5

RUNOFF

LEARNING OBJECTIVES

When you have finished reading this chapter you should have:

- An understanding of the process of runoff leading to **channel flow**.
- A knowledge of the techniques for measuring streamflow and runoff directly.
- A knowledge of techniques used to estimate streamflow.

The amount of water within a river or stream is of great interest to hydrologists. It represents the end-product of all the other processes in the hydrological cycle and is where the largest amount of effort has gone into analysis of historical records. The methods of analysis are covered in Chapter 6; this chapter deals with the mechanisms that lead to water entering the stream: the runoff mechanisms. *Runoff* is a loose term that covers the movement of water to a channelised stream, after it has reached the ground as precipitation. The movement can occur either on or below the surface and at differing velocities. Once the water reaches a stream it moves towards the oceans in a channelised form, the process referred to as **streamflow** or **riverflow**. Streamflow is expressed as **discharge**: the volume of water over a defined time period. The SI units for discharge are m^3/s (*cumecs*). A continuous record of streamflow is called a **hydrograph** (see Figure 5.1). Although we think of this as continuous measurement it is normally either an averaged flow over a time period or a series of samples (e.g. hourly records).

In Figure 5.1 there are a series of peaks between periods of steady, much lower flows. The hydrograph peaks are referred to as **peakflow**, **stormflow** or even **quickflow**. They are the water in the stream during and immediately after a significant rainfall event. The steady periods between peaks are referred to as **baseflow** or sometimes slowflow (NB this is different from **low flow**; see Chapter 6). The shape of a hydrograph, and in particular the shape of the stormflow peak, is influenced by the storm characteristics (e.g. rainfall intensity and duration)

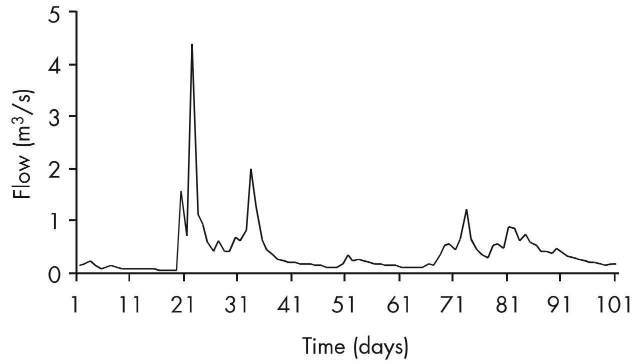

Figure 5.1 A typical hydrograph, taken from the river Wye, Wales for a 100-day period during the autumn of 1995. The values plotted against time are mean daily flow in cumecs.

and many physical characteristics of the upstream catchment. In terms of catchment characteristics the largest influence is exerted by catchment size, but other factors include slope angles, shape of catchment, soil type, vegetation type and percentage cover, degree of urbanisation and the antecedent soil moisture.

Figure 5.2 shows the shape of a storm hydrograph in detail. There are several important hydrological terms that can be seen in this diagram. The **rising limb** of the hydrograph is the initial steep part leading up to the highest or peakflow value. The water contributing to this part of the hydrograph is from *channel precipitation* (i.e. rain that falls directly onto the channel) and rapid runoff mechanisms. Some texts claim that channel precipitation shows up as a preliminary blip before the main rising limb.

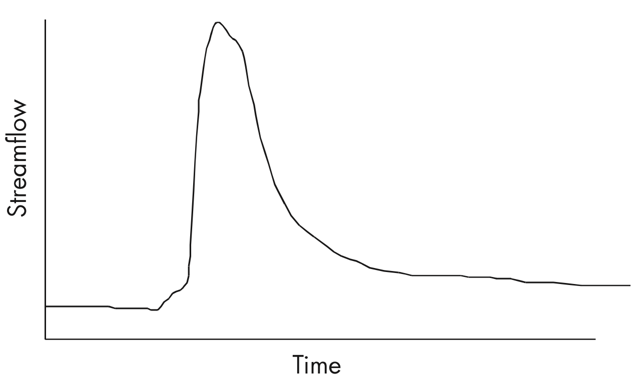

Figure 5.2 Demonstration storm hydrograph.

In reality this is very rarely observed, a factor of the complicated nature of storm runoff processes. The **recession limb** of the hydrograph is after the peak and is characterised by a long, slow decrease in streamflow until the baseflow is reached again. The recession limb is attenuated by two factors: storm water arriving at the mouth of a catchment from the furthest parts, and the arrival of water that has moved as underground flow at a slower rate than the streamflow.

Exactly how water moves from precipitation reaching the ground surface to channelised streamflow is one of the most intriguing hydrological questions, and one that cannot be answered easily. Much research effort in the past hundred years has gone into understanding runoff mechanisms; considerable advances have been made, but there are still many unanswered questions. The following section describes how it is believed runoff occurs, but there are many different scales at which these mechanisms are evident and they do not occur everywhere.

RUNOFF MECHANISMS

Figure 5.3 is an attempt to represent the different runoff processes that can be observed at the hillslope scale. **Overland flow** (Q_o) is the water which runs across the surface of the land before reaching the stream. In the subsurface, throughflow (Q_t) (some authors refer to this as **lateral flow**) occurs in the shallow subsurface, predominantly, although not always, in the unsaturated zone. Groundwater flow (Q_G) is in the deeper saturated zone. All of these are runoff mechanisms that contribute to streamflow. The relative importance of each is dependent on the catchment under study and the rainfall characteristics during a storm.

Overland flow

Some of the earliest research work on how overland flow occurs was undertaken by Robert Horton (1875–1945). In a classic paper from 1933, Horton

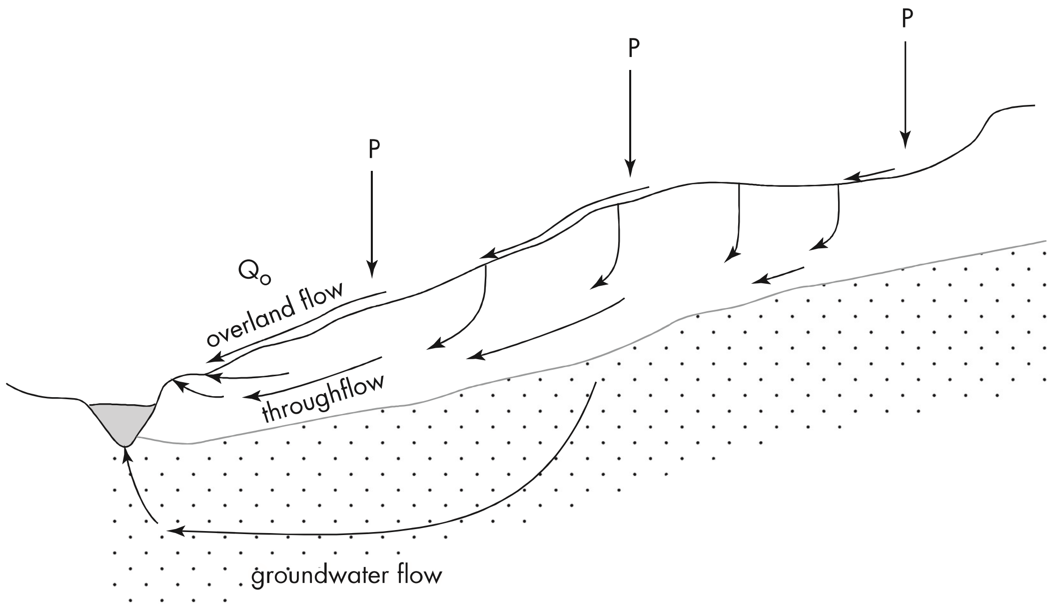

Figure 5.3 Hillslope runoff processes. See text for explanation of terms.

