTIMATION OF EVAPORATION

The difficulties in measuring evaporation using either micro-meteorological instruments (problematic when used over long time periods and at the catchment scale) or water balance techniques (accumulated errors and small scale) has led to much effort being placed on estimating evaporation rather than trying to actually measure it. Some of the techniques outlined below are complicated and this sometimes leads hydrologists to believe that they are measuring, rather than estimating, evaporation. What they are actually doing is taking climatological variables that are known to influence evaporation and simulating evaporation rates from these: an estimation technique. The majority of research effort in this field has been to produce models to estimate evaporation; however, more recently, satellite remote sensing has provided another method of estimating the evaporation flux.

The techniques described here represent a range of sophistication and they are certainly not all universally applicable. Almost all of these are concerned with estimating the potential evaporation over a land surface. As with most estimation techniques the hydrologist is required to choose the best techniques for the study situation. In order to help in this decision the various advantages and shortcomings of each technique are discussed.

Thornthwaite

Thornthwaite derived an empirical model (i.e. derived from measurement not theoretical understanding) linking average air temperature to potential evaporation. This is an inherently sensible link in that we know air temperature is closely linked to both available energy and the ability of air to absorb water vapour.

The first part of the Thornthwaite estimation technique (Thornthwaite, 1944, 1954) derives a monthly heat index (i) for a region based on the average temperature t (°C) for a month (equation 3.4).

$$i = \left(\frac{t}{5}\right)^{1.514} \tag{3.4}$$

These terms are then summed to provide an annual heat index I (equation 3.5).

$$I = \sum_{j=1}^{12} i \tag{3.5}$$

Thornthwaite then derived an equation to provide evaporation estimates based on a series of observed evaporation measurements (equation 3.6).

$$PE = 16b\left(\frac{10t}{I}\right)^{a} \tag{3.6}$$

The a and b terms in this equation can be derived in the following ways. Term b is a correction factor to account for unequal day length between months. Its value can be found by looking up tables based on the latitude of your study site. Term a is calibrated as a cubic function from the I term such as is shown in equation 3.7.

$$a = 6.7 \times 10^{-7} I^3 - 7.7 \times 10^{-5} I^2 + 0.018\, I + 0.49 \tag{3.7}$$

Case study

A LYSIMETER USED TO MEASURE EVAPORATION FROM TUSSOCK

A narrow-leafed tussock grass (*Chionochloa rigida*, commonly called 'snow' or 'tall tussock') covers large areas of the South Island of New Zealand. A field study of a catchment dominated by snow tussock (Pearce *et al*., 1984) showed high levels of baseflow (i.e. high levels of streamflow between storm events). Mark *et al*. (1980) used a percolation gauge under a single tussock plant and estimating evaporation, showed that the water balance can show a surplus. They suggested that this may be due to the tussock intercepting fog droplets that are not recorded as rainfall in a standard rain gauge (see Plate 3). The nature of a tussock leaf (long and narrow with a sharp point), would seem to be conducive to fog interception in the same manner as conifers intercepting fog. Another interpretation of the Mark *et al*. (1980) study is that the estimation of evaporation was incorrect. An understanding of the mechanisms leading to high baseflow levels is important for a greater understanding of hydrological processes leading to streamflow.

In order to investigate this further a large lysimeter was set up in two different locations. The lysimeter was 2 m in diameter and contained nine mature snow tussock plants in an undisturbed monolith, weighing approximately 8,000 kg. Percolating runoff was measured with a tipping-bucket mechanism and the whole lysimeter was on a beam balance giving a sensitivity of 0.054 mm (Figure 3.5). The rainfall was measured immediately adjacent to the lysimeter. Campbell and Murray (1990) show that although there were times when fog interception appeared to occur (i.e. the catch in the lysimeter was greater than that in the nearby rain gauge) this only accounted for 1 per cent of the total precipitation. The detailed micro-meteorological measurements showed that the tussock stomatal or canopy resistance term was very high and that the plants had an ability to stop transpiring when the water stress became too high (see earlier discussion on plant physiological response to evaporative stress). The conclusion from the study was that snow tussocks are conservative in their use of water, which would appear to account for the high baseflow levels from tussock-covered catchments (Davie *et al*., 2006).

Figure 3.5 Large weighing lysimeter at Glendhu being installed. The weighing mechanism can be seen underneath.

