Chapter 2: FLUVIAL GEOMORPHOLOGY

Geomorphology deals with the reconstruction of the history of the surface of the earth through a study of its forms, the materials of which it is made up of and the processes that shape it.

Landforms are small to medium tracts or parcels of the earth’s surface with a distinct shape and size and which are made up of specific materials through the action of specific geomorphic agents or media (like wind, surface water, groundwater, waves and glaciers) and a series of geomorphic processes (like weathering, erosion & deposition). The shaping of each landform commences at a distinct point in time, but is a dynamic process, as each landform is continually undergoing changes in form and size through (usually) very slow geomorphic actions and processes. Following the processes of weathering or breaking down of the lithosphere, the geomorphic agents erode or carry away the weathered products. Erosion therefore causes changes on the earth’s surface and is followed by deposition, the laying down of weathered (and eroded) material transported by the geomorphic agent, which is likewise responsible for changes in the landscape.

Landscapes, on the other hand, are a collection of several related landforms covering larger parcels of the earth’s surface. The evolution of the landscape describes the transformation of a part of the earth’s surface from one landform into another or the change of individual landforms once they have been formed. Landforms are modified by changes in the processes or intensities of tectonic movement of landmasses (vertically or horizontally) and also by climatic change. Each landform therefore has a history of development and change over time, which can be likened to the stages of life — youth, maturity and old age.

Changes on the surface of the earth are attributed mainly to erosion by various geomorphic agents like water, waves, wind and ice, followed by deposition, which alters the landscape by covering the land surfaces and filling basins and depressions. These depositional surfaces themselves are again subjected to weathering and mass wasting processes and subsequent erosion which shape and modify the landscape. Acting over long periods of time, these geomorphic agents produce systematic changes leading to sequential development of landforms, with each agent and process producing its own assemblage of landforms and leaving its distinct imprints on the landforms they create. The study of landforms should accordingly offer clues as to the agents and processes responsible for their formation.

The actions of geomorphic agents are usually classified according to one of two types of landforms created by them, namely erosional or destructional and depositional or constructional landscape-forming processes.

Controls determining structure and morphological properties

The type of landform that will develop as a result of the action of each geomorphic agent depends mainly on the geology, especially the rock and mineral type, including hardness, permeability and the presence of fractures, joints, folds, and so on. The following independent controls also influence the evolution of landforms:

1. stability of sea level;
2. tectonic stability of landmasses;
3. climate.

A disturbance in any of these three controlling factors can upset the systematic and sequential stages in the development and evolution of landforms.
The River or Fluvial System

Rivers are found in many different climatic zones, ranging from humid to arid, and from equatorial to arctic. Some of the larger rivers even flow across different climatic zones, such as the Nile and Colorado, both of which sustain agriculture and urban centres in desert regions, but which have their origins in more humid climates. They cut their channels into a range of bedrock and alluvial substrates.

Natural streams are open hydraulic systems in equilibrium.

Fluvial system controls

In systems terminology controls are factors that influence, or control, system. In river systems, climate, geology, vegetation cover and topography are some of the controls, while a very 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.

System variables

Fluvial system variables include hillslope angle, flow discharge and sediment transport rates. These are all quantities that change through time. Some variables (called controlling variables) act to control others (called adjustable variables). The ultimate controls are the external basin controls (climate, tectonics, base level and human activity). These influence the entire fluvial system. Complex interactions occur between internal variables as a result of feedbacks. Negative feedbacks act to mitigate the effects of a disturbance to the system, allowing a state of equilibrium to be maintained. Positive feedbacks lead to instability, exacerbating the effects of disturbances. This involves the crossing of one or more thresholds, causing the system to move towards a new equilibrium.

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.

Each river drains an area of land called its drainage basin – also known as its catchment or watershed – which supplies water and sediment to the channel (see figure 2.1). The basic unit of study in fluvial systems is the drainage basin.

Latorita River, tributary of the Lotru River

(Drainage basin)

Figure 2.1 : Catchment or drainage basin with the drainage divide (or watershed) as perimeter.

The catchment is bounded by a drainage divide, catchment boundary, or watershed - which is clearly visible as a ridge in areas of high relief, but more difficult to discern in flatter topographies. Drainage basins form a mosaic on the landscape, varying in size from a few hectares to millions of square kilometres. The outlet where the main channel exits the basin is at a lower elevation than the rest of the basin.
Stream order

Within catchments, rivers form branching networks of channels which drain runoff, transporting water and sediment from the land area to the basin outlet and which can be described by several geometrical and topological properties. The concept of stream order provides a method to classify or order the hierarchy of natural channels within a catchment.

Strahler's stream order classifies the hierarchy of streams in a catchment by regarding each stream segment of a river network as a branch in a tree, with the next segment downstream as its parent. The smallest, or primary, tributaries are assigned the order of “1” and where two first-order streams come together, they form a second-order stream. At the confluence of two second-order streams, they form a third-order stream. (see figure 2.2).

Streams of lower order joining a higher order stream do not change the order of the higher stream. Thus, if a first-order stream joins a second-order stream, it remains a second-order stream. It is not until a second-order stream combines with another second-order stream that it becomes a third-order stream.

Stream order correlates well with drainage area, but is also controlled on a regional scale by topography & geology. As streams increase in order, they also increase in length, exponentially. Worldwide, about 70-75% of stream kilometres occur as headwater (first order) streams.

Drainage patterns.

These systems commonly display distinct drainage patterns that often reflect the structure of underlying folded sedimentary beds (such patterns are called concordant). The main, or trunk, channel is fed by numerous smaller tributaries which join to form progressively larger channels. Drainage patterns, as viewed from the top, vary considerably between drainage basins. Where these patterns correlate to the geological structure and relief of the landscape over which it flows, it is termed accordant, as displayed in figure 2.3, whereas discordant drainage patterns are those which do not correlate to the underlying topography or geology.

<table>
<thead>
<tr>
<th>Deranged</th>
<th>Dendritic: (tree-like)</th>
<th>Parallel:</th>
<th>Trellis</th>
</tr>
</thead>
<tbody>
<tr>
<td>No coherent pattern evident in the river systems, in recent geologic formations where the catchments are still being shaped and not yet in equilibrium.</td>
<td>The most common form of drainage pattern with many contributing tributaries. Develop where the river channel follows the slope of the terrain.</td>
<td>Streams all flow in the same direction, parallel, usually caused by steep slopes with some relief on uniformly sloping surfaces.</td>
<td>The drainage pattern looks like a garden or vine trellis, typically found in fold mountains.</td>
</tr>
</tbody>
</table>

Figure 2.2: Strahler's hierarchical ranking of stream order. The smallest stream has order 1 and at every confluence of similar order, the downstream order increase by 1.
Rectangular
Streams consist mainly of straight line segments with tributaries and bends joining at 90 degrees. Caused by preferential erosion of joints.

Radial
Streams radiate outwards from a central high point, typically found on volcanoes.

Centripetal
When runoff drains to a central depression or pan, usually forming a pan or lake (e.g. salt flats). Typical of endorheic systems or regions.

Annular
Stream direction in concentric path, following courses of less resistant rock, resembling tree rings in cross-section. In rimming sedimentary strata of differing hardness.

Figure 2.3: Accordant drainage patterns

The evolution of drainage networks is influenced by a number of factors like geology, climate and the long-term history of the catchment. Even along one river, differences in the form, cross-section and alignment of channels can be seen, especially variations in size, ranging from the smallest headwater streams of a few centimetres wide to large rivers several hundred metres or more in width.

The size of a river channel at a given point is largely determined by the upstream supply of water, or stream discharge. This is the volume of water that passes through a given channel cross-section in a given period of time. In the upper reaches of a river, the area drained – and hence the discharge – is relatively small.

Discharge (Q):
The product of the channel width, depth, and flow velocity

\[ Q = v w d \]

where:
- \( Q \) = Flow rate (m³/s)
- \( v \) = Ave. velocity of flow at a cross-
- \( w \) = width
- \( d \) = depth

This can also be generalized to the general flow equation

\[ Q = va \]

where \( a \) = area of the cross-section (m²)

Other stream dimensions:
- \( A_2 \) = cross-sectional area
- \( P \) = wetted perimeter

Active stream channels change over time, eroding and depositing sediments in predictable patterns that are apparent in stream cross sections.

Scarp – active, eroding side of stream channel. Also known as bank.

Thalweg – path of deepest channel

Bankfull Discharge - Typically bankfull discharge equates to a roughly 2-year recurrence interval flow

Figure 2.4: Cross-section of channel system, showing some important dimensions to determine discharge

Figure 2.5: Cross section of a stream channel. The thalweg is the deepest part of the channel.
Discharge and channel size generally increase in proportion to the upstream drainage area as the river flows downstream. While many rivers follow a single channel, multiple channels are also frequently found. Those that flow in a single channel usually tend to deviate from a straight line, typically following an irregular path, or else more regular meanders.

Flowing water deposits sediment to build many depositional landforms, which, on the smallest scale, are riffles and dunes on the channel beds, while larger forms include floodplains, alluvial fans, river terraces, and lake deltas.

The land surface water cascade system

The gravitational flow of water is the main surface system contributing to landform formation. Surface water cascades from one morphological subsystem of the land surface system to another. In each morphological subsystem, flow is regulated by variables (morphological variables) like elevation, slope, rock structure, etc. The gravity flow of water on the land surface system forms a hierarchical cascade system linking the many morphological subsystems.

The land surface water cascade system is organised in drainage systems of a given hierarchy (order or ranking): therefore a 1st order stream has small catchments; 2nd order bigger ones consisting of the smaller order catchments, and so on. Watersheds are the boundaries between such drainage systems.

![Figure 2.6: The drainage basin or catchment as unit of investigation in fluvial morphology](image)

**Drainage basins and Interfluves**

A drainage system is a clearly definable subsystem of the land surface system, drained according to a particular pattern by a specific linked network of streams. Each drainage system is an open process-response system.

**TYPES OF SYSTEM WITHIN FLUVIAL MORPHOLOGY**

A system is a group of related objects that are linked by processes. Three types of system can be identified in fluvial geomorphology: morphological systems, cascading systems & process-response systems. Process-response systems describe the interactions between the processes (flows of energy and materials) of the cascading system and the forms of the morphological system. We will now investigate each of these three types of system more closely.

