2

PRECIPITATION

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the processes of precipitation formation.
- A knowledge of the techniques for measuring precipitation (rainfall and snow).
- An appreciation of the associated errors in measuring precipitation.
- A knowledge of how to analyse rainfall data spatially and for intensity/duration of a storm.
- A knowledge of some of the methods used to estimate rainfall at the large scale.
- An understanding of the process of precipitation interception by a canopy.

PRECIPITATION AS A PROCESS

Precipitation is the release of water from the atmosphere to reach the surface of the earth. The term 'precipitation' covers all forms of water being released by the atmosphere, including snow, hail, sleet and rainfall. It is the major input of water to a river catchment area and as such needs careful assessment in any hydrological study. Although rainfall is relatively straightforward to measure (other forms of precipitation are more difficult) it is notoriously difficult to measure *accurately* and, to compound the problem, is also extremely variable within a catchment area.

Precipitation formation

The ability of air to hold water vapour is temperature dependent: the cooler the air the less water vapour is retained. If a body of warm, moist air is cooled then it will become saturated with water vapour and eventually the water vapour will condense into liquid or solid water (i.e. water or ice droplets). The water will not condense spontaneously however; there need to be minute particles present in the atmosphere, called **condensation nuclei**, upon which the water or ice droplets form. The water or ice droplets that form on condensation nuclei are normally too small to fall to the surface as precipitation; they need to grow in order to have enough mass to overcome uplifting forces within a cloud. So there are three conditions that need to be met prior to precipitation forming:

1 Cooling of the atmosphere
2 Condensation onto nuclei
3 Growth of the water/ice droplets.

Atmospheric cooling

Cooling of the atmosphere may take place through several different mechanisms occurring independently or simultaneously. The most common form of cooling is from the uplift of air through the atmosphere. As air rises the pressure decreases; Boyle's Law states that this will lead to a corresponding cooling in temperature. The cooler temperature leads to less water vapour being retained by the air and conditions becoming favourable for **condensation**. The actual uplift of air may be caused by heating from the earth's surface (leading to **convective precipitation**), an air mass being forced to rise over an obstruction such as a mountain range (this leads to **orographic precipitation**), or from a low pressure weather system where the air is constantly being forced upwards (this leads to **cyclonic precipitation**). Other mechanisms whereby the atmosphere cools include a warm air mass meeting a cooler air mass, and the warm air meeting a cooler object such as the sea or land.

Condensation nuclei

Condensation nuclei are minute particles floating in the atmosphere which provide a surface for the water vapour to condense into liquid water upon. They are commonly less than a micron (i.e. one-millionth of a metre) in diameter. There are many different substances that make condensation nuclei, including small dust particles, sea salts and smoke particles.

Research into generating artificial rainfall has concentrated on the provision of condensation nuclei into clouds, a technique called **cloud seeding**. During the 1950s and 1960s much effort was expended in using silver iodide particles, dropped from planes, to act as condensation nuclei. However, more recent work has suggested that other salts such as potassium chloride are better nuclei. There is much controversy over the value of cloud seeding. Some studies support its effectiveness (e.g. Gagin and Neumann, 1981; Ben-Zvi, 1988); other authors query the results (e.g. Rangno and Hobbs, 1995), while others suggest that it only works in certain atmospheric conditions and with certain cloud types (e.g. Changnon *et al*., 1995). More recent work in South Africa has concentrated on using hygroscopic flares to release chloride salts into the base of convective storms, with some success (Mather *et al*., 1997). Interestingly, this approach was first noticed through the discovery of extra heavy rainfall occurring over a paper mill in South Africa that was emitting potassium chloride from its chimney stack (Mather, 1991).

Water droplet growth

Water or ice droplets formed around condensation nuclei are normally too small to fall directly to the ground; that is, the forces from the upward draught within a cloud are greater than the gravitational forces pulling the microscopic droplet downwards. In order to overcome the upward draughts it is necessary for the droplets to grow from an initial size of 1 micron to around 3,000 microns (3 mm). The vapour pressure difference between a droplet and the surrounding air will cause it to grow through condensation, albeit rather slowly. When the water droplet is ice the vapour pressure difference with the surrounding air becomes greater and the water vapour sublimates onto the ice droplet. This will create a precipitation droplet faster than condensation onto a water droplet, but is still a slow process. The main mechanism by which raindrops grow within a cloud is through *collision and coalescence*. Two raindrops collide and join together (coalesce) to form a larger droplet that may then collide with many more before falling towards the surface as rainfall or another form of precipitation.

Another mechanism leading to increased water droplet size is the so-called **Bergeron process**. The pressure exerted within the parcel of air, by having the water vapour present within it, is called the **vapour pressure**. The more water vapour present the greater the vapour pressure. Because there is a maximum amount of water vapour that can be held by the parcel of air there is also a maximum vapour pressure, the so-called **saturation vapour pressure**. The saturation vapour pressure is greater over a water droplet than an ice droplet because it is easier for water molecules to escape from the surface of a liquid than a solid. This creates a water vapour gradient between water droplets and ice crystals so that water vapour moves from the water droplets to the ice crystals, thereby increasing the size of the ice crystals. Because clouds are usually a mixture of water vapour, water droplets and ice crystals, the Bergeron process may be a significant factor in making water droplets large enough to become rain drops (or ice/snow crystals) that overcome gravity and fall out of the clouds.

The mechanisms of droplet formation within a cloud are not completely understood. The relative proportion of condensation-formed, collision-formed, and Bergeron-process-formed droplets depends very much on the individual cloud circumstances and can vary considerably. As a droplet is moved around a cloud it may freeze and thaw several times, leading to different types of precipitation (see Table 2.1).

Dewfall

The same process of condensation occurs in **dewfall**, only in this case the water vapour condenses into liquid water after coming into contact with a cold surface. In humid-temperate countries dew is a common occurrence in autumn when the air at night is still warm but vegetation and other surfaces have cooled to the point where water vapour coming into contact with them condenses onto the leaves and forms dew. Dew is not normally a major part of the hydrological cycle but is another form of precipitation.

PRECIPITATION DISTRIBUTION

The amount of precipitation falling over a location varies both spatially and temporally (with time). The different influences on the precipitation can be divided into static and dynamic influences. Static influences are those such as altitude, aspect and slope; they do not vary between storm events. Dynamic influences are those that do change and are by and large caused by variations in the weather. At the global scale the influences on precipitation distribution are mainly dynamic being caused by differing weather patterns, but there are static factors such as topography that can also cause major variations through a **rain shadow effect** (see case study on pp. 18–19). At the continental scale large differences in rainfall can be attributed to a mixture

Table 2.1 Classes of precipitation used by the UK Meteorological Office

Class	*Definition*
Rain	Liquid water droplets between 0.5 and 7 mm in diameter
Drizzle	A subset of rain with droplets less than 0.5 mm
Sleet	Freezing raindrops; a combination of snow and rain
Snow	Complex ice crystals agglomerated
Hail	Balls of ice between 5 and 125 mm in diameter

