Two Years

Post Graduate Degree Programme (CBCS) in Geography

Semester – III

Paper Code: GEO/DSE/FG/T-316

Paper: Fluvial Geomorphology-I (Special Paper)

Self-Learning Material

Directorate of Open and Distance Learning (DODL)
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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.) Kallol Paul, 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 –III

Paper Code: GEO/DSE/FG/T-316

Paper: Fluvial Geomorphology-I (Special Paper)

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

Unit-01	Scope, nature and significance of fluvial geomorphology; Scales in Fluvial geomorphology
Unit-02	Fluvial system: Components, input output, Variables of fluvial system: internal and external, adjustable and controlling factors
Unit-03	Linear, Areal and Altitudinal properties of drainage basin; Law of stream number and stream length, law of basin area;
Unit-04	Hydraulics of channel flow: Stream Energy; Types of flow; type of links, number of links
Unit-05	River velocity, factors and its distribution in open channels; Flow resistance, Chézy’s and Manning’s equation
Unit-06	Channel initiation and Evolution of channel pattern, importance of headward extension and branching, lateral expansion
Unit-07	Classification of natural streams by D. L. Rosgen
Unit-08	Erosion: threshold of erosion, processes of erosion, river bank erosion
Unit-09	Transportation: processes of entrainment, bedload transport dynamics; Channel competence
Unit-10	Deposition: factors controlling deposition, depositions along the channel and across the channel
Unit-11	Sediment deposits: nature and characteristics, flood plain and deltaic plain deposits
Unit-12	Fluvial processes and forms

Contents

1.1. Introduction

The term fluvial is derived from the Latin *fluvius*, meaning river. Fluvial geomorphology is the study of the interactions between river channel forms and processes at a range of space and time scales. The influence of past events is also significant in explaining the present form of river channels. Rivers are found in many different environments and show an amazing diversity of form. Rivers drain much of the land area – with the exception of regions that are hyperarid or permanently frozen – and their variety reflects the vast range of different environments in which they are found. Climate, geology, vegetation cover and topography are just some of the factors that influence river systems.

Rivers are found in many different climatic zones, ranging from humid to arid, and from equatorial to arctic. Some of the larger rivers even flow across different climatic zones, originating in a humid area before flowing through an arid region. Examples of these 'exotic' rivers include the Nile and Colorado, both of which sustain agriculture and urban centres in desert regions. Rivers are a much-cherished feature of the natural world. They perform countless vital functions in both societal and ecosystem terms, including personal water consumption, health and sanitation needs, agricultural, navigational, and industrial uses, and various aesthetic, cultural, spiritual, and recreational associations. In many parts of the world, human-induced degradation has profoundly altered the natural functioning of river systems. Sustained abuse has resulted in significant alarm for river health, defined as the ability of a river and its associated ecosystem to perform its natural functions. In a sense, river health is a measure of catchment health, which in turn provides an indication of environmental and societal health. It is increasingly recognized that ecosystem health is integral to human health and unless healthy rivers are maintained through ecologically sustainable practices, societal, cultural, and economic values are threatened and potentially compromised. Viewed in this way, our efforts to sustain healthy, living rivers provide a measure of societal health and our governance of the planet on which we live. It is scarcely surprising that concerns for river condition have been at the forefront of conservation and environmental movements across much of the planet.

This special paper of Fluvial Geomorphology aims introduces fundamental concepts of this discipline and to cultivate interest among the students to further explore this branch of geomorphology.

1.2. Learning Objectives

- Scope, nature and significance of fluvial geomorphology; Scales in Fluvial geomorphology
- Fluvial system: Components, input output, Variables of fluvial system: internal and external, adjustable and controlling factors
- Linear, Areal and Altitudinal properties of drainage basin; Law of stream number and stream length, law of basin area;
- Hydraulics of channel flow: Stream Energy; Types of flow; type of links, number of links
- River velocity, factors and its distribution in open channels; Flow resistance, Chézy's and Manning's equation
- Channel initiation and Evolution of channel pattern, importance of headward extension and branching, lateral expansion
- Classification of natural streams by D. L. Rosgen
- Erosion: threshold of erosion, processes of erosion, river bank erosion

- Transportation: processes of entrainment, bedload transport dynamics; Channel competence
- Deposition: factors controlling deposition, depositions along the channel and across the channel
- Sediment deposits: nature and characteristics, flood plain and deltaic plain deposits
- Fluvial processes and forms

1.3. Assessment of Prior Knowledge

Fundamental concepts of geomorphology may be discussed.

1.4. Learning activities

Classroom seminars/ discussions regarding various topics covered under this paper may be done

1.5. Feedback of learning activities

Class tests may be arranged.

1.6. Examples and Illustrations

UNIT-1: SCOPE, NATURE AND SIGNIFICANCE OF FLUVIAL GEOMORPHOLOGY; SCALES IN FLUVIAL GEOMORPHOLOGY

Fluvial geomorphology is concerned with the role of rivers in shaping the morphology of the Earth and as such its subject matter was central to the time bound approach to geomorphology that was initiated by W.M. Davis and dominated the first half of the twentieth century (Gregory, 2000). However this branch of geomorphology may not have advanced as cohesively as some other branches because of a major shift that occurred in its approach and also because of parallel developments in other fields including hydrology, sedimentology, palaeohydrology and limnology. The shift in approach and new developments after 1964 were styled the new fluvialism by Tinkler (1985), and these developments were largely instigated by the influence of Fluvial Processes in Geomorphology (Leopold et al., 1964). Those authors contended that 'Our emphasis on process is not intended to minimize the importance of the historical aspects of geomorphology. Unfortunately, because of the limited understanding of geomorphic processes and their associated landforms, we ourselves are unable at present to make a truly satisfactory translation from the dynamics of process to historical interpretation. Better future understanding of the relation of process and form will hopefully contribute to, not detract from, historical geomorphology' (Leopold et al., 1964, p. vii). This refocusing of fluvial geomorphology occurred when related disciplines were developing, especially hydrology, so that there was uncertainty about the precise role of fluvial geomorphology, when new techniques were being employed, models were being used more and the dominant emphasis upon processes was making the boundaries with other disciplines such as hydrology and sedimentology increasingly obscure. By 1988, when Graf (1988) defined geomorphology as the study of earth surface forms and process and commented that 'geomorphology is largely an intellectual child of the twentieth century', it was clear that fluvial geomorphology had benefited from assimilating studies of processes and had progressed in parallel with other new disciplines such as palaeohydrology. Thus Leeks et al. (1988, p. 221) commented that 'Whilst the momentum of fluvial geomorphology in the recent past has been in the field of process studies… …there have been constant reminders that historical events over a variety of timescales cannot be neglected'. Palaeohydrology was one multidisciplinary field which also profited from a widening range of techniques (Gregory, 1983) and concentrated research on particular areas including the temperate (Starkel et al., 1991) and the global (Gregory et al., 1995) scales so that it was subsequently able to focus upon the links with environmental change (Benito et al., 1998) and environmental management and global change (Gregory and Benito, 2003). Thus much of the fluvial research that was essentially geomorphological did not necessarily fall under the title of fluvial geomorphology because of the expansion of techniques and the way in which multidisciplinary research interacted with other fields. Therefore although fluvial research, including both landform development and processes, comprised 27.7% of British geomorphological research in 1975 (Gregory, 1978, 2000) and water-related research made up nearly 20% of publications submitted for the 1996 Research Assessment Exercise (RAE) in universities and colleges in the UK (Gregory, 2000), not all of this research was categorised as fluvial geomorphology. At the end of the 20th Century a number of applications of fluvial geomorphology as well as technical advances in computing (improving modeling capabilities and GIS) and data acquisition (e.g. remote sensing, GPS) served to instigate new advances of the discipline. According to Sear et al. (1995, p. 629) 'Fluvial geomorphology and river engineering are converging as each discipline realises the benefits of the other' seeing the benefits of a fluvial geomorphology approach which links site-specific boundary conditions to predictions of morphological behaviour over realistic time and spatial scales. In the conclusion of Applied Fluvial Geomorphology for River Engineering and Management (Thorne et al., 1997a) four main areas of difficulty were identified that limit the widespread application of a geomorphological approach to improved river management and engineering (Thorne et al., 1997b) namely: geomorphology is relatively new; geomorphology is unlike open channel hydraulics in that it involves the collection of a wide range of information as a field science but uses GIS and other sources; geomorphological advice is often associated with that advice given by conservationists and single issue pressure groups; geomorphology operates on a longer time frame than many applied sciences. When reviewing the background to river channel management, fluvial geomorphology was just one discipline listed for the contributions made with others including ecology, engineering, hydrology, physical geography, and environmental sciences extending to social sciences and philosophy (Downs and Gregory, 2004). Thus as fluvial geomorphology has been revitalised it is in a more multidisciplinary position so that Newson (2006, p. 1606) suggested for example that 'Fluvial geomorphology is rapidly becoming centrally involved in practical applications to support the agenda of sustainable river basin management'. In the Dictionary of Physical Geography fluvial geomorphology is defined as 'the study of the morphology of environments worked by rivers' (Thomas and Goudie, 2000). Morphology is certainly an essential ingredient but not the only one. Roy and Lane (2003) contend that geomorphologists have been content with the major upheaval in the 1950s and 1960s without undertaking the redefinition of core philosophical and methodological approaches that was done by other disciplines and other parts of geography in the 1970s and 1980s. They believe that some fluvial geomorphologists have been innovative and have changed their view of how rivers behave because there has been a 'growth of locally specific, intensive research into particular river channel reaches, as distinct from earlier research, which was grounded in the search for empirical regularities or laws' (Roy and Lane, 2003, p. 103) so that fluvial geomorphology is 'now increasingly grounded in the measurement and understanding of individual cases' (p. 103). Hence we can now accommodate the analysis of form once more because landform developments lagged behind the understanding of process in the quantitative study of earth's surface (Lane et al., 1998). However fluvial geomorphology now includes much more than study of fluvial forms and processes and Kondolf and Piegay (2003, p. 4) commented that 'We define fluvial geomorphology in its broadest sense, considering channel forms and processes, and interactions among channel, floodplain, network, and catchment… … we consider fluvial geomorphology at different spatial and temporal scales within a nested systems perspective…. Analysis of fluvial geomorphology can involve application of various approaches from reductionism to a holistic perspective, two extremes of a continuum of underlying scientific approach along which the scientist can choose tools according to the question posed'. Thus fluvial geomorphology has come to be characterised by a multidisciplinary approach, achieved by broadening of research methods and techniques thus enabling greater research achievements. Such progress (discussed in Section 3) has been achieved by the use of innovative research methods and tools (Section 2), such as the development of cellular approaches to modelling river form and process considered to represent one of the most important advances in fluvial geomorphology over the past decade (Nicholas, 2005). Section 4 of the paper introduces presentations made during the fluvial session of the 6th International Conference on Geomorphology (Zaragoza, Spain, September 2005), focusing on the papers presented in this special issue. 2. Innovative research methods and techniques Recent advances of fluvial geomorphology, in parallel with other Earth Sciences disciplines, are partly the result of new applications developed in response to progress in computing science and new techniques related to computational fluid dynamics, remote sensing, radiometric and isotopic methods for numerical dating, geophysical data acquisition and analysis, among others. Future prospects for fluvial geomorphology are very promising, with studies of long-term river evolution and palaeohydrology benefiting from new methods and techniques and becoming more robust building upon the results of five decades of advances in the study of river processes, experimental studies and numerical models. The lack of a linear response in river systems and the multiple temporal and spatial scale context required implies that to understand relationships or to solve problems typically requires application of multiple tools

(Kondolf and Piegay, 2003). These tools are required to operate at different temporal and spatial scales, from basin studies to particle movement, and from geologic timescales to instantaneous measurements during experimental studies. In this section we review some recent achievements in research methods and techniques relevant to (1) long-term fluvial changes, (2) computational fluid dynamics and sediment transport, and (3) ecological and management studies. 2.1. Analysis of long-term fluvial changes Fluvial deposits record the compound effects of channel migration, incision and aggradation involving a broad range of spatial and temporal scales. A further complexity for interpreting fluvial deposits relates to the different styles and rates of sedimentation within alluvial environments, which may respond to longterm effects of climate change and tectonic movements, or to short term intrinsic changes promoted by extreme events (e.g. floods) or anthropogenic causes. Analysis of alluvial deposits involves well established methods for sedimentology and stratigraphy (e.g. particle size, sedimentary structures, facies, provenance), geochronology (e.g. radiocarbon, OSL, cosmogenic isotopes, dendrochronology), soil science (profile description, mineralogy, soil-forming processes and soil chronosequences), surface hydrology and hydraulics (e.g. river hydrographs, peak discharge, flow velocity and stream power), and ecology (e.g. characterisation of riparian ecosystems) (see Jacobson et al., 2003). Interpretation of alluvial records presents a great challenge because the very nature of fluvial activity determines that only fragments of the deposited components usually survive for later interpretation (Lewin and Macklin, 2003). In most cases, the incomplete and partial fluvial record precludes use of standardised methodological procedures of record analysis and interpretation followed in the study of other depositional environments (e.g. lacustrine, marine) in which a more uniform deposition rate is assumed. Nevertheless, some fluvial depositional environments (e.g. floodplain lakes, abandoned channels, backswamp areas) can fulfil this requirement at least for specific time periods. Reconstruction of fluvial history from alluvial records requires the organisation of alluvial puzzle pieces into their temporal and spatial framework. Most recent advances in alluvial record interpretation are related to methodological achievements in numerical age dating (e.g. OSL, cosmogenic nuclide exposure) and geoprospection which allows higher precision in disentangling river history. The Optically Stimulated Luminescence (OSL) method (Stokes, 1999) is a dating technique which indicates the burial time of Quaternary deposits. Alluvial deposits are normally not ideal for optical dating because of inadequate sunlight exposure of the grains prior to burial. This has resulted either in an overestimation of the burial age and/or a high variability of the results with 30–50% standard error (Porat et al., 1996; Greenbaum et al., 2000). However, new developments in instrumentation, reducing the sample size to individual grains (Duller and Murray, 2000; Bøtter- Jensen et al., 2000), and with new analytical protocols such as the single-aliquot regenerative-dose (SAR) for determining the equivalent dose (Murray and Wintle, 2000; Wintle and Murray, 2006) provide very accurate numerical dating, with age uncertainties of 5–10%, even for young deposits (b300 years) during a period of major measurement errors for the radiocarbon dating technique (Ballarini et al., 2003; Duller, 2004). Recent research has also highlighted the importance of selecting suitable sample locations (Rodnight et al., 2006). Another technique recently applied to the dating of Quaternary river terraces, specifically in dryland regions, is the U-series dating of calcrete formed within alluvial terrace deposits (Candy et al., 2004). The Useries isochron approach involves the extraction of multiple subsamples from a single horizon and analysing the U/Th isotopic ratios of each subsample. Such dating of fluvial depositional environments has led to advances in understanding river changes and the operational time of depositional environments, processes and related controls (e.g. environmental, climatic, tectonic, anthropogenic). In fact, the numerical dating of fluvial deposits and landforms has become a basic tool for relating the fluvial response at different spatial and temporal scales to the climatic, tectonic and human controlling factors, which is a central theme of fluvial geomorphology (Schumm, 1977a,b). Over recent decades, radiocarbon dating of alluvial deposits has been a common methodological procedure and, as a result, large datasets are available in the literature. A standardised method to compare radiocarbon dated alluvial records was first employed across Britain (Macklin and Lewin, 2003; Macklin et al., 2005; Johnstone et al., 2006) and has subsequently been applied to Spanish and Polish data sets (Thorndycraft and Benito, 2006a,b; Starkel et al., 2006; Macklin et al., 2006) to investigate the relationships between environmental change, flooding and Holocene river dynamics. Data analysis was undertaken on catchments of different size, type and land-use history, using radiocarbon dates from units representing a modification in sedimentation style or rate (Macklin et al., 2006). The graphical representation uses the sum of the individual probability distributions resulting from the OxCal (version 3.9; Bronk Ramsey, 1995, 2001) calibration programme. The record of alluvial activity in different catchments and countries could then be compared with a range of climate proxies and land-use change indicators, demonstrating the value of the database for reconstructing past hydrological events, as well as for predicting river response to future environmental change (Macklin et al., 2006; Gregory et al., 2006a). 2.2. Computational fluid dynamics and sediment transport In recent decades, computational fluid dynamics (CFD) has taken advantage of technical developments in computing allowing lower cost and more powerful desktop computers, capable of running large and complex simulations. Fundamental equations for CFD in river simulations were developed over 40 years ago, based on the resolution of a set of partial differential equations (Ingham and Ma, 2005), although its application to model morphology is more recent (e.g. Nezu and Nakagawa, 1993; Bates et al., 2005). In hydraulics, one of the essential components of the modelling process is topographic parameterisation, which, depending on its spatial resolution, may include the appropriate representation of small scale hydraulic processes, notably in terms of the strongly related issue of model discretization and boundary roughness (Lane and Ferguson, 2005). Therefore, the two major parametric unknowns in modelling fluvial flow are topography and bed roughness (see Horrit, 2005). Recent development of geospatial methodologies like global positioning systems (GPS), digital photogrammetry and high resolution ground and airborne remote sensing techniques (e.g. ALS, SAR, LIDAR) have substantially improved the collection of consistent and accurate surface topography at reasonable cost (Casas et al., 2006). Currently, one of the best techniques for the acquisition of densely spaced and highly accurate elevation data (15 cm and horizontal accuracies within 1/1000th of the flight height) is Airborne Light Detection and Ranging (LIDAR). The LIDAR is an active sensory system mounted in an airborne platform that uses laser light to measure distances between the sensor on the airborne platform and points on the ground (or a building, tree, etc.). LIDAR systems rely on the precise kinematic positioning provided by a differential global positioning system (dGPS) and inertial altitude determination provided by an inertial measurement unit (IMU), to produce horizontally and vertically accurate elevation measurements. Standard LIDAR survey creates 1,000,000 data collection points per 1 km2. Therefore, parameterisation data can be obtained from a measurement approach when the hydraulic model scales are comparable to measurement resolution, although in practice this parameterisation can be a more complex problem (Horrit, 2005). LIDAR has also been used for levee profiling, floodplain mapping (Charlton et al., 2003), hydraulic modelling (Mark and Bates, 2000; French, 2003), and topographic mapping of environmental or hazardous areas (McKean and Roeding, 2004). In riverine areas, LIDAR has a great potential for identifying fluvial geomorphological features through the examination of the microtopography of floodplain and terrace surfaces thus providing new insight for the understanding of channel migration, mapping of floodplain sedimentary environments and archaeological prospection (Carey et al., 2006). In this respect, LIDAR may be combined with geoelectrical methods such as 2D Ground Penetrating Radar (GPR) and/or with 3D electric resistivity tomography (ERT) surveys to provide three-dimensional architecture of the sediments infilling the valley floor (Brown, 2006). GPR and ERT provide detailed visualisation of the floodplain facies and channel infills (Birkhead et al., 1996; Corbeanu et al., 2001; Gaswirth et al., 2002; Baines et al., 2002), although these methods do not provide internal channel stratigraphy which needs to be completed with hand and mechanical coring. These new geoprospection techniques may improve future analysis and spatial resolution of fluvial landforms and sedimentary facies at and below the surface in order to generate 3D geomorphic models. LIDAR and SAR (synthetic aperture radar) interferometry have the drawback of being unable to map bathymetry at high resolution scales. A new system has been developed based on swath sonar bathymetry for channel mapping which uses a phase-difference interferometric technique. This technique is capable of mapping a swath width of 10 to 15 times water depth up to 30 m, with vertical accuracy of 10 cm for swaths less than 50 m wide, reproducing channel morphology with unprecedented resolution comparable to the LIDAR systems which improve the representation of small hydraulic features (Horrit et al., 2006). These technological developments open new opportunities for investigation of fluvial morphological changes and processes in the fluvial system. In sediment transport analysis, a sub-centimetre-scale elevation error may be required to study entrainment processes and sediment transport, which can only be obtained either by direct field measurements (Church et al., 1987) or by digital photogrammetric methods (Butle et al., 1998; Lane et al., 2001). Combined field measurements and digital observations of particle clusters on gravel beds are now capable of producing inferences about sediment transport processes and morphological changes between flow events (e.g. Wittenberg and Newson, 2005). An important element for interpreting the morphology and dynamics of the river-channel system is the identification of coarse sediment connectivity (Hooke, 2003). This implies a better understanding of sediment sources, movement and delivery (Higgitt and Lu, 2001; Walling et al., 2001). Sediment connectivity has become a focus of research, first on the coupling between hillslopes and channels (e.g. Harvey, 2001; Walling et al., 2001; Benda et al., 2005); and/or reach-to-reach channel connectivity (e.g. Harvey, 2002) in some cases recognising “sedimentation zones” and “source areas” within the channel system (Church and Jones, 1982), and showing that the sediment delivery problem (Walling, 1983) continues to be an unsolved issue. The methodological approach to assess connectivity can be addressed using field mapping of fluvial landforms, sedimentological evidence and calculation of the hydraulic conditions along the channel (e.g. Hooke, 2003). 2.3. Ecological and management studies Whereas geomorphic and ecologic landscape components had been conceptualized independently these can nowbemore integrated, for example in complexity theory (Stallins, 2006) and considerable multidisciplinary progress has been made towards combining ecology and the fluvial geomorphology of stream channels. This has been demonstrated by the range of flow types associated with identification of physical biotopes (Newson and Newson, 2000), with theways inwhich habitat is defined for fish or microinvertebrates (Urban and Daniels, 2006), and with the ecological evaluation of flow regimes (Bragg et al., 2005). Specific research has now shown how substrate and aggregate flow velocity behaviour interact with the distribution of aquatic plants (e.g. Gurnell et al., 2006a) with such studies having implications for channel restoration (e.g. Gurnell et al., 2006b). 3. Recent achievements in fluvial geomorphology Recent achievements in fluvial geomorphology have been greatly aided by the range of techniques that have become available, particularly because these techniques can give more immediate results. An indication of recent achievements is given in the progress reports published in Progress in Physical Geography and Dollar (2000) followed the lead of Rhoads (1994) in suggesting that fluvial geomorphology needs to be more related to fundamental scientific issues or to the solution of pressing societal problems. The subsequent reviews (Dollar, 2002), although acknowledging the recent emphasis upon applied interdisciplinary research, focus upon fundamental fluvial science with particular reference to the ways in which channel and valley properties respond to long-term effects of climate change and tectonic activity and in the short term on how information on the intrinsic characteristics and evolution of rivers is required to explain gradual processes. Two years later it was contended that fluvial geomorphology is in a stronger position than ever, that research has broadened and strengthened, that the contribution of fluvial geomorphology to resolving complex and interdisciplinary problems is now widely recognised, all occurring when much of the accumulated process knowledge should be used to bring longer-term and broader scale perspectives of landscape change back to prominence (Dollar, 2004). One year later the progress report focused on process-based fluvial geomorphology which exemplified the way in which research in fluvial geomorphology continues to be innovative, wide-ranging and dynamic (Hardy, 2005). In the light of these extensive summaries here we highlight the current and continuing areas which are particularly productive under three headings. Firstly there has been attention accorded to fundamental core scientific issues, themes which have always been integral parts of fluvial geomorphology but which can now be enhanced as a result of developments in recent years. Thus the hydraulics of open channel flow is investigated with more insight in relation to sediment transport and to the character of flow in meander bends (e.g. Ferguson et al., 2004), and the definition of environmental flows, those necessary to maintain channel aquatic environments in a 'natural condition' has been advanced (Dollar, 2000). Such progress has aided conceptual understanding of river channel patterns in relation to sediment transport and sedimentology (e.g. Hooke, 2003), benefiting from investigations of sediment transport and long-term sediment fluxes, including slugs, pulses or sediment waves (James, 2006). Such relational studies have been paralleled by investigations of coupling and of the transfers from hillslopes to rivers, floodplain lakes and coastal waters (e.g. Benda et al., 2005). These approaches have aided the transfer of results from small scale process studies to larger spatial scales and to longer temporal periods (e.g. Brown et al., 2001). Secondly, there have been more contributions to the solution of environmental management problems, some of which have arisen directly from process-based investigations as in the way that studies of slugs or sediment waves have implications for river management (James, 2006). One approach of particular value to environmental management is river and river channel classification with emphasis upon the hierarchy of relationships between different spatial scales. In any hierarchical classification a geomorphological input is necessary. A fluvial geomorphological methodology for the design of natural stable channels has been evolved for application to river restoration and is assisted by the existence of a clear channel classification (Hey, 2006). Studies of river channel change have developed significantly over the past 4 decades so that results are now being obtained which provide valuable inputs to river and environmental management (e.g. Gregory, 2006). Thus fluvial geomorphology has now reached the stage at which methods and results can be related to river engineering and management (Thorne et al., 1997a) and can underpin an approach to river channel management in terms of sustainable catchment hydrosystems (Downs and Gregory, 2004). Such encouraging developments, thirdly, depend upon increasingly multidisciplinary involvement in both research and resulting applications. Thus in the case of river and river channel classifications it is desirable for research to be undertaken jointly with freshwater biologists or ecologists especially in relation to river habitat problems (e.g. Clarke et al., 2003) and the ecological evaluation of flow regimes (e.g. Bragg et al., 2005). Links with ecology have been particularly profitable. In a citation analysis of geomorphological literature published during 1975–2000 it was found that water-based research dominates well-cited papers, riparian research with a biological emphasis being the hottest subfield in the 1990s (Dorn, 2002), and links between geomorphology and ecology explored more generally (Urban and Daniels, 2006). Such profitable collaboration has advanced understanding of the ecology and geomorphology of fluvial systems in relation to riparian sedimentation (Steiger et al., 2003), to woody debris in channels (Gregory et al., 2003) and to the development of new subfields such as biogeomorphology and ecohydrology. Multidisciplinary approaches have also enabled theories of channel behaviour that are more physics-based and attempt to relate process and form in a predictive manner (Brooks and McDonnell, 2000) and advances in more physically based modelling such as the downstream response to imposed flow transformation (DRIFT model, King et al., 2003), or cellular modelling (Coulthard et al., 2005; Nicholas, 2005). The interdisciplinary area of palaeohydrology has included the reconstruction of long-term river regimes (Benito et al., 1998), identification of the causes and mechanisms of short-term hydrological changes (Gregory et al., 2006b) and development of palaeohydrological tools for understanding Global Change (Gregory and Benito, 2003). Many such developments, with clear applications require cooperation in a multidisciplinary team as demonstrated from river restoration at the forefront of applied science (Wohl et al., 2005) and can be prompted by introductions such as the European Water Framework Directive where fluvial geomorphology will increasingly be used to help define the physical integrity of water bodies (Newson, 2006). However, despite greater multidisciplinary research collaboration there is still a need to reduce the paradigm lock between scientists and managers and stakeholders (Gregory, 2004) and it could be that the real river management challenge is integrating scientists, stakeholders and service agencies (Rogers, 2006, (Thorndycraft and Gregory 2008).

Scale in Fluvial Geomorphology

Scale is an important consideration in fluvial geomorphology, with process–form interactions occurring over a huge range of space and time scales. At one end of this range is the long-term evolution of the landscape. At the other are small-scale processes, such as the setting in motion of an individual grain of sand resting on the bed of a channel. Space scales therefore encompass anything from a few millimetres to hundreds of kilometres. Relevant time scales stretch from a few seconds to hundreds of thousands of years or more. In order to understand how the fluvial system operates we can examine the relationships between processes and form in more detail at finer scales. This can be done by examining individual sub-systems, or sub-systems within sub-systems. When focusing in like this it is important to remember that these sub-systems are all part of an integrated whole and therefore cannot be considered in isolation from the rest of the system.

Space scales (spatial scales) In studying the fluvial system, the scale of relevance varies according to the type of investigation. At the largest, drainage basin scale, it is possible to see the form and characteristics of the drainage network and drainage basin topography. These reflect the cumulative effect of processes operating over long time scales, as well as past changes imposed by the external basin controls. At a smaller scale, the form of a reach of meandering channel can be examined in the context of drainage basin history and the influence of controlling variables at the channel scale, such as the supply of water and sediment from upstream. The way in which the form and position of the channel has changed over time scales extending to thousands of years may be preserved as floodplain deposits, which can be used in reconstructing drainage basin history. Moving in to look at an individual meander bend, process–form interactions can be observed at a smaller scale. These include flow hydraulics within the bend and associated sediment dynamics. Investigations of rates of bend migration or bank erosion processes are also carried out at this scale. Depositional channel units such as the point bar are of interest to sedimentologists, providing evidence about the flows that formed them. At a finer scale still are individual ripples on the bar surface formed by the most recent high flow and, moving even closer, the internal arrangement of grains. At the finest scale are individual grains of sediment.

Time scales (temporal scales) and equilibrium At smaller spatial scales, process–form interactions generally result in more rapid adjustments. At the largest scale, the long term evolution of channel networks occurs over time scales of hundreds of thousands of years or more, while the migration of individual meander bends can be observed over periods of years or decades, and small-scale flow-sediment interactions within minutes. The perspective of the historically oriented geomorphologist concerned with the large-scale, long-term evolution of landforms is therefore very different to that of the process geomorphologist or engineer who is interested in the operation of channel processes at much shorter time scales (Schumm, 1988). Historical studies show that the fluvial system follows an evolutionary sequence of development that is interrupted by major changes induced by the external basin controls. However, over the much shorter time periods involved in the field measurement of processes, there may be little or no significant change in fluvial landforms. This might not matter too much if flow–sediment interactions at very small scales are of interest, although basin history certainly does have an influence at the reach scale, since channel form has been shaped by past changes in flow and sediment supply. The precise definition of equilibrium is also time dependent.

Equilibrium refers to a state of balance within a system, or sub-system. Negative feedback mechanisms help to maintain the system in an equilibrium state, buffering the effect of changes in the external variables. However, different types of equilibrium may exist at different time scales. These were defined by Schumm (1977) with reference to changes in the elevation of the bed of a river channel above sea level. If you were to observe a short section of river channel over a period of a few hours you would not see any change in its form (unless there happened to be a flood), although you might see some sediment transport. Over this short time period the channel is said to be in a state of **static equilibrium**. The same river, observed over a longer time scale of a decade, would show some changes. During this time, floods of various sizes pass through the channel, scouring the bed. In the intervening periods, deposition builds up the channel bed again. As a result of these cycles of scour and fill, the elevation of the channel bed fluctuates around a constant average value and **steady state equilibrium** exists. Over longer time scales, from thousands to hundreds of thousands of years or more, erosion gradually lowers the landscape. At these time scales, the channel elevation fluctuates around a changing average condition, the underlying trend being a reduction in channel elevation. This is called **dynamic equilibrium.** As you know, the influence of the external basin controls cannot be ignored. Changes in any of these variables can lead to positive feedbacks within the system and a shift to a new equilibrium state. For example, in tectonically active regions, the section of channel might be elevated by localised uplift. Such episodes of change occur over much shorter time scales than the gradual evolution of the landscape, resulting in abrupt transitions. This type of equilibrium delights in the term **dynamic metastable equilibrium** (Charlton 2007).

References

Baines, D., Smith, D.G., Froese, D.G., Bauman, P., Nimeck, G., 2002. Electrical Resistivity Ground Imaging (ERGI): a new tool for mapping the lithology and geometry of channel-belts and valleyfills. Sedimentology 49, 441–449.

Ballarini, M., Wallinga, J., Murray, A.S., van Heteren, S., Oost, A.P., Bos, A.J.J., van Eijk, C.W.E., 2003. Optical dating of young coastal dunes on a decadal time scale. Quaternary Science Reviews 22, 1011–1017.

Bates, P.D., Lane, S.N., Ferguson, R.I., 2005. Computational fluid dynamics for environmental hydraulics. In: Bates, P.D., Lane, S.N., Ferguson, R.I. (Eds.), Computational Fluid Dynamics. Applications in Environmental Hydraulics. Wiley, Chichester, pp. 1–15.

Benda, L., Hassan, M.A., Church, M., May, C.L., 2005. Geomorphology of steepland headwaters: the transition from hillslopes to channels. Journal of the American Water Resources Association 41, 835–851.

Benito, G., Baker, V.R., Gregory, K.J. (Eds.), 1998. Palaeohydrology and Environmental Change. Wiley, Chichester. Birkhead, A.L., Heritage, G.L., White, H., Niekerk, A.W., 1996.

Ground-penetrating radar as a tool for mapping the phreatic surface, bedrock profile, and alluvial stratigraphy in the Sabie River, Kruger National Park. Journal of Soil Water Conservation 51, 234–241.

Bøtter-Jensen, L., Bulur, E., Duller, G.A.T., Murray, A.S., 2000. Advances in luminescence instrument systems. Radiation Measurements 32, 523–528.

Bragg, O.M., Black, A.R., Duck, R.W., Rowan, J.S., 2005. Approaching the physical–biological interface in rivers: a review of methods for ecological evaluation of flow regimes. Progress in Physical Geography 29, 506–531.

Bronk Ramsey, C., 1995. Radiocarbon calibration and analysis of stratigraphy: the OxCal program. Radiocarbon 37, 425–430.

Bronk Ramsey, C., 2001. Development of the radiocarbon program OxCal. Radiocarbon 43, 355–363. Brooks, S.M., McDonnell, R.A., 2000.

Research advances in geocomputation for hydrological and geomorphological modelling towards the twenty-first century. Hydrological Processes 14, 1899–1908.

Brown, A.G., 2006. The Late Pleistocene sediment geometry of confluence migration: A major process of floodplain formation. Programme Abstracts of the GLOCOPH-IGCP 518 meeting on Present and past fluvial systems: Methods and Applications, Aug. 25–Sept 2, 2006, Guarulhos and Maringá (Brazil), p. 30. Brown, A.G., Cooper, L., Salisbury, C.R., Smith, D.N., 2001.