**Morphological (form) systems**

Landforms such as channels, hillslopes and floodplains form a morphological system. The form of each component of a morphological system is related to the form of the other components in the system. For example, if the streams in the headwaters of a drainage basin are closely spaced, the hillslopes dividing them are steeper than they would be if the streams were further apart from each other. Sub-systems
could be the soil infiltration rate, drainage density etc. The morphological variables are those that govern discharge and are:

- CHANNEL WIDTH
- BOUNDARY ROUGHNESS
- SIZE & CONCENTRATION OF SEDIMENT LOAD
- DEPTH & SLOPE of CHANNEL elevation, slope, rock structure

If any of these interdependent variables change, it must be compensated for by an adjustment in the other variables.

**Cascading (process) systems**

The components of the morphological system are linked by a cascading system, which refers to the flow of water and sediment through the morphological system. Cascading systems are also called process systems or flow systems. These flows follow interconnected pathways from hillslopes to channels and through the channel network.

**Process–response systems**

The two systems (Morphological & Cascading) interact as a process–response system. Process-response describes the adjustments between the processes of the cascading system and the forms of the morphological system. There is a two-way feedback between process and form; in other words, processes shape forms and forms influence the way in which processes operate (rates and intensity). For example: a steep section of channel causes high flow velocities and increased rates of erosion, and over time erosion is focused at this steep section and the channel slope is reduced. Velocity decreases as a result, reducing rates of erosion.

**INPUTS, OUTPUTS AND STORES** of the THE FLUVIAL SYSTEM

- The basic unit of the fluvial system is the drainage basin.
- Fluvial systems are open systems, which means that energy and materials are exchanged with the surrounding environment.
- The inputs to and outputs from the system are energy and matter.

**INPUTS**

**Matter**

The main inputs are water from rainfall and runoff and sediment derived from the breakdown of the underlying rocks. Additional inputs consist of the biological material and solutes derived from the breakdown of organic material from atmospheric inputs and rock weathering.

**Energy**

Most of the energy required to drive the system is provided by the atmospheric processes that lift and condense the water that falls as precipitation over the drainage basin. Gravity then moves this water downslope, creating a flow of energy through the system. This energy is expended in moving water and sediment to the river channels network.

**OUTPUTS**

**Matter**

Water & sediment move through the system to the drainage basin outlet, and material is discharged to the ocean. Not all rivers reach the ocean; but some flow into inland lakes and seas, dry up or recharge groundwater, before reaching the ocean (e.g. Okavango River in Botswana). Such river systems are termed endorheic. This reflects another important output from fluvial systems: the loss of water by evaporation to the atmosphere.
Energy
Most of the available energy is used in overcoming the considerable frictional forces involved in moving water and sediment from hillslopes into the channel network. A lot of this energy is ‘lost’ to the atmosphere in the form of heat.

STORES
Some material is stored along the way, like water which is stored for varying periods in lakes and reservoirs, and below the ground in the soil (as soil water) and in aquifers as groundwater. Sediment is stored when it is deposited in channels, lake basins, deltas, alluvial fans and on floodplains & dams (reservoirs). Water could be released from storage at a later stage, e.g. when a channel migrates across its floodplain, eroding into formerly deposited sediments which are then carried downstream, in an intermittent fashion from season to season until it reaches the sea.

A RIVER SYSTEM CAN BE DIVIDED INTO THREE MORPHOLOGICAL SUBSYSTEMS:

The fluvial system can be divided into smaller, sub-systems on the basis of their morphological character, in order to examine its components in more detail. One way of doing this is to consider the system in terms of three zones, each of which is a process–response system with its own inputs and outputs. Such a division of the fluvial system on the basis of dominant processes operating within each zone is then the sediment production zone, sediment transfer zone and sediment deposition zone.

![Figure 2.7: Morphological subsystems of the fluvial system](image)

Certain processes dominate within each zone:

**The sediment production zone:**
Most of the sediment originates in the headwaters and is supplied to the channel network from the bordering hillslopes by erosion and mass movement of weathered rock material. The collecting system (branches) -- consists of a network of tributaries in the headwater region, which collects and funnels water and sediment to the main stream. The production zone has a predominance of erosional processes, and the production of sediment is the defining characteristic.

**The sediment transfer zone**
This sediment is then moved through the channel network in the sediment transfer zone, where the links between the channel and bordering hillslopes, and hence sediment production, are not so strong. This portion of the channel can also be regarded as the transporting system (trunk) -- the main trunk stream, which functions as a channelway through which water and sediment move from the collecting area toward the ocean. (Erosion and deposition also occur in a river’s transporting system).
Sediment deposition zone
As the river approaches the ocean, its gradient declines and the energy available for sediment transport is greatly reduced in the sediment deposition zone. It is mainly the finest sediment that reaches the ocean, as coarser sediment tends to be deposited further upstream. Only a certain proportion of all the sediment that is produced within a drainage basin actually reaches the basin outlet...This portion of the stream channel can also be regarded as the dispersing system (roots) – which consists of a network of distributaries at the mouth of a river (delta), where sediment and water are dispersed into an ocean, a lake, or a dry basin. Depositional processes dominant.

RUNNING WATER: CASCADING OF MATTER & ENERGY THROUGH THE FLUVIAL SYSTEM
Flowing water is the most important agent of erosion in most environments on the continents and stream valleys are the most common landforms shaped by these erosional processes, with rivers flowing to the oceans draining about 68 % of the Earth's land surface. The remainder of the land either is covered by ice or drains to closed or inland drainage basins. Rivers gradually sculpt the land surface, cutting through soil and rock, eroding away the surface material and forming channels, rills, streams and rivers. Streams are particularly effective landform-shaping agents, transporting material along their beds, while simultaneously keeping finer particles in suspension and carrying a load of dissolved substances. Streams wear away their channels and beds by corrosion, corrasion, and cavitation, eroding both downwards (vertically) and sideways (laterally). They deposit sediments downstream from the source as channel deposits, channel margin deposits, overbank floodplain deposits and valley margin deposits. Episodes of continued deposition and valley filling (alluviation) usually alternate with periods of erosion and valley cutting. Flowing water 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.

In climates characterised by sufficient rain, such as the humid and temperate regions, running water is considered the most important geomorphic agent in shaping the landscape. Running water on the land surface consists of two components, namely overland flow over the general land surface in a sheet formation, and channel flow as streams and rivers in valleys. Let us first look at these two types of flow. Overland flow or sheetflow follows after precipitation and infiltration: once the soil profile is saturated from rain water, water starts flowing overland in thin sheets and over a broad front, as indicated in figure 2.8.

![Figure 2.8](image)

Figure 2.8: Sheetflow occurs when the soil becomes saturated and rainfall can no longer infiltrate.

Sheet flow is initiated by friction between the flowing water and irregularities on the land surface; it is dynamically unstable, with small surface irregularities causing local concentrations of flow. Eddy formation and the diversion of flow around obstructions or into hollows initiate scouring, which results in small quantities of material being removed from the surface in the direction of flow. Overland flow therefore causes sheet erosion. The local (or micro-) relief of the land surface (i.e. distribution of and mix of particle size of the surface aggregate) causes overland flow to concentrate into narrow or wider pathways.
Overland flow or sheet flow is necessary for channel formation. The rate at which a channel forms depends on vegetation, storm intensity, slope, permeability, and erodibility of the surface.

In the early stages, down-cutting dominates during which irregularities such as waterfalls and cascades will be removed. In the middle stages, streams cut their beds slower, and lateral erosion of valley sides becomes severe. Gradually, the valley sides are reduced to lower and lower slopes. The divides between drainage basins are likewise lowered until they are almost completely flattened leaving finally, a lowland of faint relief with some low resistant remnants called monadnocks standing out here and there. This type of plain forming as a result of stream erosion is called a peneplain (an almost plain).

Figure 2.10  Sheet, rill & gully formation and drainage as inputs to the channel system

Stream flow or channel flow

Stream flow or channel flow can be defined as any surface water flow confined to a channel. The photographs in figure 2.11 are all examples of channel flow.

Figure 2.11 : Channel flow in different sized channels

HOW RIVERS SHAPE THEIR CHANNELS

Most of the erosional landforms made by running water are associated with vigorous and youthful rivers flowing over steep gradients. With time, stream channels over steep gradients adopt gentler slopes due to continued erosion, and as a consequence, streams lose their velocity, which in turn facilitates active deposition. While there may be depositional forms associated with streams flowing over steeper slopes, this occurs on a much smaller scale compared to those associated with rivers flowing over medium to gentle slopes. The gentler the river channels in gradient or slope (i.e. the flatter their slope), the greater is the deposition.

Rivers and streams continuously shape and reform their channels through erosion of the channel boundary (bed and banks) and the reworking and deposition of sediments. For example, erosion and undermining of the banks can lead to channel widening. Scouring of the channel bed deepens the channel, while sediment deposition reduces the depth and can lead to the formation of channel bars. These are just some of the ways in which channel adjustment takes place.

Natural streams are open hydraulic systems in equilibrium. As water flows downstream, energy is used in transporting and rearranging materials in both the river channel and on the flood plain.
Therefore the following can occur:

- banks may erode,
- new channels may form
- old channels can be cut off, creating backwaters
- meanders can form and migrate seasonally

![Figure 2.12: Elements of three-dimensional channel form. The planform is the shape of the river viewed from above, the channel slope is shown for the reach between points A and B, and the channel cross-section between C and D.]

We identify three zones longitudinally in a river course, based on energy & morphology: (see figure 2.7):

- **Upper zone** -- erosion, sediment & nutrient supply
- **Lowland reaches** -- sediment & nutrient transfer
- **Estuaries / deltas** -- sediment & nutrient deposition

These three zones have the following characteristics:

**Upper zone**
- erosion, sediment & nutrient supply
- high bank and bed erosion;
- bedrock exposure
- high-gradient
- low temperatures,
- high oxygen levels
- fastwater habitats
- steeper gradients,
- inaccessible,
- conservation
- less pollution
- more pristine

**Lowland reaches** –
- sediment & nutrient transfer
Bordered by a wide floodplain
Redistribution of upstream sediment
Meanders form and migrate laterally
Fine sediments can be stored on the floodplain after a flood

**Estuaries / deltas** –
- sediment & nutrient deposition
  - Habitat characteristics of river determined by:
    - range of substrates found in the higher reaches (the ‘production’ and ‘transfer’ zones)
    - the hydrological regime (Rainfall frequency, Intensity, gradient, bed and bank morphology)

![Figure 2.13: Upper zone, lowland and delta portions of a stream]
The form of a given reach of stream channel is controlled by the supply of water (flow) and also the supply of sediment to its upper reaches. Other significant controls are valley width, the channel substrate, valley slope and riparian vegetation. These controls all vary, both among streams and along the same reach within a river, creating a wide range of fluvial environments and resultant channel forms. The three-dimensional shape of a river is described in terms of its planform, slope and cross-sectional shape.