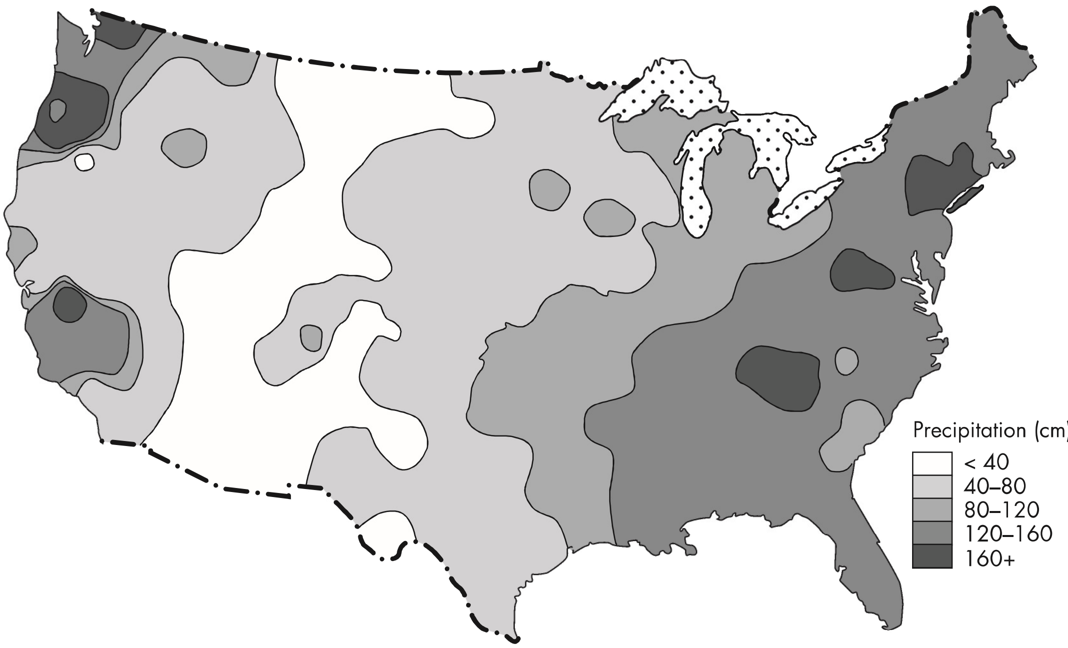


Figure 2.1 Annual precipitation across the USA during 1996.

Source: Redrawn with data from the National Atmospheric Deposition Program

of static and dynamic factors. In Figure 2.1 the rainfall distribution across the USA shows marked variations. Although mountainous areas have a higher rainfall, and also act as a block to rainfall reaching the drier centre of the country, they do not provide the only explanation for the variations evident in Figure 2.1. The higher rainfall in the north-west states (Oregon and Washington) is linked to wetter cyclonic weather systems from the northern Pacific that do not reach down to southern California. Higher rainfall in Florida and other southern states is linked to the warm waters of the Caribbean sea. These are examples of dynamic influences as they vary between rainfall events.

At smaller scales the static factors are often more dominant, although it is not uncommon for quite large variations in rainfall across a small area caused by individual storm clouds to exist. As an example: on 3 July 2000 an intense rainfall event caused flooding in the village of Epping Green, Essex, UK. Approximately 10 mm of rain fell within one hour, although there was no recorded rainfall in the village of Theydon Bois approximately 10 km to the south. This large spatial difference in rainfall was caused by the scale of the weather system causing the storm – in this case a convective thunderstorm. Often these types of variation lessen in importance over a longer timescale so that the annual rainfall in Epping Green and Theydon Bois is very similar, whereas the daily rainfall may differ considerably. For the hydrologist, who is interested in rainfall at the small scale, the only way to try and characterise these dynamic variations is through having as many **rain gauges** as possible within a study area.

Static influences on precipitation distribution

It is easier for the hydrologist to account for static variables such as those discussed below.

Altitude

It has already been explained that temperature is a critical factor in controlling the amount of water vapour that can be held by air. The cooler the air is, the less water vapour can be held. As temperature decreases with altitude it is reasonable to assume that as an air parcel gains altitude it is more likely to release the water vapour and cause higher rainfall. This is exactly what does happen and there is a strong correlation between altitude and rainfall: so-called *orographic precipitation*.

Aspect

The influence of aspect is less important than altitude but it may still play an important part in the distribution of precipitation throughout a catchment. In the humid mid-latitudes (35° to 65° north or south of the equator) the predominant source of rainfall is through cyclonic weather systems arriving from the west. Slopes within a catchment that face eastwards will naturally be more sheltered from the rain than those facing westwards. The same principle applies everywhere: slopes with aspects facing away from the predominant weather patterns will receive less rainfall than their opposites.

Slope

The influence of slope is only relevant at a very small scale. Unfortunately the measurement of rainfall occurs at a very small scale (i.e. a rain gauge). The difference between a level rain gauge on a hillslope, compared to one parallel to the slope, may be significant. It is possible to calculate this difference if it is assumed that rain falls vertically – but of course rain does not always fall vertically. Consequently the effect of slope on rainfall measurements is normally ignored.

Rain shadow effect

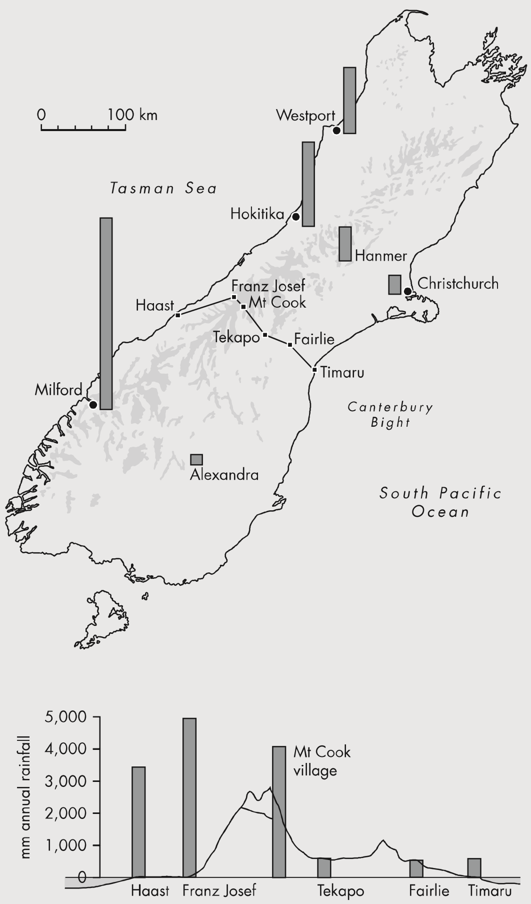
Where there is a large and high land mass it is common to find the rainfall considerably higher on one side than the other. This is through a combination of altitude, slope, aspect and dynamic weather direction influences and can occur at many different scales (see Case Study below).

Case study

THE RAIN SHADOW EFFECT

The predominant weather pattern for the South Island of New Zealand is a series of rain-bearing depressions sweeping up from the Southern Ocean, interrupted by drier blocking anticyclones. The Southern Alps form a major barrier to the fast-moving depressions and have a huge influence on the rainfall distribution within the South Island. Formed as part of tectonic uplift along the Pacific/Indian plate boundary, the Southern Alps stretch the full length of the South Island (approximately 700 km) and at their highest point are over 3,000 m above mean sea level.

The predominant weather pattern has a westerly airflow, bringing moist air from the Tasman Sea onto the South Island. As this air is forced up over the Southern Alps it cools down and releases much of its moisture as rain and snow. As the air descends on the eastern side of the mountains it warms up and becomes a föhn wind, referred to locally as a ‘nor-wester’. The annual rainfall patterns for selected stations in the South Island are shown in Figure 2.2. The rain shadow effect can be clearly seen with the west coast rainfall being at least four times that of the east. Table 2.2



also illustrates the point, with the number of rain days at different sites in a cross section across the South Island. Although not shown on the transect in Figure 2.2 recordings of rainfall further north in the Southern Alps (Cropp River inland from Hokitika) are as high as 6 m a year.

This pattern of rain shadow is seen at many different locations around the globe. It does not require as large a barrier as the Southern Alps – anywhere with a significant topographical barrier is likely to cause some form of rain shadow. Hayward and Clarke (1996) present data showing a strong rain shadow across the Freetown Peninsula in Sierra Leone. They analysed mean monthly rainfall in 31 gauges within a 20 × 50 km area, and found that the rain shadow effect was most marked during the monsoon months of June to October. The gauges in locations facing the ocean (south-west aspect) caught considerably more rainfall during the monsoon than those whose aspect was towards the north-east and behind a small range of hills.

Figure 2.2 Rainfall distribution across the Southern Alps of New Zealand (South Island). Shaded areas on the map are greater than 1,500 m in elevation. A clear rain shadow effect can be seen between the much wetter west coast and the drier east.

