9

FLUVIAL LANDSCAPES

Running water wears away molehills and mountains, and builds fans, floodplains, and deltas. This chapter covers:

- running water
- water-carved landforms
- water-constructed landforms
- fluvial landscapes and humans

Running water in action: floods

Plum Creek flows northwards over a sand bed between Colorado Springs and Denver in the USA, and eventually joins the South Platte River. On 16 June 1965, a series of intense convective cells in the region climaxed in an intense storm, with 360 mm of rain falling in four hours, and a flood (Osterkamp and Costa 1987). The flood had a recurrence interval of between 900 and 1,600 years and a peak discharge of 4,360 m^3/s, which was fifteen times higher than the 50-year flood. It destroyed the gauging station at Louviers and swept through Denver causing severe damage. The flow at Louviers is estimated to have gone from less than 5 m^3/s to 4,360 m^3/s in about 40 minutes. At peak flow, the water across the valley averaged from 2.4 to 2.9 m deep, and in places was 5.8 m deep. The deeper sections flowed at around 5.4 m/s. The flood had far-reaching effects on the geomorphology and vegetation of the valley floor. Rampant erosion and undercutting of banks led to bank failures and channel widening. The processes were aided by debris snagged on trees and other obstructions, which caused them to topple and encourage sites of rapid scouring. Along a 4.08-km study reach, the average channel width increased from 26 to 68 m. Just over half the woody vegetation was destroyed. Following a heavy spring runoff in 1973, the channel increased to 115 m and increased its degree of braiding.

FLUVIAL ENVIRONMENTS

Running water dominates fluvial environments, which are widespread except in frigid regions, where ice dominates, and in dry regions, where wind tends to be the main erosive agent. However, in arid and semi-arid areas, fluvial activity can be instrumental in fashioning landforms. Flash floods build alluvial fans and run out on to desert floors. In the past, rivers once flowed across many areas that today lack permanent watercourses.

Water runs over hillslopes as overland flow and rushes down gullies and river channels as streamflow. The primary determinant of overland flow and streamflow is runoff production. **Runoff** is a component of the land-surface water balance. In brief, runoff is the difference between precipitation and evaporation rates, assuming that soil water storage stays roughly constant. In broad terms, fluvial environments dominate where, over a year, precipitation exceeds evaporation and the temperature regime does not favour persistent ice formation. Those conditions cover a sizeable portion of the land surface. The lowest annual runoff rates, less than 5 cm, are found in deserts. Humid climatic regions and mountains generate the most runoff, upwards of 100 cm in places, and have the highest river discharges.

Runoff is not produced evenly throughout the year. Seasonal changes in precipitation and evaporation generate systematic patterns of runoff that are echoed in **streamflow**. Streamflow tends to be highest during wet seasons and lowest during dry seasons. The changes of streamflow through a year define a river regime. Each climatic type fosters a distinct river regime. In monsoon climates, for example, river discharge swings from high to low with the shift from the wet season to the dry season. Humid climates tend to sustain a year-round flow of water in **perennial streams**. Some climates do not sustain a year-round river discharge. **Intermittent streams** flow for at least one month a year when runoff is produced. **Ephemeral streams**, which are common in arid environments, flow after occasional storms but are dry the rest of the time.

FLUVIAL EROSIONAL LANDFORMS

The action of flowing water cuts rills, gullies, and river channels into the land surface.

Rills and gullies

Rills are tiny hillside channels a few centimetres wide and deep that are cut by ephemeral rivulets. They grade into gullies. An arbitrary upper limit for rills is less than a third of a metre wide and two-thirds of a metre deep. Any fluvial hillside channel larger than that is a gully. **Gullies** are intermediate between rills and arroyos, which are larger incised stream beds. They tend to be deep and long and narrow, and continuous or discontinuous. They are not as long as valleys but are too deep to be crossed by wheeled vehicles or to be 'ironed out' by ploughing. They often start at a head-scarp or waterfall. Gullies bear many local names, including dongas, *vocarocas*, ramps, and *lavakas*. Much current gullying appears to result from human modification of the land surface leading to disequilibrium in the hillslope system. **Arroyos**, which are also called **wadis**, **washes**, **dry washes**, and **coulees**, are ephemeral stream channels in arid and semi-arid regions. They often have steep or vertical walls and flat, sandy floors. Flash floods course down normally dry arroyos during seasonal or irregular rainstorms, causing considerable erosion, transport, and deposition.

Bedrock channels

River channels may cut into rock and sediment. It is common to distinguish alluvial and bedrock channels, but many river channels form in a combination of alluvium and bedrock. Bedrock may alternate with thick alluvial fills, or bedrock may lie below a thin veneer of alluvium. The three chief types of river channel are bedrock channels, alluvial channels, and semi-controlled or channelized channels.

Bedrock channels are eroded into rock. They are resistant to erosion and tend to persist for long periods. They may move laterally in rock that is less resistant to erosion. Most rivers cut into bedrock in their upper reaches, where gradients are steep and their loads coarser.

However, some rivers, such as many in Africa, flow in alluvium in their upper reaches and cut into bedrock in the lower reaches (cf. p. 108). Bedrock channels are not well researched, with most attention being given to such small-scale erosional features as scour marks and potholes in the channel bed. The long profiles of bedrock channels are usually more irregular than the long profiles of alluvial channels. The irregularities may result from the occurrence of more resistant beds, from a downstream steepening of gradient below a **knickpoint** caused by a fall of baselevel, from faulting, or from landslides and other mass movements dumping a pile of debris in the channel. Rapids and waterfalls often mark their position.

Given that many kinds of bedrock are resistant to erosion, it might seem improbable that bedrock channels would meander. However, incised meanders do form in horizontally bedded strata. They form when a meandering river on alluvium eats down into the underlying bedrock. **Intrenched meanders**, such as those in the San Juan River, Utah, USA, are symmetrical forms and evolve where downcutting is fast enough to curtail lateral meander migration, a situation that would arise when a large fall of baselevel induced a knickpoint to migrate upstream (Colour Plate 10, inserted between pages 208 and 209. **Ingrown meanders** are asymmetrical and result from meanders moving sideways at the same time as they slowly incise owing to regional warping. A **natural arch** or **bridge** forms where two laterally migrating meanders cut through a bedrock spur (p. 207).

Springs sometimes cut into bedrock. Many springs issue from alcoves, channels, or ravines that have been excavated by the spring water. The 'box canyons' that open into the canyon of the Snake River in southern Idaho, USA, were cut into basalt by the springs that now rise at the canyon heads.

Alluvial channels

Alluvial channels form in sediment that has been, and is being, transported by flowing water. They are very diverse owing to the variability in the predominant grain size of the alluvium, which ranges from clay to boulders. They may change form substantially as discharge, sediment supply, and other factors change because alluvium is normally unable to resist erosion to any great extent. In plan view, alluvial channels display four basic forms that represent a graded series – straight, meandering, braided, and anastomosing (Figure 9.1a). Wandering channels are sometimes recognized as an intermediate grade between meandering channels and braided channels. Anabranching channels are another category (Figure 9.1b).

Straight channels

These are uncommon in the natural world. They are usually restricted to stretches of V-shaped valleys that are themselves straight owing to structural control exerted by faults or joints. **Straight channels** in flat valley-floors are almost invariably artificial. Even in a straight channel, the thalweg (the trace of the deepest points along the channel) usually winds from side to side, and the long profile usually displays a series of deeper and shallower sections (pools and riffles, p. 233) much like a meandering stream or a braided stream.

Meandering channels

Meandering channels wander snake-like across a floodplain (Plate 9.1; see also Plate 9.9). The dividing line between straight and meandering is arbitrarily defined by a sinuosity of 1.5, calculated by dividing the channel length by the valley length. Water flows through meanders in a characteristic pattern (Figure 9.2). The flow pattern encourages erosion and undercutting of banks on the outside of bends and deposition, and the formation of point bars on the inside of bends. The position of meanders changes, leading to the alteration of the course through cut-offs and channel diversion (avulsions). **Avulsions** are the sudden change in the course of a river leading to a section of abandoned channel, a section of new channel, and a segment of higher land (part of the floodplain) between them. Meanders may cut down or incise. Colour Plate 10 shows the famous incised meanders of the San Juan River, southern Utah, USA. **Cut-off incised meanders** may also form.

