3

THE FLOW REGIME

The flow in river channels exerts hydraulic forces on the boundary (bed and banks). An important balance exists between the erosive force of the flow (driving force) and the resistance of the boundary to erosion (resisting force). This determines the ability of a river to adjust and modify the morphology of its channel. One of the main factors influencing the erosive power of a given flow is its discharge: the volume of flow passing through a given cross-section in a given time. Discharge varies both spatially and temporally in natural river channels, changing in a downstream direction and fluctuating over time in response to inputs of precipitation. Characteristics of the flow regime of a river include seasonal variations in discharge, the size and frequency of floods and frequency and duration of droughts. The characteristics of the flow regime are determined not only by the climate but also by the physical and land use characteristics of the drainage basin. In this chapter you will learn about:

- The pathways taken by water as it travels to the channel network.
- How the flow in a river responds to inputs of precipitation.
- Seasonal variations in flow that characterise different climatic zones.
- The size (magnitude) and frequency characteristics of floods.
- Flows that are significant in shaping the channel.

FLOW GENERATION

Hydrological pathways

Inputs of water to a drainage basin are the various forms of **precipitation** that fall over its area. These include rain, snow, sleet, hail and dew. If you look at a drainage basin on a map, you will see that river channels cover only a very small part of the total area. This means that most of the water reaching the ground surface must find its way from the hillslopes and into the channel network. A number of pathways are involved (Figure 3.1), and a given 'parcel' of water arriving at a stream channel may have taken any number of them. Not all the incoming precipitation actually makes it to the basin outlet, since a certain percentage is evaporated back to the atmosphere. Water that falls on the leaves of vegetation and artificial structures like buildings is **intercepted**. Some of this water does fall to the ground, as anyone who has sheltered under a tree will know, but much of it is evaporated (water resource managers refer to this output as 'interception loss'). The amount of interception that occurs is dependent on such factors as the extent and type of vegetation (leaf size, structure, density and the arrangement of foliage), wind speed and rainfall intensity (Jones, 1997). Evaporation also takes place from the surface layers of the soil, and from lakes and wetland areas. A related process is **transpiration**,

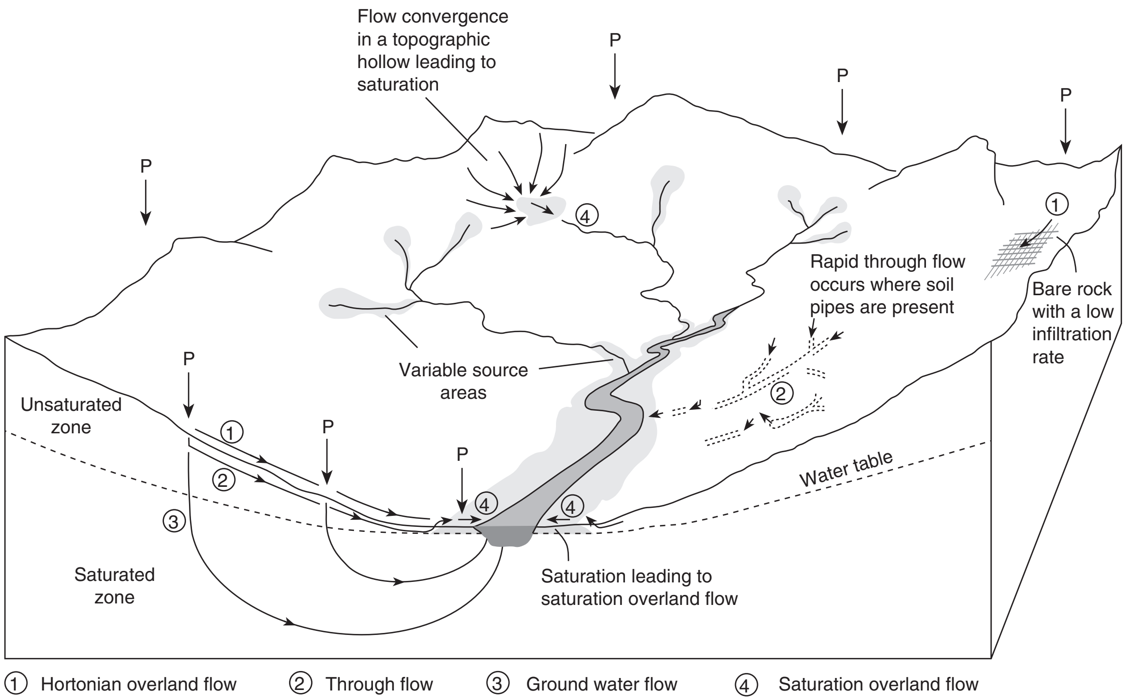

Figure 3.1 Surface and subsurface hydrological pathways.

whereby plants take up water through their roots and evaporate it through pores, called stomata, on the underside of their leaves. This means that water can be 'lost' to the atmosphere from some depth below the surface. Because evaporation and transpiration are difficult to monitor separately, they are usually considered together as the combined process of **evapotranspiration**.

Water reaching the soil surface may either enter the soil by a process called **infiltration** or remain at the surface, moving down-slope as **overland flow** (labelled 1 and 4 in Figure 3.1). Infiltrated water either travels through the soil, parallel to the surface, as **throughflow** (2), or slowly percolates downwards to the saturated zone, travelling as **groundwater flow** (3). Rates of water movement are fastest for overland flow (typically between 50 m and 500 m per hour), between 0.005 m and 0.3 m per hour for throughflow and 0.005 m to 1.5 m per *day* for groundwater flow (Ward and Robinson, 1990). How fast a river rises after rainfall is very much dependent on the relative proportion of water taking faster (surface and near-surface) pathways, and slower (subsurface) pathways. An important control on this is the rate at which water infiltrates into the soil, which is determined by the **infiltration capacity**. This is defined as the maximum rate at which water enters the soil when it is in a given condition (Horton, 1933). This is an important definition: if the rainfall intensity exceeds the infiltration capacity, the soil cannot absorb all the water and there is excess water, which ponds at the surface. This moves down slope as overland flow, more specifically, **Hortonian overland flow** (1). As Horton's definition implies, soil properties control the infiltration capacity; these include soil permeability, the presence of vegetation and plant roots and how much water is already in the soil. Following a dry period, the infiltration capacity is highest at the start of a storm, rapidly decreasing as rainfall continues, until a constant infiltration rate is reached (Horton, 1933).