Table 2.2 Average annual rainfall and rain days for a cross section across the South Island

Weather station	*Height above mean sea level*	*Annual rainfall (mm)*	*Rain days*
Haast	30	5,840	175
Mt Cook village	770	670	120
Tekapo	762	604	77
Timaru	25	541	75

Note: More details on weather differentials across the South Island of New Zealand are in Sinclair *et al.* (1996)
Source: Data from New Zealand Met. Service and other miscellaneous sources

Forest rainfall partitioning

Once rain falls onto a vegetation canopy it effectively partitions the water into separate modes of movement: **throughfall**, **stemflow** and **interception loss**. This is illustrated in Figure 2.3.

Throughfall

This is the water that falls to the ground either directly, through gaps in the canopy, or indirectly, having dripped off leaves, stems or branches. The amount of *direct throughfall* is controlled by the canopy coverage for an area, a measure of which is the leaf area index (LAI). LAI is actually the ratio of leaf area to ground surface area and consequently has a value greater than one when there is more than one layer of leaf above the ground. When the LAI is less than one you would expect some direct throughfall to occur. When you shelter under a tree during a rainstorm you are trying to avoid the rainfall and direct throughfall. The greater the surface area of leaves above you, the more likely it is that you will avoid getting wet from direct throughfall.

The amount of *indirect throughfall* is also controlled by the LAI, in addition to the **canopy storage capacity** and the rainfall characteristics. Canopy storage capacity is the volume of water that can be held by the canopy before water starts dripping as indirect throughfall. The canopy storage capacity is controlled by the size of trees, plus the area and water-holding capacity of individual leaves. Rainfall characteristics are an important control on indirect throughfall as they dictate how quickly the canopy storage capacity is filled. Experience of standing under trees during a rainstorm should tell you that intensive rainfall quickly turns into indirect throughfall (i.e. you get wet!), whereas light showers frequently do not reach the ground surface at all. In reality canopy storage capacity is a rather nebulous concept. Canopy characteristics are constantly changing and it is rare for water on a canopy to fill up completely before creating indirect

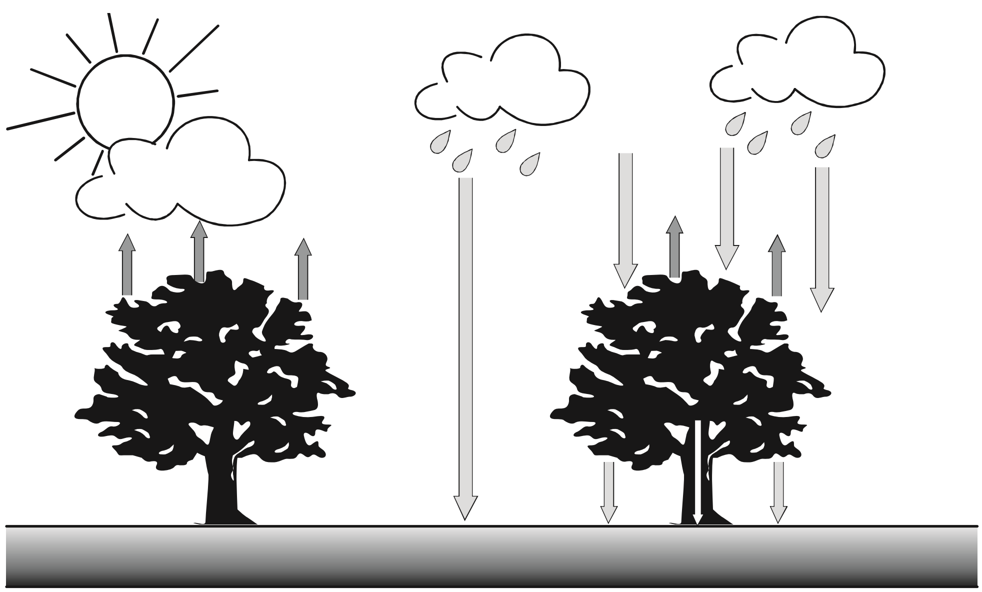


Figure 2.3 Rainfall above and below a canopy. Indicated on the diagram are stemflow (white arrow on trunk); direct and indirect throughfall (lightly hatched arrow); and interception loss (upward-facing darker arrow).

throughfall. This means that indirect throughfall occurs before the amount of rainfall equals the canopy storage capacity, making it difficult to gauge exactly what the storage capacity is.

Stemflow

Stemflow is the rainfall that is intercepted by stems and branches and flows down the tree trunk into the soil. Although measurements of stemflow show that it is a small part of the hydrological cycle (normally 2–10 per cent of above canopy rainfall; Lee, 1980) it can have a much more significant role. Durocher (1990) found that trees with smoother bark such as beech (*Fagus*) had higher rates of stemflow as the smoothness of bark tends to enhance drainage towards stemflow.

Stemflow acts like a funnel (see Figure 2.4), collecting water from a large area of canopy but delivering it to the soil in a much smaller area: the surface of the trunk at the base of a tree. This is most obvious for the deciduous oak-like tree illustrated in Figure 2.4, but it still applies for other structures (e.g. conifers) where the area of stemflow entry into the soil is far smaller than the canopy catchment area for rainfall. At the base of a tree it is possible for the water to rapidly enter the soil through flow along roots and other macropores surrounding the root structure. This can act as a rapid conduit of water sending a significant pulse into the **soil water**.

Interception loss

While water sits on the canopy, prior to indirect throughfall or stemflow, it is available for evaporation, referred to as *interception loss*. This is an evaporation process and it is discussed further in the following chapter.

Interception gain

In some circumstances it is possible that there is an interception gain from vegetation. In the Bull Run catchment, Oregon, USA it has been shown that the water yield after timber harvesting was significantly less than prior to the trees being logged (Harr, 1982; Ingwersen, 1985). This is counter to the majority of catchment studies reported by Bosch and Hewlett (1982) which show an increase in water yield as forests are logged. The reason for the loss of water with the corresponding loss of trees in Oregon

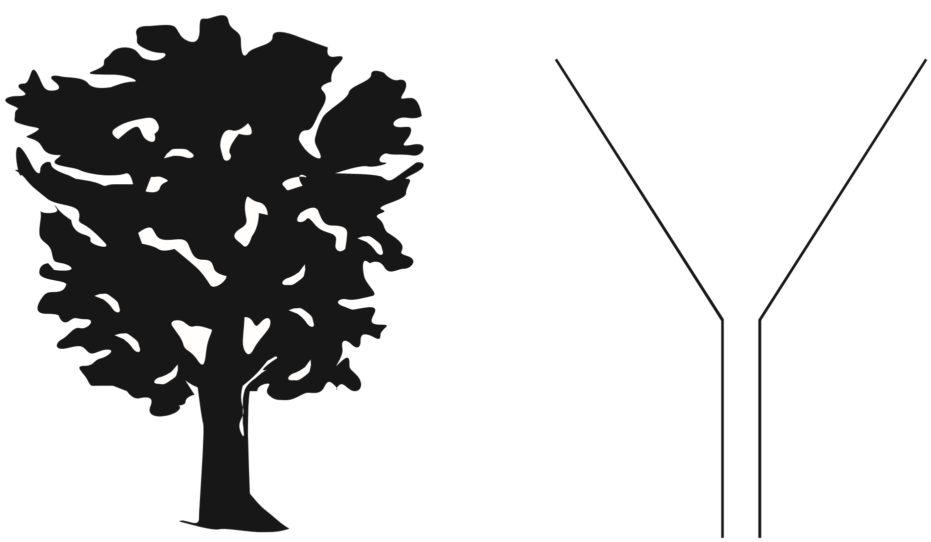


Figure 2.4 The funnelling effect of a tree canopy on stemflow.

is to do with the particular circumstances of the catchment. Fog from the cold North Pacific, with no accompanying rain, is a common feature and it is believed that the trees intercept fog particles, creating 'fog drip' which is a significant part of the water balance. Fog droplets are extremely small and Ingwersen (1985) has suggested that the sharp ends of needles on pine trees act as condensation nuclei, promoting the growth of larger droplets that fall to the ground (see an example of fogdrip from tussock leaves in Plate 3). When the trees are removed there are no condensation nuclei (or far fewer) on the resultant vegetation so the water remains in the atmosphere and is 'lost' in terms of water yield. Equally important is the influence of vegetation roughness. The turbulent mixing of air as wind passes over a rough canopy promotes rapid deposition of condensing water (directly converse to interception loss, see Chapter 3). The overall result of this is that the removal of trees leads to less water in the river; this runs counter to the evidence provided in the Case Study in Chapter 4.