Rivers continuously adjust their channels in response to fluctuations in flow and sediment supply. An important balance exists between the erosive force of the flow and the resistance of the channel boundary to erosion. Major flood events can cause significant changes to the channel form due to the increased erosional energy of the water. The resistance offered by the bed and banks determines how extensive these changes will be: for instance, channels in unconsolidated alluvium offer much less resistance to erosion than those cut in bedrock, while most flows can shape channels in sandy alluvium due to the relatively little energy required to mobilise individual sand grains. Despite their smaller size, silts and clays tend to stick together as a result of cohesive (attractive) electrochemical forces between the particles and offer more resistance to erosion than sand grains; it is for this reason that channels with a high silt and clay proportion are actually more resistant to erosion than those formed in sand and fine gravel. The role played by riparian vegetation is also significant, since this offers additional resistance to erosion. For the most, it is only extreme floods that are capable of substantially modifying bedrock channels and therefore channel adjustments tend to occur sporadically.

Alluvial channels dominated by cobbles and boulders may also be relatively unaffected by most flows, which are not powerful enough to move such coarse material. The energy needed to do geomorphological work is provided by the flow of water through the channel. For any length of channel, energy availability is dependent on two things: the flow discharge and the steepness of the channel slope. Increases in either of these will increase the stream power and therefore the potential to carry out geomorphological work.

But before sediment transport and erosion can occur, a very big amount of energy is required to merely move water in the channel, due to various types of resistance to flow, such as friction between the flowing water and the channel boundary. An estimated 95 per cent of a river’s energy is used in overcoming flow resistance, leaving only 5 per cent to do geomorphological work. Friction can be particularly high in rough, boulder bed channels but is also significant for channels formed in finer substrates. Energy is also spent when the channel is confined between valley walls, around bends and when cascading over knickpoints, rapids and waterfalls. Friction is even generated within the flow itself as a result of eddies and turbulence.

Flow and sediment supply

The flow in natural channels constantly fluctuates through a continuous series of normal flows, floods and droughts and sediment supply likewise varies through time. Rivers consequently adjust their form continuously in response to these fluctuations, which in turn influences the flow of water and sediment transport through the channel. Because the flow of water in a river provides the energy required to shape the channel, the characteristics of that flow are very important in determining channel form.

The mean discharge usually increases with the size of the upstream drainage area, but the mean discharge does not reflect the way in which flow varies through time. These variations are described by the flow regime, which can be thought of as the ‘climate’ of a river. Characteristics of the flow regime include seasonal variations in flow and the size and frequency of floods.

Processes of erosion, transport and deposition within a channel reach are influenced by the supply of sediment at the upstream end as well as sediment that is locally eroded from the bed and banks. It is not only the volume of sediment that is important, but also its size distribution. Processes of sediment transport are very different for coarse and fine sediment, so sediment supply has an important influence on channel form and behaviour.
Stream Longitudinal Profile

Longitudinal profile reflects downstream trade-off between discharge and slope in setting transport capacity (and thus ability to move sediment and incise rock).

Figure 2.14: The longitudinal profile of a river

The limiting level below which a stream cannot erode the land is called the base level of the stream. Formally the base level is the level below which a river or stream cannot incise. When a stream flows into a lake, the surface of the lake acts as a local base level. The ultimate (or absolute) base level for most streams is global sea level. Exceptions are streams that drain into closed interior basins having no outlet to the sea. Where the floor of a tectonically formed basin lies below sea level (for example, Death Valley, California and the Dead Sea in Israel), the base level coincides with the basin floor.

A local change in base level affects river profiles: knickpoints. Knickpoints therefore occur where barriers to down-cutting exist and these temporary features usually only last as long as the barrier exists (see figure 2.16):
A knickpoint is defined as a location in a river where there is a sharp change in channel slope, such as at a waterfall or lake, resulting from differential erosion above and below the knickpoint. Once developed, a scoured trough or rill is perpetuated by the continued concentration of flow and migrates upstream by headward erosion of the knickpoint, the inflection where scouring is greatest. Figure 2.16 above, illustrates that knickpoints typically occur in the landscape where resistant rock layers occur. Knickpoints are usually associated with rapids or waterfalls.

SEDIMENT SOURCES - SUMMARY

Sediment is produced in the headwater regions of the source zone by processes of weathering, mass movement and erosion. Weathering involves the physical breakdown of rocks at the Earth’s surface and produces material called regolith. This is transported downslope, under the force of gravity, by processes of mass wasting. These include rapid mass movements, such as slides and debris flows, together with the much slower processes of creep and solifluction. Sediment is also produced by the erosive action of water, ice and wind.

Processes of water erosion include rain splash, sheetwash, rilling and gullying. Soil erosion is a natural process, but it can be accelerated by human activity, with rates of soil removal exceeding rates of soil formation. Accelerated soil erosion is a major environmental problem worldwide. In order to assess rates of soil loss, various monitoring techniques are used. Models have also been developed to simulate erosion and soil loss.

SEDIMENT TRANSFER

Sediment transfer from hillslopes to channels

The term primary erosion is used to describe the initial, or in situ erosion of rock, regolith and soil. It does not include the re-erosion of material that has been deposited, for example, at the base of a hillslope. Sediment that has been transported downslope and deposited on or at the base of slopes is called colluvium. There is an important linkage between the erosion of sediment from hillslopes and its transfer to channels and valley floors.
The effectiveness of this transfer is dependent on the degree of hillslope–channel coupling. In a coupled system there is a direct transfer of sediment from slopes to channels. This is typically the case in headwater regions, where narrow valleys are bordered by steep hillslopes. Coupling also occurs when a channel erodes the valley margin. Sediment transfer to channels is much more limited in decoupled systems. Further downstream, where valleys widen and channels are bordered by floodplains, sediment is stored at the base of hillslopes or on floodplain surfaces.

Figure 2.19: Sediment stores.

In decoupled systems, colluvial sediment makes only a very small contribution to the river’s sediment load. The degree of coupling varies according to sediment size. Finer sediment is more mobile and can be transported over greater distances. For a given slope–channel system, the degree of slope–channel coupling is often stronger for fine sediment than it is for coarse material. Some examples of the locations in which sediment is stored are illustrated schematically in Figure 2.19. These include hillslopes, alluvial fans, river channels, floodplains, deltas and lake bed deposits.

WATER EROSION ON HILLSLOPES

Water erosion of the soil surface is brought about by the action of falling raindrops and surface flow, which may move as a sheet across the surface or be concentrated in rills or gullies. Subsurface flow is also significant in hillslope erosion. Soil erosion provides the main source of the fine suspended sediment that is transported by river channels (clays to fine sands). Where flow is concentrated, larger material can also be transported, for example where deep gullies erode into the coarser subsoil.

There are many interrelated variables affecting rates of erosion, including climate, parent material, relief, tectonic setting, vegetation cover and human activity. For purposes of simplification, these can be considered in terms of the erosivity of the eroding agent and the erodibility of the soil surface (figure 2.20).

Erosivity is a measure of the capacity of an eroding agent, such as rainfall or overland flow, to erode the soil surface. It is dependent on the available kinetic energy (defined below), which is determined by factors such as rainfall intensity, raindrop size, flow depth and slope angle.
Erodibility refers to the susceptibility of the soil surface to erosion and is dependent on the properties of the soil itself, such as soil texture (relative proportion of sand, silt and cohesive clay particles). As you will see later in this chapter, erodibility is also dependent on the amount and type of vegetative cover and on land use practices.

**Rain splash erosion**

Falling raindrops possess kinetic energy, which is often enough to detach soil particles when striking the soil surface, although a large proportion of the energy is used in compacting the surface and creating an impact crater. A moving object’s (like a raindrop) kinetic energy is determined by its mass (which is proportional to the raindrop size) and its velocity, as shown below:

\[
\text{Kinetic energy (K)} = mv
\]

where \( m \) = mass and \( v \) = velocity.

The terminal velocity of an object is that velocity at which the gravitational force equals the drag (or frictional) resistance: a raindrop, like any other falling object, will therefore reach a terminal, or maximum constant velocity. Raindrops do not always reach their terminal velocity before striking the ground surface: factors such as wind speed and turbulence may increase or decrease their effectiveness when they land. Wind speed is therefore a control on the effectiveness of rain splash erosion, while another important control is the presence, percentage coverage and type of vegetation. The vegetation intercepts rainfall and breaks the fall of raindrop s, reducing their kinetic energy before they reach the ground surface. Rain splash erosion is therefore most effective in areas where vegetation does not entirely cover the ground surface. When raindrops fall on a sloping surface, there is a net movement of material down-slope, which increases with slope angle, due to gravity. Another important control factor is the soil type – with soils having a high silt/clay content offering more resistance to erosion than sandy soils, because of the cohesive nature of smaller particles.

**Sheetwash erosion**

When significant overland flow occurs, water flows over the surface in thin layers as so-called sheet flow. This is a somewhat misleading term because the flow is rarely of a uniform depth, being characterised by deeper, faster threads of flow that result from micro-scale variations in the surface topography. The down-slope flow of water exerts a shear stress on the soil surface. Erosion takes place when this stress is sufficient to overcome the resistance of the soil surface. Although the erosivity of the flow increases with depth and velocity, the shallow depth of overland flow and the roughness of the soil surface mean that the shear stress is not always sufficient to erode soil particles. As a result sheetwash is only really effective on steep slopes and smooth bare soil surfaces (Morgan, 2005). However, raindrop erosion is very effective as a detachment mechanism, allowing material to be entrained (set in motion). Since the transport of soil particles requires less energy than their initial entrainment, this material can be carried by the flow until it is deposited. The combined action of raindrop erosion and sheet erosion can therefore erode a significant volume of soil from large areas of sloping land. Sheetwash tends to be dominated by fine material with a diameter of less than 1 mm (Morgan, 2005), which contributes to the suspended load of rivers. Since soil is removed in thin layers, this type of erosion may go undetected for some time.