Source: Adapted from Dunne (1978)

hypothesised that overland flow occurred when the rainfall rate was higher than the infiltration rate of a soil. Horton went on to suggest that under these circumstances the excess rainfall collected on the surface before travelling towards the stream as a thin sheet of water moving across the surface. Under this hypothesis it is the infiltration rate of a soil that acts as a controlling barrier or partitioning device. Where the infiltration capacity of a soil is low, overland flow occurs readily. This type of overland flow is referred to as **infiltration excess overland flow** or **Hortonian overland flow** although as Beven (2004) points out, Horton himself referred to it as 'rainfall excess'.

Horton's ideas were extremely important in hydrology as they represented the first serious attempt to understand the processes of storm runoff that lead to a storm hydrograph. Unfortunately, when people started to measure infiltration capacities of soils they invariably found that they were far higher than most normal rainfall rates. This is illustrated in Table 5.1 where some typical infiltration capacities and rainfall rates are shown. Other measurements confirm high infiltration capacities for soils, e.g. Selby (1970) reports infiltration capacities of between 60 and 600 mm/hour on short grazed pasture on yellow-brown pumice soils in the central North Island of New Zealand. The values were higher for ungrazed grass and under trees and are generally higher than the measured rainfall intensities (Selby, 1970).

In addition to the infiltration capacity information, it is extremely rare to find a thin sheet of water moving over the surface during a storm event. It was observations such as those by Hursh (1944) and others that led to a general revision of Horton's hypothesis. Betson (1964) proposed the idea that within a catchment there are only limited areas that contribute overland flow to a storm hydrograph. This is referred to as the **partial areas concept**. Betson did not challenge the role of infiltration excess overland flow as the primary source of stormflow, but did challenge the idea of overland flow occurring as a thin sheet of water throughout a catchment.

Hewlett and Hibbert (1967) were the first to suggest that there might be another mechanism of overland flow occurring. This was particularly based

Table 5.1 Some typical infiltration rates compared to rainfall intensities

Soil and vegetation	*Infiltration rate (mm/hr)*	*Rainfall type*	*Rainfall intensity (mm/hr)*
Forested loam	100–200	Thunderstorm	50–100
Loam pasture	10–70	Heavy rain	5–20
Sand	3–15	Moderate rain	0.5–5
Bare clay	0–4	Light rain	0.5

Source: From Burt (1987)

on the observations from the eastern USA: that during a storm it was common to find all the rainfall infiltrating a soil. Hewlett and Hibbert (1967) hypothesised that during a rainfall event all the water infiltrated the surface. This hypothesis was confirmed by a comprehensive field study by Dunne and Black (1970).

Through a mixture of infiltration and throughflow the water table would rise until in some places it reached the surface. At this stage overland flow occurs as a mixture of return flow (i.e. water that has been beneath the ground but returns to the surface) and rainfall falling on saturated areas. This type of overland flow is referred to as **saturated overland flow**. Hewlett and Hibbert (1967) suggested that the water table was closest to the surface, and therefore likely to rise to the surface quickest, adjacent to stream channels and at the base of slopes. Their ideas on stormflow were that the areas contributing water to the hydrograph peaks were the saturated zones, and that these vary from storm to storm. In effect the saturated areas immediately adjacent to the stream act as extended channel networks. This is referred to as the **variable source areas concept**. This goes a step beyond the ideas of Betson (1964) as the catchment has a partial areas response but the response area is dynamic; i.e. variable in space and time.

So who was right: Horton, or Hewlett and Hibbert? The answer is that both were. Table 5.2 provides a summary of the ideas for storm runoff generation described here. It is now accepted that saturated overland flow (Hewlett and Hibbert) is the dominant overland flow mechanism in humid, mid-latitude areas. It is also accepted that the variable source areas concept is the most valid description of stormflow processes. However, where the infiltration capacity of a soil is low or the rainfall rates are high, Hortonian overland flow does occur. In Table 5.1 it can be seen that there are times when rainfall intensities will exceed infiltration rates under natural circumstances. In arid and semi-arid zones it is not uncommon to find extremely high rainfall rates (fed by convective storms) that can lead to infiltration excess overland flow and rapid flood events; this is called **flash flooding**.

Table 5.2 A summary of the ideas on how stormflow is generated in a catchment

	Horton	*Betson*	*Hewlett and Hibbert*
Infiltration	Controls overland flow	Controls overland flow	All rainfall infiltrates
Overland flow mechanism	Infiltration excess	Infiltration excess	Saturated overland flow
Contributing area	Uniform throughout the catchment	Restricted to certain areas of the catchment	Contributing area is variable in time and space

Examples of low infiltration rates can be found with compacted soils (e.g. from vehicle movements in an agricultural field), on roads and paved areas, on heavily crusted soils and what are referred to as **hydrophobic soils**.

Basher and Ross (2001) reported infiltration capacities of 400 mm/hour in market gardens in the North Island of New Zealand and that these rates increased during the growing season to as high as 900 mm/hour. However, Basher and Ross (2001) also showed a decline in infiltration capacity to as low as 0.5 mm/hour in wheel tracks at the same site.

Hydrophobic soils have a peculiar ability to swell rapidly on contact with water, which can create an impermeable barrier at the soil surface to infiltrating water, leading to Hortonian overland flow. The cause of hydrophobicity in soils has been linked into several factors including the presence of micro-rhizal fungi and swelling clays such as allophane (Doerr *et al.*, 2007). Hydrophobicity is a temporary soil property; continued contact with water will increase the infiltration rate. For example Clothier *et al.* (2000) showed how a yellow brown earth/loam changed from an initial infiltration capacity of 2 mm/hour to 14 mm/hour as the soil hydrophobicity breaks down.

In Hewlett and Hibbert's (1967) original hypothesis it was suggested that contributing saturated areas would be immediately adjacent to stream channels. Subsequent work by the likes of Dunne and Black (1970), Anderson and Burt (1978) and others has identified other areas in a catchment prone to inducing saturated overland flow. These include hillslope hollows, slope concavities (in section) and where there is a thinning of the soil overlying an impermeable base. In these situations any throughflow is likely to return to the surface as the volume of soil receiving it is not large enough for the amount of water entering it. This can be commonly observed in the field where wet and boggy areas can be found at the base of slopes and at valley heads (hillslope hollows).

Subsurface flow

Under the variable source areas concept there are places within a catchment that contribute overland flow to the storm hydrograph. When we total up the amount of water found in a storm hydrograph it is difficult to believe that it has all come from overland flow, especially when this is confined to a relatively small part of the catchment (i.e. variable source areas concept). The manner in which the recession limb of a hydrograph attenuates the stormflow suggests that it may be derived from a slower movement of water: subsurface flow. In addition to this, tracer studies looking at where the water has been before entering the stream as stormflow have found that a large amount of the storm hydrograph consists of 'old water' (e.g. Martinec *et al.*, 1974; Fritz *et al.*, 1976). This old water has been sitting in the soil, or as fully saturated groundwater, for a considerable length of time and yet enters the stream during a storm event. There have been several theories put forward to try and explain these findings, almost all involving throughflow and groundwater.

Throughflow is a general term used to describe the movement of water through the unsaturated zone; normally this is the soil matrix. Once water infiltrates the soil surface it continues to move, either through the soil matrix or along preferential flow paths (referred to as lateral or preferential flow). The rate of soil water movement through a saturated soil matrix is described by Darcy's law (see Chapter 4) and the Richards approximation of Darcy's law when below saturation. Under normal, vertical, infiltration conditions the hydraulic gradient has a value of –1 and the saturated hydraulic conductivity is the infiltration capacity. Once the soil is saturated the movement of water is not only vertical. With a sloping water table on a hillsope, water moves down slope. However, the movement of water through a saturated soil matrix is not rapid, e.g. Kelliher and Scotter (1992) report a K_{sat} value of 13 mm/hour for a fine sandy loam. In order for throughflow to contribute to storm runoff there must be another mechanism (other than matrix flow) operating.

