Two Years

Post Graduate Degree Programme (CBCS) in Geography

Semester – IV

Paper Code: GEO/DSE/FG/T-420

PAPER: FLUVIAL GEOMORPHOLOGY-III: ANALYSIS OF DRAINAGE BASIN AND CHANNEL (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-420

Paper: FLUVIAL GEOMORPHOLOGY-III: ANALYSIS OF DRAINAGE BASIN AND CHANNEL (SPECIAL PAPER)

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

Unit-01	Fluvial hydrosystem approach: drainage basin perspective
Unit-02	Drainage basin as a unit of geomorphic study
Unit-03	Hypsometric analysis of drainage basin and its significance
Unit-04	Classification of channel links, Shreve's formula of link counting
Unit-05	Computation of topologically distinct channel network and topologically integrated channel network
Unit-06	Quantitative analysis of channel planforms and indices
Unit-07	Meander indices: shape, form, and tightness
Unit-08	Channel bed topography: identification and analysis of channel geomorphic units
Unit-09	Efficiency of channel cross-section: concept and characteristics
Unit-10	Hydrograph analysis and computation of unit hydrographs
Unit-11	Textural analysis of river sediments and pebbles
Unit-12	Vulnerability analysis of floods and riverbank erosion

Contents

INTRODUCTION:

This course builds upon fundamental principles of fluvial geomorphology to delve deeper into the complexities of drainage basin and channel analysis. By the end of this paper, students will gain a comprehensive understanding of the processes shaping river basins and channels, along with the tools required to analyze their morphology and dynamics.

LEARNING OBJECTIVES

- Understand the fundamental concepts of drainage basin morphometry and channel characteristics.
- Learn how to apply quantitative techniques to analyze drainage basins and channels.
- Explore the interactions between climate, tectonics, and surface processes in shaping fluvial landscapes.
- Gain proficiency in using GIS and remote sensing tools for drainage basin analysis.
- Evaluate the impact of human activities on fluvial systems and their implications for management and conservation.

ASSESSMENT OF PRIOR KNOWLEDGE

Students are expected to have a foundational understanding of fluvial processes and geomorphological principles gained from previous courses in physical geography and geomorphology. Familiarity with basic statistical methods, GIS software, and remote sensing techniques will be beneficial for engaging with the course material effectively.

LEARNING ACTIVITIES

- Lectures: Engage in interactive lectures covering advanced topics in drainage basin and channel analysis, including theoretical concepts and practical applications.
- Practical Sessions: Participate in hands-on exercises using GIS software and datasets to analyze real-world drainage basin characteristics and channel morphology.
- Field Trips (where possible): Undertake field trips to observe and analyze drainage basins and channels in their natural settings, integrating theoretical knowledge with field observations.
- Research Assignments: Conduct research projects on specific aspects of drainage basin analysis, presenting findings through reports or presentations.
- Group Discussions: Collaborate with peers in discussions to critically evaluate case studies and research articles related to fluvial geomorphology.

FEEDBACK OF LEARNING ACTIVITIES

- Continuous feedback will be provided through:
- In-class discussions and Q&A sessions to address queries and assess understanding.

- Regular assessments including quizzes, assignments, and presentations to gauge progress and identify areas for improvement.
- Individual feedback on research projects and practical exercises to guide learning and enhance analytical skills.

UNIT-1: FLUVIAL HYDROSYSTEM APPROACH: DRAINAGE BASIN PERSPECTIVE

Fluvial hydrosystems are the product of physical, chemical and biological processes operating throughout a river's drainage basin and over a range of time-scales from a year to tens of thousands of years. A drainage basin is the area that gathers water from precipitation and delivers it to the river (Figure 1.2). Defined by a topographic divide, the basin is occupied by a drainage network which collects the runoff from hillslopes, together with its load of sediment, particulate organic matter and solutes. Thus, a river may be seen as the artery of a drainage basin conveying water, minerals and organic matter to the sea. A drainage basin perspective is also important because the flow regime and sediment loads determine the morphology of the channel which has a strong influence on the structure and function of fluvial hydrosystems as first recognized by Hynes (1970, 1975). However, drainage basins are complex geomorphological systems with a history. This chapter describes the characteristics of drainage basins and examines the ways in which the basin influences fluvial hydrosystems over a range of time-scales.

THE RIVER BASIN AS THE BASIC UNIT

Horton (1945) first established the erosional drainage basin as the basic unit of landscape. The network of channels within a drainage basin focuses the delivery of water and material in transport from the hillslopes onto the main river and then downstream to the river mouth. As a result, rivers became viewed as dominated by longitudinal processes. 'Drainage area' and 'distance from source' have been used as scale variables for studies of river systems, but measures of drainage network composition have particular significance. The number of headwater, finger-tip, tributaries (the stream magnitude, Shreve, 1967) and the number of confluences (the stream frequency, NERC, 1975) in a basin relate closely to streamflow characteristics, but most ecological studies have employed stream ordering techniques as an objective, and widely applicable, classification system. The most commonly used system is that proposed by Strahler (1952) modified after Horton (1945). This convention (Figure 2.1a) designates all headwater streams, terminated by the first confluence, as first-order streams. Two first-order streams join to form a secondorder stream, the confluence of two second-order streams produce a third-order river, and so on.

THE BASIN SYSTEM

For any point along a river, stream ordering provides an index of scale and a measure of position within the drainage network. However, an analysis of the characteristics of a river at any point along its course requires a more dynamic perspective. To achieve this perspective a systems approach (Chorley and Kennedy, 1971) may be employed. The drainage basin comprises a set of structural units - landforms - which function both as storages and sites of material and energy conversion and transformation. Together, a set of structural units can be described as a morphologic system and the strength and direction of the connectivity between the units can be revealed by correlation analysis. In a short reach of river, for example (Figure 2.2a), the morphological properties might include such parameters as local slope, bed material size and cross-sectional form (depth:width ratio). The interrelationships between the parameters often indicate the degree to which their dynamic properties are related. Thus, an increase in bed material size is associated with an increase in local slope and a reduction of the depth:width ratio. The different structural units are linked by a range of processes. Drainage basins are open systems dependent on inputs, transfers and outputs of mass and energy. The processes - the mass and energy flows through the basin - that link the different structural units can be seen as a cascading system. Within the drainage basin, the most important of these systems is the hydrological cascade (Figure 2.1b). Solar energy and precipitation form the inputs to the basin hydrological cascade and the outputs, streamflow and evapotranspiration losses, are generated by the transfer of water through a sequence of storages within the hillslope and channel subsystems. The most important regulator is the soil store. The hydrological cascade not only describes the amount of rain that becomes runoff and the rate of runoff, but also the quality of runoff, that is, the concentration of ions, released from weathering and biological processes, and particulate material (organic and inorganic). A complete explanation of a fluvial hydrosystem requires integration of the morpholOgic and cascading systems, known as a process-response system. Thus, in Figure 2.2(b) channel morphology is seen to be determined by the discharge of water (Q) and sediment (L). These systems rarely attain exact equilibria and generally the river channel tends toward a mean form, defined as quasi-equilibrium. With a system itt equilibrium, the outputs will equal the inputs. If the inputs change, the morphologic system will adjust to a new equilibrium condition. For example, an increase in sediment load, with no change of discharge, will induce an increase in channel slope and bed material size, and a reduction of the depth:width ratio. The drainage basin of large rivers comprises a large number of small basins and Schumm (1977) divided the drainage basin system into three distinct parts (Figure 2.1a) representing different types of processresponse system. Small headwater catchments represent the production zone where runoff generation and sediment yields are determined mainly by hillslope processes. Strong lateral linkages between hillslope and channel make this zone, which can extend to include fourth-order streams, distinct. Water, with its particulate and dissolved load, is then transferred downstream through the channel network (the transfer zone) to the major alluvial river which is characterized by its storage properties. Within the storage zone the flow regime is strongly influenced by water retention on the floodplain during high flows and within the alluvial aquifer during low flows; sediments and organic matter are stored within the floodplain and nutrient cycling is highly dependent on interactions between the alluvial aquifer, the floodplain and the main river. These processes are examined in later chapters. The fluvial hydrosystem of each river sector may be described as a process-response system with point inputs (from the upstream basin) and outputs. Within each sector strong functional relationships exist between the physical, chemical and biotic components such that a quasiequilibrium state may be defined over time-scales of 10-100 years. An important characteristic of process-response systems is the role of feedback loops, particularly negative ones. The operation of these loops introduces processes which tend to oppose continued change within the system caused by variation of the cascade input. For example, the sediment load is a function of both hillslope and channel erosion at the base of the slope, increasing slope angle. The increase in sediment supply reduces channel erosion and may induce aggradation. This stabilizes the slope subsystem and reduces hillslope sediment supply. Selfregulation operates to maintain the quasi-equilibrium and to establish a new quasi-equilibrium condition if a change of the external controls (here effective precipitation or baselevel) occurs. In reality, the controls within each loop are only partial. Change is associated with many loops linking the variables involved, so that responses to changes of the external controls are usually complex, involving interactions between the different tributaries and interrelated sectors (Figure 2.2c). Within each sector of channel the morphological dynamics reflect the sediment supply from the sector immediately upstream (comprising sediment transported through, and eroded from, this sector) and the influence of local baselevel changes determined by the next downstream sector.

INFLUENCE OF SCALE

Rivers can be viewed in many different ways and not only as systems in quasi-equilibrium over an historical time-scale (10-100 years). Over short time-scales, from a day to one year, they may be seen as rather stable systems. Over geological time-scales they are viewed as systems undergoing continual change, known as a state of dynamic equilibrium. The importance of these three different perspectives, first defined by Schumm and Lichty (1965), is in the definition of cause and effect because the status of variables and the nature of their interrelationships vary according to the scale adopted. The status of variables influencing the hydrogeomorphological characteristics of rivers over different time-scales is given in Table 2.1. At a single point in time, the spatial variation of flow velocities and depths (variable to) within a short reach are determined by channel morphology (variable 9). Similarly, day to day variations of river flow result in variations of hydraulic parameters as Ihe discharge adjusts within a cross-section. Over such short time-scales channel morphology is determined and independent but over the historical time-scale, channel forms are viewed as dependent on the mean discharge of water and sediment (variable 8). The channel form is considered to be in equilibrium with the discharges and sediment loads produced by the interaction of climate (variable 6) and vegetation (variable 7) within the drainage basin upstream. However, the equilibrium channel dimensions will also reflect the valley characteristics (variable 5) which have developed over the longer (geological) time-scale. The valley gradient and the sedimentology of valley fill deposits, inherited from palaeoclimatological conditions (variable 3), are particularly important.

In the same way, the status of variables changes with spatial scale. Between-basin differences are explained by variations in climate, vegetation and relief, or by comparison of hydrological regimes, water balances or water-quality which have particular significance for fluvial hydrosystems. These are discussed in the next section. Differences between sectors reflect the declining gradient and reducing altitude; progressive changes occur in discharge and channel morphology with increasing catchment area. Large rivers also tend to be characterized by more regular and predictable variations of the abiotic variables than headwater streams and this change is particularly important for the functioning of fluvial hydrosystems. However, important differences between sectors - between fluvial hydrosystems - also reflect the different histories of each sector. This will be discussed in Section 2.4.

SOURCE AREA CHARACTERISTICS

Hydrology, channel morphology and temperature form the principal components of the abiotic milieu of river ecosystems. Their complex interactions determine, to varying degrees, species distributions and abundances of biota. In low-order headwater streams the variation of river flows, water quality and sediment transport is closely related to hillslope processes. Discharges, water quality and sediment loads vary over a range of time-scales from hours to years according to the relative contributions from the different source areas within the headwater catchments. The source areas are defined by rock/soil type, vegetation/landuse, altitude and topographic setting; drainage basin characteristics fundamentally determined by geology, climate and the way the basin has evolved over time.

2.2.1 HYDROLOGICAL REGIMES

A river's flow regime may be defined as the seasonal variation in its discharge, usually described by a graph of monthly mean flow. Beckinsale (1969) described the global variation of hydrological regimes in relation to climate and relief, and noted that within large areas of the world the hydrological regime of rivers draining small and moderately sized basins closely reflects the regional climatic rhythm (Figure 2.3a). Five major types of regime were defined: tropical rainy climates having a mean temperature over 18°C in all months; dry climates with an excess of evaporation over precipitation; warm, temperate climates; seasonally cold, snowy climates with a mean temperature of the coldest month being below -3°e; and snow and ice environments of high mountains. Variations of each type have also been defined (Figure 2.3b), reflecting the magnitude and timing of seasonal variations of flow and temperature.