Late Holocene channel changes of the Middle Trent: channel response to a thousand year flood record. Geomorphology 39, 69–82.

Butle, J.B., Lane, S.N., Chandler, J.H., 1998. Assessment of DEM quality characterising surface roughness using close range digital photogrammetry. Photogrammetric Record 16, 271–291.

Candy, I., Black, S., Sellwood, B.W., 2004. Interpreting the response of a dryland river system to Late Quaternary climate change. Quaternary Science Reviews 23, 2513–2523.

Carey, C.J., Brown, A.G., Challis, K.C., Howard, A.J., Cooper, L., 2006. Predictive modelling of multi-period geoarchaeological resources at a river confluence: a case study from the Trent-Soar, UK. Archeological Prospection 13, 241–250.

Casas, A., Hardy, R.J., Lane, S.N., Whiting, P.J., 2005. Representing bed roughness as topographic variability in a three dimensional finite volume scheme. Sixth International Conference on Geomorphology, Zaragoza 2005, p. 87. Abstracts Volume. Casas, A., Benito, G., Thorndycraft, V.R., Rico, M., 2006. The topographic data source of digital terrain models as a key element in the accuracy of hydraulic flood modelling. Earth Surface Processes and Landforms 31, 444–456.

Charlton, M.E., Large, A.R.G., Fuller, I.C., 2003. Application of airborne LiDAR in river environments: the River Coquet, Northumberland, UK. Earth Surface Processes and Landforms 28, 299–306.

Charlton, Ro. Fundamentals of Fluvial Geomorphology. Oxon: Routledge, 2007.

Church, M., Jones, D., 1982. Channel bars in gravel bed rivers. In: Hey, R.D., Bathurst, J.C., Thorne, C.R. (Eds.), Gravel-bed Rivers. Wiley, Chichester, pp. 291–324.

Church, M.A., McLean, D.G., Wolcott, J.F., 1987. River bed gravels: sampling and analysis. In: Thorne, C.R., Bathurst, J.C., Hey, R.D. (Eds.), Sediment Transport in Gravel-bed Rivers. Wiley, Chichester, pp. 43–88.

Clarke, S.J., Bruce-Burgess, L., Wharton, G., 2003. Linking form and function: towards an eco-hydromorphic approach to sustainable river restoration. Aquatic Conservation and Freshwater Ecosystems 13, 439–450.

Corbeanu, R.M., Soegaard, K., Szerbiak, R.B., Thurmond, J.B., McMechan, G.A., Wang, D., Snelgrove, S.H., Forster, C.B., Menitove, A., 2001.

Detailed internal architecture of a fluvial channel sandstone determined from outcrop, cores, and 3-D ground-penetrating radar: example from the middle Cretaceous Ferron Sandstone, east-central Utah. Bulletin of the Association of American Petroleum Geologists 85, 1583–1608.

Coulthard, T.J., Lewin, J., Macklin, M.G., 2005. Modelling differential catchment response to environmental change. Geomorphology 69, 222–241.

Dollar, E.J., 2000. Fluvial geomorphology. Progress in Physical Geography 24, 385–406. Dollar, E.J., 2002. Fluvial geomorphology. Progress in Physical Geography 26, 123–143.

Dollar, E.J., 2004. Fluvial geomorphology. Progress in Physical Geography 28, 405–450. Dorn, R.I., 2002. Analysis of geomorphology citations in the last quarter of the 20th Century. Earth Surface Processes and Landforms 27, 667–672.

Downs, P.W., Gregory, K.J., 2004. River Channel Management. Towards sustainable catchment hydrosystems. Arnold, London. Duller, G.A.T., 2004. Luminescence dating of Quaternary sediments: recent advances. Journal of Quaternary Science 19, 183–192.

Duller, G.A.T., Murray, A.S., 2000. Luminescence dating of sediments using individual mineral grains. Geologos 5, 88–106. Ferguson, R.I., Parsons, D.R., Lane, S.N., Hardy, R.J., 2004.

Flow in meander bends with recirculation at the inner bank. Water Resources Research 39, WR001965. French, J.R., 2003. Airborne LiDAR in support of geomorphological and hydraulic modeling. Earth Surface Processes and Landforms 28, 321–335.

Gaswirth, S.B., Ashley, G.M., Sheridan, R.E., 2002. Use of seismic stratigraphy to identify conduits for saltwater intrusion in the vicinity of Raritan Bay, New Jersey. Environmental and Engineering Geoscience 8, 209–218.

Graf, W.L., 1988. Fluvial Processes in Drylands. Springer Verlag, Berlin Heidelberg. Greenbaum, N., Schick, A.P., Baker, V.R., 2000. The palaeoflood record of a hyperarid catchment, Nahal Zin, Negev Desert, Israel. Earth Surface Processes and Landforms 25, 951–971.

Gregory, K.J., 1978. Fluvial processes in British basins. In: Embleton, C., Brunsden, D., Jones, D.K.C. (Eds.), Geomorphology Present Problems and Future Prospects. Oxford University Press, Oxford, pp. 40–72. Gregory, K.J. (Ed.), 1983.

Background to Palaeohydrology: A Perspective. Wiley, Chichester. Gregory, K.J., 2000. The Changing Nature of Physical Geography. Arnold, London. Gregory, K.J., 2004. Human activity transforming and designing river landscapes: a review perspective. Geographica Polonica 77, 5–20.

Gregory, K.J., 2006. The human role in changing river channels. Geomorphology 79, 172–191. Gregory, K.J., Benito, G. (Eds.), 2003.

Palaeohydrology: Understanding Global Change. Wiley, Chichester. Gregory, K.J., Starkel, L., Baker, V.R. (Eds.), 1995. Global Continental Palaeohydrology. Wiley, Chichester. Gregory, S.V., Boyer, K.L., Gurnell, A.M. (Eds.), 2003.

The ecology and management of wood in world rivers. American Fisheries Society, Bethesda, Maryland. Gregory, K.J., Benito, G., Dikau, R., Golosov, V., Johnstone, E.C., Jones, J.A.A., Macklin, M.G., Parsons, A.J., Passmore, D.G., Poesen, J., Soja, R., Starkel, L., Thorndycraft, V.R.,Walling, D.E., 2006a. Past hydrological events and global change. Hydrological Processes 20, 199–204.

Gregory, K.J., Macklin, M.G.,Walling, D.E., 2006b. Past hydrological events related to understanding Global Change: an ICSU research project. Catena 66, 2–13.

Gurnell, A.M., Oosterhout, M.P.V., De Vlieger, B., Goodson, J.M., 2006a. Reach-scale interactions between aquatic plants and physical habitat: River Frome, Dorset. River Research and Applications 22, 667–680.

Gurnell, A.M., Boitsidis, A.J., Bark, T., Morrissey, I.P., Petts, G.E., Clifford, N.J., Thompson, K., 2006b. Initial adjustments within a new river channel: Interactions between fluvial processes, colonizing vegetation, and bank profile development. Environmental Management 38, 580–596.

Hardy, R.J., 2005. Fluvial geomorphology. Progress in Physical Geography 29, 411–425. Harvey, A.M., 2001. Coupling between hillslopes and channels in upland fluvial systems: implications for landscape sensitivity illustrated from the Howgill Fells, northwest England. Catena 42, 225–250.

Harvey, A.M., 2002. Effective timescales of coupling within fluvial systems. Geomorphology 44, 175–201. Hey, R.D., 2006. Fluvial geomorphological methodology for natural stable channel design. Journal of the American Water Resources Association 42, 357–374.

Higgitt, D.L., Lu, X.X., 2001. Sediment delivery to the three gorges: 1. Catchment controls. Geomorphology 41, 143–156.

Hooke, J.M., 2003. Coarse sediment connectivity in river channel systems: a conceptual framework and methodology. Geomorphology 56, 79–94.

Horrit, M.S., 2005. Parameterisation, validation and uncertainty analysis of CFD models of fluvial and flood hydraulics in the natural environment. In: Bates, P.D., Lane, S.N., Ferguson, R.I. (Eds.), Computational Fluid Dynamics. Applications in Environmental Hydraulics. Wiley, Chichester, pp. 193–213. Horrit, M.S., Bates, P.D., Mattinson, M.J., 2006.

Effects of mesh resolution and topographic representation in 2D finite volume models of shallow water fluvial flow. Journal of Hydrology 329, 306–314.

Ingham, D.B., Ma, L., 2005. Fundamental equations for CFD in river flow simulations. In: Bates, P.D., Lane, S.N., Ferguson, R.I. (Eds.), Computational Fluid Dynamics. Applications in Environmental Hydraulics. Wiley, Chichester, pp. 19–49.

Jacobson, R., O'Connor, J.E., Oguchi, T., 2003. Surficial geologic tools in Fluvial Geomorphology. In: Kondolf, G.M., Piegay, H. (Eds.), Tools In Fluvial Geomorphology. Wiley, Chichester, pp. 25–57.

James, L.A., 2006. Bed waves at the basin scale: implications for river management and restoration. Earth Surface Processes and Landforms 31, 1692–1706. Johnstone, E., Macklin, M.G., Lewin, J., 2006.

The development and application of a database on radiocarbon-dated Holocene fluvial deposits in Great Britain. Catena 66, 14–23.

King, J., Brown, C., Sabet, H., 2003. A scenario-based holistic approach to environmental flow assessments for rivers. River Research and Applications 19, 619–639.

Kondolf, G.M., Piegay, H., 2003. Tools in Fluvial Geomorphology: Problem Statement and Recent Practice. In: Kondolf, G.M., Piegay, H. (Eds.), Tools In Fluvial Geomorphology. Wiley, Chichester, pp. 3–22.

Lane, S.N., Ferguson, R.I., 2005. Modelling reach-scale fluvial flows. In: Bates, P.D., Lane, S.N., Ferguson, R.I. (Eds.), Computational Fluid Dynamics. Applications in Environmental Hydraulics. Wiley, Chichester, pp. 217–269.

Lane, S.N., Richards, K.J., Chandler, J.H. (Eds.), 1998. Landform Monitoring, Modelling and Analysis. Wiley, Chichester. Lane, S.N., Porfiri, K., Chandler, J.H., 2001.

Monitoring river channel and flume surfaces with digital photogrammetry. ASCE Journal of Hydraulic Engineering 127, 871–877.

Leeks, G.J., Lewin, J., Newson, M.D., 1988. Fluvial geomorphology and river engineering: the case of the Afon Trannon, Mid-Wales. Earth Surface Processes and Landforms 13, 207–224. Leopold, L.B., Wolman, M.G., Miller, J.P., 1964.

Fluvial Processes in Geomorphology. Freeman, San Francisco. Lewin, J., Macklin, M.G., 2003. Preservation potential for Late Quaternary river alluvium. Journal of Quaternary Science 18, 107–120.

Macklin, M.G., Lewin, J., 2003. River sediments, great floods and centennial-scale Holocene climate change. Journal of Quaternary Science 18, 102–105.

Macklin, M.G., Johnstone, E., Lewin, J., 2005. Pervasive and longterm forcing of Holocene river instability and flooding in Great Britain by centennial-scale climate change. The Holocene 15, 937–943.

Macklin, M.G., Benito, G., Gregory, K.J., Johnstone, E., Lewin, J., Soja, R., Starkel, L., Thorndycraft, V.R., 2006. Past hydrological events reflected in the Holocene fluvial history of Europe. Catena 66, 145–154.

Mark, K., Bates, P.D., 2000. Integration of high resolution topographic data with flood-plain flow models. Hydrological Processes 14, 2109–2122.

McKean, J., Roeding, J., 2004. Objective landslide detection and surface morphology mapping using high-resolution airborne laser altimetry. Geomorphology 57, 331–351.

Murray, A.S., Wintle, A.G., 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. Radiation Measurements 32, 57–73.

Newson, M.D., 2006. 'Natural' rivers, 'hydrogeomorphological quality' and river restoration: a challenging new agenda for applied fluvial geomorphology. Earth Surface Processes and Landforms 31, 1606–1624.

Newson,M.D., Newson, C.L., 2000. Geomorphology, ecology and river channel habitat: mesoscale approaches to basin-scale challenges. Progress in Physical Geography 24, 195–217.

Nezu, I., Nakagawa, H., 1993. Turbulence in Open Channel Flows. Balkema, Rotterdam. Nicholas, A.P., 2005. Cellular modeling in fluvial geomorphology. Earth Surface Processes and Landforms 30, 645–649.

Porat, N., Wintle, A.G., Amit, R., Enzel, Y., 1996. Late Quaternary earthquake chronology from luminescence dating of colluvial and alluvial deposits of the Arava Valley, Israel. Quaternary Research 46, 107–117.

Rhoads, B.L., 1994. Fluvial geomorphology. Progress in Physical Geography 18, 103–123. Rodnight, H., Duller, G.A.T.,Wintle, A.G., Tooth, S., 2006. Assessing the reproducibility and accuracy of optical dating of fluvial deposits. Quaternary Geochronology 1, 109–120. Rogers, K.H., 2006.

The real river management challenge: integrating scientists, stakeholders and service agencies. River Research and Applications 22, 269–280.

Roy, A., Lane, S., 2003. Putting the morphology back into fluvial geomorphology: the case of river meanders and tributary junctions. In: Trudgill, S., Roy, A. (Eds.), Contemporary Meanings in Physical Geography. From What to Why? Arnold, London, pp. 103–125.

Schumm, S.A., 1977a. The Fluvial System. Wiley, New York. Schumm, S.A., 1977b. Applied fluvial geomorphology. In: Hails, J.R. (Ed.), Applied Geomorphology. Elsevier, Amsterdam, pp. 119–156.

Sear, D.A., Newson, M.D., Brookes, A., 1995. Sediment-related river maintenance: the role of fluvial geomorphology. Earth Surface Processes and Landforms 20, 629–647.

Stallins, J.A., 2006. Geomorphology and ecology: unifying themes for complex systems in biogeomorphology. Geomorphology 77, 207–216.

Starkel, L., Gregory, K.J., Thornes, J.B. (Eds.), 1991. Temperate palaeohydrology: fluvial processes in the temperate zone during the last 15,000 years. Wiley, Chichester. Starkel, L., Soja, R., Michczyñska, D.J., 2006.

Past hydrological events reflected in Holocene history of Polish rivers. Catena 66, 24–33. Steiger, J., Gurnell, A.M., Goodson, J.M., 2003. Quantifying and characterizing riparian sedimentation. River Research and Applications 19, 335–352.

Stokes, S., 1999. Luminescence dating applications in geomorphological research. Geomorphology 29, 153–171.

Thomas, D.S.G., Goudie, A. (Eds.), 2000. The Dictionary of Physical Geography, 3rd edn. Blackwell, Oxford Thorndycraft, V.R., Benito, G., 2006a.

The Holocene fluvial chronology of Spain: evidence from a newly compiled radiocarbon database. Quaternary Science Reviews 25, 223–234.

Thorndycraft, V.R., Benito, G., 2006b. Late Holocene fluvial chronology in Spain: the role of climatic variability and human impact. Catena 66, 34–41.

Thorne, C.R., Hey, R.D., Newson, M.D., 1997a. Applied Fluvial Geomorphology for River Engineering and Management. Wiley, Chichester. Thorne, C.R., Newson, M.D., Hey, R.D., 1997b. Application of applied fluvial geomorphology: problems and potential. In: Thorne, C.R., Hey, R.D., Newson, M.D. (Eds.), Applied Fluvial Geomorphology in River Engineering Management. Wiley, Chichester, pp. 365–370.

Tinkler, K.J., 1985. A short history of geomorphology. Croom Helm, London. Urban, M.A., Daniels, M., 2006. Introduction: exploring the links between geomorphology and ecology. Geomorphology 77, 203–206.

Charlton, Ro. *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge, 2007.

Thorndycraft, V R, and K J Gregory. "Fluvial geomorphology: A perspective on current status and methods." *Geomorphology*, 2008: 2-12.

Walling, D.E., 1983. The sediment delivery problem. Journal of Hydrology 65, 209–237.

Walling, D.E., Collins, A.L., Sichingabula, H.M., Leeks, G.J.L., 2001. Integrated assessment of catchment suspended sediment budgets: a Zambian example. Land Degradation and Development 12, 387–415.

Wintle, A.G., Murray, A.S., 2006. A review of quartz optically stimulated luminescence characteristics and their relevance in single-aliquot regeneration dating protocols. Radiation Measurements 41, 369–391.

Wittenberg, L., Newson, M.D., 2005. Particle clusters in gravel-bed rivers: an experimental morphological approach to bed material transport and stability concepts. Earth Surface Processes and Landforms 30, 1351–1368.

Wohl, E., Angermeier, P.L., Bledsoe, B., Kondolf, G.M., MacDonnell, L., Merritt, D.M., Palmer, M.A., Poff, N.L., Tarbon, D., 2005. River restoration. Water Resources Research 41, W10301

UNIT-2: FLUVIAL SYSTEM: COMPONENTS, INPUT OUTPUT, VARIABLES OF FLUVIAL SYSTEM: INTERNAL AND EXTERNAL, ADJUSTABLE AND CONTROLLING FACTORS

INPUTS, OUTPUTS AND STORES

The basic unit of the fluvial system is the drainage basin. Fluvial systems are open systems, which means that energy and materials are exchanged with the surrounding environment. In closed systems, only energy is exchanged with the surrounding environment. **Inputs** The main inputs to the system are water and sediment derived from the breakdown of the underlying rocks. Additional inputs include biological material and solutes derived from atmospheric inputs, rock weathering and the breakdown of organic material. Most of the energy required to drive the system is provided by the atmospheric processes that lift and condense the water that falls as precipitation over the drainage basin. The pull of gravity then moves this water downslope, creating a flow of energy through the system. This energy is expended in moving water and sediment to river channels and through the channel network. **Outputs** Water and sediment move through the system to the drainage basin outlet, where material is discharged to the ocean. Not all rivers reach the ocean; some flow into inland lakes and seas, while others, such as the Okavango River in Botswana, dry up before reaching the ocean. This reflects another important output from fluvial systems: the loss of water by evaporation to the atmosphere. Most of the available energy is used in overcoming the considerable frictional forces involved in moving water and sediment from hillslopes into channels and through the channel network. Much of this energy is 'lost' to the atmosphere in the form of heat. **Stores** A certain amount of material is stored along the way. For example, water is stored for varying lengths of time in lakes and reservoirs, and below the ground in the soil and aquifers. Sediment is stored when it is deposited in channels, lake basins, deltas, alluvial fans and on floodplains. This material may be released from storage at a later stage, perhaps when a channel migrates across its floodplain, eroding into formerly deposited sediments which are then carried downstream. Ferguson (1981) describes the channel as 'a jerky conveyor belt', since sediment is transferred intermittently seawards.

TYPES OF SYSTEM

Three types of system can be identified in fluvial geomorphology. These are morphological systems, cascading systems and process–response systems. **Morphological (form) systems** Landforms such as channels, hillslopes and floodplains form a morphological system, also referred to as a **form system**. The form of each component of a morphological system is related to the form of the other components in the system. For example, if the streams in the headwaters of a drainage basin are closely spaced, the hillslopes dividing them are steeper than they would be if the streams were further apart from each other. Relationships such as this can be quantified statistically. **Cascading (process) systems** The components of the morphological system are linked by a cascading system, which refers to the flow of water and sediment through the morphological system. Cascading systems are also called **process systems** or **flow systems**. These flows follow interconnected pathways from hillslopes to channels and through the channel network. **Process–response systems** The two systems interact as a process–response system. This describes the adjustments between the processes of the cascading system and the forms of the morphological system. There is a two-way feedback between process and form. In other words, processes shape forms and forms influence the way in which processes operate (rates and intensity). This can be seen where a steep section of channel causes high flow velocities and increased rates of erosion. Over time erosion is focused at this steep section and the channel slope is reduced. Velocity decreases as a result, reducing rates of erosion. In order to examine the components of the fluvial system in more detail, it can be divided into sub-systems, each operating as a system within the integrated whole. One way of doing this is to consider the system in terms of three zones, each of which is a process–response system with its own inputs and outputs. Within each zone certain processes dominate. The sediment **production zone** in the headwater regions is where most of the sediment originates, being supplied to the channel network from the bordering hillslopes by processes of erosion and the mass movement of weathered rock material. This sediment is then moved through the channel network in the sediment **transfer zone**, where the links between the channel and bordering hillslopes, and hence sediment production, are not so strong. As the river approaches the ocean, its gradient declines and the energy available for sediment transport is greatly reduced in the sediment **deposition zone**. It is primarily the finest sediment that reaches the ocean, as coarser sediment tends to be deposited further upstream. In fact, only a certain proportion of all the sediment that is produced within a drainage basin actually reaches the basin outlet.

F LUVIAL SYSTEM VARIABLES

Variables are quantities whose values change through time. They include such things as drainage density, hillslope angle, soil type, flow discharge, sediment yield, channel pattern and channel depth. **Internal and external variables** An important distinction exists between internal and external variables. All the examples given above are **internal variables**, which operate within the fluvial system. Internal variables are influenced by other internal variables, and also by variables that originate from outside the system. These **external variables**, such as climate, control or regulate the way in which the system operates. Unlike the internal variables, external variables operate independently, in that they are not influenced by what is going on inside the fluvial system. At the basin scale the external variables are climate, base level, tectonics and human activity. If you are considering a sub-system, such as a reach of channel in the transfer zone, the external variables would include the supply of flow and sediment to the channel. This is because these variables originate from outside the channel sub-system, even though they are internal variables at the basin scale. To avoid confusion, the 'ultimate' external variables – climate, base level, tectonics and human activity – will be referred to as **external basin controls**. **The external basin controls** The variables defined in this section act as regulators of the whole system. Any change in one of these variables will lead to a complex sequence of changes and adjustments within the fluvial system. ● ***Climate*** describes the fluctuations in average weather. Although the weather is always changing, longer-term characteristics such as seasonal and inter-annual variations can be defined. Other characteristics include how often storms of a given size can be expected to occur and the frequency and duration of droughts. Where no long-term changes are occurring in the climate, the combination of such attributes defines an envelope of 'normal' behaviour. **Climate change** occurs when this envelope shifts and a new range of climatic conditions arises. ● ***Tectonics*** refers to the internal forces that deform the Earth's crust. These forces can lead to large scale uplift, localised subsidence, warping, tilting, fracturing and faulting. Where uplift has occurred, inputs of water have to be lifted to a greater elevation, increasing energy availability; some of the highest rates of sediment production in the world are associated with areas of tectonic uplift. Valley gradients are altered by faulting and localised uplift, which may in turn affect channel pattern. Lateral (sideways) tilting can cause channel migration and affect patterns of valley sedimentation. ● ***Base level*** is the level below which a channel cannot erode. In most cases this is sea level. If there is a fall in sea level relative to the land surface, more energy is available to drive flow and sediment movement. Conversely a relative rise in base level means that less energy is available, resulting in net deposition in the lower reaches of the channel. Over time these effects may be propagated upstream through a complex sequence of internal adjustments and feedbacks. ● ***Human activity*** has had an increasing influence on fluvial systems over the last 5,000 years, especially during recent times. Activities within the drainage basin such as deforestation, agriculture and mining operations all affect the flow of water and production of sediment. These are referred to as indirect or diffuse activities. River channels are also modified directly when channel engineering is carried out. Advances in technology over the last century have meant that dam construction, channel enlargement for navigation and flood control, channel realignment, the building of flood embankments and other engineering works can now be carried out at an unprecedented scale. Today there are very few rivers that have not been affected in some way by the direct and indirect effects of human activity. It can be argued that, under some circumstances, human activity can be considered to be *both* an internal and an external variable. Many of the direct modifications described above are in response to some local human perception of the system. For example, channels are dredged because they are not deep enough for navigation, or flood defence works carried out because floods occur too frequently. Urban (2002) suggests that direct human intervention can often be classified as an internal variable, although it is more appropriate to consider *indirect* human activities as external. Some internal variables have a greater degree of independence in that they are only affected in a limited way by the fluvial system. These variables are geology, soils and vegetation and topography (which includes relief, altitude and drainage basin size). All are internal variables because they are controlled to some extent by the external basin controls, however their main influence on the operation of the fluvial system is a controlling one. **Adjustable (dependent) and controlling (independent) variables** From the discussion above it can be seen that some variables control the adjustment of other variables. For example channel pattern is, among other things, affected by the supply of sediment to the channel. In this case, channel pattern is the **adjustable** or **dependent variable** while sediment supply is the **controlling** or **independent variable**. Things can get a little confusing because controlling variables may in turn be adjusted by other variables. Extending the previous example, sediment supply is itself controlled by hillslope vegetation cover. In this case, sediment supply is the adjustable variable and vegetation cover the controlling variable. All internal variables are adjustable because their operation is ultimately regulated by the external basin controls. They are also influenced to a greater or lesser extent by other internal variables. Because the relationships between variables are so complicated, it can be very difficult to isolate the effect of one variable on another. The hierarchical nature of the fluvial system means that variables operating at larger scales tend to affect the operation of variables at smaller scales. For example, climate affects vegetation cover and hillslope erosion, which in turn determine sediment supply, which influences channel pattern, which affects the small-scale flow dynamics in the channel, which governs the movement of individual grains. This is not a one-way process, however. Over long periods of time, the cumulative effect of small-scale processes, such as the erosion and deposition of individual grains, can lead to larger scale changes. These include changes in channel pattern and, over time periods of tens to hundreds of thousands of years, can adjust the slope of the whole river valley. Time itself is an important controlling variable. Every drainage basin has a historical legacy resulting from past changes that have taken place in the basin. This includes the cumulative effect of processes such as erosion, transport and deposition over long periods of time. It also includes the far-reaching effects of changes in the external basin controls, such as the variations in climatic conditions since the Last Glacial Maximum 18,000 years ago, which have greatly affected fluvial systems worldwide. In the temperate zone, many rivers underwent a transition from a braided to a meandering form as climate conditions ameliorated, vegetation became established and sediment loads decreased. However, vast quantities of sediment still remain in formerly glaciated drainage basins, where many fluvial systems are still adjusting to this glacial legacy. **Feedbacks** A **feedback** occurs when a change in one variable leads to a change in one or more other variables, which acts to either counteract or reinforce the effects of the original change. Two types of feedback are observed: negative feedback and positive feedback. Both are initiated by a change in one of the system variables, which in turn leads to a sequence of adjustments that eventually counteract the effect of the original change (**negative feedback**) or enhance it (**positive feedback**). When there is a change in one of the external controls, negative feedbacks allow the system to recover, damping out the effect of the change. An everyday example of a negative feedback loop is a central heating system controlled by a thermostat, which switches the source of heat on and off as the room cools and warms. An equilibrium is maintained as the temperature fluctuates around an average value. A commonly cited example of negative feedback within the fluvial system occurs when a section of channel is suddenly steepened by tectonic faulting. This leads to a local increase in the flow velocity and rate of bed erosion. Over time this acts to reduce the channel slope, counteracting the effects of the original change. It should be noted that the actual sequence of events is usually rather more complex. This is because change in one part of the system can lead to complex changes, both locally and throughout the rest of the system. The nature of complex response will be discussed later in this chapter. Positive feedbacks have a very different effect. Soil erosion is a natural process and an equilibrium exists if rates of soil removal over a given period are balanced by rates of soil formation over that period. However, an external change, such as the deforestation of steep slopes, can lead to a dramatic increase in soil erosion. The upper soil layers contain the most organic matter, which is important in binding the soil together. It also increases soil permeability, allowing rainfall to soak into the soil rather than running over and eroding the soil surface. If the topsoil is removed, the lower permeability of the underlying soil layers means that more water runs over the surface, increasing erosion and removing still more soil layers. In this way several centimetres of soil can be removed by a single rain storm (Woodward and Foster, 1997). This greatly exceeds the rate of soil formation. Referred to colloquially as ‘vicious circles’ or ‘the snowball effect’, positive feedbacks involve a move away from an equilibrium state. They usually involve the crossing of a **threshold** (see below) as the system moves towards a new equilibrium. A small-scale example of positive feedback is the build up of sediment during the formation of a channel bar. Bar formation is initiated when bedload sediment is deposited at a particular location on the channel bed. This affects local flow dynamics, causing the flow to diverge over and around the initial deposit. As the flow diverges, it becomes less concentrated and therefore less able to transport the coarser sediment. Localised deposition occurs, further disrupting the flow and promoting further deposition and bar growth. Several feedbacks, both positive and negative, exist between channel form, water flow and sediment transport. The form of a channel has an important influence on the way that water and sediment move through it. For example, flow is concentrated where the channel narrows, increasing erosion potential. As you saw above, deposition may occur where the flow diverges around obstacles such as bars. The character of the channel bed is also significant, since the size and arrangement of sediment determines bed roughness and resistance to flow. Where resistance is high, the average velocity of flow in the channel is reduced. This influences hydraulic conditions near the bed of the channel, which are significant for processes of erosion and deposition. Considerable differences are seen across the channel bed, giving rise to spatial variations in erosion and deposition. These processes themselves modify the form of the channel, feeding back to influence flow.

Thresholds

Thresholds are another important concept in systems theory and you will come across many examples in fluvial geomorphology. For example, a threshold is crossed when a sand grain on the bed of the channel is entrained (set in motion). Movement is resisted by the submerged weight of the grain and friction between it and the neighbouring grains. If the driving force exerted on the grain by the flow is less than these resisting forces, no movement will occur. It is only when the driving force of the flow exceeds the submerged weight of the grain that entrainment will take place. In this example, channel flow is an external variable. When a threshold is crossed there is a sudden change in the system, for example when loose material on a slope becomes unstable and starts to move down the slope as a landslide. The gradual processes by which rock is broken down and loose material builds up on a hillslope take place over time scales of tens to hundreds of years. Why, then, does a landslide occur at a particular point in time? Such a transition can come about when a change in one of the controlling variables leads to instability within the system – as a direct result of an earthquake (tectonics) for example. Thresholds that are crossed as a result of external change are called **external thresholds**. Instabilities may also develop over time without any external change having occurred. For this reason it is possible for a major landslide to be triggered by a relatively minor rainfall event that falls well within the expected climatic norm, because instability has gradually developed over time and the system is ready 'primed'. This is an example of an **internal threshold**. Again this is something that can take place without there having been any change in the external variables. Whether or not either kind of threshold is crossed depends on how 'sensitive' the system is, in other words how close to a threshold it is. To illustrate this point, consider a pan filled with water that is heated by 10°C at normal atmospheric pressure. If the water had an initial temperature of 25°C you would not expect to see much change in its appearance. However, if the initial temperature was 90°C, the same increase in temperature would lead to a dramatic change as a threshold was crossed and the water started to boil. In the second example, the sensitivity of the system is much greater because the water temperature is closer to boiling point. Once a threshold has been crossed, the system reaches a new equilibrium. Relating this back to the sand grain example, two scenarios could be considered. In the first, the particle is resting on a flat bed and is fully exposed to the flow. In the second, it is buried beneath a layer of gravel. The threshold for movement will be much lower for the exposed grain than for the buried grain, which cannot move unless the overlying gravel is removed.

Complex response

The response of the fluvial system to change is often complex because of the many interrelationships that exist between the different components of the system. An example is the complex response of a tributary to a lowering in base level elevation at its outlet (Schumm, 1977). Here the main river, into which the tributary flows, has degraded, or lowered its channel elevation by erosion. This leads to a complex sequence of episodic erosion and deposition in the tributary as the system searches for a new equilibrium (Charlton 2007).

References:

Charlton, Ro. *Fundamentals of Fluvial Geomorphology.* Oxon: Routledge, 2007.

Ferguson, R.I., 1981. Channel form and channel changes. In: J. Lewin (ed.), *British Rivers*. Allen & Unwin, London, pp. 90–125.
Schumm, S.A., 1977. *The Fluvial System*. John Wiley & Sons, New York.
Urban, M.A., 2002. Conceptualizing anthropogenic change in fluvial systems drainage development on the Upper Embarras River, Illinois. *Professional Geographer*, 54(2): 204–18.
Woodward, J. and Foster, I.D.L., 1997. Erosion and suspended sediment transfer in river catchments. *Geography*, 82(4): 353–76.

UNIT-3: LINEAR, AREAL AND ALTITUDINAL PROPERTIES OF DRAINAGE BASIN; LAW OF STREAM NUMBER AND STREAM LENGTH, LAW OF BASIN AREA

Landscape setting is a key determinant on catchment morphometrics. Analysis of relief (change in elevation/slope), drainage density (i.e. landscape dissection) and valley width aids in the interpretation of the distribution of erosion/deposition (process zones), and sediment and water flux in catchments, thereby guiding interpretations of controls upon patterns of river character and behaviour. Catchment morphometrics (i.e. shape, area, relief and drainage density) can be measured quickly and efficiently using digital elevation models (DEMs) and geographic information systems. *Valley width* is measured as the distance between bedrock valley margins (i.e. hillslopes). It is normally measured across the top of floodplains or terraces and perpendicular to the channel.

Catchment shape

Catchment shape is a major influence upon hydrologic relationships in landscapes. Lithology and longterm landscape evolution are key controls on catchment shape. The relationship between catchment area, stream length and resultant catchment shape, can be expressed as: $L = 1.4A^{0.6}$ where L (km) is stream length measured in a straight line from the highest topographic point to the river mouth along the longest axis of the catchment and A (km2) is catchment area. The exponent 0.6 suggests that catchments elongate with increasing size and that large catchments are relatively longer than smaller catchments.

