Two Years

Post Graduate Degree Programme (CBCS) in Geography

Semester – IV

Paper Code: GEO/DSE/FG/T-419

Fluvial Geomorphology-II: CHANNEL MORPHOLOGY (Special Paper)

Self Learning Material

Directorate of Open and Distance Learning (DODL)
University of Kalyani
Kalyani, Nadia
West Bengal, India

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Director's Message

Satisfying the varied needs of distance learners, overcoming the obstacle of distance and reaching the unreached students are the threefold functions catered by Open and Distance Learning (ODL) systems. The onus lies on writers, editors, production professionals and other personnel involved in the process to overcome the challenges inherent to curriculum design and production of relevant Self Learning Materials (SLMs). At the University of Kalyani a dedicated team under the able guidance of the Hon'ble Vice-Chancellor has invested its best efforts, professionally and in keeping with the demands of Post Graduate CBCS Programmes in Distance Mode to devise a self-sufficient curriculum for each course offered by the Directorate of Open and Distance Learning (DODL), University of Kalyani.

Development of printed SLMs for students admitted to the DODL within a limited time to cater to the academic requirements of the Course as per standards set by Distance Education Bureau of the University Grants Commission, New Delhi, India under Open and Distance Mode UGC Regulations, 2020 had been our endeavour. We are happy to have achieved our goal.

Utmost care and precision have been ensured in the development of the SLMs, making them useful to the learners, besides avoiding errors as far as practicable. Further suggestions from the stakeholders in this would be welcome.

During the production-process of the SLMs, the team continuously received positive stimulations and feedback from Professor (Dr.) Amalendu Bhunia, Hon'ble Vice- Chancellor, University of Kalyani, who kindly accorded directions, encouragements and suggestions, offered constructive criticism to develop it within proper requirements. We gracefully, acknowledge his inspiration and guidance.

Sincere gratitude is due to the respective chairpersons as well as each and every member of PGBOS (DODL), University of Kalyani. Heartfelt thanks is also due to the Course Writers-faculty members at the DODL, subject-experts serving at University Post Graduate departments and also to the authors and academicians whose academic contributions have enriched the SLMs. We humbly acknowledge their valuable academic contributions. I would especially like to convey gratitude to all other University dignitaries and personnel involved either at the conceptual or operational level of the DODL of University of Kalyani.

Their persistent and co-ordinated efforts have resulted in the compilation of comprehensive, learner-friendly, flexible texts that meet the curriculum requirements of the Post Graduate Programme through Distance Mode.

All the Self Learning Materials are self-writing and collected from e-book, journals and websites.

Director
Directorate of Open and Distance Learning
University of Kalyani

Syllabus

Semester –IV

Paper Code: GEO/DSE/FG/T-419

Paper: Fluvial Geomorphology-II: CHANNEL MORPHOLOGY (Special Paper)

Internal Evaluation/ Assessment – 10; Examination/Report/ Viva Voce – 40 (Semester end Examination); Credit – 4; Marks -50

Unit-01	Hydrological properties of channels, concept of equilibrium
Unit-02	Channel form and controls on its adjustment
Unit-03	Channel shape: pools, riffles, and bars, channel asymmetry and bed asymmetry; Hydraulic geometry, bed configuration
Unit-04	Concept of most efficient channel: width, mean depth, maximum depth, channel form index
Unit-05	Channel pattern: straight, sinuous, meandering, braided, anabranching, anastomosing
Unit-06	Meander geometry
Unit-07	Channel longitudinal profile and gradient, causes of profile concavity
Unit-08	Formation of channel bars, alluvial fans, floodplains, deltas, and estuaries
Unit-09	Channel decay: causes and consequences
Unit-10	Channel change through time: causes and evidences of channel shifting
Unit-11	Construction of dams and barrages and their impact on the fluvial system
Unit-12	Stream corridor: strategies for management

Contents

INTRODUCTION:

In the field of geography and environmental science, the study of channels and their formative processes is integral to understanding landscape dynamics and hydrological systems. Channel morphology encompasses the physical characteristics and evolution of river and stream channels, including their shape, size, patterns, and changes over time.

In this module, we will delve into the fundamental concepts of channel morphology, exploring how natural and human-induced factors influence channel formation and evolution. By the end of this material, you will have gained a comprehensive understanding of The Importance of Channel Morphology: Channels serve as conduits for water flow, shaping the landscape and influencing ecological habitats. Understanding channel morphology is crucial for managing water resources, mitigating floods, and preserving aquatic ecosystems.

Key Components of Channel Morphology: We will examine the key components that define channel morphology, such as channel cross-sections, longitudinal profiles, sinuosity, bed materials, and bank characteristics.

Processes Shaping Channel Form: Learn about the dynamic processes that shape channel morphology, including erosion, sediment transport, deposition, and meander migration.

Human Impacts on Channel Morphology: Human activities like damming, channelization, and urbanization significantly alter natural channel morphology, leading to environmental consequences and management challenges.

Applications in Geomorphology and Engineering: Gain insights into how the principles of channel morphology are applied in fields like river restoration, floodplain management, and hydraulic engineering.

Throughout this self-learning material, we encourage you to engage actively with the content, reflecting on real-world examples and case studies. By the end of this module, you will be equipped with a foundational knowledge of channel morphology that can be applied to various aspects of geography, environmental science, and water resource management.

LEARNING OBJECTIVES

By the end of this self-learning module on channel morphology, you will be able to:

Define and explain the key concepts related to channel morphology, including channel cross-section, longitudinal profile, sinuosity, and sediment transport.

Understand the processes that shape channel morphology, such as erosion, deposition, and meander migration.

Analyze the impacts of natural and human-induced factors on channel morphology.

Apply knowledge of channel morphology to real-world scenarios in geography, hydrology, and environmental science.

Evaluate management strategies for mitigating the impacts of human activities on channel morphology.

ASSESSMENT OF PRIOR KNOWLEDGE

Before beginning this module, it's beneficial to assess your current understanding of related topics:

Self-Assessment Quiz: Complete a brief quiz to gauge your familiarity with basic geomorphological concepts.

Review of Previous Study: Reflect on any coursework or experience related to river systems, hydrology, or landscape evolution.

Identifying Learning Objectives: Identify which learning objectives are already familiar and which ones are new or require further exploration.

LEARNING ACTIVITIES

During this self-learning journey, engage in the following activities to enhance comprehension and retention:

Reading Assignments: Read assigned sections on channel morphology from recommended textbooks or scholarly articles.

Interactive Multimedia: Watch videos or animations demonstrating channel processes like erosion, deposition, and meander formation.

Case Studies: Analyze case studies of river restoration projects or urban stream management to understand practical applications.

Virtual Field Trips: Explore virtual field trips to rivers and streams to observe channel morphology firsthand.

Problem-Solving Exercises: Solve problems related to calculating channel cross-sectional area, sediment transport rates, or channel adjustment over time.

Discussion Forums: Participate in online discussions to share insights and ask questions about channel morphology topics.

FEEDBACK OF LEARNING ACTIVITIES

To enhance learning outcomes and address any gaps in understanding, receive feedback through:

Self-Assessment Reflection: Reflect on your progress and identify areas for improvement based on completed activities.

Peer Review: Engage in peer review sessions where you provide feedback on each other's problem-solving exercises or case study analyses.

Instructor Feedback: Seek feedback from instructors or mentors on completed assignments or discussion contributions.

Assessment Tools: Use quizzes or assessments at the end of each module to evaluate your comprehension and receive automated feedback.

UNIT-1: HYDROLOGICAL PROPERTIES OF CHANNELS, CONCEPT OF EQUILIBRIUM

The term fluvial is derived from the Latin *fluvius*, meaning river. Fluvial geomorphology is the study of the interactions between river channel forms and processes at a range of space and time scales. The influence of past events is also significant in explaining the present form of river channels. Rivers are found in many different environments and show an amazing diversity of form.

DIVERSITY OF FORM

A quick look through the photographs in this book will give you some idea of the variety that can be seen in rivers and streams worldwide. Rivers drain much of the land area – with the exception of regions that are hyperarid or permanently frozen – and their variety reflects the vast range of different environments in which they are found. Climate, geology, vegetation cover and topography are just some of the factors that influence river systems. 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, originating in a humid area before flowing through an arid region. Examples of these 'exotic' rivers include the Nile and Colorado, both of which sustain agriculture and urban centres in desert regions. Perennial rivers flow for all or most of the year, while many of those in dryland environments only transmit water at certain times. The 'trail' shown in Colour Plate 1 is actually an ephemeral channel that was photographed in South Africa during the winter dry season. A small herd of cattle in the distance provide an idea of scale. The material in which the channel is formed is called the channel substrate. An important distinction can be made between bedrock and alluvial substrates (Figure 1.1). 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 (Figure 1.1b). The ephemeral channel shown in Colour Plate 1 is formed in alluvium. Colour Plate 2 shows an example of a bedrock channel, while Colour Plate 3 shows a mixed channel which has a rock bed and alluvial banks. Most rivers flow to the oceans, although some drain to inland seas and lakes, while others dry up completely before reaching the ocean. 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 1.2). This area is bounded by a drainage divide or catchment boundary, something that is clearly visible as a ridge in mountainous areas but which can be rather difficult to discern in areas of more subdued topography. The outlet, where the main channel exits the basin, is at a lower elevation than the rest of the basin area. Drainage basins form a mosaic across the land surface, varying greatly in size from a few hectares to millions of square kilometres. Within each drainage basin is a branching network of channels. The main, or trunk, channel is fed by numerous small tributaries which join to form progressively larger channels. Drainage patterns, as viewed from the air or on maps, vary considerably between basins. The development and evolution of drainage networks is influenced by a number of factors, including geology, climate and long-term drainage basin history. Further information on the ways in which structural controls influence drainage patterns can be found in Box 1.1. In terms of the actual form of different channels, obvious differences can be seen, even along the same river. One of the more noticeable things is the variation in size, from tiny headwater streams that are just 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 discharge supplied from upstream. 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. As you move downstream, discharge and channel size generally increase with the upstream drainage area. While many rivers follow a single channel, there are also numerous examples of rivers with multiple channels. Those that flow in a single channel usually tend to deviate from a straight line, sometimes following rather an irregular path, in other cases forming more regular meanders (Colour Plates 4 and 5). Braided channels are characterised by numerous bars and islands formed by sediment deposits in the channel (Colour Plates 6 and 7). A rather different type of multiple channel are the anabranching channels shown in Colour Plates 8 and 9. Rather than being separated by bars, the individual channels are cut into the floodplain itself. When considering channel form, individual sections, or 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. As a result, different channel patterns may be found along the same channel. At the reach scale – typically a few tens to hundreds of metres or more – there is a homogeneity of form. Rivers are three-dimensional in shape and, in addition to the channel planform patterns described above, variations are also seen in cross-sectional shape and channel slope (Figure 1.3). 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. You will see later in the book how certain groupings, or assemblages, of these features are associated with different channel types.

The precise definition of **equilibrium** is also time dependent. Equilibrium refers to a state of balance within a system, or sub-system. Negative feedback mechanisms help to maintain the system in an equilibrium state, buffering the effect of changes in the external variables. However, different types of equilibrium may exist at different time scales. These were defined by Schumm (1977) with reference to changes in the elevation of the bed of a river channel above sea level. If you were to observe a short section of river channel over a period of a few hours you would not see any change in its form (unless there happened to be a flood), although you might see some sediment transport. Over this short time period the channel is said to be in a state of static equilibrium (Figure 2.5a).

The same river, observed over a longer time scale of a decade, would show some changes. During this time, floods of various sizes pass through the channel, scouring the bed. In the intervening periods, deposition builds up the channel bed again. As a result of these cycles of scour and fill, the elevation of the channel bed fluctuates around a constant average value (Figure 2.5b) and steady state equilibrium exists. Over longer time scales, from thousands to hundreds of thousands of years or more, erosion gradually lowers the landscape. At these time scales, the channel elevation fluctuates around a changing average condition, the underlying trend being a reduction in channel elevation. This is called dynamic equilibrium and is illustrated in Figure 2.5c. As you know, the influence of the external basin controls cannot be ignored. Changes in any of these variables can lead to positive feedbacks within the system and a shift to a new equilibrium state. For example, in tectonically active regions, the section of channel might be elevated by localised uplift. Such episodes of change occur over much shorter time scales than the gradual evolution of the landscape, resulting in abrupt transitions. This type of equilibrium delights in the term dynamic metastable equilibrium (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-2: CHANNEL FORM AND CONTROLS ON ITS ADJUSTMENT

The form of a channel is largely a function of the water and sediment supplied to it. Adjustments to channel form occur as a result of process feedbacks that exist between channel form, flow and sediment transport. At the reach scale, the type of adjustment that can take place is constrained by the valley setting, the nature of bed and bank materials, and bank vegetation. This gives rise to a wide diversity of different channel forms.

CONTROLS ON CHANNEL ADJUSTMENT AND FORM

Flow and sediment supply both fluctuate through time, meaning that continuous adjustment takes place through the erosion, reworking and deposition of sediment. The flow and sediment regimes are called driving variables because they drive these processes. Along a given reach, channel adjustment is constrained within certain boundaries that are imposed by local conditions. For example, a sand-bed river flowing across a wide floodplain is able to adjust its form much more readily than a bedrock channel confined within a narrow gorge. Energy availability is also important, and channel adjustments are often limited for rivers that flow over low gradients, especially where cohesive banks are protected by vegetation. These constraints are called boundary conditions and include valley confinement, channel substrate, valley slope and riparian vegetation. Figure 8.1 provides a schematic representation of the driving variables and boundary conditions that influence channel form. As discussed in Chapter 3, p. 32 a channel is said to be 'in regime' when its form fluctuates around an equilibrium condition over the time scale considered. Not all channels are in regime, and there are many examples of non-regime, or disequilibrium, channels. This may be because the channel is evolving in response to long term changes in the flow or sediment regime, caused by a change in one of the external basin controls. Examples include incising or aggrading channels and those that are undergoing a change in channel pattern. Alternatively, some bedrock and dryland channels may exist in a permanent state of disequilibrium because it is only during flood flows that adjustments take place.

In such cases, low flows have little or no influence on the overall channel form. Many empirical relationships have been developed to relate 'regime dimensions' (e.g. channel width or depth), to control variables (e.g. bankfull discharge). It is important to realise that these regime dimensions represent an average and are not applicable to all channel types or flow regimes. As you saw in Chapter 7, the available stream power along a given reach is determined by the discharge and valley slope. At the sub-reach scale there are spatial variations in energy expenditure, which result from variations in channel shape and resistance to flow. These in turn influence patterns of erosion and deposition. For example, energy and erosion potential are concentrated where the channel narrows. Conversely, flow resistance is increased by obstructions to flow such as boulders, bedforms, bars or woody debris, which can lead to localised deposition. There is therefore two-way feedback between channel form and flow hydraulics – form influences flow and flow influences form.

This point is well illustrated by the work of Ashworth and Ferguson (1986) on a glacially fed braided river in Arctic Norway. An intensive monitoring programme was carried out to make detailed measurements of channel morphology, velocity and shear stress, bedload size and transport rate, and the size of bed material. Ashworth and Ferguson's diagram of the feedbacks between channel processes, morphological changes and sediments is shown in Figure 8.2. Starting at the top left of this diagram is the discharge of the river, which is unsteady (varies over time). The irregular form of the channel creates non-uniform flow conditions over the rough channel bed. As a result, complex spatial variations are seen in velocity, which also changes over time. At any point in the channel, the bed shear stress is determined by the vertical velocity profile (Chapter 6, pp. 87–90). Rates of bedload transport are determined by bed shear stress as well as the size and amount of bed material that is available for transport. As with velocity and shear stress distributions, rates of bedload transport are spatially variable, and also change with time. Bedload transport may maintain the existing channel shape, size and pattern. Alternatively, channel form can be modified as a result of scour, fill and possible lateral migration. The nature of such changes is spatially variable, and in turn feeds back to influence the velocity distribution within the channel. Bedload transport also governs the size distribution and structure of bed sediments through selective entrainment and transport (Chapter 7). The character of the bed material determines the roughness of the channel, in turn affecting the velocity distribution in the channel. Driving variables Flow regime The flow in natural river channels is unsteady, fluctuating through time in response to inputs of precipitation to the drainage basin. Characteristics of the flow regime, discussed in Chapter 3, include seasonal variations, flood frequency–magnitude relationships and the frequency and duration of low flows. Since discharge influences stream power, velocity and bed shear stress, the characteristics of the flow regime have an important influence on channel form. Of morphological significance is the bankfull discharge. This was defined in Chapter 3 and is the discharge at which the channel is completely filled with water. The bankfull discharge marks a morphological discontinuity between within-bank and out-of-bank flows. Since the flow in natural channels is unsteady, the bankfull discharge provides a representative flow. As discussed in Chapter 3, channels are shaped by a range of flows. The geomorphological effectiveness of a given flood depends not only on its size, but also on the frequency with which it occurs. Large floods can carry out a considerable amount of geomorphological work. However, their comparative rarity means that the cumulative effect of smaller, more frequent flows may be more significant in shaping the channel. Box 3.3 provides a fuller explanation of the magnitude and frequency of channel-forming flows. Bankfull discharge (or the equivalent bankfull width) has often been used in developing statistical relationships between discharge and channel form parameters. It is important to realise that bankfull discharge is actually quite difficult to define and that its frequency of occurrence varies considerably between different rivers (see Chapter 3, p. 32).

Sediment regime

The supply of sediment varies through time. It is not only the volume of sediment that is important but also its size distribution. As you will see later in this chapter, there are significant differences in the behaviour and morphology of bedload, suspended load and mixed load channels. Fluctuations in the volume and size of sediment are brought about by variations in sediment supply from the drainage basin (Chapter 4) and processes of sediment transfer through the channel network (Chapters 5 and 7). As with the flow regime, it is the processes in the drainage basin, upstream from a given reach, that influence sediment supply. The balance between stream power and sediment supply There is an important balance between the supply of bedload at the upstream end of a channel reach and the stream power available to transport it. This is known as the Lane balance, having first been described as a qualitative equation by Lane in 1955. This balance is illustrated schematically by the pair of scales shown in Figure 8.3. The left hand side of the scales represents the volume and size of sediment supplied to a channel reach over a given period of time. Balanced against this is the stream power available to transport it. This is determined both by the volume of water that enters the reach (over the same time period), and by the slope over which it flows. If the stream power is exactly sufficient to transport the sediment load, both sides of the scales are in balance and there is no net erosion or deposition along the reach. This is not to say that there is no erosion or deposition whatsoever, because these processes do occur at a localised scale in response to local variations in hydraulic conditions. Rather it means that, on balance, neither erosion nor deposition will predominate. An imbalance will occur if there is an increase in the volume or calibre of the sediment load in relation to the available stream power (sediment calibre is important because it determines the flow competence required to transport it). This means that there is insufficient stream power to transport all the sediment, with the result that the excess is deposited along the reach. In this case, the balance tips towards aggradation, with net deposition occurring along the reach. Aggradation can be triggered in several ways, for example where the sediment supply is increased by upstream channel erosion, mass movement, or human activities such as mining. A particularly dramatic example of mining-induced aggradation along the Ok Tedi River in Papua New Guinea is shown in Colour Plate 16. Aggrading channels are characterised by numerous channel bars in a wide, shallow channel. Deposition within the channel may lead to the channel bed becoming elevated above the surface of the floodplain. This, together with reduced channel capacity, increases the incidence of flooding and also promotes channel migration. A different situation arises when the stream power exceeds what is needed to transport the sediment load through the reach. This excess energy has to be expended somehow, so it is used to entrain sediment from the bed and erode the channel boundary. In this case degradation predominates. An example of a degrading channel is shown in Plate 8.1. Degradation can be caused by an increase in discharge, perhaps caused by an increase in flood frequency, or by a decrease in sediment supply. This can occur downstream from dams or where gravel mining has removed sediment from the river bed (Chapter 5). The Lane balance is simplistic because much depends on the calibre of bed sediment within the reach. For example, no degradation can occur in a boulder-bed stream if the bed sediment is too coarse to be moved by the available stream power. This can be true even if the stream power exceeds the sediment supplied at the upstream end of the reach. Even when degradation does occur, another limitation of the equation is that it does not tell us where within the reach erosion will occur (Simon and Castro, 2003). This means that the equation cannot be used to predict the actual nature of channel change. For example, if the channel bed is more resistant to erosion than the banks, bank erosion is likely to be an initial adjustment. However, in a sand-bed channel with cohesive banks it is more likely that an initial adjustment would be scouring of the bed (Simon and Castro, 2003). Resistance to erosion can be highly variable within a given reach, as can the specific stream power along that reach. This gives rise to spatially complex adjustments along the reach, even if there is net aggradation or net degradation along the reach as a whole.

Boundary conditions

Valley slope This refers to the downstream slope of the valley floor (as opposed to the slope of the channel itself) and determines the overall rate at which potential energy is expended along a given reach. The valley slope imposed on a given reach of channel is determined by a combination of factors including tectonics, geology, the location of the reach within the drainage basin and the long-term history of erosion and sedimentation along the valley. Although the overall energy available along a given reach is largely determined by the valley slope, it is possible for adjustments to occur that increase flow resistance at different scales (channel resistance, form resistance and boundary resistance). Different types of channel and floodplain morphology are associated with low, medium and high-energy environments.

Valley confinement

A channel may be defined as confined, partly confined, or unconfined, depending on how close the valley sides are. The degree of valley confinement is important for several reasons. In confined settings (Plate 8.2) channel adjustments are restricted by the valley walls, which also increase flow resistance. In addition, valley width influences the degree of slope–channel coupling that exists. Inputs of sediment from mass movements and other slope processes may exceed transport capacity, in turn influencing channel form. The episodic nature of mass movements means that these contributions can vary considerably over time. In partly confined settings (an example is shown in Colour Plate 10), some degree of lateral migration and floodplain development is possible. However, where the river comes against the valley wall or hillslope it is prevented from migrating further, which can lead to the development of over-deepened sections of channel. Stream power is also concentrated within the narrow valley and sections of the floodplain surface may be stripped during major floods. Where the hillslopes are a long way from the channel and have relatively little influence in contributing to the channel load, the channel is described as unconfined (Colour

Plate 4). Typically these settings are found in the lower reaches of rivers where there is very little interaction between channel and hillslopes.