**Rills**

If the flow is sufficiently concentrated, a critical shear stress may be reached at which small micro-channels called rills start to form. Some well developed rills, formed in a road cutting, are shown in Plate 4.1. Rills vary in size with widths of between 50 mm and 300 mm and depths of up to 30 mm (Knighton, 1998). The critical conditions under which rills start to form can be considered in terms of a critical shear stress after Horton’s (1945) theory of slope erosion by overland flow. It should be noted that this applies mainly to sparsely-vegetated dryland environments where intense rainfall and overland flow occur on a fairly regular basis. The diagram in Figure 4.5 represents overland flow occurring on a slope. The depth of flow increases with distance from the drainage divide, as flow accumulates in a down-slope direction (this has been exaggerated for clarity in the diagram). Since shear stress increases with depth, there
must be a critical point on the slope at which the shear stress is great enough to allow incision to occur. This point is reached a **critical distance** (Xc) from the drainage divide, where the flow reaches a **critical depth** (d). Xc varies from slope to slope according to the balance between erosivity and erodibility. Above this point on the slope is a **belt of no erosion**, incision can occur below this point, and parallel rills start to form in the **belt of active erosion**. Further up the slope these features tend to be discontinuous and ephemeral, being destroyed by inter-rill erosion or wall collapse. The eroded sediment is carried down-slope by the flow, reducing the energy available for further incision. If the transport capacity of the flow is exceeded, deposition starts to occur in the form of small fan-shaped features. Horton called this the **zone of deposition**. In the field, these zones are not as clearly defined as might appear from the diagram because soils are typically very heterogeneous. Even at the micro-scale, there is considerable variation from place to place in slope, roughness, infiltration capacity, cohesiveness and other factors affecting erodibility.

As a result, complex spatial relationships exist between areas of erosion and areas of deposition. Rills may develop into more permanent features under favourable conditions. They are significant in the initiation of new stream channels when network extension occurs, or where a surface has recently been exposed, perhaps as a result of glacial retreat or volcanic eruption. In order for a permanent channel to form, a sufficient concentration of flow is required. This can happen when one rill becomes dominant over neighbouring rills, incising at a faster rate, concentrating flow at depth and leading to further incision. Even when permanent channels do not form, rill erosion, together with sheetwash and rainsplash erosion between rills (inter-rill erosion), results in the net removal of material from slopes. The concentrated flow in rills can transport larger soil particles, and even small rock fragments.

![Figure 2.21 Rills developed in a road cutting, South Africa.](image1)

![Figure 2.22 Horton's model of overland flow and rill formation. Adapted from Horton (1945).](image2)

Rills may account for much of sediment removal from a hillside, although this depends on the spacing of rills and the extent of the area affected (Morgan, 2005). This can lead to the loss of soil fertility and productivity when erosion proceeds more rapidly than soil formation. New soil is produced at a rate of a few millimetres a century, whereas a single storm can result in the removal of several centimetres of soil (Woodward and Foster, 1997).

![Figure 2.23: Examples of rill erosion](image3)
Eventually the rill grows into a gully, and if the rill **deepens to intersect the groundwater table** it will become a permanently flowing, or perennial stream. Rills develop into **wide and long gullies**, which increase in depth and width over time and distance, eventually merging to form a network of valleys.

**Gullies** are relatively permanent ephemeral channels. They are most commonly found in arid and semi-arid environments, where precipitation is highly seasonal and vegetation cover is sparse. Gullies are morphologically different to stream channels, being relatively deep and narrow, with steep sidewalls and a stepped channel slope. They typically range in size from depths of 0.5 m up to 25 m or 30 m (Soil Science Society of America, 1996) although there is no clearly defined upper limit, and the distinction between large gullies and some ephemeral stream channels can be somewhat vague (Poesen et al., 2002). Gullies are often connected to the river system and provide an effective link between upland areas and channels, allowing the rapid transmission of water and sediment into river systems. In dryland environments gully erosion is an important sediment source, contributing an average of 50-80 per cent of the overall sediment production (Poesen et al., 2002). Smaller features, intermediate in size between rills and gullies, also exist. These are called **ephemeral gullies** and are defined by the Soil Science Society of America (1996) as **small channels that are eroded by concentrated overland flow and that can easily be filled by normal tillage, only to reform in the same location by additional runoff events**.

**Figure 2.24** : Gullying

Gullies are relatively permanent ephemeral channels usually found in arid and semi-arid environments, with highly seasonal precipitation and sparse vegetation cover. Gullies are **morphologically different to stream channels, being relatively deep and narrow, with steep sidewalls and a stepped channel slope**. They typically range in size from depths of 0.5 m up to 25 m or 30 m.

Erosion is focussed at the sharp break in slope at the upslope end of the gully [gully head], where overland flow erodes the lip of the gully head before falling into the plunge pool at its base. Deepening and undercutting of the headwall take place at the plunge pool - undermining the headwall and allowing the gully head to retreat further up-slope.

The **steep sidewalls of the gully head are highly susceptible to various types of mass movement**, especially when saturated during extreme events. **Subsurface processes are also very significant in gully head retreat**. Subsurface flow moving towards the gully head can **weaken the walls**, and the development of piping is common. The collapse of pipes further contributes to gully head retreat. Under certain circumstances, gullies can extend rapidly upslope and tributary gullies may also form.

Gully erosion is common where the particle size of substrate or underlying soil or alluvium is fairly uniform.

**Arroyos**

Arroyos are gully-like features that are cut into debris-choked valleys. Evidence from many arid and semi-arid areas suggests that these features form as a result of increasing soil erosion. Associated with gullies are **badlands**: high-relief areas that are intensively dissected by gullies and are useless for agriculture. Badlands form on unconsolidated sediments, or poorly consolidated rocks in sparsely
vegetated areas and may be initiated by gully erosion. They are associated with arid and semi-arid climates but can also form in more humid climates.

Erosion is focussed at the **gully head**, the sharp break in slope at the upslope end of the gully. In the dramatic example shown in figure 2.25 overland flow is occurring over a large area (the green area is totally submerged), and ‘waterfalls’ can be seen where the flow plunges over multiple gully heads. Overland flow erodes the lip of the gully head as the water flows over it before falling into the plunge pool at its base, where deepening and undercutting of the headwall take place. This undermines the headwall and allows the gully head to retreat further up-slope.

The steep sidewalls of the gully head are highly susceptible to various types of mass movement, especially when saturated during extreme events. Subsurface processes are also very significant in gully head retreat. Subsurface flow moving towards the gully head can weaken the walls, and the development of piping is common. T

**Figure 2.25**: Arroyo

**Modes of sediment transport in river channels**

**Downstream Changes in Particle Size**

The size of river sediment normally decreases in size downstream, from boulders in mountain streams to silt and sand in major rivers because the coarse bed load is gradually reduced in size by abrasion. (see figure 2.26).

**Figure 2.26**: River sediment particles decrease in size further downstream
Furthermore, the textures of particles are modified by abrasion during wind or water transport: particles close to the source in a fluvial (or aeolian) system are angular, while those far from the source become increasingly rounded as they move further away from their source, as indicated in figure 2.27.

![Diagram of distance of transport affecting particle textures]

**Figure 2.27**: The effect of abrasion in forming smoothed surfaces and textures of particles downstream.

Chemical weathering continues during intermittent times when particles are temporarily deposited before further transport; furthermore, the percentage of very stable minerals (i.e., quartz and clays) increases with increasing transport distance.

The sediment supplied to most river channels varies greatly in size, from microscopic clay particles to large boulders. Streams transport sediment and transported material is called load.

**Classification of sediment size**

The size of sediment particles or load is usually along the scheme below:

- **Boulders** > 256 mm
- **Cobbles** 80 mm - 256 mm
- **Gravel** 2 mm - 80 mm
- **Sand** 0.05 mm - 2 mm
- **Silt** 0.002 mm - 0.05 mm
- **Clay** <0.002 mm

Lane refers to the diameter or size of aggregate as caliber.

There are three basic classes of load, namely:

- **Bed load**: sediment rolling, bouncing, and creeping along the river bed; moves during high velocity events. Sand grains and other aggregate down to... in size also form part of bed load, but these smaller particles are easily re- **Cobbles – Traction** (rolling) suspended during flow, especially turbulent flow. **Sandy Portion – Saltation** (bouncing)
- **Suspended load**: sediment that is fine enough to remain in suspension in stream (size depends on velocity and turbulence)
- **Dissolved load**: the invisible load of dissolved ions (e.g., Ca, Mg, K, HCO₃) from chemical weathering processes.

**Bedload**

Many alluvial and bedrock channels are characterised by deposits of coarse material forming the channel bars, although finer-grained sand and silt bars can also be common. The form and behaviour of bedload-dominated channels is rather different from suspended load dominated channels. Coarse material – typically coarse sands, fine gravels and larger particles – is moved close to, or along the bed of the channel as bedload because it weighs more. Particles are in therefore in continuous or regular contact with the channel bed and move discontinuously by rolling, bouncing, creeping, sliding or in a
series of hopping motions called saltation and at a slower velocity than the streamflow. See figures 2.28 & 2.29.

![Figure 2.28: Bedload transport by means of saltation](image)

![Figure 2.29: Modes of bedload sediment transport.](image)

Compared with the suspended load, bedload movement is more localised, involving much shorter travel distances and the bed load generally constitutes between 5 and 20 percent of the total load of a stream. These mechanisms have important implications for the way in which sediment of different sizes is transferred through the system.

**Suspended load**

Fine sediment particles which are usually in suspension during normal flow conditions, are referred to as suspended load. Depending on the velocity and turbulence of flow, the finer sediment – clay particles, silts and sands – are carried in the flow as suspended load and can be transported over considerable distances. This material is carried aloft, suspended above the channel bed by turbulent eddies, and is transported downstream in the main body of the flow.