2.2.2 THE WATER BALANCE

Over a selected time period, the water balance of a drainage basin can be evaluated as: $P-(E+Q)=O$ where, P is precipitation; E is evaporation plus transpiration; and Q is stream discharge. In humid temperature regions, annual precipitation increases with altitude and proximity to the sea. Thus, across Europe annual precipitation ranges from above 2000 mm to below 600 mm. Actual evapotranspiration across most of humid temperate Europe is between 500 and 600 mm. As the scale of observation decreases, the physical characteristics of the drainage basin become increasingly important in explaining variations in catchment hydrology. In particular, variable losses by evapotranspiration occur as a result of land-use differences. Thus, in comparing small forested and grassland catchments in Wales with an annual precipitation of over 2000 mm, Calder and Newson (1979) found that in the forested catchments evapotranspiration losses were doubled. In deforested catchments, reduced interception storage capacity and evapotranspiration rates lead to an increase in the proportion of rainfall becoming runoff. In one example from the Hubbard Brook experimental catchment in New Hampshire, USA, in the first year after felling runoff increased by 40% (Hornbeck et al., 1970). Similarly, urban development can reduce evapotranspiration losses. Lvovitch and Chernogaeva (1977), for example, presented a water balance for Moscow suggesting that urban development reduced evapotranspiration by 62% (310 mm). 2.2.3 WATER QUALITY The dissolved content of rivers is dominated by solutes released by weathering processes and reflecting catchment geology. Dominant solutes are bicarbonate (HC03), sulphate (S04)' calcium (Ca) and silicate (Si02). The load-weighted average total dissolved solids content of world river water has been calculated as 120 mg 1-1 (Webb and Walling, 1992). However, total dissolved solids concentrations range over five orders of magnitude from less than 10 mg 1-1 in some Amazon tributaries to over 10 000 mg 1-1 in some arid areas. Thus, at the global scale, Gibbs (1970) emphasized the influence of climate and geology on the variability of river water quality, separating Ca-HC03 (rock-dominated) rivers from those dominated by Na-Cl with low (precipitation-dominated) or high (evaporation-dominated) conductivity. The influence of climate acting through high rainfall to dilute river water gives a typical inverse relationship between discharge-weighted solute concentration and mean annual runoff. Meybeck (1982) also described global variations of dissolved organic carbon concentrations with median values of 10 mg 1-1 for taiga rivers (in high latitude areas with marshy coniferous forest), 6 mg 1-1 for tropical rivers, and 3 mg 1-1 for temperate rivers. Although indicating a general climatic control, these values probably relate to the different organic contents of the soils in the different climatic regions. The role of geology and land use again becomes clearer at a smaller scale. In a review of water quality in British rivers (Walling and Webb, 1981) specific electrical conductance was found to vary from 35 to 1200 j.LS em-I in small and unpolluted catchments. Within one medium-sized basin, the River Exe in south-west England, conductivity under low flows ranged from less than 50 j.LS cm-1 to over 1000 J.L5 em-I being dominated by geological variability within the catchment: lowest levels are associated with resistant slates and grits and highest levels with the younger, and more easily weathered, marls (Walling and Webb, 1975).

2.3 LARGE BASINS

The drainage basin of a large river can be viewed as a nested hierarchy of basins of different size. With increasing distance downstream and increasing catchment area, the influence of a particular headwater basin is diminished, and the influence of particular geologies and land uses decline. For example, the hydrological regime of the lower Rhine integrates the runoff pattern from the Alpine region where 50% of the precipitation falls as snow, and where snowmelt produces a high summer flow, with runoff from the maritime climatic region which produces high flows in winter. In such large basins, flow and water quality reflect not only the mixing of water from the different headwater catchments but also the routing of flows through the channel network, and the influence of channel, floodplain and groundwater storage. In many ways the longitudinal 'continuum' is simply a reflection of increasing scale: increasing discharge, increasing channel size and the progressive downstream regulation of the spatial and temporal variations that characterize small basins. The different transfers may be expressed over a range of time-scales and incorporate important storages with different transport rates, retention times, and turnover rates. Consequently, the changing nature of the different storages along a river and the complex interactions of these storages are important factors in determining the functional dynamics of fluvial hydrosystems. There are many different storages, both physical and biological, but three brief examples illustrate their significance.

2.3.1 DEAD ZONES

At flows below bankfull, variations of channel shape and vegetation characteristics create hydraulic boundaries within the channel which separate the main flow from backwaters. Such 'dead-zones' with near zero flow velocity (Figure 2.4a) are separated from the main flow by localized lines of lateral shear. During storage, water-quality changes take place by interactions between water, sediment, detritus and biota. Tracer experiments undertaken on the lower Severn, UK (Bevan and Carling, 1992) show typical distributions (Figure 2.4a) which suggest the existence of large-scale storage areas - dead zones - within the reach. 2.3.2 FLOOD ROUTING The passage of water from the production zone (Figure 2.1a) to the mouth of the river during a flood involves lags over short time-scales. As discharge rises, water enters each successive reach of channel, part enters storage within that reach such that there is a net reduction in the rate of increase of discharge downstream. The resulting 'attenuation' of flood peaks is illustrated in Figure 2.4(b). Attenuation is at a minimum and nearly zero at about 375 m3 S-1 (at just above bankfull), and increases for lower discharges contained within the channel and also very considerably at higher, floodplain, flows (Archer, 1989). Through floodplain sectors, this attenuation is exacerbated by overbank storage and by losses through the channel bed into the alluvial aquifer. Thus, along a 220 km reach of the River Allier, France, losses to the alluvial aquifer during the 1968 flood reduced the flood volume by 13% (Dacharry, 1974).

2.3.3 SEDIMENT DELIVERY

The downstream transport of sediment is markedly influenced by storages operating over a range of temporal scales. Sediment yield to the oceans is but a small fraction of that eroded from hillslopes. That part of the gross erosion or sediment mobilization within a catchment represented by the sediment yield at the catchment outlet is defined by the sediment delivery ratio (Walling, 1983). Values of the delivery ratio below 10% are common and an example of a catchment sediment budget is shown in Figure 2.4(c). Some of the eroded material will be deposited within the production zone, further down the slope or at the foot of the slope. Woody debris jams, macrophyte beds, the roots and overhanging branches of riparian vegetation are important retention features for leaf litter and sediment storage. Although instream deposits are periodically flushed from the system following decomposition of debris jams and/or by high floods, debris jams can retain a substantial proportion of the organic matter in channels. Typical residence times vary from a few years to a few hundred years (Gregory, 1992). Depositional landforms represent particularly important storages: bars, fans, floodplains, terraces and deltas. There are many different sedimentary environments (Happ, 1971), some retaining large amounts of organic material, such as cutoff channels and backswamps. Some floodplain sediments together with their organic deposits can remain stable for tens of thousands of years. Throughout the temperate zone, for example, the long-term storage of thicknesses of sands, gravels, alluvium and organic material has resulted from valley aggradation during the period of Holocene sea-level rise (Gregory and Maizels, 1991).

2.4 BASIN HISTORY

Many rivers of the temperate zone have a history extending for millions of years, complicated by major climatic and tectonic changes. For example, the proto-Thames had established a concordant drainage pattern in response to mid-Tertiary orogenic movements by about 20 million years ago. Its present lower course was established during the early Pleistocene when it was diverted southwards by a complex sequence of ice advances and retreats. Major efforts have been made to elucidate the response of fluvial systems to climatic change over the past 15000 years (Gregory, 1983; Gregory et ai., 1987; Starkel et ai., 1991). At the change from cold to warm climate at the beginning of the Holocene the general trend of middle river courses throughout the temperate zone has been from braided to meandering channel patterns. This change, manifested by a reduction in channel width and an increase in sinuosity, was caused by a decrease in flood frequency and sediment load following the establishment of the temperate forest. Many rivers today are flowing directly within coarsegrained gravels that have been inherited from deglaciation outwash and valley-fill deposits. However, these general trends mask a complexity of responses, both along and between rivers. 2.4.1

DRAINAGE MODIFICATION DURING THE PLEISTOCENE

Throughout much of the temperate zone, drainage patterns were modified by glaciation during the Pleistocene by: 1. Diversion of drainage lines, a characteristic of the former glacial zone, by which many rivers had their courses markedly altered by ice advances during the Pleistocene; 2. Intensification of the pre-existing drainage by valley erosion, sometimes forming flat-floored and steep-sided glacial troughs; 3. Deposition, especially by ice-sheets, which resulted in till sheets mantling large areas of low-lying landscapes upon which new stream networks have formed; 4. Glacio-isostasy, which describes the deformation of the earth's crust beneath the load of ice. Given the average densities of ice and rock a 1000 m thick ice sheet would depress the earth's crust by 267 m; 5. Glacio-eustasy, which describes the influence on sea-level of water periodically stored in, and discharged from, ice sheets. Thus, differences between rivers and between sectors along rivers may reflect their situation during the Pleistocene. Four general situations can be defined: 1. Rivers draining young valleys formed after the retreat of the ice sheet and characterized by complex valley profiles including lakes and bedrock sections; 2. Rivers draining areas of former valley glaciers; 3. River valleys dominated by former glacial meltwaters, being characterized by rivers flowing in obviously much larger valleys (known as underfit streams) and often with extensive dry-valley networks, indicating a formerly more extensive drainage system; 4. Aggraded lower valleys where the bedrock valley floor relates to low sea level during the maximum glaciation.

2.4.2 CLIMATE CHANGE AND DRAINAGE-NETWORK RESPONSE

Changes of a fluvial hydrosystem can result from changes in upstream or downstream variables. Climatic changes influence fluvial hydrosysterns by changing discharge and sediment yields in two ways: (1) by changing precipitation and solar energy inputs, and by the resulting vegetation changes; and (2) by changing baselevel. For example, one measure of drainage structure within river basins is the drainage density (D), the total channel length (L) divided by basin area (A): D = LIA Another measure of drainage character is stream frequency (F): F=NIA where N is the number of stream segments of all Strahler orders. Studies of relationships between drainage density and stream frequency for mature basins (Melton, 1958) demonstrate a high coefficient of correlation. The derived function F = 0.6941)2 shows that as drainage density increases within a constant area, it does so by an accompanying increase in stream frequency. Drainage networks develop on new land surfaces over time-scales of 104+ years. For a study of drainage development on tills of different age, Ruhe (1952) suggested that rapid initial network expansion was followed by stabilization after about 20 000 years. Experimental studies (Schumm, 1977) confirmed a third phase, first noted by Glock (1931), of drainage integration and channel loss. During the period of constant drainage density there is a progressive change of drainage pattern, characterized by the loss of low-order tributaries in the centre of the basin and the addition of first-order tributaries at the periphery. Thus, as the 'older' interior segments enter Glock's phase of integration and channel loss, network growth continues in those areas of steeper slope towards the margins of the basin. Sediment yield is high during the first phase of network expansion, but then declines rapidly over time (Figure 2.6c). In part, the decreasing sediment yield reflects the falling rate of network growth but it also relates to the progressive widening of the valleys and declining main valley slopes which increase opportunities for sediment storage within the drainage basin. Climate exerts a strong influence on drainage density (e.g. Gardiner, 1987). Under constant climatic conditions, the drainage density adjusts to an equilibrium conditioned by channel head steam erosion and depositional infill by slope processes (Calver, 1978). The primary controls on drainage density are surface geology, relief, climate and vegetation (Table 2.1). The influence of vegetation on runoff and sediment yields (Figure 2.5) is particularly important. With high runoff rates, channel headward extension and bifurcation increase channel density; a small catchment area is required to sustain a stream, and drainage density is high. Thus, in semiarid environments where high-intensity storms combine with thin soils and poor vegetation cover to generate high runoff rates and high sediment yields, drainage densities can exceed 100 km-1• In contrast, humid temperate environments with well-rounded slopes, dense vegetation cover and deep soils have low drainage densities, often less than 5 km-1.

2.4.3 EPISODES OF CHANGE.

Changes of the external controls induce adjustments to new equilibrium states but because time lags occur in the system, drainage basin evolution can be characterized by episodes of high rates of geomorphic work, evidenced by high sediment yields. With climatic change, vegetation response is relatively slow and a short period of landscape instability may be triggered (Figure 2.6a). For example, with a change from arid to humid climates, drainage density, runoff rates and sediment yields are seen to increase rapidly to a maximum before declining to a relatively low level. Similar episodes of change occur following glaciation. Church and Ryder (1972) defined the term paraglacial to cover non-glacial processes that are directly conditioned by glaciation. This concept focuses on the instability of the exposed glacial sediments with respect to the fluvial environment which succeeds the glacier spatially and temporally. The sediment yield of postglacial rivers is far in excess of the 'normal' material supply to be expected in the non-glacial environment (Figure 2.6b). Following deglaciation, vegetation succession, changes in runoff and sediment exhaustion return the sediment yield to levels conditioned by concurrent rates of debris production by primary weathering. Another important external change is baselevel lowering. Declining sea-level increases relief leading to an extension of the drainage network. Initially, there will be an episode of very high sediment yield (Figure 2.6c) as main channel incision moves progreSSively upstream from the basin mouth, scouring alluvium previously deposited in the valley. Tributaries will be rejuvenated successively as incision progresses upstream, and drainage density will increase. A new 'normal' sediment yield will be established reflecting the increased basin relief and extended drainage network. 2.4.4 COMPLEX RESPONSES Field studies of alluvial chronologies in relation to regional climatic changes during the Holocene (Schumm, 1977; Starkel et al., 1991) demonstrate that the number, magnitude, and duration of erosional and depositional events varied between valleys and along the same valley. Geomorphic histories are complicated: first, because changes of the external controls varied at the regional scale; secondly, because large basins comprise a hierarchy of process-response systems (subbasins and main channel sectors); and thirdly, because the morphological systems are also complex. With regard to regional variations of the external controls, baselevel changes, for example, have varied markedly around the North Atlantic. Following the maximum of the last glaciation, eustatic rise was complete by about 5000 years BP but isostatic movements have continued and are still continuing in some areas. About 28 000 years BP, the channel of the River Thames in what is today the outer estuary, was about 50 m below present sea level. Clark et al., (1978) predicted changes of sea-level relative to the 16 000 years BP shoreline which indicated a rise of sea-level by over 75 m along the Portuguese coast to a relative fall of 900 m around the Baltic, here reflecting the dominance of isostatic movements. Within a single, large drainage basin a complex response is indicated by studies of sediment-yield variations during drainage basin development following baselevellowering (Figure 2.6c). Typically, the decrease in sediment yield over time is punctuated by secondary sediment pulses. The morphological interactions involved in such a complex sequence of sediment-yield variations are indicated in Figure 2.2c. Tributary erosion follows main channel incision, the main channel becomes a conveyor of upstream sediment in increasing quantities, and the inevitable result is deposition and formation of a braided stream. As tributaries adjust to the new baselevels, sediment loads decrease, and a new phase of main channel erosion occurs. The channel form changes to a single channel with low width/depth ratio reflecting the decreased sediment loads. The speed with which morphological changes establish a new equilibrium or recover to the former equilibrium dimensions is known as the relaxation time (Figure 2.7a). A change of input may be a short-lived disturbance (e.g. a rare high-magnitude flood) or a sustained impact such as climate change or human interference (e.g. dam construction). Nevertheless, different components of the system respond at different rates (Figure 2.7b). An increase in runoff, for example, will lead relatively quickly to a change of the cross-sectional width and depth of alluvial channels, but the increase of drainage density, requiring the development of new fingertip channels, will occur more slowly. Thus, different landforms have different sensitivities to change (Brunsden and Thomes, 1979). The recovery of morphologic equilibrium will also vary (Figure 2.7c). Systems with low sensitivity have long relaxation times and slow recovery, such that they reflect the characteristics of relatively frequent disturbance and demonstrate progressive change (Figure 2.7cc). Those with high recovery rates tend to exhibit considerable temporal adjustment to the general magnitude of frequent processes (Figure 2.7ca). However, even within these systems, at any given point in time each part of the landscape may exhibit varying degrees of adjustment to present processes. To understand the evolutionary behaviour of river channels Schumm (1973, 1977, 1979) introduced the concepts of 'complex response' and 'geomorphic thresholds'. 'Complex response' explains many of the peculiarities of river channel evolution and recent alluvial and terrace sequences, by recognizing that a single event can trigger a complex reaction as the components of the basin morphologic system respond to change. Part of this complexity relates to a change of an external variable (climate, baselevel, isostatic response to denudation, vegetation cover); an external threshold is crossed causing changes within the affected system. However, part also relates to internal adjustments between the components of the drainage basin system. Thresholds can also be exceeded when the external variables remain constant; progressive change of the system itself renders it unstable. For example, progressive channel aggradation causes an increase of channel slope; deposition sediment storage - continues until a threshold slope is reached at which point a phase of channel incision and sediment removal is initiated. The nature of such geomorphic thresholds, determined by the control of slope on sediment transport processes, can be particularly important in explaining the different histories of valley sectors.