Channel substrate

Considerable variations are seen in the form and behaviour of channels developed in different substrates. The substrate determines how resistant the channel is to the erosive force of the flow. It also influences boundary roughness, and therefore flow resistance (Chapter 6, pp. 79–80). Alluvial channels formed in sand and gravel are generally more easily adjusted than those with cohesive silt and clay substrates. This is because the individual particles can be entrained at relatively low velocities, so non-cohesive substrates tend to be associated with wider, shallower cross-sections and faster rates of channel migration. Bedrock and mixed bedrock-alluvial channels are influenced over a range of scales by various geological controls .

Riparian vegetation

Vegetation on the banks and bed of river channels controls channel form in various ways. It often acts to protect and strengthen the banks, and research has shown that a dense network of roots can increase erosion resistance by more than a factor of ten. As a result, channels with vegetated banks are often narrower than those with non-vegetated banks under similar formative flows. This effect is most marked for densely vegetated banks (Hey and Thorne, 1986). Flow resistance can also be increased by vegetation growing on the bed and banks, as well as by woody debris (fallen trees and branches) that enters the channel from the banks. An interesting example of the influence of riparian vegetation on channel form is provided by the Slesse Creek, British Columbia, Canada, and is reported by Millar (2000) and MacVicar (1999). The Slesse Creek drains an area of 170 km2 within the Fraser River basin, flowing southwards from the United States into British Columbia. The reach shown in Plate 8.3 is approximately 2 km north of the US–Canadian border. This photograph was taken in 1940 and shows a stable meandering channel with a width of approximately 30 m and bordered by native forest vegetation. During the 1950s and 1960s logging of this stream bank vegetation took place along the British Columbia part of the channel. The second photograph (Plate 8.4) shows the same reach in 1993. Subsequent to the removal of forest, the channel has developed an unstable braided form, widening to approximately 150 km. Millar (2000) suggests that these changes are mainly attributable to reduced bank stability, resulting from logging along river banks. This is because there has been no logging in the upstream (US) part of the drainage basin, which is a protected conservation area, so it is assumed that there has been little change in the flow or sediment regimes.

Downstream changes

Downstream changes in slope, discharge, valley confinement, sediment supply and particle size give rise to different balances between erosion and deposition along different parts of the profile. This leads to downstream changes in channel and floodplain morphology. In general terms, the cumulative supply of sediment increases downstream but the available energy decreases. Figure 8.4(a) shows how selected channel parameters change through the sediment production, transfer and deposition zones. The discharge in most river channels increases in a downstream direction, as a progressively larger area is drained. In order to accommodate the growing volume of flow, channel dimensions (width and depth) typically increase downstream, and are often accompanied by a slight rise in velocity. The way in which these parameters change with increasing discharge can be described by the hydraulic geometry of the channel. Further information on hydraulic geometry relationships is provided in Box 8.1. Downstream reductions in bed material size reflect differences in the way in which coarse and fine sediment are transferred along the channel. In contrast to the relatively localised transport of bedload particles, fine material, carried in suspension, is transported over much greater distances (Chapter 5, p. 60). Observations show that there is a general decline in sediment size along the channel. The main causes of this downstream reduction are widely recognised as being abrasion and selective transport (Rice and Church, 1998). Abrasion refers to the reduction in size of individual particles by chipping, grinding and splitting. Physical and chemical weathering processes are also significant in the pre-weakening of individual particles. Selective transport refers to the longer travel distances associated with smaller grains, which are more mobile. The rate of reduction in sediment size varies considerably and downstream increases are often observed at several locations. The downstream decrease in sediment size is often disrupted by inputs of coarser material. These include material from bank erosion, inputs from tributaries, and colluvial material. Material entering the main stream from tributaries is typically coarser than that in the main channel (Knighton, 1998). This causes a sudden increase in sediment size followed by a progressive fining further downstream. Complex patterns of downstream size reduction are seen where slopechannel coupling is strong and non-alluvial supplies are dominant. These include contributions from hillslopes (e.g. mass movements), the erosion of bedrock outcrops and glacial material (Rice and Church, 1998). The channel slope (represented by the thick line in Figure 8.4a) is typically steepest in the headwaters, becoming gentler in the lower reaches. The resulting long profile of many rivers is concave in shape, although the degree of concavity varies. Downstream increases in discharge, together with a decrease in bed material size, mean that the load can be transported over progressively shallow slopes. Exceptions to this are seen in arid and semi-arid regions, where downstream conveyance losses and high rates of evaporation lead to a downstream reduction in discharge. In this case a straight or convex profile may develop, since increasingly steep slopes are needed to compensate for the downstream reduction in flow. Irregularities are often seen in the long profile, for example flatter sections are associated with lakes and reservoirs, and steeper sections at the site of waterfalls (see Figure 8.4b). In addition, there is often a change in the channel slope where tributaries join the main channel, because of the sudden increase in discharge. In tectonically active areas, where rates of uplift may be similar to erosion rates, rivers are in a state of dynamic equilibrium constantly trying to 'catch up' with tectonically driven changes. It takes time for a concave profile to develop, so the overall shape of long profiles in tectonically active regions tends to be straight rather than convex.

CHANNEL ADJUSTMENT

Time scales of adjustment

Different components of a channel's morphology (e.g. bedforms, cross-sectional shape, slope) change over different time scales (Figure 8.5). This is because some components are more readily adjusted than others. For example, bedforms in a sand-bed channel are rapidly modified by a wide range of flows. Adjustments to channel width and depth take place over months to years, planform adjustments occur over tens to hundreds of years, while changes in the long profile may take thousands of years. Morphological adjustments therefore tend to lag behind the changes that cause them. This means that it can be difficult to link processes of flow and sediment transport with channel dimensions and form. Channel form is directly controlled by flow regime and sediment supply. This chapter will deal with relatively short-term adjustments that are directly influenced by the flow and sediment regimes. Chapter 9 will examine longer-term changes that are brought about by changes in the external basin controls (climate, tectonics, base level and human activity). These all act as controls on the flow and sediment regimes and, through a complex sequence of adjustments, lead to long-term changes in channel form and behaviour How adjustments are made Channel form and behaviour reflect the driving variables and boundary conditions influencing a given channel reach. These controls also influence the ways in which channel adjustments are made. There are potentially four degrees of freedom, or variables, that can be modified: channel cross-section, slope, planform and bed roughness. Modifications to the cross-sectional size and shape are associated with changes in width and depth of the channel by processes such as bank erosion, incision of the bed, or aggradation. Channel slope can be adjusted in different ways. Negative feedback reduces the slope of steeper sections by erosion, and the slope of flatter sections is increased by deposition. Increases or decreases in channel length also affect channel slope, as illustrated in Figure 8.6. There are several different types of channel planform adjustments. These include lateral migration, meander bend development, reworking of bars, and even wholesale shifts of the channel to a new course. Finally, changes in bed roughness are brought about when the channel rearranges bed material, for example, in sand-bed channels, where bedforms are modified in response to changes in flow conditions. Mutual interrelationships exist between these variables, with adjustments made to one affecting one or more of the others. For instance, the formation of a meander cut-off alters the channel planform as well as increasing channel slope. The influence of the driving variables and boundary conditions often reduces the degrees of freedom that a particular channel has. In the case of the mixed bedrock-alluvial channel shown in Colour Plate 3, depth increases are greatly restricted by the rock bed of the channel. On the other hand, the alluvial banks allow the channel to be widened much more easily. However, reductions in cross-sectional size by deposition may be limited if the channel has degradational tendencies (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-3: CHANNEL SHAPE: POOLS, RIFFLES, AND BARS, CHANNEL ASYMMETRY AND BED ASYMMETRY; HYDRAULIC GEOMETRY, BED CONFIGURATION

Geomorphic units are features that form at the subchannel scale and can be erosional or depositional in origin. Distinctive assemblages or groupings of geomorphic units characterise the different channel types introduced in Chapter 1. For instance, braided channels contain numerous mid-channel bars, while bedrock channels are associated mainly with erosional features such as potholes and bedrock steps, although bars can also form if sufficient bed sediment is available. Geomorphic units also affect hydraulic processes, and provide a range of different habitats for in-stream flora and fauna.

Bars

Bars are in-channel accumulations of sediment which may be formed from boulders, gravel, sand or silt. Bars can be divided into two broad groups: unit bars and compound bars (Smith, 1974). Unit bars are relatively simple bar forms whose morphology is mainly determined by processes of deposition (Ashmore, 1991). The evolution of these simple bar forms into more complex forms is described by Smith (1974), who made observations of the Kicking Horse River, British Columbia, Canada.

Compound bars have a more complex history, having been shaped by many episodes of erosion and deposition. When erosion occurs, the basic shape of the bar is trimmed and dissected. Church and Jones (1982) recognise four main types of unit bars. These are illustrated in Figure 8.7. Longitudinal bars are elongated in the direction of flow. They form in the centre of the channel, typically where the channel is relatively wide. Bar growth is brought about by the accumulation of finer material, both in an upwards and in a downstream direction (Church and Jones, 1982). Longitudinal bars tend to taper off in a downstream direction (Robert, 2003). Transverse bars are lobe shaped (lobate) with relatively steep downstream faces. They are commonly found where there is an abrupt channel expansion, and downstream from confluences (Church and Jones, 1982). Transverse unit bars are not usually attached to the banks (Robert, 2003). The channel junction bars shown in Figure 8.7 are transverse bars that are associated with the flow separation that occurs at channel confluences. Point bars are a feature of most meandering channels and form on the inside of meander bends as a result of the secondary flow patterns that are associated with flow in curved channels. Point bars are elongated in the direction of flow, with a steep outer face. Diagonal bars are common in gravel-bed channels (Robert, 2003). These are bank-attached features that run obliquely across the channel. Diagonal bars may have a steep downstream front. Both longitudinal and transverse bars are closely related to mid-channel bars. The compound mid-channel bars that characterise braided channels often have a complex history (see Colour Plates 6 and 7 for examples of these compound bars). Two terms that are commonly used to describe complex bar forms are medial (or lingoid) bars and lateral bars (Robert, 2003). Medial bars are symmetrical, detached from the banks and have a characteristic lobate shape. Lateral bars are attached to one bank and have an asymmetric shape. Both types of compound bars have complex evolutionary histories. Boulder bars form in channels that are dominated by coarse bedload. As you will see later in this chapter, different morphologies are associated with the islands that are associated with anabranching channels. These include sand ridges, excavated islands, bedrock bars and vegetated bars with a bedrock core.

Benches

Benches are flat-topped, elongated, depositional features that form along one or both banks of channels. They are typically found on the inside of bends and along straight reaches, and are intermediate in height between the level of the channel bed and floodplain (Figure 8.8a). In bedrock and boulder-bed channels a boulder berm (bench composed of boulders) may form at the edge of the channel Benches can also form where flow separation occurs at the outer (concave) bank of tightly curving meander bends. This results in deposition and is illustrated in Figure 8.8(b). Erskine and Livingstone (1999) have observed sequences of adjacent benches along a bedrock-confined channel in the Hunter Valley, New South Wales, Australia. Rivers in this region have a very high flow variability, and each bench is associated with a different flow frequency. These benches are often eroded by catastrophic floods but are subsequently reconstructed by lower magnitude floods.

Riffle–pool sequences

The terms riffle and pool come from trout angling and refer to large-scale undulations in the bed topography. They are commonly found in gravel-bed channels with low to moderate channel slopes but do not tend to form in sand- or silt-bed channels (Knighton, 1998). The difference between riffles and pools is most obvious at low stages, when the flow moves rapidly over coarse sediment in the relatively steep riffle sections and more slowly through the deeper pools (Plate 8.5). The spacing from pool to pool, or riffle to riffle, is related to the width of the channel (and hence flow discharge).

In most cases this is between five and seven times the channel width (Keller and Melhorn, 1978). A longitudinal section through a riffle–pool sequence is shown in Figure 8.8(c). This illustrates the differences in bed slope, bed material size and the slope of the water surface at high and low flows. At higher flows, the differences between riffles and pools are less obvious, with less variation in the water surface slope. Riffle–pool sequences are found in straight, meandering and braided reaches. Analogous features are sometimes seen in ephemeral channels as regularly spaced accumulations of relatively coarse sediment, although there is little variation in the bed topography (Leopold et al., 1966). In ecological terms, both riffles and pools provide important habitats. For example, certain species of fish lay their eggs in the spaces between the coarse gravels in riffles, while pools provide shelter and a suitable habitat for rearing young. Various theories have been put forward to explain how riffle–pool sequences are maintained. Keller (1972) introduced a theory of velocity reversal. This suggests that the flow velocity increases at a faster rate in pool sections than in riffles as the discharge approaches bankfull. The higher shear stresses that develop in the pools lead to scouring of coarse material, which is deposited immediately downstream to form riffles. However, there is conflicting evidence to support this theory Several researchers have shown that pools have a larger cross-sectional area of flow than riffles during most flow conditions. In order to ensure continuity of flow, pools should therefore have lower cross-sectional velocities (see Chapter 6, pp. 76–77). For example, Carling (1991) made observations on the River Severn, England. These indicated that neither the cross-sectional average velocity nor the near-bed shear velocity were noticeably greater in pools than riffles during overbank/nearoverbank conditions. Instead, there was a tendency for average hydraulic variables in riffles and pools to become more similar as the discharge increased. Other theories have also been put forward. For example, field and laboratory measurements have shown that riffle surfaces tend to experience more turbulent flows. As a result, a tightly packed and interlocked bed surface develops at riffles. This is brought about by the vibration of particles and occasional particle transport during relatively low flows. In contrast, pools experience less near-bed turbulence during low flows and do not develop the same type of resistant bed structure (Robert, 2003). This means that critical bed shear stresses for sediment entrainment are higher in riffles than in pools. The riffles therefore tend to be maintained as topographic high points, while scouring occurs at pools (Robert, 2003).

Steps and pools

Steps and pools (Figure 8.8d and Plate 8.6) often characterise steep, upland channels and have been observed 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 has been observed to form part of the structure of steps. 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 resistance. There is a considerable dissipation of energy as flow cascades over each step and enters the relatively still pools (Bathurst, 1993). The spacing of steps and pools has been widely reported as being, on average, two to three times the channel width. Pools also tend to become more closely spaced as the slope increases. The height of steps appears to increase with the size of the bedload (Chin, 1999). Channels in which step–pool sequences form typically have a wide range of sediment sizes, from fine gravel to large boulders. Laboratory-based simulations indicate that step–pool sequences probably form during large floods, which mobilise the coarsest sediment. One theory suggests that, when the coarsest ‘keystones’ come to rest, they act as a barrier, leading to the accumulation of finer sediment. Downstream from this, the flow of water over the step scours a pool (Knighton, 1998).

Rapids and cascades

Like step–pool sequences, these are associated with steep channel gradients. Rapids are characterised by transverse, rib-like arrangements of coarse particles that stretch across the channel, while cascades have a more disorganised, 'random' structure. Rapids and cascades are stable during most flows because only the highest flows are competent to move the coarser cobbles and boulders that form the main structure.

Potholes

These deep, circular scour features are formed in bedrock channel reaches by abrasion (Plate 7.1). Processes of formation are described in Chapter 7 (p. 96).

Bedrock bars

In incised bedrock channels, the flow sometimes moves around bedrock bars (Figure 8.8e). These form when multiple sub-channels are incised into the bedrock substrate, leaving 'islands' or bedrock bars between them. Bedrock bars may form the core of a bedrock-alluvial bar, which becomes covered by a layer of sediment on which vegetation becomes established.

HYDRAULIC GEOMETRY RELATIONSHIPS

Discharge varies over space, usually increasing in a downstream direction, and over time in response to inputs of precipitation. Hydraulic geometry describes the way in which flow velocity, width and depth change in response to these discharge variations. Downstream hydraulic geometry refers to the changes in width, depth and flow velocity that accompany downstream increases in discharge. An example is provided in Figure 1 for three locations along a channel (top left of figure). Section A is located in the headwaters, B in the middle reaches and C further downstream. The channel cross-section at each location is shown on the right-hand side of the figure. Looking at these, the most obvious thing is the downstream increase in channel size. Notice also how the width of the channel increases faster than the depth. These downstream changes are represented by the three graphs on the right-hand side of the figure, where width, depth and velocity are each plotted against mean annual discharge. The slope of each line indicates the rate at which that variable changes downstream. As mentioned before, width increases faster than depth. The slope of the velocity graph is very gentle because downstream increases in velocity are relatively small. In some cases, velocity shows little downstream change, or even decreases slightly. The relationships shown by the graphs can be described mathematically by the equations: w = aQh d = cQf v = kQm where w, d, v and Q are width, depth, velocity and discharge and a, c, k, b, f and m are numerical constants. These equations are examples of a type of relationship called a power function, with the values for the numerical constants describing the relationship for each. The relationships are slightly different for each river, which will have its own unique values for a and b, defining the downstream relationship between width and discharge; c and f, for depth and k and m for velocity. The exponents (powers) b, f and m define the slope of each graph and, since discharge is equal to the product of width, depth and velocity, from basic algebra it follows that b + f + m = 1. Power functions appear as a straight line when plotted using a logarithmic scale. Average values for b, f and m are generally reported to be 0.5, 0.4 and 0.1 respectively. From casual observations of your local river you have probably noticed how the depth, width and velocity of flow all increase as the river rises. At-astation hydraulic geometry relationships describe variations at one particular cross-section as discharge changes through time. This is illustrated in Figure 1 for section B, with B1, B2 and B3 showing low, moderate and mean annual flood discharges. The corresponding schematic graphs are on the left hand side of the figure, using average values for the exponents (b = 0.26, f = 0.40 and m = 0.34). There is actually a fair amount of variation between different cross-sections. In wide, shallow channels the width tends to increase faster than the depth (b is greater than f), while in deeper, narrower channels it is the depth that increases at a faster rate (b is less than f). Changes in velocity are greatly influenced by elements of flow resistance at a given cross-section. In general where flow resistance is high, as in a braided channel where bars increase roughness, velocity tends to increase at a slower rate (lower value of m) than in channels with less flow resistance. Downstream hydraulic geometry relationships may be applied to identify bankfull stage and channel dimensions in ungauged basins. This assists in the design of stable natural channels for river management (Rosgen, 1994). These relationships can also be useful in studies of the effects of land-use changes on channel dimensions (Gordon et al., 2004). However, it is important to realise that there are a number of limitations associated with hydraulic geometry relationship In particular, care must be exercised when applying empirically-derived relationships beyond the range of data for which they were derived. The underlying data are representative only of the channel type and environment for which the measurements are made. This means that the relationships should not be applied beyond this, for example in predicting stable channel dimensions for a different type of river environment (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-4: CONCEPT OF MOST EFFICIENT CHANNEL: WIDTH, MEAN DEPTH, MAXIMUM DEPTH, CHANNEL FORM INDEX

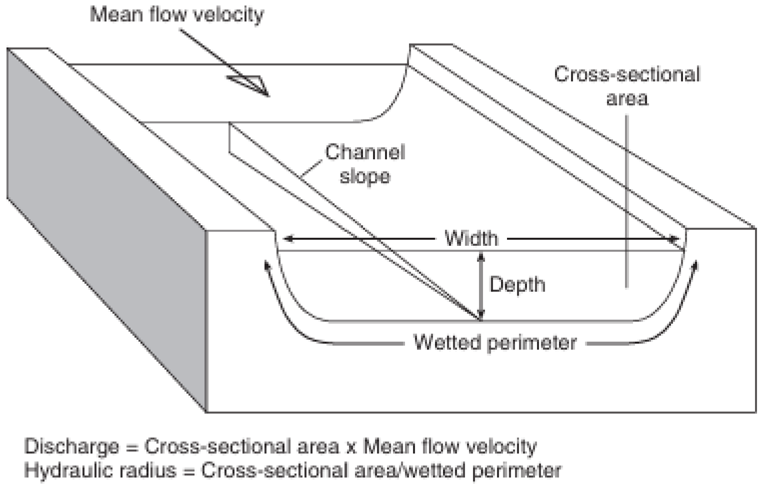
Forces driving and resisting the flow of water A force is anything that moves an object, or causes the speed or direction of a moving object to change. Forces are vector quantities, which means that they have both magnitude (size) and direction. The unit of force is the newton (N), and force magnitude is defined by the mass of the object and the acceleration produced.1 Forces are always mutual. In other words, if a force is exerted on an object, the object will react with an equal and opposite force. In most situations, several forces are involved, so the balance between driving and resisting forces is usually considered. Forces acting on an object are balanced if the object is stationary, or if it is moving at a constant velocity. The driving force causing water to flow (whether in a channel, rill, gully or overland) is the down-slope component of gravity. This acts on a given mass of water, causing it to deform (flow) and move in a downstream direction over the channel boundary (bed and banks). Opposing this movement are resisting forces. Resistance occurs because of friction between the flow and channel boundary. Also, the fluid itself resists deformation because of internal forces within the flow. As water moves down slope, it exerts a shearing force, or shear stress, on the channel boundary (shear stress is represented by the Greek letter tau, τ). The bed shear stress (t0) is expressed as a force per unit area of the bed (in N m-2) and increases with flow depth and channel steepness. This relationship is described by the du Boys equation (Box 6.1).