(Photograph courtesy of Barry Fahey)

The Thornthwaite technique is extremely useful as potential evaporation can be derived from knowledge of average temperature (often readily available from nearby weather stations) and latitude. There are drawbacks to its usage however; most notably that it only provides estimates of monthly evaporation. For anything at a smaller time-scale it is necessary to use another technique such as

Penman's (see below). There are also problems with using Thornthwaite's model in areas of high potential evaporation. The empirical nature of the model means that it has been calibrated for a certain set of conditions and that it may not be applicable outside these. The Thornthwaite model has been shown to underestimate potential evaporation in arid and semi-arid regions (e.g. Acheampong, 1986). If the model is being applied in conditions different to Thornthwaite's original calibration (humid temperate regions) it is advisable to find out if any researcher has published different calibration curves for the climate in question.

Penman

Penman was a British physicist who derived a theoretical model of evaporation. Penman's first theoretical model was for open water evaporation and is shown in equations 3.8 and 3.9 (Penman, 1948):

$$E_o = \frac{\Delta Q^* + \gamma E_a}{\Delta + \gamma} \tag{3.8}$$

where an empirical relationship states that:

$$E_a = 2.6\delta_e\left(1 + \frac{u}{1.862}\right) \tag{3.9}$$

and Q^* = net radiation (in evaporation equivalent units of mm/day)
Δ = rate of increase of the saturation vapour deficit with temperature (kPa/°C see Figure 3.6)
δ_e = vapour pressure deficit of the air (kPa)
γ = psychrometric constant (≈ 0.063 kPa/°C)
u = wind speed at 2 m elevation (m/s)

In his original formula Penman estimated net radiation from empirical estimates of short- and long-wave radiation. The formula given here requires observations of temperature, wind speed, vapour pressure (which can be derived from relative humidity) and net radiation and gives the evaporation in units of mm per day. All of these can be obtained from meteorological measurement (see p. 49). It is normal to use daily averages for these variables, although Shuttleworth (1988) has suggested that it should not be used for time steps of less than ten days. There are several different ways of presenting this formula, which makes it difficult to interpret between texts. The main difference is in whether the evaporation is a flux or an absolute rate. In the equation above terms like 'net radiation' have been divided by the amount of energy required to evaporate 1 mm of water (density of water (ρ) multiplied by the latent heat of vaporisation (λ)) to turn them into water equivalents. This means the equation derives an absolute value for evaporation rather than a flux.

Penman continued his work to consider the evaporation occurring over a vegetated surface (Penman and Scholfield, 1951), while others refined the work (e.g. van Bavel, 1966). Part of this refinement was to include a term for aerodynamic resistance (r_a) to replace E_a (equation 3.9). **Aerodynamic resistance** is a term to account for the way in which the water evaporating off a surface mixes with a potentially drier atmosphere above it through turbulent mixing. The rougher the canopy surface the greater degree of turbulent mixing that will occur since air passing over the surface is buffeted around by protruding objects. As it is a resistance term, the higher the value, the greater the resistance to mixing; therefore a forest has a lower value of r_a than smoother pasture. Some values of aerodynamic resistance for different vegetation types are given in Table 3.3.

Substituting the new aerodynamic resistance term into the Penman equation, and presenting the results as a water flux (kg of water per m^2 of area), the evaporation estimation equation can be written as equation 3.10.

$$PE = \frac{Q^*\Delta + \frac{\rho \cdot c_\rho \cdot \delta_e}{r_a}}{\lambda(\Delta + \gamma)} \tag{3.10}$$

where

Q^* = net radiation (W/m^2)
Δ = rate of increase of the saturation vapour pressure with temperature (kPa/°C) (see Figure 3.6)
ρ = density of air (kg/m^3)
c_p = specific heat of air at constant pressure (≈ 1,005 J/kg)
δ_e = vapour pressure deficit of the air (kPa)
λ = latent heat of vaporisation of water (J/kg) (see Figure 3.4)
γ = psychrometric constant (≈0.063 kPa/°C)
r_a = aerodynamic resistance to transport of water vapour (s/m) given by equation 3.11.

$$r_a = \frac{\left(\ln\left(\frac{z-d}{z_0}\right)\right)^2}{\kappa^2 u} \tag{3.11}$$

and

κ = Von Karman constant (≈ 0.41)
u = wind speed above canopy (m/s)
z = height of anemometer (m)
d = **zero plane displacement** (the height within a canopy at which wind speed drops to zero, often estimated at two-thirds of the canopy height) (m)
z_o = roughness length (often estimated at one eighth of vegetation height) (m)

