Chapter 2 : FLUVIAL GEOMORPHOLOGY VERSION 2013.2

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:

(i) stability of sea level;
(ii) tectonic stability of landmasses;
(iii) 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.***

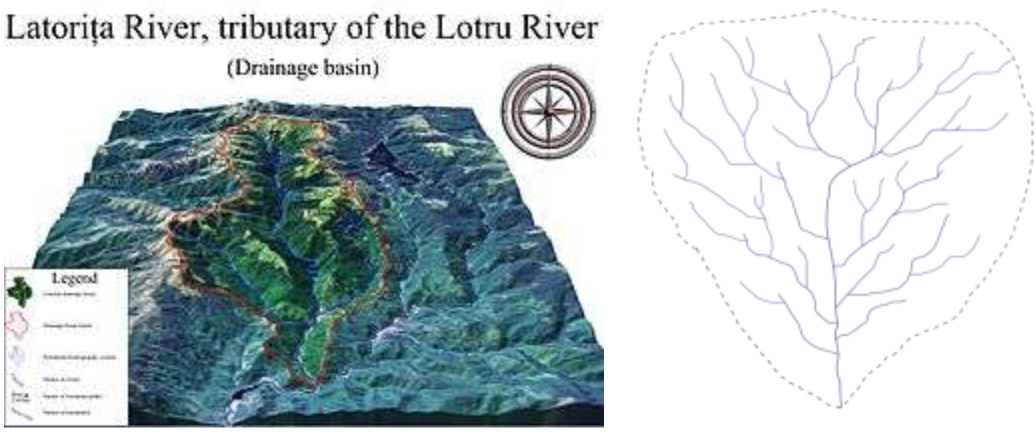
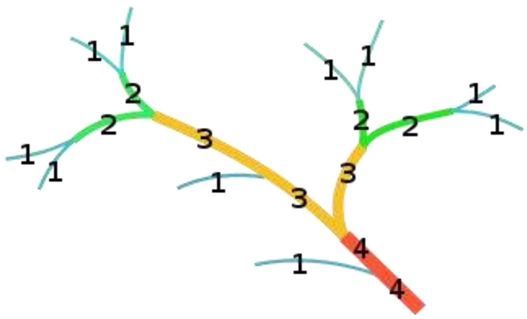


Figure 2.1 :   Catchment or drainage basin with the drainage divide (or watershed) as perimeter.

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).

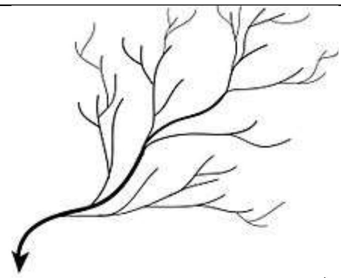
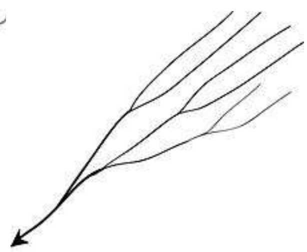
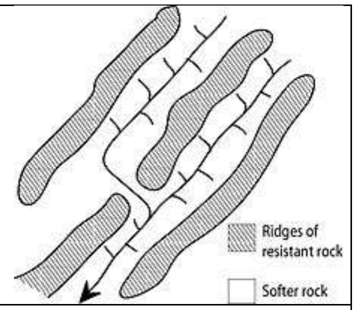
Figure 2.2 : Strahler's hierarchical ranking of *stream order.* The smallest stream has order 1 and at every confluence of similar order, the downstream order increase by 1.

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. World-wide, 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.**

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

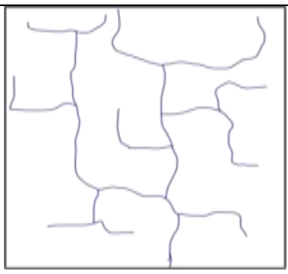
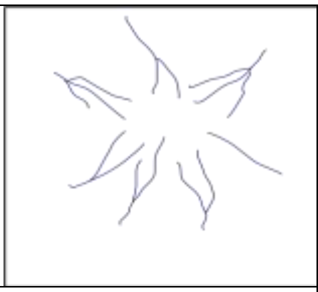
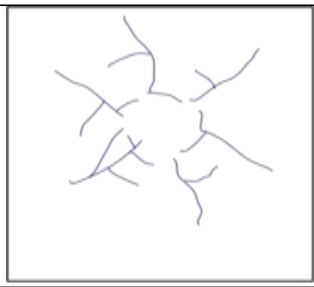
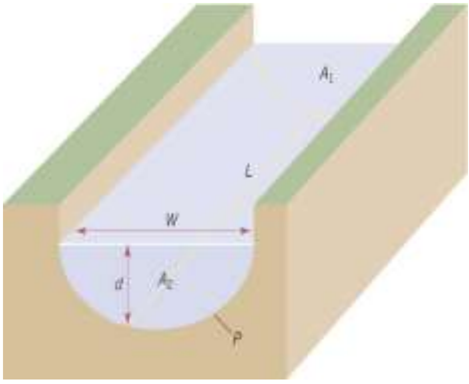
			