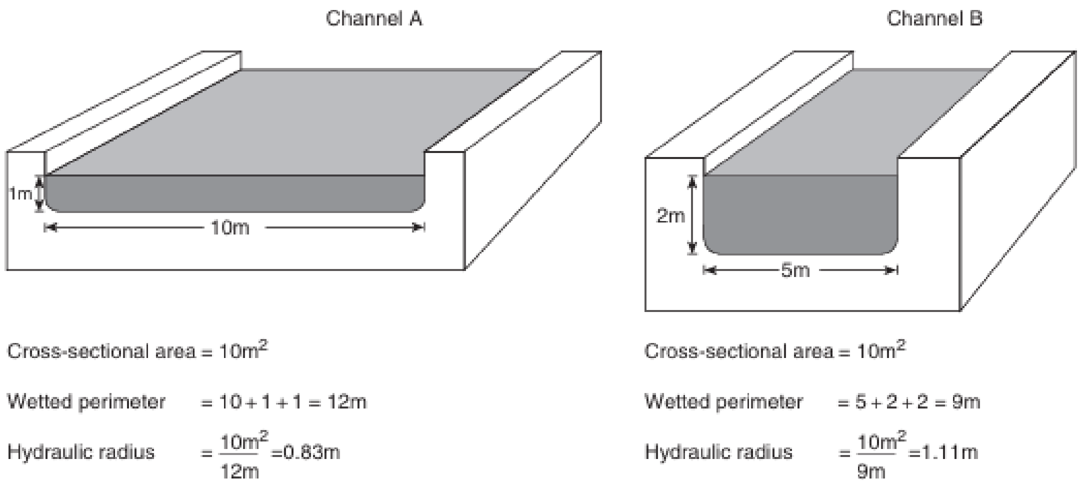
Channel parameters In order to describe the flow of water in river channels it is necessary to define some basic channel parameters, most of which are illustrated in Figure 6.1. Channel size can be defined by its cross-section: a slice taken across the channel, perpendicular to the direction of flow.

The area of the cross-section is given by the product of channel width and the mean flow depth. At a given cross-section, the cross-sectional area changes through time in response to fluctuations in discharge (defined in previous chapters). The maximum discharge that can be contained within the channel, before water starts to inundate the floodplain, is called the bankfull discharge. The width of the channel at bankfull discharge is called the bankfull width. It should be noted that there are several issues associated with the definition of bankfull discharge for many river systems (see Chapter 3, p. 32). The shape of a river channel affects its hydraulic efficiency, something that can be quantified by calculating the hydraulic radius. This is a measure of how much contact there is between the flow and channel boundary, and is calculated from:

$$\text{Hydraulic radius} = \frac{\text{Cross-sectional area (m}^2\text{)}}{\text{Wetted perimeter (m)}}$$

The wetted perimeter is the length of channel boundary that is in direct contact with the flow at a given cross-section. An example is provided in Figure 6.2, which shows two channel cross-sections. For the pur poses of this illustration, it will be assumed that the only difference between them is their shape, channel A is wide and shallow, while channel B is narrow and deep. Both have the same cross-sectional area but the wetted perimeter is larger for channel A, resulting in a lower hydraulic radius. Assuming all else is equal, the loss of energy arising from friction with the bed and banks will be greater for channel A. Channel B is therefore more hydraulically efficient. For wider channels, the hydraulic radius is very similar to the flow depth.





Channel slope is usually expressed as a gradient (difference in channel bed elevation along a given length of channel in meters divided by that length in metres). This is related to, but not necessarily the same as, the water surface slope, the downstream change in water surface elevation along the channel. Water surface slope is an important variable because it closely approximates the energy slope along a particular length of channel. As water flows through the channel, potential energy is con verted to kinetic energy. This is in turn converted to heat energy, which is generated as a result of friction,2 and 'lost' from the channel. As a result there is a downstream reduction in the total energy 'possessed' by a given parcel of water. The steepness of the energy slope reflects the rate at which energy is being expended (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-5: CHANNEL PATTERN: STRAIGHT, SINUOUS, MEANDERING, BRAIDED, ANABRANCHING, ANASTOMOSING

Most single-channel rivers and streams follow a winding path and straight channels are rare. 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 (Figure 8.11). 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. This is seen even in straight channels, and is often associated with the development of riffles, pools and alternate bars.

Braided channels

Morphology and behaviour Braided channels are very dynamic, with high rates of fluvial activity and rapid adjustments to channel form. Complex flow patterns are seen as the flow diverges and converges around the many bars. This results in large variations in bed shear stress across the channel, which influences patterns of scour and deposition. Major changes can take place at times of high flow, when rapid rates of channel migration are facilitated by high stream power and erodible banks. There can also be wholesale shifts in channel position as sub-channels are abandoned, or previously abandoned channels are reoccupied. The channel bars that typify braided channels are complex features. They are modified by processes of erosion, as well as deposition, and evolve over short periods of time. Various conditions are associated with braided channels. These are: ● An abundant bedload. This is supplied from upstream and enhanced by further contributions from bank erosion. ● Erodible banks. These allow a wide, shallow channel to form in which bars can develop. Channel subdivision continues until there is insufficient stream power to erode the banks (Leopold and Wolman, 1957). ● High stream power. Braided channels are often associated with steep slopes, although some large braided channels flow over low gradients. ● A highly variable discharge. Associated variations in flow competence lead to the sporadic movement of bedload, which is significant in bar development. However, braided channels can form under conditions of constant discharge (Leopold and Wolman, 1957; Ashmore, 1991). In the field, it is not uncommon to find channels where braided reaches alternate with meandering ones. This suggests that a variable discharge may not be of primary importance. The type of braiding and the number of sub-channels found within the main channel (braiding intensity) reflect the different environments in which braided channels are found. A number of braiding indexes have been developed, using different criteria, in an attempt to quantify the intensity of braiding. For instance, one type of index quantifies the number of active channels across the channel belt. Braiding indexes allow comparison between different channel reaches and can be used to assess channel changes over time. A distinction is often made between bars and islands, although these have the same origin and share similar morphological characteristics. While bars are only emergent at low flows, islands are more permanent, stable features, that may be vegetated and are only inundated by very high flows. The three-dimensional shape of braided channels is not obvious from studying the planform alone, as this does not show variations in depth or the important links that exist between channels and bars. Individual sub-channels are typically curved in planform where they flow around bars. Field measurements have shown that the secondary flows which develop within these curved channel sections are similar to those in meander bends. Plate 8.11 shows a bar and the downstream confluence of two curved braid channels during low flow conditions. (The channel on the right hand side of the photograph is dry.) The photo is looking upstream. What you are looking at here is similar to two meander bends, back to back. The steep faces on each side of the bar have been formed by erosion on the outside of each bend, while finer sediment has been transported to the inside of the bends and deposited on more gently sloping bars.

The initiation and development of a braided planform

Various mechanisms are involved in the development of braided channel forms, depending on the channel setting and controls. Bar formation is brought about by both depositional and erosional processes. The mechanism of central bar deposition was first described by Leopold and Wolman in 1957 from laboratory flume experiments. This process is represented schematically by the diagrams at the top left of Figure 8.15, which incorporate the later observations of Ashmore (1991). In the top diagram, localised flow convergence at the upstream end of the reach leads to scour, with the eroded material forming a thin bedload sheet that migrates along the bed of the channel. Where flow becomes locally incompetent to transport the coarsest particles, they are deposited. These coarse deposits, called lag deposits, start to alter the flow pattern. This causes more sediment to be deposited, and leads to the upward growth and downstream extension of the incipient bar. As the bar expands, the flow is forced to flow around it, attacking the banks and widening the channel, which in turn produces more bed load. In carrying out his own flume experiments, Ashmore (1991) found that a different process was dominant. He called this transverse bar conversion (see Figure 8.15). In the initial stages of this process, flow convergence leads to scour at the bed, forming a narrow, steep-sided chute. Erosion of the chute produces a large amount of sediment. Since the flow diverges as it exits the narrow chute, the flow competence rapidly declines and a lobe of sediment is deposited. This is an incipient bar. Successive bedload sheets stall (the bedload stops moving and is deposited) as they travel across the bar, building up the bar surface.

As material accumulates, a steep slip face forms where the accumulated material starts to fall over the downstream edges of the bar. As the height of the bar increases, it starts to obstruct the channel, and flow becomes diverted around an elevated central lobe. This is different from the central bar mechanism, in that the initial bar deposits are formed by the erosion and wholesale deposition of large amounts of bedload, rather than the deposition of just the coarser material that the flow is locally incompetent to transport. The central bar mechanism appears to be more typical of situations where bed shear stress is only just above the threshold for movement, meaning that small variations in depth can render the flow locally incompetent to transport the coarser sediment. When the bed shear stress is considerably greater than that required to initiate bedload movement, it is possible for the large volumes of sediment required for the second mechanism to be eroded and deposited (Ashmore, 1991). Bar development also involves erosional mechanisms, the first being the development of chute cut-offs. This involves the erosion of a channel (or chute) across a point bar, eventually separating it from the bank (Figure 8.15). Bars can also be dissected by multiple channels which form during high flow when the bar is submerged, resulting in multiple dissection (Figure 8.15). Bars are dissected during higher flow conditions, when water flows across the bar surface. Flow concentration leads to the formation of channels that cut into the bar surface. Dissected bars have the appearance, at higher stages, of being a number of smaller bars. Several examples of dissected bars can be seen in Colour Plate 6. You may also be able to identify bars that have been separated from the channel margins by chute cut-offs. Avulsion is also important in Field observations show that individual anabranches tend to have a lower width-to-depth ratio than a single channel carrying the same flow. However, narrow channels can only form in cohesive or well vegetated alluvial substrates, otherwise instability and bank collapse lead to channel widening. If the individual anabranches are unable to maintain a low width-to-depth ratio, the hydraulic advantages of anabranching are lost. 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). Wandering gravel bed channels At the high end of the anabranching energy spectrum are wandering gravel bed channels. These are intermediate between braided and meandering channels (Colour Plate 17). Wandering channels tend to develop where there are inputs of coarse sediment but where bedload transport rates are lower than for braided channels. Typical specific stream powers are within the range of 30–100 W m–2 (Church, 1983). Wandering channels are laterally active, although less so than braided channels, with fewer channels and active bars. The bars in wandering channels are more stable than those in braided channels, often being vegetated, and in most cases a dominant channel can be identified. Anabranching behaviour in wandering channels is often associated with channel avulsion, where new channels are incised within the existing floodplain. This tends to happen at higher flows, perhaps as a result of a blockage, such as an accumulation of woody debris in one of the existing channels. The supply of coarse sediment also promotes the development of bars. Anastomosing channels The term anastomosing comes from medicine and refers to the branching and rejoining of blood vessels. This is rather an appropriate description for the rivers shown in Colour Plates 8 and 9. Anastomosing channels are rare but can be found in a variety of environments. They share a number of common characteristics, which include low gradients, very low specific stream powers (less than 8 W m–2) and stable banks formed from cohesive sediment or sand stabilised by riparian vegetation. In some cases the low-energy environment is caused by tectonic subsidence or an increase in the local or regional base level. Anastomosing channels are dominated by suspended sediment, and rates of bedload transport are generally very low. This combination of low energy availability and cohesive banks limits rates of lateral activity. The dominant process of floodplain formation is the vertical accretion of fine sediment. Many examples can be found in Australia's two largest drainage basins: Lake Eyre and the MurrayDarling. Here, a combination of low gradients and an arid climate lead to a downstream decline in flow competence and a build-up of fine, cohesive sediment. Vegetation in the riparian zone is adapted to the arid climate, with deep root systems, further increasing bank stability. The mud-dominated anastomosing system of Cooper Creek in the semi-arid Channel Country of western Queensland is a well known example (Colour Plate 8). This has extensive anastomosed reaches along its length, which cover distances of over 400 km. The channel network is interesting in that there is a primary network of between one and four main channels with additional channels on the floodplain, which become active during flood flows (Knighton, 1998). Another feature of this system are numerous waterholes, that have developed along lines of preferential scour and bank erosion. The extreme flow regime for this river was discussed in Chapter 3 (see Figures 3.3b and 3.4). braid development and describes a relatively sudden switching of the flow from one channel to another (Ferguson, 1993). This occurs when chute cut-offs form, or when the flow shifts to formerly abandoned channels. Other mechanisms include the blocking of channels by aggrading bars. This leads to ponding and the formation of an overflow channel. Anabranching 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. The flow in anabranching rivers is divided into two or more separate channels. Individual anabranches are typically incised into the floodplain, the resultant islands having a similar elevation to the surrounding floodplain. The islands are stable features which are often well vegetated and can sometimes remain relatively unchanged for decades or centuries (Knighton, 1998). Flow in each anabranch is largely independent of that of its neighbours (Bridge, 1993). Although anabranching channels are fairly uncommon, they are the most diverse channel type and are found in many different environments. Examples come from climatic regions ranging from sub-arctic to tropical, and from monsoonal to semi arid (Knighton, 1998), forming in fine to coarse-grained alluvial substrates. Some anabranching channels are found in relatively high-energy environments while others have extremely low specific stream power. Despite this variety, there do appear to be some common characteristics, for example, these channels are usually characterised by flood-dominated regimes, and tend to have banks that are relatively resistant in comparison with the available stream power (Nanson and Huang, 1999). 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). This is because the combined hydraulic radius of the multiple channels is greater (more hydraulically efficient) than for a single channel carrying the same flow (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-6: 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 straight-line distance from one bend to the next (Figure 8.12a). Since the distance between successive meander bends varies, a mean wavelength is calculated for several meander bends along the reach of interest. There is a well established relationship between channel width and meander wavelength, which is usually approximately ten to fourteen times the bankfull width. (Chorley et al., 1984). 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 (Richards, 1982). Interestingly, a similar relationship is seen for other meandering systems, for example the small supraglacial streams that flow over the surface of glaciers often develop meanders, despite the absence of sediment (Plate 8.8). At a much larger scale, meanders also form in the Gulf Stream of the Atlantic. In both cases, the wavelengths of the meanders are scaled to the width of the flow in the same way as for alluvial channels.

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 (see Figure 8.12b). 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). This ratio is relatively small for tight bends and increases for bends that curve more gradually. 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 as shown in Figure 8.12(c). An asymmetric crosssection is associated with meander bends since scouring and bank steepening take place at the outside of the bend, while deposition occurs on the inside of the bend. At riffle sections, where the line of fastest flow crosses the channel, the cross-section is more symmetrical.

The regularity of meander bends varies greatly, with some following rather an irregular path, while others are highly regular. Ferguson (1979) asked whether meanders are regular or random. He concluded that meander form is a compromise between flow behaviour, which tends towards regularity, and the 'random' attributes of floodplains that lead to irregularity. These include variations in floodplain topography and sedimentology. Why do rivers meander? It might seem logical that rivers and streams should take the most direct course – a straight line – down the slope of the valley. However, most single-channel rivers show some degree of sinuosity. Over the years, a number of theories of meander formation have been developed on the basis of a growing body of theoretical, field and experimental research. However, no general agreement has been reached. The influence of channel controls is reflected in the close correlations that exist between meander wavelength, mean radius of curvature and channel width. Since channel width is related to discharge, the implication is that meanders are scaled to the range of discharges that shape the channel (Box 8.1). Meander wavelength can also be correlated with sediment load and channel slope. Lengthening of the channel reduces the slope (this was illustrated in Figure 8.6). However, bends increase flow resistance because more energy has to be used to move the flow around them. In theory, the most energy-efficient kind of bends are symmetrical meanders. These represent the path of least work, allowing the channel to lengthen, but minimise the associated energy expenditures (Langbein and Leopold, 1966). At the same time, bends allow energy expenditures to be more evenly distributed along the channel. Since flow resistance is greater at riffles than at pools, the additional energy used in turning around bends (where pools are located) balances the increased flow resistance encountered at riffles. However, reach-scale relationships such as these do not actually tell us about the processes involved. The question of how rivers meander is not fully understood but relates to interactions between the flow and the material forming the bed and banks. As water flows through a channel, spiralling secondary flow cells are set up within the flow as a result of boundary resistance (Chapter 6, p. 88). The three-dimensional nature of this flow can be significantly altered by any irregularities in the channel boundary, with the effects of any disturbance being propagated downstream. These flow variations create differences in velocity and shear-stress distributions, and hence patterns of erosion and deposition. This leads to channel-form adjustments, which in turn affect channel flow, resulting in further channel modifications. Figure 8.13 shows a conceptual model of meander formation that was developed by Keller (1972). During stage 1, alternate bars form on opposite sides of the channel as a result of alternating zones of erosion and deposition. The flow is diverted around these bars, converging as it moves towards the banks and diverging as it moves across the channel. This promotes erosion and pool formation on alternate sides of the channel. Deposition occurs at the crossing points, where riffles form (stages 2 and 3). Erosion continues to be focused at the banks, leading to the development of bends (stage 4). As these grow, and the channel extends, new riffles and pools form so that the spacing remains between five and seven times the channel width (stage 5). However, in the case of low-energy channels with resistant banks, there is no progression beyond the straight channel of stage 3. This is because the eroding force of the flow is insufficient to overcome the resistance of the banks. Lateral migration and bend development therefore cannot take place. The Keller model is not universally applicable because meanders can still form in channels without riffles and pools. While they are commonly observed in gravel bed channels, riffles and pools do not tend to form in sand and silt-bed channels. Flow and sediment transport in meander bends As water flows around a meander bend, the water tends to 'pile up' against the outer bank, resulting in a superelevation of water on that side of the channel (Figure 8.14a). As a result of this localised increase in depth, a pressure gradient develops across the channel (Chapter 6, p. 88). This leads to a compensatory flow of water across the channel bed, from the outside (high pressure) to the inside (low pressure) of the bend. Although relatively weak in comparison with the primary flow in the channel, these secondary flows are significant in moving sediment towards the inside of the bend. Point bar deposits are sorted, with coarser sediment deposited at the base of the bar, while progressively finer sediment is carried up the bar surface. The upper surface of the point bar is often draped with fine vertical accretion deposits laid down during high flows. A widely recognised feature in meander bends is the zone of high velocity that shifts from the inside to the outside of the bend with increasing distance along the bend. Dietrich et al. (1979) made detailed measurements of local boundary shear stresses at various locations along a meander bend in a sand-bed channel. These show how the zone of maximum bed shear stress also shifts from the inside to the outside of the bend. A close correspondence exists between the zone of maximum shear stress and the maximum average flow velocity (Dietrich, 1987). Measurements were also made of bedload transport and rates of bedform migration. These showed that the zones of maximum bedload transport were similar to zones of maximum bed shear stress (Dietrich et al., 1979). Meander migration Once meanders have formed, further development often takes place, as individual bends migrate, by erosion of the outer (concave) bank and compensatory deposition on the point bar at the inside of the bend (Plate 8.9). There are various ways in which meanders can migrate, some of which are illustrated in Figure 8.14(b). Lateral extension occurs during the formation of meanders and has the effect of lengthening the channel and increasing the amplitude of meander bends. The effect of valley confinement on meander development can be seen in Plate 8.10. In this case, the lateral extension of the meander bends is restricted by the valley walls. Confined meanders such as these can also form where migration is restricted by rock outcrops or artificial structures such as roads and railway embankments. Translation occurs when the fastest flow erodes the bank downstream from the bend apex, resulting in down-valley movement. Meander bends often develop an asymmetric planform when one limb of the meander migrates at a faster rate than the other, a situation that may be caused by variations in bank resistance along the channel. Meanders can develop more complex forms, for example bends with two apices (double heading) or where lobes form on existing bends (Hooke and Harvey, 1983). However, meander bends do not continue to grow indefinitely, or the channel slope would become too gentle to allow transport of the sediment load. Instead, a negative feedback mechanism comes into operation, when individual meander loops become 'short-circuited' to form a cut-off. This process shortens the channel length, with a resultant increase in the channel slope. Two main types of cut-off are observed: neck cut-offs and chute cut-offs (Figure 8.14b). Neck cut-offs are the most common (Knighton, 1998) and several examples are seen in Colour Plate 11. The cut-off at the bottom right of this photograph would previously have been a double-headed meander bend. Channel curvature is an important control on meander migration because of its influence on flow within the channel. Several researchers have found that meander migration is at a maximum for bends with an rc/ value of between 2 and 3. For example Hickin and Nanson (1975, 1984) reconstructed past rates of migration and channel curvature from former scroll bar deposits formed by the Beatton River in British Columbia, with maximum rates of meander migration occurring when the curvature was within this range (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-7: CHANNEL LONGITUDINAL PROFILE AND GRADIENT, CAUSES OF PROFILE CONCAVITY

The potential energy that is available to do work along a river (e.g. entrain/transport sediment, erode bed/banks) is a function of the balance of impelling and resisting forces at any given point in a river system. Conceptually, a range of equilibrium relationships driven by negative feedback mechanisms attempt to maintain a balance between slope, bed material size, channel size, etc. These relationships determine the consumption of energy at different positions along the longitudinal profile of a river (especially energy that is required to overcome frictional resistance). This attempt to balance impelling and resisting forces along a reach is a critical determinant of a river's behavioural regime, indicating whether there is a relative dominance of erosional or depositional tendencies at any given location along the longitudinal profile. Adjustments to the magnitude or distribution of impelling or resisting forces are primary agents of river behaviour (Chapter 11) and change (Chapter 12). The balance of impelling and resisting forces determines the potential for threshold exceedance in any given reach. Within a given set of boundary conditions, rivers may adjust to prevailing water, sediment and vegetation conditions to generate a range of river morphologies (see Chapter 10). These different types of river use their energy in efforts to balance impelling and resisting forces in different ways.

Some rivers act to maximise resistance, while others act to maximise sediment transport. Understanding these different behavioural attributes is a key step in interpreting why a river looks and behaves in the way that it does (Chapter 11). By interpreting differences in these relationships, major transitions in process–form relationships can be detected along river systems. For example, in upstream sections of catchments, resistance is imposed by forcing elements such as bedrock and so bedrock-confined rivers dominate. Adjustments to these resistance elements occur over geological timeframes. In contrast, in downstream reaches, alluvial channels are able to create and sustain their own forms of resistance elements around which adjustments occur over geomorphic timeframes. There are many enigmatic relationships in the study of river systems. One of the most puzzling scenarios to unravel is how rivers adjust to balance impelling and resisting forces at differing positions along smooth, concave-upwards longitudinal profiles. The fact that most river system 'hold together' over extensive periods of time attests to the effectiveness of this balancing act. Although rivers are forever subjected to disturbance events, whether natural or human induced, dramatic responses are the exception rather than the norm. Typically, systems internally adjust to minimise the impacts of external events. In other words, rivers adjust the balance of impelling and resisting forces in any given reach such that the system is able to use available water (flow) to transport available sediment. Of course there are exceptions to this generalisation, and this balance may be disrupted by extreme events. Indeed, some systems are especially sensitive to change, especially if they lie close to a threshold condition. In general terms, however, it is the ability of channels to alter the resistance that they provide to flow that modifies the operation of the Lane balance in any given reach.