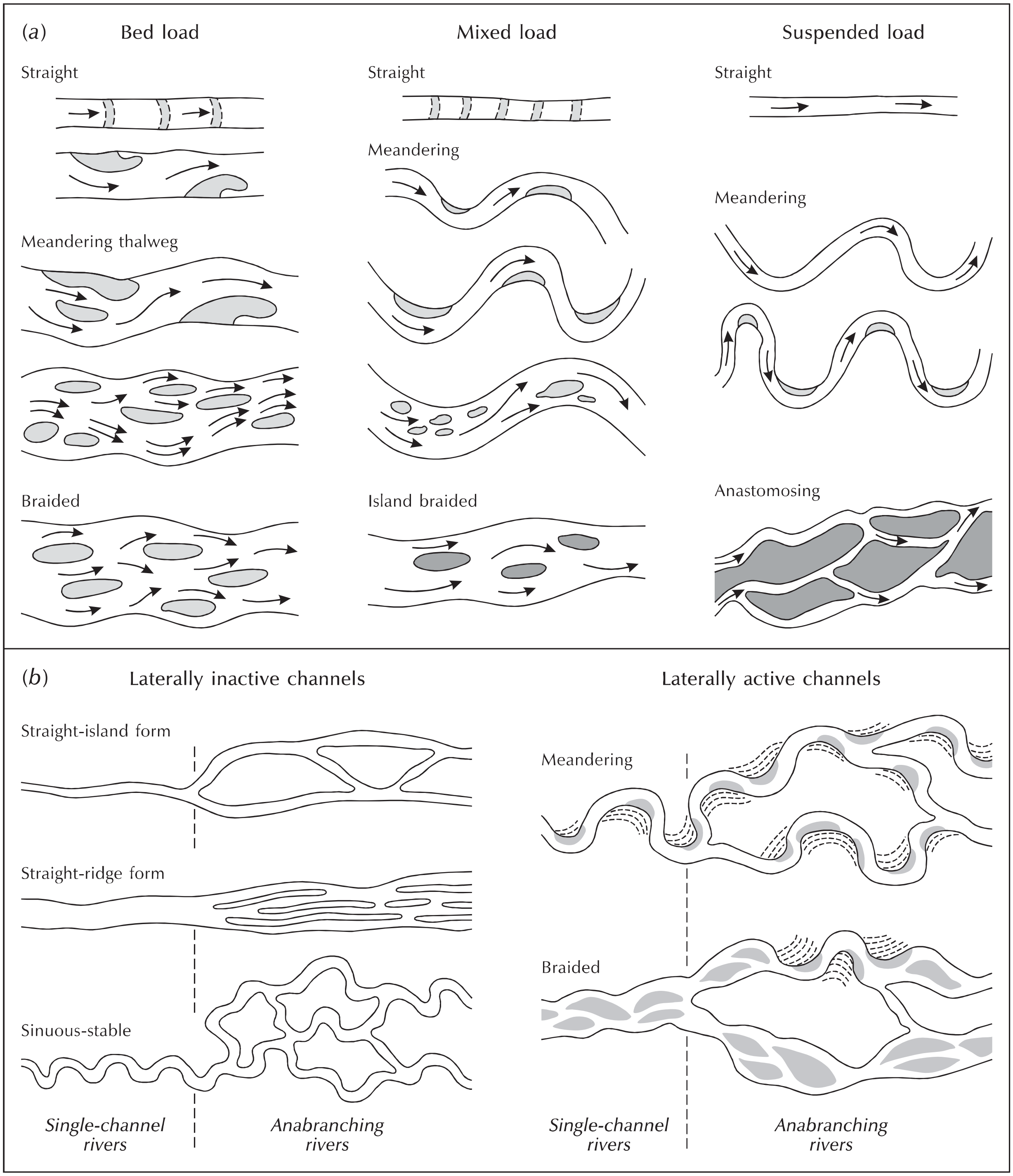

Figure 9.1 Classifications of channel patterns. (a) Channel form classified according to channel pattern (straight, meandering, braided, and anastomosing) and sediment load (suspended load, suspended-load and bed-load mix, bed load). (b) A classification of river patterns that includes single-channel and anabranching forms.
Sources: (a) Adapted from Schumm (1981, 1985b) and Knighton and Nanson (1993); (b) Adapted from Nanson and Knighton (1996)

Plate 9.1 Meanders on the River Bollin, Cheshire, England.
(*Photograph by David Knighton*)

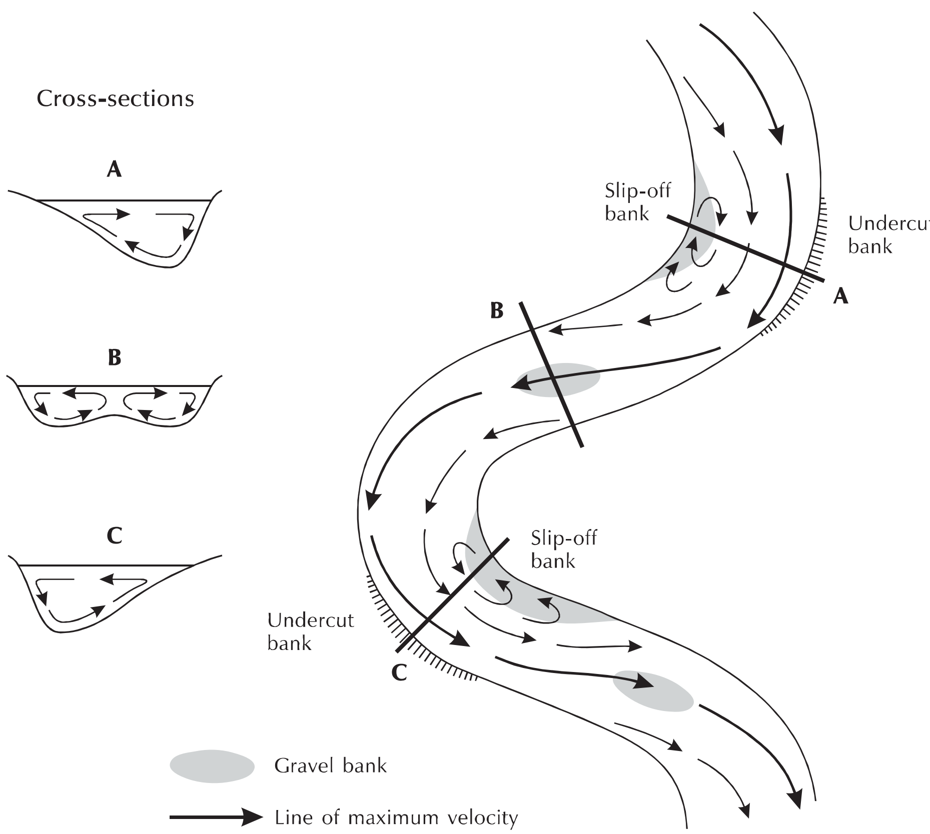

Figure 9.2 Water flow in a meandering channel.

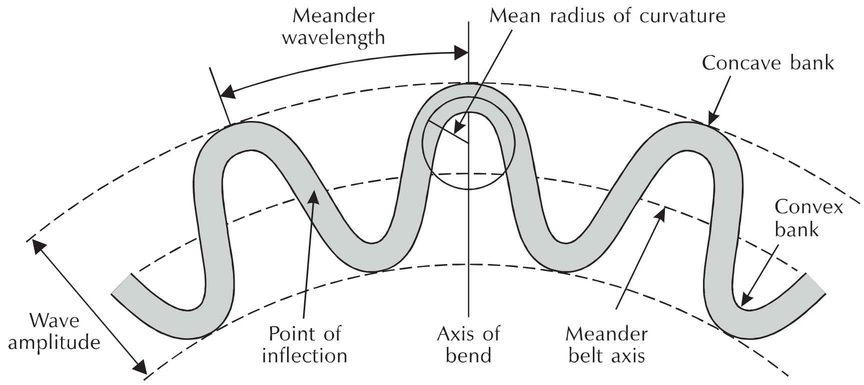

Figure 9.3 Parameters for describing meanders.

Meanders may be defined by several morphological parameters (Figure 9.3). Natural meanders are seldom perfectly symmetrical and regular owing to variations in the channel bed. Nonetheless, for most meandering rivers, the relationships between the morphometric parameters give a consistent picture: meander wavelength is about ten times channel width and about five times the radius of curvature.

Meandering is favoured where banks resist erosion, so forming deep and narrow channels. However, why rivers meander is not entirely clear. Ideas centre on: (1) the distribution and dissipation of energy within a river; (2) helical flow; and (3) the interplay of bank erosion, sediment load, and deposition. A consensus has emerged that meandering is caused by the intrinsic instabilities of turbulent water against a movable channel bank.

Braided channels

Braided channels (Plates 9.2 and 9.3) are essentially depositional forms that occur where the flow divides into a series of braids separated by islands or bars of accumulated sediment (see Best and Bristow 1993). The islands support vegetation and last a long time, while the bars are more impermanent. Once bars form in braided rivers, they are rapidly colonized by plants, so stabilizing the bar sediments and forming islands. However, counteracting the stabilization process is a highly variable stream discharge, which encourages alternate phases of degradation and aggradation in the channel and militates against vegetation establishment. Some braided rivers have twenty or more channels at one location.

Braided channels tend to form where (1) stream energy is high; (2) the channel gradient is steep; (3) sediment supply from hillslopes, tributaries, or glaciers is high and a big portion of coarse material is transported as bed load; and (4) bank material is erodible, allowing the channel to shift sideways with relative ease. They are common in glaciated mountains, where channel slopes are steep and the channel bed is very gravelly. They form in sand-bed and silt-bed streams where the sediment load is high, as in parts of the Brahmaputra River on the Indian subcontinent.

Anastomosing channels

Anastomosing channels have a set of distributaries that branch and rejoin (Plate 9.4). They are suggestive of braided channels, but braided channels are single-channel forms in which flow is diverted around obstacles in the channel, while anastomosing channels are a set of interconnected channels separated by bedrock or by stable alluvium. The formation of anastomosing channels is favoured by an aggradational regime involving a high suspended-sediment load in sites where lateral expansion is constrained. Anastomosing channels are rare: River Feshie, Scotland, is the only example in the UK.