Measures used to assess catchment shape include the circularity ratio, the elongation ratio and the form factor. The 'normal' pear-like ovoid shape of catchments can be related to circular forms to determine the *circularity ratio*:

$$\mathrm{Rc} = \frac{\mathrm{A}}{\mathrm{Ac}}$$

where *R*c is the circularity ratio, *A* is catchment area and *A*c is the area of a circle with the same circumference as the catchment.

Using this ratio, catchments with low ratios (about 0.4) are relatively elongate and are controlled primarily by geologic structure. Basins that are not controlled by structure have circularity ratios between 0.6 and 0.7 and are relatively round (i.e. the ratio is close to 1.0).

Unlike the circularity ratio that relies on the measurement of circles, the *elongation ratio* measures the catchment area to length relationship to give a measure of catchment shape:

$$Er = \frac{A0.5}{L}$$

where *E*r is the elongation ratio, *A* (km2) is the catchment area and *L* (km) is the catchment length along its axis. The closer to 1.0 the ratio is, the more round the catchment is. Catchments with elongation ratios around 0.6 are relatively elongate. In theory, the more elongate the catchment is, the slower the runoff from the basin is.

The *form factor* is another measure of the relationship between catchment area and length. However, unlike the elongation ratio that gives a measure of the shape of the catchment, the form factor provides a measure of the relationship between catchment area and catchment length and it's effect on hydrology:

$$Rh = \frac{H}{L}$$

where Rf is the form factor, A (km2) is the catchment area and L (km) is the catchment length along its axis. Catchments with a ratio of 4 have flashy flood regimes, while catchments with ratios closer to 8 tend to have lower flood intensities.

Catchment relief

Maximum catchment relief H is defined as the difference between the elevation of the catchment mouth Emin and the highest peak in the catchment Emax: $H = E - E$ max min However, this only measures the total fall of a catchment. To gain a clearer picture of the relative height over which water falls and the distance it travels, the maximum catchment relief is calibrated for catchment length using the *relief ratio*. This provides a measure of the average drop in elevation per unit length of river: R H L h = where Rh is the relief ratio, H (m) is the maximum catchment relief and L (m) is the basin length along its axis. Note: units of H and L should be the same (e.g. metres), so as to make Rh dimensionless.

The *hypsometric interval* is measured as the proportion of the catchment area that lies above and below a certain elevation. Moving upstream, elevation h progressively increases. The area a of the catchment cumulatively increases with each incremental increase in h (e.g. between contours). Relative values of h/H and a/A can be used to derive the hypsometric curve where the x and y values are dimensionless, representing proportions of the total area and height. For y = 0, all heights are above the datum plane. As such, they lie within the total area (i.e. x = 1). The area below the curve is calculated as the hypsometric interval (HI). This measure of topographic setting varies for different tectonic zones and geologic settings. For example, the top curve in Figure 3.5d typifies a relatively steep terrain in a catchment that has a significant proportion of its catchment comprising high-relief mountains that readily transfer flow/ sediment (over half the catchment area is high relief, h/H = 0.75). The bottom curve in Figure 3.5d reflects lowlying terrain in a catchment with a significant proportion of its area in rounded foothills and lowland/coastal plain plains (for half of the catchment area, h/H = 0.2).

These various measures of relief may vary markedly from terrain to terrain or for differing sub-catchments within a catchment, dependent upon the nature, extent and pattern of landscape units.

Drainage density and network extension

*Drainage density D*d is measured as the total length of stream channels per unit area of a catchment (e.g. km km−2). It provides guidance into the degree of landscape dissection, which in turn exerts a significant influence upon flow and sediment transfer through a catchment. Higher surface areas promote greater runoff and sediment generation. Average drainage density in moderately resistant lithologies range from 8.0 to 16.0. Ratios below this range are considered low. At the other end of the spectrum, dissected badlands may have drainage densities >1000. Maximum efficiency of flow and sediment transfer is achieved in these basins with complex bifurcating networks of small channels. These conditions promote rapid geomorphic responses to disturbance events. Vegetation cover and land use influence drainage density. Sparse vegetation cover leaves the landscape exposed to intense rainfall events that induce high rates of erosion and landscape dissection, maintaining and/or increasing drainage density.

Drainage networks evolve over time to generate a leaf vein pattern of streams. In initial stages, incision along the trunk stream lowers the base level along tributaries, inducing head cut development along these streamlines. Drainage density of the basin is low at this stage. Drainage extension and channel expansion progressively increase drainage density. However, once the drainage network has reached its maximum extent for a given catchment area, it is no longer possible to maintain rates of incision and erosion. As a result, drainage network integration reduces channel numbers and drainage density.

Drainage pattern

Drainage patterns describe the ways in which tributary streams are connected to each other and the trunk stream. Drainage network patterns are a product of the lithology and structure of a region. *Dendritic* drainage patterns are the most common form. They develop in areas of homogeneous terrain in which there is no distinctive geologic control. A pattern analogous to veins in a leaf is produced. Tributaries join the trunk stream at acute angles, less than 90°. The lack of structurally controlled impediments ensures that this configuration promotes relatively smooth downstream conveyance of sediment. In many other settings, however, geologic structure exerts a dominant influence on drainage pattern. For example, a *trellis* pattern is indicative of both a strong regional dip and the presence of folded sedimentary strata. Trunk streams flow along valleys created by downturned fold structures called synclines. Short tributaries enter the main channel at sharp angles approaching 90°. These tight-angle tributary junctions may induce short runout zones for debris flows. A *parallel* pattern is found in terrains with a steep regional dip or in regions where parallel, elongate outcrops of resistant rock impose a preferred drainage direction. Tributaries tend to stretch out in a parallel fashion following the slope of the landscape surface. In areas of right-angled jointing and faulting, a *rectangular* pattern is commonly observed. Streamlines are concentrated where the exposed rock is weakest. Tributaries join the trunk stream at sharp angles. *Radial* and *annular* drainage patterns develop around a central elevated point. This pattern reflects differential erosion of volcanoes and eroded structural domes respectively. *Multi-basinal* (or deranged) networks occur where the pre-existing drainage pattern has been disrupted. These networks are typically observed in limestone terrains or in areas of glacially derived materials. Finally, *contorted* drainage networks occur where the drainage network has been disrupted by neotectonic and volcanic activity.

Geologic controls on drainage network form, and river character and behaviour

Geologic controls on slope and sediment calibre exert a primary influence upon river character and behaviour. *Imposed boundary conditions* determine the relief, slope and valley morphology (width and shape) within which rivers adjust. In a sense, these factors influence the maximum potential energy conditions within which a river can operate. They also constrain the way that energy is used, through their control on valley width and, hence, the concentration (or dissipation) of flow energy. Imposed boundary conditions effectively dictate the pattern of landscape units, thereby determining the valley setting within which a river behaves and/or changes. Drainage basin evolution over millions of years often provides a significant antecedent control on contemporary river forms and processes.

Lithologic controls upon sediment calibre and volume

The calibre and volume of sediment supplied to valley floors fashion the behavioural regime of rivers. Rivers can only move the sediments available to them. Lithology influences both the calibre and volume of available sediments. The mineralogical composition of any rock determines the texture and hardness of its weathering breakdown products. Hence, the regional lithology influences whether these materials are resistant to erosion. The lithology of any given place is a product of geologic history. Minerals derived from upper mantle materials make their way to the Earth's surface either directly via volcanic events or indirectly via subsurface (endogenetic) processes and sub sequent removal of overlying materials. The enormous pressure and strain exerted by tectonic forces, and burial, induce metamorphic adjustment of igneous rocks and their reworked sedimentary counterparts. Weathering processes that break down parent rocks exert a significant influence upon the mix of grain sizes that are available to be reworked by geomorphic processes. In river environments, many sediments along valley floors are derived from reworking of upstream sediment stores that have been derived from rocks with a completely different mineralogical composition, and associated range of weathering breakdown products.

Differing lithologic settings produce rivers with differing bed material sizes. Channels that are lined with large boulders and cobbles are not found in areas where the regional lithology generates materials that are very friable. Resistant materials, such as gneiss or marble, generate coarse-bed, bedload-dominated rivers. Rivers in granitic environments have a distinctly bimodal sediment mix, with coarse granules and sand on the bed, while floodplains are made up largely of silt–clay materials. Rivers that flow through sandstone are often remarkably clear because they lack fine-grained sediments that induce turbid flow. These streams have a uniform sediment mix of sand-sized materials and are characterised by non-cohesive banks. A stark contrast is evident along rivers in basaltic terrains, where the lack of coarse-grained materials results in turbid, muddy, suspended-load streams. Flow in many limestone (karst) terrains is ephemeral, and most of the sediment load is transported in solution. Hence, the mix of available grain sizes exerts a primary control upon whether the river operates as a bedload-dominated river, a mixed-load river, a suspended-load river, or a solution-load river. The erodibility of bedrock also influences the volume of sediment that is supplied to a river system. Hard, resistant lithologies supply small amounts of sediment to rivers, resulting in supply-limited, bedrock-dominated landscapes. Such rocks often create steps along longitudinal profiles demarcated by waterfalls and over-steepened sections, along with narrow valleys. Rock hardness also affects the abrasive capacity of bed materials, influencing the rate of downstream decrease in grain size along a river. Weak, highly erosive rocks commonly oversupply a river with sediment, such that aggradation ensues in these transport-limited environments. Badland (gullied) environments commonly occur in such highly erosive rocks. The vast surface areas in these landscapes generate enormous volumes of sediment that result in aggradational valley floors (i.e. they are aggradational settings). Tributary–trunk stream relationships The spatial arrangement of tributaries in a river network exerts a primary influence upon process relationships at the catchment scale. By definition, a tributary is the smaller of two intersecting channels, and the larger is the trunk stem. The tributary–trunk stream catchment area ratio, the spacing between tributary confluences and the confluence intersection angle, among many considerations, determine the impact of tributaries upon the trunk stream. In some cases, tributary networks are too small to have a significant impact on flow and sediment inputs to the trunk stream, resulting in no change in its morphology. However, in other cases, tributary networks may have a significant impact on the morphology of the trunk stream. Tributaries that induce abrupt changes in water and sediment flux at confluence zones are called 'geomorphically significant (or effective) tributaries'. In general terms, consistent flow-related morphological changes occur at junctions where the ratio between tributary size and trunk stream size approaches 0.6 or 0.7. Intersection angles tend to be acute. However if this angle approaches 90°, the likelihood of a geomorphic effect at a confluence increases. The cumulative effect of confluences within a catchment should be proportional to the total number of geomorphically significant tributaries. The confluence density (number of geomorphically significant confluences per unit area or per unit channel length) is related to drainage density and can provide a simple measure of the net morphological effect of confluences in river networks. The drainage pattern of a catchment dictates the relative size and spacing of tributary networks. Dendritic networks in heart-shaped or pear-shaped catchments instigate confluence effects throughout the catchment. Downstream increases in catchment width promote the coalescence of hierarchically branched channels. Larger tributaries that join downstream may have a geomorphically significant effect upon the trunk stream. In contrast, narrow, rectangular catchments with trellis networks lack larger tributaries. These networks have a small number of geomorphically significant tributaries. Also, the effectiveness of these similarly sized tributaries diminishes downstream, as their size is progressively smaller relative to the trunk stream. Catchment configuration and network geometry influence the distance between geomorphically significant confluences. Large tributary junctions that are closely spaced may have confluence effects that overlap, particularly during large floods. In contrast, more widely spaced geomorphically significant tributaries exert a localised effect on factors such as downstream grain size. In general, as basin size increases, the channel length and area affected by individual confluence-related channel and valley morphological modifications increase. This measure can be used to determine how the degree and spatial extent of disturbance events in tributaries (floods and changes to sediment supply) impacts upon trunk stream dynamics. If the catchment configuration is altered, for example by emplacement or removal of blockages such as dams, the significance of tributaries to overall flow and sediment flux can be altered considerably. Stream order *Stream order* provides a measure of the relative size and pattern of channels within a drainage network. This exerts a significant influence upon the relative discharge of streams at any position in a drainage network. First-order streams have no tributaries, second-order streams only have first-order tributaries and so on. The quantitative framework of stream ordering explicitly recognises and documents the hierarchical structure of catchments. The Horton-Strahler stream order scheme, involves the following analysis:

1. Small, fingertip tributaries that occur in the headwaters (upstream most parts) of drainage networks are assigned order 1.

2. The junction of two streams of the same order *u* forms a downstream channel segment of order *u* + 1. For example, when two first-order streams come together, the segment of channel downstream of the confluence is assigned an order of 2. If two second order streams come together a third-order stream is formed downstream.

3. The junction of two streams of unequal order *u* and *v*, where *v* > *u*, creates a downstream segment with an order equal to that of the higher order stream *v*. For example, if a second-order stream meets a third-order stream, no change in order results and the segment downstream of the confluence remains as a third-order stream.

This approach does not consider the relative change in channel size and discharge that occurs when smaller, lower order tributaries meet a larger order stream. Determination of stream order is highly dependent on the scale of analysis and interpretations of where channel networks are considered to start in the headwater areas of catchments.

Three *laws of network composition* relate stream order to the number of streams, their length and their catchment area. The law of stream numbers is characterised by an inverse geometric progression whereby as stream order increases the number of streams of that order decrease. This means that there are relatively more first-order streams than second-order streams and third-order streams and so on until there is only one stream of a higher order at the catchment mouth. In catchments of relatively uniform lithology and structure, the ratio of the number of first- to second-order streams equals the ratio of the number of second- to third-order streams and so on. This is called the *bifurcation ratio R*b. The higher the bifurcation ratio, the more frequently a drainage line splits into a tributary and trunk stream and the higher the drainage density. The law of stream lengths states that as stream order increases there is a direct increase in stream length for that order, such that first-order streams tend to be relatively short compared with streams of a higher order. The rate of increase in stream length typically lies between 1.5 and 3. Finally, the law of catchment area states that catchment area increases in a smooth progression with increasing stream order. The relative increase in stream length has a ratio of between 3 and 6. These various parameters provide a descriptive summary of basin network composition. In a sense, these measures of catchment morphometrics build upon an implicit assumption that the upstream or upslope parts of landscapes are connected to downstream or downslope areas. In many instances, however, this assumption does not hold entirely true. While many landscapes are effectively connected (or coupled), some are at best partly con nected, while others may be disconnected.

Conclusion

Efforts to read the landscape build upon meaningful analysis of catchment-specific morphometrics. Differentiation of source, transfer and accumulation zones provides helpful guidance in framing analysis of river systems. Appraising relations to longitudinal profiles and associated understandings of downstream changes in slope and valley width helps to explain the balance of erosion and deposition, and resulting river forms, at different positions in landscapes. These considerations, alongside catchment shape, size and tributary–trunk relationships, fashion the flux of water and sediment through a drainage network. Tectonic, lithologic and climatic controls upon drainage density exert a primary influence upon the availability of materials to be distributed and their erodibility. Analyses of flow and sediment fluxes must consider how landscape components fit together at the catchment scale (i.e. their connectivity). Critically, site-specific investigations must be framed within their landscape and catchment context (Fryirs & Brierley, 2013).

References:

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

UNIT-4: HYDRAULICS OF CHANNEL FLOW: STREAM ENERGY; TYPES OF FLOW; TYPE OF LINKS, NUMBER OF LINKS

Introduction

When water accumulates in a channel on an inclined surface it has the ability to flow. The energy of that flow is able to perform geomorphic work such as transporting sediment or deforming channel boundaries. Slope and volume of water are key determinants of the amount of energy and the way in which that energy is used. Channel boundary factors influence these relationships: they determine the amount of seepage (and hence flow continuity), the manner/rate of energy consumption in overcoming friction and the ease with which bed and banks can be deformed. As tributaries join the trunk stream, flow volume increases. However, in general terms, slope tends to decrease. Changes to these controls affect the capacity of rivers to transport materials of differing texture or induce erosion and deposition as ways of using their available energy. This chapter outlines the primary forms of impelling and resisting forces in river systems. The chapter is structured as follows. Following summary comments on the mechanics of fluid flow, impelling and resisting forces are outlined. This is followed by a discussion of the way that energy is used in river systems in the context of the degradation–aggradation balance along longitudinal profiles, and the associated distribution of erosion and depositional processes.

Impelling and resisting forces and Lane’s balance of erosion and deposition in channels

Rivers act to move water and sediment downslope. In doing this they expend energy and perform geomorphic work. However, a critical energy level or threshold must be reached before a river can perform this work. The potential energy of flow within a channel is measured as the mass of water entering a river at a certain height above a given base level. As water moves downstream, potential energy is converted to kinetic energy. The *conservation of energy principle* states that the potential energy plus kinetic energy must remain constant within the system (i.e. no energy is lost). Hence, any loss in potential energy is matched by an equivalent gain in kinetic energy. However, rivers are nonconservative systems and friction causes much available energy to be dissipated in the form of heat, which performs no geomorphic work. Whether geomorphic work is done is dependent on the available amount of potential energy and the balance of energy expended and energy conserved at any particular location, such that erosion thresholds are, or are not, breached. Three possibilities exist: (1) a river may have more energy than that required to move its water and sediment load, in which case it has surplus energy and will adjust in the form of erosion; (2) it may have exactly that required, in which case it is stable; and (3) it may have an energy deficit, which will result in adjustment in the form of deposition.

In physics, a *force* refers to any influence that causes a free body (object) to undergo a change. In the case of river systems, water acts as the primary force by which matter in the form of sediment is moulded and shaped as it moves downstream. This largely reflects the amount of water (discharge) acting on a given slope. The proportion of erosion and deposition that occurs along a river channel is a function of the relative balance of *impelling* and *resisting* forces. The Lane balance diagram provides a key conceptualisation of this dynamic. There are four key components to the Lane balance. The left bucket depicts the volume of bed material load Q_s, with a sliding scale of median bed material size/calibre D_{50}. The right bucket depicts the volume of water in the river channel Q_w (discharge), with a sliding scale of channel slope s. The relative sizes of the buckets and their positions along the sliding scale determine whether the balance is tipped to the left or to the right and whether aggradation (deposition) or degradation (erosion) occurs. In theory, the channel acts to maintain the balance. If discharge increases proporor the channel increases its slope (e.g. a bend is cut off, so that channel length is reduced), the balance will tip to the right and degradation (erosion) results. This means that there is excess energy in the system relative to the volume and size of sediment and that energy is consumed via incision (the channel cuts into its bed). Alternatively, the same outcome occurs if the sediment load *Q*s is reduced or if the bed material size is decreased. In contrast, if excess sediment is added to the stream (i.e. *Q*s increases), especially if that bed material is coarser (i.e. *D*50 increases), the available discharge is unable to move all available material within the channel and aggradation (deposition) occurs (i.e. sediments accumulate on the bed;). Once more, the same outcome arises if discharge *Q*w decreases or channel slope decreases (i.e. channel length increases as the channel becomes more sinuous).

Critically, the Lane balance is used to describe how a channel is likely to adjust to maintain its balance in response to changes to flow and sediment conditions. In general, the water/discharge ‘bucket’ is primarily a function of climatic controls upon flow availability and variability, whereas the sediment ‘bucket’ is primarily a function of geological controls upon sediment availability (calibre and volume, determined primarily by weathering breakdown products, the erosivity of those materials and the erodibility of the landscape). The Exner equation links erosion or deposition to a deficit of, or excess in, sediment flux respectively, thereby providing a means to quantify these relationships. The equation describes conservation of mass between sediment on the channel bed and sediment in transport. It states that bed elevation increases (i.e. aggradation occurs) proportional to the amount of sediment that drops out of transport, and conversely decreases (i.e. degradation occurs) proportional to the amount of sediment that becomes entrained by the flow. As such, the Exner equation can be used to predict the occurrence of erosional and depositional forms along a reach. The equation is often used in its one-dimensional form as follows: *∂ ∂ = − ∂ ∂ n t q x* 1 0 ε s where *∂n*/*∂t* is the change in bed elevation over time, *ε*0 is the grain packing density, *q*s is the sediment discharge, *∂x* is the downstream direction. Values of *ε*0 for natural channels range from 0.45 to 0.75. The value for randomly packed spherical grains is 0.64.

Impelling forces drive adjustments through erosion and reworking of materials as a given volume of water flows over a certain slope. This is often measured in terms of the ‘energy’ and ‘efficiency’ of flow within a channel. Flow with sufficient energy is able to perform geomorphic work. To do this, it must overcome a number of threshold conditions to entrain and transport sediment, whereby it is able to erode the channel margins. Measures of *stream power* and *shear stress* are commonly used to explain how sediment is transported along a river channel.

Resisting forces reduce flow energy via friction. They determine how a channel consumes its available energy, i.e. the ability of the river to carry sediment of a given volume and calibre. These factors resist change, limiting the extent of river activity and adjustment, striving to maintain river morphology. They are commonly measured as *flow*, *boundary* and *channel resistance*. Prior to analysing and interpreting impelling and resisting forces, the mechanics of fluid flow are briefly described.

Mechanics of fluid flow

An understanding of the mechanics of fluid flow is required to quantify flow energy and the efficiency with which channels are able to use that energy. A fluid is defined as a material that deforms continuously and permanently under the application of a shearing stress, no matter how small. The inability of fluids to resist shearing stress gives them their characteristic ability to change their shape or to flow. Overcoming friction is a key characteristic of water flow. The ability of flow to overcome friction is dependent on flow volume and the nature of the surface over which it is moving. There are two fundamentally different types of flow motion:

1. *Laminar flow* refers to smooth, orderly motion in which fluid elements or particles appear to slide over each other in layers or laminae with no large-scale mixing.

2. *Turbulent flow* refers to random or chaotic motion of individual fluid particles with rapid macroscopic mixing of particles through the flow.

Flow is turbulent in natural channels. Velocity profiles represent the displacement of particles of water with respect to the bed in a given time period. The velocity gradient in turbulent flow is uniformly steep. Flow speed increases rapidly away from the boundary (channel bed), with the gradient being proportional to boundary roughness.

Isovels are contours of equal downstream velocity viewed in cross-section. The deepest part of the channel is referred to as the *thalweg*. Generally, the highest velocity filament of flow is located in this part of the channel. Velocity profiles, the pattern of isovels and the position of the thalweg vary for channels of differing shape and size (Figure 5.3a). In general terms, however, isovels are more closely packed near the channel bed than further away and they are less closely packed near to the banks than near to the bed (i.e. velocity increases as you move away from the rough boundary – channel bed and banks). The thalweg tends to sit just below the surface of the flow due to free-surface resistance.

Helicoidal flow is the anticlockwise, corkscrew-like motion of water in a meander bend (sinuous channel). This secondary flow is initiated by oscillation or perturbation in the flow and associated pressure gradient forces (Figure 5.3b). A number of secondary currents may be evident that diverge and converge at different positions in the channel.

Impelling forces in river channels

Total, specific and critical stream power *Total stream power* is an expression for the rate of potential energy expenditure against the bed and banks of a river channel per unit downstream length. It measures the rate of work done by flowing water in overcoming bed and internal flow resistance (described later), and transporting sediment. It reflects the total energy available to do work along a river channel. Total (or gross) stream power is measured as the volume of water (discharge *Q*) multiplied by the channel slope *s* and the specific weight of water:

$$\Omega = \gamma Qs$$

where Ω (W m−2) is the total stream power, *Q* (m3 s−1) is the discharge, *s* (m m−1) is the slope and *γ* is the specific weight of water (which is a function of acceleration due to gravity (9.8 m s−2) multiplied by water density (1000 kg m−3), i.e. 9800 N m−2).

Specific (or unit) stream power is a measure of energy expenditure per unit width of channel. It is measured as total stream power divided by the width of flow:

$$\omega = \Omega\ w$$

where *ω* (W m−2) is the specific stream power, Ω (W m−2) is the total stream power and *w* (m) is the water surface width at a specific discharge. Indicative thresholds of channel erosion and floodplain reworking have been defined in relation to critical values of unit stream power. For example, the thresholds for movement of pebbles, cobbles and boulder are around 1.5 W m−2, 16 W m−2 and 90

W m^{-2} respectively. The threshold level of channel instability is around 35 W m^{-2}, while 300 W m^{-2} is a threshold for floodplain stripping. These threshold values are merely indicative estimates. Real world values vary dependent upon reach- and catchment specific conditions, reflecting topographic, climatic and vegetation factors, among many considerations.

Critical stream power is the power needed to transport the average sediment load supplied to a stream. Where critical power is greater than the total stream power generated, there is insufficient energy to entrain and transport sediment. In contrast, where critical power is less than the total stream power generated there is sufficient energy available to move sediment and deposition occurs (i.e. these are supply-limited conditions).

Mean boundary shear stress

Shear stress, also referred to as tractive force, is the force applied by flowing liquid to its boundary. Put simply, shear stress describes the force of water along a channel boundary. Bedload movement and sediment transport are functions of shear stress. When the drag force of flowing water against a particle is greater than the gravitational force holding it in place the particle begins to move. Mean boundary shear stress is a measure of the force of flow per unit bed area. In other words, it is a measure of the drag exerted by the flow on the channel bed. It is computed as: $\tau\ \gamma\ 0 = Rs$ where $\tau 0$ (N m^{-2}) is the shear force per unit area of the surface (alternatively, 1 N m^{-2} = 1 kg m^{-1} s^{-2} = 1 Pa (pascal)), γ is the specific weight of water (9800 N m^{-3}), R (m) is the hydraulic radius, s (m m^{-1}) is the slope. In many cases, channel depth d is substituted for hydraulic radius, especially for channels with a high width/depth ratio. Mean boundary shear stress is used to determine the ability of flow to perform geomorphic work, especially bedload transport. It measures the force acting on the bed and banks of a channel. In general, shear stress on the banks of a channel tends to be 0.7 to 0.8 of that acting on the bed.

The concept of *critical shear stress* can be used to determine threshold conditions required to initiate bed erosion and sediment movement. Critical shear stress is computed as: $\tau\ c = k(\rho s - \rho)gD$ where τc (N m^{-2}) is the critical bed shear stress, k is a coefficient representing packing density, ρs is the sediment density (assumed to be constant at 2650 kg m^{-3}), ρ is the water density (assumed to be constant at 1000 kg m^{-3}), g is the acceleration due to gravity (9.81 m s^{-2}) and D (mm) is the characteristic grain size. For hydraulically rough beds that are common in natural streams, k ranges from 0.03 to 0.06, with 0.045 accepted for uniform spherical sediment. If $k = 0.045$ and water density and sediment density are considered as constants, it follows that: $\tau\ c = 0.73D$ (Fryirs & Brierley, 2013).

Reference:

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

UNIT - 05: RIVER VELOCITY, FACTORS AND ITS DISTRIBUTION IN OPEN CHANNELS; FLOW RESISTANCE, CHÉZY'S AND MANNING'S EQUATION

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 mean 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 (**t**0**) 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.

Channel parameters

In order to describe the flow of water in river channels it is necessary to define some basic channel parameters. 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. 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. 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. 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.

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The **wetted perimeter** is the length of channel boundary that is in direct contact with the flow at a given cross-section. 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, and 'lost' from the channel. As a result there is a downstream reduction in the total energy 'possessed' by a given parcel of water. The steepness of the energy slope reflects the rate at which energy is being expended.

Flow velocity

Flow velocity varies over both space and time in natural channels. It is determined mainly by the channel slope, roughness and cross-sectional form (remember that channel depth and cross-sectional area change with discharge). If you have ever waded out into a stream, you will know that the flow velocity, like the depth of flow, tends to increase as you move out into the channel. This is because of friction between the flow and the channel boundary, which is greatest near the bed and banks. Together with the effects of turbulence, the effects of this frictional resistance create variations in velocity distributions that are seen at different spatial and temporal scales. These are briefly discussed below.

● ***Variations with time:*** At any given point within the flow, the velocity fluctuates rapidly because of the effects of turbulence. This means that instantaneous velocities at a specific location can be much higher or lower than the **time-averaged velocity** that is recorded by a flow meter. Over periods of days, weeks or months, variations in velocity are also seen at the channel scale in response to discharge fluctuations.

● ***Variations with depth:*** These can be seen from measurements of time-averaged velocity made at different vertical heights above the channel bed (imagine a vertical line stretching upwards from a specific point on the channel bed). At the bed itself, the velocity is zero, but increases with vertical distance above the bed. The actual *rate* of increase, or **velocity gradient**, is greatest close to the bed, levelling off further away from the bed. The vertical velocity gradient at any point determines the shear stress exerted on the bed at that point.

● ***Variations across the cross-section:*** The fastest flow occurs towards the centre of the channel. At this cross-sectional scale, the average flow velocity can be calculated by making a number of measurements of velocity across the channel and at different depths. A description of how to do this is provided by Goudie (1981).

● ***Downstream variations:*** Although there is typically a decrease in channel slope along the length of a channel, the velocity generally shows little change or increases slightly. This is because the decrease in slope is often compensated for by a downstream decrease in channel roughness and an increase in hydraulic efficiency.

The concept of flow continuity

A casual observer walking alongside a natural stream channel might notice that the deeper sections are relatively slow flowing, while the shallow sections are relatively fast-flowing. The reason for this is that – assuming no tributaries join the channel and there are no significant interactions with groundwater – the same volume of water has to travel through each section in a given time. If this did not happen, the flow would start building up in some parts of the channel. Other parts would run dry as water flowed downstream faster than it was supplied from upstream. The mass of a given volume of water can be calculated by multiplying that volume by its density. According to the volumetric continuity equation above, the volume of flow does not change and, since water cannot be compressed, its density (1 kg per litre) does Therefore the mass of water passing (1) is equal to the mass of water passing (2). If this was not the case, water would be spontaneously created or destroyed somewhere along the channel.

Variations through time: steady and unsteady flow

Steady and unsteady flows are classifications of flow variations through time. In the example above, it was assumed that the discharge entering the reach did not change through time, something that is called **steady flow**. In natural channels, the flow is usually **unsteady**, varying through time as the drainage basin responds to inputs of precipitation.

Variations through space: uniform and non-uniform flow

Flows can also be classified according to variations over space. In a channel reach with a constant slope and cross-sectional shape, there will be no variation in either depth or velocity along the reach. This is called **uniform flow** and occurs in the upstream segment of the channel. The uniformity of the flow is indicated by streamlines – lines indicating the mean direction of individual 'parcels' of flow – which are parallel. Although most hydraulic equations are based on the assumption that the flow is uniform, this is rarely the case for natural channels, where the shape and dimensions of the channel vary in a downstream direction. There are also bends and obstacles to flow such as constrictions and channel bars. The flow expands into wider sections and becomes concentrated where there are constrictions. This means that the streamlines are no longer parallel, and the flow is described as **non-uniform**. It is only under uniform flow conditions that the channel bed slope, water surface slope and energy slope are the same. There are two types of non-uniform flow. **Gradually varied** flow reflects changes that occur over distances of tens of metres or more. **Rapidly varied** flow is associated with sudden changes in channel width, depth or alignment. In these situations, the streamlines cannot follow the line of the channel and something called flow separation occurs. Hydraulic jumps and drops are also associated with rapidly varied flow (Charlton, 2008).

F L O W R E S I S TANCE

A surprising amount of energy has to be used by flowing water in order to overcome flow resistance. It has been estimated that as much as 95–97 per cent of the total energy of a river is expended in this way (Morisawa, 1968). **Flow resistance formulae** express the relationship between flow velocity, channel slope, roughness and cross-sectional shape. Velocity increases with channel slope, but decreases with increasing boundary roughness. For example, a concrete-lined channel offers much less frictional resistance than a rocky, boulder-strewn channel. The hydraulic radius is also significant, since this determines the area of contact between the flow and boundary. Roughness is difficult to measure directly, so resistance formulae include an empirically-derived **friction coefficient.**

F L O W R E S I S TANCE FORMULAE

The Chezy equation is named after the eighteenth century French hydraulic engineer, Antoine de Chezy. This was later refined by the nineteenth century Irish engineer, Robert Manning. The Darcy–Weisbach equation has a long history of development and is named after two of the great hydraulic engineers of the nineteenth century. It has a sounder theoretical basis than the Manning and Chezy equations, although the Manning equation is still widely used today. **Chézy equation Manning equation Darcy–Weisbach equation** where *v* = velocity, *C* = Chezy roughness factor, *R* = hydraulic radius, *s* = channel slope, *n* = Manning roughness coefficient, *g* = acceleration due to gravity (9.8 m s−2) and *f* = Darcy–Weisbach friction factor. The Chezy coefficient (*C*) represents gravitational and frictional forces. Its value decreases with increasing roughness. Manning's roughness coefficient (*n*) is usually determined from tables. (Table 1 gives some values of Manning's '*n*' for natural channels.) Another method is to use photographs to make comparisons with channels of known roughness.