Adjustments to bed material size and the arrangement of materials on the bed, channel geometry (size and shape) and channel planform (number of channels and their sinuosity) use flow energy in differing ways, allowing the channel to minimise the extent of response to disturbance events. For example, channel contraction results in greater dissipation of energy in overbank flows that occur more frequently because the channel is smaller (and vice versa). Alternatively, increases in channel sinuosity decrease the slope of the channel, thereby reducing flow energy (and vice versa). A similar increase in roughness, and consumption of energy, arises from an increase in bed material size or an increase in channel multiplicity (and associated surface area, or boundary resistance). These various relationships adjust in different ways at differing positions along longitudinal profiles. Slope and available energy are high in steepland settings of the source area headwater compartment of river systems with smooth, concave-upward longitudinal profiles. However, catchment areas are small, minimising available discharge. Typically, infrequent high-intensity storms are re quired to trigger formative flow events. Given steep slopes, these events have significant capacity to perform geomorphic work. However, roughness elements on the valley floor consume large amounts of energy, minimising the geomorphic effectiveness of these flows. There are two primary components to the high inherent roughness in these areas. First, these bedrock-controlled rivers have irregular channel boundaries (i.e. channel geometry), irregular channel alignment (i.e. planform), steep slopes with a large number of steps, cascades and waterfalls, and many forced roughness elements induced by features such as trapped wood (log jams). Second, most fine-grained sediments are flushed through these high-energy settings, leaving behind the coarse fraction of the bed material load. This is usually comprised of boulders, cobbles and gravels. These materials are only mobilised by the highest (extreme) flows. Indeed, they are often lag deposits from the last formative flow. Critically, these materials generate considerable flow resistance to all events other than the deepest flow. As a consequence, the valley floor is relatively stable for the vast majority of the time. Infrequent high-magnitude events may bring about localised disturbance, but to all intents and purposes the channel adjusts very little. Incrementally, when critical shear stress is exceeded and the coarsest fraction is mobilised, the channel is able to cut into its bed (i.e. these are degradational systems). Transfer reaches along rivers with smooth, concaveupwards longitudinal profiles are typically found downstream of source zones along sections with lower slopes and wider valley settings. These are transition zones in river systems. The two primary components of impelling forces, slope and discharge are adjusting along differing trajectories. The rate of decrease in elevation decreases with distance from the headwaters, reducing slope with distance along the longitudinal profile. In contrast, discharge tends to increase as catchment area increases. In general terms, a peak in total stream power conditions occurs around the beginning of the transfer zone, typically for third- to fifth-order streams (see Chapter 3). However, total stream power decreases downstream, as the rate of decrease in slope is greater than the rate of increase in discharge. Just as importantly, however, the increase in valley width with distance downstream alters the way in which the channel uses its available energy. Rather than simply concentrating energy in efforts to incise into its bedrock bed, flow energy is dissipated in wider sections of valley in which floodplain pockets may form (i.e. the river is able to deposit and rework finer grained sediments in these reaches). As a consequence, the river uses its energy in a different way to upstream reaches, achieving a different configuration in which erosional and depositional processes are approximately in balance. Finally, the accumulation zone of catchments has lower slope and higher discharge conditions than upstream reaches. In these areas the channel is most able to consume its own energy through self-adjustment processes, as the channel flows within its own sediments. Hence, adjustments to channel planform and geometry are able to accommodate alterations to flow and sediment conditions more readily than elsewhere in the catchment. Impelling forces are lower than upstream, as lower slopes exert a greater influence upon total stream power than the impact of higher discharge. Typically, flow is smoother in the larger channels of accumulation zones that flow within finer grained (less rough) sediments (except where large bedforms are produced), so resisting forces are also reduced. Bed materials are reworked more frequently in these areas relative to upstream. In these parts of catchments, the channel is less able to transport all sediments made available to it, so aggradation occurs on valley floors and sediments prograde into receiving basins. In summary, flow energy is used in differing ways in source, transfer and accumulation zones, but the quest for balance is sustained throughout. Understanding of these relationships at any given site, appreciating the ways in which differing forces are applied in any given system and unravelling the ways in which one reach is connected to another are key skills in efforts to read the landscape. Geomorphologists analyse how the balance works (or does not work) along any given section of river and how the causes of that adjustment may induce on- or off-site impacts, whether locally or in upstream or downstream parts of the longitudinal profile (Fryirs & Brierley, 2013).

References:

Fryirs, K. A., & Brierley, G. J. (2013). *Geomorphic Analysis of River Systems: An Approach to Reading the Landscape.* Wiley-Backwell.

UNIT-8: FORMATION OF CHANNEL BARS, ALLUVIAL FANS, FLOODPLAINS, DELTAS, AND ESTUARIES

Bars are in-channel accumulations of sediment which may be formed from boulders, gravel, sand or silt. Bars can be divided into two broad groups: unit bars and compound bars (Smith, 1974). Unit bars are relatively simple bar forms whose morphology is mainly determined by processes of deposition (Ashmore, 1991). The evolution of these simple bar forms into more complex forms is described by Smith (1974), who made observations of the Kicking Horse River, British Columbia, Canada Compound bars have a more complex history, having been shaped by many episodes of erosion and deposition. When erosion occurs, the basic shape of the bar is trimmed and dissected. Church and Jones (1982) recognise four main types of unit bars. These are illustrated in Figure 8.7. Longitudinal bars are elongated in the direction of flow. They form in the centre of the channel, typically where the channel is relatively wide. Bar growth is brought about by the accumulation of finer material, both in an upwards and in a downstream direction (Church and Jones, 1982). Longitudinal bars tend to taper off in a downstream direction (Robert, 2003). Transverse bars are lobe shaped (lobate) with relatively steep downstream faces. They are commonly found where there is an abrupt channel expansion, and downstream from confluences (Church and Jones, 1982). Transverse unit bars are not usually attached to the banks (Robert, 2003). The channel junction bars shown in Figure 8.7 are transverse bars that are associated with the flow separation that occurs at channel confluences. Point bars are a feature of most meandering channels and form on the inside of meander bends as a result of the secondary flow patterns that are associated with flow in curved channels. Point bars are elongated in the direction of flow, with a steep outer face. Diagonal bars are common in gravel-bed channels (Robert, 2003). These are bank-attached features that run obliquely across the channel. Diagonal bars may have a steep downstream front. Both longitudinal and transverse bars are closely related to mid-channel bars. The compound mid-channel bars that characterise braided channels often have a complex history (see Colour Plates 6 and 7 for examples of these compound bars). Two terms that are commonly used to describe complex bar forms are medial (or lingoid) bars and lateral bars (Robert, 2003). Medial bars are symmetrical, detached from the banks and have a characteristic lobate shape. Lateral bars are attached to one bank and have an asymmetric shape. Both types of compound bars have complex evolutionary histories. Boulder bars form in channels that are dominated by coarse bedload. As you will see later in this chapter, different morphologies are associated with the islands that are associated with anabranching channels. These include sand ridges, excavated islands, bedrock bars and vegetated bars with a bedrock core.

Benches

Benches are flat-topped, elongated, depositional features that form along one or both banks of channels. They are typically found on the inside of bends and along straight reaches, and are intermediate in height between the level of the channel bed and floodplain

(Figure 8.8a). In bedrock and boulder-bed channels a boulder berm (bench composed of boulders) may form at the edge of the channel Benches can also form where flow separation occurs at the outer (concave) bank of tightly curving meander bends. This results in deposition and is illustrated in Figure 8.8(b). Erskine and Livingstone (1999) have observed sequences of adjacent benches along a bedrock-confined channel in the Hunter Valley, New South Wales, Australia. Rivers in this region have a very high flow variability, and each bench is associated with a different flow frequency. These benches are often eroded by catastrophic floods but are subsequently reconstructed by lower magnitude floods.

Riffle–pool sequences

The terms riffle and pool come from trout angling and refer to large-scale undulations in the bed topography. They are commonly found in gravel-bed channels with low to moderate channel slopes but do not tend to form in sand- or silt-bed channels (Knighton, 1998). The difference between riffles and pools is most obvious at low stages, when the flow moves rapidly over coarse sediment in the relatively steep riffle sections and more slowly through the deeper pools (Plate 8.5). The spacing from pool to pool, or riffle to riffle, is related to the width of the channel (and hence flow discharge).

In most cases this is between five and seven times the channel width (Keller and Melhorn, 1978). A longitudinal section through a riffle–pool sequence is shown in Figure 8.8(c). This illustrates the differences in bed slope, bed material size and the slope of the water surface at high and low flows. At higher flows, the differences between riffles and pools are less obvious, with less variation in the water surface slope. Riffle–pool sequences are found in straight, meandering and braided reaches. Analogous features are sometimes seen in ephemeral channels as regularly spaced accumulations of relatively coarse sediment, although there is little variation in the bed topography (Leopold et al., 1966). In ecological terms, both riffles and pools provide important habitats. For example, certain species of fish lay their eggs in the spaces between the coarse gravels in riffles, while pools provide shelter and a suitable habitat for rearing young. Various theories have been put forward to explain how riffle–pool sequences are maintained. Keller (1972) introduced a theory of velocity reversal. This suggests that the flow velocity increases at a faster rate in pool sections than in riffles as the discharge approaches bankfull. The higher shear stresses that develop in the pools lead to scouring of coarse material, which is deposited immediately downstream to form riffles. However, there is conflicting evidence to support this theory.

Several researchers have shown that pools have a larger cross-sectional area of flow than riffles during most flow conditions. In order to ensure continuity of flow, pools should therefore have lower cross-sectional velocities (see Chapter 6, pp. 76–77). For example, Carling (1991) made observations on the River Severn, England. These indicated that neither the cross-sectional average velocity nor the near-bed shear velocity were noticeably greater in pools than riffles during overbank/nearoverbank conditions. Instead, there was a tendency for average hydraulic variables in riffles and pools to become more similar as the discharge increased. Other theories have also been put forward. For example, field and laboratory measurements have shown that riffle surfaces tend to experience more turbulent flows. As a result, a tightly packed and interlocked bed surface develops at riffles. This is brought about by the vibration of particles and occasional particle transport during relatively low flows. In contrast, pools experience less near-bed turbulence during low flows and do not develop the same type of resistant bed structure (Robert, 2003). This means that critical bed shear stresses for sediment entrainment are higher in riffles than in pools. The riffles therefore tend to be maintained as topographic high points, while scouring occurs at pools (Robert, 2003).

Steps and pools

Steps and pools (Figure 8.8d and Plate 8.6) often characterise steep, upland channels and have been observed 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 has been observed to form part of the structure of steps. 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 resistance. There is a considerable dissipation of energy as flow cascades over each step and enters the relatively still pools (Bathurst, 1993). The spacing of steps and pools has been widely reported as being, on average, two to three times the channel width. Pools also tend to become more closely spaced as the slope increases. The height of steps appears to increase with the size of the bedload (Chin, 1999). Channels in which step–pool sequences form typically have a wide range of sediment sizes, from fine gravel to large boulders. Laboratory-based simulations indicate that step–pool sequences probably form during large floods, which mobilise the coarsest sediment. One theory suggests that, when the coarsest 'keystones' come to rest, they act as a barrier, leading to the accumulation of finer sediment. Downstream from this, the flow of water over the step scours a pool (Knighton, 1998).

Rapids and cascades

Like step–pool sequences, these are associated with steep channel gradients. Rapids are characterised by transverse, rib-like arrangements of coarse particles that stretch across the channel, while cascades have a more disorganised, 'random' structure. Rapids and cascades are stable during most flows because only the highest flows are competent to move the coarser cobbles and boulders that form the main structure.

Potholes

These deep, circular scour features are formed in bedrock channel reaches by abrasion (Plate 7.1). Processes of formation are described in Chapter 7 (p. 96). Bedrock bars In incised bedrock channels, the flow sometimes moves around bedrock bars (Figure 8.8e). These form when multiple sub-channels are incised into the bedrock substrate, leaving 'islands' or bedrock bars between them. Bedrock bars may form the core of a bedrock-alluvial bar, which becomes covered by a layer of sediment on which vegetation becomes established.

FLOODPLAIN MORPHOLOGY

Processes of floodplain formation The morphology of floodplains is intimately linked with the form and behaviour of the river channels that shape them. Various processes of deposition, reworking and erosion are involved in the formation and development of floodplains. Sediment accumulates on floodplain surfaces by various processes of accretion, the main ones being vertical, lateral and braid bar accretion (Nanson and Croke, 1992). Lateral accretion deposits are laid down by migrating rivers, which erode into the floodplain and lay down sediment in their wake. The accretion of point bar deposits can sometimes be seen as a series of concentric ridges on the inside of bends called meander scrolls. Braid bar accretion occurs when bars are abandoned and gradually become incorporated into the floodplain deposits. There are various ways in which this can happen, for example when a large flood lays down extensive bar deposits. Alternatively, bars may become abandoned when the main braid channels shift to another part of the valley. Vertical accretion deposits are composed of fine material that settles out of suspension when overbank flows inundate the floodplain. The increased area of contact, coupled with the roughness of the floodplain surface, greatly reduces flow velocities, and a thin layer of sediment is draped across the floodplain. This displays a fining-upwards sequence, where the coarser particles, which settle out first, are overlain by progressively finer material. There is also a fining of sediment away from the channel, since only the very smallest particles are carried to the edge of the inundated area. Over a number of years the cumulative effect of overbank flows leads to the development of a vertical sequence of thin layers. Other, more localised, types of accretion can also be identified. For example, counterpoint accretion is associated with the deposition of concave bank benches at confined meander bends (see section on channel geomorphic units above). As an over-tightened meander bend migrates, bench deposits become incorporated into the floodplain.

Erosional processes include floodplain stripping, where entire sections of the floodplain surface are removed by high-magnitude flood events. Floodplain stripping is most likely to occur in relatively confined valley settings, where floodplain flows are concentrated between the valley walls. Other erosional processes include the formation of flood channels, which carry water during overbank flows. Avulsion involves a shift in the position of a channel and is a common process in braided reaches where the flow frequently abandons and reoccupies sub-channels. Avulsion can also involve the diversion of flow into a newly eroded channel cut into the floodplain. This type of avulsion is important in the development of anabranching channels. The morphology and development of floodplains is controlled by the driving variables and boundary conditions. An important balance exists between the shear stress exerted by the flow and the resistance of the floodplain to erosion. Shear stress is closely related to specific stream power (Chapter 7, pp. 93–94), and therefore to such controls as flow regime, valley slope and valley confinement. On the other side of the balance, resistance to erosion is largely determined by the cohesiveness of the floodplain sediments. An energy-based floodplain classification was proposed by Nanson and Croke (1992). This recognises three main classes of floodplain: ● High-energy non-cohesive floodplains are typically found in steep upland areas where the specific stream power in the channel at bankfull flow exceeds 300 W m–2. An example is shown in Plate 8.7. Lateral migration is often prevented by the coarseness of the floodplain sediment, which builds up vertically over time. These floodplains are disequilibrium features that are partly or completely eroded by infrequent extreme events. ● Medium-energy non-cohesive floodplains are formed from deposits ranging from gravels to fine sands. Specific stream power ranges from 10 to 300 W m–2. The main processes of floodplain construction are lateral point bar accretion (meandering channels) and braid bar accretion (braided channels). These floodplains are typically in dynamic equilibrium with the annual to decadal flow regime. Some of the features associated with medium-energy non-cohesive floodplains are illustrated in Figure 8.9(a). ● Low-energy cohesive floodplains are usually associated with laterally stable single thread or anastomosing channels. Formed from silt and clay, the dominant processes are vertical accretion of finegrained sediment and infrequent channel avulsions. Specific stream power at bankfull stage is generally less than 10 W m-2. Features associated with low energy cohesive floodplains are shown in Figure 8.9(b). These classes can be further subdivided, mainly on the basis of floodplain forming processes. They can also be related to the downstream reductions in stream power and sediment size discussed on pp. 124–128. Floodplain geomorphic units Levees Levees are elongated, raised ridges that form at the channel–floodplain boundary during overbank flow events (Figure 8.9b). 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 (see Chapter 6, pp. 88–90 and Figure 6.10). This results in the preferential deposition of material at the edges of the channel. Colour Plate 9 shows an anastomosing river in flood. 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 (discussed in Chapter 10). Crevasse splays Levees can be breached by floodwaters. This may lead to the formation of a crevasse splay, a fan-shaped lobe of sediment deposited when sediment-charged water escapes and flows down the levee (Figure 8.9a). If flow is sufficiently concentrated, a new channel may be cut and deepened by scour. Backswamps 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 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.

Cut-offs

These are abandoned meander bends that have been short-circuited by the flow. Several examples can be seen in Colour Plate 11. 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 (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