Plate 9.2 The lower, braided reach of Nigel Creek, Alberta, Canada.
(*Photograph by David Knighton*)

Plate 9.3 Braiding and terraces at the junction of the Hope River (left) and Waiau River (centre), New Zealand.
(*Photograph by David Knighton*)

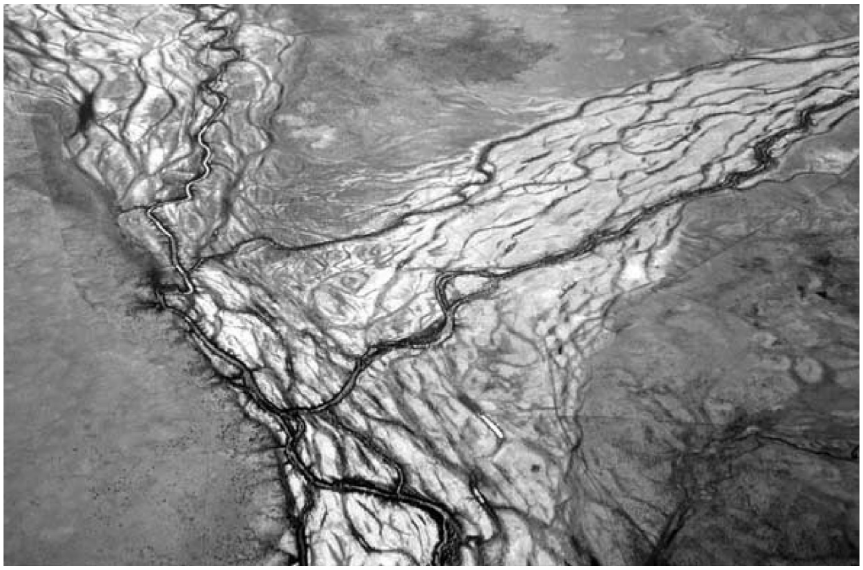

Plate 9.4 The junction of two anastomosing rivers, Queensland, Australia.
(*Photograph by David Knighton*)

Anabranching channels

Anabranching rivers consist of multiple channels separated by vegetated and semi-permanent alluvial islands or alluvial ridges. The islands are cut out of the floodplain or are constructed in channels by the accretion of sediments. Anabranching is a fairly uncommon but a widespread channel pattern that may affect straight, meandering, and braided channels alike (Figure 9.1). Conditions conducive to the development of anabranching include frequent floods, channel banks that resist erosion, and mechanisms that block or restrict channels and trigger avulsions. The anabranching rivers of the Australian interior seem to be the outcome of low-angle slopes and irregular flow regimes. Those on the alluvial plains of south-western New South Wales form a complicated network along 100 km and more of the Edward and Murray Rivers; for instance, Beveridge Island is about 10 km long and lies between two roughly equal branches of the Murray River. Those on the Northern Plains near Alice Springs appear to be a stable river pattern designed to preserve a throughput of relatively coarse sediment in low-gradient channels that characteristically have abundant vegetation in them and declining downstream discharges (Tooth and Nanson 1999).

Hydraulic geometry

The controlling influence of discharge upon channel form, resistance to flow, and flow velocity is explored in the concept of **hydraulic geometry**. The key to this concept is the discharge equation:

$$Q = wdv$$

where Q is stream discharge (m^3/s), w is the stream width (m), d is the mean depth of the stream in a cross-section (m), and v is the mean flow velocity in the cross-section (m/s). As a rule of thumb, the mean velocity and width–depth ratio (w/d) both increase downstream along alluvial channels as discharge increases. If discharge stays the same, then the product wdv does not change. Any change in width or depth or velocity causes compensating changes in the other two components. If stream width were to be reduced, then water depth would increase. The increased depth, through the relationships expressed in the Manning equation (p. 71), leads to an increased velocity. In turn, the increased velocity may then cause bank erosion, so widening the stream again and returning the system to a balance. The compensating changes are conservative in that they operate to achieve a roughly continuous and uniform rate of energy loss – a channel's geometry is designed to keep total energy expenditure to a minimum. Nonetheless, the interactions of width, depth, and velocity are indeterminate in the sense that it is difficult to predict an increase of velocity in a particular stream channel. They are also complicated by the fact that width, depth, velocity, and other channel variables respond at different rates to changing discharge. Bedforms and the width–depth ratio are usually the most responsive, while the channel slope is the least responsive. Another difficulty is knowing which stream discharge a channel adjusts to. Early work by M. Gordon Wolman and John P. Miller (1960) suggested that the bankfull discharge, which has a 5-year recurrence interval, is the dominant discharge, but recent research shows that as hydrological variability or channel boundary resistance (or both) becomes greater, then channel form tends to adjust to the less frequent floods. Such incertitude over the relationship between channel form and discharge makes reconstructions of past hydrological conditions from relict channels problematic.

Changes in hydrological regimes may lead to a complete alteration of alluvial channel form, or what Stanley A. Schumm called a '**river metamorphosis**'. Such a thoroughgoing reorganization of channels may take decades or centuries. Human interference within a catchment often triggers it, but it may also occur owing to internal thresholds within the fluvial system and happen independently of changes in discharge and sediment supply. A good example of this comes from the western USA, where channels incised when aggradation caused the alluvial valley floor to exceed a threshold slope (Schumm and Parker 1977). As the channels cut headwards, the increased sediment supply caused aggradation and braiding in downstream reaches. When incision ceased, less sediment was produced at the stream head and incision began in the lower reaches. Two or three such aggradation–incision cycles occurred before equilibrium was accomplished.

River long profiles, baselevel, and grade

The **longitudinal profile** or **long profile** of a river is the gradient of its water-surface line from source to mouth. Streams with discharge increasing downstream have concave long profiles. This is because the drag force of flowing water depends on the product of channel gradient and water depth. Depth increases with increasing discharge and so, in moving downstream, a progressively lower gradient is sufficient to transport the bed load. Many river long profiles are not smoothly concave but contain flatter and steeper sections. The steeper sections, which start at **knickpoints**, may result from outcrops of hard rock, the action of local tectonic movements, sudden changes in discharge, or critical stages in valley development such as active headward erosion. The long profile of the River Rhine in Germany is shown in Figure 9.4. Notice that the river is 1,236 km long and falls about 3 km from source to mouth, so the vertical distance from source to mouth is just 0.24 per cent of the length. Knickpoints can be seen at the Rhine Falls near Schaffhausen and just below Bingen. Most long profiles are difficult to interpret solely in terms of fluvial processes, especially in the case of big rivers, which are normally old rivers with lengthy histories, unique tectonic and other events in which may have influenced their development. Even young rivers cutting into bedrock in the Swiss Alps and the Southern Alps of New Zealand have knickpoints, which seem to result from large rock-slope failures (Korup 2006).

Baselevel is the lowest elevation to which downcutting by a stream is possible. The ultimate baselevel for any stream is the water body into which it flows – sea, lake, or, in the case of some enclosed basins, playa, or salt lake (p. 234). Main channels also prevent further downcutting by tributaries and so provide a baselevel. Local baselevels arise from bands of resistant rock, dams of woody debris, beaver ponds, and human-made dams, weirs, and so on. The complex long profile of the River Rhine has three segments, each with a local baselevel. The first is Lake Constance, the second lies below Basel, where the Upper Rhine Plain lies within the Rhine Graben, and the third lies below Bonn, where the Lower Rhine embayment serves as a regional baselevel above the mouth of the river at the North Sea (Figure 9.4).

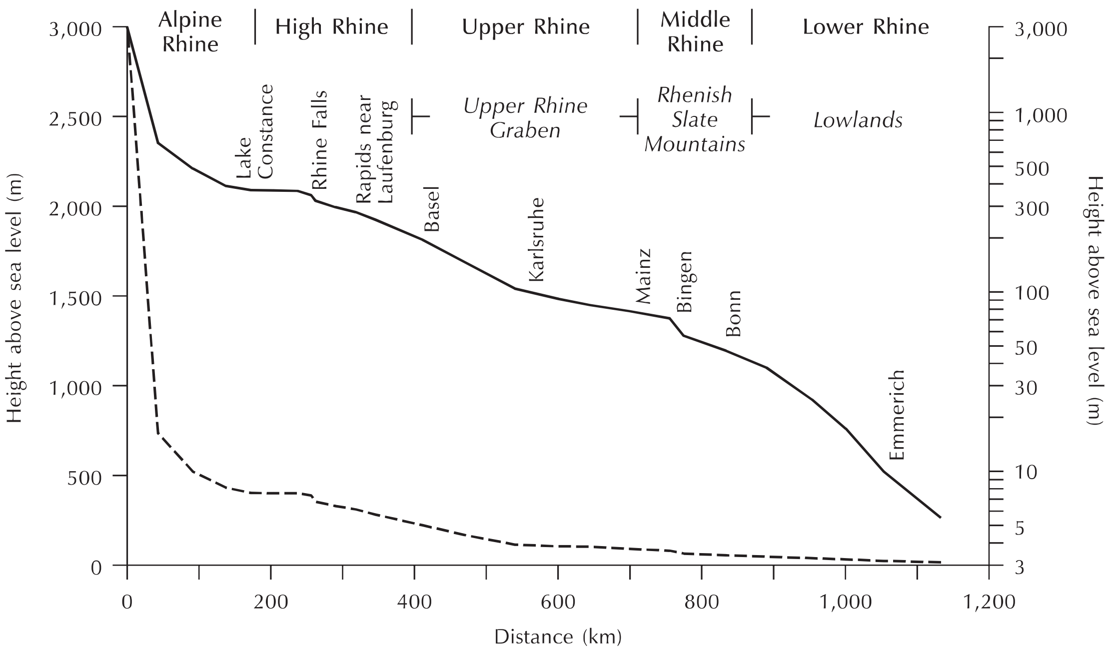

Figure 9.4 Long-profile of the River Rhine, shown on an arithmetic height scale (dashed line) and logarithmic height scale (solid line).
Source: After Ahnert (1998, 174)

Grade, as defined by J. Hoover Mackin (1948), is a state of a river system in which controlling variables and baselevel are constant:

> A graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristics, just the velocity required for the transportation of the load provided by the drainage basin. The graded stream is a system in equilibrium; its diagnostic characteristic is that any change in any of the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change.
>
> (Mackin 1948, 471)

If the baselevel changes, then streams adjust their grade by changing their channel slope (through aggradation or degradation), or by changing their channel pattern, width, or roughness. However, as the controlling variables usually change more frequently than the time taken for the channel properties to respond, a graded stream displays a quasi-equilibrium rather than a true steady state.