One of the first theories put forward concerning the contribution of throughflow to a storm hydrograph was by Horton and Hawkins (1965) (this Horton was a different person from the proposer of Hortonian overland flow). They proposed the mechanism of *translatory* or *piston flow* to explain the rapid movement of water from the subsurface to the stream. They suggested that as water enters the top of a soil column it displaces the water at the bottom of the column (i.e. old water), and the displaced water enters the stream. The analogy is drawn to a piston where pressure at the top of the piston chamber leads to a release of pressure at the bottom. The release of water to the stream can be modelled as a pressure wave rather than tracking individual particles of water. Piston flow has been observed in laboratory experiments with soil columns (e.g. Germann and Beven, 1981).

At first glance the simple piston analogy seems unlike a real-life situation since a hillslope is not bounded by impermeable sides in the same way as a piston chamber. However the theory is not as far-fetched as it may seem, as the addition of rainfall infiltrating across a complete hillslope is analogous to pressure being applied from above and in this case the boundaries are upslope (i.e. gravity) and the bedrock below. Brammer and McDonnell (1996) suggest that this may be a mechanism for the rapid movement of water along the bedrock and soil interface on the steep catchment of Maimai in New Zealand. In this case it is the hydraulic gradient created by an addition of water to the bottom of the soil column, already close to saturated, that forces water along the base where hydraulic conductivities are higher.

Ward (1984) draws the analogy of a thatched roof to describe the contribution of subsurface flow to a stream (based on the ideas of Zaslavsky and Sinai, 1981). When straw is placed on a sloping roof it is very efficient at moving water to the bottom of the roof (the guttering being analogous to a stream) without visible overland flow. This is due to the preferential flow direction along, rather than between, sloping straws. Measurements of hillslope soil properties do show a higher hydraulic conductivity in the downslope rather than vertical direction. This would account for a movement of water downslope as throughflow, but it is still bound up in the soil matrix and reasonably slow.

There is considerable debate on the role of **macropores** in the rapid movement of water through the soil matrix. Macropores are larger pores within a soil matrix, typically with a diameter greater than 3 mm. They may be caused by soils cracking, worms burrowing or other biotic activities. The main interest in them from a hydrologic point of view is that they provide a rapid conduit for the movement of water through a soil. The main area of contention concerning macropores is whether they form continuous networks allowing rapid movement of water down a slope or not. There have been studies suggesting macropores as a major mechanism contributing water to stormflow (e.g. Mosley, 1979, 1982; Wilson *et al*., 1990), but it is difficult to detect whether these are from small areas on a hillslope or continuous throughout. Jones (1981) and Tanaka (1992) summarise the role of pipe networks (a form of continuous macropores) in hillslope hydrology. Where found, pipe networks have considerable effect on the subsurface hydrology but they are not a common occurrence in the field situation.

The role of macropores in runoff generation is unclear. Although they are capable of allowing rapid movement of water towards a stream channel there is little evidence of networks of macropores moving large quantities of water in a continuous fashion. Where macropores are known to have a significant role is in the rapid movement of water to the saturated layer (e.g. Heppell *et al*., 1999) which may in turn lead to piston flow (McGlynn *et al*., 2002).

Groundwater contribution to stormflow

Another possible explanation for the presence of old water in a storm hydrograph is that it comes from the saturated zone (groundwater) rather than from throughflow. This is contrary to conventional hydrological wisdom which suggests that groundwater contributes to baseflow but not to the stormflow component of a hydrograph. Although a groundwater contribution to stormflow had been suggested before, it was not until Sklash and Farvolden (1979) provided a theoretical mechanism for this to occur that the idea was seriously considered. They proposed the capillary fringe hypothesis to explain the groundwater ridge, a rise in the water table immediately adjacent to a stream (as observed by Ragan, 1968). Sklash and Farvolden (1979) suggested that the addition of a small amount of infiltrating rainfall to the zone immediately adjacent to a stream causes the soil water to move from an unsaturated state (i.e. under tension) to a saturated state (i.e. a positive pore pressure expelling water). As explained in Chapter 4, the relationship between soil water content and soil water tension is non-linear. The addition of a small amount of water can cause a rapid change in soil moisture status from unsaturated to saturated. This provides the groundwater ridge which:

> not only provides the early increased impetus for the displacement of the groundwater already in a discharge position, but it also results in an increase in the size of the groundwater discharge area which is essential in producing large groundwater contributions to the stream.
>
> (Sklash and Farvolden, 1979: 65)

An important point to stress from the capillary fringe hypothesis is that the groundwater ridge is developing well before any throughflow may have been received from the contributing hillslope areas. These ideas confirm the variable source areas concept and provide a mechanism for a significant old water contribution to storm hydrographs. Field studies such as that by McDonnell (1990) have observed groundwater ridging to a limited extent, although it is not an easy task as often the instrument response time is too slow to detect the rapid change in pore pressure properly.

Case study

THE MAIMAI RUNOFF GENERATION STUDIES

Figure 5.4 Maimai catchments in South Island of New Zealand. At the time of photograph (1970s) five catchments had been logged and are about to be replanted with *Pinus radiata*.

The Maimai catchment study (near Reefton on West Coast of the South Island of New Zealand) was established in 1974 for research into the effects of logging native beech forest (*Nothofagus*) and replanting with different non-indigenous species (Figure 5.4). The installation of hydrological measuring equipment and the fact that rainfall and stormflow are frequently observed made it an ideal place for studying stormflow generation mechanisms in depth. The knowledge gained from detailed hydrological process studies at Maimai have played a major part in shaping thinking on stormflow generation mechanisms.

The Maimai catchment is characterised by short, steep slopes (approximately 300 m with angles of around 35°), covered in thick vegetation, with incised channels and very small valley bottoms. Annual rainfall is approximately 2,600 mm with an average of 156 rain days a year, and stormflow makes up 65 per cent of the total streamflow (Rowe *et al*., 1994; Pearce *et al*., 1986).

Mosley (1979, 1982) used Maimai to investigate the role of macropores as conduits for rapid movement of rainfall to the stream. Observations of macropore flow rates using cut soil faces and dye tracers suggested that rainfall could travel down the short steep hillslopes at Maimai in less than 3 hours (i.e. within the time frame of a storm event). Subsequent chemical and isotopic analysis of streamflow, rainfall and water exiting the cut soil pit faces showed that the majority of measured streamflow was 'old' water, suggesting that rapid, extensive macropore flow was not the main mechanism for stormflow generation (Pearce *et al*., 1986).

McDonnell (1990) investigated this further, in particular looking at possible groundwater ridging (Sklash and Farvolden, 1979) as a mechanism for large amounts of old water as saturated overland flow. Although this could be observed at Maimai, the amount of water held near the stream prior to an event was not large enough to account for all of the old water, which suggested that another mechanism (e.g. piston flow) might be working (McDonnell, 1990).

McGlynn *et al*. (2002) present a summary conceptual diagram of runoff mechanisms on Maimai hillslopes that combines many of the features described above (see Figure 5.5). In this model there is rapid infiltration of water through macropores to reach the bedrock. At this stage a form of piston flow occurs as the saturated zone at the base of the soil mantle is confined by the soil matrix above it. At the bedrock interface there may be a network of macropores or else the same situation of a confined aquifer in that the soil matrix above has a much lower hydraulic conductivity. Water is then pushed out at the bottom due to the pressure from new water arriving directly at the bedrock interface. There is also a mixing of the new water with old water sitting in bedrock hollows, creating a rapid movement of old water into the stream during storm events.

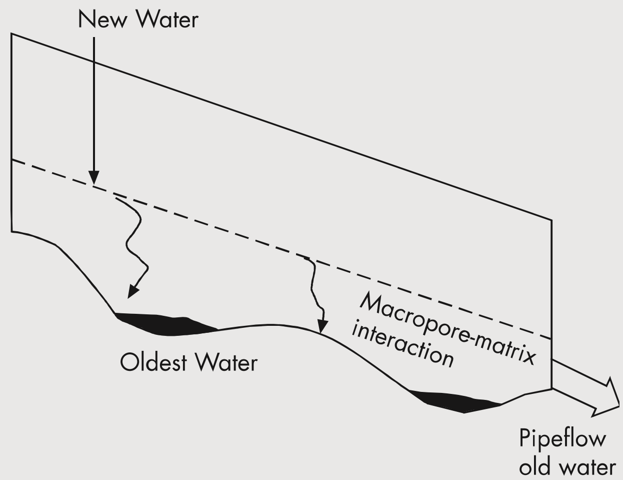

Figure 5.5 Summary hypothesis for hillslope stormflow mechanisms at Maimai. Rapid movement of water occurs through rapid infiltration to the bedrock interface and then a form of piston flow along this interface.