Chézy equation

$$v = C\sqrt{Rs}$$

Manning equation

$$v = \frac{R^{0.67} s^{0.5}}{n}$$

Darcy–Weisbach equation

$$v = \sqrt{\frac{8gRs}{f}}$$

Example application of the Manning equation: Calculate the velocity of a lowland meandering channel with riffles and pools, which has a slope of 0.001 m m–1, a wetted perimeter of 9 m and a cross-sectional area of 10 m2. *R* = 1.11 m (cross-sectional area/wetted perimeter) and '*n*' = 0.040, so: = 0.85 m so:

$$v = \frac{1.11^{0.67} \times 0.001^{0.5}}{0.040} = 0.85 \text{ m s}^{-1}$$

Problems and limitations Although widely used, these formulae have limitations. One of the main problems is that roughness is controlled by a number of different factors, including bed material size, bedforms and vegetation. This cannot be adequately represented by a single, empirically derived roughness coefficient. Flow resistance also changes with stage, being highest at low flows and lowest at bankfull stage. Once overbank flow starts to occur, the increased roughness of the floodplain surface greatly increases the overall flow resistance. **Channel resistance** At the valley scale, flow resistance increases when the channel comes into contact with the valley margins. This occurs in confined valley settings and where there are changes in valley alignment. The three-dimensional shape of the channel is also influential, since resistance is increased by irregularities in the banks, downstream changes in cross-section, and where the flow moves around bends. Bedrock-influenced channels can be highly irregular in form, with large variations in slope, width and channel cross-section. The high resistance of such channels is further increased by features such as cascades, vertical steps and potholes which increase form resistance. In a detailed investigation of variations in total flow resistance for different channel types along the Sabie River in South Africa, Heritage *et al*. (2004) reported extreme values of total flow resistance for bed rock influenced channel reaches during low flows. These values were calculated for mixed anabranching channel sections, where the flow is divided into a number of separate bedrock-dominated distributary channels under low flow conditions (the distributaries are separated by bedrock core bars that are overlain by cohesive sediment and vegetation). At low discharges, the flow in each distributary is very shallow and the highly fissured bedrock pavement means that the wetted perimeter is very large and tortuous (a large wetted perimeter means a smaller hydraulic conductivity and greater flow resistance). Numerous pools and rapids form within the fissures, with steep water surface slopes and very high rates of energy dissipation. Added to this are the effects of numerous boulders, which create obstacles to the flow. As discharge increases, a decrease in resistance is seen as these features become increasingly submerged by the flow. During flood flows, the vegetated bars separating the distributary channels become inundated, with an increase in resistance that is attributed to the increased resistance of the vegetation (Heritage *et al*., 2004). **Boundary resistance** There are two components of boundary resistance. The first of these, **grain roughness**, relates to the effects of the individual grains making up the channel boundary. **Form roughness** refers to features such as ripples and dunes, which are created when certain alluvial substrates are moulded by the flow. Grain roughness In general terms, flow resistance increases with the diameter of individual grains. However, an important factor is the depth of flow relative to the size of the particles. This can be expressed in terms of a ratio: $\frac{d}{D}$ where d is the flow depth and D is a characteristic grain size index; the median size of the bed sediment is often used. This ratio is used in many process-based equations in fluvial geomorphology and acts as a very significant control on the overall resistance in a channel (Robert, 2003). Bathurst (1993) compares the ratio of flow depth to characteristic grain size for different channel types along an idealised channel system. For a sand-bed channel, the flow depth may be over a thousand times greater than the diameter of the individual sand grains (2 mm or less). For gravel-bed channels, the ratio may be between 5 and 100, depending on the dominant grain size, which can range from cobbles (up to 250 mm) down to fine gravels (10 mm). In boulder bed channels, where most particles have diameters of 250 mm or more, the particles may project through the whole depth of flow, with a d/D ratio of less than 1. Where the stream bed consists of gravel or cobbles, grain roughness can be the dominant component of flow resistance (Knighton, 1998). However, the effect of grain roughness is often 'drowned out' as the depth increases. Grain size, and the spacing of individual grains, can also have a significant influence on the structure of turbulent flows. These effects will be discussed in the next section. Form roughness In sand-bed channels, it is possible for a wide range of flows to shape the channel bed. At different flow intensities, a sequence of **bedforms** develops. These include dunes, which are scaled to the depth of flow in the channel. Bedforms increase turbulence and can cause flow separation, leading to significant energy losses at high flows. Varying levels of resistance are associated with different types of bedforms (Simons and Richardson, 1966) and in sand bed streams the presence of these forms often exceeds grain roughness in importance (Knighton, 1998). In gravel-bed rivers, longitudinal variations in channel slope and bed roughness are often associated with periodic features called **riffles and pools**. Increased flow resistance occurs mainly as a result of ponding upstream from the shallower riffles (Hey, 1988). These are interspersed by deeper, slower moving pools, with a spacing of between five and seven times the channel width. Channel bars also increase flow resistance, particularly in braided channels. Even at higher flow stages, bars can account for between 50 and 60 per cent of total flow resistance (Prestegaard, 1983). Micro-scale variations in the bed topography of many gravel-bed channels are associated with smaller features called cluster bedforms. These consist of a single protruding obstacle, such as a large pebble, with associated accumulations of finer material immediately upstream and downstream. Cluster bedforms have a significant effect on shear stress distributions and flow resistance (Lawless and Robert, 2001). In steep, rocky channels, sequences of **steps and pools** may form. These are associated with very high rates of energy expenditure, particularly at low flows when considerable energy has to be dissipated in hydraulic jumps and pools (Bathurst, 1993). The extreme resistance reported by Heritage *et al*. (2004) for the Sabie River was associated with bedrock-influenced rapids and cataracts. This morphology is highly irregular in form, with very high energy dissipation caused by hydraulic jumps, constrictions and other disturbances to the flow. **Other controls on flow resistance** Riparian and in-channel vegetation increases flow resistance. This varies with stage, because vegetation that is upright at low flows may become flattened at higher flows. Seasonal effects are also seen when vegetation dies back during the winter months. Patchy growth can lead to considerable variations in resistance across the channel bed. In some channels woody debris builds up to create additional resistance. Sediment transport may also be of some significance. A high suspended load increases fluid viscosity, reducing turbulence and, in turn, flow resistance (Knighton, 1998). Several studies have investigated the dependence of flow resistance on bedload transport for the coarse bed materials typical of mountain rivers. However these effects appear to be small in relation to other controls (Bathurst, 1993).

FLOW BEHAVIOUR

Subcritical, critical and supercritical flow

The unsteady, gradually varied flow in most natural channels is **subcritical**. However, another type of flow behaviour, **supercritical flow**, is also observed. Within supercritical flows, turbulent mixing is less intense, with less deviation from the main downstream direction of flow. As a result, supercritical flows move rapidly and efficiently through the channel. They may overshoot tight bends and can also be highly erosive (Kay, 1998). The different types of flow behaviour can be predicted by calculating the ratio between the inertial and gravitational forces. The inertial force is given by $v2/d$; where v is the flow velocity, and d is its depth. The gravitational force is the acceleration due to gravity, g. The ratio between these forces is usually expressed in the form: $\mathrm{Fr} = v/\sqrt{gd}$ where Fr = Froude number, v = velocity, g is gravitational constant and d is depth. At Froude numbers less than 1 the gravitational forces dominate and the flow is subcritical. Conversely, when the inertial forces dominate, at Froude numbers greater than 1, the flow is supercritical. In rare cases, where the Froude number is equal to 1, the flow is described as being **critical**, or transitional. A **hydraulic drop** occurs when subcritical flow changes to supercritical flow. In this example, the increase in channel slope increases the flow velocity, resulting in a reduction in depth (the hydraulic drop). At the base of the weir, the flow changes back to subcritical, forming a **hydraulic jump**. A breaking wave indicates where this transition occurs. The sudden change in flow conditions at the hydraulic jump is caused by the decrease in slope at the base of the weir. Associated with this is a decrease in velocity and an increase in depth. The high velocity flow has considerable inertia and continues along the bed of the river before it is 'pulled' up to the surface and into the breaking wave. Turbulence is increased in this zone because of shear between the downstream and upstream movements of water. **Flow separation** Flow separation occurs where there are irregularities in the boundary. Examples include abrupt changes in bank orientation, sharp bends and obstructions at the bed such as large boulders. This results in the detachment of the boundary layer, which continues on in the direction of the flow but as a free shear layer In the flow separation zone, between the free layer and the boundary, is a 'bubble' of slow moving, recirculating fluid. The large difference in velocity between the fast moving shear layer and slowly recirculating flow in the separation zone means that large shear stresses develop. The resultant transfer of momentum leads to the free layer becoming unstable a certain distance downstream from the separation point, where it reattaches to the boundary. Increased turbulence is created by flow separation and results in a **wake** downstream from the object. Since flow separation affects shear stress distributions, it also influences processes of sediment erosion and deposition.

Laminar and turbulent flow

When considering the internal structure of fluid flow, a distinction is made between two quite different types of flow: laminar and turbulent. The British engineer Osborne Reynolds first demonstrated the existence of these two types of flow in his well known experiments on flows through pipes, carried out in the 1870s and 1880s. By injecting a thin stream of coloured dye into the water, Reynolds was able to observe patterns of movement within the flow. At low flow velocities, the dye was seen to travel as a single thread in a straight line through the tube, and was described by Reynolds as *direct* flow (now known as **laminar** or **viscous flow**). In laminar flows, the fluid moves as a series of layers, which slide over one another. This can be visualised as being somewhat similar to the way in which a pack of cards slide over each other when a shear stress is applied. Highly viscous fluids, such as oil or treacle, tend to exhibit laminar flow because of their high resistance to deformation. This can be seen from the 'smooth' way in which these fluids flow over a surface when gently poured. Water has a relatively low viscosity, so laminar flow only occurs at very low flow velocities. Reynolds found that a second, very different, type of flow occurred at higher velocities. In contrast to laminar flows, a series of horizontal and vertical swirling motions developed, dispersing the dye throughout the flow. Described as *sinuous* by Reynolds, this flow behaviour was subsequently termed **turbulent flow** by Lord Kelvin. Within the three-dimensional body of flow, movement can be in any direction: vertically up or down, sideways, upstream, downstream, or any combination of these. Reynolds found that as flows changed from laminar to turbulent, a transitional flow-type developed, with the turbulence intensity increasing as the flow became fully turbulent. From these experiments it was clear that two different types of flow behaviour existed. What was not so clear was how the transition between these flows could be predicted, as velocity is only one of a number of variables that control flow behaviour. Reynolds conducted further experiments using different fluids, and pipes with varying diameters. From these experiments, he derived an equation to define the transition from laminar to turbulent flow as a function of a single parameter, the **Reynolds number (Re)**. At low Reynolds numbers laminar flow occurs and at high values, turbulent flow. The concept of a Reynolds number is fundamental to much of modern fluid dynamics. It is calculated using the **Reynolds equation**, which expresses the ratio between **inertial** and **viscous forces** acting on the fluid. The inertia of an object – in this case a body of flowing water – is defined by its mass. Inertia determines how difficult it is to set something in motion (here: to initiate flow) but also how difficult it is to stop it, slow it down or change its direction once it has started moving. The greater the mass of water, the more inertia it has. Fluids that are denser than water (e.g. mercury) have more inertia because their mass per unit volume is greater. The inertial forces also increase with velocity. Acting against the inertial forces are viscous forces, which resist fluid deformation and flow. The viscosity of a fluid is determined by its structure at the molecular level, as work must be done to move the molecules past one another. Factors such as the regularity of molecular shapes and the strength of attraction between molecules affect the way in which the fluid responds to deformation. The more viscous a fluid is, the more it resists deformation and the less easily it flows.4 letter nu, which confusingly looks rather like a '*v*') = kinematic viscosity. At low Re numbers (less than 500), the viscous forces dominate and flow is laminar. Where the inertial forces are dominant (at Re numbers greater than 2,100), the inertia of the flowing water is much more significant than the viscous forces resisting that movement and turbulent flow occurs. The transition between laminar and turbulent flow occurs between Re values of 500 and 2,000. The Reynolds number is dimensionless: it does not have units. When velocity (in m s−1) and hydraulic radius (m) are multiplied together, the resultant units are m2 s−1. The units for

kinematic viscosity are also m2 s−1, so when the Reynolds number is calculated the units cancel out.
The boundary layer

When a fluid moves over a solid boundary, such as the wall of a pipe or the bed of a channel, it is affected by friction between the fluid and the boundary, in addition to the internal friction (viscosity) within the fluid itself. At a certain distance from the boundary, its effects are no longer 'felt' by the fluid and the flow velocity reaches a maximum or **free stream velocity**. The **boundary layer** is the thickness of flow that is affected by the boundary and is significant for several reasons. Much of the erosion and transport of sediment takes place in the boundary layer, turbulence is generated within it, and most of the plants and animals that are found in rivers live within this zone. Structure of the boundary layer A velocity gradient exists through the boundary layer, with the velocity increasing with distance from the boundary, at which it is zero. Immediately above the boundary is a thin (a millimetre or less) **viscous** or **laminar sublayer** within which the fluid movement is slowed so much by friction that the flow is laminar. Despite its limited thickness, the laminar sublayer is not insignificant. Fluid shear stresses within this layer are low and small particles of sediment that are wholly submerged within it are 'protected' from turbulent eddies that may entrain those particles that project above it. It also provides protection to small organisms. Between the laminar sublayer and the outer, turbulent, boundary layer is a transitional or **buffer layer** in which the flow structure is intermediate between laminar and turbulent. The outer boundary layer is fully turbulent and is called the **logarithmic layer**. Within this layer, the time-averaged velocity is often observed to increase logarithmically with height above the boundary (Richards, 1982). Above the boundary layer, in the **free stream layer**, there is no velocity gradient. The free stream layer is not always present because, in many cases, the boundary layer extends through the whole depth of flow. Hydraulically rough and smooth surfaces As well as influencing the overall resistance of a channel, roughness elements, such as grains of sediment, have significant effects on the structure of the boundary layer. Significant here is the height of grains relative to the thickness of the boundary layer. Where the grains of bed sediment are small enough to be totally submerged in the laminar sublayer, the flow is described as being **hydraulically smooth**. In this case the flow over the boundary is the same as it would be if the boundary was totally smooth and no grains were present. A different situation arises when grains are large enough to project through the sublayer. The rate of energy loss is increased as a result of turbulent eddy shedding. Eddies are generated as the flow moves over and around the particles, eddy size increasing with particle size (Leeder, 1999). In this situation the flow is **hydraulically rough**. Where the penetrating grains are of a similar diameter to the thickness of the sublayer the surface is described as **transitional**. To complicate matters, the thickness of the laminar sublayer decreases with increasing near-bed velocity. This means that some of the 'protected' smaller grains can become exposed at higher flows. The spacing of individual grains also has an effect, because eddy development and flow resistance are reduced when grains are closely spaced. The degree of channel bed roughness letter nu, which confusingly looks rather like a '*v*') = kinematic viscosity. At low Re numbers (less than 500), the viscous forces dominate and flow is laminar. 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Most natural channels are hydraulically rough (Robert, 2003), with roughness elements including coarse sediment, bedforms and woody debris. The form roughness associated with bedforms can be very high because of the generation of eddies associated with flow separation (Leeder, 1999). Momentum transfer, velocity distributions and fluid shear stress The fact that a velocity gradient exists within the boundary layer – or that there is a boundary layer at all – is due to the viscosity of the fluid. If the fluid were non-viscous (something called an 'ideal fluid'), it would all flow at the same velocity, with the exception of a thin layer of molecules that adhered to the boundary itself. However, most fluids, including water, are not 'ideal' and something called **momentum transfer** takes place. The momentum of a moving object, or 'parcel' of fluid, is determined by the product of its mass and velocity. Within the flow, momentum transfer allows slower moving fluid to be speeded up by faster moving fluid, and vice versa. For laminar flows this is brought about by a process called **molecular diffusion**. Molecules within a slower moving layer of fluid have less momentum than molecules in the faster moving layer above. If a molecule moves from a slower moving layer to a faster moving one, that molecule will have less momentum than the molecules surrounding it. As a result it will be speeded up and, at the same time, it slows down the surrounding molecules very slightly. Similarly, a molecule moving from a faster moving layer to a slower moving layer will be slowed down by the surrounding molecules, but will accelerate them slightly. Molecular diffusion also occurs within turbulent flows, although it is insignificant in comparison to a much more effective form of momentum transfer brought about by turbulent eddies that transfer 'lumps' of fluid within the flow. In a recognisable sequence of events, called **turbulent bursting**, an **ejection** of low momentum fluid occurs from the region close to the boundary. This moves upwards into the flow profile. Following this, an inrush, or **sweep**, of high momentum fluid moves downwards to replace the ejected fluid. This sequence of events recurs with a certain periodicity and is strongly related to flow structures called vortices, which develop within the transitional zone of the boundary layer (see Leeder, 1999, Robert, 2003, or Bridge, 2003, for more detail). Momentum transfer by turbulent eddies results in an apparent viscosity which is called **eddy viscosity**. The large size of eddies means that this type of momentum exchange is much more efficient than molecular diffusion. This means that high momentum fluid and low momentum fluid are much more thoroughly 'mixed' within the turbulent region. As a result, velocity differences between different 'layers' of flow in the turbulent part of the profile are smaller than those observed in the underlying viscous layers (laminar sublayer and buffer zone). In other words, the velocity gradient is much gentler in the fully turbulent logarithmic layer than in the underlying viscous layers. Since there is a velocity gradient, there must also be a shear stress between different layers within the flow, which are travelling at different velocities. This acts over the area of the plane of contact between the two layers. Calculating bed shear stress Much research has focussed on defining how the time averaged velocity varies with height above the bed within turbulent boundary layers. This is difficult to predict, mainly because the eddy viscosity varies with the nature of the turbulence. Also the situation is greatly complicated when the boundary is made of movable sediment. Bedforms shaped by the flow modify the geometry of the channel. This, in turn, feeds back to affect the flow. An understanding of the velocity–height relationship brings with it a number of practical applications. For example, it allows bed shear stress to be derived from measured vertical velocity profiles. Bed shear stress is used in numerous equations to describe flow, to determine boundary resistance and to estimate the volume of sediment that is being transported. Unfortunately it is extremely difficult to measure directly in the lab, let alone in natural river channels (Middleton and Southard, 1984). If the flow is steady and uniform – rarely the case in natural channels – the du Boys equation can be used. This involves measuring the channel or water surface slope using surveying techniques but is not necessarily a straightforward procedure. An alternative method is to measure the time averaged velocity at different heights throughout the turbulent profile. From these measurements, the bed shear stress can be calculated indirectly using a relationship called the logarithmic 'law of the wall'. This describes the relationship between vertical velocity distribution, boundary roughness and a 'surrogate' for bed shear stress called the shear velocity. This a measure of the velocity gradient and shear stress near the boundary (Chanson, 1999). Bed shear stress is related to the near-bed velocity gradient, increasing with the steepness of the gradient. It is also affected by the roughness of the channel bed. **Secondary flows** Secondary flows are vertical and lateral currents that develop within the channel, perpendicular to the main direction of flow. They exist at a much larger scale than turbulent eddies but are weaker than the primary (downstream) flow. This cross-channel circulation is superimposed on the primary flow to produce helical secondary flow cells. Secondary flows are caused by irregularities in the channel boundary, or where there is a difference in water surface elevation across the channel. This situation occurs at meander bends, where water 'piles up' at the outside of the bend. This sets up a pressure gradient, causing water to move downwards and across the channel bed, from the outside to the inside of the bend. Secondary flows are also observed in straight channels, with the spirals scaled to the depth of flow. **Overbank flows** Overbank flows occur when the capacity of the channel is exceeded. As mentioned in Chapter Three there is a certain amount of spatial variation along the channel in the timing and extent of inundation. Rather than a river suddenly 'bursting its banks', inundation first starts to occur in those areas where the bank topography is of a relatively low elevation. During peak flow conditions, the water mainly flows in a down-valley direction (Bridge, 2003). Overbank flows are more complex than within-channel flows, with the channel and floodplain forming what is called a compound channel. This consists of a deeper, central portion (the river channel) which is flanked on both sides by shallower floodplain flows. This is a conceptual model, developed by Knight and Shiono (Knight, 1989; Shiono and Knight, 1991), which represents the interactions between the main channel and floodplain flows. In general, the velocity of flow on the floodplain is much lower than that in the main channel because of the shallower depth of flow, which creates a very large wetted perimeter. In addition, floodplain flows are greatly retarded by the increased roughness of the floodplain surface. This is due to vegetation (especially bushes and scrub), man-made structures such as field boundaries, and variations in floodplain topography. Within the channel itself, experimental and field studies have shown that flow velocities are relatively fast (Bridge, 2003). As a result, a shear layer develops between the faster-moving channel flow and slower-moving floodplain flow. This extends laterally for a considerable distance, both across the floodplain and into the main channel. The shear layer indicates a two-way transfer of momentum between the floodplain flow and main channel flow (Knight, 1989). This comes about because the slower moving flow on the floodplain reduces the velocity of the flow in the channel. At the same time, the faster-moving flow in the channel speeds up the flow on the floodplain. Associated with the shear layer is a bank of large-scale vertical 'interface vortices', which transfer high-momentum fluid from the channel onto the floodplain (Knight and Shiono, 1996). This has important implications for sediment movement, since sediment is also routed onto the floodplain.

References:

Morisawa, M., 1968. *Streams: Their Dynamics and Morphology*. McGraw-Hill, New York.

Heritage, G.L., Large, A.R.G., Moon, B.P. and Jewitt, G., 2004. Channel hydraulics and geomorphic effects of an extreme flood event on the Sabie River, South Africa.
Catena, 58: 151–81.

Robert, A., 2003. *River Processes: An Introduction to Fluvial Dynamics*. Arnold, London.

Bathurst, J.C., 1993. Flow resistance through the channel network. In: K. Beven and M.J. Kirkby (eds), *Channel Network Hydrology*. John Wiley & Sons, Chichester, pp. 43–68.

Knighton, D.A., 1998. *Fluvial Forms and Processes: A New Perspective*. Arnold, London.

Simons, D.B. and Richardson, E.V., 1966. *Resistance to Flow in Alluvial Channels*. United States Geological Survey Professional Paper 422–J.

Hey, R.D., 1988. Bar form resistance in gravel-bed rivers. *Journal of Hydraulic Engineering*, 114: 1498–508.

Prestegaard, K.L., 1983. Bar resistance in gravel bed streams at bankfull stage. *Water Resources Research*, 19: 472–6.

Lawless, M. and Robert, A., 2001. Scales of boundary resistance in coarse-grained channels. *Geomorphology* 39, 221–38.

Kay, M., 1998. *Practical Hydraulics*. E. & F.N. Spon, London.

Richards, K.S., 1982. *Rivers: Form and Process in Alluvial Channels*. Methuen, London.

Leeder, M., 1999. *Sedimentology and Sedimentary Basins: From Turbulence to Tectonics*. Blackwell, Oxford.

Middleton, G.V. and Southard, J.B., 1984. *Mechanics of Sediment Movement.* Society of Economic Paleontologists and Mineralogists (SEPM) Short Course 3.

Chanson, H., 1999. *The Hydraulics of Open Channel Flow: An Introduction*. Arnold, London.

Knight, D.W. and Shiono, K., 1996. River channel and floodplain hydraulics. In: M.G. Anderson, D.E. Walling and P.D. Bates (eds), *Floodplain Processes*. John Wiley & Sons, Chichester, pp. 139 81.

Knight, D.W., 1989. Hydraulics of flood channels. In: K. Beven and P. Carling (eds), *Floods: Sedimentological and Geomorphological Implications*. John Wiley & Sons, Chichester, pp. 83–105.

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

Goudie A. (1981). *Geomorphological Techniques*. Allen &Unwin, London.

UNIT- 6: CHANNEL INITIATION AND EVOLUTION OF CHANNEL PATTERN, IMPORTANCE OF HEADWARD EXTENSION AND BRANCHING, LATERAL EXPANSION

Introduction

River evolution is the study of river adjustment over time. Evolution is ongoing. Even if boundary conditions remain relatively constant, adjustments occur. Appraisal of the trajectory and rate of river evolution is required to assess whether ongoing adjustments are indicative of long-term trends or whether they mark a deviation in the evolutionary pathway of that river. Such insights guide interpretation of the likelihood that the direction, magnitude and rate of change will be sustained into the future. To perform these analyses, it is important to determine how components of a river system adjust and change over differing timeframes, and assess what the consequences of those changes are likely to be. Reconstructions of the past provide a means to forecast likely future river behaviour. Instinctively, human attention is drawn to landscapes that are subject to change. Observations of bank erosion, river responses to flood events, anecdotal records of river adjustments or analyses of historical maps and aerial photographs provide compelling evidence of the nature and rate of river adjustments. Efforts to read the landscape must frame these insights in a broader context, examine their representativeness and isolate controls upon evolutionary trajectories. For example, do these adjustments reflect modifications around a characteristic state and associated equilibrium scenarios over a given timeframe? Are short-term adjustments indicative of longer term trends? Has the river been subjected to threshold-induced change? How has the balance of formative and reworking processes and controls changed over time? Is the river sensitive or resilient to disturbance? How are responses to disturbance manifest through the catchment, remembering that an erosional signal in one place is often matched by a depositional signal elsewhere? Attributes such as thalweg shift on braidplains, meander migration/translation, cutoff development or avulsion are characteristic behavioural traits for certain types of rivers. In some instances, alterations to the boundary conditions under which rivers operate may bring about river change, whereby the behavioural regime of the river is transformed, and the river is now characterised by a different set of process–form relationships. River evolution may occur in response to progressive adjustments, an instantaneous event (e.g. a major flood or an earthquake) or longer term changes to geologic and climatic boundary conditions. This distinction between behaviour and change is essentially a matter of timescale. All rivers change as they evolve over time. River change can result from alterations to impelling forces, resisting forces, or both. Resulting adjustments modify the nature, intensity and distribution of erosional and depositional processes along a reach. In some instances, predictable transitions can occur. For example, a change from a wandering gravel-bed river to an active meandering river can occur as flux boundary conditions are altered to reduce sediment load and discharge, or vegetation cover is increased. However, just because a particular type of river in a given system responds to an event of a given magnitude in a certain way, does not mean that an equivalent type of river in an adjacent catchment will respond to a similar event in a consistent manner. Even if particular cause and effect relationships are well understood, some systems may demonstrate complex (or chaotic) responses to disturbance events. More importantly, no two systems are subjected to the same set of disturbance events. Each system has its own history and its own geography (configuration), with its own cumulative set of responses to disturbance events, and associated lagged and off-site responses. The trajectory of river change may be influenced by the co-occurrence of disturbance events, such as a large flood following vegetation clearance. Such concatenations may set the system on a trajectory of change that would not have occurred if the system had not been disturbed or if these disturbances had occurred independently. Geologic and climatic factors determine the environmental setting and the nature of disturbance events to which rivers are subjected. They set the imposed and flux boundary conditions that fashion the erodibility and erosivity of a landscape, and the resulting character, behaviour and pattern of river types. Stark contrasts can be drawn, for example, between a dry, low-relief landscape with negligible vegetation cover and a high-precipitation mountainous terrain with dense forest cover. Formative processes, rates of activity (magnitude–frequency relations) and evolutionary trajectories vary markedly in these differing settings. Hence, any consideration of river evolution must be framed in relation to these geologic and climatic controls. Particular emphasis is placed upon how landscape setting influences the imposed boundary conditions (especially slope and valley width) that constrain the range of behaviour of rivers, and the flux boundary conditions (i.e.

flow and sediment regimes) that determine the mix of erosional and depositional processes along any given reach. Critically, as noted from the Lane balance diagram, alteration to either the imposed or flux boundary conditions promotes evolutionary adjustments. Geologic factors set and alter the imposed boundary conditions under which rivers operate, through their influence on lithology, relief, slope, valley morphology and erosivity and/or erodibility of a landscape. For example, tectonic activity or volcanic events may disrupt the nature and configuration of a landscape. Climate considerations play two critical roles. First, they are key determinants of the type and effectiveness of geomorphic processes (flow and sediment interactions) that shape landscapes at any given place. Second, climatic factors mediate the role of ground cover, which affects hydrologic processes and landscape responses to geomorphic processes through its influence upon surface roughness and resistance. Alterations to flux boundary conditions drive adjustments to the flow–sediment balance, prospectively modifying the evolutionary trajectory of a system. Evolutionary adjustment may take a mere moment in time (e.g. river responses to a volcanic eruption) or be lagged some time after a disturbance event. Elsewhere, landscapes may be stable or demonstrate progressive adjustment over time. Some rivers are adjusted to high coefficients of discharge variability, such that large floods are rare but not unusual – they are part of the 'formative process regime' for that particular setting. Other rivers are adjusted to smaller, more recurrent events. Many rivers flow on surfaces created by past events, or are still adjusting to past flow and sediment regimes. In these cases, geomorphic memory continues to exert a significant influence upon contemporary forms and the nature and effectiveness of processes. Understanding how contemporary processes relate to historical influences is a key challenge in efforts to read the landscape. This chapter is structured as follows. First, timescales of river change are discussed. Second, pathways and rates of geomorphic evolution are summarised for different types of rivers. Third, geologic and climatic controls on river evolution are considered. Finally, tools to interpret river evolution by reading the landscape are reviewed.

Timescales of river adjustment

Timescale of river adjustment varies from place to place, dependent upon the range of adjustment of the system (its sensitivity/resilience), the range and sequence of disturbance events and the legacy of past impacts. Both sensitive and resilient systems are prone to disturbance – responses are more likely and/or recurrent in the former relative to the latter. Analysis of river evolution frames system responses to disturbance events in relation to adjustments over geologic and geomorphic time. Geologic controls set the imposed boundary conditions within which rivers operate. Over timeframes of millions of years, tectonic setting exerts a primary control upon topography, determining slope and valley settings that influence river morphology and behaviour. Over geomorphic time, rivers adjust to climatically fashioned flux boundary conditions (flow variability, sediment availability and vegetation cover) over hundreds or thousands of years. Any disruption to flux boundary conditions may affect the evolutionary trajectory of a river. The key consideration here is whether the reach is able to accommodate adjustments while it continues to operate as the same type of river (i.e. it operates within its behavioural regime) or whether these altered conditions bring about a transition in process–form relationships (i.e. river change occurs). A continuum of responses to disturbance events may be discerned: • *No response may be detected*, as systems absorb the impacts of disturbance. Stable rivers can tolerate considerable variation in controlling factors and forcing processes. For example, gorges are resilient to adjustment or change. Alluvial systems with inherent resilience induced by the cohesive nature of valley floor deposits, or the mediating influence of riparian vegetation and wood, may demonstrate limited adjustment over thousands of years. In these cases, responses to disturbance events are short-lived or intransitive, and change does not occur. • *Part of progressive change*. Rivers may respond rapidly at first after disruption, but in a uniform direction thereafter, such that change occurs gradually over a long period. For example, progressive denudation results in gradual reduction of relief over time, as gravitationally induced processes transfer sediments from source to sink. This results in long-term changes to slope and, hence, river type. Progressive adjustments are often observed following ramp or pulse disturbance events, so long as threshold conditions are not breached. • *Change may be instantaneous* as breaching of intrinsic or extrinsic threshold conditions prompts the transition to a new state or even a new type of river. These effects tend to be long lasting or persistent. • *Change may be lagged*. Off-site impacts of major disturbances may induce a lagged response in downstream reaches (e.g. conveyance of a sediment slug). The subsequent history of disturbance events affects the nature/ rate of response and prospects for recovery. Efforts to read the landscape seek to unravel variability in forms, rates and consequences of adjustments within any given system over differing timeframes. Pathways and rates of adjustment and evolution vary markedly for different types of rivers.

Pathways and rates of river evolution

Evolutionary pathways and rates of adjustment of rivers vary in differing geologic and climatic settings. This reflects differing ways in which boundary conditions and disturbance events affect flow–sediment interactions along a river. Alternatively, disturbance events may affect the surfaces upon which these processes are acting. Evolutionary adjustments are likely to be most marked for those systems that have the greatest capacity to adjust and change. Hence, the nature and rate of evolution tend to be most pronounced in freely adjusting alluvial settings. These rivers have the greatest range in their degrees of freedom, such that pronounced disturbance events may trigger adjustments in channel planform, channel geometry (bed and bank processes), assemblages of channel and floodplain geomorphic units, and bed material organisation. The mix of water, sediment and vegetation conditions, as such, influences likely pathways of river adjustment for rivers in differing settings. Characteristic examples of evolutionary pathways are presented for rivers in differing valley settings below. Likely evolutionary pathways of rivers in confined valley settings Rivers in confined valley settings have limited capacity for adjustment. Their morphologies are largely imposed and are comprised largely of an array of imposed (bedrock) erosional forms. Steep headwater rivers progressively rework assemblages of slope-induced erosional geomorphic units as channels cut into bedrock via incisional processes over timeframes of thousands of years. In contrast, gorges are stable and resilient systems over timeframes of hundreds or thousands of years. However, progressive incision and lateral valley expansion eventually create space along the valley floor for floodplain pockets to develop in partly confined valleys. These transitions reflect changes to imposed boundary conditions. Likely evolutionary pathways of rivers in partly confined valley settings Just as gorges progressively widen to partly confined valleys with bedrock-controlled floodplain pockets over thousands of years, so sustained widening of these valleys eventually promotes the transition to partly confined valleys with planform-controlled floodplain pockets. Increased valley width and reduced valley floor slope or changes in material texture, in turn, may result in a transition in the type of planform-controlled floodplain pockets that are observed, say from a low-sinuosity variant to a meandering planform variant. Adjustments to flow–sediment relations (i.e. flux boundary conditions) may bring about a transition to adjacent types of rivers along this continuum. Reduced energy conditions induced by lower flow and/or sediment availability may transform a braided river into a wandering gravel-bed river, and vice versa. In turn, reduced energy conditions induced by lower flow and/or sediment availability may transform a wandering gravel-bed river into an active meandering river, and vice versa. Alternatively, increase in sediment load (bedload fraction) may transform a passive meandering (suspended-load) river into an active meandering (mixed-load) river, and vice versa. Various stages of evolutionary adjustments may be discerned along a discontinuous watercourse, reflecting cut and fill phases. However, should certain circumstances eventuate, the river may maintain a continuous watercourse. The examples outlined in convey progressive evolutionary adjustments. In essence, the types of rivers that are found in an adjacent position along the longitudinal profile (i.e. an energy gradient) are likely to present the next step or phase in the evolutionary adjustment of a river. This may reflect conditions of decreasing energy associated with progressive landscape denudation, or increasing energy associated with uplift (i.e. steeper slope conditions). This line of reasoning, whereby juxtaposed river types along slope-induced environmental gradients provide guidance into likely evolutionary adjustments, is a direct parallel to Walther's law of the correlation of facies: adjacent sedimentary deposits in contemporary landscapes are used to guide inferences into stacked depositional units within basin fills. Geologic and climatic controls are the primary determinants of imposed and flux boundary conditions, and the associated suites of disturbance events to which rivers are subjected. Although these geologic and climatic considerations act in tandem, they are considered separately below for simplicity.