When upward-moving currents exceed the velocity at which particles will settle to the channel bed under gravity, these particles remain in suspension. The finest fraction of the suspended load is called the wash load and consists of clay particles with diameters of less than 0.0063 mm - fine enough to remain in suspension even when water movement is barely perceptible. Wash load sediment can be carried over many kilometres in a matter of hours and eventually settle and are deposited where velocity decreases, such as in a lake or in the oceans.

![Figure 2.30: Suspended load](image)
Dissolved load

Material is also transported in solution as the dissolved load. These solutes are derived from a number of sources, including rock and soil weathering, the atmosphere, biosphere and human activity. While the material discharged to the oceans comprises predominantly of fine sediment, this is not always the case.

All stream water contains dissolved ions, which is invisible in water.

- Cations [positively charged (+)]
- Anions [negatively charged (-)]

The bulk of the dissolved content of most rivers consists of seven ionic species:

- Bicarbonate (HCO$_3^-$)
- Calcium (Ca$^{++}$)
- Sulphate (SO$_4^{--}$)
- Chloride (Cl$^-$)
- Sodium (Na$^+$)
- Magnesium (Mg$^{++}$)
- Potassium (K$^+$)

As well as dissolved silica as Si(OH)$_4$.

Determining when erosion, transportation or deposition will take place: Hjulström diagram –

Deposition of sediment by a stream is caused by a decrease in velocity: as competence is reduced, so sediment begins to drop out. To determine when erosion, transportation or deposition will occur, Hjulström did one of the first published quantitative studies of geomorphological processes, looking at soil erosion and sediment transport in the drainage basin of the river Fyrisån, in Uppsala, Sweden in the 1930s. He showed the relationship between the size of soil or sediment particles in a stream and the velocity of the stream, expressing this relationship as two curves on a graph, which has subsequently become known as the Hjulström curve or graph or diagram. This graph is used by hydrologists, engineers and sedimentologists to determine whether a river (or channel) will erode, transport or deposit sediment.

![Hjulström diagram](image)

**Figure 2.31**: Two versions of the Hjulström diagram, indicating the relationship between sediment size and flow velocity in determining when deposition, transport or erosion will take place.

It shows the velocity a stream needs to flow in order to

- pick up (erode),
- carry (transport),
- or deposit (sediment) a clast or grain in flowing water, or to estimate the speed at which deposition or sedimentation will take place when flowing water carrying sediment slows down.
The upper curve shows the critical erosion velocity in cm/s as a function of particle size in mm, while the lower curve shows the deposition velocity as a function of particle size. For cohesive sediment, like clay and silt, erosion velocity increases with decreasing grain size, as the cohesive forces are relatively more important when the particles get smaller. At particle sizes where friction is the dominating force preventing erosion, the required velocity increases with particle size – and the two curves follow each other closely. For deposition, the critical velocity depends on the settling velocity, which decreases with decreasing grain size. Sand particles of about 0.1 mm up to 1 mm require the lowest stream velocity to erode.

The axes are logarithmic. Notice that the grain size increases from .001 mm to .01 mm to .1 mm - each 10 times the previous.

**Lane’s Balance**

![Lane’s Balance diagram](image)

Figure 2.3: The Lane balance between stream power (stream slope / discharge) and sediment supply (sediment load / sediment supply). Adapted from Brierley and Fryirs (2005).

\[
Q_s d_{50} \sim Q_w S
\]

Where:
- \(Q_s\) = sediment discharge (kg/s)
- \(Q_w\) = water discharge (cm/s)
- \(d_{50}\) = sediment size (m)
- \(S\) = slope (m/m)

**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.

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.

**Valley setting**

Channel processes are driven by flow and sediment supply, although the range of channel adjustments that are possible are often restricted by the valley setting. The influence of channel substrate and vegetation on bank erosion and channel migration have already been mentioned. The valley slope is also significant, affecting the steepness of the channel, which, together with discharge, determines stream power. Channels that flow over very gentle gradients can sometimes be extremely restricted in the adjustments they can make because so little energy is available. Another control on channel adjustment is the degree of **valley confinement**. While some channels are able to migrate freely across a wide floodplain, others are confined to a greater or lesser extent by the valley walls.

**THE FORM OF A CHANNEL**

When the stream beds turn gentler due to continued erosion, downward cutting becomes less dominant, while lateral erosion of banks increases; a consequence of this is that the stream channel widens, and that hills and valleys surrounding the stream channel are reduced to plains over time. The shape and form of the channel itself also change down the course of a river, and are dependent on the gradient, flow rate and sediment load of the stream at different points. **Four main types of alluvial channel form can be identified:** straight, meandering, braided and anabranching. Bedrock channels also exhibit a wide variety of forms. However, with so many environmental variables influencing channel form, a range of different channel forms and behaviour is possible and not all rivers fit neatly into one of these categories – there are many examples of transitional rivers that have characteristics associated with more than one channel type.

**Channel substrate**: the material in which the channel is formed. An important distinction can be made between bedrock and alluvial substrates (Figure 2.33).

- **Bedrock channels**, as their name suggests, are sections of channel that are cut directly into the underlying bedrock, while
- **alluvial channels** are formed in **alluvium** – sediment that has previously been laid down in the valley floor by rivers. Alluvium can include a mixture of unconsolidated particles ranging in size from boulders, gravels and sands to finer deposits of silts and clays. Where the valley floor is wide enough, material laid down in the channel, together with silt deposited by floods, form a **floodplain** adjacent to the river channel.
ALLUVIAL CHANNEL FORM

Characteristics of alluvial channels

Channels usually lined with alluvium
Removed & carried further downstream during flood
Re-deposited during wane
Erodible channel boundaries (alluvial banks & bed)
Transport Capacity ≤ Sediment Supply
Storage can be quite high
Input ≥ Output

Figure 2.33: Alluvial channels are formed in alluvial deposits formerly laid down by fluvial processes.

Alluvial channel form

Four main types of alluvial channels are generally recognised: straight, meandering, braided and anabranching.

The previous sections have considered the controlling influence of the driving variables and boundary conditions on channel form and behaviour. Each of these controls varies across a continuous range. For instance, slopes range from steep to gentle, valleys from confined to unconfined, and sediment loads from suspended to mixed to bedload dominated. Many different combinations are possible, leading to the immense variety of fluvial forms and behaviour that is seen globally.

The continuum of alluvial channel types is illustrated in Figure 2.41 on p 30. In general terms, different channel types exist along an energy gradient, ranging from high energy braided channels through meandering and straight to low-energy anastomosing channels (a sub-set of anabranching channels). Floodouts and chains of ponds are found in low-gradient arid environments, where downstream reductions in discharge result in a dwindling supply of energy. This continuum can be related to the channel controls, since stream power integrates channel slope and flow regime. It also influences the type of load that the channel can carry, which in turn determines the substrate and stability of the channel. Like all channel classifications, the one shown in Figure 8.10 is, by necessity, a simplification of reality. While some channel reaches typify, say, a braided form, many have characteristics that are associated with more than one type. In fact, it has been suggested that channels with an intermediate form might be the norm rather than the exception (Ferguson, 1987). Although there is a continuum of forms, thresholds do exist between them. For example, there is a meandering–braiding threshold, above which rivers braid and below which they meander. Rivers that are close to this threshold, such as many of those in the South Island of New Zealand, have alternating meandering and braided reaches.

Straight channels

Most single-channel rivers and streams follow a winding path and straight channels are rare. Even where they do exist, variations are usually seen in flow patterns and bed elevation. Straight channels are relatively static, with rates of channel migration limited by a combination of low energy availability and high bank strength. This is especially true where the channel banks are formed from more resistant material, such as cohesive silts and clays.

Meandering channels

In large flood and delta plains, rivers rarely flow in straight courses. Loop-like channel patterns called meanders develop over flood and delta plains. In contrast to braided rivers, meandering rivers typically contain one channel that winds its way across the floodplain. As it flows, it deposits sediment on banks that lie on the insides of curves (point bar deposits), and erode the banks on the outside of curves.
A meander is not a landform but is only a type of channel pattern. Their formation is due to:

(i) propensity of water flowing over very gentle gradients to work laterally on the banks;

(ii) unconsolidated nature of alluvial deposits making up the banks with many irregularities which can be used by water exerting pressure laterally;

(iii) coriolis force acting on the fluid water deflecting it like it deflects the wind.

When the gradient of the channel becomes extremely low, water flows leisurely and starts working laterally. Slight irregularities along the banks slowly get transformed into a small curvature in the banks; the curvature deepens due to deposition on the inside of the curve and erosion along the bank on the outside. If there is no deposition and no erosion or undercutting, the tendency to meander is reduced. Normally, in meanders of large rivers, there is active deposition along the convex bank and undercutting along the concave bank.

Meanders occur most commonly in channels that lie in fine-grained stream sediments and have gentle gradients, but also form in a variety of bedrock substrates. Associated with moderate stream powers, alluvial meanders may develop in gravels, sands, or fine-grained silts and clays. Meanders are scaled to the size of the channel, with wider spaces (longer wavelength) for larger channels. The degree of meandering varies greatly, from channels that only deviate slightly from a straight line to sequences of highly convoluted meander bends. Variations are also seen in the regularity of meander bends. Meandering channels evolve over time as individual bends migrate across the floodplain. Erosion is usually focused at the outside of meander bends, which gradually eat into the floodplain as the channel migrates laterally. At the same time, deposition on the inside of the bend allows the channel to maintain its width. Cut-offs – short sections of abandoned channel – indicate the path of former meander bends.

Circumstances favourable for meandering to occur include a gentle slope (< 2%), a relatively low sediment load and cohesive banks. Meandering rivers change their course gradually. In the progress of their migration, they create floodplains much wider than the channel itself, as it erodes the cut bank, which is the outside loop of the meander, while deposition takes place on the point bar, which is on the inside loop of the meander. Meandering alluvial systems are also characterized by seasonal flooding and the deposition and build-up of very fertile soil
Streams generally **erode on outer (cut) bank** where velocity is greatest, and **deposit on the inner sides of bends** where velocity is less. Meanders tend to grow as the flow erodes the banks, favouring development of meandering channels. Successive point bar deposits create a floodplain, while cut banks migrate closer to each other, causing cut-offs and the formation of oxbow lakes. These processes all add to the migration of the stream and the building up of alluvial deposits that form a floodplain.