Rectangular Streams consist mainly of straight line segments with tributaries and bends joining at 90 degrees. Caused by preferential erosion of joints.	**Radial** Streams radiate outwards from a central high point, typically found on volcanoes.	**Centripetal** When runoff drains to a central depression 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) :	**Other stream dimensions:**
The product of the channel width, depth, and flow velocity $$Q = v\,w\,d$$ where: Q = **Flow rate (m^3/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 (m^2)	**A_2** cross-sectional area **P** wetted perimeter

Figure 2.4 Cross-section of channel system, showingsome important dimensions to determine discharge

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

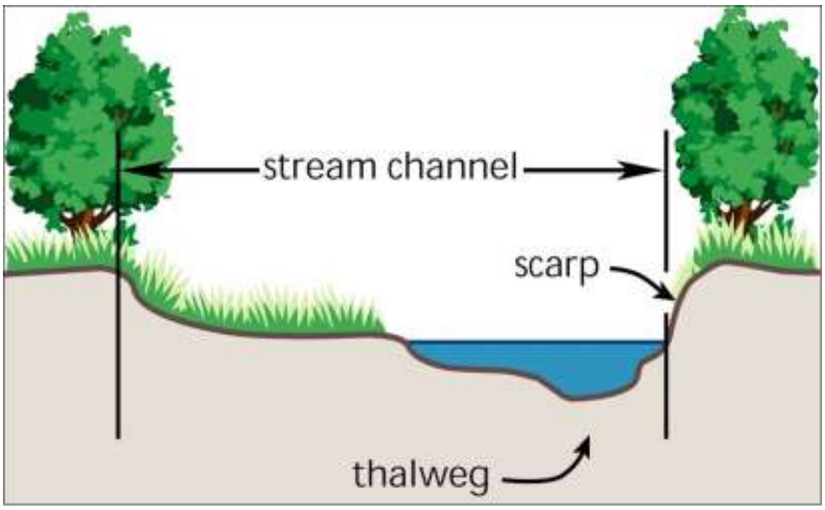


Figure.2.5 : Cross section of a stream channel. The thalweg is the deepest part of the channel.

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.**

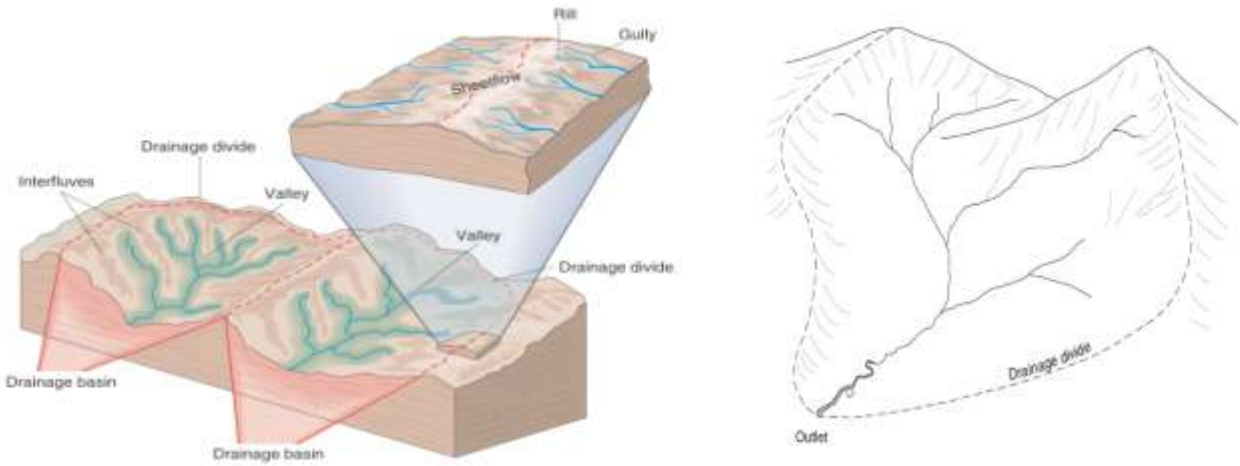


Figure 2.6 : The drainage basin or catchment as unit of investigation in fluvial morphology