UNIT-9: CHANNEL DECAY: CAUSES AND CONSEQUENCES

River evolution is the study of river adjustment over time. Evolution is ongoing. Even if boundary conditions remain relatively constant, adjustments occur. Appraisal of the trajectory and rate of river evolution is required to assess whether ongoing adjustments are indicative of long-term trends or whether they mark a deviation in the evolution ary pathway of that river. Such insights guide interpretation of the likelihood that the direction, magnitude and rate of change will be sustained into the future. To perform these analyses, it is important to determine how components of a river system adjust and change over differing timeframes, and assess what the consequences of those changes are likely to be. Reconstructions of the past provide a means to forecast likely future river behaviour. Instinctively, human attention is drawn to landscapes that are subject to change. Observations of bank erosion, river responses to flood events, anecdotal records of river adjustments or analyses of historical maps and aerial pho tographs provide compelling evidence of the nature and rate of river adjustments. Efforts to read the landscape must frame these insights in a broader context, examine their representativeness and isolate controls upon evolutionary trajectories. For example, do these adjustments reflect modifications around a characteristic state and associated equilibrium scenarios over a given timeframe? Are short-term adjustments indicative of longer term trends? Has the river been subjected to threshold-induced change? How has the balance of formative and reworking processes and controls changed over time? Is the river sensitive or resilient to disturbance? How are responses to disturbance manifest through the catchment, remembering that an erosional signal in one place is often matched by a depositional signal elsewhere? Analysis of river behaviour in Chapter 11 highlighted how different types of rivers have differing capacities to adjust, such that they respond to differing forms of disturbance event in different ways. Attributes such as thalweg shift on braidplains, meander migration/translation, cut off development or avulsion are characteristic behavioural traits for certain types of rivers. In some instances, altera tions to the boundary conditions under which rivers operate may bring about river change, whereby the behavioural regime of the river is transformed, and the river is now characterised by a different set of process–form relationships. River evolution may occur in response to pro gressive adjustments, an instantaneous event (e.g. a major f lood or an earthquake) or longer term changes to geologic and climatic boundary conditions. This distinction between behaviour and change is essentially a matter of timescale. All rivers change as they evolve over time. In essence, if the geomorphic structure of a river changes, so does everything else (i.e. process relationships and the balance of impelling and resisting forces at the reach scale (Chapter 5) encom pass adjustments to bed material organisation (Chapter 6), assemblages of instream and floodplain geomorphic units (Chapters 8 and 9) and channel geometry (Chapter 7)). River change can result from alterations to impelling forces, resisting forces, or both. Resulting adjustments modify the nature, intensity and distribution of erosional and depositional processes along a reach. In some instances, predictable transitions can occur. For example, a change from a wandering gravel-bed river to an active meandering river can occur as flux boundary conditions are altered to reduce sediment load and discharge, or vegetation cover is increased. However, just because a particular type of river in a given system responds to an event of a given magnitude in a certain way, does not mean that an equivalent type of river in an adjacent catchment will respond to a similar event in a consistent manner. Even if particular cause and effect relationships are well understood, some systems may demonstrate complex (or chaotic) responses to disturbance events (see Chapter 2). More importantly, no two systems are subjected to the same set of disturbance events. Each system has its own history and its own geography (configuration), with its own cumulative set of responses to disturbance events, and associated lagged and off-siteresponses. The trajectory of river change may be influenced by the co-occurrence of disturbance events, such as a large f lood following vegetation clearance. Such concatenations may set the system on a trajectory of change that would not have occurred if the system had not been disturbed or if these disturbances had occurred independently. Also, similar outcomes may arise from different processes and causes (the principle of convergence or equifinality; see Chapter 2). Geologic and climatic factors determine the environ mental setting and the nature of disturbance events to which rivers are subjected. They set the imposed and flux boundary conditions that fashion the erodibility and erosivity of a landscape, and the resulting character, behaviour and pattern of river types. Stark contrasts can be drawn, for example, between a dry, low-relief landscape with negligible vegetation cover and a high-precipitation mountainous terrain with dense forest cover. Formative processes, rates of activity (magnitude–frequency relations) and evo lutionary trajectories vary markedly in these differing set tings. Hence, any consideration of river evolution must be framed in relation to these geologic and climatic controls. In this chapter these considerations are appraised for differing tectonic settings and morphoclimatic regions. Particular emphasis is placed upon how landscape setting influences the imposed boundary conditions (especially slope and valley width) that constrain the range of behaviour of rivers, and the flux boundary conditions (i.e. flow and sediment regimes) that determine the mix of erosional and depositional processes along any given reach. Critically, as noted from the Lane balance diagram, alteration to either the imposed or flux boundary conditions promotes evolutionary adjustments. Geologic factors set and alter the imposed boundary conditions under which rivers operate, through their influence on lithology, relief, slope, valley morphology and erosivity and/or erodibility of a land scape. For example, tectonic activity or volcanic events may disrupt the nature and configuration of a landscape. Climate considerations play two critical roles. First, they are key determinants of the type and effectiveness of geomorphic processes (flow and sediment interactions) that shape landscapes at any given place. Second, climatic factors mediate the role of ground cover, which affects hydrologic processes and landscape responses to geomorphic proc esses through its influence upon surface roughness and resistance. Alterations to flux boundary conditions drive adjustments to the flow–sediment balance, prospectively modifying the evolutionary trajectory of a system. Evolutionary adjustment may take a mere moment in time (e.g. river responses to a volcanic eruption) or be lagged some time after a disturbance event. Elsewhere, landscapes may be stable or demonstrate progressive adjustment over time. Some rivers are adjusted to high coefficients of discharge variability (see Chapter 4), such that large floods are rare but not unusual – they are part of the 'formative process regime' for that particular setting. Other rivers are adjusted to smaller, more recurrent events. Many rivers flow on surfaces created by past events, or are still adjusting to past flow and sediment regimes. In these cases, geomorphic memory continues to exert a significant influence upon contemporary forms and the nature and effectiveness of processes. Understanding how contemporary processes relate to historical influences is a key challenge in efforts to read the landscape. This chapter is structured as follows. First, timescales of river change are discussed. Second, pathways and rates of geomorphic evolution are summarised for different types of rivers. Third, geologic and climatic controls on river evolution are considered. Then, evolutionary responses to changes in boundary conditions are outlined, and the river evolution diagram presented in Chapter 11 is used to extend analysis of river behaviour to incorporate inter pretations of the nature and capacity for river change for various types for rivers. Finally, tools to interpret river evo lution by reading the landscape are reviewed. Timescales of river adjustment Timescale of river adjustment varies from place to place, dependent upon the range of adjustment of the system (its sensitivity/resilience), the range and sequence of disturbance events and the legacy of past impacts. Both sensitive and resilient systems are prone to disturbance – responses are more likely and/or recurrent in the former relative to the latter. Analysis of river evolution frames system responses to disturbance events in relation to adjustments over geologic and geomorphic time (see Chapter 2). Geologic controls set the imposed boundary conditions within which rivers operate. Over timeframes of millions of years, tectonic setting exerts a primary control upon topography, deter mining slope and valley settings that influence river mor phology and behaviour. Over geomorphic time, rivers adjust to climatically fashioned flux boundary conditions (flow variability, sediment availability and vegetation cover) over hundreds or thousands of years. Any disruption to flux boundary conditions may affect the evolutionary trajectory of a river. The key consideration here is whether the reach is able to accommodate adjustments while it continues to operate as the same type of river (i.e. it operates within its behavioural regime) or whether these altered conditions bring about a transition in process–form relationships (i.e. river change occurs). As noted in Chapter 2, river responses to disturbance events range from gradualist (uniformitarian) adjustmentsthrough to catastrophic change. A continuum of responses to disturbance events may be discerned: • No response may be detected, as systems absorb the impacts of disturbance. Stable rivers can tolerate con siderable variation in controlling factors and forcing processes. For example, gorges are resilient to adjust ment or change. Alluvial systems with inherent resilience induced by the cohesive nature of valley floor deposits, or the mediating influence of riparian vegetation and wood, may demonstrate limited adjustment over thousands of years. In these cases, responses to disturbance events are short-lived or intransitive, and change does not occur. • Part of progressive change. Rivers may respond rapidly at f irst after disruption, but in a uniform direction there after, such that change occurs gradually over a long period. For example, progressive denudation results in gradual reduction of relief over time, as gravitationally induced processes transfer sediments from source to sink. This results in long-term changes to slope and, hence, river type. Progressive adjustments are often observed following ramp or pulse disturbance events, so long as threshold conditions are not breached. • Change may be instantaneous as breaching of intrinsic or extrinsic threshold conditions prompts the transition to a new state or even a new type of river. These effects tend to be long lasting or persistent. • Change may be lagged. Off-site impacts of major disturbances may induce a lagged response in downstream reaches (e.g. conveyance of a sediment slug). The sub sequent history of disturbance events affects the nature/ rate of response and prospects for recovery. Efforts to read the landscape seek to unravel variability in forms, rates and consequences of adjustments within any given system over differing timeframes. Pathways and rates of adjustment and evolution vary markedly for different types of rivers. Pathways and rates of river evolution Evolutionary pathways and rates of adjustment of rivers vary in differing geologic and climatic settings. This reflects differing ways in which boundary conditions and disturbance events affect flow–sediment interactions along a river. Alternatively, disturbance events may affect the sur faces upon which these processes are acting (e.g. role of fire, human disturbance – see Chapter 13). Evolutionary adjust ments are likely to be most marked for those systems that have the greatest capacity to adjust and change. Hence, the nature and rate of evolution tend to be most pronounced in freely adjusting alluvial settings. These rivers have the greatest range in their degrees of freedom, such that pro nounced disturbance events may trigger adjustments in channel planform, channel geometry (bed and bank proc esses), assemblages of channel and floodplain geomorphic units, and bed material organisation. The mix of water, sediment and vegetation conditions, as such, influences likely pathways of river adjustment for rivers in differing settings. Characteristic examples of evolutionary pathways are presented for rivers in differing valley settings below. Likely evolutionary pathways of rivers in confined valley settings Rivers in confined valley settings have limited capacity for adjustment. Their morphologies are largely imposed and are comprised largely of an array of imposed (bedrock) erosional forms. Steep headwater rivers progressively rework assemblages of slope-induced erosional geomorphic units as channels cut into bedrock via incisional processes over timeframes of thousands of years (Figure 12.1a). In con trast, gorges are stable and resilient systems over timeframes of hundreds or thousands of years. However, progressive incision and lateral valley expansion eventually create space along the valley floor for floodplain pockets to develop in partly confined valleys (Figure 12.1a). These transitions reflect changes to imposed boundary conditions. Likely evolutionary pathways of rivers in partly confined valley settings Just as gorges progressively widen to partly confined valleys with bedrock-controlled floodplain pockets over thou sands of years, so sustained widening of these valleys even tually promotes the transition to partly confined valleys with planform-controlled floodplain pockets (Figure 12.1a). Increased valley width and reduced valley floor slope or changes in material texture, in turn, may result in a transition in the type of planform-controlled floodplain pockets that are observed, say from a low-sinuosity variant to a meandering planform variant (Figure 12.1a). Likely evolutionary pathways of rivers in laterally unconfined valley settings Variants of channel planform were conveyed along a con tinuum in Chapter 10. Adjustments to flow–sediment relations (i.e. flux boundary conditions) may bring about a transition to adjacent types of rivers along this continuum. Reduced energy conditions induced by lower flow and/or sediment availability may transform a braided river into a wandering gravel-bed river, and vice versa (Figure 12.1b). In turn, reduced energy conditions induced by lower flowand/or sediment availability may transform a wandering gravel-bed river into an active meandering river, and vice versa (Figure 12.1b). Alternatively, increase in sediment load (bedload fraction) may transform a passive meandering (suspended-load) river into an active mean dering (mixed-load) river, and vice versa (Figure 12.1b). Various stages of evolutionary adjustments may be discerned along a discontinuous watercourse, reflecting cut and fill phases (Figure 12.1c). However, should certain circumstances eventuate, the river may maintain a continuous watercourse. The examples outlined in Figure 12.1 convey progressive evolutionary adjustments. In essence, the types of rivers that are found in an adjacent position along the longitudinal profile (i.e. an energy gradient) are likely to present the next step or phase in the evolutionary adjustment of a river. This may reflect conditions of decreasing energy associated with progressive landscape denudation, or increasing energy associated with uplift (i.e. steeper slope conditions). This line of reasoning, whereby juxtaposed river types along slope-induced environmental gradients provide guidance into likely evolutionary adjustments, is a direct parallel to Walther’s law of the correlation of facies (Chapter 6): adja cent sedimentary deposits in contemporary landscapes are used to guide inferences into stacked depositional units within basin fills. Geologic and climatic controls are the primary determinants of imposed and flux boundary conditions, and the associated suites of disturbance events to which rivers are subjected. Although these geologic and climatic considerations act in tandem, they are considered separately below for simplicity. Geologic controls upon river evolution Geologic setting determines the imposed boundary conditions within which rivers adjust and evolve. The nature and movement of tectonic plates is a primary determinant of the distribution and relief of terrestrial and oceanic sur faces. The nature and position of mountain belts and depo sitional basins is determined largely by the distribution of plates and geologic processes that occur at different types of plate boundaries. Landscape relief and topography are fashioned by the balance of endogenetic processes (i.e. geo logic processes that are internal to the Earth) and exogenetic processes (i.e. geomorphic processes that erode and deposit materials at the Earth’s surface). The nature, frequency and consequences of geologic disruption and disturbance events vary markedly in different tectonic set tings. This is determined largely by position relative to a plate margin and the nature of tectonic activity at that margin. Vertical and lateral displacement along fault-lines is common in some settings. Contorted strata of folded rocks attest to the incredible forces at play. Faulting, folding and tilting generate distinctive topographic controls upon slope, valley morphology and drainage patterns (Chapter 3). Volcanic activities and subsidence modify relief and availability of materials. Tectonic setting frames the long-term landscape and dynamic context of river systems. Simplified schematic rep resentations of primary tectonic settings are presented in Figure 12.2. Tectonic motion generates three primary types of plate boundaries: collisional (constructional) (Figure 12.2a and b), pull-apart (Figure 12.2c and d) and lateral displacement (Figure 12.2e). The rate of uplift, subsidence or lateral movement of a plate affects the relative stability and nature of landscape adjustment. Uplifting rivers incise into their beds, creating narrow valleys. Subsiding rivers aggrade, creating expansive floodplains. Rivers in low relief, plate-centre settings are characterised by long-term stability, as they slowly denude and rework the landscape (Figure 12.2f). Long-term changes to the position of plate boundaries affect the nature of constructional and reworking proc esses at any given locality (Figure 12.3). These geological foundations determine patterns of lithological and structural variability, affecting the erodibility and erosivity of landscapes. The convergence of continental plates generates major mountain chains. Deeply incised bedrock channels in headwater settings contrast starkly with transport-limited braided rivers, low-relief rivers atop uplifted plateau land scapes or deeply incised gorges at plateau margins (Figure 12.4a–c respectively). Uplift of supply-limited plateau land scapes may create deeply entrenched, superimposed drain age networks. For example, Figure 12.4d shows the plan form of a meandering river that previously formed on a relatively flat alluvial plain that has been retained as the landscape was uplifted, creating a deeply etched, bedrock controlled gorge in which river character and behaviour are imposed. Differing forms of constructional landscapes are generated through subduction of dense but relatively thin oceanic plate beneath a continental plate. Recurrent phases of tectonic activity produce basin and range topography comprised of mountain ranges, volcanic chains and inter vening basins, exerting a dominant imprint upon contemporary drainage networks. The imprint of landscape setting upon river character, behaviour and evolution is clearly evident in pull-apart basins. This tectonic setting is char acterised by striking alignment of lakes and straight, bedrock-controlled river systems. In some instances, basins that pulled apart in the past may retain a dominant imprint upon contemporary landscapes, forming escarpments (Figure 12.4e) and rift valleys (Figure 12.4f). Alternatively, lack of tectonic activity is a primary determinant of riverprocesses and forms in plate-centre landscapes. These low relief, low-erosion settings often have profound stability and antiquity (Figure 12.4g). Long-term changes to plate tectonic boundaries ensure that any given landscape setting has likely been subjected to differing forms and phases of tectonic activity (Figure 12.3). Geologic adjustments are sometimes imprinted atop each other. Elsewhere, the imprint of past events has been virtually erased, though metamorphosis of rocks may provide insights into former conditions. Importantly, tec tonic setting not only fashions the relief and erodibility of a landscape, it also affects the climate and, hence, the erosivity of that landscape. Climatic influences on river evolution Spatial and temporal variability in climate are genetically linked to geologic considerations, as mountain belts and other topographic factors influence temperature and pre cipitationregimes and the movement of weather systems. The distribution of landmasses and latitudinal factors fashion continental or maritime climate conditions and solar radiation effects. Topographic and climatic conditions can be combined to differentiate morphoclimatic regions (Chapter 4; Figure 12.5). Climatic controls upon river evolution are manifest in two primary ways. Direct influences reflect hydrologic con siderations and thermal conditions, expressed primarily by the flow regime. This drives the flux boundary conditions under which rivers operate. Indirect influences are manifest primarily through climatic influences upon ground cover (and rainfall–runoff associations) and resistance factors (i.e. surface roughness). Any alteration to these relation ships affects the flux boundary conditions under which rivers operate. Adjustments to the flow and sediment balance may alter the evolutionary trajectory of a river. In many settings, past climatic conditions continue to exert an influence upon the effectiveness of contemporary geo morphic processes (i.e. climatic memory). Direct and indirect impacts of climate variability vary markedly in differing morphoclimatic regions. Some tropical humid regions are characterised by high temperatures and high precipitation throughout the year, and have rain forest vegetation associations. Rivers in these regions are attuned to recurrent high flow conditions and considerable roughness on valley floors, but interannual variability in f low is limited (i.e. the coefficient of variation for discharge is low; Chapter 4). Tropical humid areas with prominent dry and monsoonal seasons are characterised by savanna vegetation. Although seasonal variability in flow and geo morphic activity is pronounced, interannual variability is limited. Rivers in these areas are especially sensitive to the effectiveness of the monsoon. Mid-latitude regions are dominated by arid and semi-arid climates. Desert and steppe landscapes have limited vegetation cover. Pronounced, highly effective geomorphic activity occurs during short storms. Desert environments with limited sediment availability are characterised by etched/sculpted bedrock rivers. Other deserts have ephemeral rivers with high sand availability, resulting in high width/depth channels because of the non-cohesive, non-vegetated nature of bank materials. Humid-temperate rivers have perennial flow. Vegetation cover exerts a primary influence upon process–form relationships. Warmer humid regions are not subjected to severe winter conditions, but summers can be hot and dry. Vegetation cover may be relatively sparse and shrub-likein Mediterranean areas, but is much more substantive in subtropical regions. There is marked variability in runoff generation and geomorphic effectiveness of floods in this morphoclimatic zone. Some areas have extremely high coefficients of variation in discharge, with significant inter annual variability in flood events. Often, river systems are attuned to extremely high, but infrequent, flows. Mediterranean rivers have seasonal discharge and variable ground cover. Ephemeral streams are subjected to irregular rework ing by flash floods. Discontinuous watercourses are prom inent. Cooler humid regions have severe winters and continental climates, with significant areas of boreal forest. Rivers freeze in winter, and there is extensive permafrost in northerly latitudes. Profound adjustments may occur during spring melt. Polar regions are dry and cold, and bedrock-dominated rivers are relatively inactive. Landscape history and climate setting bring about marked variability in flora and fauna across the globe. Faunal interactions with rivers can affect the nature, rate and effectiveness of geomorphic processes. A wide range of ecosystem engineers is evident. Ants and worms induce bioturbation in soils, impacting upon sediment supply and transfer on hillslopes. Beaver dams exert a direct impact upon channels. Hippopotamus tracks may induce channel realignment. Wombat burrows may locally enhance rates of bank erosion. Changes to these faunal interactions may alter the evolutionary trajectory of the river. Similarly, any factor that alters vegetation cover (and associated resistance/ roughness) can have a significant affect upon the evolutionary trajectory of a river. For example, the geomorphic role of fire varies markedly in differing morphoclimatic regions. Savanna and Mediterranean areas are especially prone to fire events that clear ground cover, resulting in pulsed flow and sediment inputs into river systems. Climate is a key driver of river change. It fashions the sequence of disturbance events that bring about geomorphic adjustments, influencing system dynamics and the behavioural regime of any given reach. In some instances, f loods or droughts may bring about transitions to a differ ent type of river. Impacts upon the flow regime, and changes to ground cover, alter the rate of sediment move ment in river systems, thereby affecting both sides of the Lane balance diagram. Long-term changes during the Quaternary period have been induced by glacial–interglacial cycles (Figure 12.6). At the coldest part of the last glacial maximum (15 000 18 000 yr ago), ice covered one-third of the land area of the Earth to an average depth of 2–3 km, but in places up to4 km. Ice sheets created sculpted/denuded landscapes, creating slowly adjusting bedrock-dominated rivers. Alpine glaciers carved U-shaped valleys and fiords. During the recessional stages of ice sheet activity, meltwater channels realigned many drainage networks. Significant volumes of glacially reworked materials drape many landscapes, pro viding large sediment stores that can be reworked by river systems. Hence, there are marked differences in the historical imprint upon contemporary landscapes in glaciated and non-glaciated settings. Glacial cycles also brought about significant falls in sea level (up to 120 m). This exerted a profound impact upon patterns and rates of sedimentation in lowland basins as base level adjusted. Deep canyons were carved into what are now parts of the continental shelf. These effects were propagated upstream, leaving terraces at valley margins. Subsequent sea level rise during interglacial periods created drowned valleys and ria coastlines. Floodplain, terrace and marine sediments in infilled lowland valleys and estuaries retain records of multiple phases of sea level rise and fall. Longer term glacial–interglacial cycles also brought about major river changes in arid morphoclimatic zones, altering the distribution and extent of monsoonal climatic influences. As climate changes, so too does the vegetation cover. Hence, geomorphic adjustments reflect alterations to both impelling forces (the flow regime) and resisting forces (ground cover) (see Chapter 5). Geomorphic responses to climate change are markedly variable in different parts of the world. The impact of climate change is not simply a measure of the direction or extent of change. Temperature changes from −20 °C to −30 °C may not induce a marked difference in process response, but transition from −5 °C to +5 °C certainly does. Similarly, change in annual precipitation from 9000 to 10 000 mm a^{-1} is unlikely to induce marked variability in geomorphic process activity, but changes from 500 to 1500 mm a−1 definitely would, primarily because of altered vegetation cover. Geomorphic responses to variability in climatic conditions vary markedly for different types of rivers, reflecting their sensitivity to adjustment (Chapter 11). They also vary dependent upon the condition of the system at the time of the disturbance event (especially its resistance). In many instances, contemporary landscapes have been fashioned largely by conditions from the past. Landscape memory: imprint of past geologic and climatic conditions upon contemporary river processes, forms and evolutionary trajectory Contemporary rivers flow upon, and rework, surfaces created by past events. Hence, historical influences may exert a primary influence upon the distribution, rate and effectiveness of erosional and depositional processes. This imprint from the past varies markedly in differing settings. As noted in Chapter 2, landscape memory is fashioned primarily by past geologic and climatic conditions, or events. Some landscapes also retain a prominent memory of former anthropogenic activities (discussed in Chapter 13). Geologic controls determine the relief, topography and erodibility of a landscape (see Figure 12.7a and b). The influence of elevation upon potential energy mani fests itself as impelling forces (and associated kinetic energy) driven largely by slope (i.e. erosivity). This exerts a primary control upon the effectiveness of erosional processes and the resulting degree of landscape dissection. Geologic factors also influence the nature and extent of accommodation space and associated patterns of sediment stores in landscapes. Valley setting, in turn, affectschannel–floodplain relationships, thereby influencing the contemporary capacity for adjustment of rivers. The contemporary climate regime is a primary determinant of the flux boundary conditions under which rivers operate, affecting discharge and flow energy and vegetation and/or ground cover which resist erosion processes. Critically, these relationships have changed over time. The impact of these changes is especially pronounced in those parts of the world affected by Pleistocene glacial activity. Glaciers carved deep and narrow valleys in mountain areas, constraining the range of geomorphic behaviour of con temporary channels in these settings (Figure 12.7c). Many downstream areas were draped with glacially reworked materials. In some instances these vast (paraglacial) sediment stores that reflect former climatic conditions con tinue to influence contemporary river behaviour (Figure 12.7d). The distribution of these sediment stores is influenced largely by geologic controls upon the accommodation space in landscapes, such as wider sections of valleys that store glacio-fluvial, glacio-lacustrine and alluvial fan materials. In many other settings, ice sheets stripped surface materials from vast areas, limiting contemporary rates of sediment supply across largely denuded areas (Figure 12.7e). Another form of climatic memory is that associated with floodplain deposits of underfit streams (i.e.contemporary channels are too small to have formed the valleys within which they presently flow; Figure 12.7f). These inherited forms influence contemporary river morphology and associated patterns and rates of sediment erosion, transport and deposition. In this instance, climatic memory directly reflects geologic memory, as past geologic controls induced the accommodation space along palaeovalleys within which contemporary rivers flow. Landscapes retain a selective memory of past events. Sometimes a sharp erosional boundary reflects a major disjunct in time, highlighting the removal or erasure of a significant part of the record (Chapter 6). Indeed, some landscapes may retain a very limited history of past events. Elsewhere, especially in long-term depositional basins in accretionary environments, a remarkable long-term record may be preserved (much of which is buried subsurface). Hence, different parts of a landscape retain variable records of past activity. Ultimately, changes to boundary conditions drive river evolution. River responses to altered boundary conditions The Lane balance diagram provides a simplified basis with which to interpret primary controls upon river evolution. Essentially, if the bed stability of a river changes, so will the geomorphology. In other words, the balance becomes unsettled and adjustments ensue. The two key considerations here are the amount of water acting on a given slope and the volume and texture of sediment delivered to the channel. As noted above, geology and climate are the primary determinants of these factors. The tectonic setting determines the rate of uplift (i.e. relief and sediment gen eration, and erosion rate), while lithology determines the breakdown size of weathered/eroded materials. Uplift or subsidence also alters the slope upon which geomorphic processes are acting. Climatic factors determine the flow regime and the amount of water available to do work in river systems. Evolution is driven by changes to these various controls. Davisian notions of landscape adjustment infer that rivers evolve as slopes decrease and valley floors widen over geologic time, prior to uplift kick-starting the cycle once more (Chapter 2). Such continuity in boundary conditions, and even the direction of change, is seldom observed in reality, as invariably something happens to disrupt these patterns over timeframes of millions or tens of millions of years. Disturbance events may alter the flow–sediment balance along a river, whereby changes to geologic and climatic conditions induce adjustments in process relation ships along valley floors, and resulting river morphologies. Various examples of river evolutionary adjustments in response to altered boundary conditions, disturbance events and flow–sediment fluxes are outlined below. River responses to tectonic uplift and displacement along fault-lines Uplift of a fault block, or even an entire plateau landmass within a plate, induces rejuvenation, whereby rivers are made young again and incise into underlying bedrock. If the rate of bed incision is unable to keep up with the rate of uplift, convex bulges are created along longitudinal profiles. These areas are characterised by waterfalls and/or oversteepened sections of the bed profile (see Chapter 3). In some instances, knickpoint erosion may instigate river capture, wherein flow that was previously part of a separate basin is realigned and captured as a headward-cutting channel eats through the drainage divide over time (see Figure 12.8). An underfit stream now flows within the abandoned valley (i.e. the stream is much smaller than the river that created the valley itself). Elsewhere, stepped longitudinal profiles with multiple waterfalls reflect the recur rence of uplift events and the hardness of bedrock layers through which knickpoint retreat occurs. The pulsed nature of bed incision and knickpoint retreat in tectonically active settings is often accompanied by dramatic influxes of sediment from hillslope failures, some of which dam the river with variable longevity. Alternatively, lateral displacement along fault-lines during earthquake eventscan realign and/or reconfigure river systems. This can occur in a lateral dimension (Figure 12.9a) or vertical dimension (Figure 12.9b). River responses to long-term changes in valley setting Rivers are products of the valleys in which they flow. Long term changes to valley morphology reflect geologic con trols (see Figure 12.2c). For example, progressive knickpointretreat along trunk and tributary rivers at the plate margin creates series of dissected gorges in escarpment-dominated landscapes at the margins of pull-apart basins. These valleys cut backwards and incise far more rapidly than they widen. Changes to valley floor slope and valley width over millions of years induce transitions from a gorge to a partly con fined valley with bedrock-controlled discontinuous flood plains (Figure 12.1b) and subsequently to a partly confined valley with planform-controlled discontinuous floodplains (Figure 12.1c).River responses to major sediment inputs Rivers respond to marked increases in sediment load by aggrading. In some instances this may bring about pro found landscape responses. For example, volcanic eruptions can drape vast volumes of material across a landscape, transforming incised bedrock streams into highly sediment charged systems that may infill valleys to considerable depth, promoting the development of braided rivers (Figure 12.10). These localised and irregular disturbance events are relatively spatially constrained (i.e. they occur in semi-predictable places, determined primarily by tectonic setting). Volcanic disruptions to river systems occur primarily in subduction and pull-apart settings and in response to hot spot activity (i.e. areas of thin crust through which molten materials from the upper mantle are released at the Earth’s surface). Volcanic events are generally recurrent (i.e. they occur at the same place on repeated occasions, and resulting materials build up over time). Landscape responses are fashioned by the magnitude of an eruption, resulting sediment inputsand the interval between events (i.e. the length of time over which sediment reworking occurs). In general terms, volcanic landscapes that have not experienced an eruption for a significant period tend to become deeply etched bedrock-controlled systems. These rivers are resilient to change during flood events. However, eruptions bring about dramatic transformations, altering all attributes of the river. Lahars and debris flow deposits line valley floors. Aggradation induces braided rivers with an array of mid channel depositional geomorphic units. Typically, these are short- to medium-term adjustments post-eruption, as the river progressively adapts to prevailing flow–sediment con ditions by incising into its bed (i.e. flow conditions remain relatively consistent over time, while the rate of sediment production is not maintained). Incision and reworking promote a transition back to increasingly imposed river morphologies. Downstream transfer of materials accentu ates bed incision and the deeply etched character of the landscape. The imprint of volcanic events brings about a range of localised and off-site impacts. Tephra deposits may create a significant drape of materials over vast areas. Rivers sub sequently flow within very light, low-density, highly porous materials, such that coarse bed material is readily conveyed within the channel (often as suspended load). In other set tings, ignimbrite flows may infill valleys and create plateau like landscapes with caps of extremely resistant materials. Valley incision and headward retreat subsequently demar cate these materials as knickpoints and waterfalls along longitudinal profiles. Long-term erosion of volcanic landscapes can create inverted relief. This occurs when lava flows infill valleys, f lattening out the ground surface (Figure 12.11). As the thicker basaltic materials are often more resistant to erosion than the surrounding country rock, long-term progressive erosion may result in basalt-peaked caps derived from materials previously deposited on valley floors as the high points in these landscapes. While volcanic events induce massive sediment inputs into riverscapes over irregular but infrequent timescales, more recurrent but much smaller sediment inputs occur in response to landslides and associated hillslope instability events (Figure 12.12). A range of outcomes may occur, dependent upon the amount of sediment input, the size of the valley and the capacity of the river to rework these deposits. In extreme instances the valley may become blocked, forming a dam and lake. This alters the base level of the trunk stream, resulting in aggradation and delta growth within the lake. Downstream, the channel responds to reduced sediment loads by incising. Eventually the dam may break. This results initially in extensive flooding and erosion of downstream reaches. Subsequently, the massive influx of deposits induces aggradation as a sediment slugmoves through the system. Extreme landscape responses to landslide events are especially pronounced following earth quakes or extreme storms (cyclones). Such scenarios are especially pronounced in steep, dissected terrains close to plate margins in regions with (sub)tropical climates. In others settings, hillslope-derived materials may be stored along valley floors for a considerable period of time. This is primarily determined by valley width, and associated hillslope–valley floor connectivity and the space for sediment storage (Chapter 14). If these deposits are not accessible to the channel, they may have a negligible impact upon river behaviour and change. River responses to climate change (flow regime and ground cover changes) Impacts of climate change and variability may be manifest through localised extreme events, semi-regular, systematic changes (e.g. glacial–interglacial cycles) or long-term adjust ments associated with the movement of tectonic plates (the distribution of land masses is a primary control upon weather patterns; Figure 12.3). Of primary concern here isthe impact of changing boundary conditions and disturbance events upon the way in which rivers operate, and their evolution. Examples of system responses to climate induced alterations to the flow–sediment balance and ground cover are outlined below. Climate change induces marked variability in the char acter, behaviour and evolution of river systems in glaciated and non-glaciated landscapes. Phases of glacial activity in mountainous terrain induce extensive erosion and sculpt ing of landscapes. The mountains themselves are etched and denuded, while valleys are carved. Stripped surficial materials and bedrock are broken down and conveyed considerable distances from source. As a consequence, the boundary conditions upon which rivers operate are transformed. Transitional climatic phases at the ends of ice ages are periods of intensive geomorphic activity. This period is referred to as the paraglacial interval. Melting glaciers and ice sheets result in pronounced discharge variability. Hills lopes are unstable, as previously supporting ice has melted, and vegetation cover is negligible. This results in extensive sediment movement, aggrading valley floors and the formation of large alluvial fans. Steep slopes, abundant bedload-calibre material and fluctuating discharge result in braided river systems. Extensive braidplains (or valley sandar) are evident at the margins of many contemporary glaciers or ice sheets. Over time, discharge is reduced and streams incise into their beds, creating extensive terrace sequences (Figure 12.7d). Rivers retain extensive sediment loads, and braided channel planforms extend well beyond the mountain front. Amelioration of climatic conditions over thousands of years results in less variable discharges, diminishing sediment loads and increases in vegetationcover on hillslopes and valley floors. Rivers respond by changing to wandering gravel-bed or active meandering systems (Figure 12.1b). In some instances, post-glacial climate changes may gen erate some truly epic landscapes, inducing profound altera tions to river systems. This is exemplified by breaching of ice-dammed lakes, which release vast volumes of water in truly catastrophic flows (termed jokulhlaups). These floods may etch and sculpt vast terrains, fashioning future drain age networks and resulting river morphologies (Figure 12.13). Elsewhere, streams beneath glaciers or significantmeltwater flow can realign drainage networks (a form of river capture). Many landscapes retain a significant cli matic memory from these post-glacial events. Although non-glaciated terrains are not subjected to paraglacial sedimentation and breaching of ice-dammed lakes, dramatic landscape responses to changing climatic conditions may occur in these settings. For example, former river courses in some desert landscapes have been draped by sand dunes in response to drier climates and reduced vegetation cover during glacial periods (Figure 12.14). Some non-glaciated landscapes have been subjected to progressive drying over hundreds of thousands of years. Marked reductions in channel geometry, along with notabledecreases in sinuosity and bed material size, result in rivers that are clearly undersized for the valleys within which they f low (i.e. these rivers are underfit; see Figure 12.15). Figure 12.16 conveys marked transitions in the post glacial history of a river in a non-glaciated landscape. In this evolutionary sequence, sparsely vegetated slopes and f luctuating discharge conditions induced a braided river planform at the last glacial maximum. Climate amelioration and vegetation growth brought about dramatic trans formation of the flow–sediment balance, whereby the energy of the system was diminished to such an extent that the river became a discontinuous watercourse with a fine grained swamp that accumulates suspended-load deposits.These valley floor deposits have subsequently been incised to create a continuous channel. Glacial–interglacial cycles induce significant sea level change (eustasy). This may bring about geomorphic adjust ment along the lower course of rivers. Sea levels may be lowered by 120–150 m during glacial maxima, essentially extending river courses onto what is now the continental shelf. The nature of geomorphic adjustment varies for differing fluvial–marine interactions (i.e. whether a delta or estuary is present) and the nature/extent of the continental shelf itself. Profound adjustments are noted along the lowland plains of large rivers, where incised valleys and fills develop significant terrace sequences (Figure 12.17). These terraces, in turn, constrain subsequent channel responses during periods of rising sea levels. Alternatively, the pro found weight of accumulated deposits along the lowland plains of rivers, or in inland-draining (endorheic) basins, may induce subsidence via isostatic adjustment. Given the very low slope of these settings, avulsion may be experienced along these low-energy, suspended-load rivers. Ongoing climate changes associated with global warming are bringing about marked geomorphic transitions for some rivers. For example, melting permafrost has increased discharges and the erosive potential of many rivers that drain into polar regions. Impacts of ice flows following spring melt have been accentuated. This exemplifies region ally specific patterns and trends in the evolutionary adjust ment of rivers. Finally, the impacts of climate changes upon rivers must be related to the magnitude–frequency relations of forma tive events, especially the geomorphic effectiveness of extreme floods. As noted previously, there is significant variability in response in differing landscape and climatic settings. This reflects the sensitivity/resilience of a river, and the extent to which the river is attuned to seasonal and interannual variability in discharge. In some instances, extreme floods may exert a profound imprint or memory upon the system, whereby the river is subsequently unable to adjust its boundaries. Depending upon the condition of the system at the time of the event, and associated availability of sediment, flows may be highly erosive or highly depo sitional. Either way, transformation of channel boundaries exerts a significant influence upon the subsequent evolutionary adjustments of the river. These various pathways and rates of geomorphic evolution are meaningfully cap tured using the river evolution diagram.