Source: adapted from McGlynn *et al*. (2002)

How relevant are the Maimai stormflow generation studies?

The studies that have taken place at Maimai have been extremely important in influencing hydrological thinking around the world. However, an argument can be made that the conditions at Maimai are far from generally applicable elsewhere. The main study catchment (M8) has short, steep slopes and is in an area of high, and frequent, rainfall. The soils are extremely porous (infiltration rates in excess of 1,600 mm/hour have been measured) and remain within 10 per cent of saturation for most of the year (Mosley, 1979). These conditions are not common and it would be difficult to generalise the concepts beyond Maimai. One of the really important concepts that Maimai has shown is that under conditions ideal for stormflow generation the mechanisms are still extremely complex and spatially variable. This is true wherever in the world the study is taking place.

Summary of storm runoff mechanisms

The mechanisms that lead to a storm hydrograph are extremely complex and still not fully understood. Although this would appear to be a major failing in a science that is concerned with the movement of water over and beneath the surface, it is also an acknowledgement of the extreme diversity found in nature. In general there is a reasonable understanding of possible storm runoff mechanisms but it is not possible to apply this universally. In some field situations the role of throughflow and piston flow are important, in others not; likewise for groundwater contributions, overland flow and pipeflow. The challenge for modern hydrology is to identify quickly the dominant mechanisms for a particular hillslope or catchment so that the understanding of the hydrological processes in that situation can be used to aid management of the catchment.

The processes of storm runoff generation described here are mostly observable at the hillslope scale. At the catchment scale (and particularly for large river basins) the timing of peak flow (and consequently the shape of the storm hydrograph) is influenced more by the channel drainage network and the precipitation characteristics of a storm than by the mechanisms of runoff. This is a good example of the problem of scale described in Chapter 1. At the small hillslope scale storm runoff generation mechanisms are important, but they become considerably less so at the much larger catchment scale.

Baseflow

In sharp contrast to the storm runoff debate, there is general consensus that the major source of baseflow is groundwater – and to a lesser extent throughflow. This is water that has infiltrated the soil surface and moved towards the saturated zone. Once in the saturated zone it moves downslope, often towards a stream. A stream or lake is often thought to occur where the regional water table intersects the surface, although this may not always be the case. In Chapter 4 the relationship between groundwater and streamflow has been explained (see Figure 4.9). However in general it can be said that baseflow is provided by the slow seepage of water from groundwater into streams. This will not necessarily be visible (e.g. springs) but can occur over a length of streambank and bed and is only detectable through repeated measurement of streamflow down a reach.

Channel flow

Once water reaches the stream it will flow through a channel network to the main river. The controls over the rate of flow of water in a channel are to do with the volume of water present, the gradient of the channel, and the resistance to flow experienced at the channel bed. This relationship is described in uniform flow formulae such as the Chezy and Manning equations (see p. 92). The resistance to flow is governed by the character of the bed surface. Boulders and vegetation will create a large amount of friction, slowing the water down as it passes over the bed.

In many areas of the world the channel network is highly variable in time and space. Small channels may be ephemeral and in arid regions will frequently only flow during flood events. The resistance to flow under these circumstances is complicated by the infiltration that will be occurring at the water front and bed surface. The first flush of water will infiltrate at a much higher rate as it fills the available pore space in the soil/rock at the bed surface. This will remove water from the stream and also slow the water front down as it creates a greater friction surface. Under a continual flow regime the infiltration from the stream to ground will depend on the hydraulic gradient and the infiltration capacity.

MEASURING STREAMFLOW

The techniques and research into the measurement of streamflow are referred to as **hydrometry**. Streamflow measurement can be subdivided into two important subsections: instantaneous and continuous techniques.

Instantaneous streamflow measurement

Velocity–area method

Streamflow or discharge is a volume of water per unit of time. The standard units for measurement of discharge are m^3/s (cubic metres per second or *cumecs*). If we rewrite the units of discharge we can think of them as a water velocity (m/s) passing through a cross-sectional area (m^2). Therefore:

$$m^3/s = m/s \times m^2 \quad (5.1)$$

The **velocity–area method** measures the stream velocity, the stream cross-sectional area and multiplies the two together. In practice this is carried out by dividing the stream into small sections and measuring the velocity of flow going through each cross-sectional area and applying equation 5.2.

$$Q = v_1 a_1 + v_2 a_2 + \ldots v_i a_i \quad (5.2)$$

where Q is the streamflow or discharge (m^3/s), v is the velocity measured in each trapezoidal cross-sectional area (see Figure 5.6), and a is the area of the trapezoid (usually estimated as the average of two depths divided by the width between).

The number of cross-sectional areas that are used in a discharge measurement depend upon the width and smoothness of stream bed. If the bed is particularly rough it is necessary to use more cross-sectional areas so that the estimates are as close to reality as possible (note the discrepancy between the broken and solid lines in Figure 5.6).

The water velocity measurement is usually taken with a flow meter (Figure 5.7). This is a form of propeller inserted into the stream which records the number of propeller turns with time. This reading can be easily converted into a stream velocity using the calibration equation supplied with the flow meter.

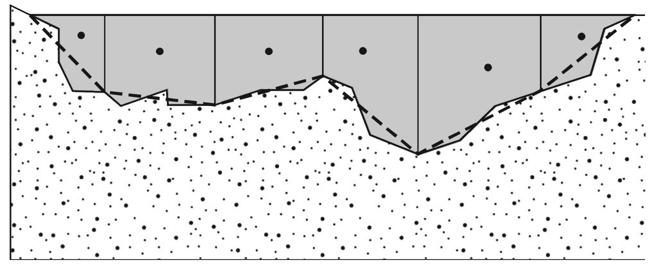

Figure 5.6 The velocity–area method of streamflow measurement. The black circles indicate the position of velocity readings. Dashed lines represent the triangular or trapezoidal cross-sectional area through which the velocity is measured.

Figure 5.7 Flow gauging a small stream.

In the velocity–area method it is necessary to assume that the velocity measurement is representative of all the velocities throughout the cross-sectional area. It is not normally possible to take multiple measurements so an allowance has to be made for the fact that the water travels faster along the surface than nearer the stream bed. This difference in velocity is due to friction exerted on the water as it passes over the stream bed, slowing it down. As a general rule of thumb the sampling depth should be 60 per cent of the stream depth – that is, in a stream that is 1 m deep the sampling point should be 0.6 m below the surface or 0.4 m above the bed. In a deep river it is good practice to take two measurements (one at 20 per cent and the other at 80 per cent of depth) and average the two.

Where there is no velocity meter available it may be possible to make a very rough estimate of stream velocity using a float in the stream (i.e. the time it takes to cover a measured distance). When using this method allowance must be made for the fact that the float is travelling on the surface of the stream at a faster rate than water closer to the stream bed.

The velocity–area method is an effective technique for measuring streamflow in small rivers, but its reliability is heavily dependent on the sampling strategy. The technique is also less reliable in small, turbid streams with a rough bed (e.g. mountain streams). Under these circumstances other streamflow estimation techniques such as **dilution gauging** may be more applicable (see streamflow estimation section).

Continuous streamflow measurement

The methods of instantaneous streamflow measurement described above only allow a single measurement to be taken at a location. Although this can be repeated at a future date it requires a continuous measurement technique to give the data for a hydrograph. There are three different techniques that can be used for this method: stage discharge relationships, flumes and weirs, and ultrasonic flow gauging.