Table 3.3 Estimated values of aerodynamic and stomatal resistance for different vegetation types

Vegetation type	*Aerodynamic resistance (r_a) (s/m)*	*Canopy resistance (r_c) (s/m)*
Pasture	30	50
Forest	6.5	112
Scrub	6.5	160
Tussock	7.0	120

Source: from Andrew and Dymond (2007).
NB although the values of canopy resistance are presented as fixed they actually vary considerably throughout a day and season

Although this formula looks complicated it is actually rather simple. It is possible to split the equation into two separate parts that conform to the understanding of evaporation already discussed. The available energy term is predominantly assessed through the net radiation (Q^*) term. Other terms in the equation relate to the ability of the atmosphere to absorb the water vapour (Δ, ρ, c_p, δ_e, λ, γ, this is referred to as the sensible heat transfer function) and the rate at which diffusion will absorb the water vapour into the atmosphere (κ, u, z_o, etc.).

Figure 3.6 shows the relationship between the saturated vapour pressure and temperature. The slope of this curve (Δ) is required in the Penman equation and its derivatives. This can be estimated from equation 3.12 using average air temperature (T, °C):

$$\Delta = \frac{2053.058\exp^{\frac{17.27T}{T+237.3}}}{(T+237.3)^2} \tag{3.12}$$

When using the Penman equation there are only four variables requiring measurement: net radiation, wind speed above the canopy, atmospheric humidity and temperature, which when combined will provide vapour pressure deficit (see Figures 3.6–3.8).

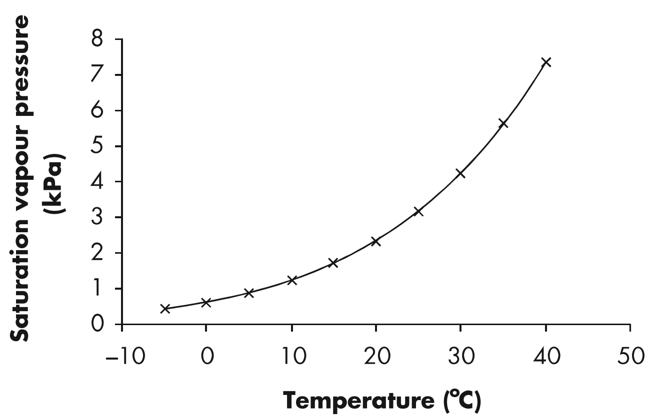

Figure 3.6 The relationship between temperature and saturation vapour pressure. This is needed to calculate the rate of increase of saturation vapour pressure with temperature (Δ). Equation 3.12 describes the form of this relationship.

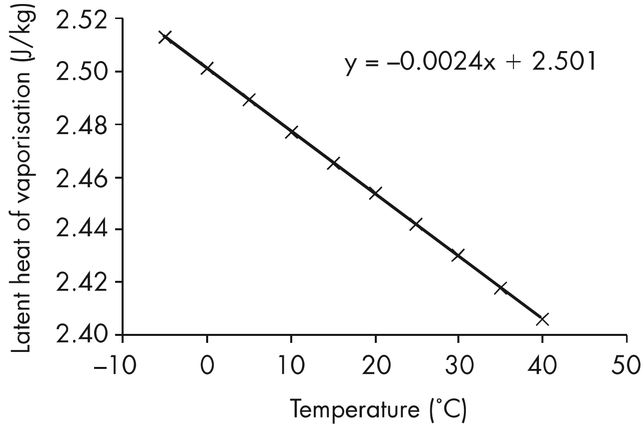

Figure 3.7 The relationship between temperature and latent heat of vaporisation.

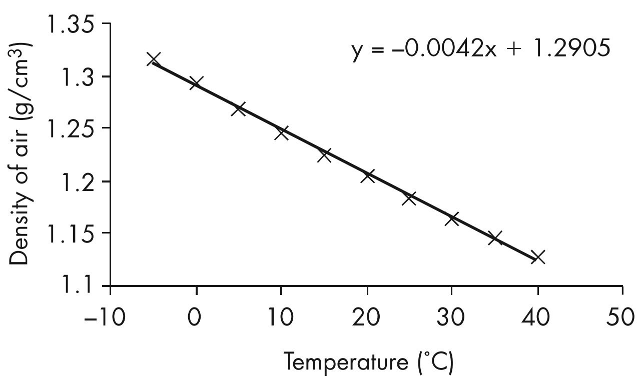

Figure 3.8 The relationship between air temperature and the density of air.