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.**

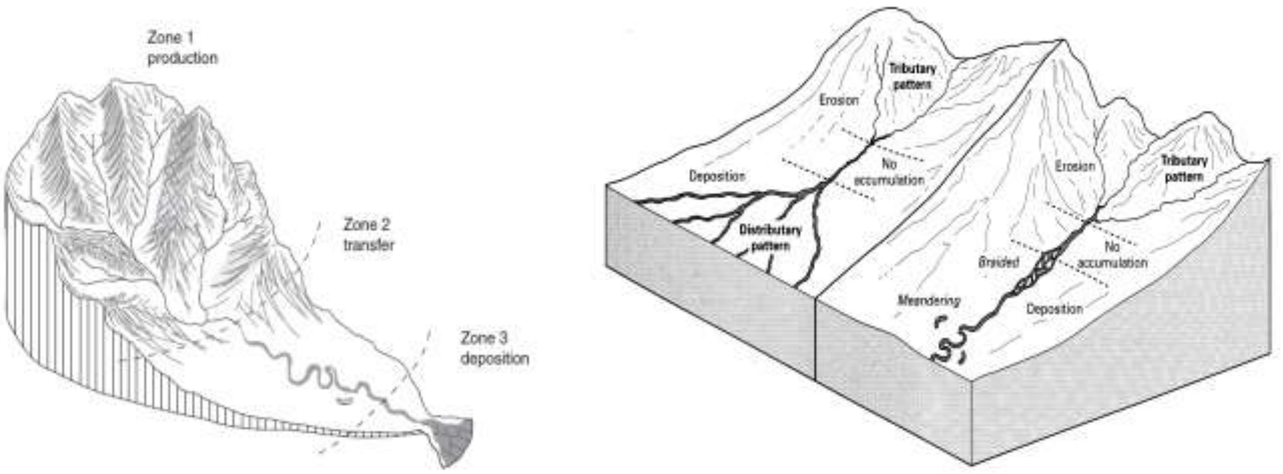


Figure 2.7 : Morphological subsystems of the fluvial system

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

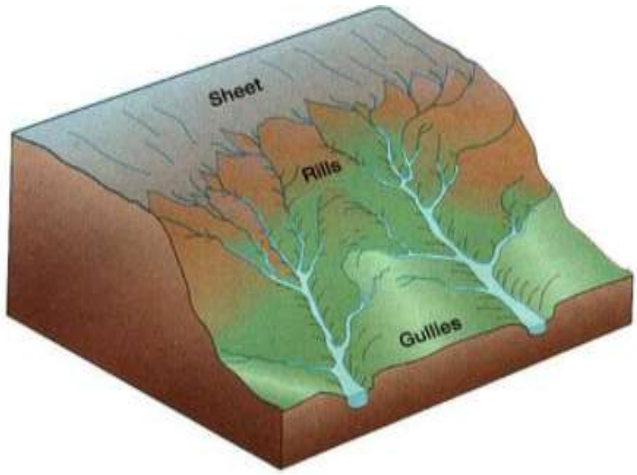
Flowing **water is the most important agent of erosion** in most environments on the continents and **stream valleys** are the **most common landforms shaped by these erosional processes,** with rivers flowing to the oceans draining about 68 % of the Earth's land surface. The remainder of the land either 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, r**unning 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 flgure 2.8.

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

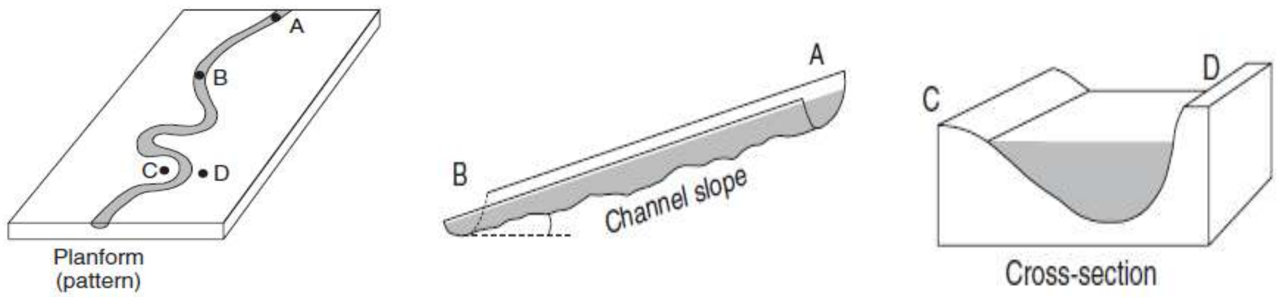
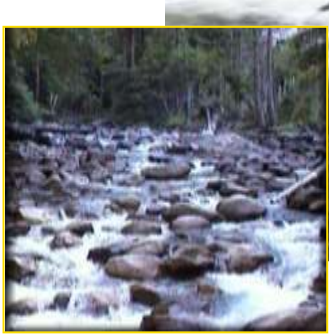


Figure 2.12 : Elements of three-dimensional channel form. The planform is the shape of the river viewed from above, the channel slope is shown for the reach between points A and B, and the channel cross-section between C and D.

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).