A second type of overland flow, **saturation overland flow** (4), is generated when the soil is totally saturated and therefore cannot take any more water (Kirkby and Chorley, 1967). Where the water table is relatively shallow, for example in valley bottoms, rainfall can cause it to rise to the ground surface. Where this occurs, a saturated area forms around the channel, increasing in extent as the storm progresses. The saturated area acts as an extension to the channel network, meaning that a significant volume of water is transferred in a short time. These saturated areas are known as **variable source areas** or **dynamic contributing areas** and are highly significant in humid environments, where this is the main way in which storm run-off is generated (Hewlett and Hibbert, 1967; Dunne and Black, 1970). Hortonian overland flow is rarely observed in humid environments, unless the surface has a low infiltration rate, for example where there are outcrops of bare rock or artificially paved surfaces. However, in dryland environments, a combination of factors mean that Hortonian overland flow is the dominant mechanism. Rainfall, when it occurs, typically has a high intensity and exceeds the low infiltration capacity of sparsely vegetated soils (Dunne, 1978). In addition, dryland soils often develop a crust at the surface, which further reduces the infiltration capacity.

Water that infiltrates the soil may travel at various depths below the surface. While it is fairly obvious that water should move downwards under the force of gravity, it might seem counter-intuitive that it also flows through the soil as throughflow. This happens because a preferential flow path is set up. Soil permeability decreases with depth, meaning that the downward movement of water is slowed. During rainfall, this leads to a backing up effect in the more permeable surface layers. This has been likened to the flow of water down a thatched roof: it is easier for the water to move parallel to the slope of the roof, along the stems of the straw, than to move vertically downwards through it (Ward, 1984; Zaslavsky and Sinai, 1981).

Where they exist, **soil pipes** provide a very rapid throughflow mechanism, and rates of flow can be comparable to those in surface channels. Soil pipes are hydraulically formed conduits that can be up to a metre or more in diameter. They are found in a wide range of environments and are sometimes several hundreds of metres in length (Jones, 1997). As such, they can act as an extension to the channel network, allowing the drainage basin to respond rapidly to precipitation inputs (Jones, 1979).

The schematic diagram in Figure 3.1 represents the headwaters of a humid zone river, where the channel intersects the water table surface (it should be noted that aquifers are not always present, for example aquifer development is extremely limited in headwater areas that are underlain by impermeable rocks and characterised by thin soils). In this example, groundwater contributions are made to the channel flow from the underlying aquifer and the channel is called a **gaining stream**. Even during rainless periods, flow will be maintained, as long as the level of the water table does not fall below that of the channel bed. A rather different situation exists for many dryland channels, where the water table may be several metres, or tens of metres, below the surface. In this case, the direction of flow is reversed, as water is lost through the bed and banks of the channel, percolating downwards to recharge the aquifer. This is termed a **losing stream** and, although not exclusive to drylands, many examples are found in these environments. It is not unusual for a river to be gaining and losing flow to groundwater along different parts of its course, while seasonal fluctuations in water table levels can mean that losing streams become gaining streams for part of the year. The Euphrates in Iraq provides a good example, with most of the flow being generated in the headwaters in northern Iraq, Turkey and Syria. Further downstream, at the Hit gauging station (150 km west of Baghdad) the river loses flow to groundwater for much of the year (Wilson, 1990). The loss of flow from a channel, due to downward percolation and high evaporation rates, is referred to as **transmission loss**.

The storm hydrograph and drainage basin response

Flow discharge (also known as ***Q***) is the volume of water passing through a given channel cross-section in a given time. The units of discharge most commonly used are cubic metres per second ($m^3\ s^{-1}$), known as 'cumecs', although for very small flows litres per second may be used. In the United States cubic feet per second, or 'cusecs', are used instead of cumecs. Box 3.1 explains how discharge is monitored.

Box 3.1

THE MEASUREMENT OF STREAMFLOW

The channel discharge is the volume of water flowing through a given channel cross-section in a given time. A number of different methods have been developed to measure discharge. These can be grouped into **instantaneous measurements**, where discharge is measured at a particular point in time, and **continuous measurements** for a record of discharge variations through time.

The velocity–area method

Discharge is measured in cubic metres per second (m^3s^{-1}). It increases with the area of the channel cross section and with the velocity of flow. Discharge can be calculated for a given channel cross-section by measuring its cross-sectional area and the mean flow velocity:

$$Q = A \times v$$

where Q = discharge, A = cross-sectional area and v = mean flow velocity.

Figure 1(a) illustrates the method used. The first stage is to stretch a tape measure across the width of the channel. The channel is then divided into a number of sub-sections. This varies according to the width of the river, but discharge is usually gauged at twenty or more subsections. Ideally, the discharge flowing through each sub-section should be similar, so these are more closely spaced where the flow is deeper and faster.

Velocity measurements are then made using a flow meter. A commonly used design is a propeller mounted on a rod, which is lowered into the flow. The flow velocity is directly proportional to the rate at which the propeller is turned by the flow. A digital readout shows the velocity in m s^{-1} or, if you are not so lucky, the number of rotations per minute. (This can then be converted to the equivalent velocity using a simple formula.) One velocity measurement is made for each of the sub-sections, at the points indicated in Figure 1(a). Since flow velocity increases from zero at the channel bed to a maximum near the water surface, a representative average flow needs to be measured. It can be shown that the flow velocity at a height of $0.4d$ above the bed is representative of the average flow velocity, where d is the total depth of flow. Velocity measurements should therefore be made at a distance below the water surface that is 0.6 of the total depth (i.e. 0.4 of the depth from the bed). Thus if the depth of flow was 1 m, you would measure the velocity at a depth of 0.6 m below the surface (or 0.4 m above the bed).

The width and depth of each sub-section are measured as shown in the diagram. The discharge flowing through each sub-section can be calculated by multiplying the sub-section width (w), depth (d) and velocity (v). In order to calculate the discharge for the whole cross-section, the discharges for each of the sub-sections are added together. The discharge flowing through the ‘left over’ triangles adjacent to each bank is usually assumed to be negligible and is not included in this calculation.

This method is not very suitable for steep, turbulent, rocky streams where accurate current meter measurements are hard to obtain. A more appropriate technique in this case is to use **dilution gauging**. A chemical ‘tracer’ substance such as salt or dye is released into the flow, either as a single large ‘gulp’ or ‘slug’, or by continuously injecting it into the flow. Changes in concentration are monitored further downstream. Since the amount of dilution increases with discharge, it is possible to relate the change in concentration to stream discharge.