Every other term in the equation is either a constant, a simple relationship from another variable or can be measured once. Of these four variables net radiation is the hardest to obtain from meteorological stations as net radiometers are not common. There are methods of estimating net radiation from measurements of incoming solar radiation, surface **albedo** (or reflectivity) and day length (see Oke, 1987).

The modified Penman equation provides estimates of potential evaporation at a surface for time intervals much less than the monthly value from Thornthwaite. This makes it extremely useful to hydrology and it is probably the most widely used method for estimating potential evaporation values.

However, there are problems with the Penman equation which make it less than perfect as an estimation technique. The assumption is made that the soil heat flux is unimportant in the evaporation energy budget. This is often the case but is an acknowledged simplification that may lead to some overall error, especially when the time step is less than one day. It is normal practice to use Penman estimates at the daily time step; however, in some modelling studies they are used at hourly time steps.

One major problem with the Penman equation relates to its applicability in a range of situations and in particular in the role of advection, as discussed on p. 38. This is where there are other energy sources available for evaporation that cannot be assessed from net radiation. Calder (1990) shows the results from different studies in the UK uplands where evaporation rates vastly exceed the estimates provided by the Penman equation. The cause of this discrepancy is the extra energy provided by cyclonic storms coming onto Britain from the Atlantic Ocean, something that is poorly accounted for in the Penman equation. The part of the Penman equation dealing with the ability of the atmosphere to absorb the water vapour (sensible heat transfer function) does account for some advection but not if it is a major energy source driving evaporation and it is highly sensitive to the aerodynamic resistance term. This does not render the Penman approach invalid; rather, in applying it the user must be sure that net radiation is the main source of energy available for evaporation or the aerodynamic resistance term is well understood.

Simplifications to Penman

There have been several attempts made to simplify the Penman equation for widespread use. Slatyer and McIlroy (1961) separated out the evaporation caused by sensible heat and advection from that caused by radiative energy. Priestly and Taylor (1972) derived a simplified Penman formula for use in the large-scale estimation of evaporation, in the order of 'several hundred kilometres' where it can be argued that large-scale advection is not important. Their formula for potential evaporation is shown in equation 3.13.

$$PE = \alpha \frac{\left(Q^* - Q_G\right)\Delta}{\lambda\left(\Delta + \gamma\right)} \qquad (3.13)$$

where Q_G is the soil heat flux term (often ignored by Penman but easily included if the measurements are available) and α is the Priestly–Taylor parameter, all other parameters being as defined earlier. The α term is an approximation of the sensible heat transfer function and was estimated by Priestly and Taylor (1972) to have a value of 1.26 for saturated land surfaces, oceans and lakes – that is to say, the sensible heat transfer accounts for 26 per cent of the evaporation over and above that from net radiation. This value of α has been shown to vary away from 1.26 (e.g. α = 1.21 in Clothier *et al*., 1982) but to generally hold true for large-scale areas without a water deficit.

Penman–Monteith

Monteith (1965) derived a further term for the Penman equation so that actual evaporation from a vegetated surface could be estimated. His work involved adding a canopy resistance term (r_c) into the Penman equation so that it takes the form of equation 3.14.

$$E_t = \frac{Q^*\Delta + \rho c_p \delta_e / r_a}{\lambda\left(\Delta + \gamma\left(1 + \frac{r_c}{r_a}\right)\right)} \qquad (3.14)$$

Looking at the Penman–Monteith equation you can see that if r_c equals zero then it reverts to the Penman equation (i.e. actual evaporation equals potential evaporation). If the canopy resistance is high the actual evaporation rate drops to less than potential. Canopy resistance represents the ability of a vegetation canopy to control the rate of transpiration. This is achieved through the opening and closing of stomata within a leaf, hence r_c is sometimes referred to as stomatal resistance. Various researchers have established canopy resistance values for different vegetation types (e.g. Szeicz *et al*., 1969), although they are known to vary seasonally and in some cases diurnally. Rowntree (1991) suggests that for grassland under non-limiting moisture conditions the range of r_c should fall somewhere between 60 and 200 s/m. The large range is a reflection of canopy resistance being influenced by a plant's physiological response to variations in climatological conditions (see earlier discussion of stomatal control p. 40). Some values of canopy resistance for different vegetation types are given in Table 3.3.