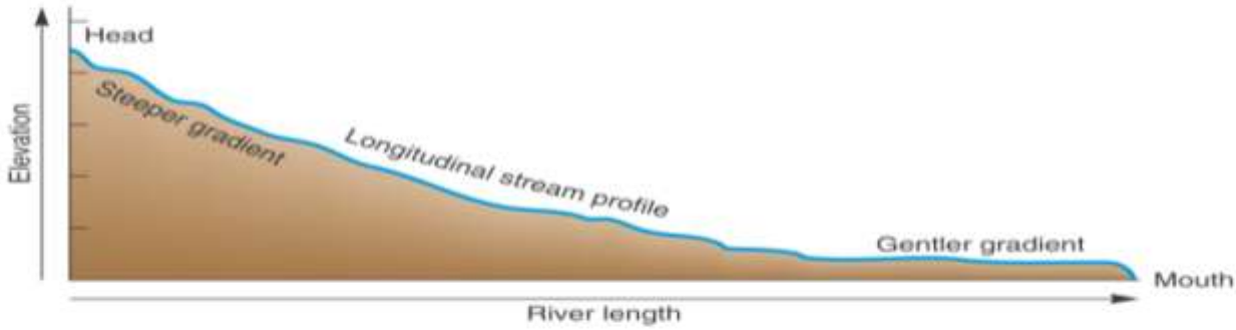


Figure 2.14 : The longitudinal profile of a river

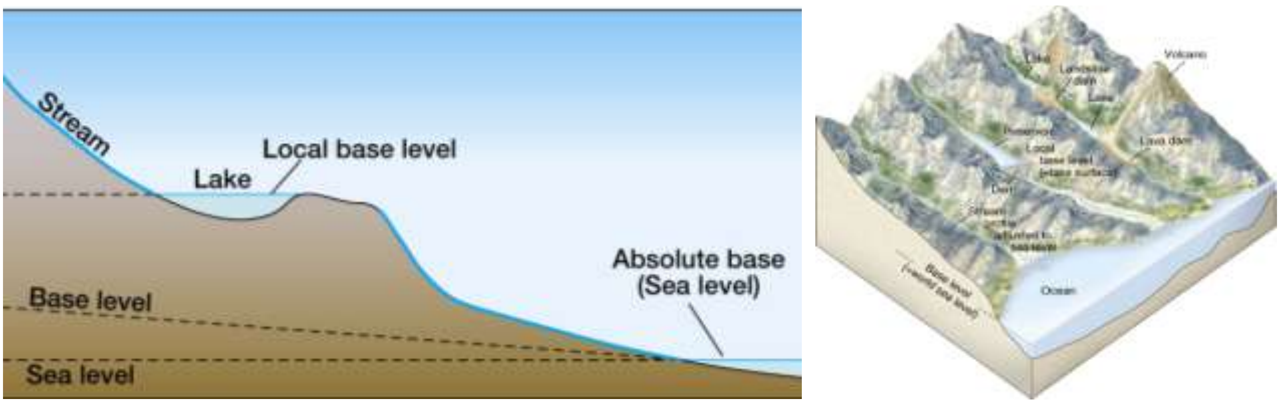


Figure 2.15 : The local and absolute base level of a typical river.

The limiting level below which a stream cannot erode the land is called the base level of the stream. Formally the **base level** is t***he 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):