Continuous streamflow measurement

Although laborious, the velocity-area method only gives you one measurement of discharge at a particular point in time. In order to plot hydrographs like the ones in Figure 3.2, it is necessary to have a continuous record of flow.

Discharge is difficult to measure directly. However, it is related to **stage**, the water level or height, which is much easier to record. Figure 1(b) shows a gauging

Box 3.1

THE MEASUREMENT OF STREAMFLOW—CONT'D

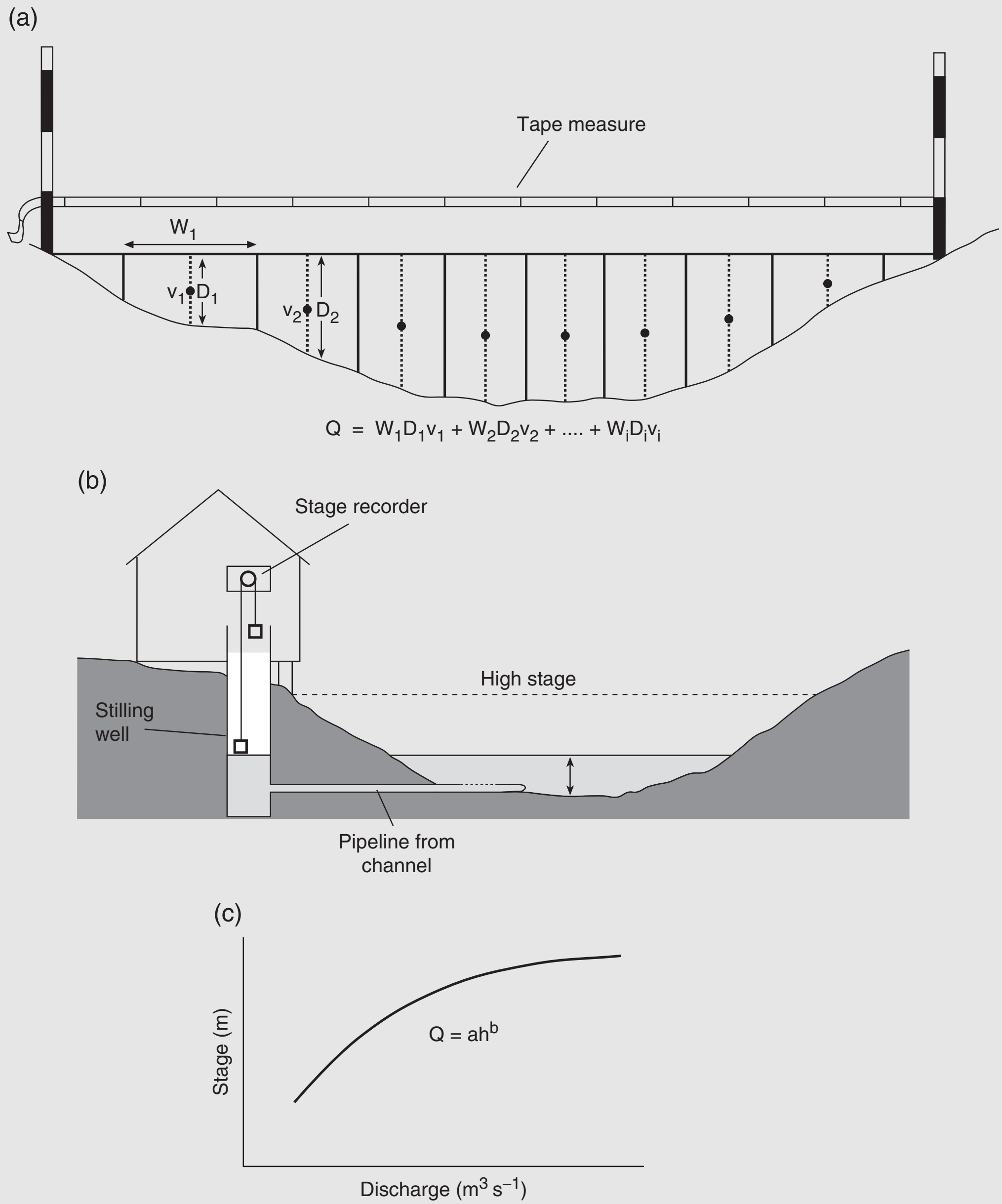

Figure 1 (a) Measuring discharge using the velocity-area method. Q = discharge, D = depth, w = width and v = velocity. (b) Gauging station at a rated section. (c) Rating curve. Q = discharge, h = stage, a and b are coefficients.

(Continued)

Box 3.1

THE MEASUREMENT OF STREAMFLOW—CONT'D

station situated on a natural cross-section. Rather than measuring the water level in the river itself, a pipeline leads to a **stilling well** which damps out the effects of surface waves and turbulence. The level of water in the stilling well is monitored using a **stage recorder**. Traditionally this would have consisted of a float attached to a counterweight via a pulley (shown in Figure 1b). As the water level rises and falls the float moves up and down, turning the pulley. This is attached to a pen, which draws a trace on a chart mounted on a rotating drum. The disadvantage is that the charts then have to be read manually. Digital measurements of water level can be made by sending an infrared signal down the stilling well. This is reflected by the water surface back to a receiver, with the time delay measured.

Stage is converted to discharge using a **rating curve** (Figure 1c). Using this, the discharge corresponding to a given stage can either be read off from the graph or calculated using the rating equation. This is of the form $Q = ah^b$, where Q = discharge, h = stage and a and b are coefficients, which describe the unique relationship between stage and discharge for the cross-section. This relationship is affected by the shape of the cross-section. At lower flows, small increases in discharge result in relatively large increases in stage because the channel is narrow near its bed. As the flow increases, a greater increase in discharge is needed to produce the same increase in depth, because the channel widens above the bed. Therefore a wide, shallow channel will have a different relationship (relatively small increases in stage with rising discharge) to a narrow, deep channel (relatively large increases in stage with rising discharge).

The rating curve is derived from discharge measurements made at different stages. Measurements are difficult to make at high flows, so most rating curves tend to be less reliable at high flows. Another problem is that if a large flood alters the shape of the cross-section, the rating curve has to be re-calibrated. This can be avoided using a gauging structure such as a **flume** or **weir**. These have a regular cross-sectional area and the flow velocity is controlled by the structure. The rating curve can be derived using a combination of measurements and hydraulic theory.