Geologic controls upon river evolution

Geologic setting determines the imposed boundary conditions within which rivers adjust and evolve. The nature and movement of tectonic plates is a primary determinant of the distribution and relief of terrestrial and oceanic surfaces. The nature and position of mountain belts and depositional basins is determined largely by the distribution of plates and geologic processes that occur at different types of plate boundaries. Landscape relief and topography are fashioned by the balance of endogenetic processes (i.e. geologic processes that are internal to the Earth) and exogenetic processes (i.e. geomorphic processes that erode and deposit materials at the Earth's surface). The nature, frequency and consequences of geologic disruption and disturbance events vary markedly in different tectonic settings. This is determined largely by position relative to a plate margin and the nature of tectonic activity at that margin. Vertical and lateral displacement along fault-lines is common in some settings. Contorted strata of folded rocks attest to the incredible forces at play. Faulting, folding and tilting generate distinctive topographic controls upon slope, valley morphology and drainage patterns. Volcanic activities and subsidence modify relief and availability of materials. Tectonic setting frames the long-term landscape and dynamic context of river systems. These geological foundations determine patterns of lithological and structural variability, affecting the erodibility and erosivity of landscapes. The convergence of continental plates generates major mountain chains. Deeply incised bedrock channels in headwater settings contrast starkly with transport-limited braided rivers, low-relief rivers atop uplifted plateau landscapes or deeply incised gorges at plateau margins. Uplift of supply-limited plateau landscapes may create deeply entrenched, superimposed drainage networks. For example, 12.4d shows the planform of a meandering river that previously formed on a relatively flat alluvial plain that has been retained as the landscape was uplifted, creating a deeply etched, bedrock controlled gorge in which river character and behaviour are imposed. Differing forms of constructional landscapes are generated through subduction of dense but relatively thin oceanic plate beneath a continental plate. Recurrent phases of tectonic activity produce basin and range topography comprised of mountain ranges, volcanic chains and intervening basins, exerting a dominant imprint upon contemporary drainage networks. The imprint of landscape setting upon river character, behaviour and evolution is clearly evident in pull-apart basins. This tectonic setting is characterised by striking alignment of lakes and straight, bedrock-controlled river systems. In some instances, basins that pulled apart in the past may retain a dominant imprint upon contemporary landscapes, forming escarpments and rift valleys. Alternatively, lack of tectonic activity is a primary determinant of river processes and forms in plate-centre landscapes. These lowrelief, low-erosion settings often have profound stability and antiquity. Long-term changes to plate tectonic boundaries ensure that any given landscape setting has likely been subjected to differing forms and phases of tectonic activity. Geologic adjustments are sometimes imprinted atop each other. Elsewhere, the imprint of past events has been virtually erased, though metamorphosis of rocks may provide insights into former conditions. Importantly, tectonic setting not only fashions the relief and erodibility of a landscape, it also affects the climate and, hence, the erosivity of that landscape.

Climatic influences on river evolution

Spatial and temporal variability in climate are genetically linked to geologic considerations, as mountain belts and other topographic factors influence temperature and precipitation regimes and the movement of weather systems. The distribution of landmasses and latitudinal factors fashion continental or maritime climate conditions and solar radiation effects. Topographic and climatic conditions can be combined to differentiate morphoclimatic regions. Climatic controls upon river evolution are manifest in two primary ways. *Direct* influences reflect hydrologic considerations and thermal conditions, expressed primarily by the flow regime. This drives the flux boundary conditions under which rivers operate. *Indirect* influences are manifest primarily through climatic influences upon ground cover (and rainfall–runoff associations) and resistance factors (i.e. surface roughness). Any alteration to these relationships affects the flux boundary conditions under which rivers operate. Adjustments to the flow and sediment balance may alter the evolutionary trajectory of a river. In many settings, past climatic conditions continue to exert an influence upon the effectiveness of contemporary geomorphic processes (i.e. climatic memory). Direct and indirect impacts of climate variability vary markedly in differing morphoclimatic regions. Some tropical humid regions are characterised by high temperatures and high precipitation throughout the year, and have rainforest vegetation associations. Rivers in these regions are attuned to recurrent high flow conditions and considerable roughness on valley floors, but interannual variability in flow is limited. Tropical humid areas with prominent dry and monsoonal seasons are characterised by savanna vegetation. Although seasonal variability in flow and geomorphic activity is pronounced, interannual variability is limited. Rivers in these areas are especially sensitive to the effectiveness of the monsoon. Mid-latitude regions are dominated by arid and semi-arid climates. Desert and steppe landscapes have limited vegetation cover. Pronounced, highly effective geomorphic activity occurs during short storms. Desert environments with limited sediment availability are characterised by etched/sculpted bedrock rivers. Other deserts have ephemeral rivers with high sand availability, resulting in high width/depth channels because of the non-cohesive, non-vegetated nature of bank materials. Humid-temperate rivers have perennial flow. Vegetation cover exerts a primary influence upon process–form relationships. Warmer humid regions are not subjected to severe winter conditions, but summers can be hot and dry. Vegetation cover may be relatively sparse and shrub-like in Mediterranean areas, but is much more substantive in subtropical regions. There is marked variability in runoff generation and geomorphic effectiveness of floods in this morphoclimatic zone. Some areas have extremely high coefficients of variation in discharge, with significant interannual variability in flood events. Often, river systems are attuned to extremely high, but infrequent, flows. Mediterranean rivers have seasonal discharge and variable ground cover. Ephemeral streams are subjected to irregular reworking by flash floods. Discontinuous watercourses are prominent. Cooler humid regions have severe winters and continental climates, with significant areas of boreal forest. Rivers freeze in winter, and there is extensive permafrost in northerly latitudes. Profound adjustments may occur during spring melt. Polar regions are dry and cold, and bedrock-dominated rivers are relatively inactive. Landscape history and climate setting bring about marked variability in flora and fauna across the globe. Faunal interactions with rivers can affect the nature, rate and effectiveness of geomorphic processes. A wide range of ecosystem engineers is evident. Ants and worms induce bioturbation in soils, impacting upon sediment supply and transfer on hillslopes. Beaver dams exert a direct impact upon channels. Hippopotamus tracks may induce channel realignment. Wombat burrows may locally enhance rates of bank erosion. Changes to these faunal interactions may alter the evolutionary trajectory of the river. Similarly, any factor that alters vegetation cover (and associated resistance/ roughness) can have a significant affect upon the evolutionary trajectory of a river. For example, the geomorphic role of fire varies markedly in differing morphoclimatic regions. Savanna and Mediterranean areas are especially prone to fire events that clear ground cover, resulting in pulsed flow and sediment inputs into river systems. Climate is a key driver of river change. It fashions the sequence of disturbance events that bring about geomorphic adjustments, influencing system dynamics and the behavioural regime of any given reach. In some instances, floods or droughts may bring about transitions to a different type of river. Impacts upon the flow regime, and changes to ground cover, alter the rate of sediment movement in river systems, thereby affecting both sides of the Lane balance diagram. At the coldest part of the last glacial maximum (15 000– 18 000 yr ago), ice covered one-third of the land area of the Earth to an average depth of 2–3 km, but in places up to 4 km. Ice sheets created sculpted/denuded landscapes, creating slowly adjusting bedrock-dominated rivers. Alpine glaciers carved U-shaped valleys and fiords. During the recessional stages of ice sheet activity, meltwater channels realigned many drainage networks. Significant volumes of glacially reworked materials drape many landscapes, providing large sediment stores that can be reworked by river systems. Hence, there are marked differences in the historical imprint upon contemporary landscapes in glaciated and non-glaciated settings. Glacial cycles also brought about significant falls in sea level (up to 120 m). This exerted a profound impact upon patterns and rates of sedimentation in lowland basins as base level adjusted. Deep canyons were carved into what are now parts of the continental shelf. These effects were propagated upstream, leaving terraces at valley margins. Subsequent sea level rise during interglacial periods created drowned valleys and ria coastlines. Floodplain, terrace and marine sediments in infilled lowland valleys and estuaries retain records of multiple phases of sea level rise and fall. Longer term glacial–interglacial cycles also brought about major river changes in arid morphoclimatic zones, altering the distribution and extent of monsoonal climatic influences. As climate changes, so too does the vegetation cover. Hence, geomorphic adjustments reflect alterations to both impelling forces (the flow regime) and resisting forces (ground cover). Geomorphic responses to climate change are markedly variable in different parts of the world. The impact of climate change is not simply a measure of the direction or extent of change. Temperature changes from −20 °C to −30 °C may not induce a marked difference in process response, but transition from −5 °C to +5 °C certainly does. Similarly, change in annual precipitation from 9000 to 10 000 mm a^{-1} is unlikely to induce marked variability in geomorphic process activity, but changes from 500 to 1500 mm a^{-1} definitely would, primarily because of altered vegetation cover. Geomorphic responses to variability in climatic conditions vary markedly for different types of rivers, reflecting their sensitivity to adjustment. They also vary dependent upon the condition of the system at the time of the disturbance event (especially its resistance). In many instances, contemporary landscapes have been fashioned largely by conditions from the past.

Landscape memory: imprint of past geologic and climatic conditions upon contemporary river processes, forms and evolutionary trajectory

Contemporary rivers flow upon, and rework, surfaces created by past events. Hence, historical influences may exert a primary influence upon the distribution, rate and effectiveness of erosional and depositional processes. This imprint from the past varies markedly in differing settings. Geologic controls determine the relief, topography and erodibility of a landscape. The influence of elevation upon potential energy manifests itself as impelling forces (and associated kinetic energy) driven largely by slope (i.e. erosivity). This exerts a primary control upon the effectiveness of erosional processes and the resulting degree of landscape dissection. Geologic factors also influence the nature and extent of accommodation space and associated patterns of sediment stores in landscapes. Valley setting, in turn, affects channel–floodplain relationships, thereby influencing the contemporary capacity for adjustment of rivers. The contemporary climate regime is a primary determinant of the flux boundary conditions under which rivers operate, affecting discharge and flow energy and vegetation and/or ground cover which resist erosion processes. Critically, these relationships have changed over time. The impact of these changes is especially pronounced in those parts of the world affected by Pleistocene glacial activity. Glaciers carved deep and narrow valleys in mountain areas, constraining the range of geomorphic behaviour of contemporary channels in these settings. Many downstream areas were draped with glacially reworked materials. In some instances these vast (paraglacial) sediment stores that reflect former climatic conditions continue to influence contemporary river behaviour. The distribution of these sediment stores is influenced largely by geologic controls upon the accommodation space in landscapes, such as wider sections of valleys that store glacio-fluvial, glacio-lacustrine and alluvial fan materials. In many other settings, ice sheets stripped surface materials from vast areas, limiting contemporary rates of sediment supply across largely denuded areas. Another form of climatic memory is that associated with floodplain deposits of underfit streams. These inherited forms influence contemporary river morphology and associated patterns and rates of sediment erosion, transport and deposition. In this instance, climatic memory directly reflects geologic memory, as past geologic controls induced the accommodation space along palaeovalleys within which contemporary rivers flow. Landscapes retain a selective memory of past events. Sometimes a sharp erosional boundary reflects a major disjunct in time, highlighting the removal or erasure of a significant part of the record. Indeed, some landscapes may retain a very limited history of past events. Elsewhere, especially in long-term depositional basins in accretionary environments, a remarkable long-term record may be preserved (much of which is buried subsurface). Hence, different parts of a landscape retain variable records of past activity. Ultimately, changes to boundary conditions drive river evolution.

River responses to altered boundary conditions

The Lane balance diagram provides a simplified basis with which to interpret primary controls upon river evolution. Essentially, if the bed stability of a river changes, so will the geomorphology. In other words, the balance becomes unsettled and adjustments ensue. The two key considerations here are the amount of water acting on a given slope and the volume and texture of sediment delivered to the channel. As noted above, geology and climate are the primary determinants of these factors. The tectonic setting determines the rate of uplift (i.e. relief and sediment generation, and erosion rate), while lithology determines the breakdown size of weathered/eroded materials. Uplift or subsidence also alters the slope upon which geomorphic processes are acting. Climatic factors determine the flow regime and the amount of water available to do work in river systems. Evolution is driven by changes to these various controls. Davisian notions of landscape adjustment infer that rivers evolve as slopes decrease and valley floors widen over geologic time, prior to uplift kick-starting the cycle once more. Such continuity in boundary conditions, and even the direction of change, is seldom observed in reality, as invariably something happens to disrupt these patterns over timeframes of millions or tens of millions of years. Disturbance events may alter the flow–sediment balance along a river, whereby changes to geologic and climatic conditions induce adjustments in process relationships along valley floors, and resulting river morphologies. Various examples of river evolutionary adjustments in response to altered boundary conditions, disturbance events and flow–sediment fluxes are outlined below. River responses to tectonic uplift and displacement along fault-lines Uplift of a fault block, or even an entire plateau landmass within a plate, induces rejuvenation, whereby rivers are made young again and incise into underlying bedrock. If the rate of bed incision is unable to keep up with the rate of uplift, convex bulges are created along longitudinal profiles. These areas are characterised by waterfalls and/or oversteepened sections of the bed profile. In some instances, knickpoint erosion may instigate river capture, wherein flow that was previously part of a separate basin is realigned and captured as a headward-cutting channel eats through the drainage divide over time. An underfit stream now flows within the abandoned valley (i.e. the stream is much smaller than the river that created the valley itself). Elsewhere, stepped longitudinal profiles with multiple waterfalls reflect the recurrence of uplift events and the hardness of bedrock layers through which knickpoint retreat occurs. The pulsed nature of bed incision and knickpoint retreat in tectonically active settings is often accompanied by dramatic influxes of sediment from hillslope failures, some of which dam the river with variable longevity. Alternatively, lateral displacement along fault-lines during earthquake events can realign and/or reconfigure river systems. This can occur in a lateral dimension or vertical dimension. River responses to long-term changes in valley setting Rivers are products of the valleys in which they flow. Longterm changes to valley morphology reflect geologic controls. For example, progressive knickpoint retreat along trunk and tributary rivers at the plate margin creates series of dissected gorges in escarpment-dominated landscapes at the margins of pull-apart basins. These valleys cut backwards and incise far more rapidly than they widen. Changes to valley floor slope and valley width over millions of years induce transitions from a gorge to a partly confined valley with bedrock-controlled discontinuous floodplains and subsequently to a partly confined valley with planform-controlled discontinuous floodplains. River responses to major sediment inputs Rivers respond to marked increases in sediment load by aggrading. In some instances this may bring about profound landscape responses. For example, volcanic eruptions can drape vast volumes of material across a landscape, transforming incised bedrock streams into highly sedimentcharged systems that may infill valleys to considerable depth, promoting the development of braided rivers. These localised and irregular disturbance events are relatively spatially constrained (i.e. they occur in semi-predictable places, determined primarily by tectonic setting). Volcanic disruptions to river systems occur primarily in subduction and pull-apart settings and in response to hot spot activity (i.e. areas of thin crust through which molten materials from the upper mantle are released at the Earth's surface). Volcanic events are generally recurrent (i.e. they occur at the same place on repeated occasions, and resulting materials build up over time). Landscape responses are fashioned by the magnitude of an eruption, resulting sediment inputs and the interval between events (i.e. the length of time over which sediment reworking occurs). In general terms, volcanic landscapes that have not experienced an eruption for a significant period tend to become deeply etched bedrock-controlled systems. These rivers are resilient to change during flood events. However, eruptions bring about dramatic transformations, altering all attributes of the river. Lahars and debris flow deposits line valley floors. Aggradation induces braided rivers with an array of midchannel depositional geomorphic units. Typically, these are short- to medium-term adjustments post-eruption, as the river progressively adapts to prevailing flow–sediment conditions by incising into its bed (i.e. flow conditions remain relatively consistent over time, while the rate of sediment production is not maintained). Incision and reworking promote a transition back to increasingly imposed river morphologies. Downstream transfer of materials accentuates bed incision and the deeply etched character of the landscape. The imprint of volcanic events brings about a range of localised and off-site impacts. Tephra deposits may create a significant drape of materials over vast areas. Rivers subsequently flow within very light, low-density, highly porous materials, such that coarse bed material is readily conveyed within the channel (often as suspended load). In other settings, ignimbrite flows may infill valleys and create plateaulike landscapes with caps of extremely resistant materials. Valley incision and headward retreat subsequently demarcate these materials as knickpoints and waterfalls along longitudinal profiles. Long-term erosion of volcanic landscapes can create inverted relief. This occurs when lava flows infill valleys, flattening out the ground surface. As the thicker basaltic materials are often more resistant to erosion than the surrounding country rock, long-term progressive erosion may result in basalt-peaked caps derived from materials previously deposited on valley floors as the high points in these landscapes. While volcanic events induce massive sediment inputs into riverscapes over irregular but infrequent timescales, more recurrent but much smaller sediment inputs occur in response to landslides and associated hillslope instability events. A range of outcomes may occur, dependent upon the amount of sediment input, the size of the valley and the capacity of the river to rework these deposits. In extreme instances the valley may become blocked, forming a dam and lake. This alters the base level of the trunk stream, resulting in aggradation and delta growth within the lake. Downstream, the channel responds to reduced sediment loads by incising. Eventually the dam may break. This results initially in extensive flooding and erosion of downstream reaches. Subsequently, the massive influx of deposits induces aggradation as a sediment slugthrough the system. Extreme landscape responses to landslide events are especially pronounced following earthquakes or extreme storms (cyclones). Such scenarios are especially pronounced in steep, dissected terrains close to plate margins in regions with (sub)tropical climates. In others settings, hillslope-derived materials may be stored along valley floors for a considerable period of time. This is primarily determined by valley width, and associated hillslope–valley floor connectivity and the space for sediment storage. If these deposits are not accessible to the channel, they may have a negligible impact upon river behaviour and change. River responses to climate change (flow regime and ground cover changes) Impacts of climate change and variability may be manifest through localised extreme events, semi-regular, systematic changes (e.g. glacial–interglacial cycles) or long-term adjustments associated with the movement of tectonic plates. Of primary concern here is the impact of changing boundary conditions and disturbance events upon the way in which rivers operate, and their evolution. Examples of system responses to climateinduced alterations to the flow–sediment balance and ground cover are outlined below. Climate change induces marked variability in the character, behaviour and evolution of river systems in glaciated and non-glaciated landscapes. Phases of glacial activity in mountainous terrain induce extensive erosion and sculpting of landscapes. The mountains themselves are etched and denuded, while valleys are carved. Stripped surficial materials and bedrock are broken down and conveyed considerable distances from source. As a consequence, the boundary conditions upon which rivers operate are transformed. Transitional climatic phases at the ends of ice ages are periods of intensive geomorphic activity. This period is referred to as the paraglacial interval. Melting glaciers and ice sheets result in pronounced discharge variability. Hillslopes are unstable, as previously supporting ice has melted, and vegetation cover is negligible. This results in extensive sediment movement, aggrading valley floors and the for mation of large alluvial fans. Steep slopes, abundant bedload-calibre material and fluctuating discharge result in braided river systems. Extensive braidplains (or valley sandar) are evident at the margins of many contemporary glaciers or ice sheets. Over time, discharge is reduced and streams incise into their beds, creating extensive terrace sequences (12.7d). Rivers retain extensive sediment loads, and braided channel planforms extend well beyond the mountain front. Amelioration of climatic conditions over thousands of years results in less variable discharges, diminishing sediment loads and increases in vegetation cover on hillslopes and valley floors. Rivers respond by changing to wandering gravel-bed or active meandering systems. In some instances, post-glacial climate changes may generate some truly epic landscapes, inducing profound alterations to river systems. This is exemplified by breaching of ice-dammed lakes, which release vast volumes of water in truly catastrophic flows (termed jokulhlaups). These floods may etch and sculpt vast terrains, fashioning future drainage networks and resulting river morphologies. Elsewhere, streams beneath glaciers or significant meltwater flow can realign drainage networks (a form of river capture). Many landscapes retain a significant climatic memory from these post-glacial events. Although non-glaciated terrains are not subjected to paraglacial sedimentation and breaching of ice-dammed lakes, dramatic landscape responses to changing climatic conditions may occur in these settings. For example, former river courses in some desert landscapes have been draped by sand dunes in response to drier climates and reduced vegetation cover during glacial periods. Some non-glaciated landscapes have been subjected to progressive drying over hundreds of thousands of years. Marked reductions in channel geometry, along with notable decreases in sinuosity and bed material size, result in rivers that are clearly undersized for the valleys within which they flow. Climate amelioration and vegetation growth brought about dramatic transformation of the flow–sediment balance, whereby the energy of the system was diminished to such an extent that the river became a discontinuous watercourse with a finegrained swamp that accumulates suspended-load deposits. These valley floor deposits have subsequently been incised to create a continuous channel. Glacial–interglacial cycles induce significant sea level change (eustasy). This may bring about geomorphic adjustment along the lower course of rivers. Sea levels may be lowered by 120–150 m during glacial maxima, essentially extending river courses onto what is now the continental shelf. The nature of geomorphic adjustment varies for differing fluvial–marine interactions (i.e. whether a delta or estuary is present) and the nature/extent of the continental shelf itself. Profound adjustments are noted along the lowland plains of large rivers, where incised valleys and fills develop significant terrace sequences. These terraces, in turn, constrain subsequent channel responses during periods of rising sea levels. Alternatively, the profound weight of accumulated deposits along the lowland plains of rivers, or in inland-draining (endorheic) basins, may induce subsidence via isostatic adjustment. Given the very low slope of these settings, avulsion may be experienced along these low-energy, suspended-load rivers. Ongoing climate changes associated with global warming are bringing about marked geomorphic transitions for some rivers. For example, melting permafrost has increased discharges and the erosive potential of many rivers that drain into polar regions. Impacts of ice flows following spring melt have been accentuated. This exemplifies regionally specific patterns and trends in the evolutionary adjustment of rivers. Finally, the impacts of climate changes upon rivers must be related to the magnitude–frequency relations of formative events, especially the geomorphic effectiveness of extreme floods. As noted previously, there is significant variability in response in differing landscape and climatic settings. This reflects the sensitivity/resilience of a river, and the extent to which the river is attuned to seasonal and interannual variability in discharge. In some instances, extreme floods may exert a profound imprint or memory upon the system, whereby the river is subsequently unable to adjust its boundaries. Depending upon the condition of the system at the time of the event, and associated availability of sediment, flows may be highly erosive or highly depositional. Either way, transformation of channel boundaries exerts a significant influence upon the subsequent evolutionary adjustments of the river. These various pathways and rates of geomorphic evolution are meaningfully captured using the river evolution diagram.

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

These determinations reflect the imposed boundary conditions within which a river operates, as shown by the outer band of the river evolution diagram. The width of the outer band increases from confined through partly confined to laterally unconfined settings, as the potential range of variability increases. Rivers can more readily adopt differing morphologies and behavioural attributes if there is space for the channel to adjust on the valley floor. A similar degree of variability is evident in the width of the inner band on these figures. This reflects the natural capacity for adjustment as determined by flux boundary conditions. The width of this inner band represents the range of states that the river can adopt while still being considered to be the same type of river (i.e. retaining a consistent set of core geomorphic attributes that reflect the character and behaviour of that river type). In a sense, this is a measure of the sensitivity of the river, as it records the ease with which the river is able to adjust. As indicated for the potential range of variability, the width of the inner band is greatest in laterally unconfined settings. River responses to disturbance events are indicated by the arrows shown at the top of the inner band on the river evolution diagram. The spacing of the arrows indicates their frequency, while the size of an arrow indicates its magnitude. In most instances, disturbance events promote river adjustments but the reach remains within the inner band (i.e. perturbations fall within the natural capacity for adjustment). River adjustment within the inner band may breach intrinsic threshold conditions, marking a shift in the way energy is used (either concentrated or dispersed). Typically, this reflects an adjustment in the character or distribution of resisting forces (e.g. bed resistance, form resistance, resistance induced by riparian vegetation or wood). These internal adjustments alter the assemblage of erosional and depositional landforms on the valley floor, yet fall within the behavioural regime of the river. In other instances, changes to the prevailing flux boundary conditions and/or severe disturbances may bring about changes to the formative processes that fashion river morphology (i.e. river change has occurred). This scenario is highlighted by the shift in the position of the inner band. In these instances, altered stream power relationships reflect differing energy use in relation to prevailing flux boundary conditions. Reaches now operate within a different inner band on the river evolution diagram, with altered energy conditions. The shape of the pathway for adjustment, shown by the jagged line within the inner band, has a different form for the new river type, depicting a change in process–form associations along the valley floor, such that there is a change in river morphology. The new configuration represents a different type of river, with a different appearance and set of formative processes (behaviour). Inevitably, there may be some overlap in the position of former and contemporary bands, and some geomorphic units may be evident in both situations. However, the assemblage of geomorphic units in the two bands differs, reflecting a change to river character and behaviour. The shift in the position of the inner band can be induced by a press disturbance that exceeds an extrinsic threshold. This usually reflects alteration to flux boundary conditions, as modified flow and sediment transfer regimes (i.e. impelling forces) drive river change. In this case, the time that is required for recovery following perturbation is longer than the recurrence interval of disturbance events. Effectively, the previous configuration of the river was unable to cope with changes to the magnitude and rate of applied stress. Rare floods of extreme magnitude, or sequences of moderate-magnitude events that occur over a short interval of time, may breach extrinsic threshold conditions, transforming river character and behaviour. Dependent on the subsequent set of process–form associations adopted by the river, the natural capacity for adjustment may widen or contract as the new type of river adjusts to different flux boundary conditions. The position of the inner band within the potential range of variability (the outer band) indicates whether the change in river type marks a transition to a higher energy state (an upwards adjustment) or a lower energy state (downward adjustment). Changes to the amplitude, frequency and shape of the pathway of adjustment within the inner band indicate how the river responds to pulse disturbances of varying magnitude and frequency. In some cases, change may occur during a threshold breaching flood. In this instance the two inner bands are located adjacent to each other and the date of the disturbance event is noted. In other cases change may be lagged or occur progressively over time. In these instances, the space between the two inner bands is widened to depict whether change occurred over years or decades. In more complex situations, transitional river types are depicted on the diagram. For simplicity, only a major shift between one river type and another is shown in the examples outlined below. Emphasis is placed upon the nature of evolutionary changes to the river, timeframes over which these changes occur and evolutionary trajectory. Disturbance events that have the capacity to induce changes in other settings are unable to bring about significant geomorphic adjustments along confined rivers, as the inherent resilience of the system is too strong. Perturbations to the flow and sediment regime are accommodated by instream adjustments to hydraulic resistance, such as the nature and distribution of bedforms, dissipating flow energy. Adjustments to river character and behaviour are negligible and the river type remains the same. Millions of years of valley widening may allow for out-of channel deposition and generation of floodplain pockets, but the assemblage of erosional and depositional geomorphic units along the reach is likely to remain consistent over tens of thousands of years (at least). A different pattern of responses to changes in external stimuli may be experienced in partly confined valley settings, where the potential range of variability is somewhat broader than in confined valleys. This enables a greater range of possible river morphologies to develop. Antecedent controls and prevailing flux boundary conditions shape the contemporary configuration of the river. A bedrock-controlled discontinuous floodplain river has negligible capacity for adjustment because of the bedrock-imposed setting. Valley widening over tens of thousands of years results in progressive transition to a planform-controlled situation. The example demonstrates potential adjustments in this situation, as there is greater capacity for adjustment because of the greater degrees of freedom. Local areas of the channel are able to adjust their planform within the partly confined valley. For example, lateral migration may form ridges and floodchannels within the vertically accreted silty floodplain. In this instance, the natural capacity for adjustment has shifted to a lower energy river type. This is indicated on the river evolution diagram by a downward shift in the position of the inner band (the natural capacity for adjustment) within the outer band (the potential range of variability). In addition, the range of river behaviour has been reduced (i.e. the width of the inner band has narrowed; note the logarithmic scale). Rivers are more sensitive to change in laterally unconfined valley settings relative to partly confined and confined valleys (i.e. the potential range of variability and the natural capacity for adjustment are greatest in laterally unconfined valley settings). Changes are shown from a braided configuration to a meandering mixed-load system, from a mixed-load meandering to a suspended-load meandering river, from a gravel-bed braided to a low-sinuosity sand-bed river and from a braided to a fine-grained discontinuous watercourse. These changes reflect alterations to both the impelling forces that promote change (i.e. less variability in flow, less coarse-sized material on the valley floor, etc.) and internal system adjustments that modify the pattern and extent of resistance. A major shift in the assemblage of geomorphic units ensues, resulting in altered patterns of mid-channel and bank-attached geomorphic units, and processes of floodplain formation and reworking. Channel geometry and bedform assemblages are transformed as well. This reflects the adoption of a lower energy river type within the same landscape setting. In some instances, an increase in resistance increases the capacity of the system to trap finer grained materials, thereby aiding the transition to a single-channelled or discontinuous channel configuration. Increased stability enhances prospects for vegetation development on the valley floor. As a result, the natural capacity for adjustment is narrower, reflecting a reduction in the range of behaviour. Changes to energy relationships reflect the consumption of energy, altering the pathway of adjustment. For example, the transition from braided to meandering configurations is marked by a switch from tight chaotic oscillations reflecting recurrent reworking of materials on the channel bed to a jagged shape that reflects the occasional formation of cut-offs and subsequent readjustment of channel geometry, planform and slope. Post-glacial adjustments to flow and sediment fluxes commonly induced changes from a braided to a meandering channel planform. In the early post-glacial interval, abundant sediment, highly variable flows and negligible vegetation cover promoted the development of braided rivers. A wide range of mid-channel bars and shifting channels of varying size characterised these bedload dominated systems. Progressive reduction in sediment availability in the post-glacial era, along with reduced variability in discharge and progressive encroachment of vegetation onto the valley floor, brought about the transformation of many of these braided rivers into mixed-load meandering systems by the mid-Holocene. These rivers are now characterised by laterally migrating single channels with point bars and associated instream geomorphic units, and an array of laterally and vertically accreted floodplain forms. This shows the transformation from a mixed-load laterally migrating channel into a slowly migrating suspended-load river with a much smaller channel capacity. This transition reflects a decline in fluvial activity driven by changes to the discharge regime. Different pathways and rates of adjustment may be experienced by different types of rivers subjected to similar climatically induced changes to prevailing flow and sediment fluxes. A range of tools and approaches used to analyse and interpret river evolution is outlined in the following section.

Reading the landscape to interpret river evolution

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

Conclusion

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

References:

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

UNIT-7: CLASSIFICATION OF NATURAL STREAMS BY D. L. ROSGEN

It has long been a goal of individuals working with rivers to define and understand the processes that influence the pattern and character of river systems. The differences in river systems, as well as their similarities under diverse settings, pose a real challenge for study. One axiom associated with rivers is that what initially appears complex is even more so upon further investigation. Underlying these complexities is an assortment of interrelated variables that determines the dimension, pattern, and profile of the present-day river. The resulting physical appearance and character of the river is a product of adjustment of its boundaries to the current streamflow and sediment regime.

River form and fluvial process evolved simultaneously and operate through mutual adjustments toward self-stabilization. Obviously, a classification scheme risks oversimplification of a very complex system. While this may appear presumptuous, the effort to categorize river systems by channel morphology is justified in order to achieve, to some extent, the following objectives:

1. Predict a river's behavior from its appearance;

2. Develop specific hydraulic and sediment relations for a given morphological channel type and state;

3. Provide a mechanism to extrapolate site-specific data collected on a given stream reach to those of similar character;

4. Provide a consistent and reproducible frame of reference of communication for those working with river systems in a variety of professional disciplines.

2. **Stream classification review**

A definition of classification was offered by Platts (1980) where "classification in the strictest sense means ordering or arranging objects into groups or sets on the basis of their similarities or relationships." The effort to classify streams is not new. Davis (1899) first divided streams into three classes based on relative stage of adjustment: youthful, mature, and old age. Additional river classification systems based on qualitative and descriptive delineations were subsequently developed by Melton (1936) and Matthes (1956).

Straight, meandering, and braided patterns were described by Leopold and Wolman (1957). Lane (1957) developed quantitative slope-discharge relationships for braided, intermediate, and meandering streams. A classification based on descriptive and interpretive characteristics was developed by Schumm (1963) where delineation was partly based on channel stability (stable, eroding, or depositing) and mode of sediment transport (mixed load, suspended load, and bedload).

A descriptive classification was also developed by Culbertson et al. (1967) that utilized depositional features, vegetation, braiding patterns, sinuosity, meander scrolls, bank heights, levee formations, and floodplain types. Thornbury (1969) developed a system based on valley types. Patterns were described as antecedent, superposed, consequent, and subsequent. The delineative criteria of these early classification systems required qualitative geomorphic interpretations creating delineative inconsistencies. Khan (1971) developed a quantitative classification for sand-bed streams based on sinuosity, slope, and channel pattern.

To cover a wider range of stream morphologies, a descriptive classification scheme was developed for and applied on Canadian Rivers by Kellerhals et al. (1972, 1976), Galay et al. (1973), and Mollard (1973). The work of these Canadian researchers provides excellent description and interpretation of fluvial features. This scheme has utility both for aerial photo delineation and for describing gradual transitions between classical river types. and to date offers the most detailed and complete list of channel and valley features. The large number of possible interpretative delineations, however, makes this scheme quite complex for general planning objectives.