Stage vs discharge relationship

River **stage** is another term for the water level or height. Where multiple discharge measurements have been taken (i.e. repeat measurements using velocity–area method) it is possible to draw a relationship between river stage and discharge: the so-called **rating curve**. An example of a rating curve is shown in Figure 5.8. This has the advantage of allowing continuous measurement of river stage (a relatively simple task) that can then be equated to the actual discharge. The stage discharge relationship is derived through a series of velocity–area measurements at a particular site while at the same time recording the stage with a stilling well (see Figure 5.9). As can be seen in Figure 5.8, the rating curve is non-linear, a reflection of the river bank profile. As the river fills up between banks it takes a greater volume of water to cause a change in stage than at low levels.

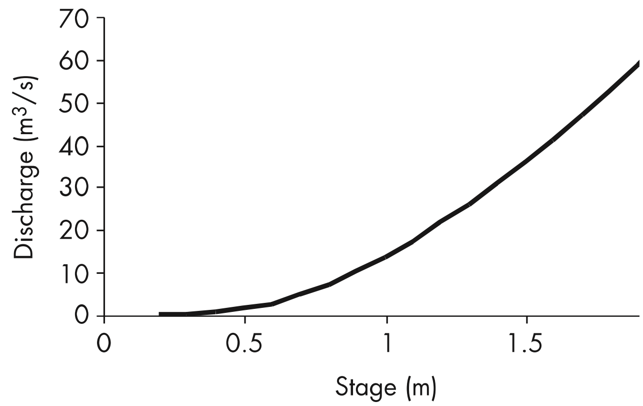

Figure 5.8 A rating curve for the river North Esk in Scotland based on stage (height) and discharge measurements from 1963–1990.

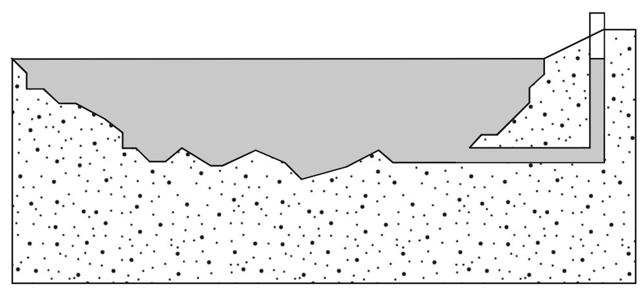

Figure 5.9 Stilling well to provide a continuous measurement of river stage (height). The height of water is measured in the well immediately adjacent to the river.

An accurate stage vs discharge relationship is dependent on frequent and accurate measurement of river discharge, and a static river bed profile. If the river bed profile changes (e.g. during a large flood event it may get scoured out or new sediment deposited), the stage vs discharge relationship will change and the historic relationship will no longer be valid. This assumption of a static river bed profile can sometimes be problematic, leading to the installation of a concrete structure (e.g. flume or weir) to maintain stability.

One of the difficulties with the stage vs discharge relationship is that the requirement of frequent measurements of river discharge lead to many measurements taken during periods of low and medium flow but very few during flood events. This is for the double reason that: floods are infrequent and unlikely to be measured under a regular monitoring programme; and the danger of streamflow gauging during a flood event. The lack of data at the extreme end of the stage vs discharge curve may lead to difficulties in interpreting data during peak flows. The error involved in estimating peak discharge from a measured stage vs discharge relationship will be much higher at the high flow end of curve.

When interpreting data derived from the stage discharge vs relationship it is important that the hydrologist bears in mind that it is stream stage that is being measured and from this stream discharge is inferred (i.e. it is not a direct measurement of stream discharge).

Flumes and weirs

Flumes and weirs utilise the stage–discharge relationship described above but go a step further towards providing a continuous record of river discharge. If we think of stream discharge as consisting of a river velocity flowing through a cross-sectional area (as in the velocity profile method) then it is possible to isolate both of these terms separately. This is what flumes and weirs, or *stream gauging structures*, attempt to do.

The first part to isolate is the stream velocity. The way to do this is to slow a stream down (or, in some rare cases, speed a stream up) so that it flows with constant velocity through a known cross-sectional area. The critical point is that in designing a flume or weir the river flows at the same velocity (or at least a known velocity) through the gauging structure irrespective of how high the river level is. Although this seems counter-intuitive (rivers normally flow faster during flood events) it is achievable if there is an area prior to the gauging structure that slows the river down: a stilling pond.

The second part of using a gauging structure is to isolate a cross-sectional area. To achieve this a rigid structure is imposed upon the stream so that it always flows through a known cross-sectional area. In this way a simple measure of stream height through the gauging structure will give the cross-sectional area. Stream height is normally derived through a stilling well, as described in Figure 5.9, except in this case there is a regular cross-sectional area.

Once the velocity and cross-sectional area are kept fixed the rating curve can be derived through a mixture of experiment and hydraulic theory. These relationships are normally power equations dependent on the shape of cross-sectional area used in the flume or weir. There is an international standard for manufacture and maintenance of weirs (ISO 1438) that sets out theoretical ratings curves for different types of structures. The general formula for a V notch weir is shown in equation 5.3.

$$Q = 0.53.\sqrt{2g}.C.\tan\left(\frac{\theta}{2}\right).h^{2.5} \tag{5.3}$$

where Q is discharge (m^3/s); g is the acceleration due to gravity (9.81 m/s^2); C is coefficient of discharge (see Figure 5.10); θ is the angle of V-notch (°); h is the height of water or stage (m). The coefficient of discharge can be estimated from figure 5.10 for a certain angle of V-notch. For a 90° V-notch the coefficient of discharge is 0.578 and the rating equation becomes:

$$Q = 1.366h^{2.5} \tag{5.4}$$

There is a similar type of equation for rectangular weirs, based on the width of the rectangular exit and another version of the coefficient of discharge relationship.

The shape of cross-sectional area is an important consideration in the design of flumes and weirs. The shape of permanent structure that the river flows

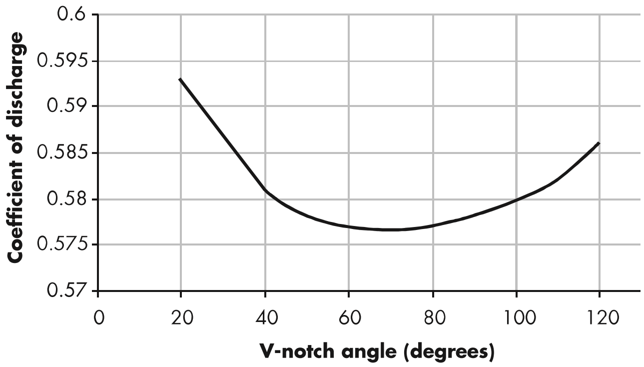

Figure 5.10 Coefficient of discharge for V-notch weirs (ISO 1438).

Figure 5.11 A V-notch weir. The water level in the pond behind the weir is recorded continuously.

through is determined by the flow regime of the river and the requirements for the streamflow data. A common shape used is based on the V-structure (see Figure 5.11). The reason for this is that when river levels are low, a small change in river flow will correspond to a significant change in stage (measured in the stilling well). This sensitivity to low flows makes data from this type of flume or weir particularly suitable for studying low flow hydrology. It is important that under high flow conditions the river does not overtop the flume or weir structure. The V shape is convenient for this also because as discharge increases the cross-sectional area flowed through increases in a non-linear fashion. The angle of the V-notch will vary depending on the size of stream being measured and the sensitivity required (90° and 120° V-notch weirs are both commonly used).

One of the difficulties in maintaining gauging structures is that by slowing the river down in a stilling pond any sediment being carried by the water may be deposited (see **Hjulstrom curve** in Chapter 7), which in time will fill the stilling pond and lessen its usefulness. Because of this the stilling pond needs to be dredged regularly, particularly in a high energy environment such as mountain streams. To overcome this difficulty there is a design of trapezoidal flume that speeds the stream up rather than slows it down (see Figure 5.12). The stream is forced to go down a steep section immediately prior to the gauging structure. In this way any sediment is flushed out of the weir, removing the need for

Figure 5.12 A trapezoidal flume. The stream passes through the flume and the water level at the base of the flume is recorded continuously.

regular dredging. This is really only possible for small streams as the power of large rivers at high velocities would place enormous strains on the gauging structure.