MEASUREMENT

For hydrological analysis it is important to know how much precipitation has fallen and when this occurred. The usual expression of precipitation is as a vertical depth of liquid water. Rainfall is measured by millimetres or inches depth, rather than by volume such as litres or cubic metres. The measurement is the depth of water that would accumulate on the surface if all the rain remained where it had fallen (Shaw, 1994). Snowfall may also be expressed as a depth, although for hydrological purposes it is most usefully described in water equivalent depth (i.e. the depth of water that would be present if the snow melted). This is a recognition that snow takes up a greater volume (as much as 90 per cent more) for the same amount of liquid water.

There is a strong argument that can be made to say that there is no such thing as precipitation measurement at the catchment scale as it varies so tremendously over a small area. The logical end-point to this argument is that all measurement techniques are in fact precipitation estimation techniques. For the sake of clarity in this text precipitation measurement techniques refer to the methods used to quantify the volume of water present, as opposed to estimation techniques where another variable is used as a surrogate for the water volume.

Rainfall measurement

The instrument for measuring rainfall is called a *rain gauge*. A rain gauge measures the volume of water that falls onto a horizontal surface delineated by the rain gauge rim (see Figure 2.5). The volume is converted into a rainfall depth through division by the rain gauge surface area. The design of a rain gauge is not as simple as it may seem at first glance as there are many errors and inaccuracies that need to be minimised or eliminated.

Figure 2.5 A rain gauge sitting above the surface to avoid splash.

There is a considerable scientific literature studying the accuracy and errors involved in measuring rainfall. It needs to be borne in mind that a rain gauge represents a very small point measurement (or sample) from a much larger area that is covered by the rainfall. Any errors in measurement will be amplified hugely because the rain gauge collection area represents such a small sample size. Because of this amplification it is extremely important that the design of a rain gauge negates any errors and inaccuracies.

The four main sources of error in measuring rainfall that need consideration in designing a method for the accurate measurement of rainfall are:

1 Losses due to evaporation
2 Losses due to wetting of the gauge
3 Over-measurement due to splash from the surrounding area
4 Under-measurement due to turbulence around the gauge.

Evaporation losses

A rain gauge can be any collector of rainfall with a known collection area; however, it is important that any rainfall that does collect is not lost again through evaporation. In order to eliminate, or at least lessen this loss, rain gauges are funnel shaped. In this way the rainfall is collected over a reasonably large area and then any water collected is passed through a narrow aperture to a collection tank underneath. Because the collection tank has a narrow top (i.e. the funnel mouth) there is very little interchange of air with the atmosphere above the gauge. As will be explained in Chapter 3, one of the necessary requirements for evaporation is the turbulent mixing of saturated air with drier air above. By restricting this turbulent transfer there is little evaporation that can take place. In addition to this, the water awaiting measurement is kept out of direct sunlight so that it will not be warmed; hence there is a low evaporation loss.

Wetting loss

As the water trickles down the funnel it is inevitable that some water will stay on the surface of the funnel and can be lost to evaporation or not measured in the collection tank. This is often referred to as a *wetting loss*. These losses will not be large but may be significant, particularly if the rain is falling as a series of small events on a warm day. In order to lessen this loss it is necessary to have steep sides on the funnel and to have a non-stick surface. The standard UK Meteorological Office rain gauge is made of copper to create a non-stick surface, although many modern rain gauges are made of non-adhesive plastics.

Rain splash

The perfect rain gauge should measure the amount of rainfall that would have fallen on a surface if the gauge was not there. This suggests that the ideal situation for a rain gauge is flush with the surface. A difficulty arises, however, as a surface-level gauge is likely to over-measure the catch due to rain landing adjacent to the gauge and splashing into it. If there was an equal amount of splash going out of the gauge then the problem might not be so severe, but the sloping sides of the funnel (to reduce evaporative losses) mean that there will be very little splash-out. In extreme situations it is even possible that the rain gauge could be flooded by water flowing over the surface or covered by snow. To overcome the splash, flooding and snow coverage problem the rain gauge can be raised up above the ground (Figure 2.5) or placed in the middle of a non-splash grid (see Figure 2.6).

Turbulence around a raised gauge

If a rain gauge is raised up above the ground (to reduce splash) another problem is created due to air turbulence around the gauge. The rain gauge presents an obstacle to the wind and the consequent aerodynamic interference leads to a reduced catch (see Figure 2.7). The amount of loss is dependent on

Figure 2.6 Surface rain gauge with non-splash surround.

Figure 2.8 Baffles surrounding a rain gauge to lessen the impact of wind turbulence. The gauge is above ground because of snow cover during the winter.

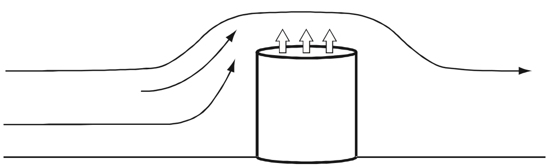


Figure 2.7 The effect of wind turbulence on a raised rain gauge. An area of reduced pressure (and uplift) develops above the gauge in a similar manner to an aircraft wing. This reduces the rain gauge catch.

both the wind speed and the raindrop diameter (Nešpor and Sevruk, 1999). At wind speeds of 20 km/hr (Beaufort scale 2) the loss could be up to 20 per cent, and in severe winds of 90 km/hr (Beaufort scale 8) up to 40 per cent (Bruce and Clark, 1980; Rodda and Smith, 1986). The higher a gauge is from the surface the greater the loss of accuracy. This creates a major problem for gauges in areas that receive large snowfalls as they need to be raised to avoid surface coverage.

One method of addressing these turbulence difficulties is through the fitting of a shield to the rain gauge (see Figure 2.8). A rain gauge shield can take many forms but is often a series of batons surrounding the gauge at its top height. The shield acts as a calming measure for wind around the gauge and has been shown to greatly improve rain gauge accuracy.

The optimum rain gauge design

There is no perfect rain gauge. The design of the best gauge for a site will be influenced by the individual conditions at the site (e.g. prevalence of snowfall, exposure, etc.). A rain gauge with a non-splash surround, such as in Figure 2.6, can give very accurate measurement but it is prone to coverage by heavy snowfall so cannot always be used. The non-splash surround allows adjacent rainfall to pass through (negating splash) but acts as an extended soil surface for the wind, thereby eliminating the turbulence problem from raised gauges. This may be the closest that it is possible to get to measuring the amount of rainfall that would have fallen on a surface if the rain gauge were not there.

The standard UK Meteorological Office rain gauge has been adopted around the world (although not everywhere) as a compromise between the factors influencing rain gauge accuracy. It is a brass-rimmed rain gauge of 5 inches (127 mm) diameter standing 1 foot (305 mm) above the ground. The lack of height above ground level is a reflection of the low incidence of snowfall in the UK; in countries such as Russia and Canada, where winter snowfall is the norm, gauges may be raised as high as 2 m above the surface. There is general recognition that the UK standard rain gauge is not the best design for hydrology, but it does represent a reasonable compromise. There is a strong argument to be made against changing its design. Any change in the measurement instrument would make an analysis of past rainfall patterns difficult due to the differing accuracy.

Siting of a rain gauge

Once the best measurement device has been chosen for a location there is still a considerable measurement error that can occur through incorrect siting. The major problem of rain gauge siting in hydrology is that the scientist is trying to measure the rainfall at a location that is representative of a far greater area. It is extremely important that the measurement location is an appropriate surrogate for the larger area. If the area of interest is a forested catchment then it is reasonable to place your rain gauge beneath the forest canopy; likewise, within an urban environment it is reasonable to expect interference from buildings because this is what is happening over the larger area. What is extremely important is that there are enough rain gauges to try and quantify the spatial and temporal variations.