Reference evaporation

The Penman–Monteith equation is probably the best evapotranspiration estimation method available. However for widespread use there is a need to have the stomatal resistance and aerodynamic resistance terms measured for a range of canopy covers at different stages of growth. To overcome this, the idea of reference evaporation has been introduced. This is the evaporation from a particular vegetation surface and the evaporation rate for another surface is related to this by means of crop coefficients. The Food and Agriculture Organisation (FAO) convened a group of experts who decided that the best surface for reference evaporation is close-cropped, well-watered grass. This is described in Allen *et al*. (1998) as a hypothetical reference crop with an assumed crop height of 0.12 m, a fixed canopy resistance of 70 s/m and an albedo of 0.23. Using these fixed values within the Penman–Monteith equation the reference evaporation (ET_o in mm/day) can be calculated from equation 3.15.

$$ET_o = \frac{0.408\Delta\left(Q^* - Q_G\right) + \gamma \cdot \frac{900}{T + 273} u \cdot \delta_e}{\Delta + \gamma\left(1 + 0.34_u\right)} \qquad (3.15)$$

where

Q^* is net radiation at the crop surface (MJ/m^2/day)

Q_G is soil heat flux density (MJ/m^2/day)
T is mean daily air temperature at 2 m height (°C)
u is wind speed at 2 m height (m/s)
δ_e is the saturation vapour pressure deficit (kPa)
Δ is the slope of the vapour pressure curve (kPa/°C)
γ is the psychrometric constant (kPa/°C)

The reference evapotranspiration, provides a standard to which evapotranspiration at different periods of the year or in other regions can be compared and evapotranspiration of other crops can be related (Allen *et al*., 1998). Scotter and Heng (2003) have investigated the sensitivity of the different inputs to the reference evaporation equation in order to show what accuracy of measurement is required.

Table 3.4 outlines some crop coefficients as set out by FAO (Allen *et al*., 1998). At the simplest level the evapotranspiration for a particular crop can be estimated by multiplying the crop coefficient with the reference evapotranspiration although there are more complex procedures outlined in Allen *et al*. (1998) which account for growth throughout a season and climatic variability. Where the crop coefficient values shown in Table 3.4 are higher than 1.0 it is likely that the aerodynamic roughness of the canopy makes for higher evaporation rates than for short grass. Where the values are less than 1.0 then the plants are exerting stomatal control on the transpiration rate.

Simple estimation of E_t from *PE* and soil moisture

Where there is no stomatal control exerted by plants (e.g. in a pasture) the relationship between actual evaporation (E_t) and potential evaporation (*PE*) is by and large driven by the availability of water. Over a land surface the availability of water can be estimated from the soil moisture content (see Chapter 4). At a simple level it is possible to estimate the relationship between potential and actual evaporation using soil moisture content as a measured variable (see Figure 3.9). In Figure 3.9 a value of 1 on the y-axis corresponds to actual precipitation equalling potential evaporation (i.e. available water is not a limiting factor on the evaporation rate). The exact position where this occurs will be dependent on the type of soil and plants on the land surface, hence the lack of units shown on the x-axis and the two different curves drawn. This type of simple relationship has been effective in determining actual evaporation rates in a model of soil water budgeting (e.g. Davie *et al*., 2001) but cannot be relied on for accurate modelling studies. It provides a very crude estimate of actual evaporation from knowledge of soil moisture and potential evaporation.

Table 3.4 Crop coefficients for calculating evapotranspiration from reference evapotranspiration

Crop type	*Crop coefficient (K_c)*	Comment
Beans and peas	1.05	Sometimes grown on stalks reaching 1.5 to 2 metres in height. In such cases, increased K_c values need to be taken.
Cotton	1.15–1.20	
Wheat	1.15	
Maize	1.15	
Sugar Cane	1.25	
Grapes	0.7	
Conifer forests	1.0	Confers exhibit substantial stomatal control. The K_c can easily reduce below the values presented, which represent well-watered conditions for large forests.
Coffee	0.95	

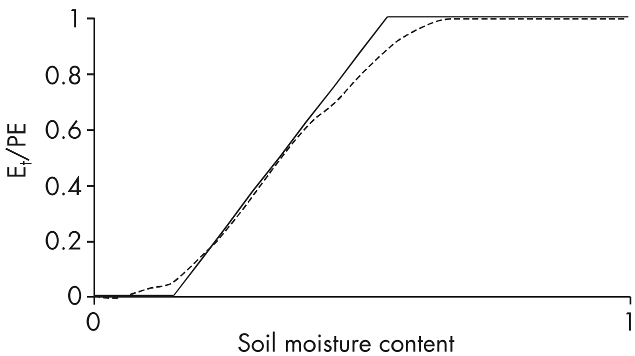

Figure 3.9 A hypothetical relationship between the measured soil moisture content and the ratio of actual evaporation to potential evaporation.