The difference between flumes and weirs

Although flumes and weirs perform the same function – measuring stream discharge in a continuous fashion – they are not the same. In a weir the water is forced to drop over a structure (the weir – Figure 5.11) in the fashion of a small waterfall. In a flume (Figure 5.12) the water passes through the structure without having a waterfall at the end.

Ultrasonic flow gauging

Recent technological developments have led to the introduction of a method of measuring stream discharge using the properties of sound wave propagation in water. The method actually measures water velocity, but where the stream bed cross-sectional area is known (and constant) the instrumentation can be left in place and combined with measurements of stage to provide a continuous measurement of river discharge. There are two types of **ultrasonic flow gauges** that work in slightly different ways.

The first method measures the time taken for an ultrasonic wave emitted from a transmitter to reach a receiver on the other side of a river. The faster the water speed, the greater the deflection of the wave path and the longer it will take to cross the river. Sound travels at approximately 1,500 m/s in water (dependent on water purity and depth) so the instrumentation used in this type of flow gauging needs to be extremely precise and be able to measure in nanoseconds. This type of flow gauging can be installed as a permanent device but needs a width of river greater than 5 m and becomes unreliable with a high level of suspended solids.

The second method utilises the Doppler effect to measure the speed of particles being carried by the stream. At an extremely simple level this is a measurement of the wavelength of ultrasonic waves that bounce off suspended particles – the faster the particle the shorter the wavelength. This type of instrument works in small streams (less than 5 m width) and requires some suspended matter.

Measuring hillslope runoff

The measurement of runoff may be required to assess the relative contribution of different hillslope runoff processes; i.e. throughflow, overland flow, etc. There are no standard methods for the measurement of runoff processes; different researchers use different techniques according to the field conditions expected and personal preference.

Overland flow

The amount of water flowing over the soil surface can be measured using collection troughs at the bottom of hillslopes or runoff plots. A runoff plot is an area of hillslope with definite upslope and side boundaries so that you can be sure all the overland flow is generated from within each plot. The upslope and side boundaries can be constructed by driving metal plates into the soil and leaving them protruding above the surface. It is normal to use several runoff plots to characterise overland flow on a slope as it varies considerably in time and space. This spatial and temporal variation may be overcome with the use of a rainfall simulator.

Throughflow

Measurement of throughflow is fraught with difficulty. The only way to measure it is with throughflow troughs dug into the soil at the appropriate height. The problem with this is that in digging, the soil profile is disturbed and consequently the flow characteristics change. It is usual to insert troughs into a soil face that has been excavated and then refill the hole. This may still overestimate throughflow as the reconstituted soil in front of the troughs may encourage flow towards it as an area allowing rapid flow.

ESTIMATING STREAMFLOW

In the past thirty years probably the greatest effort in hydrological research has gone into creating numerical models to simulate streamflow. With time these have developed into models simulating all the processes in the hydrological cycle so that far more than just streamflow can be estimated. However, it is often streamflow that is seen as the end-product of a model, a reflection of the importance streamflow has as a hydrological parameter. These models are described in Chapter 6; this section concentrates on direct estimates of streamflow.

Physical or geomorphological estimation techniques

The geomorphological approach to river systems utilises the idea that the river channel is in equilibrium with the flow regime. This suggests that measures of the channel (e.g. depth/width ratio, **wetted perimeter**, height to **bankfull discharge**) can be used to estimate the streamflows in both a historical and contemporary sense. Wharton (1995) provides a review of these different techniques. This is not a method that can be used to estimate the discharge in a river at one particular time, but it can be used to estimate parameters such as the mean annual **flood**. Important parameters to consider are the stream diameter, wetted perimeter and average depth. This is particularly for the area of channel that fills up during a small flooding event: so-called bankfull discharge.

It is possible to estimate the average velocity of a river stretch using a kinematic wave equation such as Manning's (equation 5.5).

$$v = \frac{k.R^{2/3}.\sqrt{s}}{n} \tag{5.5}$$

where v is velocity (m/s); k is a constant depending on which units of measurement are being used (1 for SI units, 1.49 for Imperial); R is the **hydraulic radius** (m); s is the slope (m/m); and n is the Manning roughness coefficient. Hydraulic radius is the wetted perimeter of a river divided by the cross-sectional area. In very wide channels this can be approximated as mean depth (Goudie *et al*., 1994). The Manning roughness coefficient is estimated from knowledge of the channel characteristics (e.g. vegetation and bed characteristics) in a similar manner to Chezy's roughness coefficient in Table 5.3. Tables of Manning roughness coefficient can be found in Richards (1982), Maidment (1992), Goudie *et al*. (1994), and in other fluvial geomorphological texts.

Dilution gauging

Dilution gauging works on the principle that the more water there is in a river the more it will dilute a solute added into the river. There is a well-

Table 5.3 Chezy roughness coefficients for some typical streams

Type of channel	*Chezy roughness coefficient for a hydraulic radius of 1 m*
Artificial concrete channel	71
Excavated gravel channel	40
Clean regular natural channel <30 m wide	33
Natural channel with some weeds or stones <30 m wide	29
Natural channel with sluggish weedy pools <30 m wide	14
Mountain streams with boulders	20
Streams larger than 30 m wide	40

Source: Adapted from Richards (1982)

established relationship between the amount of the tracer found naturally in the stream (C_o), the concentration of tracer put into the river (C_t), the concentration of tracer measured downstream after mixing (C_d), and the stream discharge (Q). The type of tracer used is dependent on the equipment available; the main point is that it must be detectable in solution and non-harmful to the aquatic flora and fauna. A simple tracer that is often used is a solution of table salt (NaCl), a conductivity meter being employed to detect the salt solution.

There are two different ways of carrying out dilution gauging that use slightly different equations. The first puts a known volume of tracer into the river and measures the concentration of the 'slug' of tracer as it passes by the measurement point. This is referred to as gulp dilution gauging. The equation for calculating flow by this method is shown in equation 5.6.

$$Q = \frac{C_t V}{\sum \left(C_d - C_o\right) \Delta t} \tag{5.6}$$

where Q is the unknown streamflow, C is the concentration of tracer either in the slug (t), downstream (d), or background in the stream (o); Δt is the time interval. The denominator of this equation is the sum of measured concentrations of tracer downstream.

The second method uses a continuous injection of tracer into the river and measures the concentration downstream. The continuous injection method is better than the slug injection method as it measures the concentration over a greater length of time, however it requires a large volume of the tracer. Using the formula listed below the stream discharge can be calculated using equation 5.7.

$$Q = q \frac{C_t - C_d}{C_d - C_0} \tag{5.7}$$

where q is the flow rate of the injected tracer (i.e. injection rate) and all other terms are as for the gulp injection method.