Linking river evolution to the natural capacity for adjustment: adding river change to the river evolution diagram

The river evolution diagram introduced in Chapter 11 can be used to evaluate pathways of river evolution in relation to the behavioural regime for any given type of river. In general terms, the greater the range of variability in geo morphic behaviour demonstrated by a river (i.e. the greater the degrees of freedom and capacity for adjustment), the greater the likelihood that evolutionary adjustments and geomorphic change will occur over shorter timeframes. Conversely, the more resilient the river the longer the time frame for discernible geomorphic adjustment. These deter minations reflect the imposed boundary conditions within which a river operates, as shown by the outer band of the river evolution diagram. The width of the outer band increases from confined through partly confined to later ally unconfined settings, as the potential range of variability increases. Rivers can more readily adopt differing morphologies and behavioural attributes if there is space for the channel to adjust on the valley floor. A similar degree of variability is evident in the width of the inner band on these figures. This reflects the natural capacity for adjustment as determined by flux boundary conditions. The width of this inner band represents the range of states that the river can adopt while still being considered to be the same type of river (i.e. retaining a consistent set of core geomorphic attributes that reflect the character and behaviour of that river type). In a sense, this is a measure of the sensitivity of the river, as it records the ease with which the river is able to adjust. As indicated for the potential range of variability, the width of the inner band is greatest in laterally unconfined settings. River responses to disturbance events are indicated by the arrows shown at the top of the inner band on the river evolution diagram. The spacing of the arrows indi cates their frequency, while the size of an arrow indicates its magnitude. In most instances, disturbance events promote river adjustments but the reach remains within the inner band (i.e. perturbations fall within the natural capacity for adjustment). River adjustment within the inner band may breach intrinsic threshold conditions, marking a shift in the way energy is used (either concentrated or dispersed). Typically, this reflects an adjustment in the character or distribution of resisting forces (e.g. bed resistance, form resistance, resistance induced by riparian vegetation or wood). These internal adjustments alter the assemblage of erosional and depositional landforms on the valley floor, yet fall within the behavioural regime of the river. In other instances, changes to the prevailing flux bound ary conditions and/or severe disturbances may bring about changes to the formative processes that fashion river mor phology (i.e. river change has occurred). This scenario is highlighted by the shift in the position of the inner band. In these instances, altered stream power relationships reflect differing energy use in relation to prevailing flux boundary conditions. Reaches now operate within a differ ent inner band on the river evolution diagram, with alteredenergy conditions. The shape of the pathway for adjust ment, shown by the jagged line within the inner band, has a different form for the new river type, depicting a change in process–form associations along the valley floor, such that there is a change in river morphology. The new con f iguration represents a different type of river, with a differ ent appearance (character) and set of formative processes (behaviour). Inevitably, there may be some overlap in the position of former and contemporary bands, and some geomorphic units may be evident in both situations. However, the assemblage of geomorphic units in the two bands differs, reflecting a change to river character and behaviour. The shift in the position of the inner band can be induced by a press disturbance that exceeds an extrinsic threshold. This usually reflects alteration to flux boundary conditions, as modified flow and sediment transfer regimes (i.e. impel ling forces) drive river change. In this case, the time that is required for recovery following perturbation is longer than the recurrence interval of disturbance events. Effectively, the previous configuration of the river was unable to cope with changes to the magnitude and rate of applied stress. Rare floods of extreme magnitude, or sequences of moderate-magnitude events that occur over a short inter val of time, may breach extrinsic threshold conditions, transforming river character and behaviour. Dependent on the subsequent set of process–form associations adopted by the river, the natural capacity for adjustment may widen or contract as the new type of river adjusts to different flux boundary conditions. The position of the inner band within the potential range of variability (the outer band) indicates whether the change in river type marks a transition to a higher energy state (an upwards adjustment) or a lower energy state (downward adjust ment). Changes to the amplitude, frequency and shape of the pathway of adjustment within the inner band indicate how the river responds to pulse disturbances of varying magnitude and frequency. In some cases, change may occur during a threshold breaching flood. In this instance the two inner bands are located adjacent to each other and the date of the disturbance event is noted. In other cases change may be lagged or occur progressively over time. In these instances, the space between the two inner bands is widened to depict whether change occurred over years or decades. In more complex situations, transitional river types are depicted on the diagram. For simplicity, only a major shift between one river type and another is shown in the examples outlined below. Figures 12.18–12.23 build upon the interpretation of river behaviour for various river types presented in Chapter 11, using documented examples of river evolution from the literature. Emphasis is placed upon the nature of evolutionary changes to the river, timeframes over which these changes occur and evolutionary trajectory. Figure 12.18 represents an imposed river configuration such as a gorge. Disturbance events that have the capacity to induce changes in other settings are unable to bring about significant geomorphic adjustments along confined rivers, as the inherent resilience of the system is too strong. Perturbations to the flow and sediment regime are accommodated by instream adjustments to hydraulic resistance, such as the nature and distribution of bedforms, dissipating flow energy. Adjustments to river character and behaviour are negligible and the river type remains the same. Millions of years of valley widening may allow for out-of channel deposition and generation of floodplain pockets, but the assemblage of erosional and depositional geomorphic units along the reach is likely to remain consistent over tens of thousands of years (at least). A different pattern of responses to changes in external stimuli may be experienced in partly confined valley set tings, where the potential range of variability is somewhat broader than in confined valleys (Figure 12.19). This enables a greater range of possible river morphologies to develop. Antecedent controls and prevailing flux boundary conditions shape the contemporary configuration of the river. A bedrock-controlled discontinuous floodplain river has negligible capacity for adjustment because of thebedrock-imposed setting. Valley widening over tens of thousands of years results in progressive transition to a planform-controlled situation. The example demonstrates potential adjustments in this situation, as there is greater capacity for adjustment because of the greater degrees of freedom. Local areas of the channel are able to adjust their planform within the partly confined valley. For example, lateral migration may form ridges and floodchannels within the vertically accreted silty floodplain. In this instance, the natural capacity for adjustment has shifted to a lower energy river type. This is indicated on the river evolution diagram by a downward shift in the position of the inner band (the natural capacity for adjustment) within the outer band (the potential range of variability). In addition, the range of river behaviour has been reduced (i.e. the width of the inner band has narrowed; note the logarithmic scale).

Rivers are more sensitive to change in laterally uncon f ined valley settings relative to partly confined and confined valleys (i.e. the potential range of variability and the natural capacity for adjustment are greatest in laterally unconfined valley settings). Changes are shown from a braided configu ration to a meandering mixed-load system (Figure 12.20), from a mixed-load meandering to a suspended-load mean dering river (Figure 12.21), from a gravel-bed braided to a low-sinuosity sand-bed river (Figure 12.22) and from a braided to a fine-grained discontinuous watercourse (Figure 12.23). These changes reflect alterations to both the impelling forces that promote change (i.e. less variability in flow, less coarse-sized material on the valley floor, etc.) and internal system adjustments that modify the pattern and extent of resistance. A major shift in the assemblage of geomorphic units ensues, resulting in altered patterns of mid-channel and bank-attached geomorphic units, andprocesses of floodplain formation and reworking. Channel geometry and bedform assemblages are transformed as well. Critically, these adjustments occur over much shorter timeframes than those indicated for the examples shown in confined and partly confined settings in Figures 12.18 and 12.19. The four transitions in river character and behaviour shown for laterally unconfined valley settings in Figures 12.20–12.23 all show a downward shift in the natural capacity for adjustment within the potential range of vari ability. This reflects the adoption of a lower energy river type within the same landscape setting. This transition is especially pronounced in Figure 12.23. In some instances, an increase in resistance increases the capacity of the system to trap finer grained materials, thereby aiding the transition to a single-channelled or discontinuous channel configuration. Increased stability enhances prospects for vegetation development on the valley floor. As a result, the natural capacity for adjustment is narrower, reflecting a reduction in the range of behaviour. Changes to energy relationships reflect the consumption of energy, altering the pathway of adjustment. For example, the transition from braided to meandering configurations shown in Figure 12.20 is marked by a switch from tight chaotic oscillations reflecting recurrent reworking of materials on the channel bed to a jagged shape that reflects the occasional formation of cut-offs and subsequent readjustment of channel geometry, planform and slope. Post-glacial adjustments to flow and sediment fluxes com monly induced changes from a braided to a meanderingchannel planform (Figure 12.20). In the early post-glacial interval, abundant sediment, highly variable flows and negligible vegetation cover promoted the development of braided rivers. A wide range of mid-channel bars and shift ing channels of varying size characterised these bedload dominated systems. Progressive reduction in sediment availability in the post-glacial era, along with reduced variability in discharge and progressive encroachment of vegetation onto the valley floor, brought about the transformation of many of these braided rivers into mixed-load meandering systems by the mid-Holocene. These rivers are now characterised by laterally migrating single channels with point bars and associated instream geomorphic units, and an array of laterally and vertically accreted floodplain forms. The impacts of long-term drying upon the planform and geometry of rivers in non-glaciated environments shown in Figure 12.15 are reconstructed using the river evolution diagram in Figure 12.21. This shows the transformation from a mixed-load laterally migrating channel into a slowly migrating suspended-load river with a much smaller channel capacity. This transition reflects a decline in fluvial activity driven by changes to the discharge regime. Different pathways and rates of adjustment may be experienced by different types of rivers subjected to similar climatically induced changes to prevailing flow and sediment fluxes. In Figure 12.22, a stable low-sinuosity sand bed river with a vertically accreted floodplain has replaced a gravel-bed braided river. Figure 12.23 builds upon the example shown in Figure 12.16, showing the transition from a braided configuration at the last glacial maximum to the development of a fine-grained discontinuous water course in a cut-and-fill river. A range of tools and approaches used to analyse and interpret river evolution is outlined in the following section.