**Meander geometry**

Various methods are used to quantify the geometric characteristics of meandering channels. These are based on measurements that can be made in the field, from maps, aerial photographs and, increasingly, satellite images. The spacing of meander bends, or **meander wavelength** ($l$), can be determined by measuring the channel width $\approx 10$ to $14 \times$ wavelength as in figure below.

The **sinuosity ratio** gives an indication of how ‘bendy’ a channel is and can be worked out by measuring the length of a channel reach and dividing this by the straight line distance along the valley. Channels with a sinuosity ratio of less than 1.1 are described as **straight**, those between 1.1 and 1.5 are **sinuous**, and **meandering** channels have a ratio of more than 1.5.
Although widely used, these descriptions are somewhat arbitrary, since they are not based on any physical differences. There is a tendency for the thalweg, or line of fastest flow, to shift from side to side along the channel (See figure 2.37). This is seen even in straight channels, and is often associated with the development of riffles, pools and alternate bars.

Since the distance between successive meander bends varies, a mean wavelength is calculated for several meander bends along the reach of interest. Meander wavelength is more strongly related to channel width than to bankfull discharge. This is because secondary circulation within the channel, which is significant in meander development, is controlled by channel size. Interestingly, a similar relationship is seen for small supraglacial streams that flow over the surface of glaciers: they often develop meanders, despite the absence of sediment. There is a well established relationship between channel width and meander wavelength, which is usually ≈ ten to fourteen times the bankfull width (See figure 2.38).

Figure 2.38 : The relationship between channel width and meander wavelength.

Meander wavelength can also be influenced by the channel substrate, and longer wavelengths are associated with gravel channels than for silt and clay channels of a similar size. The reason for this is that cohesive banks allow the development of a narrower cross-section with tighter bends (Schumm, 1968). An indication of the ‘tightness’ of individual bends can be determined by fitting a circle to the centre line of a meander bend: the radius of this circle is called the radius of curvature (rc). To allow comparison between channels of different sizes, the tightness of bends is usually expressed as the ratio between the radius of curvature and the channel width at the bend (rc/w). Observations have shown that many bends develop an rc/w ratio of 2 to 3. For bends that are tighter than this, flow separation leads to increased energy losses (Bagnold, 1960).

In cross-section, the form of the channel varies along its length. meander bends are associated with asymmetric cross sections since scouring and bank steepening occur at the outside of the bend, while deposition occurs on the inside of the bend. The cross-section is more symmetrical at riffle sections.

**Braided channels**

Braided channels have substantial inputs of bedflow from upstream catchments. When sediment supply exceeds the transport energy in a stream, it is unable to move all the available load and tends to deposit the coarsest sediment, causing one or more central bars to form, which divert the flow towards the banks. This flow in turn increases lateral erosion on the banks, causing more deposition and bar formation. Braided channels tend to form in streams with a highly variable discharge, easily erodible banks, and/or a high sediment load and they are characterised by numerous channels that split off and rejoin each other to give a braided appearance.

Figure 2.39 : Braided channel segments of Son (left) & Gandak (right) rivers.
They usually carry coarse-grained sediment down a steep gradient and have a high total load, with bedload typically being more than 11% of the total load. Braided rivers are usually wide and relatively shallow (they have a high channel width to depth ratio). As the valley widens, the water column is reduced and more and more sediment gets deposited as islands and lateral bars, resulting in the development of separate channels of flow.

Deposition and lateral erosion of banks are essential for the formation of a braided stream pattern. The appearance of a braided channel varies with changing flow conditions, with many bars becoming partly or wholly submerged during high flows, giving the appearance of a single wide channel. At low flows, extensive areas of bar surface may be exposed. Braided rivers, are associated with high rates of energy expenditure, which is involved in the transport of large volumes of sediment. They often have steep channel slopes, although there are several examples of large braided rivers that flow over low gradients. Erodible banks are also required for the channel to become wide enough to allow for the growth and development of channel bars. Braided channels are highly dynamic, with frequent shifts in channel position. Modifications, such as the dissection and reworking of bars and the formation and growth of new bars, occur over relatively short periods of time (days to years). The presence of bars leads to complex patterns of flow within the channel, and there can be sudden shifts in the location of sub-channels. Individual channels can be abandoned or reoccupied in the space of a few days.

Anabranching channels

With anabranching channels the flow separates into two or more distinct channels, and this channel form is rare in comparison to braided and meandering channels. The separate channels, called anabranches, are typically cut into the floodplain, dividing it up into a number of large islands with an elevation similar to the surrounding floodplain. Usually the island positions remain relatively fixed over many years or decades and are well vegetated, where climatic conditions allow this. Individual anabranches can themselves be straight, meandering or braided, with the flow in each ranch reasonably independent of the surrounding branches. Unlike braided channels, rates of lateral channel migration are usually very low, but new channels can be cut when floodwaters breach the channel boundary and spill out on to the floodplain. Other channels are abandoned as the flow is diverted elsewhere, or when they become infilled with sediment.

Anabranching channels occur in flood-dominated regimes, and usually have banks that are relatively resistant in comparison with the available stream power. It is the most diverse channel type which is found in many different environments – cold, dry, wet, semi-arid, tropical and sub-arctic, and it occurs in substrates ranging from coarse to fine-grained alluvium and in various energy regimes. They are nevertheless comparatively scarce.

Colour Plates 8 and 9 both show examples of a subset of low-energy anabranching channel that is referred to as anastomosing. Although most research on anabranching channels has so far been focused on anastomosing channels, anabranching channels represent the most diverse of the four main channel types.

When considering channel form, individual reaches of channel are usually considered. This is because of the downstream changes in channel size and shape that are brought about by factors such as increasing drainage area and variations in channel substrate. Therefore different channel patterns may be found along the same channel. At the reach scale – (a few metres to a few hundreds of metres) – there is a homogeneity of form. In addition to the variation in the channel planform patterns described above, variations are also seen in cross-sectional shape and channel slope. For example, braided channels tend to be relatively wide and shallow in comparison to meandering channels, which have a narrower, deeper cross-section. Headwater streams in mountainous areas typically flow in steep channels, with frequent waterfalls, pools and rapids. This is in contrast to rivers that flow across lowland floodplains, where the channel slope is much more gentle. Fluvial forms also exist at the sub-channel scale. These include channel bars, pools excavated by localised scour, and periodic features such as dunes and ripples that form on the bed of sandy channels. Certain groupings, or assemblages, of these features are associated with different channel types.
Why do rivers anabranch?

Anabranching channels are often associated with very low slopes and, because little energy is available, the range of possible adjustments is somewhat limited. Although the slope cannot be increased, form adjustments can lead to a reduction in flow resistance, thus increasing the energy available for transporting sediment. It has been demonstrated that two or more channels with a low width-to-depth ratio (narrow and deep) are more hydraulically efficient than a single channel (Nanson and Huang, 1999), because the combined hydraulic radius of the multiple channels is greater (more hydraulically efficient) than for a single channel carrying the same flow.

Anabranching channels are usually formed by erosion, when avulsion leads to the incision of a new channel into the floodplain. Avulsion occurs during high flows, when one of the banks is breached and water spills out onto the floodplain. If flow is sufficiently concentrated, a new channel can be incised, eventually rejoining another channel further downstream. Some anabranches are only active during flood flows, acting as a distributory system for dispersing and storing water and sediment (Nanson and Huang, 1999). Individual anabranches are abandoned when they become infilled with sediment, perhaps as a result of a blockage or because the flow is diverted elsewhere. In some cases, anabranching may develop as a result of sediment deposition, when flow is concentrated by the development of bars or ridges within a relatively inefficient channel (Wende and Nanson, 1998).

Anastomosing channels

There is some confusion surrounding the terminology of these multi-channelled forms, which are sometimes referred to as anastomosing. The nomenclature used by Nanson and Knighton (1996) will be used here, where the term anabranching is used to describe all planforms that are characterised by more than one separate channel. Anastomosing will be used to describe one particular subgroup of low-energy anabranching channels.

General controls on channel morphology

![Figure 2.40: Alluvial channel type as function of bedload supply and diameter, and stream energy](image)

Straight channels typically steep, low-sediment supply
Braided channels typically high sediment supply
Meandering channels typically low to medium slope and moderate sediment supply
Anastomosed channels typically reflect vegetation influences

**Straight Channels**

Straight channels are rare. They form where streams are confined by topography or follow geologic structures. Generally mountain streams.

**Braiding favoured by:**

- High slope
- High sediment load
- High discharge variability
- Erodible (non-cohesive) banks
To determine whether a given length stream will have a straight, meandering, braided, or anastomosed form, the following figure is helpful:

*Figure 2.41 The continuum of variants of channel planform. After Brierley and Fryirs (1992), adapted from Church (1992) and Schumm (1977).*
DEPOSITION

Sediment particles are deposited when there is a reduction in the competence and capacity of the flow. The process takes place at a very small scale and involves individual grains, although depositional forms can be observed over a wide range of spatial scales, from the smallest bedforms to vast floodplains and deltas. Thresholds for deposition are associated with the fall (or settling) velocity; the deposition of suspended sediment occurs when the fall velocity dominates over turbulent diffusion. Fall velocity is also affected by the viscosity and density of water. Because there is also a close relationship between fall velocity and particle size, the coarsest sediment is usually deposited first, leading to sediment sorting, which is both a vertical and horizontal gradation of sediment, from coarse to fine. These are both influenced by changes in suspended sediment concentration. In addition, finer material can be transported as agglomerations of sediment called flocs, which have a greater fall velocity than the individual particles forming them. In the case of bedload transport, the near-bed flow conditions are significant. Bedload deposition occurs where the bed shear stress drops below the critical shear stress (Shields’s parameter) required to transport particles of a given size.

The different circumstances that lead to sediment deposition include:

- **Reductions in flow discharge** which are seen as flows recede, or along dryland rivers, where downstream losses are caused by high rates of evaporation and percolation.
- **Decreases in slope** which can be localised, or involve a gradual reduction over a longer length of channel and cause a reduction in average flow velocity and stream power.
- **Increases in cross-sectional area** cause the flow to diverge and become less concentrated. Flow resistance increases because there is more contact between the flow and channel boundary. There is a large increase in cross-sectional area when overbank flows occur.
- **Increases in boundary resistance** are associated with vegetation and coarse bed sediment. When overbank flows occur, velocity is reduced by the increased roughness of the floodplain surface, leading to the deposition of suspended sediment.
- **Flow separation**, which causes sediment to become decoupled from the flow.
- **Obstructions to flow**. Sediment often accumulates behind obstructions. These include boulders, outcrops or islands of bedrock, woody debris and man-made structures such as bridge piers, dams and flow control structures. Changes in the supply of sediment are also important. For example, sediment tends to accumulate immediately downstream from scour zones caused by flow convergence, when the material scoured from the channel bed is deposited immediately downstream. At a larger scale, increases in the supply of sediment to a channel reach are caused by changes within the upstream drainage area (Chapter 5).