An attempt to classify rivers in the great plains region using sediment transport, channel stability, and measured channel dimensions was developed by Schumm (1977). Classifying stream systems on the basis of stability is often difficult because of the qualitative criteria can vary widely among observers leading to inconsistencies in the classification. Similarly, data on ratio of bedload to total sediment load as needed in this classification, while useful, often is not readily available to those who need to classify streams.

Brice and Blodgett (1978) described four channel types of: braided, braided pointbar, wide-bend point-bar, and equi-width point-bar. A descriptive inventory of alluvial river channels is well documented by Church and Rood (1983). This data set can be very useful for many purposes including the grouping of rivers based on similar morphological characteristics. Nanson and Croke (1992) presented a classification of flood plains that involved particle size, morphology of channels, and bank materials. This classification has some of the same criteria of channel type as presented in this paper, but is restricted to flood plains. Pickup (1984) describes the relation of sediment source and relative amounts of sediment to various aspects of river type, but is not a classification of channels. Recent documentation by Selby (1985) showed a relationship between the form and gradient of alluvial channels and the type, supply and dominant textures (particle sizes) of sediments. This relationship utilizes the Schumm (1977) classification in that an increase in the ratio of bed material load to total sediment load with a corresponding increase in channel gradient leads to a decrease in stability causing channel patterns to shift from a meandering to braided channel form. In his classification, Selby (1985) treats anastomosed and braided channel patterns similarly. However, the anastomosed rivers are not similar to braided rivers in slope, adjustment processes, stability, ratio of bed material to total load or width/depth ratios as shown by (Smith and Smith, 1980).

Typically, theoretically derived schemes, often do not match observations. To be useful for extrapolation purposes, restoration designs, and prediction, classification schemes should generally represent the physical characteristics of the river. With certain limitations, most of these classification and/or inventory systems met the objectives of their design. However, the requirement for more detailed, reproducible, quantitative applications at various levels of inventory over wide hydrophysiographic provinces has led to further development of classification schemes.

2. **Stream classification concepts**

The morphology of the present day channel is governed by the laws of physics through observable stream channel features and related fluvial processes. Stream pattern morphology is directly influenced by eight major variables including channel width, depth, velocity, discharge, channel slope, roughness of channel materials, sediment load, and sediment size (Leopold et al., 1964). A change in any one of these variables sets up a series of channel adjustments which lead to a change in the others, resulting in channel pattern alteration. Because stream morphology is the product of this integrative process, the variables that are measurable should be used as stream classification criteria.

The directly measurable variables that appear from both theory and experience to govern channel morphology have been included in the present classification procedure. These "delineative criteria" interact with one another to produce a stream's dominant features. The present classification system has evolved from field observation of hundreds of rivers of various sizes in all the climatic regions of North America, experience in stream restoration, extensive teaching, and practical applications of the classification system by many hydrologists, geomorphologists, fisheries experts, and plant ecologists. Initial efforts to develop the classification procedure began in 1973, and a preliminary version was presented to the scientific community (Rosgen, 1985).

3. Stream classification system

The classification of rivers is an organization of data on stream features into discreet combinations. The level of classification should be commensurate with the initial planning level objective. Because these objectives vary, a hierarchy of stream classification and inventories is desirable because it allows an organization of stream inventory data into levels of resolution from very broad morphological characterizations to discreet, measured, reach-specific descriptions. Each level should include appropriate interpretations that match the inventory specificity. Further, general descriptions and characteristics of stream types should be able to be divided into even more specific levels. The more specific levels should provide indications of stream potential, stability, existing "states", etc., to respond to higher resolution data and interpretations when planning needs change.

Current river "state" and influences on the modern channel by vegetation, flow regime, debris, depositional features, meander patterns, valley and channel confinement, streambank erodibility, channel stability, etc., comprise additional parameters that are considered critical to evaluate by stream type at a more detailed inventory level (Level III). However, for the sake of brevity and clarity, this paper will focus on the first two levels, the broad geomorphic characterization (Level I) and the morphological description (Level II) which incorporates the general character of channel form and related interpretations. Portions of the data used for detailed assessment levels are contained in the sub-type section of the earlier classification paper (Rosgen, 1985).

4.1. Geomorphic characterization (level I)

The purpose of delineation at this level is to provide a broad characterization that integrates the landform and fluvial features of valley morphology with channel relief, pattern, shape, and dimension. Level I combines the influences of climate, depositional history, and life zones (desert shrub, alpine, etc.) on channel morphology. The presence, description, and dimensions of floodplains, terraces, fans, deltas and outwash plains are a few examples of valley features identified. Depositional and erosional history overlay channel patterns at this level. Generalized categories of "stream types" initially can be delineated using broad descriptions of longitudinal profiles, valley and channel cross-sections, and plan-view patterns.

Longitudinal profiles

The longitudinal profile, which can be inferred from topographic maps, serves as the basis for breaking the stream reaches into slope categories that reflect profile morphology. For example, the stream types of Aa + (Fig. 1) are very steep, (greater than 10%), with frequently spaced, vertical drop/scour-pool bed features. They tend to be high debris transport streams, waterfalls, etc. Type A streams are steep (4–10% slope), with steep, cascading, step/pool bed features. Type B streams are riffle-dominated types with "rapids" and infrequently spaced scour-pools at bends or areas of constriction. The C, DA, E and F stream types are gentle-gradient riffle/pool types. Type G streams are "gullies" that typically are step/pool channels. Finally, the D type streams are braided channels of convergence/divergence process that lead to localized, frequently spaced scour/depositional bed forms.

Bed features are consistently found to be related to channel slope. Grant et al. (1990) described bed features of pools, riffles, rapids, cascades, and steps as a function of bed-slope gradient. Using their bed form descriptions, the above described stream types were plotted against the corresponding slope ranges reported by Grant et al. (1990). "Groupings", (Fig. 2), were apparent for riffle/pool stream types (C, E, and F) at less than 2%, rapids at 2–4% in "B" and "G", cascades in slopes 4–10% in type A streams, and steps for slopes 4–40% in types A and Aa + streams. Because gradient and bed-feature relationships are integral to the delineation of stream type categories, "stream types" are more than just "arbitrary units". Bed morphology can be predicted from stream type by using bed-slope indices.

Cross-section morphology

The shape of the cross-section that would indicate a narrow and deep stream as opposed to a wide and shallow one can be inferred at this broad level. The manner in which the channel is incised in its valley can also be deduced at this level as well as information concerning floodplains, terraces, colluvial slopes, structural control features, confinement (lateral containment), entrenchment (vertical containment), deep, confined, and, entrenched. The width of the channel and valley are similar. This contrasts with type C streams, where the channel is wider and shallower with a well-developed floodplain and a very broad valley. Type E streams have a narrow and deep channel (low width/depth ratio) but have a very wide and well developed floodplain. Type F streams have wide and shallow channels, but are an entrenched meandering channel type with little to no developed floodplain. Type G channels have low width/depth ratio channels similar to type E streams except they are well entrenched (no floodplain), are steeper, and less sinuous than type E streams.

Plan view morphology

The pattern of the river is classed as relatively straight (A stream types), low sinuosity (B stream types), meandering (C stream types), and tortuously meandering (E stream types). Complex stream patterns are found in the multiple channel, braided (D) and anastomosed (DA) stream types. Sinuosity can be calculated from aerial photographs and often, like slope, serves as a good initial delineation of major stream types. These river patterns have integrated many processes in deriving their present form and thus, provide interpretations of their associated morphology.

Even at this broad level. of delineation, consistency of dimension and associated pattern can be observed by broad stream types. Meander width ratio (belt width/ bankfull surface width) was calculated by general categories of stream types for a wide variety of rivers. Early work by Inglis (1942) and Lane (1957) discussed meander width ratio but the values were so divergent among rivers that the ratio appeared to have little value. When stratified by general stream types, however, the variability appears to be explained by the similarities of the morphological character of the various stream types. This has value not only for classification and broad-level delineations, but also for describing the most probable state of channel pattern **in** stream restoration work.

Discussion

Interpretations of mode of adjustment — either vertical, lateral, or both — and energy distribution can often be inferred in these broad types. Many variables that are not discrete delineative variables integrate at this level to produce an observable morphology. A good example of this is the influence of a deep sod-root mass on type E streams that produces a low width/depth ratio, low meander length, low radius of curvature, and a high meander width ratio. Vegetation is not singled out for mapping at this level, but is implicit in the resulting morphology. If this vegetation is changed, the width/depth ratio and other features will result in adjustments to the type C stream morphology. Detailed vegetative information, however, is obtained at the channel state level.

Delineating broad stream types provides an initial sorting within large basins and allows a general level of interpretation. This leads to organization and prioritization for the next more detailed level of stream classification.

4.2. The morphological description (level II) General description

This classification scheme is delineated initially into the major, broad, stream categories of A–G as shown in Fig. I and Table 2. The stream types are then broken into discreet slope ranges and dominant channel-material particle sizes. The stream types are given numbers related to the median particle size diameter of channel materials such that 1 is bedrock, 2 is boulder, 3 is cobble, 4 is gravel, 5 is sand, and 6 is silt/clay. This initially produces 42 major stream types as shown in.

A range of values for each criterion is given in the key to classification for 42 major stream types. The range of values chosen to represent each delineative criterion is based on data from a large assortment of streams throughout the United States, Canada and New Zealand. A recent data set of 450 rivers was statistically used to refine and test previous ranges of delineative criteria as described in the author's earlier publication (Rosgen, 1985).

Histograms were drawn of the distribution of values of each delineative criterion for each channel type. From the histograms of 5 criteria for 42 major stream types, the mean and "frequent range" of values were recorded. The most frequently observed values seemed to group into a recognizable "river form" or morphology. When values were outside of the range of the "most frequently observed" condition, a distinctly different morphology was identified. As a result, the delineation of unique stream types representing a range of values amongst several variables were established.

The classification can be applied to ephemeral as well as perennial channels with little modification. Bankfull stage can be identified in most perennial channels through observable field indicators. Although, these bankfull stage indicators, are often more elusive in ephemeral channels.

The morphological variables can and do change even in short distances along a river channel, due to such influences of change as geology and tributaries. Therefore, the morphological description level incorporates field measurements from selected reaches, so that the stream channel types used here apply only to individual reaches of channel. Data from individual reaches are not averaged over entire basins to describe stream systems. A category may apply to a reach only a few tens of meters or may be applicable to a reach of several kilometers.

Data is obtained from field measurements of representative or "reference reaches." The resultant stream type as delineated can then be extrapolated to other reaches where detailed data is not readily available. In similar valley and lithological types, stream types can often be delineated using these reference reaches through the use of aerial photos, topographic maps, etc.

Continuum concept

When the variables which make up the range of values within a stream type change, there is more often than not, a change in stream type. Exceptions occur infrequently, where values of one variable may be outside of the range for a given stream type.

This level recognizes and describes a continuum of river morphology within and between stream types. The continuum is applied where values outside the normal range are encountered but do not warrant a unique stream type. Often the general appearance of the stream and the associated dimensions and patterns of the stream do not change with a minor value change in one of the delineative criteria. For example, slope values as shown in Fig. 5, using the continuum concept, are not "lumped", but rather are sorted by sub-categories of: a + (steeper than 0.10), a (0.04-0.10), b (0.02- 0.039), c (less than 0.02) and c- (less than 0.001).

The application of this concept allows an initial classification of a C4 stream type (a gravel bed, sinuous, high width/depth ratio channel with a well-developed floodplain. If the slope of this stream was less than 0.001, then the stream type would be a C4c-.

Rivers do not always change instantaneously, under a geomorphic exceedance or "threshold". Rather, they undergo a series of channel adjustments over time to accommodate change in the "driving" variables. Their dimensions, profile and pattern reflects on these adjustment processes which are presently responsible for the form of the river. The rate and direction of channel adjustment is a function of the nature and magnitude of the change and the stream type involved. Some streams change very rapidly, while others are very slow in their response.

Delineative criteria

At this level of inventory each reach is characterized by field measurements and validation of the classification. The delineation criteria and ranges for various stream types are shown in Fig. 5. This classification key also represents the sequential process for classification. The classification process starts at the top of the chart (single or multiple thread channels), and proceeds downward through channel materials and slope ranges.

Entrenchment

An important element of the delineation is the interrelationship of the river to its valley and/or landform features. This interrelationship determines whether the river is deeply incised or entrenched in the valley floor or in the deposit feature. Entrenchment is defined as the vertical containment of river and the degree to which it is incised in the valley floor (Kellerhals et al., 1972). This makes an important distinction of whether the flat adjacent to the channel is a frequent floodplain, a terrace (abandoned floodplain) or is outside of a flood-prone area. A quantitative expression of this feature, "entrenchment ratio" was developed by the author so that various mappers could obtain consistent values. The entrenchment ratio is the ratio of the width of the flood-prone area to the bankfull surface width of the channel. The flood-prone area is defined as the width measured at an elevation which is determined at twice the maximum bankfull depth. Field observation shows this elevation to be a frequent flood (50 year return period) or less, rather than a rare flood elevation. These categories were empirically derived based on hundreds of streams. As with other criteria, the measured entrenchment ratio value may lie somewhat outside of the classification range. When this occurs, the author applies the continuum concept which allows for a category description where the entrenchment is either greater or less than the most frequently observed value for a given morphology. The continuum allows for a change of \pm 0.2 units where the corresponding delineative criteria still match the range of variables consistent for that type. In this case, all of the other attributes must be considered before assigning a stream type.

Width/depth ratio

The width/depth ratio describes the dimension and shape factor as the ratio of bankfull channel width to bankfull mean depth. Bankfull discharge is defined as the momentary maximum peak flow; one which occurs several days in a year and is often delineation and significance of bankfull discharge are found in Leopold et al. (1964), Dunne and Leopold (1978), and Andrews (1980). Hydraulic geometry and sediment transport relations rely heavily on the frequency and magnitude of bankfull discharge.

Osborn and Stypula (1987) utilized width/depth ratio to characterize stream channels for hydraulic relations using channel boundary shear as a function of channel shape.

For this classification, values of low width/depth ratio are those less than 12. Values greater than 12 are moderate or high. Average values and ranges are shown in the stream type summaries. As in the continuum concept, applied to entrenchment ratio, there is an occasion where width/depth ratio values can vary by ±2 units without showing a different morphology. This does not occur very frequently, but the continuum allows for some flexibility to fit the stream type into a "dominant" morphology.

Sinuosity

Sinuosity is the ratio of stream length to valley length. It can also be described as the ratio of valley slope to channel slope. Mapping sinuosity from aerial photos is often possible, and interpretations can often be made of slope, channel materials, and entrenchment once sinuosity is determined. Values of sinuosity appear to be modified by bedrock control, roads, channel confinement, specific vegetative types, etc. Generally speaking, as gradient and particle size decreases, there is a corresponding increase in sinuosity. The continuum as mentioned earlier also applies and adjustments of + or -0.2 can be applied to this delineative criteria. Meander geometry characteristics are directly related to sinuosity following minimum expenditure of energy concepts. Initial studies by Langbein and Leopold (1966) suggested that a sine generated curve describes symmetrical meander paths. From this observation they predicted the radius of curvature of meander bends from meander wavelength and channel sinuosity. In comparing observed versus predicted values of radius of curvature for 79 streams, Williams (1986) found this relation to be highly correlated when applied to an expanded data set. This demonstrates the interrelationship of sinuosity to meander geometry. Based on such relations and the relative ease of determination, sinuosity was selected as one of the delineative criteria for stream classification.

Channel materials

The bed and bank materials of the river is not only critical for sediment transport and hydraulic influences but also modifies the form, plan and profile of the river. Interpretations of biological function and stability also require this information. Often a good working knowledge of the soils associated with various landforms can predict the channel materials at the broad delineation level. Reliable estimates of the soil characteristics for glacial till, glacial outwash, alluvial fans, river terracgs, lacustrine and eolian deposits, and residual soils can be derived from mapped lithology.

Field determination of channel materials for this classification system utilizes the "pebble count" method developed by Wolman (1954), with a few modifications to account for bank material and for sand and smaller sizes. This is a determination the frequency distribution of particle sizes that make up the channel. The pebble count data is plotted as cumulative percent and percent of total distribution. The dominant particle size is identified in the cumulative percent curve as the median size of channel materials or size that 50% of the population is of the same size or finer *(D50).* This data is used in biological evaluation, sediment supply assessment, and other interpretative applications. *Slope* Water surface slope is of major importance to the morphological character of the channel and its sediment, hydraulic, and biological function. It is determined by measuring the difference in water surface elevation per unit stream length. Typically, slope is often measured through at least 20 channel widths or two meander wavelengths. As observed with the other delineative variables, slope values less or greater than the most frequently observed ranges can occur. These can occur without a significant change in the other delineative criteria for that stream type. The most frequently observed slope categories and applications of the continuum concept for slope is shown in Fig. 5. In broad-level delineations, slopes can often be estimated by measuring sinuosity from aerial photos and measuring valley slope from topographic maps (valley slope/sinuosity = channel slope). The basin and associated landform relief can also be used to estimate stream slope ranges, as for example terraces and slopes of alluvial fans.

5. Application

Past observations of adjustments of stream systems often provide insight into sensitivity and consequence of change. Stream system changes can be due to flow, sediment, or many of the interrelated variables that have produced the modern channel. If changes produce disequilibrium, similar stream types receiving similar impacts may be expected to respond the same. If the observer knows the stream type of the disturbed reach, and has cross-section, bank erosion, sediment data, riparian vegetation and fisheries data, this information can be used predicatively to evaluate the risk and sensitivity to disturbance.

5.1. Evolution of stream types

In reviewing historical aerial photos, observations can be made of progressive stages in channel adjustment. These adjustments occur partially as a result of change in stream-flow magnitude and/or timing, sediment supply and/or size, direct disturbance, and vegetation changes. These observed changes in channel morphology over time can be communicated in terms of stream type changes. For example, due to streambank instability, and a resultant increase in bank erosion rate, the stream increased its width/depth ratio; decreased sinuosity; increased slope; established a bimodal particle size distribution; increased bar deposition; accelerated bank erosion; and decreased the meander width ratio. These changes can be described more simply as a series of progressive changes of channel adjustment in stream type from an E4 to C4 to C4 (bar-braided) to D4. Another example of channel adjustment where morphological patterns are changed sufficient to indicate a shift in stream type is shown in Fig. 9. In this scenario, a change in streambank stability led to an increase in width/depth ratio and slope, and a decrease in sinuosity and meander width ratio. As the slope steepened along with a high width/depth ratio, chute cutoffs occurred across large point bars creating a gulley. The stream abandoned its floodplain, decreased the width/depth ratio, steepened the slope and decreased sinuosity. This resulted in a change in base level as all of the tributaries draining into this stream were over-steepened. Sediment from both channel degradation and bank erosion was increased. As the banks continued to erode, the width/depth ratio and sinuosity both increased with a corresponding decrease in slope. The channel was still deeply entrenched, but eventually started to develop a floodplain at a new elevation. This stream eventually evolved under a changed sediment and flow regime into a sinuous, low gradient, low width/depth ratio channel with a well developed floodplain which matched the original morphology, except now exists at a lower elevation in the valley. This case is shown more simply in Fig. 9 as a shift from an E4 stream type to C4 to G4 to F4 and back to an E4 type. These changes have been well documented throughout western North America due to various reasons including climate change and adverse watershed impacts. The knowledge provided by observing these historical adjustments and the understand ing of the tendency of rivers to regain their own stability can assist those restoring disturbed river systems. Often the works of man try to "restore" streams back to a state that does not match the dimension, pattern and slope of the natural, stable form. As stream types change, there are a large number of interpretations associated with these "morphological shifts". Stream types can imply much more than what is initially described in its alphanumeric title. *5.2. Fish habitat* When physical structures are installed in channels to improve the fish habitat, the adjustment processes that occur sometimes create more damage than habitat. For example, Trail Creek in southeast Colorado, a C4 stream type, had a gabion check dam installed at 80% of the bankfull stage to create a plunge pool for fish. The results were; decreased upstream gradient; width/depth ratio increase; decreased mean bed particle diameter; and decreased competence of the stream to move its own sediment. The longitudinal profile of the river changed creating headward aggradation. With a decrease in slope, there was a corresponding increase in sinuosity that resulted in accelerated lateral channel migration and increased bank erosion. Subsequently, the stream abandoned the original channel and created a "headcut gulley" with a gradient that was twice the valley slope. This converted the C4 stream type to a G4 type in a period of approximately two years. The "new" stream type has abandoned its floodplain, is rejuvenating tributaries headward and creating excess sediment from stream degradation and bank erosion. This disequilibrium caused by the check dam is long-term and has deteriorated the habitat that the structure was initially designed to improve. Unfortunately, structures like this continue to be installed by well-meaning individuals without a clear understanding of channel adjustment processes. To prevent similar problems and to assist biologists in the selection and evaluation of commonly used in-channel structures, guidelines by stream type were developed (Rosgen and Fittante, 1986). In the development of these guidelines hundreds of fish habitat improvement structures were evaluated for effectiveness and channel response. A stream classification was made for each reach containing a structure. From this data, the authors rated various structures from "excellent" to "poor" for an extensive range of stream types. These guidelines provide "warning flags" of potential adverse adjustments to the river so that technical assistance may be obtained. In this manner, structures may be better designed to not only meet their objectives, but help maintain the stability and function of the river. Fisheries habitat surveys presently integrate this stream classification system (USDA, 1989). The objective for this integration is to determine the potential of the stream reach, current state, and a variety of hydraulic and sediment relations that can be utilized for habitat and biological interpretations.

5.3. Flow resistance

Application of the Manning's equation and the selection of a roughness coefficient N value to predict mean velocity is a common methodology used by engineers and hydrologists. The lack of consistent criteria for selection of the correct N values, however, creates great variability in the subsequent estimate of flow velocity. Barnes (1967), and Hicks and Mason (1991) produced photographs and a variety of stream data which was primarily a visual comparison approach for the selection of roughness coefficients. However, using these books for a visual estimate of roughness, actually involves looking at various stream types. The author classified each of the 128 streams described in both publications, noted the occurrence of vegetation influence, and plotted the bankfull stage N values by stream type. The remarkable similarity of N values by stream type for two data bases from two countries revealed another application for estimating a bankfull stage roughness coefficient using stream classification. This may help in developing more consistent roughness estimates and provide an approach for improving stream discharge estimates by using the manning's equation. The Hicks and Mason (1991) work is exemplary in terms of evaluating and displaying variations in N with changes in stream discharge. These variations can potentially be developed as a rate of change index for changes in stage by stream type. The influence of vegetation is shown to cause a marked adjustment in values by stream type. As would be expected, this relationship suggests the vegetation influence on roughness is diminished as channel gradient and bed material particle size increase. Stream types essentially integrate those variables affecting roughness, such as; gradient, shape and form resistance, particle size, and relative depth of bankfull discharge to the diameter of the larger particles in the channel. Rather than looking at discrete predictors, stream types integrate the many variables that influence resistance. Another recommended application to roughness estimation is to develop specific relations of roughness and associated velocity as recently developed for "mountain streams" by Jarrett (1984, 1990). In this method, equations were stratified for steeper slopes and cobble/boulder channel materials, using hydraulic radius and slope in the equations. Jarrett's results were valuable in that they produced values much different from most published equations. This work could be even more effective if the stream data were further stratified into stream types and size of stream. In this manner, much like the Manning's N values, equations could be developed using the integrating effects of stream types and thereby advance the state of the art of applications.

5.4. Hydraulic geometry relations

The original work of Leopold and Maddock (1953) made a significant contribution to the applied science in the development of hydraulic geometry relations. The variables of; depth, velocity, and cross-sectional area were quantitatively related to discharge as simple power functions for a given river cross-section. Their findings prompted numerous research efforts over the years. To refine average values of exponents, and to demonstrate the potential for applications of hydraulic geometry relations by stream types, this author assembled stream dimensions, slopes, and hydraulic data for six different stream types having the same discharge and channel materials. The objective was to demonstrate how the shape (width/depth ratio), profile (gradient), plan view (sinuosity), and meander geometry affect the hydraulic geometry relations. For example channel width increases faster than mean depth, with increasing discharge in high width/depth ratio channels. The opposite is true in low width/depth ratio channels. Streamflow values from baseflow of approximately 4 cfs up to bankfull values of 40 cfs were compared for each cross-section, and the corresponding widths, depths, velocities, and cross-sectional area for each stream type were computed. The A3, B3, C3, D3, E3 and F3 stream types selected for comparisons all had a cobble dominated bed-material size. The resultant hydraulic geometry relations for the selected array of stream types at the described flow ranges are shown in Fig. 11. Except for the E3 stream type for the plot of width/discharge, the slope of the plotted relations did not significantly change nearly as much as the intercept values.

6. Shear stress/velocity relations

Using the same data from the six stream types described previously, a "lumped" data base for all stream types from low to high flow was made for the corresponding shear stress *(T = 'yRS)* (Shields, 1936) vs. mean velocity, where; T = shear stress, *D.L. R o s g e n / Catena 22 (1 994) 169- 199 189* = density of water, *R* = hydraulic radius, and *S* = channel slope. As expected, a meaningful relation was not found. However, plotting shear stress and velocity stratification by stream type provided a trend that did shows promise (Fig. 12). While more data are needed to establish mathematical and statistical relationships, the comparisons arranged by stream type may have potential for future applications.

6.1. Critical shear stress estimates

Previous investigations of the magnitude of shear stress required to entrain various particle diameters from the stream-bed material have produced a wide range of values.

6.2. Sediment relations s

Stream types have been used to characterize sediment rating curves that reflect sediment supply in relation to stream discharge. For example, a sediment rating curve regression relation for an A2 stream type would have a characteristic low slope and intercept. The sediment rating curve for the C4 stream type, however, has a higher intercept and steeper slope. The author has used this procedure for both suspended and bedload rating curves. These relationships were initially plotted as a function of channel stability ratings as developed by Pfankuch (1975). Applications for cumulative effects analysis for non-point sediment sources utilized this approach (USEPA, 1980). Subsequent comparisons of data with stream type delineations indicated similar relations. The ratio of bedload to total sediment load can also be stratified by stream type where measured data is available. Ranges of less than 5% bedload to total sediment load for C3 stream types have been reported, but values greater than 75% bedload to total load for G4 stream types have also been measured (Williams and Rosgen, 1989). The "high ratio" bedload streams are the A3, A4, A5, D3, D4, D5, F4, F5, G3, G4, and G5 stream types.

6.3. Management interpretations

The ability to predict a river's behavior from its appearance and to extrapolate information from similar stream types helps in applying the interpretive information in Table 3. These interpretations evaluate various stream types in terms of; sensitivity to disturbance, recovery potential, sediment supply, vegetation controlling influence, and streambank erosion potential. Application of these interpretations can be used for; potential impact assessment, risk analysis, and management direction by stream type. For example, livestock grazing effects were related to stream stability and sensitivity using stream types (Meyers and Swanson, 1992). They summarized their study results on streams in northern Nevada that "... range managers should consider the stream type when setting local standards, writing management objectives, or determining riparian grazing management strategies." This interpretive information by stream type can also apply to establishment of watershed and streamside management guidelines dealing with; silvicultural standards, surface disturbance activities, surface disturbance activities, gravel and surface mining activities, riparian management guidelines, debris management, floodplain management, cumulative effects analysis, flow regulation from reservoirs/ diversions, etc. An example of the implementation of these guidelines by stream type are shown in the Land and Resource Management Plan (USDA, 1984). Applications for riparian areas (USDA, 1992), have utilized the stream classification system into their recently developed "Integrated Riparian Evaluation Guide" - Intermountain Region. The classification system was used to help stratify and classify riparian areas based on natural characteristics and existing conditions. It is also used to evaluate the potential risks and sensitivities of riparian areas.

6.4. Restoration

The morphologic variables that interact to form the dimensions, profile and patterns of modern rivers are often the same variables that have been adversely impacted by development and land use activities. To restore the "disturbed" river, the natural stable tendencies must be understood to predict the most probable form. Those who undertake to restore the "disturbed" river must have knowledge of fluvial process, morphology, channel and meander geometry, and the natural tendencies of adjustment toward stability in order to predict the most effective design for long-term stability and function. If one works against these tendencies, restoration is generally not successful. Restoration applications using stream classification and the previously discussed principles are documented in the "Blanco River" case study (National Research Council, 1992).

7. **Summary**

Rivers are complex natural systems. A necessary and critical task towards the understanding of these complex systems is to continue the river systems research. In the interim, water resource managers must often make decisions and timely predictions without the luxury of a complex and thorough data base. Therefore, a goal for researchers and managers is to properly integrate what has been learned about rivers into a management decision process that can effectively utilize such knowledge. There is often more data collected and available on rivers than is ever applied. Part of the problem is the large number of "pieces" that this data comprises and the difficulty of putting these pieces into meaningful form. , The objective of this stream classification system presented here is to assist in bringing together these "pieces" and the many disciplines working with rivers *D.L. Ro s g en / Catena 22 (1994) 169-199 195* under a common format - a central theme for comparison, a basis for extrapolation, prediction, and communication. The stream classification system can assist in organizing the observations of river data and of molding the many pieces together into a logical, useable, and reproducible system. With the recent emphasis on "natural" river restoration or "naturalization" throughout Europe and North America, understanding the potential versus the existing stream type is always a challenge. The dimension of rivers related to the flow, and the patterns, which in turn are related to the dimensions, have to be further stratified by discrete stream types. In this way, the arrangement of the variables that make up the plan, profile and section views of stable stream types that are integrated within their valley's can be emulated. This also involves re-creation of the corresponding appropriate bed morphology associated with individual stream types with the observed sequence of step/pool and/or riffle pool bed features as a function of the bankfull width. The use of meander width ratios by stream type helps to establish the minimum, average and ranges of lateral containment of rivers. This often helps the design engineer/hydrologist determine appropriate widths that need to be accommodated when natural, stable rivers are re-constructed within their valleys. River and floodplain elevations, which need to be constructed, can be often determined by the used of the entrenchment ratio, which depicts the vertical containment of rivers in the landform. Using these integrative, morphological relations by stream type, can avoid the problematic "works" done on streams which create changes in the dimensions, pattern and profile of rivers which are not compatible with the tendencies of the natural stable form. A classification system is particularly needed to stratify river reaches into groups that may be logically compared. Such stratification reduces scatter that might appear to come from random variation, whereas the scatter often results from attempting to compare items generically different. For example, data developed from empirical relations associated with process oriented research in natural channels such as tractive force relations, resistance and sediment transport equations, etc., can be stratified by stream type. This can help reduce the scatter when applied to stream types different than those from which the relations were developed. Utilizing quantitative channel morphological indices for a classification procedure insures for consistency in defining stream types among observers for a great aversity of potential applications. The classification presented here may be Ae first approximation of a system that undoubtedly will be refined over the years with continued experience and knowledge. This stream classification system hopefully can be a vehicle to provide better communication among those studying river systems and promote a better understanding of river processes, helping put principles into practice. *196 D.L. Rosgen / Catena 22 (1994) 169-199*

References

Andrews, E.D., 1980. Effective and bankfull discharges of streams in the Yampa River Basin, Colorado and Wyoming. J. Hydrol., 46: 311-330.

Andrews, E.D., 1983. Entrainment of gravel from natural sorted riverbed material. Geol. Soc. Am. Bull., 94:1225-1231. Andrews, E.D., 1984. Bed-material entrainment and hydraulic geometry of gravel-bed rivers in Colorado. Geol. Soc. Am. Bull., 95: 371-378.

Baker, V.R., 1974. Paleohydraulic interpretation of Quaternary alluvium near Golden, Colorado. Quat. Res., 4: 94-112. Baker, V.R_ and Ritter, D.F., 1975. Competence of rivers to transport coarse bedload material. Geol. Soc. Am. Bull., 86: 975-978.

Barnes, H.H., 1967. Roughness characteristics of natural channels. U.S. Geol. Survey, Water Supply Paper 1849, 213 pp. Bradley, W.C. and Mears, A.I., 1980. Calculations of flows needed to transport coarse fraction of Boulder Creek alluvium at Boulder, Colorado. Geol. Soc. Am. Bull., Part II, 91: 1057-1090.

Brice, J.C. and Blodgett, J.C., 1978. Counter Measures for Hydraulic Problems at Bridges. Vol. 1, Analysis and Assessment. Report No. FHWA-RD-78-162, Fed. Highway Admin., Washington, DC, 169 pp.

Church, M., 1978. Paleohydrological reconstructions from a Holocene valley. In: A.D. Miall (Editor.), Fluvial Sedimentology. Can. Soc. Pet. Geol. Mem., 5: 743-772.

Church, M. and Rood, K., 1983. Catalogue of alluvial river channel regime data. Natural Sciences and Engineering Research Council of Canada. Dept. Geology Univ. British Columbia, Vancouver, B.C., Edition 1.0. Culbertson, D.M., Young, L.E. and Brice, J.C., 1967. Scour and fill in alluvial channels. U.S. Geological Survey, Open File Report, 58 pp.

Davis, W.M., 1899. The geographical cycle. Geogr. J., 14: 481-504.

Dunne, T. and Leopold, L.B., 1978. Water in Environmental Planning. Freeman, San Francisco, CA, 818 pp. Fahnestock, R.K., 1963. Morphology and hydrology of a glacial stream - White River. Mount Rainier, Washington. U.S. Geol. Survey Prof. Paper 422-A, 70 pp.

Galay, V.J., Kellerhals, R. and Bray, D.I., 1973. Diversity of river types in Canada. In: Fluvial Process and Sedimentation. Proceedings of Hydrology Symposium. National Research Council of Canada, pp. 217- 250.

Grant, G.E., Swanson, F.J. and Wolman, M.G., 1990. Pattern and origin of stepped-bed morphology in high-gradient streams, Western Cascades, Oregon. Geol. Soc. Am. Bull., 102: 340-352.