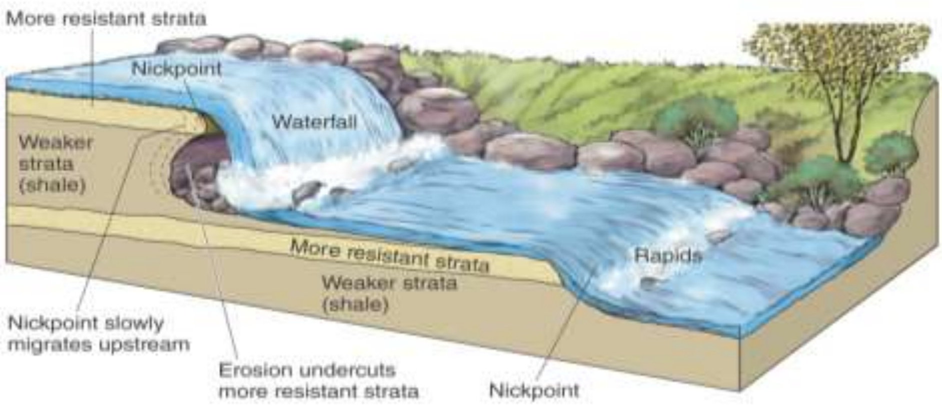


Figure 2.16 : Knickpoint

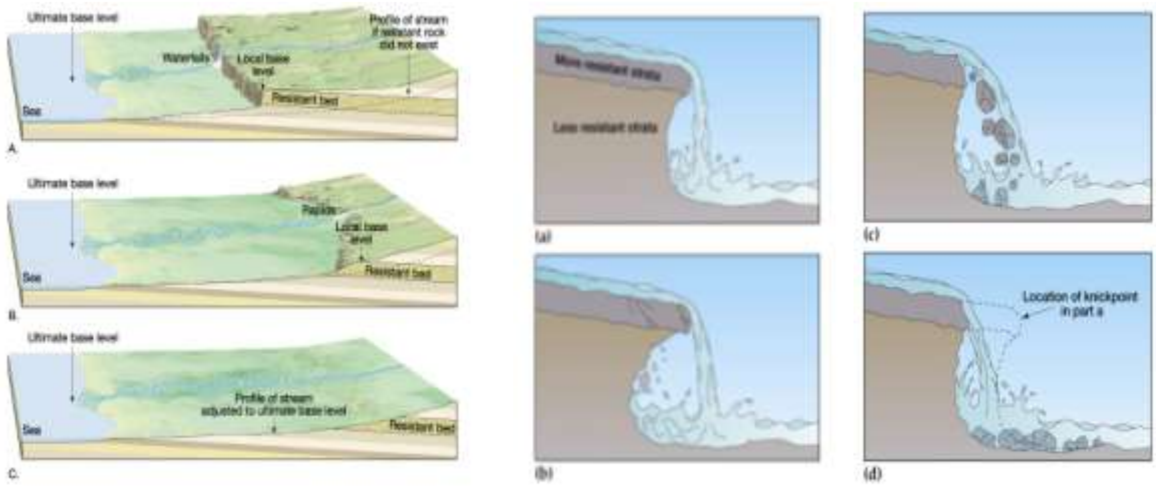


Figure 2.17 : Flattening of landscape and knickpoint migration due to continued erosion over time.

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 down-slope, 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.

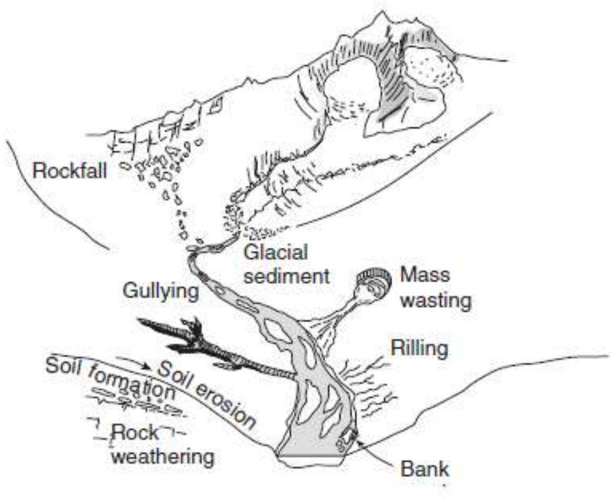


Figure 2.18 : Different sediment sources.

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.

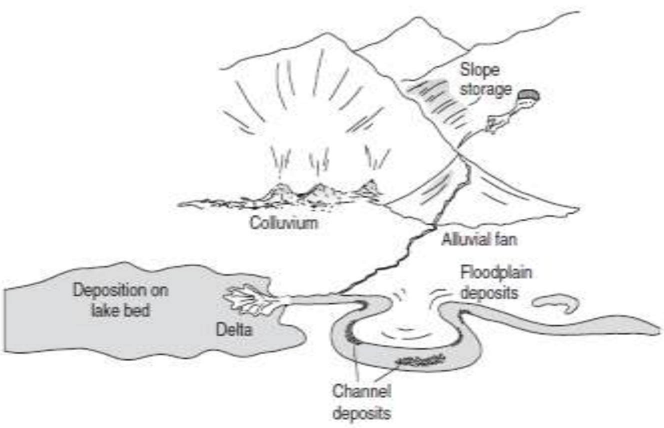


Figure 2.19 : Sediment stores.

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.

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.

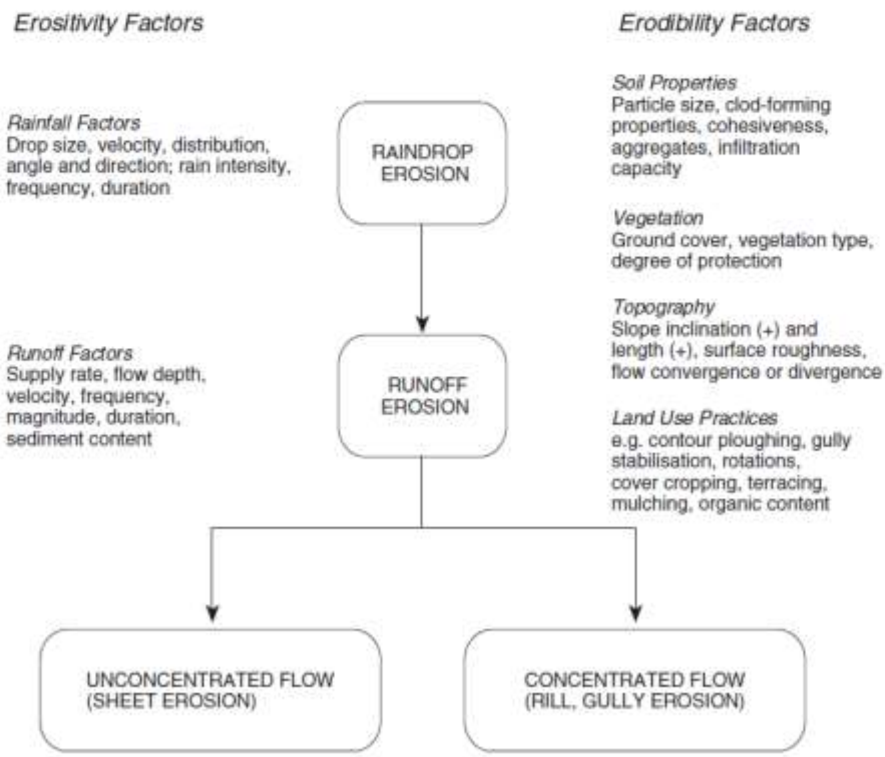


Figure 2.20 : The main factors controlling water erosion on hillslopes and sediment yield to river channels. After Cooke and Doornkamp (1990).