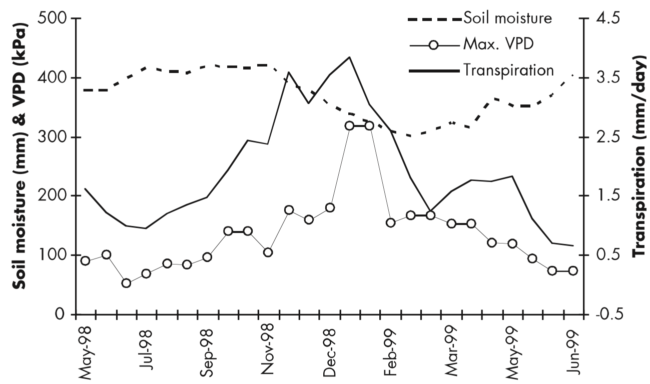

Figure 3.10 Time series of measured transpiration, measured soil moisture and estimated vapour pressure deficit for a forested site, near Nelson, New Zealand. NB as a Southern Hemisphere site the summer is from December until February.

Source: Data courtesy of Rick Jackson

The relationship between actual evaporation and soil moisture is not so simple where there is a vegetation type that exerts stomatal control on the evaporation rate (e.g. coniferous forest). In this case the amount of evaporation will be related to both soil moisture (available water) and the vapour pressure deficit (ability of the atmosphere to absorb water vapour). This is illustrated by Figure 3.10, a time series of soil moisture, transpiration and vapour pressure deficit for a stand of *Pinus radiata* in New Zealand. Transpiration was measured using sapflow meters on a range of trees; soil moisture was measured with a neutron probe and vapour pressure deficit was estimated from a nearby meteorological station. At the start of the summer period (Oct.–Nov. 1998) the soil moisture level is high and the transpiration rate climbs rapidly to a peak. Once it has reached the peak, the transpiration rate plateaus, despite the maximum vapour pressure deficit continuing to climb. During this plateau in transpiration rate the forest is exerting some stomatal control so that the transpiration doesn't increase by as much as the vapour pressure deficit. From January 1999 (the height of the Southern Hemisphere summer) the transpiration rate drops markedly. Initially this matches a drop in the maximum vapour pressure deficit but the transpiration rate continues to drop below early summer rates (with similar VPD values). This is the time that the lack of soil moisture is starting to limit the tree transpiration. Figure 3.10 illustrates the complex relationship between evaporation from a vegetated surface, the soil moisture conditions and the atmospheric conditions.

Remote sensing of evaporation

Water vapour is a greenhouse gas and therefore it interferes with radiation (i.e. absorbs and reradiates) from the earth's surface. Because of this the amount of water vapour in the atmosphere can be estimated using satellite remote sensing, particularly using passive microwave sensors. The difficulty with using this information for hydrology is that it is at a very large scale (often continental) and is concerned with the whole atmosphere not the near surface. In order to utilise satellites for estimation of evaporation a combined modelling and remote sensing approach is required. Burke *et al*. (1997) describe a combined Soil–Vegetation–Atmosphere–Transfer (SVAT) model that is driven by remotely sensed data. This type of approach can be used to estimate evaporation rates over a large spatial area relatively easily. Mauser and Schädlich (1998) provide a review of evaporation modelling at different scales using remotely sensed data.

Mass balance estimation

In the same manner that evaporation pans and lysimeters estimate evaporation rates, evaporation at the large scale (catchment or lake) can be estimated through the water balance equation. This is a relatively crude method, but it can be extremely effective over a large spatial and/or long temporal scale. The method requires accurate measurement of precipitation and runoff for a catchment or lake. In the case of a lake, change in storage can be estimated through lake-level recording and knowledge of the surface area. For a catchment it is often reasonable to assume that change in storage is negligible over a long time period (e.g. one year) and therefore the evaporation is precipitation minus runoff.