The rule-of-thumb method for siting a rain gauge is that the angle when drawn from the top of the rain gauge to the top of the obstacle is less than 30° (see Figure 2.9). This can be approximated as at least twice the height of the obstacle away from the gauge. Care needs to be taken to allow for the future growth of trees so that at all times during the rainfall record the distance apart is at least twice the height of an obstacle.

Gauges for the continuous measurement of rainfall

The standard UK Meteorological Office rain gauge collects water beneath its funnel and this volume is read once a day. Often in hydrology the data needs to be measured at a finer timescale than this, particularly in the case of individual storms which often last much less than a day. The most common modern method for measuring continuous rainfall uses a tipping-bucket rain gauge. These are very simple devices that can be installed relatively cheaply, although they do require a data-logging device nearby. The principle behind the tipping-bucket rain gauge is that as the rain falls it fills up a small ‘bucket’ that is attached to another ‘bucket’ on a balanced cross arm (see Figure 2.10). The ‘buckets’ are very small plastic containers at the end of each cross arm. When the bucket is full it tips the balance so that the full bucket is lowered down and empties out. At the time of tipping a magnet attached to the balance arm closes a small reed switch which sends an electrical signal to a data-logging device. This then records the exact time of the tipped bucket. If the rain continues to fall it fills the bucket on the other end of the cross

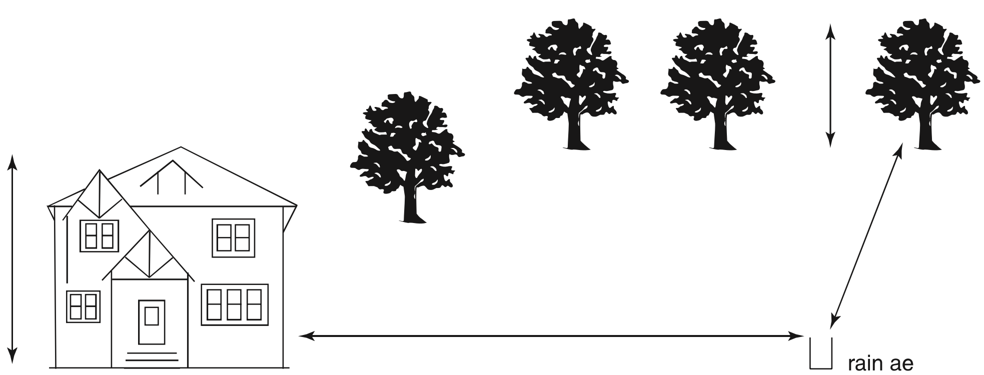


Figure 2.9 Siting of a rain gauge away from obstructions.

Figure 2.10 The insides of a tipping-bucket rain gauge. The 'buckets' are the small white, triangular reservoirs. These are balanced and when full they tip over bringing the black arm past the other stationary arm. In doing so a small electrical current is passed to a data logger.

arm until it too tips the balance arm, sending another electrical impulse to the data logger. In this way a near continuous measurement of rainfall with time can be obtained.

It is important that the correct size of tipping bucket is used for the prevailing conditions. If the buckets are too small then a very heavy rainfall event will cause them to fill too quickly and water be lost through overspill while the mechanism tips. If the buckets are too large then a small rainfall event may not cause the cross arm to tip and the subsequent rainfall event will appear larger than it actually was. The tipping-bucket rain gauge shown in Figure 2.10 has an equivalent depth of 0.2 mm of rain which works well for field studies in south-east England.

Snowfall measurement

The measurement of snowfall has similar problems to those presented by rainfall, but they are often more extreme. There are two methods used for measuring snowfall: using a gauge like a rain gauge; or measuring the depth that is present on the ground. Both of these methods have very large errors associated with them, predominantly caused by the way that snow falls through the atmosphere and is deposited on the gauge or ground. Most, although not all, snowflakes are more easily transported by the wind than raindrops. When the snow reaches the ground it is easily blown around in a secondary manner (drifting). This can be contrasted to liquid water where, upon reaching the ground, it is either absorbed by the soil or moves across the surface. Rainfall is very rarely picked up by the wind again and redistributed in the manner that drifting snow is. For the snow gauge this presents problems that are analogous to rain splash. For the depth gauge the problem is due to uneven distribution of the snow surface: it is likely to be deeper in certain situations.

Rain gauge modification to include snowfall

One modification that needs to be made to a standard rain gauge in order to collect snowfall is a heated rim so that any snow falling on the gauge melts to be collected as liquid water. Failure to have a heated rim may mean that the snow builds up on the gauge surface until it overflows. Providing a heated rim is no simple logistic exercise as it necessitates a power source (difficult in remote areas) and the removal of collected water well away from the heat source to minimise evaporation losses.

A second modification is to raise the gauge well above ground level so that as snow builds up the gauge is still above this surface. Unfortunately the raising of the gauge leads to an increase in the turbulence errors described for rain gauges. For this reason it is normal to have wind deflectors or shields surrounding the gauge.

Snow depth

The simplest method of measuring snow depth is the use of a core sampler. This takes a core of snow, recording its depth at the same time, that can then be melted to derive the water equivalent depth. It is this that is of importance to a hydrologist. The major difficulties of a core sample are that it is a non-continuous reading (similar to daily rainfall measurement), and the position of coring may be critical (because of snow drifting).

A second method of measuring snow depth is to use a **snow pillow**. This is a method for measuring snow accumulation, a form of water storage, hence it is described in Chapter 4 (p. 75).

Forest rainfall measurement

The most common method of assessing the amount of canopy interception loss is to measure the precipitation above and below a canopy and assume that the difference is from interception. Stated in this way it sounds a relatively simple task but in reality it is fraught with difficulty and error. Durocher (1990) provides a good example of the instrumentation necessary to measure canopy interception, in this case for a deciduous woodland plot.

Above-canopy precipitation

To measure above-canopy precipitation a rain gauge may be placed on a tower above the canopy. The usual rain gauge errors apply here, but especially the exposure to the wind. As described in Chapter 3, the top of a forest canopy tends to be rough and is very good for allowing turbulent transfer of evaporated water. The turbulent air is not so good for measuring rainfall! An additional problem for any long-term study is that the canopy is not static; the tower needs to be raised every year so that it remains above the growing canopy.

One way around the tower problem is to place a rain gauge in a nearby clearing and assume that what falls there is the same amount as directly above the canopy nearby. This is often perfectly reasonable to assume, particularly for long-term totals, but care must be taken to ensure the clearing is large enough to avoid obstruction from nearby trees (see Figure 2.9).

Throughfall

Throughfall is the hardest part of the forest hydrological cycle to measure. This is because a forest canopy is normally variable in density and therefore throughfall is spatially heterogeneous. One common method is to place numerous rain gauges on the forest floor in a random manner. If you are interested in a long-term study then it is reasonable to keep the throughfall gauges in fixed positions. However, if the study is investigating individual storm events then it is considered best practice to move the gauges to new random positions between storm events. In this way the throughfall catch should not be influenced by gauge position. To derive an average throughfall figure it is necessary to come up with a spatial average in the same manner as for areal rainfall estimates (see below).

To overcome the difficulty of a small sampling area (rain gauge) measuring something notoriously variable (throughfall), some investigators have used either troughs or plastic sheeting. Troughs collect over a greater area and have proved to be very effective (see Figure 2.11). Plastic sheeting is the ultimate way of collecting throughfall over a large area, but has several inherent difficulties. The first is purely logistical in that it is difficult to install and maintain, particularly to make sure there are no rips. The second is that by having an impervious layer above the ground there is no, or very little, water entering the soil. This might not be a problem for a short-term study but is over the longer term, especially if the investigator is interested in the total

Figure 2.11 Throughfall troughs sitting beneath a pine tree canopy. This collects rain falling through the canopy over the area of the trough. It is sloping so that water flows to a collection point.

water budget. It may also place the trees under stress through lack of water, thus leading to an altered canopy.