The top graph in Figure 3.2 (solid line) shows an **annual hydrograph** of daily flows, which might be observed for a river in the temperate zone over the course of a year. The lower graph is a single **storm hydrograph**, which shows the response of the drainage basin to one precipitation event. During a particular rainfall event, there is a delay between the onset of rainfall and the time at which the discharge starts to increase. The initial increase is due to water falling directly into the channel and close to it, though as the storm progresses, water travelling from greater distances reaches the channel. Water taking the fastest pathways – overland flow, shallow throughflow and pipe flow – is rapidly transferred to the channel, contributing to the **quick flow** component of the storm hydrograph. **Base flow** contributions come from water taking the slower subsurface routes, taking much longer to reach the channel. This means that water continues to enter the channel as base flow for some time after rainfall has ceased, keeping rivers flowing during dry periods. Glaciers, lakes, reservoirs and wetlands also contribute to base flow. The relative proportions of quick flow and base flow determine the size of the hydrograph peak and the time delay, or **lag time**, between peak rainfall and peak flow. For instance, where quick flow dominates, lag times are relatively short and peak flows relatively high. Rivers dominated by base flow respond more slowly and the peak flow is lower. The two lines shown in Figure 3.2(a) represent two temperate zone rivers, identical in every respect apart from the underlying geology: basin 1 is underlain by impermeable rocks

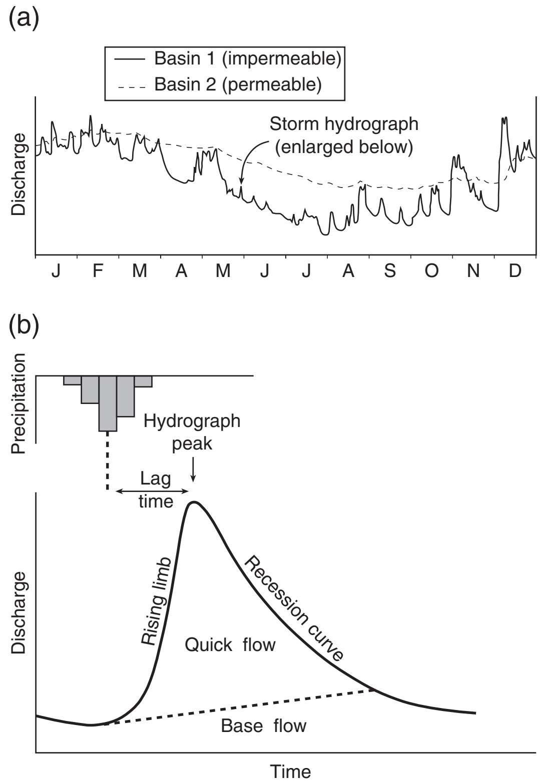

Figure 3.2 (a) Typical annual hydrograph of a temperate zone river whose drainage basin is underlain by an impermeable geology (solid line). Also shown is the annual hydrograph of a basin with a more permeable geology (dotted line). (b) Characteristics of a storm hydrograph.

and basin 2 by permeable rock. From the graphs you will see that basin 1 has a flashy response, with marked storm peaks, whereas basin 2 has a more damped response. This is because most of the precipitation falling over basin 2 infiltrates and takes slower subsurface pathways to the stream channel. There is also a marked difference in the summer low flows, with the greater base flow component for basin 2, which sustains higher summer flows.

The hydrological response of a river to discrete inputs of precipitation through time, as indicated by the shape of its hydrograph, is determined by the drainage basin characteristics and climatic factors shown in Table 3.1.

ANNUAL FLOW REGIMES

The **annual flow regime** of a river describes the seasonal variations in flow that are observed during an 'average' year. As you might expect, this is influenced by the seasonal distribution of rainfall, and the balance between rainfall and evaporation at different times of year. For example, some tropical rivers experience a marked wet and dry season, drying up completely for part of the year and carrying high flows during the wet season. Climate also has an important influence on the type and density of vegetation, soils and land use, all of which act as controls on the processes of runoff generation (Table 3.1).

Several climate characteristics are important in determining the flow regime. These include whether it is humid or arid, if it is predominantly warm or cold, the annual range of temperatures, and whether precipitation is seasonal or occurs all year round. At high latitudes and in some mountain environments, the timing and length of glacial ablation and snowmelt is a dominant factor. Figure 3.3 shows a selection of typical flow regimes, which characterise different climatic zones. These come from a classification scheme developed by Beckinsale (1969) from an existing climate classification. The different regimes are categorized using a system of two letters. The first letter relates to the mean annual precipitation and annual temperature range:

A: Warm, moist tropical climates, where the mean temperature exceeds 18°C for all months of the year.

B: Dry climates, where rates of potential evaporation[1] exceed annual precipitation.

C: Warm moist temperate climates.

D: Seasonally cold climates with snowfall, where the mean temperature is less than –3°C during the coldest month.

The second letter indicates the seasonal distribution of precipitation:

F: Appreciable rainfall all year round.

W: Marked winter low flow.

S: Marked summer low flow.

For example, the regime of the Pendari River (Figure 3.3A) is influenced by a tropical climate with a marked winter low flow, and would be classified as AW.

Table 3.1 Factors affecting hydrological response of a basin

Soils and geology

- *Soil type and thickness*. Soil texture (relative proportion of sand, silt and clay particles) affects infiltration rates. Sandy soils have high permeability whereas clay soils do not. In arid areas a crust can form on the soil surface, decreasing the permeability. Soil thickness affects how much water the soil can absorb.
- *Geology*. Drainage basins underlain by a permeable geology tend to have a slower response to precipitation, although the flow is sustained for a longer time during dry periods. Drainage basins underlain by impermeable materials have a faster, or more 'flashy' response.

Vegetation and land use

- *Vegetation type and density*. Vegetation reduces the impact of raindrops and allows a more 'open' soil structure, meaning that infiltration rates are higher. Vegetation also affects interception rates and evapotranspiration losses from the basin.
- *Urban areas.* Depends on the proportion of the drainage basin that is urbanised. Large areas of paved surfaces, drains and culverts rapidly transmit water to river channels, leading to an increase in peak flow and a shorter lag time.
- *Grazing and cultivation.* When deforestation occurs, rates of overland flow tend to increase. Heavy machinery and trampling by animals compact the soil, reducing permeability, although ploughing can increase infiltration rates. Flow may be concentrated in plough furrows that run up and down the slope.
- *Land drainage.* The installation of field drains allows rapid transfer of runoff into the nearest stream channel.