2.5 BIOLOGICAL RESPONSES

For biota, rivers and streams are hazardous environments characterized by stresses (the velocity or shear stress of the flowing water) and disturbances: erosion, abrasion, siltation and burial, and desiccation as well as extremes of water quality (temperature, dissolved oxygen, pH, toxic metals). Early biological research on rivers attempted to define zones on the basis of site-specific, or biotope-specific, characteristics and related these to the longitudinal succession of fish species, benthic invertebrate taxa and algae (see Hawkes, 1975 for a review). The stream classifications (see Table 8.1) used various physical parameters such as the types of stream bed, channel slope, valley cross-sectional form, and annual temperature profiles. However, rivers were viewed merely as canals isolated from drainage basin processes. Hynes' (1970) pioneering approach to running-water ecology led to the rapid development of interactions between biologists, hydrologists and geomorphologists. Subsequently, the 'catchment ecosystem approach' (Bormann and Likens, 1979) has become accepted for a variety of studies. Thus, Lotspeich (1980, p. 585) concluded: The problem of classifying natural areas, especially aquatic systems, would be greatly simplified if the watershed (=drainage basin) were the base for classifying, since it is the watershed, functioning in response to external forces, that controls aquatic systems. Biological communities became viewed as deterministic systems having evolved toward the most probable conditions of channel morphology and flow, and the drainage basin became the basic unit in lotic (running water) ecology (Cummins, 1992). 2.5.1 HYDRAULIC STRESS The structure and function of most aquatic communities, and many rivermargin ones, is related to hydraulic conditions (Gore 1994): to flow patterns at scales ranging from the whole river to the individual bedform (Chapters 5, 6, 7 and 8). Variations of hydraulic variables at a point or averaged for a cross-section are particularly important in studies of fluvial hydrosystems and can usefully be described in relation to changes of discharge (Figure 2.8) - the 'hydraulic geometry' approach (Leopold and Maddock, 1953). This approach may also be used to describe the variation of hydraulic variables along a river (the downstream hydraulic geometry) in relation to a standard flow, such as the median (50th percentile), the bankfull, or some intermediate, reference flow (such as the 5th percentile flow). At a local scale (0.1-100 m2) hydraulic conditions, especially flow velocity, are seen as having the dominant influence on the pattern of species distributions within a site or sector. This has been stressed especially for benthic macroinvertebrates (Statzner and Higler, 1986; Statzner et al., 1988) but it is also implicit in studies of habitat suitability for fish (Stalnaker et al., 1989). Velocity is particularly important because it influences metabolism and a large variety of behavioural characteristics in animals living in running waters. 2.5.2

DISTIJRBANCES Those concerned with riparian zones and floodplains also recognize the role of hydraulics in determining the patchy distributions of species, even though the patch dimensions are larger than the aquatic ones, typically varying from 10 to 1000 m2 (Naiman and Decamps, 1990; Petts, 1990). Physical disturbance plays a key role in structuring these systems. For example, rivers maintain vegetation typical of early succession due to periodic scouring and disturbance influences competition by creating gaps for exploitation by less competitive species. However, in order to understand the long-term influence of such disturbances, the nature of recovery mechanisms (or negative feedback processes) must also be considered. Pickett and White (1985) define disturbances as any relatively discrete event in time that disrupts ecosystem, community, or population structure and that changes resources, ava!1ability of substratum, or the physical environment. The roles of disturbance in stream ecology have been discussed by Resh et al., (1988) and specifically with regard to land-water ecotones, by Naiman and Decamps (1990). However, a disturbance, such as a flood, can not be defined solely in terms of the event magnitude and frequency. The severity of impact will also relate (a) to the timing of the disturbance in relation to stability thresholds, physical and biological, and (b) to the effectiveness of recovery processes (Milner, 1994). Two groups of disturbances should be distinguished: relatively frequent and low magnitude events that are part of the 'normal' regime and more or less predictable; and relatively rare disturbances that are outside the predictable range. In reality, these two groups are the end members of a continuum because the frequency of low magnitude disturbances influences the rate of system recovery to unpredictable disturbance by extreme events (Figure 2.7c). Over medium timescales (1~100 years) most river systems may be viewed as quasi-equilibrium states. The long-term effectiveness of an event in disturbing this quasi-equilibrium condition varies widely, depending on the system component under consideration (such as benthic invertebrates or channel morphology) and the effectiveness of the recovery processes (Figure 2.7c). For example, the insect species of semiarid streams recover quickly to catastrophic wash-out by floods because they have short life cycles and continuous emergence, ensuring rapid recolonization by adults from adjacent riparian areas (Fisher et al., 1982). In contrast, the morphology of these streams may experience progressive change because of the weak physical recovery processes to flood-induced erosion (Wolman and Gerson, 1978). Most river ecosystems are characterized by strong recovery processes and this is important when considering river management (Chapter 12).

2.6 THE FLUVIAL HYDROSYSTEM APPROACH

Chorley et al., (1984, p. 7) describe 'The palimpsest view of the landform system ... composed of a nested hierarchy of subsystems each having different levels of sensitivity and recovery, the whole being subject to a temporal stream of input (i.e. process) changes'. This view may be extended to consider the fluvial hydrosystem and is complicated further by biological processes, especially succession. Whereas functional explanations of fluvial hydrosystems can be developed for short time-scales, short-term functional relationships must be seen as superimposed on the longer-term character of the dynamic drainage basin system. Inter-adjustments of the different structural units within a fluvial hydrosystem involve directional changes, thresholds, feedback effects and time-lags. Within the drainage basin, different landforms evolve in different ways and at different rates and the basin may be viewed as a patchwork of morphological units each bearing the imprint of processes of different antiquity. Tune lags within the fluvial system operate at two levels: lags in the upstream and downstream transmission of perturbations, and lags between changes of physical habitat and biological responses. As illustrated in Figure 2.2(c), such responses are made complex by interactive adjustments between adjacent sectors along the main river and between the main river and its tributaries. Superimposed on the overall downstream gradient of ecological conditions, the fluvial hydrosystem approach recognizes discrete reaches (channel sectors), differentiated by discontinuities, within which a quasiequilibrium condition may be defined and functional relationships established over time-scales of 10-100 years. The discontinuities between sectors reflect the heterogeneous character of large basins: 1. The complex geologies, spatially variable climates and different vegetation zones which cause different headwater basins to produce different discharges, water qualities and sediment loads; 2. The complex geolOgic history of large basins which produces sectors with different valley dimensions and alluvial fills that vary in thickness and sedimentology; and 3. The variable rates of change of the different components of the morpholOgiC system and of biological succession. Thus, fluvial hydrosystems reflect the interaction of hydrological, geomorphological and biological processes. They are characterized by a patchwork of aquatic and semiaquatic habitats, the arrangement of which continually changes as a result of disturbance (Chapter 5) and succession (Chapter 10). In summary, the fluvial hydrosystem approach views the ecological characteristics of any site on a river (a) as part of the larger drainage basin system, modified by (b) the historical legacy of environmental change that may have affected the whole catchment or have been specific to each sector, and (c) as determined not only by longitudinal fluxes but also by lateral and vertical exchanges between the channel and its floodplains and alluvial aquifer, respectively (Petts & Bravard, 1998).

References:

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UNIT-2: DRAINAGE BASIN AS A UNIT OF GEOMORPHIC STUDY

Drainage basin is an area drained by the stream and its tributaries. It is bounded by a divide. Drainage basin is also sometimes called watershed or catchment area. It can be thought of as an open system that receives energy or input from the atmosphere and sun over the basin and loses energy or output through the water and sediment mainly through the basin mouth or outlet (Strahler 1964). The present form of any drainage basin is the result of the processes that have operated in the past on the material available locally. These processes at the basin level are the precipitation and runoff, sediment yield and rate of erosion. However, these processes in the past may not be the same in their relative importance as the ones that operate in the drainage basin at present. The importance of studying the drainage basin characteristics derives from the need of studying forms of channels and channel networks as they are related to physical characteristics of the drainage basin, and also from the need of relating physical characteristics of the basin to flow characteristics and sediment yield. The drainage pattern is the arrangement and length of small, medium and large streams in the basin. Two aspects of the development of drainage basins have been studied. In earlier years, the drainage pattern development in relation to the structure and lithology of the underlying rocks was studied. This was essentially qualitative in nature. In the recent times drainage patterns have been treated more as geometric patterns and attempts have been made to derive relationships for them (Horton 1945). The drainage pattern acquired at any time is the result of the combined effect of lithology, precipitation pattern and climate, and their variation with respect to space and time. Since the sediment eroded from the drainage basin along with water causing erosion, flows through the tributaries and the main stream, the drainage net is intimately associated with the hydraulic geometry of the stream channels and their longitudinal profile. As suggested by Schumm (1977) the drainage basin is primarily a sediment production area where climate, diastrophism and land use act as the upstream controls.

Glock (1932) assumed that the drainage pattern is initiated on an essentially smooth plane due to the uplift. According to him the drainage pattern goes through the following developmental stages: initiation, elongation (headward growth of the main stream), elaboration (filling in of the previously undissected areas by small tributaries), maximum extension (the maximum development of the drainage pattern) and abstraction (loss of tributaries as the elevation is reduced through time). This sequence takes a long time in geologic sense. During this sequence the sediment yield first increases to a maximum and then decreases. However, such erosional development cannot be observed. Hence several drainage basins in different stages of development are studied at a given time. Thus what is to be observed in time domain is studied in space domain assuming the process to be ergodic. The topographic characteristics of the drainage basin can be visualised either for the basin or for the drainage network. The most important topographic characteristics for the basin are its area, length, shape and relief. The corresponding characteristics for the drainage network are area tributary to stream channels, drainage density, stream length, network shape or drainage pattern, and network relief.

DRAINAGE PATTERNS AND TEXTURE

Drainage pattern is the general arrangement of channels in a drainage basin. Drainage patterns reflect the influence of such factors as initial slope, inequalities in rock hardness, structural controls, recent diastrophism, and recent geomorphic and geologic history of the drainage basin. Because drainage patterns are influenced by many factors, they are quite useful in the interpretation of geomorphic features and their study represents one of the more practical approaches to the understanding of the structural and lithologic controls on landform evolution. Looking at them in the most general manner, one can classify drainage patterns into the following categories: Figure 2.1 (a) shows dendritic or branch-like pattern that is probably the most common drainage pattern. This is characterised by irregular branching of tributary streams in many directions and at almost any angle usually less than 90o . Dendritic patterns develop on rocks of uniform resistance and indicate a complete lack of structural control. This pattern is more likely to be found on nearly horizontal sedimentary rocks or on areas of massive igneous rocks. They may also be seen on complex metamorphosed rocks. Trellised or lattice-like pattern shown in Fig. 2.1 (b) displays a system of sub-parallel streams, usually along the strike of the rock formations or between parallel or nearly parallel topographic features recently deposited by wind or ice. Radial pattern shown in Fig. 2.1 (c) is usually found on the flanks of domes or volcanoes and various other types of isolated conical and sub conical hills. Parallel drainage pattern shown in Fig. 2.1 (d) is usually found in regions of pronounced slope or structural controls that lead to regular spacing of parallel or near parallel streams. Rectangular drainage pattern shown in Fig. 2.1 (e) has the main stream and its tributaries displaying right-angled bends. This is common in areas where joints and faults intersect at right angle. The streams are thus adjusted to the underlying structure. Deranged drainage pattern, see Fig. 2.1 (f) indicates a complete lack of structural or bed rock control. Here the preglacial drainage has been affected by glaciation and new drainage has not had enough time to develop any significant degree of integration. It is marked by irregular stream courses that flow into and out of lakes and swamps and have only a few short tributaries.

Centripetal pattern shown in Fig. 2.1 (g) is encountered locally. Here the drainage lines converge into a central depression. These are found on sinkholes, craters and other basin like depressions. Highly violent pattern shown in Fig. 2.1 (h) is characteristic of areas of complex geology. The complex drainage patterns observed in nature are a result of differing lithology, regional slopes, presence of joints and faults, and geologic activities such as glaciation, volcanism and limestone solution. Zernitz (1932), Howard (1967) and Thornbury (1969) have given full description of commonly occurring drainage patterns and their interpretation.