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:

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 raindrops, 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.

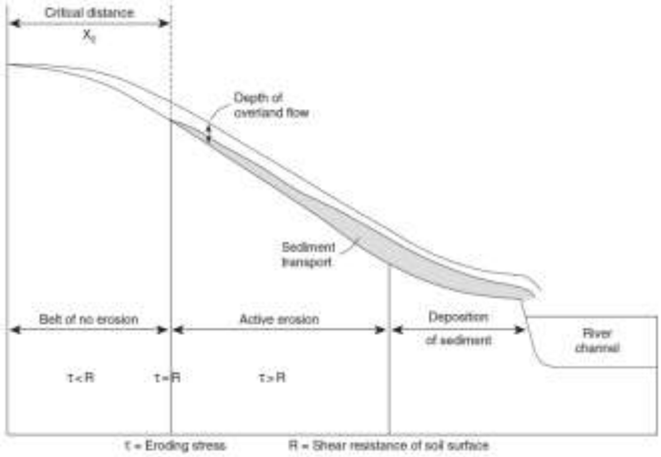


Figure 2.22 Horton's model of overland flow and rill formation. Adapted from Horton (1945).

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

Eventually the rill grows into a gully, and if the **gully 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**.

Gullying

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.

Figure 2.25 : Arroyo

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

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).

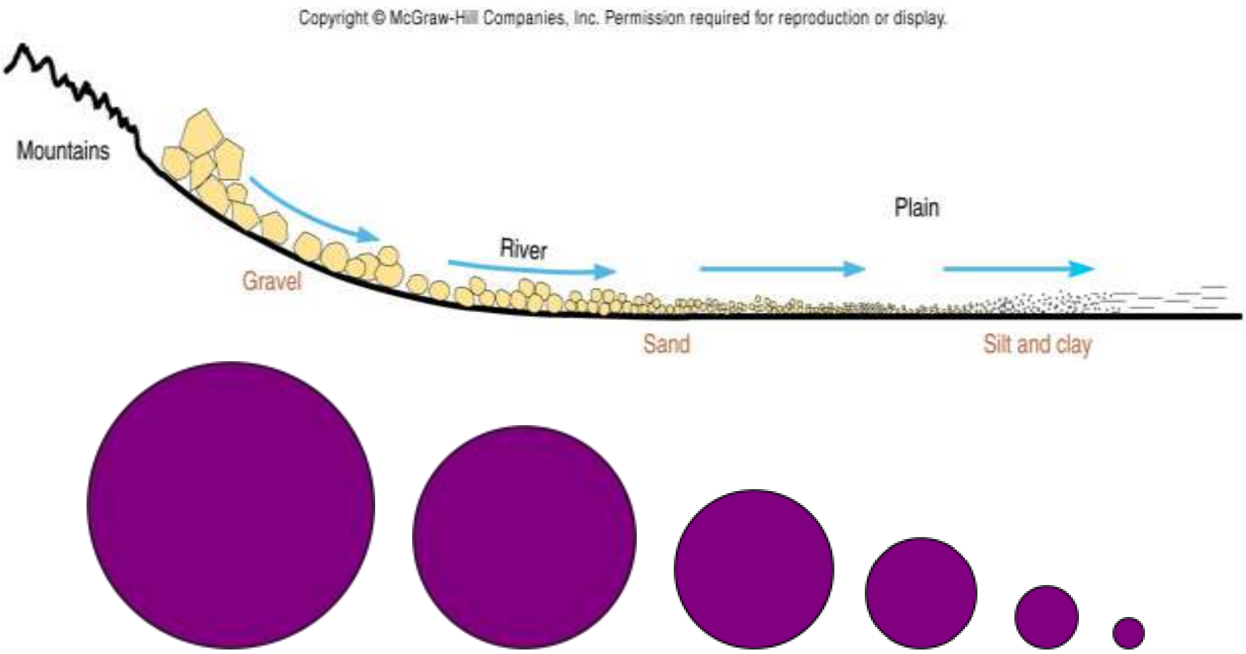


Figure 2.26 : River sediment particles decrease in size further downstream

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.