Probably the most difficult part of dilution gauging is calculating the distance downstream between where the tracer is injected and the river concentration measuring point (the mixing distance). This can be estimated using equation 5.8.

$$L = 0.13 C_z \left(\frac{0.7 C_z + w}{g} \right) \left(\frac{w^2}{d} \right) \tag{5.8}$$

where L = mixing distance (m)
C_Z = Chezy's **roughness coefficient** (see Table 5.3)
w = average stream width (m)
g = gravity constant (≈9.8 m/s^2)
d = average depth of flow (m)

FLOODS

The term *flood* is difficult to define except in the most general of terms. In a river a flood is normally considered to be an inundation of land adjacent to a river caused by a period of abnormally large discharge or encroachment by the sea (see cover photograph, Figure 5.13, and Plate 6), but even this definition is fraught with inaccuracy. Flooding may occur from sources other than rivers (e.g. the sea and lakes), and 'abnormal' is difficult to pin down, particularly within a timeframe. Floods come to our attention through the amount of damage that they cause and for this reason they are often rated on a cost basis rather than on hydrological criteria. Hydrological and monetary assessments of flooding often differ markedly because the economic valuation is highly dependent on location. If the area of land inundated by a flooding river is in an expensive region with large infrastructure then the cost will be considerably higher than, say, for agricultural land. Two examples of large-scale floods during the 1990s illustrate this point. In 1998 floods in China caused an estimated US$20 billion of damage with over 15 million people being displaced and 3,000 lives lost (Smith, 2001). This flood was on a similar scale to one that occurred in the same region during 1954. A much larger flood (in a hydrological sense)

Figure 5.13 Images of flood inundation in Fiji, 2007.

in the Mississippi and Missouri rivers during 1993 resulted in a similar economic valuation of loss (US$15–20 billion) but only 48 lives were lost (USCE, 1996). The flood was the highest in the hydrological record and had an average recurrence interval of between 100 and 500 years (USCE, 1996). The difference in lives lost and relative economic loss (for size of flood) is a reflection of the differing response to the flood in two economically contrasting countries.

As described in Chapter 2 for precipitation, flooding is another example where the *frequency–magnitude relationship* is important. Small flood events happen relatively frequently whereas the really large floods occur rarely but cause the most damage. The methods for interpreting river flows that may be used for flood assessment are discussed in Chapter 6. They provide some form of objective flood size assessment, but their value is highly dependent on the amount of data available.

Floods are a frequently occurring event around the world. At the time of preparing of this chapter (June and July 2007) there were eleven large flood events reported in the news media (see Table 5.4). These floods were caused by varying amounts of rainfall, and occurred in different seasons of the year but all caused significant damage and in many cases loss of lives. There are numerous reasons why a river will flood and they almost always relate back to the processes found within the hydrological cycle. The main cause of river floods is when there is too much rainfall for the river to cope with. Other, more special causes of floods are individual events like dam bursts, **jökulhlaups** (ice-dam bursts) or snow melt (see pp. 72–75).

Influences on flood size

The extent and size of the flood can often be related to other contributing factors that increase the effect of high rainfall. Some of these factors are described here but all relate back to concepts introduced in earlier chapters detailing the processes found within the hydrological cycle. Flooding provides an excellent example of the importance of scale, introduced in Chapter 1. Many of the factors discussed here have an influence at the small scale (e.g. hillslopes or small research catchments of less than 10 km^2) but not at the larger overall river catchment scale.

Antecedent soil moisture

The largest influence on the size of a flood, apart from the amount and intensity of rainfall, is the wetness of the soil immediately prior to the rainfall or snow melt occurring. As described on p. 59, the amount of infiltration into a soil and subsequent storm runoff are highly dependent on the degree of saturation in the soil. Almost all major flood events are heavily influenced by the amount of

Table 5.4 Flooding events in news reports during June–July 2007

Location (date)	*Rainfall or flood statistics*	*Effect*
Midlands and Yorkshire, UK (June 2007)	1 location 103 mm of rainfall in 24 hours; many places recorded over 50 mm of rain in 12 hours	30,000+ houses affected; estimated £1.5bn damage
New South Wales, Australia (June 2007)	300 mm rainfall in 3 days	9 lives lost, 5,000 evacuated
Bangladesh (June 2007)	400 mm cumulative rainfall in places	130 lives lost, 10,000 evacuated
India (June 2007)	475 mm rain in 4 days	57 lives lost, 100,000 people evacuated
China (June–July 2007)	300 mm rainfall in 4 days	88 lives lost, 500,000 people evacuated; 56,000 homes destroyed; 91,800 ha crops destroyed
Mid-West, USA (July 2007)	305 mm rainfall in 7 days	17 lives lost
Pakistan (July 2007)	105 mm rainfall in 12 hours; 30 year record	110 lives lost, 200,000 homeless
Southern Japan (July 2007)	200 mm rainfall in 4 days	3,400 evacuated
Sudan (July 2007)	At several sites the Nile was more than a metre higher than in 1988 (a previous record level)	59 lives lost; 30,000 homes evacuated
Northland, New Zealand (July 2007)	270 mm rain total; 213 mm rain in 24 hours; 1 in 150 year storm	23 houses destroyed. Estimated damage \$80M (≈ US\$60M)
Midlands, England, UK (July 2007)	121 mm of rain in 24 hours; wettest May–July since records began in 1766	7 people killed, estimated £2bn damage

rainfall that has occurred prior to the actual flood-causing rainfall.

Deforestation

The effect of trees on runoff has already been described, particularly with respect to water resources. There is also considerable evidence that a large vegetation cover, such as forest, decreases the severity of flooding. There are several reasons for this. The first has already been described, in that trees provide an intercepting layer for rainfall and therefore slow down the rate at which the water reaches the surface. This will lessen the amount of rainfall available for soil moisture and therefore the antecedent soil moisture may be lower under forest than for an adjoining pasture (NB this is not always the case, it is dependent on the time of year). The second factor is that forests often have a high organic matter in the upper soil layers which, as any gardener will tell you, is able to absorb more water. Again this lessens the amount of overland flow, although it may increase the amount of throughflow. Finally, the infiltration rates under forest soils are often higher, leading again to less saturation excess overland flow.

The removal of forests from a catchment area will increase the propensity for a river to flood and also increase the severity of a flood event. Conversely the planting of forests on a catchment area will decrease the frequency and magnitude of flood events. Fahey and Jackson (1997) show that after conversion of native tussock grassland to exotic pine plantations a catchment in New Zealand showed a decrease in the mean flood peaks of 55–65 per cent. Although data of this type look alarming they are almost always taken from measurements at the small research catchment scale. At the larger scale the influence of deforestation is much harder to detect (see Chapter 8).

Urbanisation

Urban areas have a greater extent of impervious surfaces than in most natural landforms. Consequently the amount of infiltration excess (Hortonian) overland flow is high. In addition to this, urban areas are often designed to have a rapid drainage system, taking the overland flow away from its source. This network of gutters and drains frequently leads directly to a river drainage system, delivering more flood water in a faster time. Where extensive urbanisation of a catchment occurs; flood frequency and magnitude increases. Cherkauer (1975) shows a massive increase in flood magnitude for an urban catchment in Wisconsin, USA when compared to a similar rural catchment (see pp. 169–170). Urbanisation is another influence on flooding that is most noticeable at the small scale. This is mostly because the actual percentage area covered by impermeable urban areas in a larger river catchment is still very small in relation to the amount of permeable non-modified surfaces.

River channel alterations

Geomorphologists traditionally view a natural river channel as being in equilibrium with the river flowing within it. This does not mean that a natural river channel never floods, but rather that the channel has adjusted in shape in response to the normal discharge expected to flow through it. When the river channel is altered in some way it can have a detrimental effect on the flood characteristics for the river. In particular, **channelisation** using rigid structures can increase flood risk. Ironically, channelisation is often carried out to lessen flood risk in a particular area. This is frequently achieved, but in doing so water is passed on downstream at a faster rate than normal, increasing the flood risk further downstream. If there is a natural floodplain further downstream this may not be a problem, but if there is not, downstream riparian zones will be at greater risk.

Land drainage

It is common practice in many regions of the world to increase agricultural production through the drainage of 'swamp' areas. During the seventeenth and eighteenth centuries huge areas of the fenlands of East Anglia in England were drained and now are highly productive cereal and horticultural areas. The drainage of these regions provides for rapid removal of any surplus water, i.e. not needed by plants. Drained land will be drier than might be expected naturally, and therefore less storm runoff might be assumed. This is true in small rainfall events but the rapid removal of water through subsurface and surface drainage leads to flood peaks in the river drainage system where normally the water would have been slower to leave the land surface. So, although the drainage of land leads to an overall drying out of the affected area it can also lead to increased flooding through rapid drainage. Again this is scale-related, as described further in Chapter 8.