Drainage Texture An important geomorphic concept about the drainage pattern is the drainage texture by which one means relative spacing of drainage lines. Drainage texture is commonly expressed as fine, medium or coarse. Climate affects the drainage texture both directly and indirectly. The amount and type of precipitation influence directly the quantity and character of runoff. In areas where the precipitation occurs primarily in the form of thunder showers, a larger percentage of rainfall will result in runoff immediately and hence, other factors remaining the same, there will be more surface drainage lines. The climate affects the drainage texture indirectly by its control on the amount and types of vegetation present which, in turn, influences the amount and rate of surface runoff. With similar conditions of lithology and geologic structure, semiarid regions have finer drainage structure than humid regions, even though major streams may be more widely spaced in semiarid than in humid regions. It is also noticed that drainage lines are more numerous over impermeable materials than over permeable areas. The initial relief also affects drainage structure; drainage lines develop in larger number upon an irregular surface than on the one that lacks conspicuous relief. Bad-land topography promotes fine drainage structure. Impermeable clays and shales, sparse vegetation and existence of thundershowers are responsible for very fine drainage structure. Coarse drainage structure is in particular found on sand and gravel outwash plains. Gravel plains have fewer drainage lines on them than adjacent till plains underlain by relatively impermeable clay till. The drainage structure can be qualitatively related to a parameter known as drainage density (see section 2.9) first defined by Horton (1932) as total length of streams per unit of drainage area. Drainage density varies from about 0.93 km/km2 on steep impervious areas to nearly zero for highly permeable basins. It varies from about 2.0 to 0.60 km/km2 in humid regions. As indicated by Smith (1950) and Strahler (1957), coarse drainage structure corresponds to drainage density less than 5.0 km/km2 , medium drainage structure to drainage density value between 5 and 15 km/km2 and fine drainage structure to drainage density between 15 and 150 km/km2 .

STREAM ORDER

A stream net or river net is the interrelated drainage pattern formed by a set of streams in a certain area. A junction is the point where two channels meet. A link is any unbroken stretch of the river between two junctions; this is then known as the interior link. If it is between the source and first junction, it is called the exterior link. Quantitative analysis of the stream network really started with Horton (1945). This analysis has been developed to facilitate comparison between different drainage basins, to help obtain relations between various aspects of drainage patterns, and to define certain useful properties of drainage basins in significant terms. According to Horton (1945) the main stream in the river net should be denoted by the same order number all the way from its mouth to its headwaters. Thus, at every junction where the order changes, one of the lower order streams is renumbered to the higher order and the process repeated. Thus in Fig. 2.2 (a) the main stream is shown as the fourth order stream right back to its source. The third order streams which are tributary to the fourth order stream are also extended back to their farthest source as the third order streams and so on. The streams joining the third order stream are second order stream and they can be extended backward. It can be immediately realized that a certain amount of subjectivity is involved in the ordering of streams according to Horton's method. In Strahler's (1952) system, see Fig. 2.2 (b), the headwater streams that receive no tributary are called first order streams. Two first order streams unite to give a second order stream. Two second order streams unite to give a third order stream and so on. When two streams of different order unite, the combined stream retains the order of the higher order stream. A combination of two streams of lower order, say $(u - 1)$, with a stream of given order u increases the order of the latter by one integer, that is $(u + 1)$. The result of this system of ordering is that it does not reflect any increments except approximately doubling the discharge if one assumes that streams of the same order in the same drainage basin carry approximately equal discharges. Scheideggar (1965) defines the order x after two streams of order u1 and u2 by x = log2 2 2 1 2 u u d i ´ ...(2.1) His system of ordering is shown in Fig. 2.2 (c). Shreve (1967) has suggested a system of ordering streams in which, the order numbers of two streams contributing to the junction are added to arrive at the order number below the junction, see Fig. 2.2 (d). Thus each exterior link or head tributary has a magnitude 1. If links of magnitude u1 and u2 join, then the resultant downstream link has the order (u1 + u2). If we assume that the first order streams are approximately of the same magnitude and that the discharge is neither lost nor gained from any source other than the tributaries (which is not completely true) then Shreve number is roughly proportional to the discharge in the segment of the stream. It may be mentioned that Strahler's system of ordering has been more commonly used than the other methods and the same is utilised herein. The analysis of drainage basin considering stream orders is often known as morphometry. The morphometric analysis of drainage basins carried out by Horton (1945), Strahler (1952), Rzhanitsyn (1960) and others is based on the premise that for the given conditions of lithology, climate, rainfall, and other relevant parameters in the basin, the river net, the slope and the surface relief tend to reach a steady state in which the morphology is adjusted to transmit the sediment and excess flow produced. If there are any major climatic or hydrologic changes in the region, the steady state morphologic characteristics will naturally be modified. In other words, the river net is the definite response of the drainage basin to the complex physical processes taking place over the drainage basin.

AREAS OF DRAINAGE BASINS

Basin area is hydrologically important because it directly affects the size of the storm hydrograph, and the magnitude of mean and peak flows. Amount of sediment eroded from the drainage basin is also related to the basin area. In fact, since almost every watershed characteristic is correlated with area, the area is the most important parameter in the description of form and processes of the drainage basin. The area Au of a basin of given order u is defined as the total area projected upon a horizontal plane, which contributes overland flow to the channel segment of a given order and all the tributaries of the lower order. Thus area of the basin of the third order, A3 will be the sum of areas of first and second order basins, plus all additional areas, known as inter-basin areas, contributing directly to channels of order higher than the first.

BASIN SHAPE

Basin shape affects the hydrologic characteristics of the basin, namely hydrograph shape. As mentioned earlier a long narrow basin having high bifurcation ratio gives a low but sustained peak whereas round basins with low bifurcation ratio would give a sharply peaked hydrograph.

The circularity and elongation ratios can be of practical utility in predicting certain hydrological characteristics of the drainage basin. Elongation ratio has been used in the studies of sediment eroded from the basins. In general drainage basins tend to become more elongated with strong relief and steep slopes. Available data indicate that the drainage basin gets relatively elongated as its size increases.

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VEGETATION

Vegetation including grass, shrubs, and forests plays an important role in the hydrologic cycle and catchment erosion. Hence, its effect is of prime importance to those working on river morphology and river dynamics. Studies by various investigators have shown that water and sediment yield, flood peaks and the time of their occurrence, and the velocity of travel of the flow peak are strongly influenced by the nature and extent of vegetation. When the pressure on the land, because of increase in the population and human activity, was not heavy, there were marginal changes in the forests, and minor disturbances in their coverage were soon made up naturally. However, because of increase in the population and industrial growth and consequent increase in food, space and energy requirements of nations, there has been indiscriminate deforestation in some parts of the world. In the early eighties, most of the tropical forests were estimated as being altered by man at around 12 million hectares per year, see Bruijnzeel (1990). Many investigators consider this as an underestimate. Deforestation includes cutting of trees for fuel, timber and other industrial uses, deforestation caused by great and small forest fires, shifting of zoom cultivation, construction activity related to logging such as creation of access roads, skid tracks and landings, clearing areas for habitation and developing industry, surface mining and similar activities. Generally speaking, a forest subjected to some of the above mentioned disturbances may recover to its previous state if left alone for a sufficiently long period. However, this is not the case when the forest is converted to permanent agriculture such as grazing, cropping or extractive tree crops. The effects of partial or complete removal of forest on climate, water yield and its seasonal distribution and on sediment production have been studied by many investigators. Below are given the salient features of these effects, (see Bruijnzeel (1990)). Rainfall and Water Yield 1. Tropical forests reflect about 12 percent of the incident short wave radiation while agricultural crops reflect 15 to 20 percent. Hence a different partitioning of energy between warming up of the boundary layer and evaporation is to be expected when tropical forests are converted to grass lands or agricultural crops. 2. As a result of extensive studies during the past four decades, it is found that the extent of forests has definite effect on rainfall. 3. It has been found that in humid tropics removal of natural forests cover may result in considerable initial increase in water yield, the increase depending on the amount of rain received. The initial increase in water yield, after removal of the forest cover, gradually decreases with the passage of years and may return to pre cut flows in about eight years in case of natural regrowth.

Stream Flow Regime 1. Geological, topographical and vegetative cover play an important role on floods and hence isolation of effect of vegetation becomes rather difficult. 2. If geology is favourable, cutting vegetation shifts infiltration flow to surface flow and therefore peaks will enhance. 3. Also in the absence of retarding effect, the peak is likely to occur earlier. 4. Change in evapo-transpiration and infiltration opportunities associated with change in forest cover will govern the dry season flows. If infiltration opportunities after forest removal have decreased to the extent that the increase in amount of water leaving the area as stream flow exceeds the gain in base flow associated with reduced evapo-transpiration, then the dry season flow is reduced. If on the other hand, the surface infiltration characteristics are maintained over most of the area by deliberate soil conservation practices or by some other method, then the effect of reduced evapo-transpiration after clearing will show up as increased base flow or dry seasonal flow.

Sediment Production and Yield When dealing with the effect of change in forest cover on erosion and sedimentation it is helpful to distinguish between surface erosion (i.e., splash, sheet and rill erosion), gully erosion, and mass movements because the ability of vegetation cover to control the various forms of erosion is rather different. It is well known that only part of the material eroded from hill side will enter drainage network, the rest may move into temporary storage such as depressions, foot slopes, small alluvial fans or in small tributaries, behind debris basins or flood plains. The stored material may be released during large storms or caught by vegetation, or form stable topographic elements. Since these storage opportunities tend to increase with increase in area, sediment delivery ratio, which is defined as the amount of sediment passing a given section during a given time divided by amount of sediment eroded from upstream in the same time, is found to decrease with increase in catchment area. It may be years before sediment stored in temporary storages is released and its effect felt several kilometers downstream from the region of erosion. Sediment yield, which is rate of sediment passing a given section, is discussed in detail, in Chapter–III. This was found to be the case on the Brahmaputra river in Assam (India) after

1950 strong earthquake, see Goswami (1985). During August 1950 earthquake, apparently one of the most severe ever recorded, massive landslides occurred which temporarily blocked many major tributaries. Bursting of these dams after several days not only produced devastating floods downstream, but also brought down enormous volume of sediment thereby raising the beds of these rivers considerably. The mean annual suspended load and water discharge between 1955–1963 were 750 000 m3 and 16 530 m3 /s as against 130 000 m3 and 14 850 m3 /s during 1969–1976. Also during the former period the river reach upstream of Pandu was aggrading, whilst it was degrading during 1969–1976.

Surface and Gully Erosion Studies by Wiersum (1984) have indicated that erosion is minimum (0.10 to 0.60 tonnes/km2 /year) in those areas where soil surface is adequately protected by a well developed litter and herb layer. When this layer is destroyed or removed, erosion rates rise dramatically to 500 to 5000 tonnes/km2 /year. Hence protection from tree stands lays not so much in the ability of tree canopy to break the power of rain drops but rather in developing and maintaining a litter layer. When rills are formed and they grow into gullies, their lateral and head ward extension through scouring, undercutting and subsequent collapse of walls cause a large increase in sediment production. Mass Wasting Some of the highest reported natural erosion rates from rain-forested areas have been related to intense mass wasting under conditions of steep topography, tectonic activity, and intense rainfall. In mass wasting, steep slopes in combination with geological and climatic factors are more important than land use. Prasad (1975) after ten years of observations of seismic activity, rainfall and occurrence of land slides in eastern Nepal concluded that intense precipitation and associated saturation of soil were apparently more important than seismic shocks. Starkel (1972) has opined that the role of vegetation in preventing shallow slope failures (less than 3 m) is very important; Manandhar and Khanal (1988) have confirmed this in south of Khatmandu. As regards the influence of tall vegetation on slope stability, the net effect is considered positive, the major factor being the mechanical reinforcement of the soil by tree roots.

RELIEF ASPECTS

The relief is the difference in elevation between given points. Maximum basin relief is the difference in elevation between the basin mouth and the highest point on the basin perimeter. Alternative definition of maximum relief is the basin relief along the longest dimension of the basin parallel to the principal drainage line. Relief ratio Rk is the ratio of maximum basin relief to the horizontal distance along the longest dimension of basin parallel to the principal drainage line (Schumm 1956). Melton (1958) defines the relative relief as the maximum relief H divided by the basin perimeter P while Maxwell defines the relative relief as H divided to basin diameter. Use of the perimeter as the horizontal length dimension solves the difficulty of locating a suitable axial line in the basin. Two other parameters involving maximum relief have been defined by Strahler (1957). The ruggedness number is the product of maximum relief H and the drainage density Dd i.e. (H Dd). The geometry number is defined as (H Dd /S) where S is the ground slope. Both these parameters are dimensionless. Observed values of ruggedness number vary from 0.05 to about 1. Strahler (1964) found that the geometry number varies over a relatively narrow range viz. 0.40 to 1.0.

Hypsometric Curves

Let “a” be the horizontal projected area of a drainage basin at an elevation h (see Fig. 2.7) and A be the total projected area. Then one can prepare a curve between relative height h/H and relative area a/A as shown in the Figure 2.7. Such a curve is known as hypsometric curve (Garde, 2006).