Drainage basins and river channel networks

A river system can be considered as a network in which **nodes** (stream tips and stream junctions) are joined by **links** (streams). Stream segments or links are the basic units of stream networks. **Stream order** is used to denote the hierarchical relationship between stream segments and allows drainage basins to be classified according to size. Stream order is a basic property of stream networks because it relates to the relative discharge of a channel segment. Several stream-ordering systems exist, the most commonly used being those devised by Arthur N. Strahler and by Ronald L. Shreve (Figure 9.5). In **Strahler's ordering system**, a stream segment with no tributaries that flows from the stream source is denoted as a first-order segment. A second-order segment is created by joining two first-order segments, a third-order segment by joining two second-order segments, and so on. There is no increase in order when a segment of one order is joined by another of

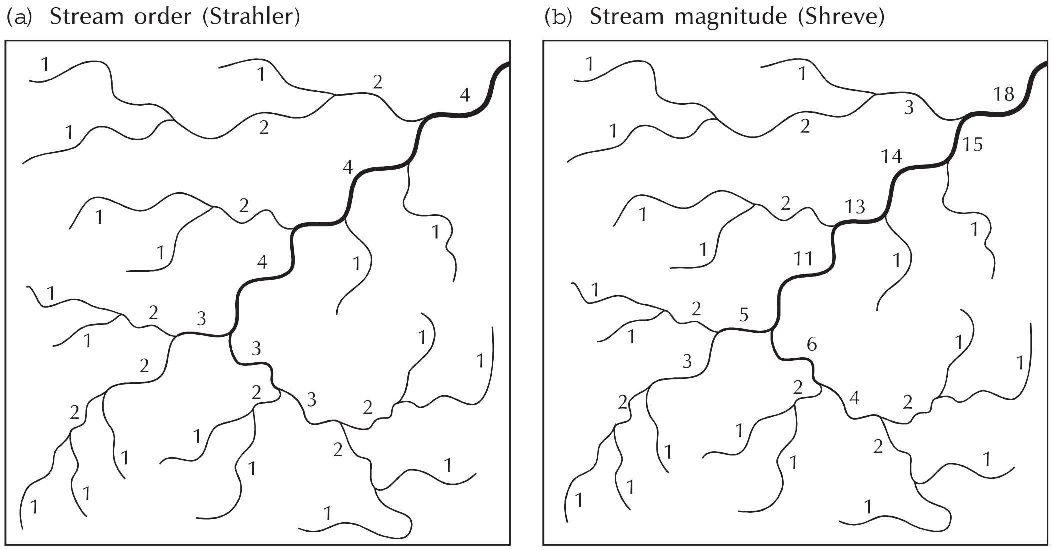

Figure 9.5 Stream ordering. (a) Strahler's system. (b) Shreve's system.

a lower order. Strahler's system takes no account of distance and all fourth-order basins are considered as similar. **Shreve's ordering system**, on the other hand, defines the magnitude of a channel segment as the total number of tributaries that feed it. Stream magnitude is closely related to the proportion of the total basin area contributing runoff, and so it provides a good estimate of relative stream discharge for small river systems.

Strahler's stream order has been applied to many river systems and it has been proved statistically to be related to a number of drainage-basin morphometry elements. For instance, the mean stream gradients of each order approximate an inverse geometric series, in which the first term is the mean gradient of first-order streams. A commonly used topological property is the **bifurcation ratio**, that is, the ratio between the number of stream segments of one order and the number of the next-highest order. A mean bifurcation ratio is usually used because the ratio values for different successive basins will vary slightly. With relatively homogeneous lithology, the bifurcation ratio is normally not more than five or less than three. However, a value of ten or more is possible in very elongated basins where there are narrow, alternating outcrops of soft and resistant strata.

The main geometrical properties of stream networks and drainage basins are listed in Table 9.1. The most important of these is probably drainage density, which is the average length of channel per unit area of drainage basin. **Drainage density** is a measure of how frequently streams occur on the land surface. It reflects a balance between erosive forces and the resistance of the ground surface, and is therefore related closely to climate, lithology, and vegetation. Drainage densities can range from less than 5 km/km^2 when slopes are gentle, rainfall low, and bedrock permeable (e.g. sandstones), to much larger values of more than 500 km/km^2 in upland areas where rocks are impermeable, slopes are steep, and rainfall totals are high (e.g. on unvegetated clay 'badlands' – Plate 9.5). Climate is important in basins of very high drainage densities in some semi-arid environments that seem to result from the prevalence of surface runoff and the relative ease with which new channels are created. Vegetation density is influential in determining drainage density, since it binds the surface layer preventing overland flow from concentrating along definite lines and from eroding small rills, which may develop into stream channels. Vegetation slows the rate of overland flow and effectively stores some of the water for short time periods. Drainage density also relates to the length of overland flow,

Table 9.1 Selected morphometric properties of stream networks and drainage basins

Property	*Symbol*	*Definition*
Network properties		
Drainage density	D	Mean length of stream channels per unit area
Stream frequency	F	Number of stream segments per unit area
Length of overland flow	L_g	The mean upslope distance from channels to watershed
Areal properties		
Texture ratio	T	The number of crenulations in the basin contour having the maximum number of crenulations divided by the basin perimeter length. Usually bears a strong relationship to drainage density
Circulatory ratio	C	Basin area divided by the area of a circle with the same basin perimeter
Elongation ratio	E	Diameter of circle with the same area as the drainage basin divided by the maximum length of the drainage basin
Lemniscate ratio	k	The square of basin length divided by four times the basin area
Relief properties		
Basin relief	H	Elevational difference between the highest and lowest points in the basin
Relative relief	R_{hp}	Basin relief divided by the basin perimeter
Relief ratio	R_h	Basin relief divided by the maximum basin length
Ruggedness number	N	The product of basin relief and drainage density

Source: Adapted from Huggett and Cheesman (2002, 98)

Plate 9.5 High drainage density in the Zabriskie Point badlands, Death Valley, California, USA. (*Photograph by Kate Holden*)

which is approximately equal to the reciprocal of twice the drainage density. And, importantly, it determines the distance from streams to valley divides, which strongly affects the general appearance of any landscape.

Early studies of stream networks indicated that purely random processes could generate fluvial systems with topological properties similar to natural systems (Shreve 1975; Smart 1978). Such random-model thinking has been extremely influential in channel network studies. However, later research has identified numerous regularities in stream network topology. These systematic variations appear to be a result of various factors, including the need for lower-order basins to fit together, the sinuosity of valleys and the migration of valley bends downstream, and the length and steepness of valley sides. These elements are more pronounced in large basins, but they are present in small catchments.

Valleys

Valleys are so common that geomorphologists seldom defined them and, strangely, tended to overlook them as landforms. **True valleys** are simply linear depressions on the land surface that are almost invariably longer than they are wide with floors that slope downwards. Under special circumstances, as in some overdeepened glaciated valleys (p. 255), sections of a valley floor may be flat or slope upwards. Valleys occur in a range of sizes and go by a welter of names, some of which refer to the specific types of valley – gully, draw, defile, ravine, gulch, hollow, run, arroyo, gorge, canyon, dell, glen, dale, and vale.

As a rule, valleys are created by fluvial erosion, but often in conjunction with tectonic processes. Some landforms that are called 'valleys' are produced almost entirely by tectonic processes and are not true valleys – Death Valley, California, which is a half-graben, is a case in point. Indeed, some seemingly archetypal fluvial landforms, including river valleys, river benches, and river gorges, appear to be basically structural landforms that have been modified by weathering and erosion. The Aare Gorge in the Bernese Oberland, the Moutier–Klus Gorge in the Swiss Jura, the Samaria Gorge in Crete, hill-klamms in the Vienna Woods, Austria, and the Niagara Gorge in Ontario and New York state all follow pre-existing faults and clefts (Scheidegger and Hantke 1994). Erosive processes may have deepened and widened them, but they are essentially endogenic features and not the product of antecedent rivers.

Like the rivers that fashion them, valleys form **networks** of main valleys and tributaries. Valleys grow by becoming deeper, wider, and longer through the action of running water. Valleys deepen by hydraulic action, corrasion, abrasion, potholing, corrosion, and weathering of the valley floor. They widen by lateral stream erosion and by weathering, mass movements, and fluvial processes on the valley sides. They lengthen by headward erosion, by valley meandering, by extending over newly exposed land at their bottom ends, and by forming deltas.

Some valley systems are exceptionally old – the Kimberly area of Australia had been land throughout the Phanerozoic and was little affected by the ice ages (Ollier 1991, 99). The drainage system in the area is at least 500 million years old. Permian, Mesozoic, Mid- to Late Cretaceous, and Early Tertiary drainage has also been identified on the Australian continent.

FLUVIAL DEPOSITIONAL LANDFORMS

Alluvial bedforms

Riverbeds develop a variety of landforms generated by turbulence associated with irregular cross-channel or vertical velocity distributions that erode and deposit alluvium. The forms are **riffle–pool sequences** (Box 9.1) and **ripple–antidune sequences** (Figure 9.7). **Step–pool sequences** are large-scale and created by, for example, the dam-building activities of beavers.

Floodplains

Most rivers, save those in mountains, are flanked by an area of moderately flat land called a **floodplain**, which is formed from debris deposited when the river is in flood. Small floods that occur frequently cover a part of the floodplain, while rare major floods submerge the entire area. The width of floodplains is roughly proportional to river discharge. The active floodplain of the lower Mississippi River is some 15 km across. Adjacent floodplains in regions of subdued topography may coalesce to form **alluvial plains**.

Convex floodplains

The low-gradient floodplains of most large rivers, including those of the Rivers Mississippi, Amazon, and Nile, are broad and have slightly convex cross-sections, the land sloping away from the riverbank to the valley sides (Figure 9.8a). The convexity is primarily a product of sedimentation. Bed load and suspended sediment are laid down in the low-water channel and along its immediate edges, while only suspended materials are laid down in the flood basins and backswamps. Bed load accumulates more rapidly than suspended load, and deposition is more frequent in and near to the channel than it is in overbank sites. In consequence, the channel banks and levees grow faster than the flood basins and may stand 1–15 m higher.