Stemflow

The normal method of measuring stemflow is to place collars around a tree trunk that capture all the water flowing down the trunk. On trees with smooth bark this may be relatively simple but is very difficult on rough bark such as found on many conifers. It is important that the collars are sealed to the tree so that no water can flow underneath and that they are large enough to hold all the water flowing down the trunk. The collars should be sloped to one side so that the water can be collected or measured in a tipping-bucket rain gauge. Maintenance of the collars is very important as they easily clog up or become appropriate resting places for forest fauna such as slugs!

MOVING FROM POINT MEASUREMENT TO SPATIALLY DISTRIBUTED ESTIMATION

The measurement techniques described here have all concentrated on measuring rainfall at a precise location (or at least over an extremely small area). In reality the hydrologist needs to know how much precipitation has fallen over a far larger area, usually a catchment. To move from point measurements to a spatially distributed estimation it is necessary to employ some form of spatial averaging. The spatial averaging must attempt to account for an uneven spread of rain gauges in the catchment and the various factors that we know influence rainfall distribution (e.g. altitude, aspect and slope). A simple arithmetic mean will only work where a catchment is sampled by uniformly spaced rain gauges and where there is no diversity in topography. If these conditions were ever truly met then it is unlikely that there would be more than one rain gauge sampling the area. Hence it is very rare to use a simple averaging technique.

There are different statistical techniques that address the spatial distribution issues, and with the growth in use of **Geographic Information Systems (GIS)** it is often a relatively trivial matter to do the calculation. As with any computational task it is important to have a good knowledge of how the technique works so that any shortcomings are fully understood. Three techniques are described here: **Thiessen's polygons**, the **hypsometric method** and the **isohyetal method**. These methods are explored further in a Case Study on p. 31.

Thiessen's polygons

Thiessen was an American engineer working around the start of the twentieth century who devised a simple method of overcoming an uneven distribution of rain gauges within a catchment (very much the norm). Essentially Thiessen's polygons attach a representative area to each rain gauge. The size of the representative area (a polygon) is based on how close each gauge is to the others surrounding it.

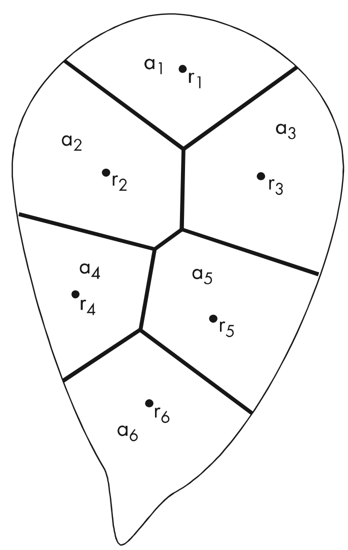


Figure 2.12 Thiessen's polygons for a series of rain gauges (r_i) within an imaginary catchment. The area of each polygon is denoted as a_i. Locations of rain gauges are indicated by bullet points.

Each polygon is drawn on a map; the boundaries of the polygons are equidistant from each gauge and drawn at a right angle (orthogonal) to an imaginary line between two gauges (see Figure 2.12). Once the polygons have been drawn the area of each polygon surrounding a rain gauge is found. The spatially averaged rainfall (R) is calculated using formula 2.1:

$$R = \sum_{i=1}^{n} \frac{r_i a_i}{A} \tag{2.1}$$

where r_i is the rainfall at gauge i, a_i is the area of the polygon surrounding rain gauge i, and A is the total catchment area.

The **areal rainfall** value using Thiessen's polygons is a weighted mean, with the weighting being based upon the size of each representative area (polygon). This technique is only truly valid where the topography is uniform within each polygon so that it can be safely assumed that the rainfall distribution is uniform within the polygon. This would suggest that it can only work where the rain gauges are located initially with this technique in mind (i.e. *a priori*).

Hypsometric method

Since it is well known that rainfall is positively influenced by altitude (i.e. the higher the altitude the greater the rainfall) it is reasonable to assume that knowledge of the catchment elevation can be brought to bear on the spatially distributed rainfall estimation problem. The simplest indicator of the catchment elevation is the hypsometric (or hypsographic) curve. This is a graph showing the proportion of a catchment above or below a certain elevation. The values for the curve can be derived from maps using a planimeter or using a digital elevation model (DEM) in a GIS.

The hypsometric method of calculating spatially distributed rainfall then calculates a weighted average based on the proportion of the catchment between two elevations and the measured rainfall between those elevations (equation 2.2).

$$R = \sum_{j=1}^{m} r_j p_j \tag{2.2}$$

where r_j is the average rainfall between two contour intervals and p_j is the proportion of the total catchment area between those contours (derived from the hypsometric curve). The r_j value may be an average of several rain gauges where there is more than one at a certain contour interval. This is illustrated in Figure 2.13 where the shaded area (a_3) has two gauges within it. In this case the r_j value will be an average of r_4 and r_5.

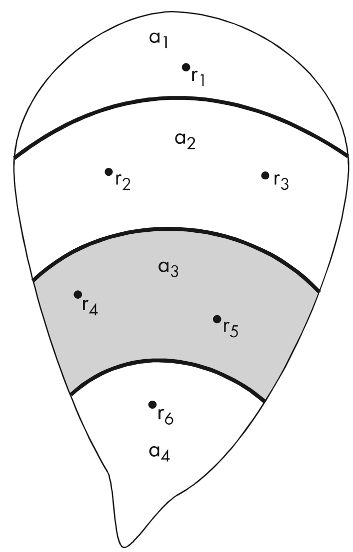


Figure 2.13 Calculation of areal rainfall using the hypsometric method. The shaded region is between two contours. In this case the rainfall is an average between the two gauges within the shaded area. Locations of rain gauges are indicated by bullet points.

Intuitively this is producing representative areas for one or more gauges based on contours and spacing, rather than just on the latter as for Thiessen's polygons. There is an inherent assumption that elevation is the only topographic parameter affecting rainfall distribution (i.e. slope and aspect are ignored). It also assumes that the relationship between altitude and rainfall is linear, which is not always the case and warrants exploration before using this technique.

Isohyetal and other smoothed surface techniques

Where there is a large number of gauges within a catchment the most obvious weighting to apply on a mean is based on measured rainfall distribution rather than on surrogate measures as described above. In this case a map of the catchment rainfall distribution can be drawn by interpolating between the rainfall values, creating a smoothed rainfall surface. The traditional isohyetal method involved drawing isohyets (lines of equal rainfall) on the map and calculating the area between each isohyet. The spatial average could then be calculated by equation 2.3

$$R = \sum_{i=1}^{n} \frac{r_i a_i}{A} \tag{2.3}$$

where a_i is the area between each isohyet and r_i is the average rainfall between the isohyets. This technique is analogous to Figure 2.13, except in this case the contours will be of rainfall rather than elevation.

With the advent of GIS the interpolating and drawing of isohyets can be done relatively easily, although there are several different ways of carrying out the interpolation. The interpolation subdivides the catchment into small grid cells and then assigns a rainfall value for each grid cell (this is the smoothed rainfall surface). The simplest method of interpolation is to use a nearest neighbour analysis, where the assigned rainfall value for a grid square is proportional to the nearest rain gauges. A more complicated technique is to use **kriging**, where the interpolated value for each cell is derived with knowledge on how closely related the nearby gauges are to each other in terms of their co-variance. A fuller explanation of these techniques is provided by Bailey and Gatrell (1995).

An additional piece of information that can be gained from interpolated rainfall surfaces is the likely rainfall at a particular point within the catchment. This may be more useful information than total rainfall over an area, particularly when needed for numerical simulation of hydrological processes.