References:

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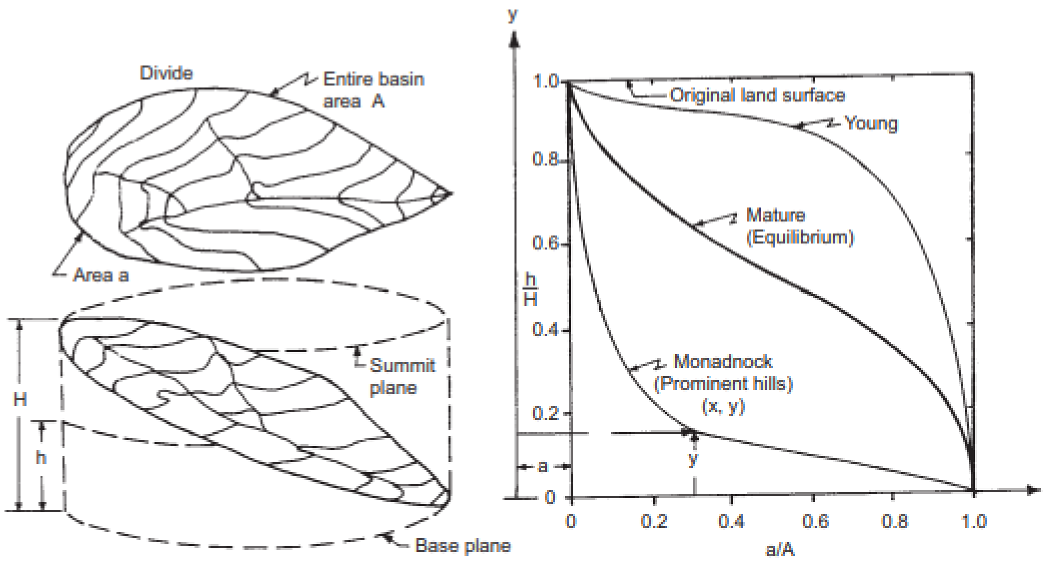
UNIT-3: HYPSOMETRIC ANALYSIS OF DRAINAGE BASIN AND ITS SIGNIFICANCE

Hypsometric Curves Let “a” be the horizontal projected area of a drainage basin at an elevation h (see Fig. 2.7) and A be the total projected area. Then one can prepare a curve between relative height h/H and relative area a/A as shown in the Figure 2.7. Such a curve is known as hypsometric curve. The analysis of large drainage basins using such curves was first done by Langbein (1947) and later used by Strahler and others. The hypsometric curve, in general, will change with time because of the gradual erosion of areas at higher levels and hence the relative position of the hypsometric curve on a/H vs h/H graph gives an idea about the stage of development of the basin landscape. Figure 2.7 shows young and mature stages of topography. This figure also shows monadnock phase in which the resistant rock in the basin may form prominent hills at isolated places giving a distorted hypsometric curve. Sometimes the integral o 1 z f (x) dx where f (x) = h/H and x = a/A is used as an index of evolution of the topography of the basin. This integral represents the rock mass that is still to be eroded. Young phase would correspond to a high value of the integral while mature phase would correspond to a relatively small value. Strahler (1964) has indicated that most of the hypsometric curves can be represented by an equation of the form Here y and x are as shown in Fig 2.7. The exponent z increases as the topography becomes more mature. Hypsometric curves are also related to hydrologic characteristics of the drainage basin. Thus distribution of elevation in a drainage basin is closely related to the amount of flood storage available, the effect of which is to make the rising limb of hydrograph less steep, increase the time lag and make the peak lower and less pronounced. Knowledge of hypsometric curve is also useful in better estimates of rainfall, snowfall and evaporation in the basin.



UNIT-4: CLASSIFICATION OF CHANNEL LINKS, SHREVE'S FORMULA OF LINK COUNTING

A stream net or river net is the interrelated drainage pattern formed by a set of streams in a certain area. A junction is the point where two channels meet. A link is any unbroken stretch of the river between two junctions; this is then known as the interior link. If it is between the source and first junction, it is called the exterior link. Quantitative analysis of the stream network really started with Horton (1945). This analysis has been developed to facilitate comparison between different drainage basins, to help obtain relations between various aspects of drainage patterns, and to define certain useful properties of drainage basins in significant terms. According to Horton (1945) the main stream in the river net should be denoted by the same order number all the way from its mouth to its headwaters. Thus, at every junction where the order changes, one of the lower order streams is renumbered to the higher order and the process repeated. Thus in Fig. 2.2 (a) the main stream is shown as the fourth order stream right back to its source. The third order streams which are tributary to the fourth order stream are also extended back to their farthest source as the third order streams and so on. The streams joining the third order stream are second order stream and they can be extended backward. It can be immediately realized that a certain amount of subjectivity is involved in the ordering of streams according to Horton's method. In Strahler's (1952) system, see Fig. 2.2 (b), the headwater streams that receive no tributary are called first order streams. Two first order streams unite to give a second order stream. Two second order streams unite to give a third order stream and so on. When two streams of different order unite, the combined stream retains the order of the higher order stream. A combination of two streams of lower order, say (u – 1), with a stream of given order u increases the order of the latter by one integer, that is (u + 1). The result of this system of ordering is that it does not reflect any increments except approximately doubling the discharge if one assumes that streams of the same order in the same drainage basin carry approximately equal discharges. Scheideggar (1965) defines the order x after two streams of order u1 and u2 by x = log2 2 2 1 2 u u d i ´ ...(2.1) His system of ordering is shown in Fig. 2.2 (c). Shreve (1967) has suggested a system of ordering streams in which, the order numbers of two streams contributing to the junction are added to arrive at the order number below the junction, see Fig. 2.2 (d). Thus each exterior link or head tributary has a magnitude 1. If links of magnitude u1 and u2 join, then the resultant downstream link has the order (u1 + u2). If we assume that the first order streams are approximately of the same magnitude and that the discharge is neither lost nor gained from any source other than the tributaries (which is not completely true) then Shreve number is roughly proportional to the discharge in the segment of the stream. It may be mentioned that Strahler's system of ordering has been more commonly used than the other methods and the same is utilised herein. The analysis of drainage basin considering stream orders is often known as morphometry. The morphometric analysis of drainage basins carried out by Horton (1945), Strahler (1952), Rzhanitsyn (1960) and others is based on the premise that for the given conditions of lithology, climate, rainfall, and other relevant parameters in the basin, the river net, the slope and the surface relief tend to reach a steady state in which the morphology is adjusted to transmit the sediment and excess flow produced. If there are any major climatic or hydrologic changes in the region, the steady state morphologic characteristics will naturally be modified. In other words, the river net is the definite response of the drainage basin to the complex physical processes taking place over the drainage basin.

UNIT-5: COMPUTATION OF TOPOLOGICALLY DISTINCT CHANNEL NETWORK AND TOPOLOGICALLY INTEGRATED CHANNEL NETWORK

As the properties of drainage networks1 are strongly influenced by differences in rock resistance and in surface slope, a knowledge of such properties will in principle be useful in the interpretation of landforms and geologic structure, both past and present. The earliest method of classifying drainage networks was by patterns, or qualitative features that could be easily identified and associated with some type of geologic structure. Thus, dendritic patterns occur in regions of uniform lithology, rectangular patterns occur in regions of orthogonally intersecting faults and joints, and so forth. Zernitz (1932) gives a detailed review of drainage patterns and their interpretation. Horton (1932, 1945) made an entirely different approach to the study of drainage networks. Unlike previous investigators, Horton was interested in a quantitative analysis of drainage basins, and his work was based upon a method he had devised for classifying individual streams and networks. Horton's classification procedure has been generally superseded by a modification proposed by Strahler (1952), and in this paper Strahler's method will be used exclusively. In Strahler's procedure, channels that originate at an external source are classified as first order; when two streams of order co join, a stream of order (a; + 1) is created. The order of a channel network is that of its highest order stream. For a complete discussion of Horton and Strahler ordering methods, see Shreve (1966, p. 20-23). In his 1945 paper, Horton proposed his two well-known laws of drainage composition. Stated in terms of Strahler ordering, they are:

Law of stream numbers (1) where Na is the number of streams of order co in a network of order 0 and RB is the Strahler bifurcation ratio; and Law of stream lengths La, = RL (M1)Li, (2) where La is the mean length of streams of 1 In this paper, the term "drainage network" means all stream channels in a given region. A "channel net- work" is a particular element of a drainage network and consists of all channels upstream from a given point in the drainage network. In general, the terminology is the same as in Shreve (1966, p. 20). order co in the network and Ri is the Strahler stream-length ratio. As indicated by the approximate equality signs, these laws are not intended to be exact but rather to give the central tendency for a large number of networks. The values of the parameters RB and Ri, are generally determined by plotting Na and La versus a; on semilog paper and fitting the points by straight lines; the slopes of the lines are —log RB and log Ri, respectively. For actual stream systems, RB generally falls in the range 2.5 to 5.0 with the modal value near 4; RL generally lies between 1.5 and 3.0. Horton's investigations marked a radical departure from the course of previous work, partly because of their emphasis on quantitative description and partly because they introduced topological considerations. Topology is of course not concerned either with the drainage pattern or with measurements of length and area but with the way in which the various channels are connected. As pointed out by Melton (1959) and Shreve (1966), a channel network with NI sources has (Ni—l) junctions. The topological properties of the network are completely specified when it is known how the (Ni—l) junctions are connected to each other and to the sources. The bifurcation ratio RB gives some information about this connectivity. For example, if RB ~ 2, streams of a given order in general combine only with streams of the same order. If RB ~ 4, the typical value, half of the streams of a given order combine with others of the same order and half terminate in streams of higher order. Horton (1945, p. 290, 300-306) apparently felt that RB would be affected by relief and drainage density but would be relatively insensitive to geological controls. Investiga- tions by Strahler and his students (reviewed in Strahler, 1964) have generally confirmed the latter supposition but not the former. The relation between stream numbers, bifurcation ratios, and network topology was made clear in an important paper by Shreve (1966). Shreve suggested that, in the absence of geological controls, a natural population of channel networks is topologically random, that is, all topologically distinct channel net- works (TDCN) with a given number of sources are equally likely. He showed that for N\ sources the number of topologically distinct configurations is 1 2JVi- 1 (3

given set of stream numbers NI, N% ..., N a — 1, l = nV.-*W N*~2 } »-i \Na - 2Na+l/ ' (4) These two equations completely determine the statistical properties of sets of stream numbers and functions of stream numbers (such as bifurcation ratios) for topologically random networks. In particular, the probability of occurrence of a given set of stream numbers is p(Ni, (5) The value of W(Ni) rapidly becomes astronomical as NI increases; for example, i^(60) = 4.06 X 1032. However, the classification by stream numbers combines different TDCN into a much smaller number of categories. One of the most important results of Shreve's analysis is that the most probable sets of stream numbers have bifurcation ratios in the range 3 to 5, thus explaining Horton's law of stream numbers. As an example, Table 1 shows explicit results for NI = 60. The 4.06 X 1032 TDCN can be represented by 730 sets of stream numbers, of which the 10 most probable are listed in the table. The bifurcation ratios for these 10 sets, which by themselves account for just under half the possible results for NI. = 60, all lie in the range 3 to 5, and clearly the Strahler bifurcation ratio Rg will also lie in this range. Shreve noted that the most probable networks do show a systematic deviation from Horton's law of stream numbers in that the bifurcation ratios for a set of stream numbers tend to decrease with increasing order; this effect is also noted in natural systems (Schumm, 1956, p. 603; Maxwell, 1960, p. 12; Shreve, 1966, p. 24-26). Smart (1968a) has discussed the effect of topological randomness on the statistics of stream lengths. By accepting Shreve's hy- pothesis and introducing the assumption that link lengths are independent random variables drawn from the same population, he was able to derive an approximate relation for mean stream lengths which is an improvement over Horton's law. k = II (Na-i - 1)/(2JV« - 1) , co > 2 , (6) where /t is the mean interior link length. This expression is particularly satisfactory because it does not involve empirical parameters but rather exhibits in a quantitative fashion the complex relation between stream numbers and mean stream lengths (compare Melton, 1958, p. 46). The question of random development of drainage networks was approached in another way by Leopold and Langbein (1962). They proposed a method of simulating drainage networks by random walks on a square lattice. The channel networks produced in this way are strikingly similar to those of natural systems, particularly in regard to the distribution of dimensionless parameters, such as the bifurcation and stream-length ratios. Carrying out the Leopold-Langbein simulation procedure by hand is quite laborious, but Schenck (1963) and Smart and others (1968) have shown how the games can be performed and analyzed on a digital computer and have in general verified and extended Leopold and Langbein's earlier conclusions. The success of the random-walk model lends added support to the idea that natural drainage basins are topologically random. Howard (1968, personal commun.) has used the random-walk simulation to study the effects of stream capture and the role of headward growth in determining network configurations. Thus in the past twenty years the focus of interest on drainage networks has shifted from qualitative properties that are intimately related to geological controls to topological and geometrical properties that can apparently be explained in large part by the simple assumption of randomness. The considerable success of the random model naturally raises the question of whether or not there is anything further to be learned from studying network topology. I believe that there is, for reasons that can be illustrated by extending Shreve's analogy between channel networks and perfect gases. The physics of gases began with experimental observations that led to an empirical equation of state (Boyle's law). On the theoretical side it was clear that a straightforward application of the laws of mechanics was impracticable because of the large number of variables involved. The first successful theory of gases came with the introduction of statistical mechanics with its basic assumption of randomness; this approach provided a good over-all description of gas behavior and, in particular, gave a derivation of Boyle's law. In somewhat the same way, Shreve's and Smart's application of statistics (not statistical mechanics) to the network problem provides a good general description of channel networks and an explanation of Horton's laws of drainage composition. (To digress for a moment, it may be noted that the reason why Horton's laws are much less precise than Boyle's law lies in the small number of "particles" (sources or links) in a typical network. The large-number limits of Horton's laws are given by Shreve (1967).) Although, as already stated, statistical mechanics did explain the essential properties of gases, more and more precise experiments began to disclose systematic disagreements between theory and observation. It was recognized quite early that the principal source of this disagreement was interactions between the gas particles, a phenomenon not included in the perfect-gas model. A combined ex- perimental and theoretical attack based on this point of view then led to a much improved understanding of the interactions between gas molecules. In the same way, one might hope that investigations of systematic deviations of channel networks from the random model would both improve our understanding of the interactions and processes responsible for their formation and maintenance and provide at least a supplementary tool for the interpretation of geological structure. In one sense this proposed study will be more difficult than the corresponding one for gases; because of the small size of the systems involved, the natural scatter in observations will be large, thus making it hard to detect small systematic deviations from random behavior. The general remedy that statistical theory offers for this ailment is of course large sample sizes.