Climate change

In recent years any flooding event has led to a clamour of calls to explain the event in terms of climatic change. This is not easy to do as climate is naturally so variable. What can be said though is that river channels slowly adjust to changes in flow regime which may in turn be influenced by changes

Case study

MOZAMBIQUE FLOODS OF 2000

During the early months of 2000 world news was dominated by the catastrophic flooding that occurred in southern Africa and Mozambique in particular. The most poignant image from this time was the rescuing of a young mother, Sophia Pedro, with her baby Rosita, born up a tree while they sought refuge from the flood waters. The international media coverage of the devastating flood damage and the rescue operation that followed has ensured that this flood will be remembered for a long time to come. It has given people the world over a reminder that flooding is a hydrological hazard capable of spreading devastation on a huge scale.

The floods of Mozambique were caused by four storms in succession from January through to March 2000. The first three months of the year are the rainy season (or monsoon) for south-eastern Africa and it is usual for flooding to occur, although not to the scale witnessed in 2000. The monsoon started early in southern Mozambique; the rainfall in Maputo was 70 per cent above normal for October–November 1999. This meant that any heavy rainfall later in the rainy season would be more likely to cause a flood.

The first flood occurred during January 2000 when the Incomáti and Maputo rivers (see Figure 5.14) both burst their banks causing widespread disruption. The second flood occurred in early February, as the waters started to recede, except that now Cyclone Connie brought record rainfall to southern Mozambique and northern South Africa. The Limpopo river was as high as ever recorded (previous high was in 1977) and major communication lines were cut. The third flood, 21 February until the end of February, occurred when Cyclone Eline moved inland giving record rainfall in Zimbabwe and northern South Africa, causing record-breaking floods. The Limpopo was 3 m higher than any recorded flood and for the first time in recorded history the Limpopo and Incomáti rivers joined together in a huge inundation. The extent of the flooding can be seen in the satellite images (see Plates 7 and 8). The fourth flood was similar in size to the second and occurred following Cyclone Glória in early March (Christie and Hanlon, 2001).

There is no doubt that the Mozambique floods were large and catastrophic. How large they are, in terms of return periods or average recurrence intervals (see Chapter 6) is difficult to assess. The major difficulty is to do with paucity of streamflow records and problems with measuring flows during flood events. On the Incomáti river the flow records go back to 1937, and this was the largest flood recorded. For the Limpopo there is some data back to the 1890s, and again this was the largest recorded flow event. On the Maputo river to the south the flood levels were slightly lower than a 1984 event. The difficulties in measuring riverflow during large flood events is well illustrated by the failure of many gauging stations to function properly, either through complete inundation or being washed away. Christie and Hanlon (2001) quote an estimate of the flood on the Limpopo having a 100-year average recurrence interval, although this is difficult to verify as most gauges failed. Smithers *et al*. (2001) quote an unpublished report by Van Bladeren and Van der Spuy (2000) suggesting that upstream tributaries of the Incomáti river exceeded the 100-year return period. Smithers *et al*. (2001) provide an analysis of the 1–7 day rainfall for the Sabie catchment (a tributary of the Incomáti) which shows that in places the 200-year return period was exceeded. (NB this is an analysis of rainfall records not riverflow.)

The reasons for the flooding were simple, as they are in most cases: there was too much rainfall for the river systems to cope with the resultant

Figure 5.14 Location of the Incomáti, Limpopo and Maputo rivers in southern Africa.

stormflow. The river catchments were extremely wet (i.e. high antecedent soil moisture values) prior to the extreme rainfall, due to a prolonged and wet monsoon. One possible explanation for the severity of the rainfall is linked in with the ENSO (El Niño: Southern Oscillation) ocean-weather patterns in the Pacific. Christie and Hanlon (2001) present evidence that during a La Niña event (extreme cold temperatures in the western Pacific Ocean) it is common to see higher rainfall totals in Mozambique. However, this is not a strong relationship and certainly could not be used to make predictions. Figure 5.15 shows the monsoon rainfall at Maputo (averaged over two rainy seasons) and associated La Niña events. There may be some link here but it is not immediately obvious, particularly when you consider 1965–66 which had high rainfall despite it being an El Niño event (often associated with drought in southern Africa).

What was unusual about the 2000 floods was that the tropical Eline cyclone (called typhoons or hurricanes elsewhere) moved inland, taking extremely high rainfall to Zimbabwe and northern

South Africa. This is not normal behaviour for this type of storm and in so doing it created large floods in the headwaters of rivers draining into Mozambique. Flood warnings were issued by Zimbabwe and South Africa but the poor state of communications in Mozambique (exacerbated by the previous floods cutting communication lines) meant that they were not available to warn people on the ground. In all 700 people died as a result of the floods and 45,000 people were displaced. It is estimated that it will cost US$450 million to repair damage to the infrastructure in Mozambique (Christie and Hanlon, 2001). This is not the total cost of the flood, which is far higher when loss of income and loss of private property are included. These costs will never be fully known as in many lesser-developed countries the costs are borne by individuals without any form of insurance cover.

In many ways there are no new lessons to learn from the Mozambique floods of 2000. It is well known that adequate warning systems are needed (but expensive) and that people should be restricted from living in flood-prone areas; but this is difficult to achieve in a poor country such as Mozambique. The cause of the flood was a huge amount of rainfall and the severity was influenced by the antecedent wetness of the ground due to a very wet monsoon.

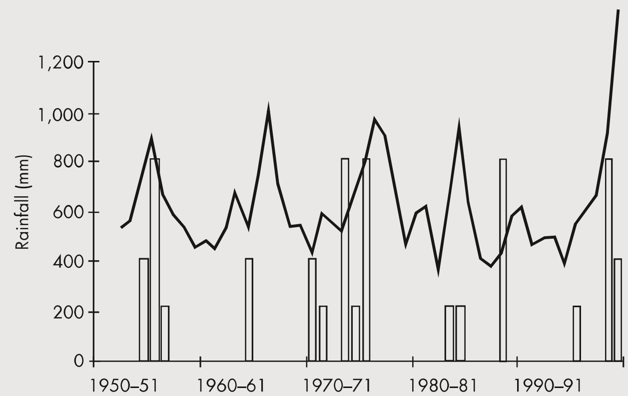

Figure 5.15 Rainfall totals during the rainy season (smoothed with a two-year average) at Maputo airport, with vertical bars indicating the strength of La Niña events (on a scale of three: strong, medium, weak).

Sources: Rainfall data from Christie and Hanlon (2001); La Niña strength from NOAA

in climate. Many studies have suggested that future climate change will involve greater extremes of weather (IPCC, 2007), including more high-intensity rainfall events. This is likely to lead to an increase in flooding, particularly while a channel adjusts to the differing flow regime (if it is allowed to).

RUNOFF IN THE CONTEXT OF WATER QUALITY

The route that water takes between falling as precipitation and reaching a stream has a large influence on water quality. The nutrient level of water is heavily influenced by the length of time water spends in contact with soil. Water that moves quickly into a river (e.g. overland flow) is likely to have a lower nutrient level than water that moves slowly through the soil as throughflow and/or groundwater. However, water that has travelled as overland flow may have a higher level of suspended solids picked up from the surface, so it may appear less pure.

In considering issues of land-use change and water quality, an important consideration is the time taken for water to reach the stream. It is important to realise that groundwater is frequently operating as a pressure wave response to rainfall recharge. Where groundwater responds to a rainfall event by emitting water into a stream it is a pressure wave response, i.e. the water entering the stream is not the same water that infiltrates and causes the response. This means that water entering the stream may be several years (or more) older and unaffected by the current land use change.

SUMMARY

The water flowing down a river is the end-product of precipitation after all the other hydrological processes have been in operation. The sub-processes of overland flow, throughflow and groundwater flow are well understood, although it is not easy to estimate their relative importance for a particular site, particularly during a storm event. The measurement of river flow is relatively straightforward and presents the fewest difficulties in terms of sampling error, although there are limitations, particularly during periods of high flow and floods.

FURTHER READING

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