The difficulty in moving from the point measurement to a spatially distributed average is a prime example of the problem of scale that besets hydrology. The scale of measurement (i.e. the rain gauge surface area) is far smaller than the catchment area that is frequently our concern. Is it feasible to simply scale up our measurement from point sources to the overall catchment? Or should there be some form of scaling factor to acknowledge the large discrepancy? There is no easy answer to these questions and they are the type of problem that research in hydrology will be investigating in the twenty-first century.

RAINFALL INTENSITY AND STORM DURATION

Water depth is not the only rainfall measure of interest in hydrology; also of importance is the **rainfall intensity** and **storm duration**. These are simple to obtain from an analysis of rainfall records using frequency analysis. The rainfall needs to be recorded at a short time interval (i.e. an hour or less) to provide meaningful data.

Figure 2.15 shows the rainfall intensity for a rain gauge at Bradwell-on-Sea, Essex, UK. It is evident from the diagram that the majority of rain falls at very low intensity: 0.4 mm per hour is considered as light rain. This may be misleading as the rain gauge recorded rainfall every hour and the small amount of rain may have fallen during a shorter period than an hour i.e. a higher intensity but lasting for less than an hour. During the period of measurement there were recorded rainfall intensities greater than 4.4 mm/hr (maximum 6.8 mm/hr) but they were so few as to not show up on the histogram scale used in Figure 2.15. This may be a reflection of only two years of records being analysed, which

Case study

RAINFALL DISTRIBUTION IN A SMALL STUDY CATCHMENT

It is well known that large variations in rainfall occur over quite a small spatial scale. Despite this, there are not many studies that have looked at this problem in detail. One study that has investigated spatial variability in rainfall was carried out in the Plynlimon research catchments in mid-Wales (Clarke *et al*., 1973). In setting up a hydrological monitoring network in the Wye and Severn catchments thirty-eight rain gauges were installed to try and characterise the rainfall variation. The rainfall network had eighteen rain gauges in the Severn catchment (total area 8.7 km^2) and twenty gauges in the Wye (10.55 km^2).

The monthly data for a period between April 1971 and March 1973 were analysed to calculate areal average rainfall using contrasting methods. The results from this can be seen in Figure 2.14. The most startling feature of Figure 2.14 is the lack of difference in calculated values and that they follow no regular pattern. At times the arithmetic mean is greater than the others while in other months it is less. When the total rainfall for the two-year period is looked at, the Thiessen's calculation is 0.3 per cent less than the arithmetic mean, while the isohyetal method is 0.4 per cent less.

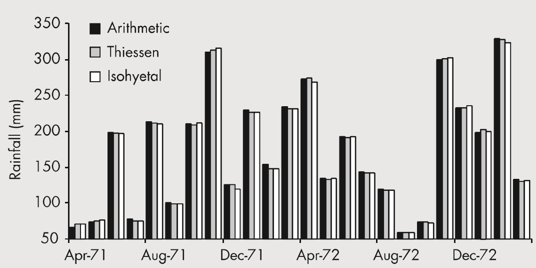


Figure 2.14 Areal mean rainfall (monthly) for the Wye catchment, calculated using three different methods.

Source: Data from Clarke *et al*. (1973)

When the data were analysed to see how many rain gauges would be required to characterise the rainfall distribution fully it was found that the number varied with the time period of rainfall and the season being measured. When monthly data were looked at there was more variability in winter rainfall than summer. For both winter and summer it showed that anything less than five rain gauges (for the Wye) increased the variance markedly.

A more detailed statistical analysis of hourly mean rainfall showed a far greater number of gauges were required. Four gauges would give an accuracy in areal estimate of around 50 per cent, while a 90 per cent accuracy would require 100 gauges (Clarke *et al*., 1973: 62).

The conclusions that can be drawn from the study of Clarke *et al*. (1973) are of great concern to hydrology. It would appear that even for a small catchment a large number of rain gauges are required to try and estimate rainfall values properly. This confirms the statement made at the start of this chapter: although rainfall is relatively straightforward to measure it is notoriously difficult to measure accurately and, to compound the problem, is also extremely variable within a catchment area.

introduces an extremely important concept in hydrology: the **frequency–magnitude** relationship. With rainfall (and runoff – see Chapters 5 and 6) the larger the rainfall event the less frequent we would expect it to be. This is not a linear relationship; as illustrated in Figure 2.15 the curve declines in a non-linear fashion. If we think of the relative frequency as a probability then we can say

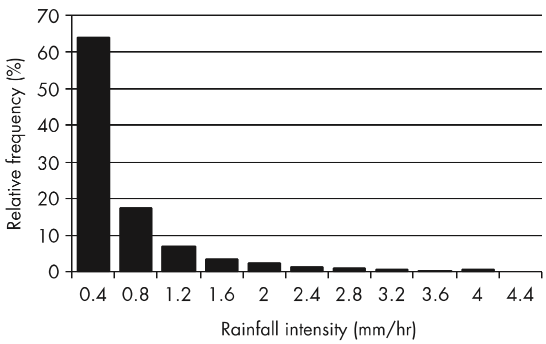


Figure 2.15 Rainfall intensity curve for Bradwell-on-Sea, Essex, UK. Data are hourly recorded rainfall from April 1995 to April 1997.

that the chances of having a low rainfall event are very high: a low magnitude–high frequency event. Conversely the chances of having a rainfall intensity greater than 5 mm/hr are very low (but not impossible): a high magnitude–low frequency event.

In Figure 2.16 the storm duration records for two different sites are compared. The Bradwell-on-Sea site has the majority of its rain events lasting one hour or less. In contrast the Ahoskie site has only 20 per cent of its storms lasting one hour or less but many more than Bradwell-on-Sea that last four hours or more. When the UK site rainfall and intensity curves are looked at together (i.e. Figures 2.15 and 2.16) it can be stated that Bradwell-on-Sea experiences a predominance of low intensity, short duration rainfall events and very few long duration, high intensity storms. This type of information is extremely useful to a hydrologist investigating the likely runoff response that might be expected for the rainfall regime.

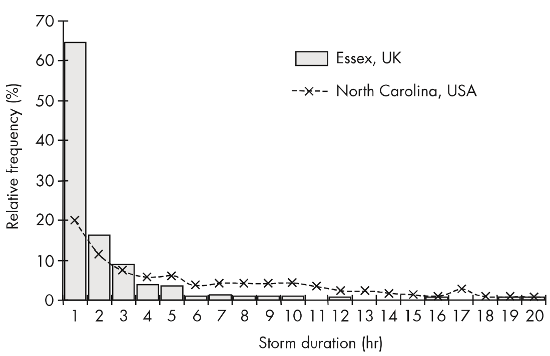


Figure 2.16 Storm duration curves. The bars are for the same data set as Figure 2.12 and the broken line for Ahoskie, North Carolina.

Source: Ahoskie data are redrawn from Wanielista (1990)

SURROGATE MEASURES FOR ESTIMATING RAINFALL

The difficulties in calculation of a spatially distributed precipitation value from point measurements make the direct estimation of areal precipitation an attractive proposition. There are two techniques that make some claim to achieving this: radar, and **satellite remote sensing**. These approaches have many similarities, but they differ fundamentally in the direction of measurement. Radar looks from the earth up into the atmosphere and tries to estimate the amount of precipitation falling over an area. Satellite remote sensing looks from space down towards the earth surface and attempts to estimate the amount of precipitation falling over an area.

Radar

The main use of ground-based radar is in weather forecasting where it is used to track the movement of rain clouds and fronts across the earth's surface. This in itself is interesting but does not provide the hydrological requirement of estimating how much rain is falling over an area.

There are several techniques used for radar, although they are all based on similar principles. Radar is an acronym (*r*adio *d*etection *a*nd *r*anging). A wave of electromagnetic energy is emitted from a unit on the ground and the amount of wave reflection and return time is recorded. The more water there is in a cloud the more electromagnetic energy is reflected back to the ground and detected by the radar unit. The quicker the reflected wave reaches back to ground the closer the cloud is to the surface. The most difficult part of this technique is in finding the best wavelength of electromagnetic radiation to emit and detect. It is important that the electromagnetic wave is reflected by liquid water in the cloud, but not atmospheric gases and/or changing densities of the atmosphere. A considerable amount of research effort has gone into trying to find the best wavelengths for ground-based radar to use. The solution appears to be that it is somewhere in the microwave band (commonly c-band), but that the exact wavelength depends on the individual situation being studied (Cluckie and Collier, 1991).