Ambilateral Classification of Network^ Topology In looking for deviations from topologically random behavior, the most straightforward procedure would be to obtain a large sample of networks of given magnitude (N\) and see whether or not each TDCN appears with the same frequency, as predicted by the model. For our previously quoted example of NI = 60, however, where W(N$ = 4.06 X 1032, such a test is obviously impossible, and, in fact, it is practical only for very small N\, say 6 or less (W(6) = 42). For larger NI, in order to make tests that do not require an unreasonable amount of data, some method of grouping the TDCN into a manageable number of classes must be found. Probably the most efficient method of specifying the topological properties of a channel network is the binary digit representation proposed by Shreve (1967) and Scheidegger (1967). Complete information about the topology of a network of magnitude NI is contained in a string of NI ones and (Ni —1) zeroes; for example, Liao and Scheidegger (1968) have shown how to obtain the stream numbers from such a string. I have not, however, found this representation particularly useful in further grouping, and would like to propose here a method of characterizing network topology which, although less abstract and efficient than the binary string, is more closely related to hydrologic and geomorphic considerations. The first step is to specify the Strahler stream numbers. The second step is to indicate, starting with the stream of highest order, where streams of order S2 — 1 are to be attached, then streams of order Si — 2, and on down to streams of first order. For each successive order w, 2Na+i streams will be required to form those of next higher order, and the remaining Ru = Nu — 2Nu+i "excess" streams can be attached in various ways. The last step is to specify the right-left arrangement at each junction; when this is done the topological properties are completely determined and a particular one of the W (Ni) TDCN has been constructed. Figure 1 shows an example of this procedure for the case NI = 28, N% = 8, A/3 = 3, W4 = 1. Previous discussions of the topological properties of natural networks have stopped at the first level, with classification on the basis of stream numbers. It appears now that a classification at the second level might be use- ful, since such a procedure would be inter- mediate to the rather crude description by stream numbers and the detailed description by the binary string. By classifying at the second level, the only topological feature that is ignored is the right-left arrangement at the junctions. Stated in another way, two TDCN belong to the same class if one can be converted into the other by reversals of the right-left order at one or more junctions. I propose here the tentative term "ambilateral" for this classification. Figure 2 shows the three ambi- lateral classes for NI = 5.

The ambilateral classification method appears to be more closely related to the hydrologic properties of stream networks than does the stream-number classification. First, although the hydrologic variables (discharge, slope, and so forth) at a given point in a drainage network may be expected to depend to some extent upon the topologic properties of the channel network above that point, there appears little reason to believe that they would be very sensitive to the right-left order of subnetworks at the junctions. Also, all members of a given ambilateral class have the same set of link magnitudes. (The magnitude of an interior link is the sum of the magnitudes of the two links contributing to it.) An example is shown in Table 2 for the six ambilateral classes for NI = 6; the first column specifies the class, the second column gives the number of members of each class, the third column gives the magnitudes /*,-, and the fourth column the values of 2m — l.The magnitudejttj is the number of sources upstream from the z'th link and (2jUj —1) is the number of links upstream from the z'th link (including itself). The values of (2/ij — 1) are closely correlated with the area upstream from a given link; if the drainage density is constant and if all links have the same length, then (2ju» — 1) is an exact measure of area. This fact is relevant because the values of many hydrologic and geomorphic variables at a point in a network (for example, discharge, slope) can be expressed empirically as a function of the area above that point. It therefore appears useful to have a classification scheme that separates networks into groups of identical (2n — 1) — values. For any but very small basins this set of n — values may be too unwieldy for practical use; however, they may prove advantageous in general arguments and derivations. A detailed discussion of the ambilateral classes and suggestions for their uses will appear in another publication. In this paper, the ambilateral classification will be used as an aid in detecting deviations from topologically random behavior. There are two principal reasons why systematic deviations from randomness might be expected to occur in natural channel networks. The first, already mentioned, is geological controls. If some close correlations between geological structure and topological and geometrical properties of networks could be established, then the geologist and geomorphologist would be provided with an important new approach to the description and interpretation of land forms. The results of previous investigations have not been encouraging in this respect, however. Few such correlations have been observed and where reported tend to be redundant. An example is the large values of Bz that occur in narrow strike valleys confined between long ridges (Strahler, 1964, p. 4-44).

The general physical laws for open systems may also cause some deviation from random behavior. Shreve (1967, p. 27) has previously remarked that the hypothesis of randomness is "by no means self-evident, for the possibility exists that certain of the topologically distinct networks might be promoted or inhibited . . . by the subtle interaction of the links in networks, . . . mediated by the interdependence of slope and channel processes." Leopold and Langbein (1962) and Langbein and Leopold (1964) have specifically suggested that the hydraulic geometry of a natural river is strongly influenced, although not completely determined, by certain general rules regarding the way in which gravitational potential energy is dissipated throughout the system. This idea can also be applied to the topologic properties of river networks. We shall use a highly idealized model to indicate how the topologic and hydrologic properties of a network may be related. Assume that we have a channel network for which (1) the drainage density is uniform, (2) all links have the same length, and (3) the area drained directly by each link contributes an amount Q0 to the total discharge (which is then (2Ni - !)£>„). This model will be used to calculate the total rate of doing work and the spatial variation of the rate of doing work, two energy expenditure parameters which Langbein and Leopold (1964) suggest are important in determining the properties of natural stream channels. Following Langbein and Leopold, we further assume that (4) the rate of doing work in the zth link is proportional to QiSi, where Qi is the discharge and s> the mean slope, (5) Si oc Qf, and (6) the rate of doing work per unit area of channel is proportional to (QiSi/Qf) = Qil+z" b . Then for any given network and WAi = Qil+*-<>, (8a) (8b) where WT is the total rate of work and WA is the rate per unit area of channel. From as- sumption 3, Qi = (2Mi - 1)0,. (9) The sums are over all links in the network and factors of proportionality in each equation have arbitrarily been set equal to unity. The procedure described above was suggested by some previous work of Howard (1968, personal commun.). Table 2 lists WT and a\WA} for each of the six ambilateral classes for NI = 6, where we have taken z = —0.75 and b = 0.5 as typical values of the exponents (Smart, 1969).

References:

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UNIT-6: QUANTITATIVE ANALYSIS OF CHANNEL PLANFORMS AND INDICES

The previous sections have considered the controlling influence of the driving variables and boundary conditions on channel form and behaviour. Each of these controls varies across a continuous range. For instance, slopes range from steep to gentle, valleys from confined to unconfined, and sediment loads from suspended to mixed to bedload dominated. Many different combinations are possible, leading to the immense variety of fluvial forms and behaviour that is seen globally. The continuum of alluvial channel types is illustrated in Figure 8.10. In general terms, different channel types exist along an energy gradient, ranging from highenergy braided channels through meandering and straight to low-energy anastomosing channels (a sub-set of anabranching channels). Floodouts and chainsof-ponds are found in low-gradient arid environments, where downstream reductions in discharge result in a dwindling supply of energy. This continuum can be related to the channel controls, since stream power integrates channel slope and flow regime. It also influences the type of load that the channel can carry, which in turn determines the substrate and stability of the channel. Like all channel classifications, the one shown in Figure 8.10 is, by necessity, a simplification of reality. While some channel reaches typify, say, a braided form, many have characteristics that are associated with more than one type. In fact, it has been suggested that channels with an intermediate form might be the norm rather than the exception (Ferguson, 1987). Although there is a continuum of forms, thresholds do exist between them. For example, there is a meandering– braiding threshold, above which rivers braid and below which they meander. Rivers that are close to this threshold, such as many of those in the South Island of New Zealand, have alternating meandering and braided reaches. For example, the Rangitata River rises in the Southern Alps, passing through a bedrock gorge as it leaves the mountains, before flowing across the Canturbury Plains. Above the gorge, the channel is braided. Below the gorge, and with a reduced sediment load, the Rangitata meanders, in contrast to the other, braided, rivers that flow across the Canterbury Plains. However, further downstream, the river cuts into unconsolidated Pleistocene terrace deposits and reverts to a braided form (Schumm, 1979). There are a number of management implications associated with such thresholds. Rivers that are close to a threshold may be particularly sensitive to variations in flow or sediment supply, or to channel engineering works, which could result in dramatic changes in form and behaviour. Careful management can (intentionally) convert a transitional braided reach to a more stable meandering form, for example by stabilising banks with vegetation or isolating sediment supplies. In qualitative terms, it is possible to identify rivers that are close to the braided–meandering threshold, since they display a number of forms that can be recognised as transitional. Conversely, a ‘typical’ meandering channel that does not have transitional features can be assumed to be some distance from the threshold. There have been a number of attempts to define this threshold in quantitative terms, by developing relationships using key variables such as bankfull discharge and channel slope (which control stream power) and the size of bedload. However, because channel form is influenced by so many different variables, this is extremely difficult to do. For a review of quantitative definitions of the braided–meandering threshold see Thorne (1997). Straight and meandering channels 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. 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.

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/w 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, 2007).

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UNIT-7: MEANDER INDICES: SHAPE, FORM, AND TIGHTNESS

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).

UNIT-8: CHANNEL BED TOPOGRAPHY: IDENTIFICATION AND ANALYSIS OF CHANNEL GEOMORPHIC UNITS

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.

UNIT-9: EFFICIENCY OF CHANNEL CROSS-SECTION: CONCEPT AND CHARACTERISTICS

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: 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 purposes 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 converted 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.

UNIT-10: HYDROGRAPH ANALYSIS AND COMPUTATION OF UNIT HYDROGRAPHS

One of the most important tasks in hydrology is to analyse streamflow data. These data are continuous records of discharge, frequently measured in permanent structures such as flumes and weirs (see Chapter 5). Analysis of these data provides us with three important features: • description of a flow regime • potential for comparison between rivers, and • prediction of possible future river flows. There are well-established techniques available to achieve these, although they are not universally applied in the same manner. This chapter sets out three important methods of analysing streamflow: hydrograph analysis, flow duration curves and frequency analysis. These three techniques are explained with reference to worked examples, all drawn from the same data set. The use of data from within the same study area is important for comparison between the techniques.

HYDROGRAPH ANALYSIS

A hydrograph is a continuous record of stream or river discharge (see Figure 5.1). It is a basic working unit for a hydrologist to understand and interpret. The shape of a hydrograph is a response from a particular catchment to a series of uniqueconditions, ranging from the underlying geology and catchment shape to the antecedent wetness and storm duration. The temporal and spatial variations in these underlying conditions make it highly unlikely that two hydrographs will ever be the same. Although there is great variation in the shape of a hydrograph there are common characteristics of a storm hydrograph that can be recognised. These have been described at the start of Chapter 5 where terms such as rising limb, falling limb, recession limb and baseflow are explained. Hydrograph separation The separation of a hydrograph into baseflow and stormflow is a common task, although never easy. The idea of hydrograph separation is to distinguish between stormflow and baseflow so that the amount of water resulting from a storm can be calculated. Sometimes further assumptions are made concerning where the water in each component has come from (i.e. groundwater and overland flow) but, as explained in the previous chapter, this is controversial. The simplest form of hydrograph separation is to draw a straight, level line from the point where the hydrograph starts rising until the stream discharge reaches the same level again (dashed line in Figure 6.1). However, this is frequently problematic as the stream may not return to its pre-storm level before another storm arrives. Equally the storm may recharge the baseflow enough so that the level is raised after the storm (as shown in Figure 6.1).

To overcome the problem of a level baseflow separation a point has to be chosen on the receding limb where it is decided that the discharge has returned to baseflow. Exactly where this point will be is not easy to determine. By convention the point is taken where the recession limb fits an exponential curve. This can be detected by plotting the natural log (ln) of discharge (Q) and noting where this line becomes straight. The line drawn between the start and ‘end’ of a storm may be straight (dotted line, see Figure 6.1) or curved (thin solid line, see Figure 6.1) depending on the preference of hydrologist Arnold et al. (1995)

provides a summary of different automated techniques. In very large catchments equation 6.1 can be applied to derive the time where stormflow ends. This is the fixed time method which gives the time from peak flow to the end of stormflow:

$$T=D^n$$

where D is the drainage area and n is a recession constant. When D is in square miles and in days, the value of n has been found to be approximately 0.2. The problem with hydrograph separation is that the technique is highly subjective. There is no physical reasoning why the 'end' of a storm should be when the recession limb fits an exponential curve; it is pure convention. Equally the shape of the curve between start and 'end' has no physical reasoning. It does not address the debate covered in Chapter 5: where does the stormflow water come from? Furey and Gupta (2001) have recently provided a hydrograph separation technique that ties into physical characteristics of a catchment and therefore is not as subjective as other techniques, although it still requires considerable interpretation by the user. What hydrograph separation does offer is a means of separating stormflow from baseflow, something that is needed for the use of the unit hydrograph (see pp. 103–106), and may be useful for hydrological interpretation and description.