Physiographic characteristics

- *Drainage basin size and shape.* In larger basins the travel times are longer, as flow has to travel greater distances to reach the outlet. The total volume of runoff increases with the drainage area. Elongated drainage basins have a response that is initially more rapid but with a lower, more gentle peak.
- *Drainage density.* Where the density of stream channels is high, the average distance over which water has to travel to reach the channel network is reduced, leading to a more rapid response.
- *Drainage basin topography.* Travel times are increased over steep slopes. In upland areas, steep slopes are often associated with thin soils and the response tends to be flashy. Rainfall may be affected by altitude and aspect with respect to storm tracks.

Channel characteristics

- *Channel and floodplain resistance.* The velocity of flow in river channels is affected by the roughness of the bed and banks and the shape of the channel. Overbank flows are slowed by the roughness of the floodplain surface.
- *Floodplain storage.* When channel capacity is exceeded, water spills out onto the surrounding floodplain, where it is stored until the floodwaters recede. If floodplain storage is limited, a greater volume of water travels downstream.
- *Conveyance losses.* In dryland environments the channel may lose flow due to high rates of evaporation and 'leakage' by exfiltration through the channel boundary.

Meteorological factors

- *Antecedent conditions.* The conditions in the drainage basin prior to the onset of precipitation. Where recent or prolonged previous rainfall has occurred, the soil may be near saturation, meaning that a relatively small input of rainfall could lead to a rapid runoff response. Where snow is lying on the ground, subsequent rainfall can cause it to melt, which may lead to flooding downstream.
- *Rainfall intensity.* Rainfall intensity is expressed in millimetres per hour (mm h^{-1}). The more intense the rainfall, the more likely it is that the infiltration capacity of the soil will be exceeded.
- *Rainfall duration.* This is the period of time over which a given rainfall event takes place. As the storm progresses, runoff contributing areas at greater and greater distances from the channel network become active. The channel network may also extend upstream as normally dry channels start to carry flow.

The Arno (Figure 3.3C) has a warm temperate rainy climate, with summer low flows, and is classified as CS.

On first looking at Figure 3.3, it might seem strange that the graph for dry climates (B) has the biggest peak. Bear in mind that these graphs indicate annual flow variability, rather than actual monthly discharges. The value for each month is the ratio of the monthly mean flow to the overall (annual) mean, which is shown by the dotted line on each graph. Regimes associated with **tropical, rainy climates** (Figure 3.3A) are affected by seasonal shifts in the inter-tropical convergence zone. Near the equator, runoff occurs year-round, with peaks at the equinoxes (Lobaye River). Further north and south, there are marked wet and dry seasons (Pendjari River). The annual runoff for rivers in **dry climates** (Figure 3.3B) is low, but extremely variable. For much of the time there is little or no flow, but extreme floods can also occur. Flood pulses on the Cooper Creek are highly erratic and do not occur every year, since they are associated with El Niño disturbances to the monsoon. In contrast, there is much less variability for rivers with **temperate rainy climates** (Figure 3.3C). These have higher flows in winter and relatively low flows in summer, a pattern which is accentuated in Mediterranean climates (Arno). Snowmelt peaks are seen for rivers with **seasonally cold, snowy climates** (Figure 3.3D). The timing of the peak is dependent on altitude, latitude and seasonal patterns of rainfall. In **high mountain environments**

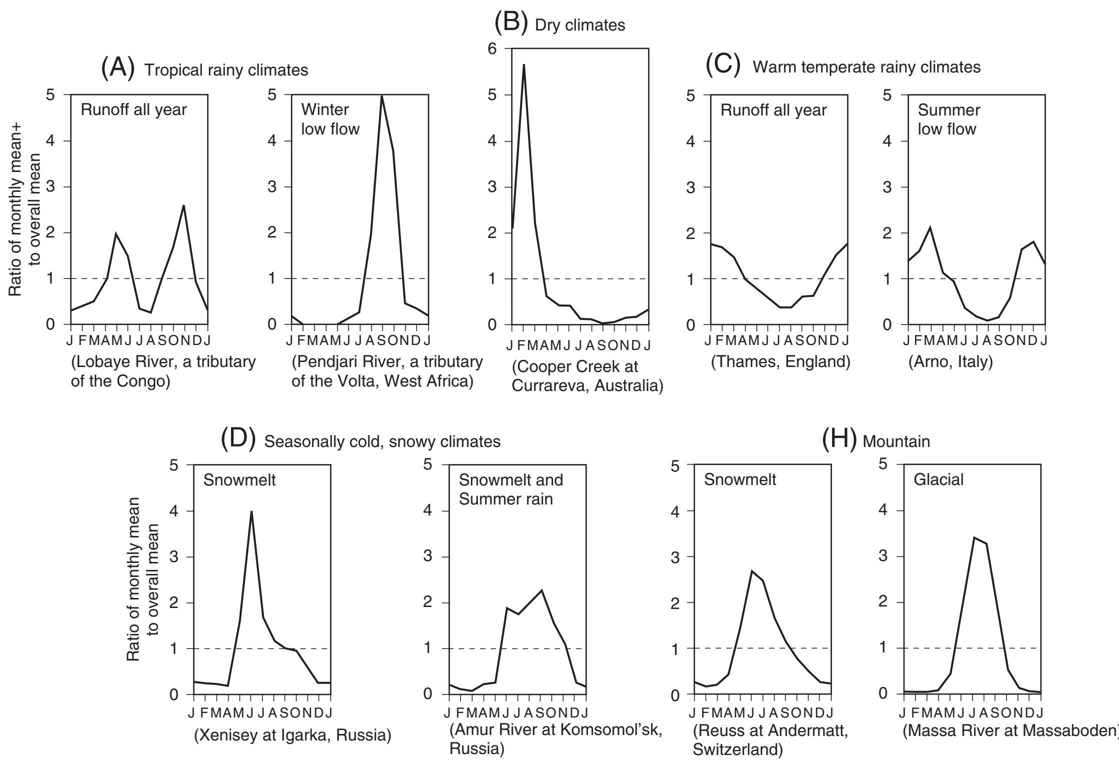

Figure 3.3 Flow regimes for selected rivers. (A) Tropical rainy climates: Lobaye River (a tributary of the Congo) and Pendjari River (a tributary of the Volta in West Africa). (B) Dry climates: Cooper Creek at Currareva, Australia. (C) Warm, temperate, rainy climates: Thames, England, and Arno, Italy. (D) Seasonally cold, snowy climates: Xenisey at Igarka and Amur River at Komsomol'sk, both in Russia. (H) Mountain climates: Reuse at Andermatt and Mass River at Massaboden, both in Switzerland. Source: Adapted from Beckinsale (1969); (B) after Knighton and Nanson (1997).