Studies have shown a good correlation between reflected electromagnetic waves and rainfall intensity. Therefore, this can be thought of as a surrogate measure for estimating rainfall. If an accurate estimate of rainfall intensity is required then a relationship has to be derived using several calibrating rain gauges. Herein lies a major problem: with this type of technique: there is no universal relationship that can be used to derive rainfall intensity from cloud reflectivity. An individual calibration has to be derived for each site and this may involve several years of measuring point rainfall coincidentally with cloud reflectivity. This is not a cheap option and the cost prohibits its widespread usage, particularly in areas with poor rain gauge coverage.

In Britain the UK Meteorological Office operates a series of fifteen weather radar with a 5 km resolution that provide images every 15 minutes. This is a more intensive coverage than could be expected in most countries. Although portable radar can be used for rainfall estimation, their usage has been limited by the high cost of purchase.

Satellite remote sensing

The atmosphere-down approach of satellite remote sensing is quite different from the ground-up approach of radar – fundamentally because the sensor is looking at the top of a cloud rather than the bottom. It is well established that a cloud most likely to produce rain has an extremely bright and cold top. These are the characteristics that can be observed from space by a satellite sensor. The most common form of satellite sensor is passive (this means it receives radiation from another source, normally the sun, rather than emitting any itself the way radar does) and detects radiation in the visible and infrared wavebands. **LANDSAT**, **SPOT** and **AVHRR** are examples of satellite platforms of this type. By sensing in the visible and infrared part of the electromagnetic spectrum the cloud brightness (visible) and temperature (thermal infrared) can be detected. This so-called 'brightness temperature' can then be related to rainfall intensity via calibration with point rainfall measurements, in a similar fashion to ground-based radar. One of the problems with this approach is that it is sometimes difficult to distinguish between snow reflecting light on the ground and clouds reflecting light in the atmosphere. They have similar brightness temperature values but need to be differentiated so that accurate rainfall assessment can be made.

Another form of satellite sensor that can be used is passive microwave. The earth emits microwaves (at a low level) that can be detected from space. When there is liquid water between the earth's surface and the satellite sensor (i.e. a cloud in the atmosphere) some of the microwaves are absorbed by the water. A satellite sensor can therefore detect the presence of clouds (or other bodies of water on the surface) as a lack of microwaves reaching the sensor. An example of a study using a satellite platform that can detect passive microwaves (SSM/I) is in Todd and Bailey (1995), who used the method to assess rainfall over the United Kingdom. Although there was some success in the method it is at a scale of little use to catchment scale hydrology as the best resolution available is around 10 × 10 km grid sizes.

Satellite remote sensing provides an indirect estimate of precipitation over an area but is still a long way from operational use. Studies have shown that it is an effective tool where there is poor rain gauge coverage (e.g. Kidd *et al.*, 1998), but in countries with high rain gauge density it does not improve estimation of areal precipitation. What is encouraging about the technique is that nearly all the world is covered by satellite imagery so that it can be used in sparsely gauged areas. The new generation of satellite platforms being launched in the early twenty-first century will have multiple sensors on them so it is feasible that they will be measuring visible, infrared and microwave wavebands simultaneously. This will improve the accuracy considerably but it must be borne in mind that it is an indirect measure of precipitation and will still require calibration to a rain gauge set.

PRECIPITATION IN THE CONTEXT OF WATER QUANTITY AND QUALITY

Precipitation, as the principal input into a catchment water balance, has a major part to play in water quantity and quality. By and large it is the spatial and temporal distribution of precipitation that drives the spatial and temporal distribution of available water. Rainfall intensity frequently controls the amount of runoff during a storm event (see Chapter 5) and the distribution of rain through the year controls the need for irrigation in an agricultural sytem.

The exception to this is in large river basins where the immediate rainfall distribution may have little bearing on the water flowing down the adjacent river. A good example is the Colorado River which flows through areas of extremely low rainfall in Utah and Arizona. The lack of rainfall in these areas has little bearing on the quantity of water in the Colorado River, it is governed by the precipitation (both rain and snow) falling in the Rocky Mountains well to the north-east.

The influence of precipitation on water quantity directly affects water quality through dilution. Where water quantity is high there is more water available to dilute any contaminants entering a river or groundwater system. It does not follow that high water quantity equates with high water quality but it has the potential to do so.

Precipitation also has a direct influence on water quality through scavenging of airborne pollutants which are then dissolved by the rain. The complex nature of a forest topography means that trees act as recipient surfaces for airborne pollutants. As rain falls onto the tree, salts that have formed on leaves and branches may be dissolved by the water, making the stemflow and throughfall pollutant-rich. This has been observed in field studies, particularly near the edge of tree stands (Neal *et al.*, 1991).

The best known example of pollutant scavenging is **acid rain**. This is where precipitation in areas polluted by industrial smokestacks dissolves gases and absorbs particles that lower the acidity of the rain. Naturally rain is slightly acidic with pH between 5 and 6, due to the dissolving of carbon dioxide to form a weak carbonic acid. The burning of fossil fuels adds nitrogen oxides and sulphur oxides to the atmosphere, both of which are easily dissolved to form weak nitric and sulphuric acids. The burning of coal is particularly bad through the amount of sulphur dioxide produced, but any combustion will produce nitrogen oxides by the combination of nitrogen and oxygen (both already in the atmosphere) at high temperatures. In areas of the Eastern United States and Scandinavia rainfall has been recorded with pH values as low as 3 (similar to vinegar). In some situations this makes very little difference to overall water quality as the soil may have enough acid-buffering capability to absorb the acid rain. This is particularly true for limestone areas where the soil is naturally alkaline. However many soils do not have this buffering capacity due to their underlying geology (e.g. granite areas in the north-east of North America). In this situation the streams become acidic and this has an extremely detrimental effect on the aquatic fauna. The major reason for the impact on fish life is the dissolved aluminium that the acidic water carries; this interferes with the operation of gills and the fish effectively drown.

It is worth noting that the dissolving of nitrous oxide can have a positive benefit to plant life through the addition to the soils of nitrate which promotes plant growth (see Chapter 7). To give some scale to the impact of large atmospheric nitrogen inputs, Löye-Pilot *et al*. (1990) estimate that atmospheric nitrogen input into the Mediterranean Sea is of the same order of magnitude as the riverine input. It is estimated that 25 per cent of nitrogen inputs to the Baltic Sea come from the atmosphere (BSEP, 2005).

SUMMARY

Precipitation is the main input of water within a catchment water balance. Its measurement is fraught with difficulties and any small errors will be magnified enormously at the catchment scale. It is also highly variable in time and space. Despite these difficulties precipitation is one of the most regularly measured hydrological variables, and good rainfall records are available for many regions in the world. A forest canopy partitions rainfall into components that move at different rates towards the soil surface. The nature of the canopy (leaf size distribution and leaf area index) determines the impact that a canopy has on the water balance equation.

Analysis of rainfall can be carried out with respect to trying to find a spatial average or looking at the intensity and duration of storm events. Although there are techniques available for estimating precipitation their accuracy is not such that it is superior to a good network of precipitation gauges.

ESSAY QUESTIONS

1 **Describe the different factors affecting the spatial distribution of precipitation at differing scales.**
2 **How are errors in the measurement of rainfall and snowfall minimised?**
3 **Compare and contrast different techniques for obtaining a spatially averaged precipitation value (including surrogate measures).**
4 **Why is scale such an important issue in the analysis of precipitation in hydrology?**
5 **Describe a field experiment (including equipment) required to measure the water balance beneath a forest canopy.**
6 **Discuss the role of spatial scale in assessing the importance of a forest canopy within a watershed.**

FURTHER READING

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