Reading the landscape to interpret river evolution

Interpretations of river evolution by reading the landscape can be complemented by sediment analysis and use of dating techniques, process measurements, appraisal of his torical records and modelling applications (see Table 12.1). In some instances, ergodic reasoning (space for time sub stitution) can be used to construct evolutionary sequences(Chapter 2). Assessment of river evolution at any given locality must be framed in its spatial context (within catchment position and in relation to regional patterns/ trends), alongside broader scale geologic and climatic considerations (i.e. tectonic setting and records of climate change). Typically, topographic maps, geology maps, remotely sensed images and resources such as Google Earth® are analysed prior to going into the field. Controls upon the contemporary character and behaviour of the river (Chapters 10 and 11) must be assessed before mean ingful interpretations of evolutionary adjustments can be performed. This entails analysis of river forms and proc esses in relation to geologic and climatic controls upon imposed and flux boundary conditions, and the associated range of disturbance events to which the river is subjected. Questions asked in these preliminary investigations include: • What is the landscape setting – geology, climate, vegetation cover? Is this a glaciated landscape, a desert, a melt water channel, an urban stream, a tropical rainforest, the flanks of a volcano? How does the setting impact upon the erodibility/erosivity of this landscape, and associated flow–sediment fluxes? • How does position in the landscape/catchment, and associated slope, catchment area and valley width affect the nature and effectiveness of erosional and depositional processes (i.e. is this a source, transfer or accumulation zone)? • How is the reach affected by downstream or upstream controls? How connected are hillslopes to the valley f loor? Building upon these geographic relationships, field anal yses of river evolution interpret the range and pattern of geomorphic units observed in a given setting. Analysis of the sedimentary record involves interpretation of the inter nal structure and characteristics of sedimentary sequences for a given landform (Chapter 6). Spatial relationships between landforms provide a basis to interpret depositional and erosional histories at the reach scale. By inter preting the sequences of sediments preserved in basin fills, stages or phases of evolutionary adjustments can be differentiated and formative events can be appraised. Inevitably, any landscape retains an incomplete record of past activities and events. Bare bedrock in confined valleys and supply-limited landscapes is indicative of erosional surfaces. Cosmogenic dating techniques can be used to determine exposure dates of differing surfaces, from which erosion rates can be determined. Reworking of deposits provides a partial preservation record in partly confined valley settings. More complete depositional sequences are evident in laterally unconfined settings and transport-limited landscapes where basin fills may record activities over long timescales. Much of the record may be buried. Terraces and floodplains often preserve records of deposition and reworking that extend back over thousands or tens of thousands of years. Insight into reworking events can be gleaned from erosional surfaces (discontinuities or unconformities) in the sediment record. Are boundaries onlapping (depositional) or erosional (e.g. local scour or f loodplain reworking)? Do they indicate changes in river behaviour (e.g. change in type of floodplain deposit)? Are depositional sequences in bank exposures consistent with deposits laid down by the contemporary river, or are they indicative of change? Disjuncts (unconformities) in the depositional record are indicative of erosive events. Linking sediment sequences to their chronology is vital in determining phases and rates of activity. Assessment of the preservation potential of deposits provides guidance into what may be missing (erasure) and the record of events that may have been obliterated by later erosion. Juxtaposition of units often represents a hiatus and/or change in process relationships. When combined with dating tech niques, phases of river evolution can be interpreted and rates of change determined. Dating tools can be used to generate age estimates of depositional features, providing insight into the time they were laid down (or reworked), the timeframe of disjuncture between eroded units and the period of time that has been lost from the depositional record. Ideally, the erosional/depositional history in one reach is related directly to evolutionary adjustments in upstream and downstream reaches. These interpretations can be supplemented by process measurements to assess the rate and effectiveness of geo morphic process activity. From this, magnitude–frequency relationships can be derived to assess how much work is likely to be performed for an event of a given magnitude. These relationships are extremely important in deriving rating curves that estimate sediment transport (Chapter 6) or formative flows that fashion channel geometry (Chapter 7). A range of logistical problems besets field-based measurement of geomorphic processes. First and foremost, the representativeness of the data (in space and time) must be assessed. How accurate/precise are the data themselves? How reliably can they be extrapolated to other situations? In many instances, measurement techniques may disturb the observed processes. As yet, many processes and phenomena cannot be observed or measured directly or even indirectly. Real-time or lapse observations and measure ments may be extremely helpful in interpreting frequent low-magnitude events, but instruments are often destroyed in catastrophic high-magnitude events. Ironically, these events may well be the primary agents of landscape adjust ment. All too often, the timescale of human observation is much shorter than that of the phenomenon under study. There are remarkably few, sustained programmes of longer term (decadal) process measurement. As such, it is difficult to discern magnitude–frequency relationships in a comprehensive manner.In some instances, stages of landscape evolution can be appraised through reasoning by analogy (ergodic reason ing), which is the recognition of similarity among different things. Comparative frameworks can be used to relate states (or stages) of evolutionary adjustment in different areas that have a similar landscape configuration (i.e. equivalent features are produced by the same set of proc esses under an equivalent set of conditions). This is referred to as space for time substitution. Time slices can be used to interpret the pathway of adjustment that is likely to be experienced for reaches of the same river type. The reliability of predictions is dependent on the similarity of the places that are being compared and the range and rate of processes and disturbance events to which they are sub jected. Similar outcomes may arise from different processes and causes (the principle of convergence or equifinality). A common origin or equivalent causality is a prerequisite for effective comparison. Increasingly, geomorphologists simulate real-world under standing as a basis to interpret process understandings, identify key controls upon process–form linkages, assess rates of activity and predict evolutionary trajectories through the use of physical and numerical models and experimental procedures (e.g. flume studies). This provides an important platform to assess understandings of real- world situations. Hypotheses and future scenarios can be tested. While modelling provides a critical basis to assess magnitude–frequency relationships for individual proc esses, it is difficult to 'scale up' processes and interactions in a way that meaningfully captures landscape-scale dynamics at the catchment scale. Models cannot generally take account of the intrinsically random or chaotic disturbances that drive landscape change, or their non-linear and complex responses. Concerns arise about the selection of input parameters and the transferability of insights from one system to another. Hence, significant questions remain about the representativeness and replicability of modelled output to real-world situations. Field verification provides the critical test of our understanding. Tools such as reading the landscape are required to meaningfully adapt findings from modelling applications to real-world conditions, circumstances and situations, linking field interpretations to theoretical understandings.

Tips for reading the landscape to interpret river evolution

Step 1. Identify individual landforms and their process–form associations Critical insight into river evolution can be gained by assess ing whether the landforms that make up any given reach are products of the contemporary behavioural regime of the river or they reflect former conditions. Palaeo landforms provide a 'glimpse' into the past. For example, the morphology, position and sedimentary structure of ter races and palaeochannels can guide insight into past flow and sediment regimes. This aids determination of whether the bed is aggrading or degrading, former channel dimensions and planform, and the sediment-transporting regime (e.g. bedload, mixed-load or suspended-load river). Step 2. Interpret river change at the reach scale The key to analysing river change is to determine whether a wholesale change in river character and behaviour has occurred such that a new river type occurs. Critically, what has changed, and why? Analysis of the pattern/position of landforms can aid interpretations into likely sequences of events. It is also important to determine the timeframe over which change has occurred (last year, decade, century, 1000 yr, 10 000 yr, million years) and to assess stages/phases of evolution as a series of evolutionary time slices. Analysis of changes to the geomorphic unit assemblage provides insight into the altered process regime, indicating how erosional or depositional processes, and their relative balance/effectiveness have changed over time. For example, terraces, inset floodplains and knickpoints are indicative of changing geomorphic conditions and associated phases of activity. Similarly, differing packages of units may be evident across the valley floor (e.g. ridge and swale topography, cut-offs, avulsion). Beyond this, any indications of adjustments to channel geometry and associated bank erosion/deposition processes must be unraveled, deter mining whether these attributes fit with the contemporary f low and sediment regime. Is the size and shape of the channel a function of the current flow and sediment regime or a remnant from the past? Is there any indication of changes to channel planform? Interpretation of erosional or depositional boundaries between geomorphic units provides insight into the reworking of these features and phases/sequences of events that characterise reach evolution. Step 3. Explain controls on river change at the reach scale It is often exceedingly difficult to isolate the impacts of past events upon river evolution. Some events or phases of geomorphic impacts leave a dominant imprint upon the contemporary landscape, essentially overriding (overprint ing) the geomorphic signal of previous events. While the records of these events may persist, they may erase signals of previous activity. In some instances, relatively trivial events may be selectively preserved, whereas impacts ofother formative events may have been entirely erased. Erasure creates a disjunct in time in the features that are preserved along the valley floor. However, accumulation zones may retain a near-continuous record of sediment preservation over time. In simple terms, whatever fashioned the valley controls the river. In some instances, valley morphology may have been shaped by past glacial activity. Elsewhere it may have been superimposed following tectonic uplift, or it may be produced by river capture. These long-term landscape con trols fashion contemporary process relationships on valley f loors. Evidence used to determine the evolutionary sequence can be used to interpret changes to the boundary conditions and associated disturbance events that triggered the history of adjustments. How and why has channel geometry changed? Have alterations to channel alignment or enlargement/contraction modified the use of energy along the reach, changing the mix of bed and/or bank processes and channel–floodplain linkages? Has the assemblage of instream geomorphic units changed (e.g. have transitions occurred between mid-channel and bank-attached features, reflecting alteration of the erosional/depositional balance of the reach)? From this, it is important to ask why formative processes are different. What factors altered the f low–sediment balance? Has available energy increased or decreased over time? Have resistance elements been altered (e.g. vegetation cover, wood loading)? Floodplain sediments may provide an indication of past depositional environments. Some floodplains were formed by different sets of processes that operated under very different conditions to those that occur today. Is a hiatus evident – is the con temporary floodplain forming in the same way it did in the past (e.g. transitions from lateral to vertical accretion)? Are formative and reworking processes consistent over time? If not, why not? Is there any evidence of palaeo-forms or transitions in floodplain type? The Lane balance diagram can be used to relate river changes to altered imposed and flux boundary conditions. Is there any evidence that the rate of change over time has been altered? Was evolution progressive or did the exceed ance of threshold conditions bring about dramatic/rapid change? What was the role of catastrophic events? Alterna tively, was change lagged after the disturbance

(Chapter 2)? What kinds of disturbance events brought about evolutionary adjustments to the river? Did geologic controls such as uplift, faulting, folding and tilting induced by earthquakes, volcanic activity or subsidence events bring about these adjustments, or were climatic factors responsible (e.g. cyclones, intense local storms, drought)? Are there any indicators of drivers of river evolution, such as flood debris (slackwater deposits), volcanic ash, fault scarp, mega landslides, earthquakes, etc.? These analyses must be framed in relation to river sensitivity. Marked variability in the type, ease and rate of adjustments is evident for bedload, mixed-load and suspended-load rivers. Similarly, reaction and relaxation times after disturbance vary. Step 4. Explain how catchment-scale relationships affect river evolution Putting it all together at the landscape scale entails explanation of catchment-wide river responses to changes in imposed and flux boundary conditions, and associated dis turbance events that drive evolution. Evolutionary assess ments must appraise what is going on at any given location in relation to what is going on elsewhere in the system. In some instances, contemporary landscapes reflect lagged responses to disturbance events elsewhere in the catch ment. Response gradients record how linkages and con nectivity transmit upstream signals of disturbance response to downstream reaches. In essence, reading the landscape entails linking these spatial and temporal considerations in any given system, determining how what happened in the past or elsewhere in the system affects what is observed in any given reach today. Pathways and rates of river evolution are determined primarily by geologic and climatic controls upon landscape setting (Figures 12.2 and 12.5), the memory or imprint of the past (Figure 12.7) and the combination of disturbance events to which a river is subjected. Constructing evolutionary sequences is a system-specific exercise. Each river must be placed in its geologic/tectonic and climatic context to interpret drivers of river evolution. This entails assess ment of adjustments to boundary conditions and the nature/effectiveness of disturbance events over time. River location must be appraised in its tectonic context, assessing position relative to plate margins and the type/rate of activity at that margin. The erodibility and erosivity of a landscape exert key controls upon river types and their evolutionary adjustment. Long-term topographic changes that affect relief, slope, drainage pattern, drainage density, valley width, etc. must be assessed to interpret how imposed boundary conditions may have changed. In some instances, underfit streams may have developed, wherein the contemporary river flows within a valley created by much larger f lows. This may reflect river capture or the influence of meltwater channels. Other forms of geologic imprint (memory) include responses to disturbance events such as earthquakes or volcanic eruptions. Inevitably, geologic controls must be viewed in relation to changing climatic conditions and associated variability in flow, sediment and vegetation interactions (i.e. flux boundary conditions). Increases or decreases in discharge may alter the available energy of formative flows. However,associated changes to vegetation cover and surface rough ness alter the effectiveness of these flows to do geomorphic work. Non-linear responses are common. Appraisal of the effectiveness of disturbance events and associated magnitude–frequency relations is a key consideration in determination of evolutionary adjustments. In some instances a dominant imprint from the past may influence contemporary process–form interactions (i.e. climatic memory). For example, sediment availability fashioned by paraglacial processes is a primary determinant of river character and behaviour in many glaciated landscapes. The key issue here is determining how changes to climatic factors have influenced the natural range of behaviour and evolutionary trajectory of the river, assessing how and why the river has responded to changes in flux boundary conditions in the way in which it has. Perhaps the key question to address here is whether the river is well adjusted to its contemporary setting (slope, discharge regime, etc.) or whether certain attributes are products of former geologic and climatic considerations. Evolutionary considerations must be framed in relation to within-catchment position, appraising the ways in which disturbance responses are conveyed through a landscape, and their consequences. Changes to base level may induce significant evolutionary adjustments, whether as a product of progressive knickpoint retreat, the role of resistant lithologies, landslide-induced dams (however temporary) or sea level changes. Also, alterations to tributary–trunk stream relationships may bring about adjustments to river character and behaviour. Comparison of system-specific evolutionary histories is required to appraise the transferability of insights from one situation to another. Are inferred evolutionary records consistent from locality to locality within a catchment and across a region? If so, this can be related to broad-scale climate variability. Is the direction and rate of change local or regional; is it systematic, or is it subject to local-scale variability in controls? Informed appraisals from the regional record can be used to generate a more complete picture of phases or stages of river evolution and their timing. Catchment-specific applications should be related to regional trends to see whether scenarios are similar or different from adjacent catchments. These records should be analysed in relation to climate patterns, histories of f lood events and responses to known disturbance events. The challenge here lies in unravelling these various spatial and temporal considerations.

Conclusion:Long-term river evolution is fashioned largely by tectonic setting and geologic history. This determines the imposed boundary conditions within which contemporary proc esses operate. Inset within this, climatic controls determine the mix of water, sediment and vegetation interactions that occur in any given landscape. Changes to flux boundary conditions drive the evolutionary trajectory of a river, pro spectively inducing river change to a different river type. Many landscapes are products of recent adjustments. Else where, landscapes may reflect great antiquity, such that rivers retain the imprint of antecedent geologic or climatic conditions. Geomorphological interpretation of river evo lution unravels how a river has adjusted and changed over various timeframes and the range of disturbances (causes) that induced these changes. However, reading the land scape does not end there! In the next chapter, various forms of human disturbance that may modify river character, behaviour and evolution are outlined(Fryirs & Brierley, 2013).

References:

Blum, M.D. and Tornqvist, T.E. (2000) Fluvial responses to climate and sea-level change: a review and look forward. Sedimentology 47(Suppl 1), 2–48.

Brierley, G.J. (2009) Landscape memory: the imprint of the past in contemporary landscape forms and processes. Area 42(1), 76–85.

Brooks, A.P. and Brierley, G.J. (2002) Mediated equilibrium: the influence of riparian vegetation and wood on the long term evolution and behaviour of a near pristine river. Earth Surface Processes and Landforms 27(4), 343–367.

Brunsden, D. and Thornes, J.B. (1979) Landscape sensitivity and change. Transactions of the Institute of British Geographers NS4, 463–484.

Church, M. and Slaymaker, O. (1989) Disequilibrium of Holocene sediment yield in glaciated British Columbia. Nature 337(6206), 452–454.

Fryirs, K. A., & Brierley, G. J. (2013). *Geomorphic Analysis of River Systems: An Approach to Reading the Landscape.* West Sussex: Wley-Blackwell.

Knox, J.C. (1995) Fluvial systems since 20,000 years BP. In: Gregory, K.J., Starkel, L. and Baker, V.R. (eds) Global Con tinental Palaeohydrology. John Wiley and Sons, Inc., New York. pp. 87–108.

Lewin, J. and Macklin, M.G. (2003) Preservation potential for late Quaternary river alluvium. Journal of Quaternary Science 18(2), 107–120.

Merritts, D.J., Vincent, K.R. and Wohl, E.E. (1994) Long river profiles, tectonism, and eustasy: a guide to interpreting f luvial terraces. Journal of Geophysical Research 99(B7), 14,031–14,050.

Nott, J., Young, R. and McDougall, I. (1996) Wearing down, wearing back, and gorge extension in the long-term denu dation of a highland mass: quantitative evidence from the Shoalhaven catchment, southeast Australia. Journal of Geology 104(2), 224–232.

Paine, D.M. (1985) 'Ergodic' reasoning in geomorphology: time for a review of the term? Progress in Physical Geography 9(1), 1–15. Simon, A. (1992) Energy, time, and channel evolution in catastrophically disturbed fluvial systems. Geomorphology 5(3 5), 345–372.

Whipple, K.X., Kirby, E. and Brocklehurst, S.H. (1999) Geo morphic limits to climate-induced increases in topographic relief. Nature 401(6748), 39–43.