Hicks, D.M. and P.D. Mason, 1991. Roughness characteristics of New Zealand rivers. New Zealand Dept. of Scientific and Industrial Research, Marine and Freshwater, Natural Resources Survey, Wellington, 329 pp.

Inglis, C.C., 1942. Rapid westerly movement of the Kosi River. Central Board of Irrigation, India (Technical), Poona, Bombay, pp. 7 and 8.

Jarrett, R.D., 1984. Hydraulics of high-gradient streams. J. Hydraul. Eng., 110(11): 1519-1539. *D.L. Rosgen / Catena 22 (1994) 169-199* 197

Jarrett, R.D., 1990. Hydrologic and hydraulic research in mountain rivers. Water Res. Bull. WARBAQ, 26(3): 419-429.

Kellerhals, R., Neill, C.R. and Bray, D.I., 1972. Hydraulic and geomorphic characteristics of rivers in Alberta. Research Council of Alberta, River Engineering and Surface Hydrology Report 72-1, p. 52.

Kellerhals, R., Church, M. and Bray, D.I., 1976. Classification and analysis of river processes. J. Hydraul. Div., ASCE, 102(HY7): 813-829.

Khan, H.R., 1971. Laboratory studies of alluvial river channel patterns. Ph.D. Dissertation, Dept. of Civil Engineering Department, Colorado State University, Fort Collins, CO.

Lane, E.W., 1957. A study of the shape of channels formed by natural streams flowing in erodible material. Missouri River Division Sediment Series No. 9, U.S. Army Engineer Division, Missouri River, Corps of Engineers, Omaha, NE. Langbein, W.B. and Leopold, L.B., 1966. River meanders - Theory of minimum variance. U.S. Geol. Survey Prof. Paper No. 422-H, 15 pp.

Leopold, L.B. and Maddock, T., 1953. The hydraulic geometry of stream channels and some physiographic implications. U.S. Geol. Survey Prof. Paper No. 252. Supt. of Docs, U.S. Government Printing Office, Washington, DC, 57 pp.

Leopold, L.B. and Wolman, M.G., 1957. River channel patterns: braided, meandering, and straight. U.S. Geological Survey Prof. Paper 282-B.

Leopold, L.B., Wolman, M.G. and Miller, J.P., 1964. Fluvial processes in geomorphology. Freeman, San Francisco, CA, 522 pp. Matthes, G., 1956. River engineering. In: P.O. Abbott (Editor), American Civil Engineering Practice. Wiley, New York, Vol. 11, pp. 15-56.

Melton, F.A., 1936. An empirical classification of flood-plain streams. Geogr. Rev., 26. Mollard, J.D., 1973. Air photo interpretation of fluvial features. In: Fluvial Processes and Sedimentation. Research Council of Canada, pp. 341-380.

Meyers, T.J. and Swanson, S., 1992. Variation of stream stability with stream type and livestock bank damage in Northern Nevada. Water Resources Bull. AWRA, 28(4): 743-754.

Nanson, G.C. and Croke, J.C., 1992. A genetic classification of floodplains. In: G.R. Brakenridge and J. Hagedorn (Editors), Floodplain Evolution. Geomorphology, 4: 459-486.

National Research Council, 1992. Restoration of aquatic ecosystems. Restoration case studies. National Academy Press, Washington, DC, pp. 470-477.

Osborn, J.F. and Stypula, J.M., 1987. New models of hydrological and stream channel relationships. In: Erosion and sedimentation in the Pacific Rim. IAHS Publ. No. 165, Oregon State Univ., Corvallis, OR.

Pfankuch, D.J., 1975. Stream reach inventory and channel stability evaluation. USDA Forest Service, R1- 75-902. Government Printing Office #696-260/200, Washington DC, 26 pp.

Pickup, G., 1984. Geomorphology of tropical rivers. 1. Landforms, hydrology and sedimentation in the fly and lower Purari, Papua New Guinea. In: A.P. Schick (Editor), Channel Processes, Water, Sediment, Catchment Controls. Catena Suppl., 5: 1-17.

Platts, W.S., 1980. A plea for fishery habitat classification. Fisheries, 5(1): 1-6. Ritter, J.R., 1967. Bed-material movement. Middle Fork Eel River, California. U.S. Geol. Survey Prof. Paper 575-C, pp. C219-C221.

Rosgen, D.L., 1985. A stream classification system. In: Riparian Ecosystems and Their Management. First North American Riparian Conference. Rocky Mountain Forest and Range Experiment Station, RM-120, pp. 91-95.

Rosgen, D.L. and Fittante, B.L., 1986. Fish habitat structures - A selection guide using stream classification. In: J.G. Miller, J.A. Arway and R.F. Carline (Editors), 5th Trout Stream Habitat Improvement Workshop, Lock Haven Univ., Lock Haven, PA. Penn. Fish Comm. Publics, Hamsburg, PA, pp. 163- 179.

Schumm, S.A., 1963. A tentative classification of alluvial river channels. U.S. Geological Survey Circular 477. Washington, DC. Schumm, S.A., 1977. The Fluvial System. Wiley, New York, 338 pp.

Selby, M.J., 1985. Earth's Changing Surface: an Introduction to Geomorphology. Oxford University Press, Oxford. 198 *D.L. Rosgen / Catena 22 (1994) 169-199*

Shields, A., 1936. Application of similarity principles and turbulence research to bed-load movement. In: W.P. Ott and J.C. Uchelen (translators), Mitt. Preuss. Verschsanst., Berlin. Wasserbau Schiffbau. California Institute of Technology, Pasadena, CA, Report No. 167, 43 pp.

Simons, D.B. and Senturk, F., 1977. Sediment Transport Technology. Water Resource Publications, Fort Collins, CO, 807 pp.

Simons, D.B. and Li, R.M., 1982. Engineering Analysis of Fluvial Systems. SLA, P.O. Box 1816, Fort Collins, CO. Smith, D.G. and Smith, N.D., 1980. Sedimentation in anastomosing river systems: examples from alluvial valleys near Banff, Alberta. J. Sediment. Petrol., 50: 157-164.

Thornbury, W.D., 1969. Principles of Geomorphology. 2nd Edition. Wiley, New York. USDA (U.S. Department of Agriculture), Forest Service, 1984.

Land and Resource Management Plan - Arapaho and Roosevelt National Forests and Pawnee Grassland. Fort Collins, CO. USDA (U.S. Department of Agriculture), Forest Service, 1989. Fisheries Habitat Surveys Handbook. Region 4 - FSH 2609.23.

Ogden, UT. USDA (U.S. Department of Agriculture), Forest Service, 1992. Integrated Riparian Evaluation Guide. Intermountain Region - Ogden, UT. USEPA (U.S. Environmental Protection Agency), 1980. An approach to water resources evaluation of nonpoint silvicultural sources. Chap. VI, Total Potential Sediment, D. Rosgen. EPA-600/8-80-012, Athens, GA, pp. 39-41.

Williams, G.W., 1986. River meanders and channel size. J. Hydrol., 88: 147-164.

Williams, G.P. and Rosgen, D.L., 1989. Measured total sediment loads (suspended loads and bedloads) for 93 United States streams. Open File Report 89-67. U.S. Geol. Survey, Denver, CO, 128 pp.

Wolman, M.G., 1954. A method of sampling coarse river-bed material. Trans. Am. Geophys. Union, 35: 951-956. *D.L. Rosgen / Catena 22 (1994) 169-199* 199

UNIT- 8: EROSION: THRESHOLD OF EROSION, PROCESSES OF EROSION, RIVER BANK EROSION

The Concept of Stream Power

Sediment entrainment, transport and deposition all involve the interaction of forces. **Work** is carried out when a force moves an object, the amount of work being defined by the size of the force and the distance over which the object moves. Work is involved in moving water through the channel and in eroding and transporting sediment. **Energy** is the capacity or ability to do work, and the same units, joules (J), are used for both.1 **Power** defines the rate at which work is done and is measured in watts (W), or joules per second. The concept of power can be illustrated by considering the transport of a piece of gravel between two points. This could be accomplished in a short period of time by a large force (high power), or over a longer period by a smaller force (low power). Although the same amount of work is involved in each case, it is carried out at different rates. **Stream power** is measured in watts per unit length of stream channel, usually W m−1. Stream power determines the **capacity** of a given flow to transport sediment. This is the maximum volume of sediment that can be transported past a given point per unit time. The available stream power is related to the water surface slope (*S*) and discharge (*Q*) of the channel. It is also affected by the gravitational constant, *g*, and the mass density of the fluid (1,000 kg m2 for water), which is represented by the Greek letter rho (ρ). These are combined in the equation below, where stream power is represented by the Greek capital letter omega (V): V = rgQS Stream power is often defined in terms of the **specific stream power**, or stream power per unit area of the bed (per m2). Specific stream power (lower-case omega, ω) is calculated by dividing the stream power per metre length of channel by the width of the channel. v = V/W Where *W* is channel width. This is useful for making comparisons between rivers, or different reaches of the same river, because it reduces the scale effects of large and small channels. For British rivers, the specific stream power ranges from less than 10 W m2 for lowland channels in parts of the south-east, to 1,000 W m2 for rivers in the north and west, which drain steep upland areas with high rainfall (Ferguson, 1981). Specific stream power can be related to bed shear stress (τ0) and (cross-sectional) average flow velocity (v): v = r_0v This means that the power per unit area of the bed is equal to the product of the average bed shear stress and the average flow velocity. Flow **competence** is the ability of a given flow to entrain sediment of a certain size and increases with bed shear stress.

PROCESSES OF EROSION IN BEDROCK CHANNELS

The morphology of bedrock channels is mainly influenced by processes of erosion because the supply of sediment is often limited. Three types of erosion are significant: block quarrying, abrasion and corrosion. **Block quarrying** is the dominant process (Hancock *et al*., 1998) and involves the removal of blocks of rock from the bed of the channel by drag and lift forces. The size of the quarried blocks can be considerable. Tinkler (1993) reports blocks of sandstone 1.2 m × 1.45 m × 0.11 m and 1.0 m × 0.5 m × 0.05 m being removed from the bed of Twenty Mile Creek, Niagara Peninsula, Ontario, during normal winter flows, when the flow depth was less than 0.4 m. Before blocks can be entrained by the flow, a certain amount of ‘preparation’ is required to loosen them. Subaerial weathering and other weakening processes play an important role in this. Weakening processes described by Hancock *et al*. (1998) include the bashing of exposed slabs by particles carried in the load and a previously undocumented process termed ‘wedging’, which leads to the enlargement of cracks in the bedrock substrate. This is thought to occur when small bedload particles are able to enter cracks that are momentarily widened by fluid forces. The particles then become very firmly lodged and prevent the crack from narrowing again. As time progresses, further widening of the crack can be sustained as larger particles fall into it, and may ultimately lead to block detachment. Under conditions of very high flow velocity, sudden changes in pressure can generate shock waves that weaken the bed by the process of **cavitation**. This effect is caused by the sudden collapse of vapour pockets within the flow (Knighton, 1998). **Abrasion** is the process by which the channel boundary is scratched, ground and polished by particles carried in the flow. Erosion is often concentrated where there are weaknesses and irregularities in the rock bed, which allow abrasion to take place at an accelerated rate. This can lead to the development of **potholes**, deep circular scour features that often form in bedrock reaches. Once a pothole starts to develop, the flow is affected, focusing further erosion. Any coarse material that collects in the pothole is swirled around by the flow, deepening and enlarging it, and literally drills down into the channel bed. Over time potholes may coalesce, leading to a lowering of the bed elevation. Plate 7.1 shows how potholes have contributed to bed lowering near the site of a waterfall. Scouring by finer material carried by the flow, such as sand, leads to the development of **sculpted forms**. These include flutes and ripple-like features, which reflect structures within the flow. These are commonly observed on the crests of large boulders and other protrusions into the flow, where flow separation takes place and fine sediment is decoupled from the flow (Hancock *et al*., 1998). The rock boundary may also be polished by fine material carried in suspension. Bedrock channels formed in soluble rock are also susceptible to erosion by **corrosion**, especially where the presence of joints and bedding planes allows solutional enlargement. Solutional features such as scallops may also be seen. These spoon-shaped hollows often cover the walls of cave streamways. Their length is related to the formative flow velocity, ranging from a few millimetres (relatively fast flow) to several metres (relatively slow). Although the actual processes of erosion operate at a small scale, their effects can be seen over scales ranging from millimetres to kilometres. There are several controls on rates of erosion, which influence the processes described above. These include micro-scale (millimetres to centimetres) variations in the rock structure, the larger scale effects of bedding, joints and fractures, and basin-scale influences such as regional geology and base level history (Wohl, 1998).

BANK EROSION IN ALLUVIAL CHANNELS

Processes of bank erosion are important in the development and evolution of different channel forms, while the migration of river channels across their floodplains involves a combination of bank erosion and deposition. Bank erosion can also create management problems when bridges, buildings and roads are undermined or destroyed. Large volumes of sediment can be generated, leading to problems of aggradation further downstream. Land disputes may also arise where boundaries lie along actively migrating river channels. Rather than being a process in itself, bank erosion is brought about by a number of different processes which can be considered in three groups: 1 *Pre-weakening processes* such as repeated cycles of wetting and drying, which 'prepare' the bank for erosion. 2 *Fluvial processes*, where individual particles and aggregates are removed by direct entrainment. 3 *Processes of mass failure*, which include the collapse, slumping or sliding of bank material into the channel. Bank material that has been detached remains at the base of the bank until it is broken down *in-situ* or entrained and transported downstream. A balance exists between the rate of sediment accumulation and its rate of removal, which acts as an important control on rates of bank erosion (Carson and Kirkby, 1972). If material accumulates at the base of the bank at a faster rate than it is removed then, to a certain extent, the bank is protected from further erosion. When the opposite situation applies, with bank material being removed faster than it accumulates, bank erosion will continue, sometimes at an increased rate. A third possibility is that rates of supply are the same as rates of removal. The relative rates of accumulation and removal are dependent on the available stream power and the controls on bank erosion discussed below. **Bank materials and weakening processes** The moisture content of the bank is significant, particularly for cohesive bank materials whose strength varies with the level of saturation. A certain amount of water is held in the pores, against the force of gravity, by matric suction forces. These result from surface tension effects, and a negative pore water pressure (less than atmospheric) develops when the soil is not completely saturated. As the soil dries, the strength of the matric suction forces increases as all but the smallest pores are emptied. These forces can be considerable and several authors have observed an increase in the resistance of the bank material to erosion at high matric suctions. However, it has also been suggested that desiccation can lead to higher rates of bank retreat, because the shrinking of clay particles causes cracking and shedding of loose material at the bank surface. The process of **slaking** occurs when banks are rapidly immersed by floodwaters and air becomes trapped and compressed within the pores. The resultant pressure causes material to become dislodged (Thorne and Osman, 1988). At high flows, banks may become saturated with water from the channel. Saturation also occurs when there is a rise in the water table or during prolonged rainfall. Under these conditions a positive pore water pressure exists between the grains. This weakens the cohesive forces, acting as a lubricant and reducing inter-granular friction. During cold conditions, the growth of lenses, wedges, and crystals of ice can significantly reduce resistance to erosion, especially where freeze–thaw cycles occur. In temperate regions, the growth of ice needles occurs during moderately sub-zero temperatures. These are elongated crystals of ice that start to grow as the temperature of the air in contact with the bank decreases, growing in the direction of cooling (into the bank). The crystals often lift and incorporate material which then moves downslope or remains as a 'sediment drape' when the ice melts (Lawler, 1988). In colder regions, where rivers freeze over in winter, cantilevers of ice can cause significant damage (Church and Miles, 1982). Where permafrost exists, thermoerosion niches are cut into frozen banks by the relatively warm water in the channel. While not a process in itself, the presence of vegetation influences the resistance to bank erosion in various ways. Root networks are particularly important and vegetated banks tend to have a more open structure and be better drained. Vegetation also acts to bind the soil together and increase the shear strength of the bank material. Unlike soil, roots have a very high tensile strength, which means that they are able to resist tension (stretching forces). **Bank erosion by fluvial processes** For any given situation the relative importance of direct entrainment and mass failure is mainly determined by the composition of the bank, although other factors can also be important. Banks composed of sand and coarser particles are non-cohesive and this material is usually detached grain by grain. Although cohesive forces do not exist between the particles, movement is resisted by inter-particle friction and the packing structures holding the grain in place. However, the selective entrainment of finer sands and gravels often leads to a weakening of the overall structure, which may lead to collapse. In the case of cohesive banks, it tends to be aggregates and crumbs that are detached rather than individual particles. The weakening processes described above are of great importance in assisting fluid forces to detach and entrain aggregates. Once entrained into the main flow, aggregates tend to disintegrate fairly rapidly. **Bank failure mechanisms** Bank failure occurs when bank material becomes unstable and falls or slides to the base of the bank. There are several types of failure, and different failure mechanisms are observed for cohesive and non-cohesive bank materials. Also important are bank height, bank angle, moisture content and the effects of vegetation.. The stability of banks is determined by the balance between the shear stress exerted by the down-slope component of gravity (driving force) and shear strength of the bank material (resisting force). In cohesive banks, failure occurs across a **failure plane**, the surface within the bank across which shear stress exceeds shear strength. Failure planes can be almost planar (flat) or curved. Typically the failure surface is almost planar and vertical, parallel to the bank surface. Where bank angles are less steep, the failure plane is usually curved and located deep within the bank. Cohesive banks are often most susceptible to failure after a flood wave has passed, when the saturated banks are no longer supported by the pressure of flow in the channel. Non-cohesive banks tend to fail along shallow slip surfaces. Mixed banks are common, typically with fine cohesive sediment overlying non-cohesive material. Undercutting of the non-cohesive material by fluvial processes leads to instability of the overlying material.

References:

Hancock, G.S., Anderson, R.S. and Whipple, K.X., 1998. Beyond power: bedrock river incision process and form. In: K.J. Tinkler and E.E. Wohl (eds), *Rivers over Rock: Fluvial Processes in Bedrock Channels*. Geophysical Monograph Series. American Geophysical Union, Washington DC: 35–60.

Tinkler, K.J., 1993. Fluvially sculpted bedforms in Twenty Mile Creek, Niagara Peninsula, Ontario. *Canadian Journal of Earth Science*, 30: 945–53.

Wohl, E.E., 1998. Bedrock channel morphology in relation to erosional processes. In: K.J. Tinkler and E.E. Wohl (eds), *Rivers over Rock: Fluvial Processes in Bedrock Channels*. Geophysical Monograph Series. American Geophysical Union, Washington DC, pp. 133–51.

Carson, M.A. and Kirkby, M.J., 1972. *Hillslope Form and Process*. Cambridge University Press, Cambridge.

Thorne, C.R. and Osman, A.M., 1988. Riverbank stability analysis II: applications. *Journal of the Hydraulic Division, Proceedings of the American Society of Civil Engineers*, 114: 151–72.

Lawler, D.M., 1988. Environmental limits of needle ice: a global survey. *Arctic and Alpine Research*, 20: 137–59.

Church, M. and Miles, M.J., 1982. Processes and mechanisms of bank erosion. In: R.D. Hey, J.C. Bathurst and C.R. Thorne (eds), *Gravel Bed Rivers*. John Wiley & Sons, Chichester, pp. 221–71.

UNIT-9: TRANSPORTATION: PROCESSES OF ENTRAINMENT, BEDLOAD TRANSPORT DYNAMICS; CHANNEL COMPETENCE

The process of particle entrainment

Whether or not a given particle is set in motion depends on the balance between the forces driving and resisting its movement. The resisting force is the immersed weight of the particle (in this simple example the effects of neighbouring grains will be ignored). The driving force is provided by the combined effect of two **fluid forces** exerted on the particle by the flow: a drag force and a lift force. The fluid **drag force** acts in the same direction as the flow and can be thought of as the 'force of the flow' that is felt when you wade out into the current of a stream. It comes about because the pressure exerted on an object by the flow is greater on its upstream side than on its more sheltered downstream side. The second fluid force, the **lift force**, acts vertically upwards and is caused by a pressure difference above and below the particle. Water flowing over the particle has to move faster. According to the Bernouilli principle, an increase in velocity results in a decrease in pressure above the particle, while the pressure below it stays the same. This difference in pressure generates lift. In theory, if this force exceeds the gravitational force, the particle will be lifted from the bed. In practice, the presence of other particles complicates matters considerably. Sediment transported as bedload is generally gravelsize and larger, although coarse sands may also form part, or all, of the bedload component. Finer bedload, which is too coarse to be transported in suspension, is moved along the bed in a series of short jumps by **saltation**. Saltating grains are lifted from the bed at a relatively steep angle by the combined forces of lift and drag. As a grain moves upwards into the flow, the lift force decreases and it starts to fall back towards the bed. The falling grain is carried downstream by the drag force, following a shallow trajectory towards the bed. Larger particles, which cannot be lifted, are rolled or dragged along the bed. This movement is usually sporadic because of variations in bed shear stress. In addition, particles tend to become lodged behind other particles or obstacles on the bed. The weight of smaller particles carried in suspension is supported by turbulence. Descending saltating grains may also be temporality lifted upwards by turbulent movements. This is called incipient suspension. **Size-selective theories of sediment transport** A considerable amount of research has focused on deriving critical flow or **entrainment thresholds** from easily measured flow parameters. There are various practical reasons why we might want to know the flow conditions that will move particles of a certain size. These include the planning of reservoir releases to flush out fine sediment from fish spawning grounds (without removing the gravel), or determining when structures such as bridge piers are at risk of being undermined by erosion. The threshold conditions for the entrainment of particles of a given size can be defined according to a critical mean flow velocity (i.e. cross-sectional average) or a critical bed shear stress. Using the mean flow velocity is an indirect method, since it is actually the hydraulic conditions near the bed of the channel that are significant. However, both relationships show similar basic trends. An explanation will first be given in terms of a critical mean flow velocity since this relationship is conceptually easier to understand. The critical mean flow velocity curves were derived from a large amount of experimental data accumulated by Filip Hjulstrøm in the 1930s. They show the entrainment and fall (or settling) velocities for particles of different sizes, from fine clay to coarse gravel and small boulders. Note that a logarithmic scale is used on both axes to cover the wide range of particle sizes and the corresponding range of flow velocities. The upper curve on the graph shows the **entrainment velocity** required to set different particle sizes in motion. Sand grains, with a diameter of between 0.2 mm and 0.7 mm, are the easiest to entrain. In the case of larger particles, which have a greater immersed weight, the entrainment velocity increases with particle size as might be expected. However, the relationship is rather different for particles smaller than 0.2 mm, since the entrainment velocity actually *increases* as the particle size decreases from fine sand to silt and clay. Reasons for this include the fact that these small particles tend to be partly or wholly enclosed within the laminar sublayer during most flows. Drag forces are lower within this layer, and particles are not exposed to turbulent lift forces. In addition, the cohesive forces between clay particles further increases the force required to set them in motion. An alternative approach, which is more relevant to modern sediment transport theory, was devised by the American engineer Albert Shields in 1936. This defines the critical bed shear stress necessary to set particles of a given size in motion. The critical bed shear stress is actually defined in a dimensionless form. The dimensionless critical bed shear stress is often referred to as the **Shields parameter**. It appears in a number of sediment transport equations and is represented by the Greek letter theta (θc – the subscript is short for 'critical'). Critical bed shear stress increases with particle size but also depends on bed roughness. Shields related the dimensionless bed shear stress to the boundary Reynolds number (Re). The boundary Reynolds number is proportional to the ratio between grain size and laminar sublayer thickness. Where Re is less than about 5, the grains are small enough to be fully submerged within the laminar sublayer. As these sheltered particles get smaller, the shear stress needed to entrain them increases. For hydraulically rough surfaces, the critical bed shear stress is independent of the boundary Reynolds number and the critical bed shear stress reaches a constant value of 0.06 (Richards, 1982). The lowest critical bed shear stress is associated with sand grains in the size range 0.2 mm to 0.7 mm (Knighton, 1998). It is important to note that the Hjulstrom and Shields experiments were carried out using well sorted bed sediment of a single size. This is not representative of the conditions on the bed of many channels, where there is a mixture of grain sizes. The arrangement of grains on the bed and the mixture of grain sizes is very significant, affecting both the entrainment of individual grains and overall transport rates. **Sediment transport in mixed beds** The mobility of individual particles is greatly affected by the size and arrangement of the particles surrounding them. In most natural channels the mixture of sediment sizes, and an irregular bed surface, makes the situation rather more complicated. This can be defined in terms of a **friction angle**, which is greatest where small particles overlie larger ones, meaning that a greater force is required to pivot smaller particles away from the bed. Larger grains can also shelter smaller grains from flows that would otherwise be competent to entrain them. The degree of **sorting** reflects the range of particle sizes in a particular sample of bed material. Well-sorted sediments have a narrow range of particle sizes, whereas poorly sorted material shows a much wider range. In gravel-bed rivers, particles may also be wedged together in various types of packing arrangements which act to resist bed shear stresses and again make it much harder for individual grains to be entrained. As a result, fine sediment is removed from the bed leaving a layer of coarse sediment, usually about one particle diameter in thickness. This armour layer protects the finer material beneath from subsequent high flows. Once a bed is armoured, a much higher critical threshold is required to break it up. Bathurst (1987a) defined **'twophase' flow** for armoured channels. During phase 1 flow an armour layer is present and rates of bedload transport are low (although finer sediment can still be supplied from further upstream). Once the armour layer breaks up phase 2 transport takes place, with a dramatic increase in transport rates as the finer sediment becomes available. This can lead to complex variations in bedload transport through time. For example, where two high magnitude flow events occur in close succession, the initial rate of transport is often much higher for the second event than for the first, which breaks up the armour layer. Research has shown that ephemeral channels do not tend to develop an armour layer because, in the absence of low flows, there is no mechanism for removing fine sediment to create the armour layer. This could mean that rates of bedload transport are greater for ephemeral channels than for channels in humid settings (Nanson *et al*., 2002). **The theory of equal mobility transport** On a level bed with a uniform sediment size, all the particles might be expected to begin moving under approximately the same flow conditions (Reid *et al*., 1997).

However, on mixed beds, the relative size of a given sediment particle determines its degree of exposure to the flow. As a result, larger particles shelter smaller particles, which then require a higher shear stress for entrainment than would otherwise be the case. In contrast, coarser grains are more easily entrained when surrounded by fine grains. This is because they are relatively more exposed to the forces of entrainment (Andrews, 1983). Particles of an intermediate size are relatively unaffected by the sheltering/hiding effects. Empirically, this 'reference size' has been shown to approximate the median size (*D*50) (Bathurst, 1987b). On the basis of field data, Parker *et al*. (1982) introduced a theory of **equal mobility** for channel beds composed of a mixture of sediment sizes. This states that the threshold condition for each size fraction is not dependent on the grain size. In other words, the movement of particles of different sizes can be initiated under similar critical flow conditions. The theory of equal mobility transport therefore challenges the size-selective transport theory of Shields (1936). However, any deviation away from an equal mobility condition represents some degree of size-selective transport. Under conditions of equal mobility, the bedload transport rate could be calculated from a single representative grain diameter such as the median size, *D*50 (Parker *et al*., 1982). Equal mobility transport is the subject of some debate, however. Field investigations into the occurrence of equal-mobility transport have mainly been carried out in gravel-bed channels, where the largest grains are cobble size or smaller (for example Andrews, 1983; Ashworth and Ferguson, 1989). Measurements made in rivers during steady uniform flows have shown that the transport of mixed sediment is only weakly size selective at low shear stresses. At higher shear stresses, sediment transport approaches equal mobility (Parker *et al*., 1982; Andrews, 1983; Marion and Weirach, 2003). However, observations made over a wider range of flows (e.g. Ashworth and Ferguson, 1989; Wilcock, 1992) have emphasised the size-selective nature of gravel transport. Only during the highest flows does sediment transport approach equal mobility. For example, Wilcock (1992) observed a progressive shift away from unequal to equal mobility transport with increasing shear stress, although equal mobility was not observed until the shear stress was over twice the critical stress required to initiate motion. One of the biggest problems associated with these investigations is obtaining sufficient field data to include a representative range of flow conditions (Reid *et al*., 1997).

BEDLOAD TRANSPORT

Bedload transport does not necessarily take place all the time, and rates may approach zero during low flows. Even when transport is occurring, it is likely that only part of the bed will be mobile at any one time. Part of the reason for this is the uneven distribution of bed shear stresses, which is directly controlled by variations in turbulent fluctuations. Large differences are observed across small areas of the bed and over short periods of time. Sweep fluid motions, inrushes of high momentum fluid from the outer zone of the boundary layer, are particularly effective at entraining bedload particles (Robert, 2003). The ejection of low momentum fluid away from the bed also allows finer sediment to be lifted up away from the bed and into the turbulent profile, maintaining it in suspension. The availability of bed sediment has an important influence on overall rates of bedload transport in a given reach of channel, and many bedload-dominated channels are transport limited. This means that transport rates might be lower than expected at a particular flow because of a lack of available sediment. This 'lack' does not necessarily refer to the total volume of bed sediment in a reach, more relevant is the availability of sediment of a certain size or calibre. Thus a flow that is competent to transport only fine gravels will not be able to entrain the larger material in a boulder-bed stream, no matter how abundant this is. The supply of bedload can be especially limited in bedrock channels and the flow capacity often exceeds that required to transport the available load. **Bedforms** Bedforms in sand-bed channels In sand-bed channels the sand grains can be transported at both high and low flows because of their low entrainment threshold. As a result the bed is easily shaped by flows to form periodic features known as bedforms. These have been intensively studied, both in the laboratory and in natural channels, and a recognisable sequence of bedforms develops in response to changing flow conditions. For the purpose of explanation, the starting point is assumed to be a plane bed, something that is rare in natural channels because even the smallest flows start to shape the bed. When water starts to flow over a flat bed the sand grains start to move, individually at first and then in patches, until periodic **ripples** develop, with crests perpendicular to the direction of flow. Field and laboratory research suggests that the wavelength, or spacing, of ripples is mainly dependent on particle size and is typically between 150 mm and 450 mm. As the flow intensity increases, ripples start to give way to **dunes**, larger features with rounded crests. Dunes are common in alluvial channels and are continuous along the bed for hundreds of kilometres in large rivers like the Mississippi and Niger. Dunes vary greatly in size, being scaled with the depth of flow and ranging from a few centimetres to a few metres in height. Dune wavelengths also vary, from tens of centimetres to more than a hundred metres in the largest rivers. Ripples and dunes migrate downstream over time, as the flow moves sand grains up the more gentle upstream slope towards the crest, from where sediment falls down the steeper downstream slope. A critical mechanism in this process is the deposition of coarse grains at the crest, where flow separation occurs. At higher flows, dunes become unstable and are 'washed out' because the flow velocity is too great to sustain deposition at the dune crest. Dunes then give way to a **plane bed**, but one that is rather different from the initial flat bed. Above the bed is a clearly defined zone of suspended sediment within which 'dust storm conditions' prevail (Leopold *et al*., 1964). This marks the transition to the **upper flow regime**, where the Froude number (ratio of inertial and gravitational forces) is greater than 1 and flow becomes supercritical. Upper flow regime bedforms include **standing wave antidunes**, where the sediment is moving but the waves themselves are stationary. This is because rates of deposition on the upstream side are matched by erosion on the downstream side. The position of standing waves is marked by waves at the water surface, the sand and water waves being in phase with each other. At higher flows sediment is thrown up from the downstream side of the bedforms at a faster rate than it can be replenished, which results in **antidunes**. These migrate *up*stream, while the sediment continues to move downstream. At very high flows, a series of **chutes and pools** develop. Chutes have a near-plane bed and shooting flow, which enters downstream pools: deeper sections that are marked by hydraulic jumps. Bedforms in gravel and mixed sand–gravel channels Bed structures also form in gravel-bed channels and have been a focus of research over recent decades. **Pebble clusters** are commonly found in this type of channel and form when a single large particle acts as an obstacle, protruding into the flow and encouraging the accumulation of coarse material on its upstream side. This upstream material may have an imbricated structure, increasing stability and requiring larger lift and drag forces to entrain the constituent particles. Finer particles are found on the downstream side of the obstacle, where shelter is provided from lift and drag forces. **Transverse ribs** are another type of gravel bedform and consist of regularly spaced ridges of coarser pebbles, cobbles or boulders that lie transverse to the flow. Like sand bedforms, these features affect flow resistance as well as rates of bedload transport. Where the bed is composed of a mixture of sand and gravel, the different mobility of the constituent particles can lead to some interesting effects. For example, longitudinal ribbons of sand have been observed to travel downstream, snaking from side to side over immobile gravel beds. Bedload can also move in thin sheets as an elongated procession of sediment with a thickness of one to two grain diameters. The coarsest sediment accumulates at the leading edge and there is a progressive fining of sediment behind it. Bedload sheets appear to be fairly common in mixed channels and are related to rates of sediment supply, becoming less frequent and reduced in extent as supply rates are reduced (Dietrich *et al*., 1989). **Assessing rates of bedload transport** From the preceding discussion you will have some idea of just how complex and variable bedload transport is. There is a general paucity of data on rates of transport because the available techniques can be expensive and time-consuming to employ. These include the collection of bedload over a period of time using portable samplers or traps excavated in the bed. Another approach is to track the movement of individual particles.