Canopy interception loss estimation

Empirical models that link rainfall to interception loss based on regression relationships of measured data sets have been developed for many different types of vegetation canopy (see Zinke (1967) and Massman (1983) for examples and reviews of these types of model). Some of these models used logarithmic or exponential terms in the equations but they all rely on having regression coefficients based on the vegetation type and climatic regime.

A more detailed modelling approach is the Rutter model (Rutter *et al.*, 1971, 1975) which calculates an hourly water balance within a forest stand. The water balance is calculated, taking into account the rate of throughfall, stemflow, interception loss through evaporation and canopy storage. In order to use the model a detailed knowledge of the canopy characteristics is required. In particular the canopy storage and drainage rates from throughfall are required to be known; the best method for deriving these is through empirical measurement. The Rutter model treats the canopy as a single large leaf, although it has been adapted to provide a three-dimensional canopy (e.g. Davie and Durocher, 1997) that can then be altered to allow for changes and growth in the canopy.

At present, remote sensing techniques are not able to provide reasonable estimates of canopy interception. They do provide some useful information that can be incorporated into canopy interception models but cannot provide the detailed difference between above- and below-canopy rainfall. In particular, satellites can give good information on the type of vegetation and its degree of cover. Particular care needs to be taken over the term 'leaf area index' when reading remote sensing literature. Analysis of remotely sensed images can provide a good indication of the percentage vegetation cover for an area, but this is not necessarily the same as leaf area index – although it is sometimes referred to as such. Leaf area index is the surface area of leaf cover above a defined area divided by the surface area defined. As there are frequently layers of vegetation above the ground, the leaf area index frequently has a value higher than one. The percentage vegetation cover cannot exceed one (as a unitary percentage) as it does not consider the third dimension (height).

EVAPORATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Evaporation, as the only loss away from the surface in the water balance equation, plays a large part in water quantity. The loss of water from soil through direct evaporation and transpiration has a direct impact on the amount of water reaching a stream during high rainfall (see Chapter 5) and also the amount of water able to infiltrate through into groundwater (see Chapter 4). The impact of evaporation on water quantity is not as great as for precipitation but it does have a significant part to play in the quantity and timing of water flowing down a river.

The influence of evaporation on water quality is mostly through the impurities left behind after water has evaporated. This may lead to a concentration of impurities in the water remaining behind (e.g. the Dead Sea between Israel and Jordan) or a build up of salts in soils (salination). This is discussed in more detail in Chapter 8.

SUMMARY

The evaporation process involves the transfer of water from a liquid state into a gaseous form in the atmosphere. For this to happen requires an available energy source, a water supply and the ability of the atmosphere to receive it. Evaporation is difficult to measure directly and there are various estimation techniques. These range from water budget techniques, such as evaporation pans and lysimeters, to modelling techniques, such as the Penman–Monteith equation. As a process, evaporation suffers from the same problems with measurement and estimation as does precipitation (i.e. extreme variability in space and time). This variability leads to difficulties in moving from point measurements to areal estimates such as are required for a catchment study. These can be overcome by using spatial averaging techniques or using evaporation estimations that assume a large base area (e.g. Priestly–Taylor). Forests have an important role to play in evaporation, particularly through interception loss. In general, more water is lost from a forested catchment than a non-forested catchment. This is through evaporation off wet leaves, but this is not always the case – there are cases where a tree canopy leads to more water in the catchment. The importance of canopy interception in a catchment water balance is dependent on the size and extent of vegetation cover found within a watershed.

ESSAY QUESTIONS

1. **Give a detailed account of the factors influencing evaporation rate above a forest canopy.**
2. **Compare and contrast the use of evaporation pans and lysimeters for measuring evaporation.**
3. **Outline the major evaporation estimation techniques and compare their effectiveness for your local environment.**
4. **Describe the factors that restrict actual evaporation (evapotranspiration) from equalling potential evaporation in a humid-temperate climate.**

FURTHER READING

Allen, R.G., Pereira L.S., Raes, D. and Smith, D. (1998) Crop evapotranspiration – Guidelines for computing crop water requirements. *FAO Irrigation and drainage paper 56* (available at www.fao.org/documents).

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Calder, I.R. (1990) *Evaporation in the uplands*. J. Wiley & Sons, Chichester.
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Cheng, M. (2003) Forest hydrology: an introduction to water and forests. CRC Press, Boca Raton, Florida.
An overview of forest hydrological processes, including evaporation and interception loss.