Flat floodplains

The majority of small floodplains are flat or gently concave in cross-section (Figure 9.8b). On these flat floodplains, natural levees are small or absent and the

Box 9.1

POOLS AND RIFFLES

River channels, even initially straight ones, tend to develop deeper and shallower sections. These are called **pools** and **riffles** respectively (Plate 9.6). Experiments in flumes, with water fed in at a constant rate, produce pool-and-riffle sequences, in which the spacing from one pool to the next is about five times the channel width (Figure 9.6). Continued development sees meanders forming with alternate pools migrating to opposite sides. The meander wavelength is roughly two inter-pool spacings of ten channel widths, as is common in natural rivers.

Plate 9.6 Riffles and pools in a straight section of the River Dean, Adlington Hall, Cheshire, England. A pool may be seen in the foreground, a riffle to the right of the middle-ground bar, with other pools and riffles beyond. (*Photograph by David Knighton*)

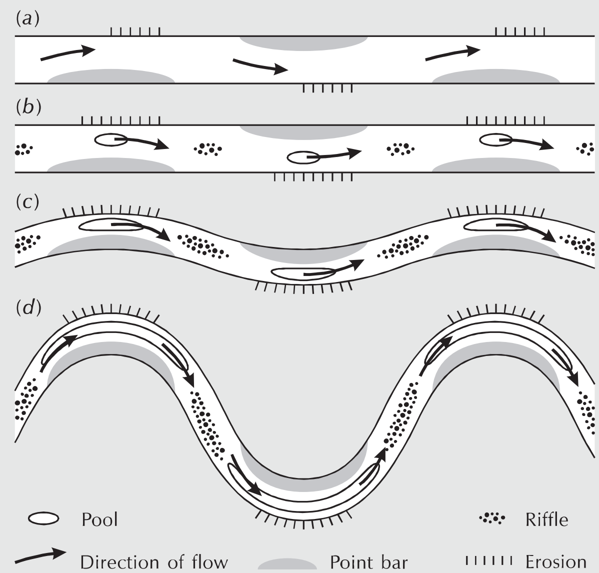

Figure 9.6 Pool-and-riffle sequences in river channels. (a) Alternating zones of channel erosion and accretion in response to faster and slower flow. (b) Pool spacing influencing the evolution of a straight channel into a meandering channel. (c) Additional pools form as the meandering channel lengthens. (d) Development of meandering channel with pools and riffles.
Source: Adapted from Dury (1969)

alluvial flats rise gently to the valley sides. The concave form is encouraged by a small floodplain area that is liable to continual reworking by the stream. Most medium-sized rivers, and many major rivers, have flat floodplains formed chiefly by lateral accretion (sedimentation on the inside of meander bends). Flat floodplains may also form by alluviation in braided streams.

Alluvial fans

An **alluvial fan** is a cone-shaped body that forms where a stream flowing out of mountains debouches on to a plain (Plate 9.7). The alluvial deposits radiate from the **fan apex**, which is the point at which the stream emerges from the mountains. Radiating channels cut into the fan. These are at their deepest near the apex and shallow with increasing distance from the apex, eventually converging with the **fan surface**. The zone of deposition on the fan runs back from the break of slope between the fan surface and the flat land in front of the **fan toe**. It was once thought that deposition was induced by a break of slope in the stream profile at the fan apex, but it has been shown that only rarely is there a break of slope at that point. The steepness of the fan slope depends on the size of the stream and the coarseness of the load, with the steepest alluvial fans being associated with small streams and coarse loads. Fans are common in arid and semi-arid areas but occur in all climatic zones. They range greatly in size. Some in Queensland, Australia, are plain to see on topographic maps or satellite images, but cannot be recognized on the ground because they have radii of about 100 km and are so flat.

Playas

Playas are the flattest and the smoothest landforms on the Earth (Plate 9.8). A prime example is the Bonneville salt flats in Utah, USA, which is ideal for high-speed car racing, although some playas contain large desicca-tion cracks so caution is advised. Playas are known as

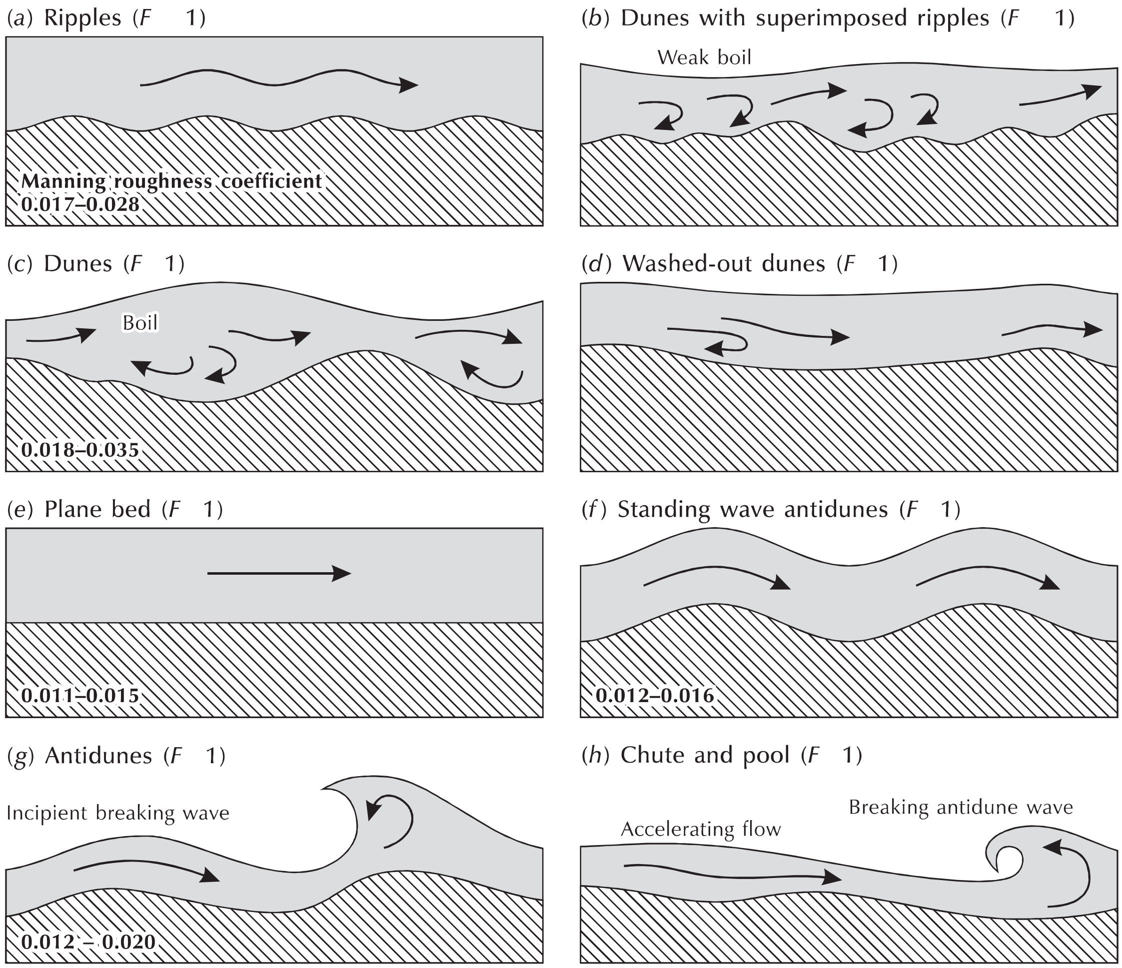

Figure 9.7 Bedforms in a sandy alluvial channel change as the Froude number, F, changes. At low flow velocities, ripples form that change into dunes as velocity increases. A further increase of velocity planes off bed undulations, and eventually a plane bed forms. The plane bed reduces resistance to flow, and sediment rates increase. The channel then stands poised at the threshold of subcritical and supercritical flow. A further increase of velocity initiates supercritical flow, and standing antidunes form. Flow resistance is low at this stage because the antidunes are in phase with the standing waves. The antidunes move upstream because they lose sediment from their downstream sides faster than they gain it through deposition. At the highest velocities, fast-flowing and shallow chutes alternate with deeper pools.
Source: Adapted from Simons and Richardson (1963) and Simons (1969)

salinas in South America and *sabkhas* or *sebkhas* in Africa. They occur in closed basins of continental interiors, which are called bolsons in North America. The **bolsons** are surrounded by mountains out of which floodwaters laden with sediment debouch into the basin. The coarser sediment is deposited to form alluvial fans, which may coalesce to form complex sloping plains known as **bajadas**. The remaining material – mainly fine sand, silt, and clay – washes out over the playa and settles as the water evaporates. The floor of the playa accumulates sediment at the rate of a few centimetres to a metre in a millennium. As water fills the lowest part of the playa, deposited sediment tends to level the terrain. Playas typically occupy about 2–6 per cent of the depositional area in a bolson. Many bolsons contained perennial lakes during the Pleistocene.