(Figure 3.3H), class HN denotes drainage basins with a nival regime (where snow patches exist) and HG those with a glacial regime. H stands for *Höhenklima*, which translates from the German as 'highland climate'.

For any drainage basin, inter-annual variations will occur between wet and dry years, so the flow regime is described using long-term averages. Even within a particular climate zone, the flow regime can differ markedly between drainage basins. This is because the physical characteristics of a drainage basin also play an important role, most notably the underlying geology and soil characteristics. These determine how much storage is available within the drainage basin in natural reservoirs, such as groundwater, lakes and wetlands.

Downstream variations in discharge

As well as varying through time, discharge also changes along the course of a river. At any location, channel form is dependent on the discharge and supply of sediment from upstream. In most cases, discharge increases downstream as the area of the drainage basin increases and tributaries join the main channel. There is also a general increase in the size of the channel, with discharge acting as a control on the gross dimensions (Knighton, 1998). The quantitative description and understanding of the nature of these downstream changes have been the focus of much research and is explored further in Chapter 8. Although there is a *general* downstream increase in channel dimensions, local influences lead to considerable variation, even over short distances.

Downstream changes in dryland channels can be very pronounced (Tooth, 2000). For example, infrequent flooding occurs along the ephemeral streams draining the Barrier Range in arid western New South Wales, Australia. Away from the uplands, high transmission losses lead to a rapid downstream reduction in discharge and channel size (Dunkerley, 1992). Downstream reductions in discharge and cross-sectional area are also observed in the piedmont and lowland zones of rivers draining the northern plains of arid central Australia (Tooth, 2000). There is an eventual termination of channel flow and bedload transport. However, during large floods, flows continue out across extensive unchannelled surfaces called 'floodouts' (Tooth, 2000). Tooth highlights the complex interactions between discharge, sediment transport, channel slope, tributary inflows, bank sediments and vegetation. These give rise to considerable variations in the downstream channel changes observed for dryland rivers.

FLOODS

Although the flow regime shows seasonal variations in river flow, it does not provide detailed information on the magnitude (size) and frequency of floods and droughts. Floods are of most interest here because they are capable of carrying out large amounts of geomorphological work and are thus significant in shaping the channel.

The term 'flood' is hard to define. In general terms, a flood is a relatively high flow that exceeds the capacity of the channel. While more frequent flows are confined within the channel, periodic high flows overtop the banks and spill out onto the surrounding floodplain. Significant here is the **bankfull discharge** (Q_b), defined as 'that discharge at which the channel is completely full' (Knighton, 1998). Although these definitions may sound straightforward enough, it is actually quite difficult to define bankfull discharge in the field because the height of the banks varies, even over short distances. This means that overtopping of the banks does not occur simultaneously at all points along the channel. Floodplain relief can be quite variable, with variations of between 1.7 m and 3.3 m observed on three Welsh floodplains (Lewin and Manton, 1975). Along the Alabama River, United States, flooding has been observed to occur more frequently at the apexes of actively migrating meander bends. This is associated with the development of floodplain features called levees. These are raised ridges that form along the banks when material is deposited during overbank flows (see Figure 8.9). Levee development is impeded at actively migrating bends because the deposits are eroded as the channel migrates. Levees are better developed (higher) along less actively migrating sections of channel, where flooding occurs less frequently (Harvey and Schumm, 1994).

Box 3.2

CALCULATING FLOOD RETURN PERIODS

Flood return period should ideally be calculated on the basis of at least thirty years worth of flow data. If possible a longer record should be used as this will contain a larger number of flood events and will provide a more representative sample of all flood events. The first step is to identify the peak flow for each year in the record to produce an **annual maximum series**. The **mean annual flood** is given by the mean of this annual maximum series. Mathematical analyses have shown that the recurrence interval of the mean annual flood is 2.33 years (Leopold *et al.*, 1964). In other words, this flow will be exceeded by the highest flow of the year once every 2.33 years on average.

All the floods in the annual maximum series are then ranked in order of magnitude, with the largest event ranked first and progressively smaller events given higher numbers. A simple formula is then used to assess the return period in years:

$$T = \frac{(n+1)}{m}$$

where T = return period in years, n = rank and m = number of years in record

If this is plotted using a logarithmic scale for flood magnitude the flood frequency curve is transformed to a straight line. It is possible to extrapolate (extend) this line to estimate the size of floods with larger return periods than those on record, although the difficulties associated with fitting a best-fit line through the existing data mean that a small difference in its gradient could make a big difference to the estimated size of the flood. There are also a number of practical difficulties associated such estimates and errors of flood discharges and errors in flood discharge estimates are generally considered to be in the range of 10 per cent to 100 per cent (Benito *et al.*, 2004). Large floods are very difficult to record accurately, since gauging stations can be damaged or even destroyed, leading to critical gaps in the flood record. Estimates of flood discharges are also dependent on the quality of the rating curve (relationship between stage and discharge – see Box 3.1).

Flood magnitude and frequency

Floods of different sizes are defined in terms of high water levels or discharges that exceed certain arbitrary limits. The height of the water level in a river is called its **stage**. For a given river, there is a relationship between the size of a flood (in terms of its maximum stage or discharge) and the frequency with which it occurs. Floods of different sizes do not occur with the same regularity: large floods are rarer than smaller floods. In other words, the larger the flood, the less often it can be expected to occur. Floods are therefore defined in terms of their magnitude (size) and frequency (how often a flood of a given size can be expected to occur).

You have probably heard reference to the ‘twenty-year flood’ or the ‘100-year flood’. This **return period** is an estimate of how often a flood of a given size can be expected to occur and, since less frequent floods are more extreme, the 100-year event would be bigger than the twenty-year flood.