Depositional environments

Deposition dominates in the deposition zone, where there is a decline in gradient and low energy levels, but limited deposition also occurs in the production and transfer zones of the fluvial system. Large-scale deposition leads to the development of characteristic landforms, including floodplains, alluvial fans and deltas. Within channels, bars represent smaller-scale depositional features, which are commonly found on the inside of meander bends, along the edges of channels, and where tributaries join the main channel. Braided channels are characterised by numerous mid-channel bars.

The floodplain

The **floodplain** is the lateral extent of the river at high flow. A **floodplain** is the flat land immediately surrounding a stream channel and submerged at times of high flow and is formed from a mixture of in-channel and overbank deposits. Their development and evolution is governed by a number of factors, including the supply of sediment (volume and calibre or diameter), the energy environment of the
channel, and the valley setting. Sediment is laid down by rivers as they migrate across the floodplain, being deposited on the inside of meander bends or when braid bars are abandoned. These channel deposits are relatively coarse in comparison with the much finer sediment that is laid down by overbank flows, further away from the channel. Processes of erosion can also be significant in reworking sediment or in removing part, or all, of the floodplain surface.

Figure 2.42: Floodplain construction by bedload deposition

Point bar deposit grows laterally through time, and it forms a floodplain in this process: deposition develops a floodplain just as erosion makes valleys. Flood plains are built up as flat areas adjacent to main channel is subjected to periodic flooding and suspended sediment is deposited, but they can also be formed by deposition of bedload as the channel migrates across its valley.

Floodplains are major landforms caused by river deposition. Large sized materials are deposited first when a stream channel breaks into a gentle slope. Thus, normally, fine sized materials like sand, silt and clay are carried by relatively slow moving waters in gentler channels usually found in the plains and deposited over the bed and when the waters spill over the banks during flooding above the bed. A river bed made of river deposits is the active floodplain. The floodplain above the bank is inactive floodplain. Inactive floodplain above the banks basically contain two types of deposits — flood deposits and channel deposits. In plains, channels shift laterally and change their courses occasionally leaving cut-off courses which get filled up gradually. Such areas over flood plains built up by abandoned or cut-off channels contain coarse deposits. The flood deposits of spilled waters carry relatively finer materials like silt and clay. The flood plains in a delta are called delta plains.

Figure 2.43: The effect of lateral channel migration with increasing meander sinuosity in forming or carving out a floodplain

Alluvial fans

Alluvial fans are typically found in situations where an upland drainage basin flows out onto a wide plain. They are fan-shaped deposits of water-transported material (alluvium) and they typically form at the base of any topographic feature where there is a marked break in slope. Consequently, alluvial fans tend to be coarse-grained, especially at their mouths, while they can be relatively fine-grained at their edges.
The sudden change from confined to unconfined conditions leads to flow divergence, while mean flow velocity is decreased by the reduction in slope. The resultant deposition leads to the formation of a conical feature with a convex cross-profile. Most fans have a radius of less than 8 km, but can be more than 100 km wide in some cases. Fans are commonly found in dry mountain regions, where an abundant sediment supply is associated with extreme discharges and frequent mass movements. Alluvial fans in humid areas show normally low cones with gentle slope from head to toe and they appear as high cones with steep slope in arid and semi-arid climates.

Figure 2.44 : (a) An alluvial fan deposited by a hill stream on the way to Amarnath, Jammu and Kashmir. (b) Alluvial fan in Death Valley, California

**Bajada**

Where a number of individual fans develop along a mountain front, they may grow laterally and coalesce to form a sloping apron of sediment called a **bajada**.

Figure 2.45 : Bajada formed from coalescing alluvial fans in Death Valley, California

**Deltas**

**Deltas** are found where sediment-charged water flows into a body of still water, dumping and spreading its load into the water body. They extend outwards from shorelines where rivers enter lakes, inland seas and oceans. In coastal areas deltas form where the supply of sediment is greater than the rate of marine erosion, although sediment is redistributed by coastal processes. The influence of fluvial processes tends to dominate in the case of lake deltas.

Unlike in alluvial fans, the deposits making up deltas are very well sorted with clear stratification. The coarsest materials settle out first and the finer fractions like silts and clays are carried out into the sea. As the delta grows, the river distributaries continue to increase in length (figure 2.46) and the delta continues to build up into the water body.
**FLOODPLAIN GEOMORPHIC UNITS**

**Levees**

Levees are found along the banks of large rivers and form when debris-laden floodwater overflows the channel and slows down as it moves onto the floodplain. They are elongated, raised linear ridges of relatively coarse deposits that run parallel to the stream channel and which form at the channel–floodplain boundary during overbank flow events (figure 2.47). During flooding as the water spills over the bank, the velocity of the water comes down and large sized and high specific gravity materials get dumped in the immediate vicinity of the bank as ridges. They are high nearer the banks and slope gently away from the river. The levee deposits are coarser than the deposits spread by flood waters away from the river. When rivers shift laterally, a series of natural levees can form. Moving across the boundary from channel to floodplain, there is a sudden loss of momentum because of the interaction between fast channel flow and slow floodplain flow, resulting in the preferential deposition of material along the edges of the channel. The photograph left shows a meandering river in flood.

The boundary between channel and floodplain may be the site of a natural levee (a broad, low ridge of alluvium built along the side of a channel by debris-laden floodwater). Levees are clearly visible as the raised strips of land running along the channel margins. The height of levees is scaled to the size of the channel and their presence implies a relatively stable channel location (Brierley and Fryirs, 2005). These natural levees should not be confused with the artificial levees that are constructed along river banks for purposes of flood control.
Point bars

Point bars are also known as meander bars. They are found on the convex side of meanders of large rivers and are sediments deposited in a linear fashion by flowing waters along the bank. They are almost uniform in profile and in width and contain mixed sizes of sediments. If there more than one ridge, narrow and elongated depressions are found in between the point bars.

Rivers build a series of them depending upon the water flow and supply of sediment. As the rivers build the point bars on the convex side, the bank on the concave side will erode actively.

![Figure 2.48: Natural levee and point bars](image)

Specific landforms associated with the floodplain of a meandering stream

Oxbow lakes

As meanders grow into deep loops, the same may get cut-off due to erosion at the inflection points and are left as ox-bow lakes. Oxbow lakes are therefore crescent-shaped lakes formed in an abandoned river bend which has become separated from the main stream by a change in the course of the river. Old channels abandoned as a river meanders across its floodplain form oxbows.

![Figure 2.49: The formation of oxbow lakes through the cut-off of mature meanders](image)

Splays or Crevasse splays

Another landform associated with meandering streams is a splay, which is a fan-shaped lobe of sediment deposited when sediment-charged water breaks the bank of the levee and flows beyond the levee. A splay is therefore a deposit of coarse material resulting from a levee breach during a flood. If flow is sufficiently concentrated, a new channel may be cut and deepened by scour.
Backswamps

A low area of swampy ground beyond a river’s natural levees. The build-up of sediment in the channel may mean that the channel is at a higher elevation than the surrounding floodplain. When levees are overtopped, water can enter the lower-lying area on the other side of the levee. This may be a depression or a swamp area characterised by wetland vegetation (Figure 8.9b). Colour Plate 9 shows some good examples of backswamps. These are not exclusively associated with anabranching rivers and can also form at the valley margins of other channel types.

Flood channels

Flood channels are relatively straight channels that bypass the main channel. They have a lesser depth than the main channel and are dry for much of the time, only becoming filled with water as the flow approaches bankfull.

Floodouts

Floodouts are associated with dryland channels. They occur where floodwaters leave the main channel and branch out onto the floodplain in a number of distributory channels. This happens where low gradients, downstream transmission losses and high rates of evaporation lead to a downstream reduction in channel capacity. Channels may re-form downstream from the floodout if flow concentration is sufficient, forming a discontinuous channel. Alternatively the floodout may mark the channel terminus. Floodouts can also form where the channel is blocked by bedrock outcrops, fluvial, or aeolian deposits such as sand dunes (Tooth, 1999).

Meander scroll bars

A meander scroll consists of long, curving, parallel ridges (scrolls) that during stages of high water have been aggraded against the inner bank of the meandering channel, while the opposite bank experienced erosion. In some cases, former point bar deposits can be seen in the surface topography of the floodplain as scroll bars, with each scroll representing a former location of the point bar (Figure 8.9a). The undulating ridge and swale topography that results consists of higher ridges separated by topographic lows called swales. Meander scroll bars can be seen as a series of vegetated ridges on the point-bar deposits in the foreground of Colour Plate 10. Migrating meanders do not always form scroll bars and the surface topography of these deposits may be relatively featureless.

![Diagram of meander scroll bars and other meander floodplain landforms](image)

Figure : 2.50 : Meander scroll bars and other meander floodplain landforms

Cut-offs

These are abandoned meander bends that have been short-circuited by the flow. Cut-offs become infilled over time by a process of abandoned channel accretion.
Palaeochannels

Palaeochannels are longer sections of abandoned channel (Colour Plate 18). Like active channels, palaeochannels exhibit a wide range of different planforms. As time goes by, they gradually become infilled by abandoned channel accretion, the degree of infilling reflecting the age of the channel. The rate at which infilling occurs is dependent on factors such as the geometry of the palaeochannel and its position on the floodplain in relation to overbank events.