SUSPENDED LOAD TRANSPORT PROCESSES

Particles carried in suspension are kept aloft by turbulent eddies and will remain in suspension as long as their weight is supported by the upward component of turbulent eddies. In a fluid at rest, a suspended particle will fall through the fluid column. The rate of fall, or **fall velocity**, is a function of the density, size and shape of the particle. It is also determined by the viscosity and density of the transporting fluid. Since the falling particle displaces fluid, its movement is resisted by an equal and opposite fluid drag force. If sufficient depth is available, the falling particle will accelerate until it reaches a terminal velocity. In channels, the fall velocity is further affected by flow turbulence and the interactions of surrounding particles (Chanson, 1999). Considerable variation is seen between particles of different sizes. The fall velocity for the finest wash load component is very low, meaning that this sediment can be transported over considerable distances. For example the terminal fall velocity of a silt grain (0.001 mm) is approximately 0.004 cm s-1, but increases to 34 cm s-1 for a 10 mm gravel particle (Chanson, 1999). Suspended sediment is transported by processes of **advection** and **turbulent diffusion**. Advection is the transport of sediment within the flow, where the sediment moves with the flow itself. Turbulent diffusion refers to the mixing of sediment through the depth profile by turbulent eddies. Within the depth profile, the greatest concentration of suspended sediment is found towards the bed of the channel. Although there is continuous movement of individual suspended grains, the overall concentration and average grain size generally decrease rapidly away from the bed. This is due to interaction between the fall velocity and the vertical component of flow associated with turbulent eddying (Knighton, 1998). The upward migration of sediment to zones of lower concentration is both an advective and a diffusion process. A related process, which is called **convection**, involves the entrainment of sediment by large-scale vortices. For example, sediment is suspended in vortices generated as a result of flow separation in the toughs of ripples and dunes (Bridge, 2003). Large-scale vortices also occur where there are sudden drops in bed elevation, at hydraulic jumps and during overbank flows. Interactions between channel and floodplain flows were discussed in Chapter 6 Vortices created within the shear zone between the faster moving channel flow and slower flow on the floodplain result in lateral transfers of water. From Figure 6.10 you will see that there is a nearsurface flow out onto the floodplain and a return flow back towards the channel. Sediment carried in the flow is also transferred from channel to floodplain. Most of the coarser sediment is deposited near to the channel margins because of the rapid deceleration of flow and reduced turbulence (Bridge, 2003). This explains the origin of alluvial ridges, or natural levees, that are found along the margins of some channels. Finer suspended sediment, especially the wash load component, is carried out onto the floodplain. **Sediment supply and transport rates** The main sources of suspended sediment include material washed in from hillslope erosion and the release of fine material and aggregates from bank erosion. The supply of fine sediment is a major control on rates of suspended sediment transport. Most suspended transport, particularly the wash load, is supply limited. This means that the supply of fine sediment often has a greater influence on the sediment concentration than flow conditions in the channel. The rate of supply varies during individual events, between events, seasonally and annually. These variations are controlled by a number of variables, including antecedent conditions, rainfall intensity, hydrograph shape and vegetation growth. High discharges tend to be associated with greater concentrations of suspended sediment. This is because the supply is increased by storm-induced erosion of hillslopes and channel banks, and the release of fine sediment from storage. With all this in mind, it is hardly surprising that no simple relationship exists between suspended sediment concentration and flow discharge for a given crosssection.

References:

Andrews, E.D., 1983. Entrainment of gravel from naturally sorted riverbeds material. *Geological Society of America Bulletin*, 94: 1225–31.

Ashworth, P.J. and Ferguson, R.I., 1989. Size selective entrainment of bedload in gravel bed streams. *Water Resources Research*, 25: 627–34.

Bathurst, J.C., 1987a. Measuring and modelling bedload transport in channels with coarse bed materials. In: K.S. Richards (ed.), *River Channels: Environment and Process*. Blackwell, Oxford, pp. 272–94.

Bridge, J.S., 2003. *Rivers and Floodplains: Forms, Processes, and Sedimentary Record*. Blackwell, Oxford.

Chanson, H., 1999. *The Hydraulics of Open Channel Flow: An Introduction*. Arnold, London.

Knighton, D.A., 1998. *Fluvial Forms and Processes: A New Perspective*. Arnold, London.

Leopold, L.B., Wolman, M.G. and Miller, J.R., 1964. *Fluvial Processes in Geomorphology*. Freeman, San Francisco. Dietrich, W.E., Kirchener, J.W., Ikeda, H. and Iseya, F., 1989. Sediment supply and the development of the coarse surface layer in gravel-bedded rivers. *Nature*, 340: 215–17.

Nanson, G.C., Tooth, S. and Knighton, D., 2002. A global perspective on dryland rivers: perceptions, misconceptions and distinctions. In: L.J. Bull and M.J. Kirkby (eds), *Dryland Rivers: Hydrology and Geomorphology of Semi-arid Channels*. John Wiley & Sons, Chichester, pp. 17–54.

Parker, G., Klingeman, P.C. and McLean, D.G., 1982. Bedload and size distribution in paved gravelbed streams. *Journal of the Hydraulic Division, ASCE* 108(4): 544–71.

Parker, G., Klingeman, P.C. and McLean, D.G., 1982. Bedload and size distribution in paved gravelbed streams. *Journal of the Hydraulic Division, ASCE* 108(4): 544–71.

Reid, I., Bathurst, J.C., Carling, P.A., Walling, D.E. and Webb, B.W., 1997. Sediment erosion, transport and deposition. In: C.R. Thorne, R.D. Hey and M.D. Newson (eds), *Applied Fluvial Geomorphology for River Engineering and Management*. John Wiley & Sons, Chichester, pp. 95–135.

Richards, K.S., 1982. *Rivers: Form and Process in Alluvial Channels*. Methuen, London.

Wilcock, P.R., 1992. Experimental investigation of the effect of mixture properties on transport dynamics. In: P. Billi, R.D. Hey, C.R. Thorne and P. Tacconi (eds), *Dynamics of Gravel-bed Rivers*. John Wiley & Sons, Chichester, pp. 109–39.

UNIT- 10: DEPOSITION: FACTORS CONTROLLING DEPOSITION, DEPOSITIONS ALONG THE CHANNEL AND ACROSS THE CHANNEL

Sediment particles are deposited when there is a reduction in the competence and capacity of the flow. The process itself takes place at a very small scale and involves individual grains, although depositional forms can be observed over a wide range of spatial scales, from the smallest bedforms to vast floodplains and deltas. The construction and development of depositional forms might be likened to the building of an anthill. The process of building the anthill involves individual ants carrying soil one crumb at a time to the site of the ant hill. Although this process takes place at a small scale, the resulting feature is much bigger than the individual ants and crumbs of soil that created it. Thresholds for deposition are associated with the fall (or settling) velocity defined earlier. The deposition of suspended sediment takes place when the fall velocity dominates over turbulent diffusion. Since the fall velocity is closely related to particle size, the coarsest sediment tends to be deposited first. This leads to sediment sorting, a vertical and horizontal gradation of sediment, from coarse to fine. It should be noted that the fall velocity is also affected by the viscosity and density of the fluid. These are both influenced by changes in suspended sediment concentration. In addition, finer material can be transported as agglomerations of sediment called flocs. These have a greater fall velocity than the individual particles forming them. In the case of bedload transport, the near-bed flow conditions are significant. Bedload deposition occurs where the bed shear stress drops below the critical shear stress (Shields's parameter) required to transport particles of a given size. Local patterns of sediment sorting are well known, for example a downstream reduction in bed particle size is commonly observed along channel bars (e.g. Bluck, 1982; Smith, 1974).

Where sediment is deposited

There are a number of different circumstances that lead to deposition. These include:

● ***Reductions in flow discharge*** which are seen as flows recede, or along dryland rivers, where downstream losses are caused by high rates of evaporation and percolation.

● ***Decreases in slope*** which can be localised, or involve a gradual reduction over a longer length of channel and cause a reduction in average flow velocity and stream power.

● ***Increases in cross-sectional area*** cause the flow to diverge and become less concentrated. Flow resistance increases because there is more contact between the flow and channel boundary. There is a large increase in cross-sectional area when overbank flows occur.

● ***Increases in boundary resistance*** are associated with vegetation and coarse bed sediment. When overbank flows occur, velocity is reduced by the increased roughness of the floodplain surface, leading to the deposition of suspended sediment.

● ***Flow separation,*** which causes sediment to become decoupled from the flow.

● ***Obstructions to flow.*** Sediment often accumulates behind obstructions. These include boulders, outcrops or islands of bedrock, woody debris and man made structures such as bridge piers, dams and flow control structures. Changes in the supply of sediment are also important. For example, sediment tends to accumulate immediately downstream from scour zones caused by flow convergence, when the material scoured from the channel bed is deposited immediately downstream. At a larger scale, increases in the supply of sediment to a channel reach are caused by changes within the upstream drainage area.

Depositional environments

Although deposition does occur in the production and transfer zones of the fluvial system it dominates in the deposition zone, where there is a decline in gradient and energy availability. Large-scale deposition leads to the development of characteristic landforms, including floodplains, alluvial fans and deltas. Within channels, **bars** represent smaller-scale depositional features. They are commonly found on the inside of meander bends, along the edges of channels, and where tributaries join the main channel. Braided channels are characterised by numerous midchannel bars. **Floodplains** border the channels of alluvial rivers and are formed from a mixture of in-channel and overbank deposits. Their development and evolution, is governed by a number of factors, including the supply of sediment (volume and calibre), the energy environment of the channel, and the valley setting. Sediment is laid down by rivers as they migrate across the floodplain, being deposited on the inside of meander bends or when braid bars are abandoned. These channel deposits are relatively coarse in comparison with the much finer sediment that is laid down by overbank flows. Processes of erosion can also be significant in reworking sediment or in removing part, or all, of the floodplain surface. **Alluvial fans** are typically found in situations where an upland drainage basin flows out onto a wide plain. The sudden change from confined to unconfined conditions leads to flow divergence, while mean flow velocity is decreased by the reduction in slope. The resultant deposition leads to the formation of a conical feature with a convex cross-profile. Most fans have a radius of less than 8 km, but can be more than 100 km wide in some cases. Where a number of individual fans develop along a mountain front, they may grow laterally and coalesce to form a sloping apron of sediment called a **bajada**. Fans are commonly found in dry mountain regions, where an abundant sediment supply is associated with extreme discharges and frequent mass movements. Frequent shifts are often seen in the position of the braided channels that cross the fan surface, although only part of the fan surface may be active during a major flood event. In long profile, the slope is steepest at the fan head, progressively decreasing along the length of the fan. There is also a down-slope reduction in sediment size, although deposits are coarse and poorly sorted. Incision and fan head trenching is associated with decreases in sediment supply, or increases in slope. Such changes can be caused by tectonics, climatic variations, a fall in regional or local base level, or human activity. In the absence of external change, the progressive lowering of the landscape will also result in a decline in sediment yield over time. Arid fans are generally smaller and steeper than those found in humid regions, a large-scale humid example being the Kosi Fan on the southern Himalayan mountain front. This covers an area of 15,000 km2 and formed where the Kosi River descends onto the wide alluvial plain of the Indus. It has a very low gradient, only averaging 1 m km-1 at its head, with further decreases downstream (Summerfield, 1991). **Deltas** are found where sediment-charged flowing water enters a body of still water. They extend outwards from shorelines where rivers enter lakes, inland seas and oceans. In coastal areas deltas form where the supply of sediment is greater than the rate of marine erosion, although sediment is redistributed by coastal processes. The influence of fluvial processes tends to dominate in the case of lake deltas.

References:

Bluck, B.J., 1982. Texture of gravel bars in braided streams. In: R.D. Hey, J.C. Bathurst and C.R. Thorne (eds) *Gravel-bed Rivers*, John Wiley & Sons, Chichester, pp. 339–55.

Smith, N.D., 1974. Sedimentology and bar formation in the upper Kicking Horse River, a braided outwash stream. *Journal of Geology*, 82: 205–23.

Summerfield, M.A., 1991. *Global Geomorphology: An Introduction to the Study of Landforms*. Longman, Harlow.

UNIT-11: SEDIMENT DEPOSITS: NATURE AND CHARACTERISTICS, FLOOD PLAIN AND DELTAIC PLAIN DEPOSITS

Processes of floodplain formation The morphology of floodplains is intimately linked with the form and behaviour of the river channels that shape them. Various processes of deposition, reworking and erosion are involved in the formation and development of floodplains. Sediment accumulates on floodplain surfaces by various processes of accretion, the main ones being vertical, lateral and braid bar accretion (Nanson and Croke, 1992). **Lateral accretion** deposits are laid down by migrating rivers, which erode into the floodplain and lay down sediment in their wake. The accretion of point bar deposits can sometimes be seen as a series of concentric ridges on the inside of bends called **meander scrolls**. **Braid bar accretion** occurs when bars are abandoned and gradually become incorporated into the floodplain deposits. There are various ways in which this can happen, for example when a large flood lays down extensive bar deposits. Alternatively, bars may become abandoned when the main braid channels shift to another part of the valley. **Vertical accretion** deposits are composed of fine material that settles out of suspension when overbank flows inundate the floodplain. The increased area of contact, coupled with the roughness of the floodplain surface, greatly reduces flow velocities, and a thin layer of sediment is draped across the floodplain. This displays a fining-upwards sequence, where the coarser particles, which settle out first, are overlain by progressively finer material. There is also a fining of sediment away from the channel, since only the very smallest particles are carried to the edge of the inundated area. Over a number of years the cumulative effect of overbank flows leads to the development of a vertical sequence of thin layers. Other, more localised, types of accretion can also be identified. For example, **counterpoint accretion** is associated with the deposition of concave bank benches at confined meander bends (see section on channel geomorphic units above). As an over-tightened meander bend migrates, bench deposits become incorporated into the floodplain. Erosional processes include **floodplain stripping**, where entire sections of the floodplain surface are removed by high-magnitude flood events. Floodplain stripping is most likely to occur in relatively confined valley settings, where floodplain flows are concentrated between the valley walls. Other erosional processes include the formation of flood channels, which carry water during overbank flows. **Avulsion** involves a shift in the position of a channel and is a common process in braided reaches where the flow frequently abandons and reoccupies sub-channels. Avulsion can also involve the diversion of flow into a newly eroded channel cut into the floodplain. This type of avulsion is important in the development of anabranching channels. The morphology and development of floodplains is controlled by the driving variables and boundary conditions. An important balance exists between the shear stress exerted by the flow and the resistance of the floodplain to erosion. Shear stress is closely related to specific stream power, and therefore to such controls as flow regime, valley slope and valley confinement. On the other side of the balance, resistance to erosion is largely determined by the cohesiveness of the floodplain sediments. An energy-based floodplain classification was proposed by Nanson and Croke (1992). This recognises three main classes of floodplain:

● ***High-energy non-cohesive floodplains*** are typically found in steep upland areas where the specific stream power in the channel at bankfull flow exceeds 300 W m–2. An example is shown in Plate 8.7. Lateral migration is often prevented by the coarseness of the floodplain sediment, which builds up vertically over time. These floodplains are disequilibrium features that are partly or completely eroded by infrequent extreme events.

● ***Medium-energy non-cohesive floodplains*** are formed from deposits ranging from gravels to fine sands. Specific stream power ranges from 10 to 300 W m–2. The main processes of floodplain construction are lateral point bar accretion (meandering channels) and braid bar accretion (braided channels). These floodplains are typically in dynamic equilibrium with the annual to decadal flow regime.

● ***Low-energy cohesive floodplains*** are usually associated with laterally stable single thread or anastomosing channels. Formed from silt and clay, the dominant processes are vertical accretion of finegrained sediment and infrequent channel avulsions. Specific stream power at bankfull stage is generally less than 10 W m-2.

Floodplain geomorphic units

Levees

Levees are elongated, raised ridges that form at the channel–floodplain boundary during overbank flow events. Moving across the boundary from channel to floodplain, there is a sudden loss of momentum because of the interaction between fast channel flow and slow floodplain flow. This results in the preferential deposition of material at the edges of the channel. Levees are clearly visible as the raised strips of land running along the channel margins. The height of levees is scaled to the size of the channel and their presence implies a relatively stable channel location (Brierley and Fryirs, 2005). These natural levees should not be confused with the artificial levees that are constructed along river banks for purposes of flood control.

Crevasse splays

Levees can be breached by floodwaters. This may lead to the formation of a crevasse splay, a fan-shaped lobe of sediment deposited when sediment-charged water escapes and flows down the levee. If flow is sufficiently concentrated, a new channel may be cut and deepened by scour.

Backswamps

The build-up of sediment in the channel may mean that the channel is at a higher elevation than the surrounding floodplain. When levees are overtopped, water can enter the lower-lying area on the other side of the levee. This may be a depression or a swamp area characterised by wetland vegetation. These are not exclusively associated with anabranching rivers and can also form at the valley margins of other channel types. Flood channels Flood channels are relatively straight channels that bypass the main channel. They have a lesser depth than the main channel and are dry for much of the time, only becoming filled with water as the flow approaches bankfull.

Floodouts

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

Meander scroll bars

In some cases, former point bar deposits can be seen in the surface topography of the floodplain as **scroll bars**, with each scroll representing a former location of the point bar. The undulating **ridge and**

swale topography that results consists of higher ridges separated by topographic lows called swales. Migrating meanders do not always form scroll bars and the surface topography of these deposits may be relatively featureless.

Cut-offs

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

Palaeochannels

Palaeochannels are longer sections of abandoned channel. Like active channels, palaeochannels exhibit a wide range of different planforms. As time goes by, they gradually become infilled by abandoned channel accretion, the degree of infilling reflecting the age of the channel.

References:

Brierley, G.J. and Fryirs, K.A., 2005. *Geomorphology and River Management: Applications of the River Styles Framework*. Blackwell, Oxford.

Nanson, G.C. and Croke, J.C., 1992. A genetic classification of floodplains. *Geomorphology*, 4: 459 86.

Tooth, S., 1999. Floodouts in central Australia. In: A.J. Miller and A. Gupta (eds), *Varieties of Fluvial Form*. John Wiley & Sons, Chichester, pp. 219–47.

UNIT-12: FLUVIAL PROCESSES AND FORMS

Controls on Channel Adjustment and Form

Flow and sediment supply both fluctuate through time, meaning that continuous adjustment takes place through the erosion, reworking and deposition of sediment. The flow and sediment regimes are called **driving variables** because they drive these processes. Along a given reach, channel adjustment is constrained within certain boundaries that are imposed by local conditions. For example, a sand-bed river flowing across a wide floodplain is able to adjust its form much more readily than a bedrock channel confined within a narrow gorge. Energy availability is also important, and channel adjustments are often limited for rivers that flow over low gradients, especially where cohesive banks are protected by vegetation. These constraints are called **boundary conditions** and include valley confinement, channel substrate, valley slope and riparian vegetation. A channel is said to be 'in regime' when its form fluctuates around an equilibrium condition over the time scale considered. Not all channels are in regime, and there are many examples of non-regime, or disequilibrium, channels. This may be because the channel is evolving in response to long term changes in the flow or sediment regime, caused by a change in one of the external basin controls. Examples include incising or aggrading channels and those that are undergoing a change in channel pattern. Alternatively, some bedrock and dryland channels may exist in a permanent state of disequilibrium because it is only during flood flows that adjustments take place. In such cases, low flows have little or no influence on the overall channel form. Many empirical relationships have been developed to relate 'regime dimensions' (e.g. channel width or depth), to control variables (e.g. bankfull discharge). It is important to realise that these regime dimensions represent an average and are not applicable to all channel types or flow regimes. At the sub-reach scale there are spatial variations in energy expenditure, which result from variations in channel shape and resistance to flow. These in turn influence patterns of erosion and deposition. For example, energy and erosion potential are concentrated where the channel narrows. Conversely, flow resistance is increased by obstructions to flow such as boulders, bedforms, bars or woody debris, which can lead to localised deposition. There is therefore two-way feedback between channel form and flow hydraulics – form influences flow and flow influences form. This point is well illustrated by the work of Ashworth and Ferguson (1986) on a glacially fed braided river in Arctic Norway. An intensive monitoring programme was carried out to make detailed measurements of channel morphology, velocity and shear stress, bedload size and transport rate, and the size of bed material. Starting at the top left of this diagram is the discharge of the river, which is unsteady (varies over time). The irregular form of the channel creates non-uniform flow conditions over the rough channel bed. As a result, complex spatial variations are seen in velocity, which also changes over time. At any point in the channel, the bed shear stress is determined by the vertical velocity profile. Rates of bedload transport are determined by bed shear stress as well as the size and amount of bed material that is available for transport. As with velocity and shear stress distributions, rates of bedload transport are spatially variable, and also change with time. Bedload transport may maintain the existing channel shape, size and pattern. Alternatively, channel form can be modified as a result of scour, fill and possible lateral migration. The nature of such changes is spatially variable, and in turn feeds back to influence the velocity distribution within the channel. Bedload transport also governs the size distribution and structure of bed sediments through selective entrainment and transport. The character of the bed material determines the roughness of the channel, in turn affecting the velocity distribution in the channel. **Driving variables** Flow regime The flow in natural river channels is unsteady, fluctuating through time in response to inputs of precipitation to the drainage basin. Characteristics of the flow regime, include seasonal variations, flood frequency–magnitude relationships and the frequency and duration of low flows. Since discharge influences stream power, velocity and bed shear stress, the characteristics of the flow regime have an important influence on channel form. Of morphological significance is the bankfull discharge. The bankfull discharge marks a morphological discontinuity between within-bank and out-of-bank flows. Since the flow in natural channels is unsteady, the bankfull discharge provides a representative flow. The geomorphological effectiveness of a given flood depends not only on its size, but also on the frequency with which it occurs. Large floods can carry out a considerable amount of geomorphological work. However, their comparative rarity means that the cumulative effect of smaller, more frequent flows may be more significant in shaping the channel. Bankfull discharge (or the equivalent bankfull width) has often been used in developing statistical relationships between discharge and channel form parameters. It is important to realise that bankfull discharge is actually quite difficult to define and that its frequency of occurrence varies considerably between different rivers. Sediment regime The supply of sediment varies through time. It is not only the volume of sediment that is important but also its size distribution. As you will see later in this chapter, there are significant differences in the behaviour and morphology of bedload, suspended load and mixed load channels. Fluctuations in the volume and size of sediment are brought about by variations in sediment supply from the drainage basin and processes of sediment transfer through the channel network. As with the flow regime, it is the processes in the drainage basin, upstream from a given reach, that influence sediment supply. The balance between stream power and sediment supply There is an important balance between the supply of bedload at the upstream end of a channel reach and the stream power available to transport it. This is known as the Lane balance, having first been described as a qualitative equation by Lane in 1955. The left hand side of the scales represents the volume and size of sediment supplied to a channel reach over a given period of time. Balanced against this is the stream power available to transport it. This is determined both by the volume of water that enters the reach (over the same time period), and by the slope over which it flows. If the stream power is exactly sufficient to transport the sediment load, both sides of the scales are in balance and there is no net erosion or deposition along the reach. This is not to say that there is no erosion or deposition whatsoever, because these processes do occur at a localised scale in response to local variations in hydraulic conditions. Rather it means that, on balance, neither erosion nor deposition will predominate. An imbalance will occur if there is an increase in the volume or calibre of the sediment load in relation to the available stream power (sediment calibre is important because it determines the flow competence required to transport it). This means that there is insufficient stream power to transport all the sediment, with the result that the excess is deposited along the reach. In this case, the balance tips towards **aggradation**, with net deposition occurring along the reach. Aggradation can be triggered in several ways, for example where the sediment supply is increased by upstream channel erosion, mass movement, or human activities such as mining. Aggrading channels are characterised by numerous channel bars in a wide, shallow channel. Deposition within the channel may lead to the channel bed becoming elevated above the surface of the floodplain. This, together with reduced channel capacity, increases the incidence of flooding and also promotes channel migration. A different situation arises when the stream power exceeds what is needed to transport the sediment load through the reach. This excess energy has to be expended somehow, so it is used to entrain sediment from the bed and erode the channel boundary. In this case **degradation** predominates. Degradation can be caused by an increase in discharge, perhaps caused by an increase in flood frequency, or by a decrease in sediment supply. This can occur downstream from dams or where gravel mining has removed sediment from the river bed. The Lane balance is simplistic because much depends on the calibre of bed sediment within the reach. For example, no degradation can occur in a boulder-bed stream if the bed sediment is too coarse to be moved by the available stream power. This can be true even if the stream power exceeds the sediment supplied at the upstream end of the reach. Even when degradation does occur, another limitation of the equation is that it does not tell us *where* within the reach erosion will occur (Simon and Castro, 2003). This means that the equation cannot be used to predict the actual nature of channel change. For example, if the channel bed is more resistant to erosion than the banks, bank erosion is likely to be an initial adjustment. However, in a sand-bed channel with cohesive banks it is more likely that an initial adjustment would be scouring of the bed (Simon and Castro, 2003). Resistance to erosion can be highly variable within a given reach, as can the specific stream power along that reach. This gives rise to spatially complex adjustments along the reach, even if there is net aggradation or net degradation along the reach as a whole. **Boundary conditions** Valley slope This refers to the downstream slope of the valley floor (as opposed to the slope of the channel itself) and determines the overall rate at which potential energy is expended along a given reach. The valley slope imposed on a given reach of channel is determined by a combination of factors including tectonics, geology, the location of the reach within the drainage basin and the long-term history of erosion and sedimentation along the valley. Although the overall energy available along a given reach is largely determined by the valley slope, it is possible for adjustments to occur that increase flow resistance at different scales (channel resistance, form resistance and boundary resistance). Different types of channel and floodplain morphology are associated with low, medium and high-energy environments. Valley confinement A channel may be defined as confined, partly confined, or unconfined, depending on how close the valley sides are. The degree of valley confinement is important for several reasons. In **confined** settings channel adjustments are restricted by the valley walls, which also increase flow resistance. In addition, valley width influences the degree of slope–channel coupling that exists. Inputs of sediment from mass movements and other slope processes may exceed transport capacity, in turn influencing channel form. The episodic nature of mass movements means that these contributions can vary considerably over time. In **partly confined** settings, some degree of lateral migration and floodplain development is possible. However, where the river comes against the valley wall or hillslope it is prevented from migrating further, which can lead to the development of over-deepened sections of channel. Stream power is also concentrated within the narrow valley and sections of the floodplain surface may be stripped during major floods. Where the hillslopes are a long way from the channel and have relatively little influence in contributing to the channel load, the channel is described as **unconfined**. Typically these settings are found in the lower reaches of rivers where there is very little interaction between channel and hillslopes. Channel substrate Considerable variations are seen in the form and behaviour of channels developed in different substrates. The substrate determines how resistant the channel is to the erosive force of the flow. It also influences boundary roughness, and therefore flow resistance. Alluvial channels formed in sand and gravel are generally more easily adjusted than those with cohesive silt and clay substrates. This is because the individual particles can be entrained at relatively low velocities, so non-cohesive substrates tend to be associated with wider, shallower cross-sections and faster rates of channel migration. Bedrock and mixed bedrock-alluvial channels are influenced over a range of scales by various geological controls. Riparian vegetation Vegetation on the banks and bed of river channels controls channel form in various ways. It often acts to protect and strengthen the banks, and research has shown that a dense network of roots can increase erosion resistance by more than a factor of ten. As a result, channels with vegetated banks are often narrower than those with non-vegetated banks under similar formative flows. This effect is most marked for densely vegetated banks (Hey and Thorne, 1986). Flow resistance can also be increased by vegetation growing on the bed and banks, as well as by woody debris (fallen trees and branches) that enters the channel from the banks. An interesting example of the influence of riparian vegetation on channel form is provided by the Slesse Creek, British Columbia, Canada, and is reported by Millar (2000) and MacVicar (1999). The Slesse Creek drains an area of 170 km2 within the Fraser River basin, flowing southwards from the United States into British Columbia.

Downstream changes

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

CHANNEL ADJUSTMENT

Time scales of adjustment

Different components of a channel’s morphology (e.g. bedforms, cross-sectional shape, slope) change over different time scales. This is because some components are more readily adjusted than others. For example, bedforms in a sand-bed channel are rapidly modified by a wide range of flows. Adjustments to channel width and depth take place over months to years, planform adjustments occur over tens to hundreds of years, while changes in the long profile may take thousands of years.

Morphological adjustments therefore tend to lag behind the changes that cause them. This means that it can be difficult to link processes of flow and sediment transport with channel dimensions and form. Channel form is directly controlled by flow regime and sediment supply.

How adjustments are made

Channel form and behaviour reflect the driving variables and boundary conditions influencing a given channel reach. These controls also influence the ways in which channel adjustments are made. There are potentially four **degrees of freedom**, or variables, that can be modified: channel cross-section, slope, planform and bed roughness. Modifications to the cross-sectional size and shape are associated with changes in width and depth of the channel by processes such as bank erosion, incision of the bed, or aggradation. Channel slope can be adjusted in different ways. Negative feedback reduces the slope of steeper sections by erosion, and the slope of flatter sections is increased by deposition. There are several different types of channel planform adjustments. These include lateral migration, meander bend development, reworking of bars, and even wholesale shifts of the channel to a new course. Finally, changes in bed roughness are brought about when the channel rearranges bed material, for example, in sand-bed channels, where bedforms are modified in response to changes in flow conditions. Mutual interrelationships exist between these variables, with adjustments made to one affecting one or more of the others. For instance, the formation of a meander cut-off alters the channel planform as well as increasing channel slope.

CHANNEL GEOMORPHIC UNITS

Geomorphic units are features that form at the subchannel scale and can be erosional or depositional in origin. 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. **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). **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. 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. 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. 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. 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. 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). 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. 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/near over bank 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 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 Bedrock bars** In incised bedrock channels, the flow sometimes moves around bedrock bars. 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.

References:

Ashworth, P.J. and Ferguson, R.I., 1986. Interrelationships of channel processes, changes and sediments in a proglacial braided river. *Geografisker Annaler*, 68A: 361–71.

Lane, E.W., 1955. The design of stable channels. *Transactions of the American Society of Civil Engineers*, 120: 1234–60.

Simon, A. and Castro, J., 2003. Measurement and analysis of alluvial channel form. In: G.M. Kondolf and H. Piegay (eds), *Tools in Fluvial Geomorphology*. John Wiley & Sons, Chichester: 291–322.

Hey, R.D. and Thorne, C.R., 1986. Stable channels with mobile gravel beds. *Journal of Hydraulic Engineering*, 112: 671–89.

Millar, R.G., 2000. Influence of bank vegetation on alluvial channel patterns. *Water Resources Research*, 36(4): 1109–18.

MacVicar, B.J., 1999. '*Application of a Rational Model of Stream Equilibrium for Predicting ChannelAdjustments'*. M.A. Sci. thesis, University of British Columbia, Vancouver, BC. Rice, S. and Church, M., 1998. Grain size along two gravel-bed rivers: statistical variation, spatial pattern and sedimentary links. *Earth Surface Processes and Landforms*, 23: 345–63.

Knighton, D.A., 1998. *Fluvial Forms and Processes: A New Perspective*. Arnold, London.

Smith, N.D., 1974. Sedimentology and bar formation in the upper Kicking Horse River, a braided outwash stream. *Journal of Geology*, 82: 205–23.

Ashmore, P.E., 1991. How do gravel bed rivers braid? *Canadian Journal of Earth Sciences*, 28: 326–41.

Church, M. and Jones, D., 1982. Channel bars in gravel bed rivers. In: R.D. Hey, J.C. Bathurst and C.R. Thorne, (eds), *Gravel Bed Rivers*. John Wiley & Sons, Chichester, pp. 291–338.

Robert, A., 2003. *River Processes: An Introduction to Fluvial Dynamics*. Arnold, London. Erskine, W.D. and Livingstone, E.A., 1999. In-channel benches: the role of floods in their formation and destruction on bedrock-confined rivers. In: A.J. Miller and A. Gupta (eds), *Varieties of Fluvial Form*. John Wiley & Sons, Chichester, pp. 445–76.

Keller, E.A. and Melhorn, N., 1978. Rhythmic spacing and origin of pools and riffles. *Geological Society of America Bulletin*, 89: 723–30.

Leopold, L.B., Emmett, W.W. and Myrick, R.M., 1966. *Channel and Hillslope Processes in a Semi-arid Area, New Mexico*. United States Geological Survey Professional Paper 323G.

Keller, E.A., 1972. Development of alluvial stream channels: a five stage model. *Geological Society of America Bulletin*, 83: 1531–6.

Carling, P.A., 1991. An appraisal of the velocity-reversal hypothesis for stable pool–riffle sequences in the River Severn. *Earth Surface Processes and Landforms*, 16: 19–31.

Bathurst, J.C., 1993. Flow resistance through the channel network. In: K. Beven and M.J. Kirkby (eds), *Channel Network Hydrology*. John Wiley & Sons, Chichester, pp. 43–68.

Chin, A., 1999. The morphologic structure of step-pools in mountain streams. *Geomorphology*, 27: 191–204.

Knighton, D.A., 1998. *Fluvial Forms and Processes: A New Perspective*. Arnold, London.

1.7. Self Assessment Test

- What is the meaning of the term fluvial? Discuss the significance of fluvial geomorphology.

- Differentiate between internal and external variables in fluvial systems. Discuss in detail the concept of space scale, time scale and equilibrium in fluvial geomorphology.
- Define river basin. Discuss the area ratio and Law of basin area.
- What is flow resistance? What are the controls of flow resistance?
- Define stream power. Discuss the bank erosion processes in alluvial channels.
- What is lift force? Explain the process of particle entrainment.
- Differentiate between bed load and suspended load. Give an account of the geomorphic features created by sediment deposits.

1.8 Study Tips

- Charlton, R. (2008). *Fundamentals of Fluvial Geomorphology.* New York: Routledge.

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

- Singh, S. (2005). *Geomorphology.* Allahabad: Prayag Pustak Bhawan.

- Thornbury, W. D. (1958). *Principles of Geomorphology.* NewYork: Wiley.

Disclaimer: This self-learning material is compiled from different books, journals and web-sources.