The return period (T) can also be expressed as a probability (P) by taking the inverse of the return period, i.e.:

$$P = \frac{1}{T}$$

Using this, the probability of a 100-year flood taking place in any one year can be calculated as

0.01 (i.e. 1 per cent), and for the twenty-year flood, 0.05 (5 per cent). The probability that a flood with a particular return period will occur is the same every year and does not depend how long it was since a flood of this size last occurred – the twenty-year does not occur like clockwork every twenty years. However, if a period of several years is considered, the likelihood of a given flood occurring during this time increases. For example, if someone bought a house on the 100 year floodplain and lived there for thirty years, the probability of that property being flooded in any one year would be 0.01. This increases to 0.3 (probability × number of years), or 30 per cent, for the thirty-year period. Box 3.2 explains how return periods are estimated. As with any odds, flood probabilities are *estimates*, and a number of underlying assumptions are made when deriving them. It is assumed that runoff is randomly distributed through time and that the data set holds a representative sample of these random events. Estimates are therefore more reliable when a longer record is available, since a larger number of flood events will be included in it. Another assumption is that there are no long-term trends in the data, which is not the case when climate change is occurring.

The frequency of bankfull discharge

Although bedrock channels are mainly influenced by high magnitude flows, those formed in alluvium can be adjusted by a much greater range of flows (see Chapter 1, pp. 5–6). This is reflected by the morphology and size of alluvial channels. Over the years, much research has focused on the bankfull discharge (defined above), since it represents a distinct morphological discontinuity between in-bank and overbank flows. Leopold and Wolman (1957) suggested that the channel cross-section is adjusted to accommodate a discharge that recurs with a certain return period. From an examination of active floodplain rivers, they found that the bankfull discharge had a return period of between one and two years. This is corroborated by later observations made for stable alluvial rivers (for example, Andrews, 1980; Carling, 1988). However, the concept of a universal return period for bankfull discharge that can be applied to all rivers is controversial. Williams (1978) observed wide variations in the frequency of bankfull discharge, which ranged from 1.01 to 32 years, and concluded that this was too variable to assume a uniform return period for all rivers. Even along the same river, there can be marked variations in the frequency of bankfull discharge (Pickup and Warner, 1976).

The concept of a uniform frequency for bankfull discharge assumes that all channels are '**in regime**'. This means that the morphological characteristics of a given channel, such as size, fluctuate around a mean condition over the time scale considered (Pickup and Reiger, 1979). This is not true for all rivers and there are many examples of **non-regime**, or **disequilibrium**, channels. An example would be where channel incision is taking place through erosion of the channel bed. This results in a deeper channel, which requires a larger, and therefore less frequent, discharge to fill it. The Gila River in Arizona, United States, was greatly enlarged when past events had led to large floods. The enlarged channel is not adjusted to the contemporary flow regime, which means that the bankfull discharge for the enlarged channel has a much lower frequency (Stevens *et al.*, 1975). The material forming the bed and banks is also significant. In cases where the boundary is very erodible, the bankfull discharge may simply reflect the most recent flood event (Pickup and Warner, 1976).

The geomorphological effectivenes of floods

Given that many rivers exceed their channel capacity and flood on a fairly regular basis, it would not be unreasonable to ask why they do not shape channels that are large enough to convey all the flows supplied to them. While it is true that high-magnitude events lead to significant changes in channel morphology, the comparative rarity of these large floods must also be taken into account. The cumulative effect of smaller, more frequent floods can also be significant in shaping the channel. The effectiveness of any given discharge over a period of time is therefore something of a compromise between its size and how often it occurs. The basic question is: are a number of smaller floods as effective as one large flood? This concept is explored further in Box 3.3.

Box 3.3

THE FREQUENCY AND MAGNITUDE OF CHANNEL FORMING FLOWS

For a given channel, the geomorphological effectiveness of a particular flood event is dependent on the magnitude or size of that flood. This is because larger floods have more potential to erode and transport sediment. However, smaller floods, although less effective on an individual basis, occur more frequently. Over a period of time – decades to centuries – the cumulative volume of sediment transported by a number of these smaller floods can be greater than for one or two major floods (Wolman and Miller, 1960).

This concept is represented by Wolman and Miller's graphs, which are shown in Figure 1. The discharge frequency curve shows the frequency distribution of different flows ranging from droughts and low flows (left hand side) to large floods (right hand side). It can be seen that normal (non-flood) flows prevail for most of the time, while low flows and floods occur less often. The frequency of a given flood decreases with its magnitude.

The sediment transport rate curve represents the volume of sediment transported by individual floods of a given magnitude. This shows that the sediment transport rate increases with flood magnitude. The third curve, shown as a dotted line, indicates the cumulative effectiveness of a given flow over time. It shows the product of the sediment transport rate for that flow and its frequency of occurrence. The indication is that moderate, high frequency floods are most geomorphologically effective.

It has been suggested that channel parameters such as the spacing of meander bends, channel width and bankfull discharge are scaled to a single, **dominant discharge**. Indeed the frequency of bankfull discharge is similar to the flow that cumulatively transports the greatest volume of sediment. At least, that is *generally* the case for alluvial channels in humid temperate regions, where bankfull discharge typically occurs every one to two years.

However, despite the attractiveness of such a theory in understanding and modelling the adjustment of channel form, it does have serious limitations. For example, many channel geomorphic units such as bars and bedforms are adjusted by normal flows. Heritage *et al.* (2001), working on mixed bedrock–alluvial channels, found that while channel dimensions are scaled to flood flows, features such as bars within the channel are adjusted to lower flows. Thus, rather than being adjusted to a single, 'dominant' discharge, channels tend to adjust to a range of flows.

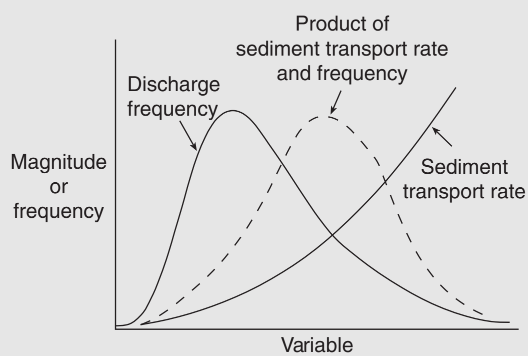

Figure 1 Relationships between discharge frequency, sediment transport rate and dominant discharge. Adapted from Wolman and Miller (1960).

Regional flood frequency curves

The flood frequency–magnitude relationship differs between regions. Despite the low annual rainfall in dryland environments, precipitation can be highly variable and the twelve largest floods ever recorded in the United States all occurred in semi-arid or arid areas (Costa, 1987). During flash floods, such as the one shown in Colour Plate 14, floodwaters rapidly inundate the dry channel. Not all dryland rivers are prone to flash flooding however, and there is considerable variation in the size, type and duration of flooding.