UNIT-10: CHANNEL CHANGE THROUGH TIME: CAUSES AND EVIDENCES OF CHANNEL SHIFTING

River evolution is the study of river adjustment over time. Evolution is ongoing. Even if boundary conditions remain relatively constant, adjustments occur. Appraisal of the trajectory and rate of river evolution is required to assess whether ongoing adjustments are indicative of long-term trends or whether they mark a deviation in the evolutionary pathway of that river. Such insights guide interpretation of the likelihood that the direction, magnitude and rate of change will be sustained into the future. To perform these analyses, it is important to determine how components of a river system adjust and change over differing timeframes, and assess what the consequences of those changes are likely to be. Reconstructions of the past provide a means to forecast likely future river behaviour. Instinctively, human attention is drawn to landscapes that are subject to change. Observations of bank erosion, river responses to flood events, anecdotal records of river adjustments or analyses of historical maps and aerial photographs provide compelling evidence of the nature and rate of river adjustments. Efforts to read the landscape must frame these insights in a broader context, examine their representativeness and isolate controls upon evolutionary trajectories. For example, do these adjustments reflect modifications around a characteristic state and associated equilibrium scenarios over a given timeframe? Are short-term adjustments indicative of longer term trends? Has the river been subjected to threshold-induced change? How has the balance of formative and reworking processes and controls changed over time? Is the river sensitive or resilient to disturbance? How are responses to disturbance manifest through the catchment, remembering that an erosional signal in one place is often matched by a depositional signal elsewhere? Analysis of river behaviour in Chapter 11 highlighted how different types of rivers have differing capacities to adjust, such that they respond to differing forms of disturbance event in different ways. Attributes such as thalweg shift on braidplains, meander migration/translation, cutoff development or avulsion are characteristic behavioural traits for certain types of rivers. In some instances, alterations to the boundary conditions under which rivers operate may bring about river change, whereby the behavioural regime of the river is transformed, and the river is now characterised by a different set of process–form relationships. River evolution may occur in response to progressive adjustments, an instantaneous event (e.g. a major flood or an earthquake) or longer term changes to geologic and climatic boundary conditions. This distinction between behaviour and change is essentially a matter of timescale. All rivers change as they evolve over time. In essence, if the geomorphic structure of a river changes, so does everything else (i.e. process relationships and the balance of impelling and resisting forces at the reach scale (Chapter 5) encompass adjustments to bed material organisation (Chapter 6), assemblages of instream and floodplain geomorphic units (Chapters 8 and 9) and channel geometry (Chapter 7)). River change can result from alterations to impelling forces, resisting forces, or both. Resulting adjustments modify the nature, intensity and distribution of erosional and depositional processes along a reach. In some instances, predictable transitions can occur. For example, a change from a wandering gravel-bed river to an active meandering river can occur as flux boundary conditions are altered to reduce sediment load and discharge, or vegetation cover is increased. However, just because a particular type of river in a given system responds to an event of a given magnitude in a certain way, does not mean that an equivalent type of river in an adjacent catchment will respond to a similar event in a consistent manner. Even if particular cause and effect relationships are well understood, some systems may demonstrate complex (or chaotic) responses to disturbance events (see Chapter 2). More importantly, no two systems are subjected to the same set of disturbance events. Each system has its own history and its own geography (configuration), with its own cumulative set of responses to disturbance events, and associated lagged and off-site responses. The trajectory of river change may be influenced by the co-occurrence of disturbance events, such as a large flood following vegetation clearance. Such concatenations may set the system on a trajectory of change that would not have occurred if the system had not been disturbed or if these disturbances had occurred independently. Also, similar outcomes may arise from different processes and causes (the principle of convergence or equifinality; see Chapter 2). Geologic and climatic factors determine the environmental setting and the nature of disturbance events to which rivers are subjected. They set the imposed and flux boundary conditions that fashion the erodibility and erosivity of a landscape, and the resulting character, behaviour and pattern of river types. Stark contrasts can be drawn, for example, between a dry, low-relief landscape with negligible vegetation cover and a high-precipitation mountainous terrain with dense forest cover. Formative processes, rates of activity (magnitude–frequency relations) and evolutionary trajectories vary markedly in these differing settings. Hence, any consideration of river evolution must be framed in relation to these geologic and climatic controls. In this chapter these considerations are appraised for differing tectonic settings and morphoclimatic regions. Particular emphasis is placed upon how landscape setting influences the imposed boundary conditions (especially slope and valley width) that constrain the range of behaviour of rivers, and the flux boundary conditions (i.e. flow and sediment regimes) that determine the mix of erosional and depositional processes along any given reach. Critically, as noted from the Lane balance diagram, alteration to either the imposed or flux boundary conditions promotes evolutionary adjustments. Geologic factors set and alter the imposed boundary conditions under which rivers operate, through their influence on lithology, relief, slope, valley morphology and erosivity and/or erodibility of a landscape. For example, tectonic activity or volcanic events may disrupt the nature and configuration of a landscape. Climate considerations play two critical roles. First, they are key determinants of the type and effectiveness of geomorphic processes (flow and sediment interactions) that shape landscapes at any given place. Second, climatic factors mediate the role of ground cover, which affects hydrologic processes and landscape responses to geomorphic processes through its influence upon surface roughness and resistance. Alterations to flux boundary conditions drive adjustments to the flow–sediment balance, prospectively modifying the evolutionary trajectory of a system. Evolutionary adjustment may take a mere moment in time (e.g. river responses to a volcanic eruption) or be lagged some time after a disturbance event. Elsewhere, landscapes may be stable or demonstrate progressive adjustment over time. Some rivers are adjusted to high coefficients of discharge variability (see Chapter 4), such that large floods are rare but not unusual – they are part of the 'formative process regime' for that particular setting. Other rivers are adjusted to smaller, more recurrent events. Many rivers flow on surfaces created by past events, or are still adjusting to past flow and sediment regimes. In these cases, geomorphic memory continues to exert a significant influence upon contemporary forms and the nature and effectiveness of processes. Understanding how contemporary processes relate to historical influences is a key challenge in efforts to read the landscape. This chapter is structured as follows. First, timescales of river change are discussed. Second, pathways and rates of geomorphic evolution are summarised for different types of rivers. Third, geologic and climatic controls on river evolution are considered. Then, evolutionary responses to changes in boundary conditions are outlined, and the river evolution diagram presented in Chapter 11 is used to extend analysis of river behaviour to incorporate interpretations of the nature and capacity for river change for various types for rivers. Finally, tools to interpret river evolution by reading the landscape are reviewed. Timescales of river adjustment Timescale of river adjustment varies from place to place, dependent upon the range of adjustment of the system (its sensitivity/resilience), the range and sequence of disturbance events and the legacy of past impacts. Both sensitive and resilient systems are prone to disturbance – responses are more likely and/or recurrent in the former relative to the latter. Analysis of river evolution frames system responses to disturbance events in relation to adjustments over geologic and geomorphic time (see Chapter 2). Geologic controls set the imposed boundary conditions within which rivers operate. Over timeframes of millions of years, tectonic setting exerts a primary control upon topography, determining slope and valley settings that influence river morphology and behaviour. Over geomorphic time, rivers adjust to climatically fashioned flux boundary conditions (flow variability, sediment availability and vegetation cover) over hundreds or thousands of years. Any disruption to flux boundary conditions may affect the evolutionary trajectory of a river. The key consideration here is whether the reach is able to accommodate adjustments while it continues to operate as the same type of river (i.e. it operates within its behavioural regime) or whether these altered conditions bring about a transition in process–form relationships (i.e. river change occurs). As noted in Chapter 2, river responses to disturbance events range from gradualist (uniformitarian) adjustments through to catastrophic change. A continuum of responses to disturbance events may be discerned: • No response may be detected, as systems absorb the impacts of disturbance. Stable rivers can tolerate considerable variation in controlling factors and forcing processes. For example, gorges are resilient to adjustment or change. Alluvial systems with inherent resilience induced by the cohesive nature of valley floor deposits, or the mediating influence of riparian vegetation and wood, may demonstrate limited adjustment over thousands of years. In these cases, responses to disturbance events are short-lived or intransitive, and change does not occur. • Part of progressive change. Rivers may respond rapidly at first after disruption, but in a uniform direction thereafter, such that change occurs gradually over a long period. For example, progressive denudation results in gradual reduction of relief over time, as gravitationally induced processes transfer sediments from source to sink. This results in long-term changes to slope and, hence, river type. Progressive adjustments are often observed following ramp or pulse disturbance events, so long as threshold conditions are not breached. • Change may be instantaneous as breaching of intrinsic or extrinsic threshold conditions prompts the transition to a new state or even a new type of river. These effects tend to be long lasting or persistent. • Change may be lagged. Off-site impacts of major disturbances may induce a lagged response in downstream reaches (e.g. conveyance of a sediment slug). The subsequent history of disturbance events affects the nature/ rate of response and prospects for recovery. Efforts to read the landscape seek to unravel variability in forms, rates and consequences of adjustments within any given system over differing timeframes. Pathways and rates of adjustment and evolution vary markedly for different types of rivers. Pathways and rates of river evolution Evolutionary pathways and rates of adjustment of rivers vary in differing geologic and climatic settings. This reflects differing ways in which boundary conditions and disturbance events affect flow–sediment interactions along a river. Alternatively, disturbance events may affect the surfaces upon which these processes are acting (e.g. role of fire, human disturbance – see Chapter 13). Evolutionary adjustments are likely to be most marked for those systems that have the greatest capacity to adjust and change. Hence, the nature and rate of evolution tend to be most pronounced in freely adjusting alluvial settings. These rivers have the greatest range in their degrees of freedom, such that pronounced disturbance events may trigger adjustments in channel planform, channel geometry (bed and bank processes), assemblages of channel and floodplain geomorphic units, and bed material organisation. The mix of water, sediment and vegetation conditions, as such, influences likely pathways of river adjustment for rivers in differing settings. Characteristic examples of evolutionary pathways are presented for rivers in differing valley settings below. Likely evolutionary pathways of rivers in confined valley settings Rivers in confined valley settings have limited capacity for adjustment. Their morphologies are largely imposed and are comprised largely of an array of imposed (bedrock) erosional forms. Steep headwater rivers progressively rework assemblages of slope-induced erosional geomorphic units as channels cut into bedrock via incisional processes over timeframes of thousands of years (Figure 12.1a). In contrast, gorges are stable and resilient systems over timeframes of hundreds or thousands of years. However, progressive incision and lateral valley expansion eventually create space along the valley floor for floodplain pockets to develop in partly confined valleys (Figure 12.1a). These transitions reflect changes to imposed boundary conditions. Likely evolutionary pathways of rivers in partly confined valley settings Just as gorges progressively widen to partly confined valleys with bedrock-controlled floodplain pockets over thousands of years, so sustained widening of these valleys eventually promotes the transition to partly confined valleys with planform-controlled floodplain pockets (Figure 12.1a). Increased valley width and reduced valley floor slope or changes in material texture, in turn, may result in a transition in the type of planform-controlled floodplain pockets that are observed, say from a low-sinuosity variant to a meandering planform variant (Figure 12.1a). Likely evolutionary pathways of rivers in laterally unconfined valley settings Variants of channel planform were conveyed along a continuum in Chapter 10. Adjustments to flow–sediment relations (i.e. flux boundary conditions) may bring about a transition to adjacent types of rivers along this continuum. Reduced energy conditions induced by lower flow and/or sediment availability may transform a braided river into a wandering gravel-bed river, and vice versa (Figure 12.1b). In turn, reduced energy conditions induced by lower flow and/or sediment availability may transform a wandering gravel-bed river into an active meandering river, and vice versa (Figure 12.1b). Alternatively, increase in sediment load (bedload fraction) may transform a passive meandering (suspended-load) river into an active meandering (mixed-load) river, and vice versa (Figure 12.1b). Various stages of evolutionary adjustments may be discerned along a discontinuous watercourse, reflecting cut and fill phases (Figure 12.1c). However, should certain circumstances eventuate, the river may maintain a continuous watercourse. The examples outlined in Figure 12.1 convey progressive evolutionary adjustments. In essence, the types of rivers that are found in an adjacent position along the longitudinal profile (i.e. an energy gradient) are likely to present the next step or phase in the evolutionary adjustment of a river. This may reflect conditions of decreasing energy associated with progressive landscape denudation, or increasing energy associated with uplift (i.e. steeper slope conditions). This line of reasoning, whereby juxtaposed river types along slope-induced environmental gradients provide guidance into likely evolutionary adjustments, is a direct parallel to Walther's law of the correlation of facies (Chapter 6): adjacent sedimentary deposits in contemporary landscapes are used to guide inferences into stacked depositional units within basin fills. Geologic and climatic controls are the primary determinants of imposed and flux boundary conditions, and the associated suites of disturbance events to which rivers are subjected. Although these geologic and climatic considerations act in tandem, they are considered separately below for simplicity. Geologic controls upon river evolution Geologic setting determines the imposed boundary conditions within which rivers adjust and evolve. The nature and movement of tectonic plates is a primary determinant of the distribution and relief of terrestrial and oceanic surfaces. The nature and position of mountain belts and depositional basins is determined largely by the distribution of plates and geologic processes that occur at different types of plate boundaries. Landscape relief and topography are fashioned by the balance of endogenetic processes (i.e. geologic processes that are internal to the Earth) and exogenetic processes (i.e. geomorphic processes that erode and deposit materials at the Earth's surface). The nature, frequency and consequences of geologic disruption and disturbance events vary markedly in different tectonic settings. This is determined largely by position relative to a plate margin and the nature of tectonic activity at that margin. Vertical and lateral displacement along fault-lines is common in some settings. Contorted strata of folded rocks attest to the incredible forces at play. Faulting, folding and tilting generate distinctive topographic controls upon slope, valley morphology and drainage patterns (Chapter 3). Volcanic activities and subsidence modify relief and availability of materials. Tectonic setting frames the long-term landscape and dynamic context of river systems. Simplified schematic representations of primary tectonic settings are presented in Figure 12.2. Tectonic motion generates three primary types of plate boundaries: collisional (constructional) (Figure 12.2a and b), pull-apart (Figure 12.2c and d) and lateral displacement (Figure 12.2e). The rate of uplift, subsidence or lateral movement of a plate affects the relative stability and nature of landscape adjustment. Uplifting rivers incise into their beds, creating narrow valleys. Subsiding rivers aggrade, creating expansive floodplains. Rivers in lowrelief, plate-centre settings are characterised by long-term stability, as they slowly denude and rework the landscape (Figure 12.2f). Long-term changes to the position of plate boundaries affect the nature of constructional and reworking processes at any given locality (Figure 12.3). These geological foundations determine patterns of lithological and structural variability, affecting the erodibility and erosivity of landscapes. The convergence of continental plates generates major mountain chains. Deeply incised bedrock channels in headwater settings contrast starkly with transport-limited braided rivers, low-relief rivers atop uplifted plateau landscapes or deeply incised gorges at plateau margins (Figure 12.4a–c respectively). Uplift of supply-limited plateau landscapes may create deeply entrenched, superimposed drainage networks. For example, Figure 12.4d shows the planform of a meandering river that previously formed on a relatively flat alluvial plain that has been retained as the landscape was uplifted, creating a deeply etched, bedrockcontrolled gorge in which river character and behaviour are imposed. Differing forms of constructional landscapes are generated through subduction of dense but relatively thin oceanic plate beneath a continental plate. Recurrent phases of tectonic activity produce basin and range topography comprised of mountain ranges, volcanic chains and intervening basins, exerting a dominant imprint upon contemporary drainage networks. The imprint of landscape setting upon river character, behaviour and evolution is clearly evident in pull-apart basins. This tectonic setting is characterised by striking alignment of lakes and straight, bedrock-controlled river systems. In some instances, basins that pulled apart in the past may retain a dominant imprint upon contemporary landscapes, forming escarpments (Figure 12.4e) and rift valleys (Figure 12.4f). Alternatively, lack of tectonic activity is a primary determinant of river processes and forms in plate-centre landscapes. These lowrelief, low-erosion settings often have profound stability and antiquity (Figure 12.4g). Long-term changes to plate tectonic boundaries ensure that any given landscape setting has likely been subjected to differing forms and phases of tectonic activity (Figure 12.3). Geologic adjustments are sometimes imprinted atop each other. Elsewhere, the imprint of past events has been virtually erased, though metamorphosis of rocks may provide insights into former conditions. Importantly, tectonic setting not only fashions the relief and erodibility of a landscape, it also affects the climate and, hence, the erosivity of that landscape (Fryirs & Brierley, 2013).

References:

Fryirs, K. A., & Brierley, G. J. (2013). *Geomorphic Analysis of River Systems: An Approach to Reading the Landscape.* West Sussex: Wiley- Blackwell.

UNIT-11: CONSTRUCTION OF DAMS AND BARRAGES AND THEIR IMPACT ON THE FLUVIAL SYSTEM

Since ancient times dams have been constructed on streams and reservoirs have been formed. Generally dams and reservoirs serve many purposes such as flood control, power generation, supplying water for irrigation, drinking and industrial use, navigation and recreation. Since independence, India has witnessed rapid growth in the construction of large dams and elaborate canal networks. Over 4000 projects have been constructed in last five decades and 700 projects have been proposed to meet increased demand for power and achieve larger irrigation potential (Raghuvanshi et al. 2000). This activity has caused several social, ecological and economic problems. As a result of completion of such water resources projects, the society is greatly benefited in terms of dependable and clean drinking water, greater availability of food, better health, sanitation and increased per capita income; availability of increased power has also resulted in greater industrial activity and better living standards. Tourism and recreational facilities created by water resources projects have led to social and cultural improvements e.g., Brindavan gardens, Ramganga garden, Kalinadi Kunj, Jaikwadi garden and Gobindsagar reservoir. (Goel and Agarwal, 2000). Flocking of rare species of birds and increase in wild life have also been reported near Ramganga, Rihand and Matatila reservoirs. As against these beneficial effects, a number of adverse effects have also been reported. Formation of reservoirs due to construction of dams submerges large areas including those of forests and a number of people are ousted from submerged areas. Table 14.1 gives some data on submergence, ousted number of people and installed power capacity. Thus, it can be seen that as a result of construction of these dams, large areas including forests have been submerged and ousted a large number of people from their homes. Resettlement and compensation to inhabitants of the submerged areas include determination of areas that will be submerged, evaluating compensation for their properties, selection of alternative sites for settlement and distribution of land. Such considerations were not made in the case of the Volga lake in Ghana which submerged about 8200 km2 area. The submergence area was greatly underestimated. Similarly, in the case of the Roseires reservoir on the Blue Nile river in Sudan, only a few months before submergence people were asked to move and submergence was underestimated by two metres (Murthy 1976). Submergence of forest areas affects the habitat of many wild life species as can be seen from Table–14.2.

Two other environmental effects of construction of large dams for water storage and its utilization are water logging and salinization, and water-borne diseases. About 2 to 3 million ha of land every year is going out of production due to salinity problems. Water logging results primarily from inadequate drainage and over irrigation, and to a lesser extent, from seepage from canals and ditches. Water logging concentrates salts, drawn up from lower portions of the soil in the plant's rooting zone. The build up of sodium in the soil is particularly detrimental form of salinization which is difficult to rectify. The irrigation-induced salinity can arise as a result of use of any irrigation water, irrigation of saline soils, and rising levels of saline ground water combined with inadequate leaching. Water-borne or water related diseases are commonly associated with the introduction of irrigation. The diseases most directly linked with irrigation are malaria, bilharzias, filaria, cholera, gastroenteritis, viral encephalitis and goitre. Other irrigation related health risks include those associated with increased use of fertilizers, herbicides and pesticides. The occurrence of these diseases in the population have been noticed in Ghana, Nigeria, Egypt, Ethiopia and other countries. Large versus Small Dams Construction of large dams and reservoirs usually entails enormous costs and displacement of a large number of people as can be seen from Table 14.1. Resettlement of these oustees is many times neglected and hence leads often to protests, hunger strikes, stoppage of work and costly litigations that further delay the work. Unfortunately, people who are likely to be affected are not consulted or taken into confidence. Further, we do not have a National Rehabilitation Policy and credible implementing and monitoring procedures for rehabilitation. There is always a complaint that these people do not get fair compensation and a guaranteed share in the prosperity that the project brings. These conditions need immediate improvement. Considering these effects, many environmentalists argue that instead of building one large dam, a few small dams should be built. However, this idea is not feasible for the following reasons (Indiresan 2000). i) A number of small dams cannot control floods or generate electricity as a high dam can. ii) Per 1000 m3 of storage, the capital cost for large, medium and small dams varies in the ratio of approximately 1:3:6; hence it will be costlier to build smaller dams than a single large dam for achieving the same storage. iii) Other things being equal, doubling height of dam increases the storage by about eight times and power potential sixteen times; hence it is better to build large dams when feasible. iv) Since rainfall in India is erratic and occurs in 3 or 4 months, water needs to be stored to meet irrigation, water supply and power needs especially when drought occurs. This is unlike in Europe where precipitation occurs all through. Hence large dams are needed. v) Evaporation loss in India is about 1.2 to 1.4 m annually; hence, storage required has to be large. Further, it has to be realized that providing food, drinking water and power to millions of people is more important than preventing displacement of a few thousand people. This is not to say that the legitimate needs and aspirations of the oustees should be overlooked. Similarly, legitimate actions have to be taken to protect the environment. In the three gorges project about one third of the investment has been set aside for rehabilitation and environmental protection. Considering all these aspects it may prudent to have good mix of large and small dams for the development of water resources in the country. Reservoir Induced Seismicity It has been found all over the world that in some cases, after the reservoir is filled, the adjacent areas are subjected to reservoir-induced earthquakes (Kolhi and Bhandari 1991, Gupta 1992). Such reservoir– induced earthquakes have occurred after impoundment of the Shivajisagar lake formed by Koyna dam and at Bhatsa dam in Maharashtra, and at Sriramsagar dam on the Godavari in India. Such earthquakes have also occurred at Hsinfengkian reservoir in China, at lake Mead formed by Hoover dam on the Colorado river in U.S.A., Nurek and Tokgotul reservoirs in Russia, Aswan dam in Egypt and at many other places. A few details about reservoir-induced earthquakes at Koyna dam can be given. The Koyna dam of height 103 m and the Shivajisagar reservoir are located in peninsular India about 200 km from Pune.

Soon after impoundment of the reservoir in 1962, the nearby area started experiencing earth tremors and the frequency of these tremors increased from the middle of 1963 onwards. These tremors were accompanied by sounds similar to those of blasting. Between 1963 and 1967, five earthquakes occurred which were strong enough to be recorded by many seismological observatories in India. The major earthquake at Koyna occurred on December 10, 1967, which had a focal depth of 10 km ± 2 km and had a magnitude of 6.0. This earthquake claimed about 200 lives, injured over 1500 people and rendered thousands homeless. It also caused damage to hoist tower of the dam and developed horizontal cracks on both the upstream and downstream faces of a number of monoliths, and damaged a large number of houses, bridges and culverts. Realising the socio-economic importance of reservoir induced seismicity, UNESCO formed a working group on these phenomena and since then a number of symposia on reservoir-induced seismicity have been organized. A number of theories/explanations have been suggested to explain why and under what conditions the seismicity is caused. Investigation of fluid injection-induced earthquakes at the Rocky Mountain Arsenal near Denver, Colorado, (U.S.A.) during 1960's and Evan's work on the mechanism of triggering earthquakes by increase of fluid pressure have helped in understanding the phenomenon of reservoir-induced seismicity. Gough and Gough have explained triggering of earthquakes due to incremental stress caused by water load in the reservoir. Gupta et al. (see Gupta 1992) identified the rate of increase of water level, duration of loading, maximum levels reached and the duration of retention of high water levels among the important factors affecting the frequency and magnitude of reservoir-induced earthquakes. Other studies by Nyland, and Bell and Nur have also indicated that the three main effects of reservoir loading relevant to inducing earthquakes are (i) the elastic stress increase that follows the filling of the reservoir; (ii) the increase in pore fluid pressure in saturated rocks due to decrease in pore volume caused by compaction in response to elastic stress increase; (iii) and pore pressure changes related to fluid migration. It is also found that reservoir-induced earthquakes are associated with shear fracturing of rocks. The shear strength of rocks is related to the ratio of the shear stress along the fault to the normal effective stress across the fault, the latter being equal to normal stress minus the pore pressure. Hence, increase in pore pressure can trigger earthquake if rocks are under initial shear stress. During the past four decades, scientists have gained some knowledge about RIS but a lot more needs to be learned. It may be mentioned that the largest reservoirimpoundment triggered earthquakes have exceeded magnitude of six. On the basis of RIS observations on a number of dams, it has been well established that major RIS events are produced by enhanced foreshock activity. Such analysis has indicated that, if two earthquakes of magnitude greater than 4 occur at RIS site within a short interval of say 2-3 weeks, there is an enhanced probability for occurrence of earthquake of magnitude greater than 5. Studies have suggested that in the case of a large reservoir (volume in excess of 1000 Mm3 usually impounded behind a dam height greater than 100 m) it is desirable to carry out geological mapping for the entire reservoir area to determine faults and competence of rocks (Garde, 2006).

References:

Garde, R. J. (2006). *River Morphology.* New Delhi: New Age Internal(P) Limited Publishers.

UNIT-12: STREAM CORRIDOR: STRATEGIES FOR MANAGEMENT

Stream restoration and mitigation is a process that involves recognizing natural and human induced disturbances that degrade the form and function of the stream and riparian ecosystems or prevent its recovery to a sustainable condition. Restoration includes a number of activities designed to enable stream corridors to recovers dynamic equilibrium and function to maintain channel dimensions, pattern and profile so that over a period of time the stream channel does not degrade or aggrade. FISRWG (1998) identifies three levels of stream improvement: (a) restoration (b) rehabilitation and (c) reclamation. Restoration is defined as the establishment of the structure and function of ecosystems. Ecological restoration involves returning an ecosystem as closely as possible to the pre-disturbance conditions and function. Restoration also implies that it will provide the highest level of aquatic and biological diversity possible. The basic principles of stream restoration include:

1. analysis of channel history and evolution; 2. analysis of cause and effect of change; 3. analysis of current condition; 4. development of specific restoration goals and objectives prior to design; 5. holistic approach to account for channel process, riparian and aquatic function; 6. consideration of passive practices such as fencing against livestock; 7. natural channel design to restore function. Rehabilitation is defined as a procedure for making the land useful again after a disturbance. It involves the recovery of ecosystem functions and processes in a degrading habitat. Rehabilitation establishes geological and hydrologically stable landscapes that support biological diversity. Reclamation is defined as a series of activities intended to change the function of an ecosystem, such as changing wetland to farmland. Restoration principles, practices and methods of monitoring are being evolved on the basis of studies on small and medium sized streams in some western countries such as U.S.A. and U.K. The structures used in stream restoration include vegetation, wood, and constructed rock and wood structures. In U.K. (see Brookes 1995) river restoration project (RRP) was formed to promote restoration of rivers for conservation, recreation and amenity. The project utilizes the expertise of river ecologists, engineers, planners, fisheries biologists and geomorphologists to establish demonstration projects to show how restoration techniques can be utilized to recreate natural ecosystem in damaged river corridors (Brookes 1995). Research needs to be carried out to study their effectiveness in degrading and aggrading streams, and to extend the methods to larger streams (Charlton, 2008).

References:

Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge.

SELF ASSESSMENT TEST:

What is channel morphology?

What processes shape channel morphology?

How does sinuosity relate to channel morphology?

Which human activity can alter channel morphology?

Study Tips:

- Charlton, R., 2016. Fundamentals of Fluvial Geomorphology, 2nd edition, Routledge, London. Garde, R.J., 2006. River Morphology, New Age International Ltd.
- Knigton, D., 1998. Fluvial forms and Processes: A New Perspective, Arnold, London.
- Leopold, L.B., Wolman, M.G. and Miller, J. P.,1964, 1992. Fluvial Processes in Geomorphology, Dover Publications, Inc. New York. Maiti, R., 2016. Modern Approaches to fluvial geomorphology, Primus Books.
- Morisawa, M., 1985. Rivers, Longman, London.
- Schumm et al.,1987. Introduction to Fluvial Processes, Wiley–Blackwell.
- Schumm, S. A., 2003. The Fluvial System, The Blackburn Press.

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