Bedrock channels

Bedrock channels also show a wide diversity of form. In comparison with alluvial channels, bedrock and mixed bedrock–alluvial rivers have received relatively little attention until recently. These channels often behave in a different way to alluvial channels, being strongly influenced by the resistant nature of their substrate. Structural controls, such as joints, bedding planes and the underlying geological strata can all have a significant effect on flow processes and river morphology. As with alluvial channels, the flow may follow single or multiple channels. Straight reaches are often associated with structural controls, for example where the channel follows the line of a fault or joint. However, flow characteristics also have an influence in shaping the channel. Colour Plate 13 shows the regularly undulating walls of a slot canyon, which have been shaped by flash floods. Meanders can also form in bedrock-influenced channels, as can be seen from the spectacular incised meanders of the Colorado (Colour Plate 5). Because of the resistant substrate, bedrock meanders tend to be scaled to higher flows than their alluvial counterparts.

![Image of bedrock channels]

Figure 2.51 Bedrock channels

Channels floored by bedrock and lacking an alluvial bed cover.

An example of a multi-channel bedrock river is shown in Colour Plate 2. The individual channels have cut their course to flow around bedrock bars. In some mixed bedrock–alluvial channels, bedrock bars may form a core that becomes covered in alluvial deposits, giving the appearance of an alluvial channel.

Indicates transport capacity >> sediment supply.

Bedrock channels

- Fixed channel boundaries (bedrock banks and bed)
- High transport capacity
- Low Storage
- Input ≈ Output
Bedrock vs. Alluvial Channels

Bedrock: sediment supply < transport capacity
Alluvial: sediment supply ≥ transport capacity

**Long time (>10^3 yr):**
mountain channels = bedrock
floodplain channels = alluvial.

**Shorter time (<10^3 yr):** material on bed surface defines bedrock vs. alluvial.

Steps and pools

Steps and pools often characterise steep, upland channels in a wide range of humid and arid environments. The steps are formed from coarser material and form vertical drops over which the flow plunges into the deeper, comparatively still water of the pool immediately downstream. Steps are relatively permanent features and consist of a framework of larger particles that is tightly packed with finer material. In forested catchments, woody debris forms part of the structure of steps, while steps and pools can also form in bedrock channels. Like riffles and pools, step–pool sequences are most apparent during low-flow conditions as they tend to be drowned out at higher flows. It is also during low-flow conditions that step–pool systems offer the most flow.

PROCESSES OF EROSION IN BEDROCK CHANNELS

The morphology of bedrock channels is mainly influenced by processes of erosion because the supply of sediment is often limited. Three types of erosion are significant: block quarrying, abrasion and corrosion.

**Block quarrying** is the dominant process (Hancock *et al.*, 1998) and involves the removal of blocks of rock from the bed of the channel by drag and lift forces. The size of the quarried blocks can be considerable. Tinkler (1993) reports blocks of sandstone $1.2 \times 1.45 \times 0.11$ m and $1.0 \times 0.5 \times 0.05$ m being removed from the bed of Twenty Mile Creek, Niagara Peninsula, Ontario, during normal winter flows, when the flow depth was less than 0.4 m.

Before blocks can be entrained by the flow, a certain amount of ‘preparation’ is required to loosen them. Subaerial weathering and other weakening processes play an important role in this. Weakening processes described by Hancock *et al.* (1998) include the bashing of exposed slabs by particles carried in the load and a previously undocumented process termed ‘wedging’, which leads to the enlargement of cracks in the bedrock substrate. This is thought to occur when small bedload particles are able to enter cracks that are momentarily widened by fluid forces. The particles then become very firmly lodged and prevent the crack from narrowing again. As time progresses, further widening of the crack can be sustained as larger particles fall into it, and may ultimately lead to block detachment. Under conditions of very high flow velocity, sudden changes in pressure can generate shock waves that weaken the bed by the process of **cavitation**. This effect is caused by the sudden collapse of vapour pockets within the flow (Knighton, 1998).

**Abrasion** is the process by which the channel boundary is scratched, ground and polished by particles carried in the flow. Erosion is often concentrated where there are weaknesses and irregularities in the rock bed, which allow abrasion to take place at an accelerated rate. This can lead to the development of **potholes**, deep circular scour features that often form in bedrock reaches. Once a pothole starts to develop, the flow is affected, focusing further erosion. Any coarse material that collects in the pothole is swirled around by the flow, deepening and enlarging it, and literally drills down into the channel bed. Over time potholes may coalesce, leading to a lowering of the bed elevation. Plate 7.1 shows how potholes have contributed to bed lowering near the site of a waterfall.
Scouring by finer material carried by the flow, such as sand, leads to the development of sculpted forms. These include flutes and ripple-like features, which reflect structures within the flow (Plate 7.2). These are commonly observed on the crests of large boulders and other protrusions into the flow, where flow separation takes place and fine sediment is decoupled from the flow (Hancock et al., 1998). The rock boundary may also be polished by fine material carried in suspension.

Bedrock channels formed in soluble rock are also susceptible to erosion by corrosion, especially where the presence of joints and bedding planes allows solutional enlargement. Solutional features such as scallops may also be seen. These spoon-shaped hollows often cover the walls of cave streamways. Their length is related to the formative flow velocity, ranging from a few millimetres (relatively fast flow) to several metres (relatively slow). Although the actual processes of erosion operate at a small scale, their effects can be seen over scales ranging from millimetres to kilometres. There are several controls on rates of erosion, which influence the processes described above. These include micro-scale (millimetres to centimetres) variations in the rock structure, the larger scale effects of bedding, joints and fractures, and basin-scale influences such as regional geology and base level history (Wohl, 1998).

EROSIONAL LANDFORMS

Valleys

Valleys start as small and narrow rills; the rills will gradually develop into long and wide gullies; the gullies will further deepen, widen and lengthen to give rise to valleys. Depending upon dimensions and shape, many types of valleys like V-shaped valley, gorge, canyon, etc. can be recognised.

A gorge is a deep valley with very steep to straight sides (figure 2.52) and a canyon is characterised by steep step-like side slopes (Figure 2.53) and may be as deep as a gorge. A gorge is almost equal in width at its top and its base.

In contrast, a canyon is wider at its top compared to its base (figure 2.53). In fact, a canyon is a variant of gorge. Valley types depend upon the type and structure of rocks in which they form. For example, canyons commonly form in horizontal bedded sedimentary rocks and gorges form in hard rocks.

Potholes and Plunge Pools

Over the rocky beds of hill-streams more or less circular depressions called potholes form because of stream erosion aided by the abrasion of rock fragments. Once a small and shallow depression forms, pebbles and boulders get collected in those depressions and get rotated by flowing water and
consequently the depressions grow in dimensions. A series of such depressions eventually join and the stream valley gets deepened. At the foot of waterfalls also, large potholes, quite deep and wide, form because of the sheer impact of water and rotation of boulders. Such large and deep holes at the base of waterfalls are called plunge pools. These pools also help in the deepening of valleys. Waterfalls are also transitory like any other landform and will recede gradually and bring the floor of the valley above waterfalls to the level below.

**Incised or Entrenched Meanders**

In streams that flow rapidly over steep gradients, normally erosion is concentrated on the bottom of the stream channel. Also, in the case of steep gradient streams, lateral erosion on the sides of the valleys is not much when compared to the streams flowing on low and gentle slopes. Because of active lateral erosion, streams flowing over gentle slopes, develop sinuous or meandering courses. It is common to find meandering courses over floodplains and delta plains where stream gradients are very gentle. But very deep and wide meanders can also be found cut in hard rocks. Such meanders are called incised or entrenched meanders. Meander loops develop over original gentle surfaces in the initial stages of development of streams and the same loops get entrenched into the rocks normally due to erosion or slow, continued uplift of the land over which they start. They widen and deepen over time and can be found as deep gorges and canyons in hard rock areas. They give an indication on the status of original land surfaces over which streams have developed.

**River Terraces**

River terraces are surfaces marking old valley floor or floodplain levels. They may be bedrock surfaces without any alluvial cover or alluvial terraces consisting of stream deposits. River terraces are basically products of erosion as they result due to vertical erosion by the stream into its own depositional floodplain. There can be a number of such terraces at different heights indicating former river bed levels. The river terraces may occur at the same elevation on either side of the rivers in which case they are called paired terraces (figure 2.54).

![Figure 2.54: Paired and unpaired river terraces](image)
When a terrace is present only on one side of the stream and with none on the other side or one at quite a different elevation on the other side, the terraces are called unpaired terraces. Unpaired terraces are typical in areas of slow uplift of land or where the water column changes are not uniform along both the banks. The terraces may result due to (i) receding water after a peak flow; (ii) change in hydrological regime due to climatic changes; (iii) tectonic uplift of land; (iv) sea level changes in case of rivers closer to the sea.

Many stream valleys contain one or more relatively flat alluvial terraces that lie above the floodplain. A terrace is a remnant of an abandoned floodplain. Streams may create depositional landforms (especially floodplains) and then start to down-cut in response to uplift.

**Anabranching bedrock channels**

Multiple bedrock channels can also be found. Preferential erosion along lines of weakness, such as joints and fractures, can lead to the development of anabranching reaches. Tooth and McCarthy (2004) report examples from arid and humid regions including South and North America, India and South Africa. (A small-scale South African example is shown in Colour Plate 2.) This type of anabranching is also referred to in the literature as anastomosing or erosional braiding. The aerial photograph in Plate 8.14 shows an anabranching bedrock reach of the semi-arid Sabie River, Mpumalanga Province, South Africa. Extensive bedrock outcrops are found along the Sabie, which has incised a wide macro-channel in which all but the most extreme floods are contained. Significant downstream variations are seen in channel morphology in response to changes in the distribution and thickness of sediment deposits. The channel changes several times from a single to a multiple-channel form and various different bedrock, alluvial and mixed-reach types have been identified (Heritage et al., 1999). Some of the anabranching reaches are characterised by extensive bedrock pavements, whereas along other reaches, deposition has created alluvial islands with bedrock core bars.

Vegetation plays an important part in stabilising these deposits (Plate 8.15). The aerial photograph was taken seven months after a major flood event, which took place in February 2000. This had an estimated return period of 200 years (Heritage et al., 2004). Given the extreme flood distribution associated with semi-arid regions, this was a huge event which resulted in major modifications to parts of the channel. Patterns of change were complex, with some deposits being largely unaffected by the flood. However, many of the mixed anastomosing reaches experienced widespread sediment stripping. The bedrock core bars and individual anabranches can be seen clearly in Plate 8.14.

Preferential erosion along lines of weakness (joints and fractures) can lead to the development of anabranching reaches in bedrock channels.