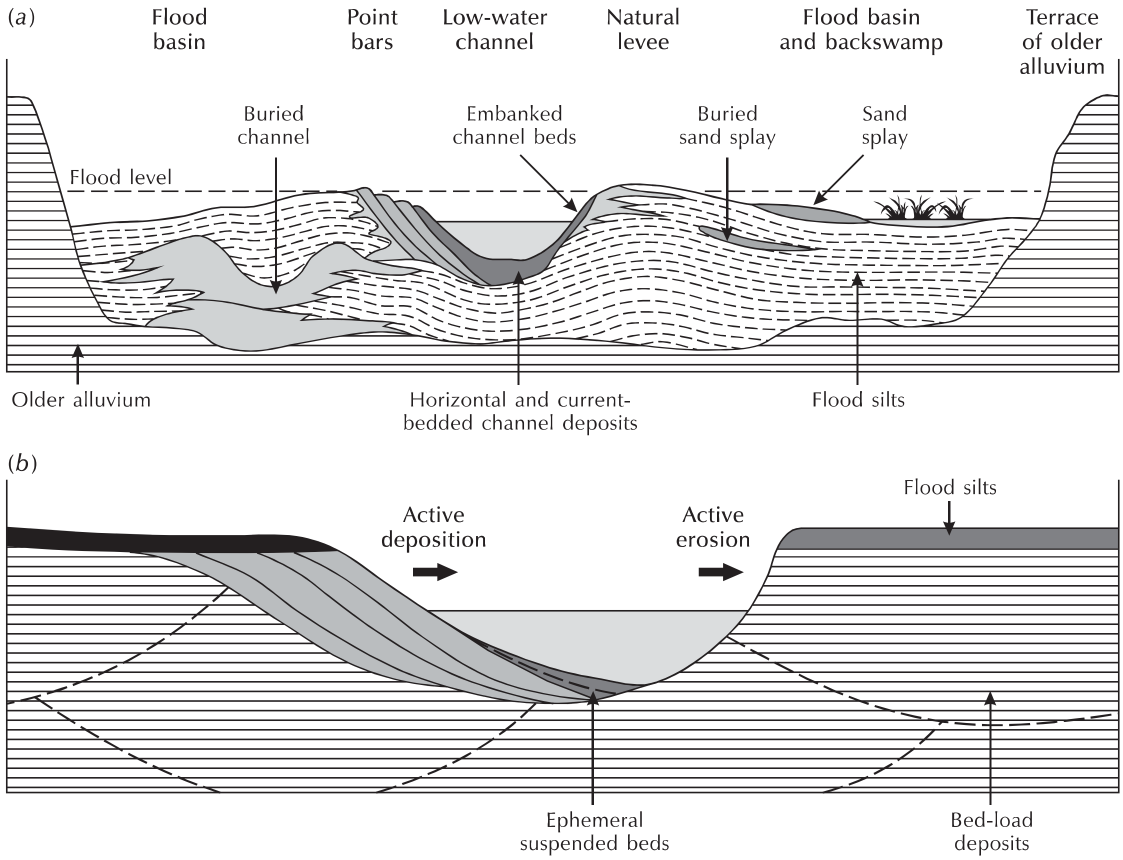

Figure 9.8 Sections through floodplains. (a) A convex floodplain. Point-bar deposits occur on inside meander bends and rarely opposite developing levees. The vertical exaggeration is considerable. (b) A flat floodplain.
Source: After Butzer (1976, 155, 159)

River terraces

A **terrace** is a roughly flat area that is limited by sloping surfaces on the upslope and downslope sides. River terraces are the remains of old valley floors that are left sitting on valley sides after river downcutting. Resistant beds in horizontally lying strata may produce flat areas on valley sides – **structural benches** – so the recognition of terraces requires that structural controls have been ruled out. River terraces slope downstream but not necessarily at the same grade as the active floodplain. **Paired terraces** form where the vertical downcutting by the river is faster than the lateral migration of the river channel (Figure 9.9a). **Unpaired terraces** form where the channel shifts laterally faster than it cuts down, so terraces are formed by being cut in turn on each side of the valley (Figure 9.9b).

The floor of a river valley is a precondition for river terrace formation. Two main types of river terrace exist that correspond to two types of valley floor: bedrock terraces and alluvial terraces.

Bedrock terraces

Bedrock or **strath** terraces start in valleys where a river cuts down through bedrock to produce a V-shaped valley, the floor of which then widens by lateral erosion (Figure 9.10). A thin layer of gravel often covers the flat, laterally eroded surface. Renewed downcutting into

Plate 9.7 Alluvial fan at base of 200-m-high ridge in the Himalayan Foothills, Tibet.
(*Photograph by Tony Waltham Geophotos*)

Plate 9.8 Playa in Panamint Valley, California, USA. A bajada can be seen rising towards the mountains in the background.
(*Photograph by Tony Waltham Geophotos*)

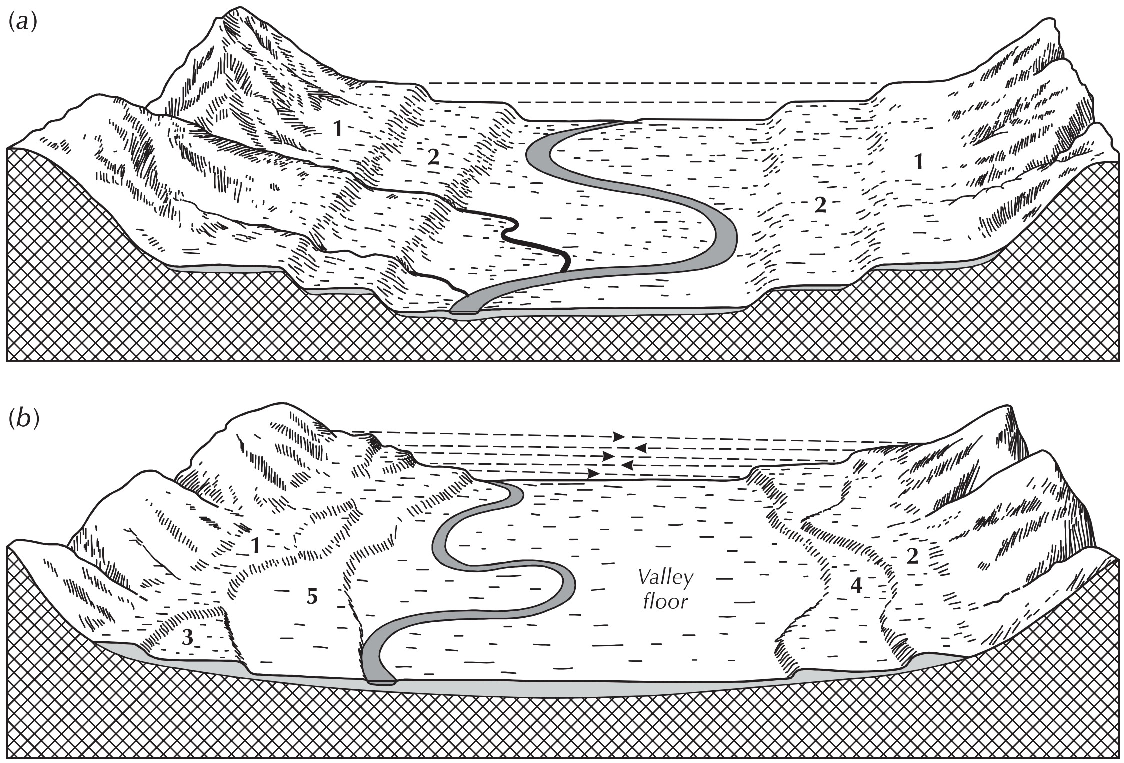

Figure 9.9 Paired and unpaired terraces. (a) Paired, polycyclic terraces. (b) Unpaired, noncyclic terraces. The terraces are numbered 1, 2, 3, and so on.
Sources: Adapted from Sparks (1960, 221–23) and Thornbury (1954, 158)

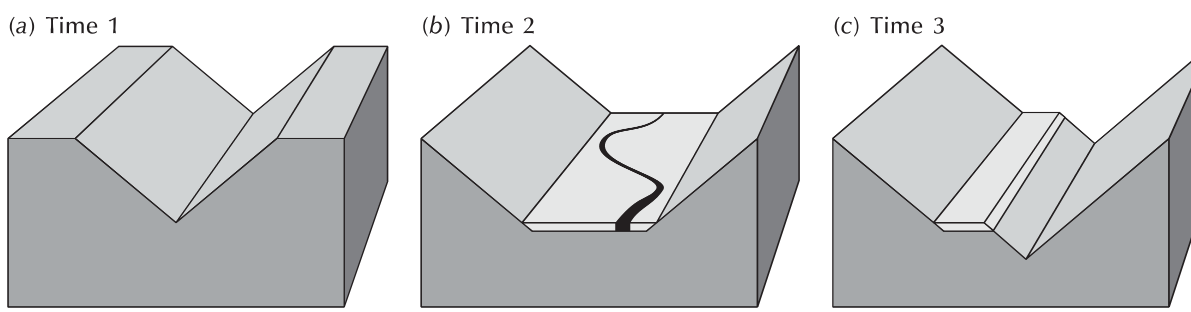

Figure 9.10 Strath (bedrock) terrace formation. (a) Original V-shaped valley cut in bedrock. (b) Lateral erosion cuts a rock-floored terrace. (c) Renewed incision cuts through the floor of the terrace.

this valley floor then leaves remnants of the former valley floor on the slopes of the deepened valley as rock-floored terraces. Rock-floored terraces are pointers to prolonged downcutting, often resulting from tectonic uplift. The rock floors are cut by lateral erosion during intermissions in uplift.

Alluvial terraces

Alluvial or **accumulation** terraces are relicts of alluvial valley floors (Plate 9.9). Once a valley is formed by vertical erosion, it may fill with alluvium to create a floodplain. Recommenced vertical erosion then cuts through the alluvium, sometimes leaving accumulation terraces stranded on the valley sides. The suites of alluvial terraces in particular valleys have often had complicated histories, with several phases of accumulation and downcutting that are interrupted by phases of lateral erosion. They often form a **staircase**, with each tread (a terrace) being separated by risers. A schematic diagram of the terraces of the upper Loire River, central France, is shown in Figure 9.11.

Terrace formation and survival

Four groups of processes promote river terrace formation: (1) crustal movement, especially tectonic and isostatic movements; (2) eustatic sea-level changes; (3) climatic changes; and (4) stream capture. In many cases, these factors work in combination. River terraces formed by stream capture are a special case. If the upper reach of a lower-lying stream captures a stream with a high baselevel, the captured stream suddenly has a new and lower baselevel and cuts down into its former valley floor. This is a one-off process and creates just one terrace level. Crustal movements may trigger bouts of downcutting. Eustatic falls of sea level may lead to headward erosion from the coast inland if the sea-floor is less steep than the river. Static sea levels favour lateral erosion and valley widening. Rising sea levels cause a different set of processes. The sea level rose and fell by over 100 m during the Pleistocene glacial–interglacial cycles, stimulating the formation of suites of terraces in many coastal European river valleys, for instance.