The unit hydrograph

The idea of a unit hydrograph was first put forward by Sherman, an American engineer working in the 1920s and 1930s. The idea behind the unit hydro graph is simple enough, although it is a somewhat tedious exercise to derive one for a catchment. The fundamental concept of the unit hydrograph is that the shape of a storm hydrograph is determined by the physical characteristics of the catchment. The majority of those physical characteristics are static in time, therefore if you can find an average hydro graph for a particular storm size then you can use that to predict other storm events. In short: two identical rainfall events that fall on a catchment with exactly the same antecedent conditions should produce identical hydrographs. With the unit hydrograph a hydrologist is trying to predict a future storm hydrograph that will result from a particular storm. This is particularly useful as it gives more than just the peak runoff volume and includes the temporal variation in discharge. Sherman (1932) defines a unit hydrograph as 'the hydrograph of surface runoff resulting from effective rainfall falling in a unit of time such as 1 hour or 1 day'. The term effective rainfall is taken to be that rainfall that contributes to the storm hydrograph. This is often assumed to be the rainfall that does not infiltrate the soil and moves into the stream as overland flow. This is infiltration excess or Hortonian overland flow. Sherman's ideas fitted well with those of Horton. Sherman assumed that the 'surface runoff is produced uniformly in space and time over the total catchment area'. Deriving the unit hydrograph: step 1 Take historical rainfall and streamflow records for a catchment and separate out a selection of typical single-peaked storm hydrographs. It is important that they are separate storms as the method assumes that one runoff event does not affect another. For each of these storm events separate the baseflow from the stormflow; that is, hydrograph separation (see p. 102). This will give you a series of storm hydro graphs (without the baseflow component) for a corresponding rainfall event. Deriving the unit hydrograph: step 2 Take a single storm hydrograph and find out the total volume of water that contributed to the storm. This can be done either by measuring the area under the stormflow hydrograph or as an integral of the curve. If you then divide the total volume in the storm by the catchment area, you have the runoff as a water equivalent depth. If this is assumed to have occurred uniformly over space and time within the catchment then you can assume it is equal to the effective rainfall. This is an important assumption of the method: that the effective rainfall is equal to the water equivalent depth of storm runoff. It is also assumed that the effective rainfall all occurred during the height of the storm (i.e. during the period of highest rainfall intensity). That period of high rainfall intensity becomes the time for the unit hydrograph. Deriving the unit hydrograph: step 3 The unit hydrograph is the stormflow that results from one unit of effective rainfall. To derive this you need to divide the values of stormflow (i.e. each value on the storm hydrograph) by the amount of effective rainfall (from step 2) to give the unit hydrograph. This is the discharge per millimetre of effective rainfall during the time unit. Deriving the unit hydrograph: step 4 Repeat steps 2 and 3 for all of the typical hydro graphs. Then create an average unit hydrograph by merging the curves together. This is achieved by averaging the value of stormflow for each unit of time for every derived unit hydrograph. It is also possible to derive different unit hydrographs for different rain durations and intensities, but this is not covered here (see Maidment, 1992, or Shaw, 1994, for more details).

Using the unit hydrograph

The unit hydrograph obtained from the steps described here theoretically gives you the runoff that can be expected per mm of effective rainfall in one hour. In order to use the unit hydrograph for predicting a storm it is necessary to estimate the 'effective rainfall' that will result from the storm rainfall. This is not an easy task and is one of the main hurdles in using the method. In deriving the unit hydrograph the assumption has been made that 'effective rainfall' is the rainfall which does not infiltrate but is routed to the stream as overland flow (Hortonian). The same assumption has to be made when utilising the unit hydrograph. To do this it is necessary to have some indication of the infiltration characteristics for the catchment concerned, and also of the antecedent soil moisture conditions. The former can be achieved through field experimentation and the latter through the use of an ante cedent precipitation index (API). Engineering textbooks give examples of how to use the API to derive effective rainfall. The idea is that antecedent soil moisture is controlled by how long ago rain has fallen and how large that event was. The wetter a catchment is prior to a storm, the more effective rainfall will be produced. Once the effective rainfall has been established it is a relatively simple task to add the resultant unit hydrographs together to form the resultant storm hydrograph. The worked example shows how this procedure is carried out. Limitations of the unit hydrograph The unit hydrograph has several assumptions that at first appearance would seem to make it inapplicable in many situations. The assumptions can be summarised as: • The runoff that makes up stormflow is derived from infiltration excess (Hortonian) overland flow. As described in Chapter 5, this is not a reasonable assumption to make in many areas of the world. • That the surface runoff occurs uniformly over the catchment because the rainfall is uniform over the catchment. Another assumption that is difficult to justify. • The relationship between effective rainfall and surface runoff does not vary with time (i.e. the hydrograph shape remains the same between the data period of derivation and prediction). This would assume no land-use change within the catchment, as this could well affect the storm hydrograph shape. Given the assumptions listed above it would seem extremely foolhardy to use the unit hydrograph as a predictive tool. However, the unit hydrograph has been used successfully for many years in numerous different hydrological situations. It is a very simple method of deriving a storm hydrograph from a relatively small amount of data. The fact that it does work (i.e. produces meaningful predictions of storm hydrographs), despite being theoretically flawed, would seem to raise questions about our under standing of hydrological processes. The answer to why it works may well lie in the way that it is applied, especially the use of effective rainfall. This is a nebulous concept that is difficult to describe from field measurements. It is possible that in moving from actual to effective rainfall there is a blurring of processes that discounts some of the assumptions listed above. The unit hydrograph is a black box model of stormflow (see end of this chapter) and as such hides many different processes within. The simple concept that the hydrograph shape is a reflection of the static characteristics and all the dynamic processes going on in a catchment makes it highly applicable but less able to be explained in terms of hydrological theory. The synthetic unit hydrograph The synthetic unit hydrograph is an attempt to derive the unit hydrograph from measurable catchment characteristics rather than from flow data. This is highly desirable as it would give the opportunity to predict stormflows when having noaround the world. The Institute of Hydrology in the UK carried out an extensive study into producing synthetic unit hydrographs for catchments, based on factors such as the catchment size, degree of urbanisation, etc. (NERC, 1975). They produced a series of multiple regression equations to predict peak runoff amount, time to peak flow, and the time to the end of the recession limb based on the measurable characteristics. Although this has been carried out relatively successfully it is only applicable to the UK as that is where the derivative data was from. In another climatic area the hydrological response is likely to be different for a similar catchment. The UK is a relatively homogeneous climatic area with a dense network of river flow gauging, which allowed the study to be carried out. In areas of the world with great heterogeneity in climate and sparse river monitoring it would be extremely difficult to use this approach(Davie, 2002).

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UNIT-11: TEXTURAL ANALYSIS OF RIVER SEDIMENTS AND PEBBLES

Sediment calibre refers to the size of material that is available to be carried by a river. Grain size exerts an influence upon particle entrainment, modes of sediment transport, distances travelled and patterns of deposition. Rivers sort their load in longitudinal, lateral and vertical directions, giving rise to characteristic morphological traits. The Wentworth scale is the primary framework that is used to define sediment calibre (Table 6.1). This scale differentiates among boulders, cobbles, gravel, sand, silt and clay. For materials coarser than sand-sized particles, the b-axis or intermediate axis, i.e. the axis perpendicular to the longest (orthogonal) axis, is usually measured in the field. Of the three primary axes, the b-axis most closely reflects the weight of a particle. Point samples can be obtained by defining a grid for a particular depositional feature (locale) on the bed/bar surface and systematically measuring b-axes for 100 particles or more at each node of the grid. Alternatively, a bulk sample can be removed from the bed/bar and sieved at 0.5ø interval in the field. Each bulk sample should be sufficiently large such that the largest stone in the sample is not more than 1 % of the total sample weight. Visual-estimation grain-size charts can be used to assess mean and maximum grain sizes for sand-size particles at 0.5ø grain-size intervals. Alternatively, samples are returned to the laboratory for more detailed grain-size analysis. For silt and clay fractions, field texturing can be used to differentiate among grain-size classes (Table 6.2), or an array of laboratory procedures can be applied. Unlike coarse sediments, cohesive sediments are not amenable to classification by grain size and distribution. Rather, complex particle bonds affect the properties of cohesive sediments, and associated behaviour in terms of erosion, deposition and resuspension. Physical, electrochemical and biological effects interact to affect these behavioural properties. Physical factors affecting erodibility include clay content, water content, clay type, temperature, bulk density and pore pressure. Physico-chemical properties of the overlying fluid also affect erodibility. The chemistry of the eroding fluid and the pore fluid (e.g. pH, salinity, cation exchange capacity, sodium adsorption ratio) influences the valence of clay particles. As such, it plays a critical role in interparticle bonding, thereby influencing the erosion rate and the degree of flocculation of the clay– water suspension. Natural organic matter, measured as the percentage of organic carbon adsorbed to clay particles in the sediment, can increase the interparticle bonding, thereby increasing resistance to erosion. Sediment size is often quoted as a characteristic grain diameter D, such as the median D50 or the D84 (the grain diameter at which 84 % of the material is finer). Field measurements are required to derive the grain-size distribution plot from which the D50, D84 and other grain-size statistics can be determined. The geometric standard deviation of sediment sizes σg provides a useful guide to the mix of grain sizes (i.e. the degree of sorting). This is computed as (D84/D16) 0.5. If σg < 1.3, the sediment mix is considered uniform. For more detailed investigations, cumulative frequency plots of grain-size distributions are produced (Figure 6.1). This commonly entails a combination of field and laboratory procedures to textural analysis. Alluvial rivers can be broadly divided into two types: sand-bed streams have surface median size D50 in the range 0.0625–2 mm, while for gravel-bed streams 2 < D50 < 256 mm. For simplicity, cobble- and boulder-bed streams are integrated into the latter category. The dividing line between sand- and gravel-bed streams is not arbitrary; streams with a characteristic size between 2 and 16 mm (pea gravel) are relatively rare. Sand and silt often move through a gravel-bed river as throughput load during floods, with little interplay with the beds beyond partial filling of the interstices of newly deposited gravels. When the concentrations of these 'fines' are too high, or when the flow velocities are too low to prevent excess accumulation within the gravel framework, the gravels can become polluted with fines. The grain-size distributions of most sandbed streams are unimodal and can often be approximated with a normal distribution function. However, many gravelbed rivers have bimodal grain-size distributions, with both a gravel mode and a sand mode, but a paucity of pea-gravel size sediment (2–16 mm). Bed material load is that part of the sediment load that exchanges with the bed (and thus contributes to morphodynamics). Wash load is transported without exchange with the bed. In other words, the wash load of a river consists of sediment moving in suspension that is too fine to be present in measurable fractions in the bed. Wash load is more properly termed 'floodplain material load' because it exchanges with the floodplain. In rivers, material finer than 0.0625 mm (silt and clay; i.e. mud with D < 0.062 mm) is often approximated as wash load. Bed material load is further subdivided into bedload and suspended load. Bedload refers to movement of material by sliding, rolling or saltating in a trajectory just above the channel bed (see below). Turbulence plays an indirect role in this motion. In contrast, suspended load is subjected to the direct dispersive effect of turbulent eddies within flow, such that particles may be moved high into the water column. Low-slope sand-bed rivers move their bed material load (typically sand) as both bedload and suspended load, but suspended load far dominates bedload at the flood conditions that transport most of the sediment. In most large, low-slope sand-bed streams, mud comprises the great majority of the sediment transported on a mean annual basis. Bedload, suspended-load and mixed-load rivers are differentiated on the basis of the relative proportion of grain sizes carried within the flow or along the channel bed (Schumm, 1968). Bedload rivers carry more than 11 % of their load as sand-sized grains or larger along the channel bed. These rivers tend to have a high width/depth ratio with non-cohesive banks and loose, coarse materials on the bed. Suspended-load rivers carry the vast proportion of their load as fine sand, silt and clay materials that are suspended in the body of the flow. Bedload is Cohesive sediments are closely linked to water quality. Many pollutants, such as heavy metals, pesticides and nutrients, preferentially adsorb to cohesive sediments. In addition to the contaminants absorbed to the sediments, the sediments themselves are sometimes a water quality concern. The turbidity caused by sediment particles can restrict the penetration of sunlight and decrease food availability, thus affecting aquatic life. Sediment coarser than 62 μm is coarse, non-cohesive material. Sediment sizes smaller than 2 μm (clay) are generally considered cohesive sediment. Silt (2–62 μm) is considered to be between cohesive and non-cohesive sediment. Indeed, the cohesive properties of silt are considered due to the existence of clay. Thus, in practice, silt and clay are both considered to be cohesive sediment. For sediment containing more than approximately 10 % clay, the clay particles control the sediment properties. Cohesive sediments consist of inorganic minerals and organic material. Inorganic minerals consist of clay minerals (e.g. silica, alumina, montmorillonite, illite and kaolinite) and nonclay minerals (e.g. quartz, carbonates, feldspar and mica, among others). The organic materials may exist as plant and animal detritus and bacteria.

Phases of sediment movement along rivers: the Hjulström diagram Bed materials move intermittently and recurrently through river systems, in a similar manner to a 'jerky conveyor belt'. Sediment is subjected to three phases of movement: entrainment, transport and deposition. The Hjulström diagram conceptualises the circumstances under which each of these phases operates for sediments of different calibre under variable velocity conditions (Figure 6.2). Entrainment is the process by which grains are picked up or plucked from the bed of a river. Transport is defined as the movement of sediment on the channel bed or within flow. Deposition occurs when energy is no longer sufficient to maintain transport and sediment is stored along the river. The threshold between entrainment and transport demarcates the flow velocity and sediment size conditions under which sediment is either picked up (entrained) or remains stationary. Given the dynamics of flow in natural channels, the entrainment threshold spans a range of flow velocities and is represented by a band on the Hjulström diagram. A sharply dipping threshold separates transport and deposition domains. This threshold depicts the grain size and velocity conditions under which sediment is removed from transport and is deposited. This is defined in relation to the fall velocity, the velocity at which sediments fall out of flow and are deposited (see below). The low point for the threshold between entrainment and transport occurs for medium-sized sands. These noncohesive grains are the most readily entrained sediments in river channels, with a threshold velocity of around 0.2 $m\ s^{-1}$. Other sand-sized materials are entrained at velocities between 0.2 and 0.4 $m\ s^{-1}$. Velocities greater than 1 $m\ s^{-1}$ are required to entrain coarser clasts. Electrochemical cohesive properties ensure that the silt–clay fraction also requires high flow velocities to be entrained. Once entrained, turbulence maintains silt and clay materials in transport under a wide range of flow velocities, such that they can be transported considerable distances. A significant drop in velocity (or even standing water with velocities.