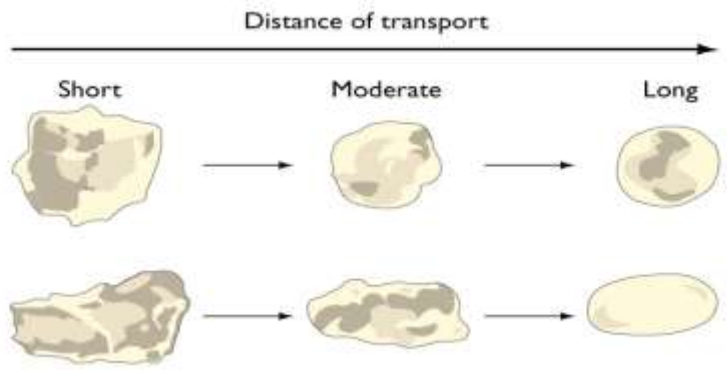


Figure 2.27 : The effect of abrasion in forming smoothed surfaces and textures of particles downstream.

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, HCO_3) 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.

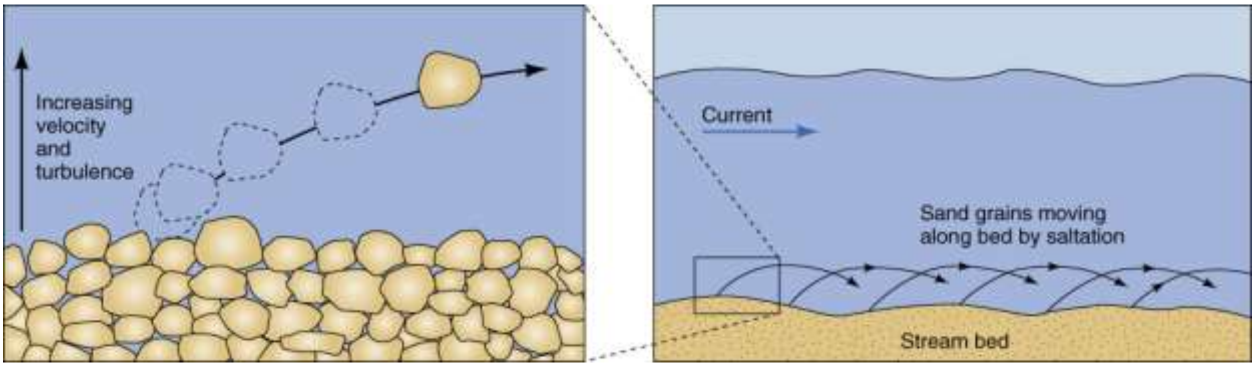


Figure 2.28 : Bedload transport by means of saltation

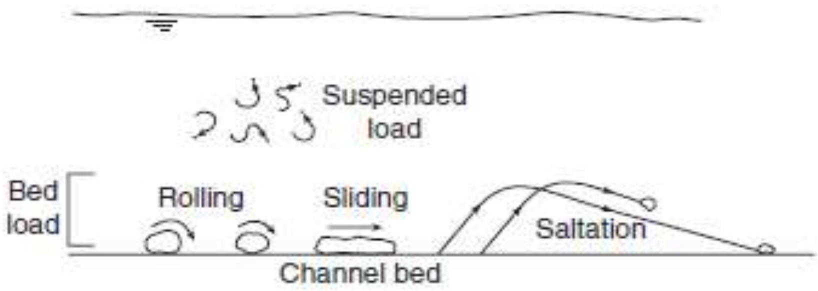


Figure 2.29 : Modes of bedload sediment transport.

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 tota l 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.

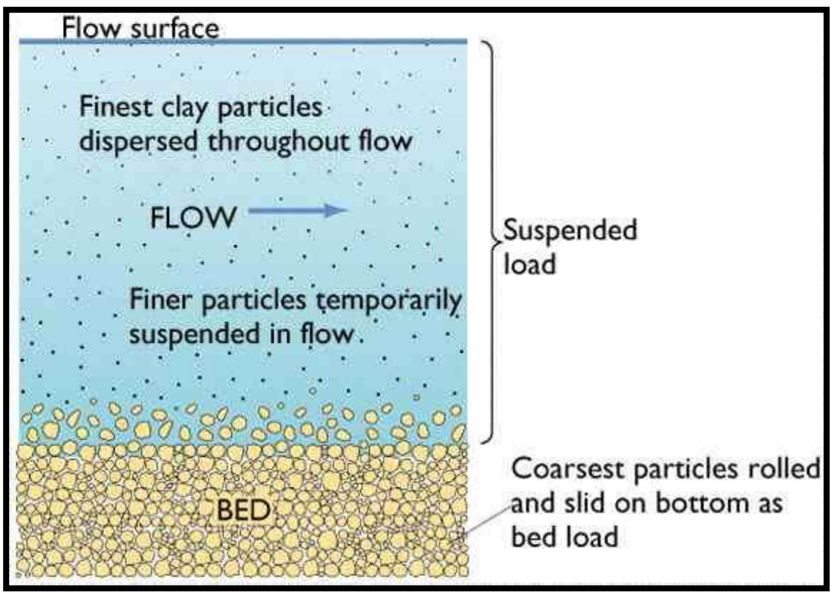


Figure 2.30 : Suspended load

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.