Climatic changes affect stream discharge and the grain size and volume of the transported load (Figure 9.12). The classic terrace sequences on Rivers Iller and Lech,

Plate 9.9 Alluvial terraces along the Broken River, Castle Hill, New Zealand. (*Photograph by David Knighton*)

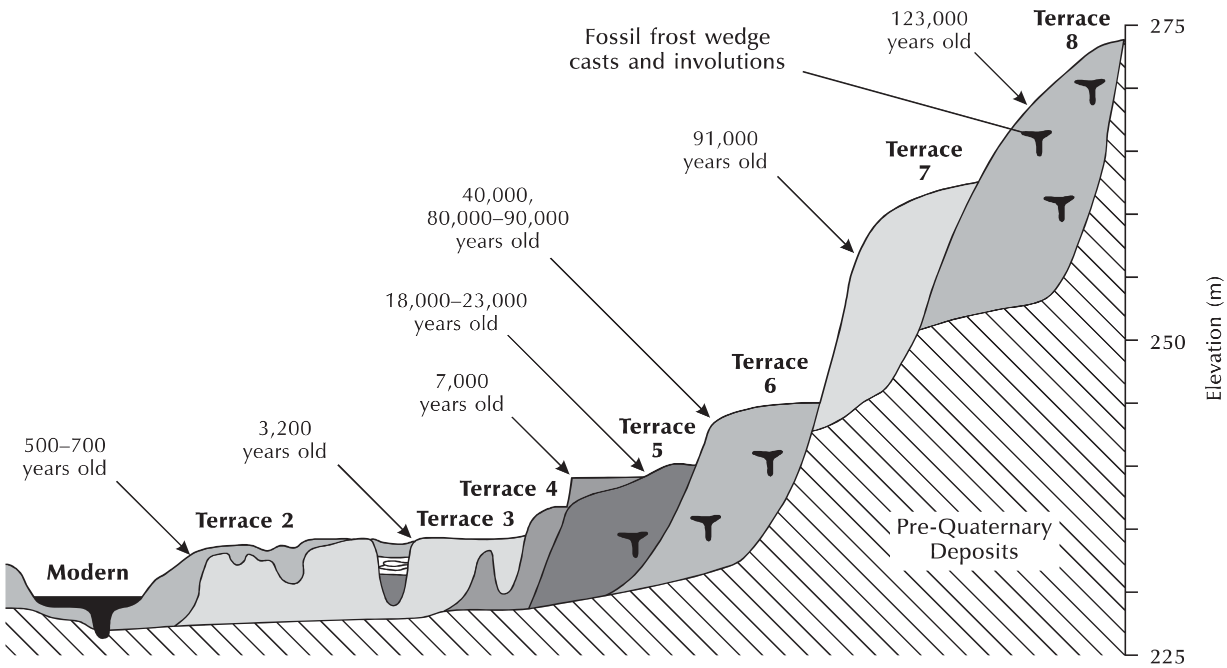

Figure 9.11 Terraces on the upper Loire River, France (diagrammatic).
Source: Adapted from Colls *et al.* (2001)

in the Swabian–Bavarian Alpine foreland, are climatically controlled terraces produced as the climate swung from glacial to interglacial states and back again. The rivers deposited large tracts of gravel during glacial stages, and then cut into them during interglacial stages. Semi-arid regions are very susceptible to climatic changes because moderate changes in annual precipitation may produce material changes in vegetation cover and thus a big change in the sediment supply to streams. In the south-west USA, arroyos (ephemeral stream channels) show phases of aggradation and entrenchment over the last few hundred years, with the most recent phase of entrenchment and terrace formation lasting from the 1860s to about 1915.

Terraces tend to survive in parts of a valley that escape erosion. The slip-off slopes of meanders are such a place. The stream is directed away from the slip-slope while it cuts down and is not undercut by the stream. Spurs at the confluence of tributary valleys also tend to avoid being eroded. Some of the medieval castles of the middle Rhine, Germany – the castles of Gutenfels and Maus, for example – stand on small rock-floored terraces protected by confluence spurs on the upstream side of tributary valleys.

Lacustrine deltas

Lacustrine or **lake deltas** are accumulations of alluvium laid down where rivers flow into lakes. In moving from a river to a lake, water movement slows and with it the water's capacity and competence to carry sediment. Providing sediment is deposited faster than it is eroded, a lacustrine delta will form.

HUMAN IMPACTS ON THE FLUVIAL SYSTEM

Human agricultural, mining, and urban activities have caused changes in rivers. This section will consider three topics: the increased flux of fluvial sediments; the effect of dams on streamflow, sediment transfer, and channels; and river modification and management.

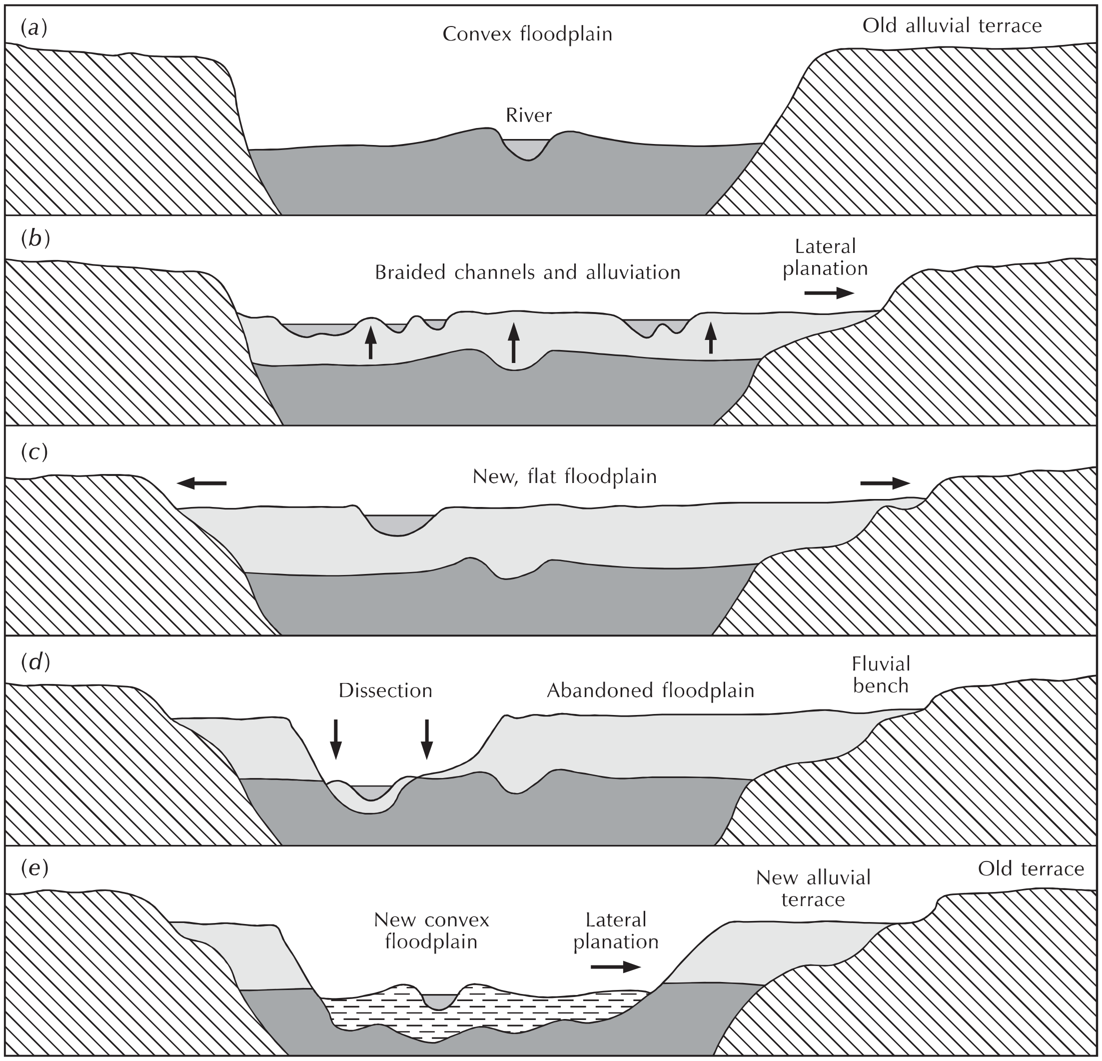

Figure 9.12 Alluvial terrace formation. (a) An initial convex floodplain. (b) Burial of the initial floodplain by coarser sediments through rapid alluviation of braided channels. (c) A stable, flat floodplain forms by alluviation and some lateral planation. (d) Another environmental change leads to dissection of alluvium and the abandonment of the flat floodplain. (e) A new convex floodplain is established by the alluviation of fine sediments and lateral planation.
Source: After Butzer (1976, 170)

River sediment increase

In North America, agricultural land-use typically accelerates erosion tenfold to a hundredfold through fluvial and aeolian processes. Much of this high sediment yield is stored somewhere in the river system, mainly in channels, behind dams, and as alluvium and colluvium. Many other reports in the literature support this conclusion. With the maturation of farmlands worldwide, and with the development of better soil conservation practices, it is probable that the human-induced erosion is less than it was several decades ago (e.g. Trimble 1999). Overall, however, there has been a significant anthropogenic increase in the mobilization of sediments through fluvial processes. Global estimates of the quantities vary considerably: one study gave a range of 24–64 billion tonnes per year of bulk sediments, depending on the scenario used (Stallard 1998); another study calculated that as much as 200 billion tonnes of sediment move every year (Smith *et al.* 2001).