Entrainment of sediment in river channels

Entrainment processes detach grains from a surrounding surfaces making them available to be transported. Detachment occurs via a number of mechanisms. Corrasion is the mechanical (hydraulic and abrasive) action of water on a surface. Cavitation occurs when pressure differentials caused by shock waves generated by the collapse of vapour pockets detach particles from a surface. Entrainment occurs at higher flow velocities than both transport and deposition. This is because the impelling forces of the flow must overcome the resistance forces acting on the grain, including friction (the weight of particle, its roughness and interlocking), and cohesion (the electrochemical and surface tension forces of the grain) (Figure 6.3). Grain density also exerts a critical influence upon entrainment. The two primary impelling forces that act to entrain particles are fluid drag and lift force. Fluid drag is exerted by an erosive agent (i.e. water) exerting a force in the direction of flow. Horizontal drag is affected by the velocity of the flow and flow density. The greater the velocity and viscosity of flow, the greater the potential for drag to exert a force on the grain. In addition, a vertical lift force is required. Vertical lift is affected by flow turbulence and buoyancy of the particle. The pressure gradient induced by the difference in flow velocity between the top and bottom of a grain acts to lift the particle vertically. Turbulent eddying may also induce vertical velocity components close to the bed. However, these forces decrease rapidly away from the bed as velocity and pressure gradients diminish. Therefore, grains that protrude from the channel bed have more vertical lift exerted upon them than particles which are embedded or interlocked in surrounding substrate. The downslope component of movement is affected by particle weight and slope angle. Once the combined lift and drag forces exceed the cohesion and friction forces, grains are raised off the channel bed and move into transport. The conditions under which initiation of motion occurs is called the entrainment threshold and can be measured as a critical velocity v_{cr}, as depicted in the Hjulstrom diagram (Figure 6.2), or as a critical shear stress τ_c (Figure 6.4). In general, the critical bed shear stress τ_c required to move grains increases with grain size (Figure 6.4). However, the shear stress at which the grain will actually move can differ by an order of magnitude. For example, a grain that is 100 mm in size could move under a shear stress as low as 20 N m^{-2} , but is more likely to move at shear stresses >100 N m^{-2} . This variance reflects instantaneous stresses within the flow, bed roughness and whether the grain is loose or embedded.

Frictional resistance influences the efficiency of entrainment in natural stream beds. For coarse sediment, grain size is the key control. Other controls include bed packing, armouring and hiding (see below). For fine-grained sediment, cohesion is the key control on the efficiency of entrainment, but sheltering within the laminar layer also occurs. Other controls include the presence of in-channel roughness elements (e.g. vegetation, wood) that reduce energy and/or increase bed strength. These additional resistance forces must be overcome before entrainment can occur. Higher velocity or shear stress is required to entrain sediments that are well protected by these elements. The Shields number is a dimensionless parameter with which to quantify sediment mobility. The Shields equation is often used to calculate the entrainment threshold of grains of various sizes. This equation relates dimensionless critical shear stress τ_c (i.e. bed drag force acting on the flow per unit bed area; see Chapter 5) to grain size and density of grain packing (i.e. embeddedness in the channel bed or submersion in the laminar layer). For grains to move on a channel bed, the boundary shear stress τ_0 must exceed the critical shear stress τ_c for a grain of a given size. However, this relationship cannot be used for silt- and clay-sized materials because these sediments are affected by electrostatic forces. The Shields equation is: τ τ ρ ∗ = c RgD where τ * is the Shields parameter, τc (N m^{-2}) is the critical bed shear stress, R is the submerged specific gravity, measured as (ρs/ρ) − 1, where ρs is the density of sediment (assumed to be constant at 2650 kg m^{-3}) and ρ is the density of flow (1000 kg m^{-3}), g is the acceleration due to gravity (9.81 m s^{-2}) and D (mm) is the characteristic grain size. The Shields parameter can be interpreted as a ratio scaling the impelling force of flow drag acting on a particle to the force resisting motion acting on the same particle. The threshold of motion for a river bed composed of grains of characteristic size D and submerged specific gravity R and subjected to bed shear stress τb is quantified by the modified Shields curve: τc p Re ∗ = +× − − − 0 5 0 22 0 06 10 0 6 7 7 0 6 .(. .) . . Re . p where τc ∗ is the critical Shields number above which motion starts and Rep is the particle Reynolds number, given as: Rep = RgD D v where R is the submerged specific gravity (where sediment density ρs is assumed to be constant at 2650 kg m^{-3} and water density ρ = 1000 kg m^{-3}), g is the acceleration due to gravity (9.8 m s^{-2}) and v (m s^{-1}) is velocity. When first mobilised, the particles roll, slide or saltate close to the bed as bedload. The Shields diagram (Figure 6.5) shows how the entrainment threshold varies for bedload, mixed-load and suspended-load rivers. On a smooth surface with a low Reynolds number (Re), small grains (i.e. suspended load particles For coarser bedload fractions, grain roughness is often the dominant component of resistance on a channel bed, acting as an impediment to transport. This is particularly evident if the bed material consists of gravel (2–64 mm) or cobbles (64–256 mm). A simple measure of the ability of flow to transport sediment of a certain calibre is represented by the ratio: d D where D (m) is the characteristic grain size and d (m) is the flow depth. As flow depth increases, the effect of grain roughness is drowned out such that entrainment and transport can occur. In small channels d/D < 1, such that flow depth is less than the size of the bed material and flow occurs around large clasts. Individual cobbles and boulders protrude from the water surface and mobilisation of bed sediment is extremely unlikely. In these instances the bed material may be organised as a series of packing arrangements that further inhibit mobility (e.g. step–pool sequences and cascades – see Chapter 8). In intermediate channels 10 > d/D > 1, such that flow depth is up to 10 times greater than bed material size. Individual clasts on the channel bed are submerged and bed material can be mobilised under a range of flows. In these settings, the bed is typically organised in a series of pools and riffles, the position and morphology of which reflect adjustments in river planform. In large channels d/D > 10 and the depth of flow is over 10 times greater than the bed material size. Under these conditions, bed material is mobile and a range of well-defined morphologies results. The pattern of units and the ability of flow to rework sediments reflect the calibre and volume of sediment supply and discharge. Surface erosion of cohesive sediments occurs as individual particles or small aggregates are removed from the body of materials by hydraulic (hydrodynamic) forces such as drag and lift. The ability of a cohesive sediment to resist surface erosion is known as erosional strength. Resistance to surface erosion differs from resistance to mass erosion. Mass erosion is determined by the undrained strength of the sediment, or yield strength. Mass erosion occurs when the yield strength is exceeded. Examples of this mechanism include slip failure of a streambank (see Chapter 7) or when large chunks of sediment are eroded from the streambed. There is a difference of one to three orders of magnitude between erosional strength and yield strength. Biological effects work alongside physical and electrochemical effects in determining the behaviour of cohesive materials. Biogenic stabilisation or biostabilisation refers to situations whereby biological action directly or indirectly induces a decrease in sediment erodibility. Discrete particles may become covered by bacteria and diatom growth, causing cohesion to increase and roughness to decrease. Further binding is caused by bacterial secretion that forms cohesive networks between the diatoms. In this way, originally non-cohesive materials may become biostabilised. Some organisms may grow on the bed surface, filling the interparticle voids and forming a microbial mat, thereby creating a smooth, protective biofilm and reducing the hydraulic roughness. This decreases the stress in the nearbed margin, thereby strengthening the bed by effectively increasing the velocity at which particles are entrained. Alternatively, some organisms rework sediments by bioturbation or create uneven surfaces with protrusions that increase hydraulic roughness. Burrowing organisms may have positive or negative effects on the stability of surficial soils. For example, Oligochaeta (burrowing worms) may reduce the critical stress for erosion 10-fold. Chironomids (common midges with burrowing larva) also have a negative effect on sediment stability, but this influence may diminish over time as the organisms excrete mucus and develop tube houses, cementing the bed and making it less erosive. Elsewhere, burrowing organisms may strengthen the bed by locally increasing the critical shear stress of sediments. These are complex biogeochemical interactions. Site-specific variability is common and, indeed, may change over differing timeframes (Fryirs & Brierley, 2013).

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UNIT-12: VULNERABILITY ANALYSIS OF FLOODS AND RIVERBANK EROSION

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

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

You have probably heard reference to the 'twenty-year flood' or the '100-year flood'. This return periodis an estimate of how often a flood of a given size can be expected to occur and, since less frequent floods are more extreme, the 100-year event would be bigger than the twenty-year flood. The return period (T) can also be expressed as a prob ability (P) by taking the inverse of the return period, i.e.:

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

Using this, the probability of a 100-year flood taking place in any one year can be calculated as0.01 (i.e. 1 per cent), and for the twenty-year flood, 0.05 (5 per cent). The probability that a flood with a particular return period will occur is the same every year and does not depend how long it was since a flood of this size last occurred – the twenty-year does not occur like clockwork every twenty years. However, if a period of several years is considered, the likelihood of a given flood occurring during this time increases. For example, if someone bought a house on the 100 year floodplain and lived there for thirty years, the probability of that property being flooded in any one year would be 0.01. This increases to 0.3 (probability × number of years), or 30 per cent, for the thirty-year period. Box 3.2 explains how return periods are estimated. As with any odds, flood probabilities are estimates, and a number of under lying assumptions are made when deriving them. It is assumed that runoff is randomly distributed through time and that the data set holds a representative sample of these random events. Estimates are therefore more reliable when a longer record is available, since a larger number of flood events will be included in it. Another assumption is that there are no long-term trends in the data, which is not the case when climate change is occurring.

The frequency of bankfull discharge

Although bedrock channels are mainly influenced by high magnitude flows, those formed in alluvium can be adjusted by a much greater range of flows (see Chapter 1, pp. 5–6). This is reflected by the morphology and size of alluvial channels. Over the years, much research has focused on the bankfull discharge (defined above), since it represents a distinct morphological discontinuity between in-bank and overbank flows. Leopold and Wolman (1957) suggested that the channel cross-section is adjusted to accommodate a discharge that recurs with a certain return period. From an examination of active floodplain rivers, they found that the bankfull discharge had a return period of between one and two years. This is corroborated by later observations made for stable alluvial rivers (for example, Andrews, 1980; Carling, 1988). However, the concept of a universal return period for bankfull dis charge that can be applied to all rivers is controversial. Williams (1978) observed wide variations in the frequency of bankfull discharge, which ranged from 1.01 to 32 years, and concluded that this was too variable to assume a uniform return period for all rivers. Even along the same river, there can be marked variations in the frequency of bankfull discharge (Pickup and Warner, 1976). The concept of a uniform frequency for bankfull dis charge assumes that all channels are 'in regime'. This means that the morphological characteristics of a given channel, such as size, fluctuate around a mean condition over the time scale considered (Pickup and Reiger, 1979). This is not true for all rivers and there are many examples of non-regime, or disequilibrium, channels. An example would be where channel incision is taking place through erosion of the channel bed. This results in a deeper channel, which requires a larger, and therefore less frequent, discharge to fill it. The Gila River in Arizona, United States, was greatly enlarged when past events had led to large floods. The enlarged channel is not adjusted to the contemporary flow regime, which means that the bankfull discharge for the enlarged channel has a much lower frequency (Stevens et al., 1975). The material forming the bed and banks is also significant. In cases where the bound ary is very erodible, the bankfull discharge may simply reflect the most recent flood event (Pickup and Warner, 1976).

The geomorphological effectivenes of floods

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

Regional flood frequency curves

The flood frequency–magnitude relationship differs between regions. Despite the low annual rainfall in dry land environments, precipitation can be highly variable and the twelve largest floods ever recorded in the United States all occurred in semi-arid or arid areas (Costa, 1987). During flash floods, such as the one shown in Colour Plate 14, floodwaters rapidly inundate the dry channel. Not all dryland rivers are prone to flash flooding however, and there is considerable variation in the size, type and duration of flooding. Regional flood frequency curves are shown in Figure 3.4. The return period is plotted on the horizontal axis using a logarithmic scale, with the relative flood magnitude on the vertical axis. A relative flood magnitude has been used to allow comparison between floods for a number of rivers in different regions. Because these all drain different areas, a direct compari son of flood magnitudes would not be very meaningful. Instead, for each river included in the analysis, the ratio between the magnitude of each flood on record and a low magnitude 'reference flow' – the mean annual flood – has been used. This is defined in Box 3.2 and has a return period of 2.33 years (i.e. the flow that will be equalled or exceeded on average once every 2.33 years2). The steepness of each curve reflects the variability of the flow, with arid zone rivers showing a much greater increase in relative flood magnitude at higher return periods. This reflects the extreme flow variability observed in these rivers and has important implications for the morphology of dryland channels, as will be seen in later chapters.

Reconstructing past floods

Palaeoflood hydrology is a new and developing area of hydrology and geomorphology, which reconstructs past flood events in order to extend the flow record. Due to problems associated with monitoring major floods and the relatively short duration of most gauged records, extreme floods are very rare in the observational record. By reconstructing palaeofloods, the flood record can be extended, allowing increased accuracy in the estimation of floods for risk analysis (Box 3.2). Evidence of pastflood events is provided by geological indicators such as flood deposits, silt lines and erosion lines along the channel and valley walls (Benito et al., 2004). Historical records are also used and include documents, chronicles and flood marks inscribed on bridges and buildings. Using this evidence, it is possible to determine the size of the largest flood events over periods of time ranging from decades to thousands of years (Benito et al., 2004). As well as identifying the largest floods, evidence of floods above or below above specified flow stages can also be reconstructed (Stedinger and Baker, 1987). Although time-consuming, it is possible to reconstruct a completerecord, chronicling the largest flood, together with the size and number of intermediate palaeofloods (Benito et al., 2004). Chapter 9 discusses some of the techniques that are used in reconstructing past flood events(Charlton, 2008).

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