Regional flood frequency curves are shown in Figure 3.4. The return period is plotted on the horizontal axis using a logarithmic scale, with the relative flood magnitude on the vertical axis. A *relative* flood magnitude has been used to allow comparison between floods for a number of rivers in different regions. Because these all drain different areas, a direct comparison of flood magnitudes would not be very meaningful. Instead, for each river included in the analysis, the ratio between the magnitude of each flood on record and a low magnitude 'reference flow' – the mean annual flood – has been used. This is defined in Box 3.2 and has a return period of 2.33 years (i.e. the flow that will be equalled or exceeded on average once every 2.33 years[2]). The steepness of each curve reflects the variability of the flow, with arid zone rivers showing a much greater increase in relative flood magnitude at higher return periods. This reflects the extreme flow variability observed in these rivers and has important implications for the morphology of dryland channels, as will be seen in later chapters.

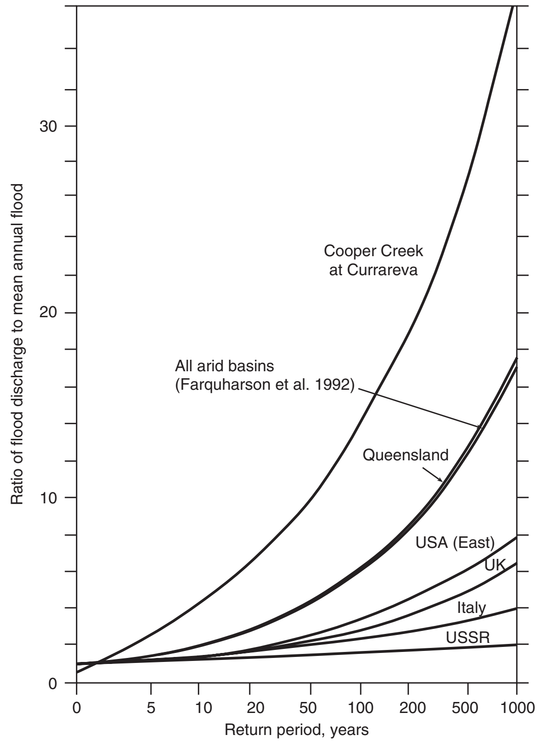

Figure 3.4 Regional growth curves, showing the relation between flood magnitude and frequency for various regions. For selected flood events the flood magnitude, relative to the mean annual flood, is plotted against the return period of that flood. After Knighton and Nanson (1997).

Reconstructing past floods

Palaeoflood hydrology is a new and developing area of hydrology and geomorphology, which reconstructs past flood events in order to extend the flow record. Due to problems associated with monitoring major floods and the relatively short duration of most gauged records, extreme floods are very rare in the observational record. By reconstructing palaeofloods, the flood record can be extended, allowing increased accuracy in the estimation of floods for risk analysis (Box 3.2). Evidence of past flood events is provided by geological indicators such as flood deposits, silt lines and erosion lines along the channel and valley walls (Benito *et al.*, 2004). Historical records are also used and include documents, chronicles and flood marks inscribed on bridges and buildings. Using this evidence, it is possible to determine the size of the largest flood events over periods of time ranging from decades to thousands of years (Benito *et al.*, 2004). As well as identifying the largest floods, evidence of floods above or below above specified flow stages can also be reconstructed (Stedinger and Baker, 1987). Although time-consuming, it is possible to reconstruct a complete record, chronicling the largest flood, together with the size and number of intermediate palaeofloods (Benito *et al.*, 2004). Chapter 9 discusses some of the techniques that are used in reconstructing past flood events.

CHAPTER SUMMARY

Inputs of precipitation falling over the area of a drainage basin are transferred to the channel via a number of different pathways. These include surface overland flow, throughflow (through the soil) and deeper groundwater flow. Rates of movement vary considerably: overland flow and shallow throughflow are generally much more rapid than groundwater flow. An important control is the infiltration capacity, which determines how quickly water can be absorbed by the soil. If the rainfall intensity is greater than the infiltration capacity, excess water builds up at the surface, leading to overland flow. Overland flow also occurs when rain falls on saturated areas. The volume of water flowing through a given cross-section in a given time (discharge) fluctuates in response to inputs of precipitation. A hydrograph shows how discharge changes over the course of a year (annual hydrograph), or one rainfall event (storm hydrograph). The shape of the hydrograph is affected by the physical, land use and climatic characteristics of the drainage basin. These variables all determine the relative proportion of water taking faster and slower pathways to reach the channel. Climate is a very important control on the annual flow regime of a river, which reflects the precipitation amount, seasonal distribution and annual temperature variations. Another important characteristic of the flow regime is the frequency and magnitude (size) of flood events. As the size of a flood increases, the frequency with which it occurs (return period) decreases. The relationship between frequency and magnitude differs from region to region. In dryland environments, large, low frequency floods are much more extreme than those with a similar return period in humid areas. The bankfull discharge is that flow at which the channel is completely filled. Wide variations are seen in the frequency with which the bankfull discharge occurs, although it *generally* has a return period of one to two years for many stable alluvial rivers. The geomorphological work carried out by a given flow depends not only on its size but also on its frequency of occurrence over a given period of time.

FURTHER READING

Introductory texts

Davie, T., 2003. *Fundamentals of Hydrology*. Routledge, London. Provides more detail on runoff generation and hydrograph analysis without getting too technical.

Jones, J.A.A., 1997. *Global Hydrology.* Longman, Harlow. Good on processes of runoff generation, measurement, and the effects of land use change.

Manning, J.C., 1997. *Applied Principles of Hydrology.* Third edition. Prentice Hall, Englewood Cliffs NJ. Written for students with a limited science background, this provides a very readable and accessible introduction.

Ward, R.C. and Robinson, M., 1990. *Principles of Hydrology.* McGraw-Hill, London. A well established hydrology textbook, which includes an interesting discussion on the development of runoff generation theory.

Websites

Dryland Rivers Research website, http://www.drylandrivers.com. An informative web site that covers many aspects of dryland rivers, with definitions, images, features, new research and useful links.

Hyperlinks in Hydrology for Europe and the Wider World, http://www.nerc-wallingford.ac.uk/ih/devel/wmo. Numerous links to websites of organisations active in various aspects of hydrology, including many educational sites.

CEH Wallingford, http://www.ceh.ac.uk/data/index.html. National River Flow Archive for the UK. Time series download facility (daily flows), hydrological summaries.