River channels and dams

Dams impose changes in streamflow and the transfer of sediment. A study of the impacts of 633 of the world's largest reservoirs (with a maximum storage capacity of 0.5 km^3 or more), and the potential impacts of the remaining >44,000 smaller reservoirs reveals the strong influence of dams on streamflow and sediment flux (Vörösmarty *et al.* 2003). It uses the residence time change (the time that otherwise free-flowing river water stays in a reservoir), in conjunction with a sediment retention function, as a guide to the amount of incoming sediment that is trapped. Across the globe, the discharge-weighted mean residence time change for individual impoundments is 0.21 years for large reservoirs and 0.011 years for small reservoirs. The large reservoirs intercept more than 40 per cent of global river discharge, and approximately 70 per cent of this discharge maintains a theoretical sediment-trapping efficiency in excess of 50 per cent. Half of all discharge entering large reservoirs shows a local sediment trapping efficiency of 80 per cent or more. Between 1950 and 1968, global sediment trapping in large reservoirs tripled from 5 per cent to 15 per cent; it doubled to 30 per cent between 1968 and 1985, but then stabilized. Several large basins such as the Colorado and Nile show almost complete trapping due to large reservoir construction and flow diversion. From the standpoint of sediment retention rates, the most heavily regulated drainage basins lie in Europe. Large reservoirs also strongly affect sediment retention rates in North America, Africa, and Australia–Oceania. Worldwide, artificial impoundments potentially trap more than 50 per cent of basin-scale sediment flux in regulated basins, with discharge-weighted sediment trapping due to large reservoirs of 30 per cent, and an additional contribution of 23 per cent from small reservoirs. Taking regulated and unregulated basins together, the interception of global sediment flux by all 45,000 registered reservoirs is at least 4–5 billion tonnes per year, or 25–30 per cent of the total. There is an additional but unknown impact due to the still smaller 800,000 or so unregistered impoundments. The study shows that river impoundment is a significant component in the global fluxes of water and sediment.

Changes in streamflow and sediment transfer caused by dams lead to downstream changes in channel form. The degradation of rivers downstream of dams is a concern around the world. It has proved difficult to generalize about responses of channels downstream of dams. Figure 9.13 displays expected responses over a timescale of about 50 years to a reduction in sediment load (Figure 9.13a) and a reduction in flood magnitude (Figure 9.13b). Figure 9.13c shows the special case in which a tributary confluence is involved. In all cases, a change in a single process may produce any one of four channel responses.

River modification and management

Fluvial environments present humans with many challenges. Many European rivers are complex managed entities. In the Swiss Jura, changes in some rivers to improve navigation destabilized the channels and a second set of engineering works was needed to correct the impacts of the first (Douglas 1971). Within the Rhine Valley, the river channel is canalized and flows so swiftly that it scours its bed. To obviate undue scouring, a large and continuous programme of gravel replenishment is in operation. The Piave river, in the eastern Alps of Italy, has experienced remarkable channel changes following decreased flows and decreased sediment supply (Surian 1999). The width of the channel has shrunk to about 35 per cent of its original size, and in several reaches the pattern has altered from braided to wandering. In England, the channelization of the River Mersey through the south of Manchester has led to severe bank erosion downstream of the channelized section, and electricity pylons have had to be relocated (Douglas and Lawson 2001).

By the 1980s, increasing demand for environmental sensitivity in **river management**, and the realization that hard engineering solutions were not fulfilling their design life expectancy, or were transferring erosion problems

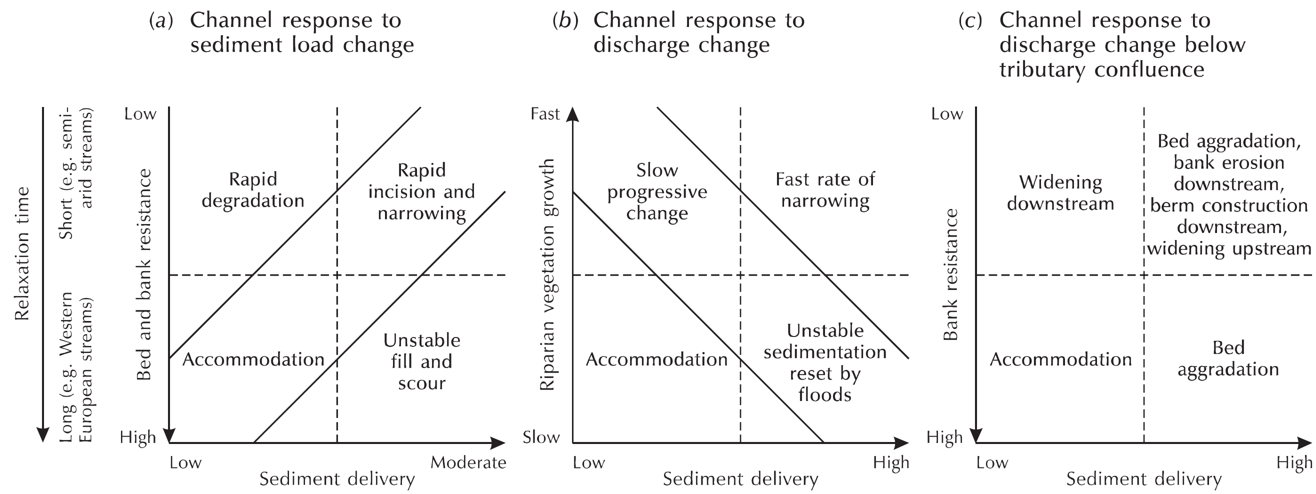

Figure 9.13 Domains of channel change in response to changing sediment load and discharge in different regions. Responses are to (a) a dominant reduction in sediment loads, (b) a dominant reduction in floods. (c) The special case of channel change below a tributary confluence in a regulated river dominated by flood reduction.
Source: Adapted from Petts and Gurnell (2005)

elsewhere in river systems, produced a spur for changes in management practices. Mounting evidence and theory demanded a geomorphological approach to river management (e.g. Dunne and Leopold 1978; Brookes 1985). Thus, to control bank erosion in the UK, two major changes in the practices and perceptions of river managers took place. First, they started thinking about bank erosion in the context of the sediment dynamics of whole river systems, and began to examine upstream and downstream results of bank protection work. Second, they started prescribing softer, more natural materials to protect banks, including both traditional vegetation, such as willow, osier, and ash, and new geotextiles to stimulate or assist the regrowth of natural plant cover (Walker 1999). River management today involves scientists from many disciplines – geomorphology, hydrology, and ecology – as well as conservationists and various user groups, such as anglers (e.g. Douglas 2000). Thus, in Greater Manchester, England, the upper Mersey basin has a structure plan that incorporates flood control, habitat restoration, and the recreational use of floodplains; while, in the same area, the Mersey Basin Campaign strives to improve water quality and river valley amenities, including industrial land regeneration throughout the region (Struthers 1997).

SUMMARY

Flowing water is a considerable geomorphic agent in most environments, and a dominant one in fluvial environments. It carves many erosional landforms, including rills and gullies, bedrock channels, and alluvial channels. River profiles, drawn from source to mouth, are normally concave, although they often possess knickpoints marked by steeper gradients. Rivers form networks that may be described by several geometrical and topological properties. Valleys are an overlooked erosional landform. Flowing water deposits sediment to build many depositional landforms. The smallest of these are features on channel beds (riffles and dunes, for example). Larger forms are floodplains, alluvial fans, playas, river terraces, and lake deltas. Flowing water is sensitive to environmental change, and especially to changes of climate, vegetation cover, and land-use. Many river valleys record a history of changing conditions during the last 10,000 years, induced by changing climates and changing land-use, that have produced adjustments in the fluvial system. Human agricultural, mining, and urban activities cause changes in rivers. Overall, they increase the flux of fluvial sediments. Dams affect streamflow, sediment transfer, and channel form downstream. Human actions modify many rivers, which need managing. Fluvial geomorphology lies at the heart of modern river management.

ESSAY QUESTIONS

1. **How would you convince a sceptical friend that rivers carved the valleys they flow through?**
2. **Why do river channel patterns vary?**
3. **To what extent have humans modified fluvial landscapes?**

FURTHER READING

Acreman, M. (2000) *The Hydrology of the UK: A Study of Change*. London: Routledge.
Not strictly geomorphology, but highly relevant to the subject.

Bridge, J. S. (2003) *Rivers and Floodplains: Forms, Processes, and Sedimentary Record*. Oxford: Blackwell Science.
A useful text for more advanced readers.

Brookes, A. J. and Shields, F. D. (1996) *River Channel Restoration: Guiding Principles for Sustainable Projects*. Chichester: John Wiley & Sons.
If you are interested in applied fluvial geomorphology, try this.

Jones, J. A. A. (1997) *Global Hydrology: Process, Resources and Environmental Management*. Harlow, Essex: Longman.
Gives a hydrological context for fluvial processes.

Knighton, A. D. (1998) *Fluvial Forms and Processes: A New Perspective*, 2nd edn. London: Arnold.
A top-rate book on fluvial geomorphology.

Kondolf, M. and Pigay, H. (2002) *Methods in Fluvial Geomorphology*. New York: John Wiley & Sons.
Discusses an integrated approach to river restoration.

Leopold, L. B., Wolman, M. G., and Miller, J. P. (1964) *Fluvial Processes in Geomorphology*. San Francisco, Calif., London: W. H. Freeman. (Published by Dover Publications, New York, 1992.)
The book that process geomorphologists used to rave about. Worth dipping into but not always easy reading.

Robert, A. (2003) *River Processes: An Introduction to Fluvial Dynamics*. London: Arnold.
A very good treatment of physical processes in alluvial channels.

Thorne, C. R., Hey, R. D., and Newson, M. D. (1997) *Applied Fluvial Geomorphology for River Engineering and Management*. Chichester: John Wiley & Sons.
Another book that considers applied aspects of the subject.