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.*

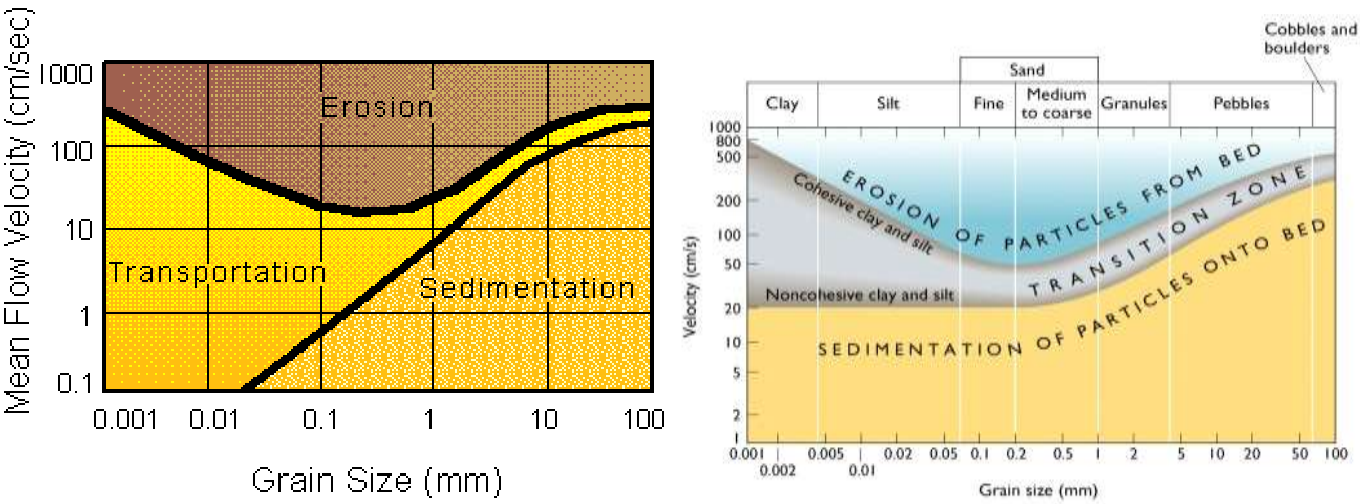


Figure 2.31 : Two versions of the Hjulström diagram, indicating the relationship between sediment size and flow velocity in determining when deposition, transport or erosion will take place.

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

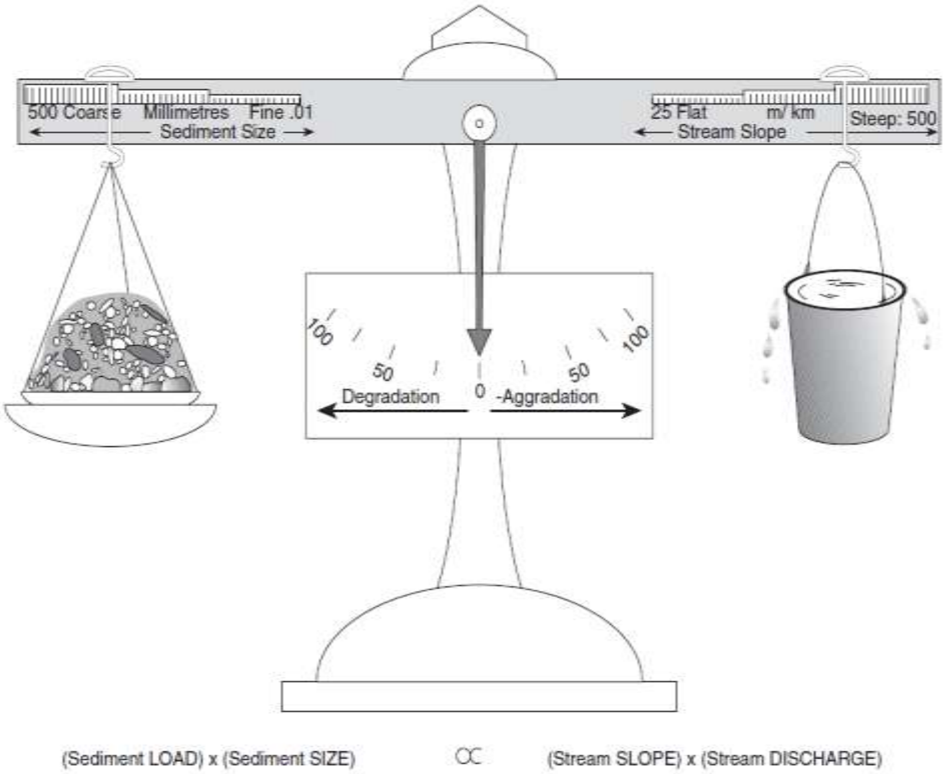


Figure 2.32: The Lane balance between stream power (stream slope / discharge) and sediment supply (sediment load / sediment supply). Adapted from Brierley and Fryirs (2005).

$\mathbf{Q_s * d_{50} \sim Q_w * S}$

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.