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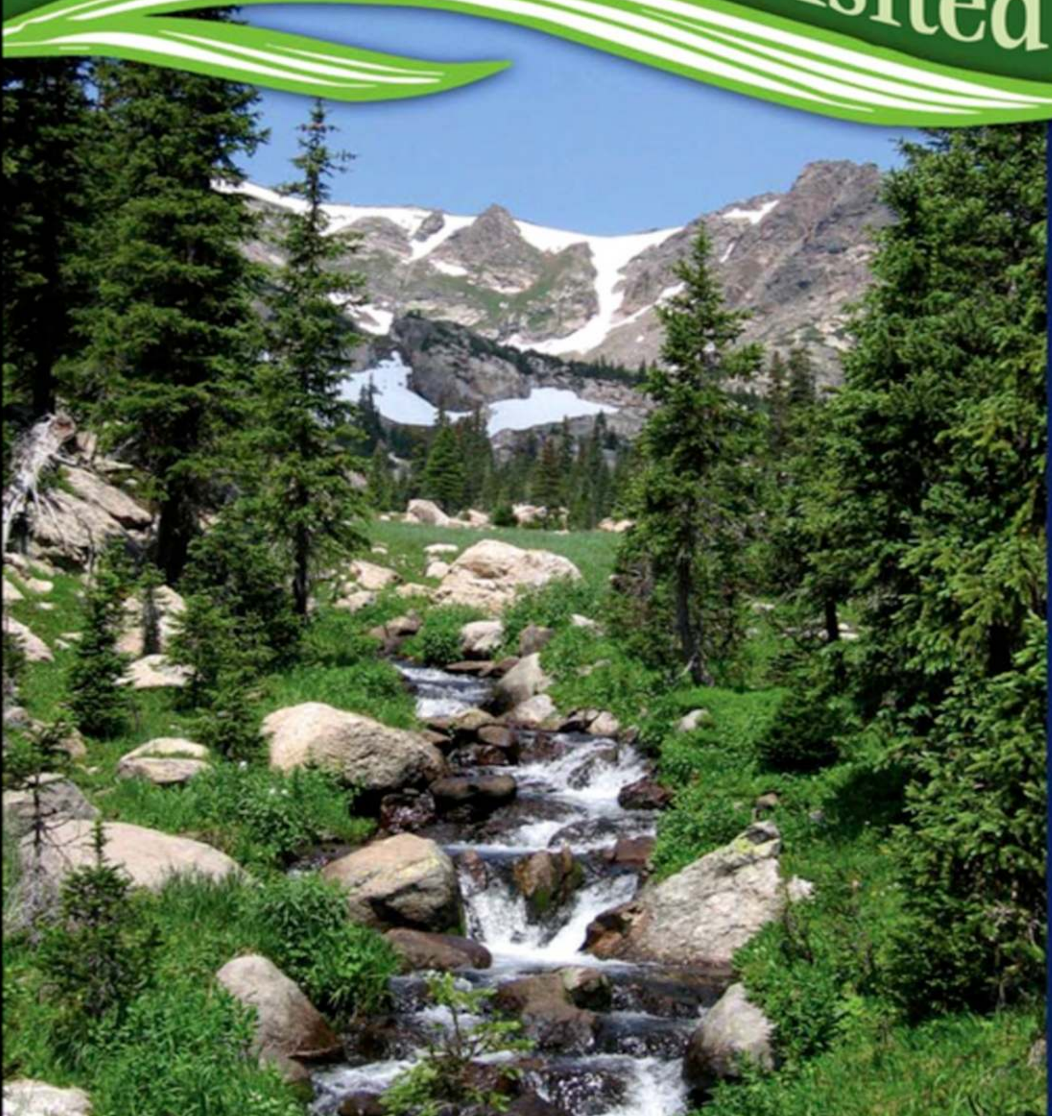
Mountain Rivers Revisited



Ellen Wohl

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MOUNTAIN RIVERS REVISITED

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Cover photographs by Ellen Wohl. (front) Tributary of Tonahutu Creek, Rocky Mountain National Park, Colorado. (back) Rio Sardinal in Braulio Carillo National Park, Costa Rica.

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CONTENTS

PREFACE

I wrote the first edition of this book, published in 2000, in response to a need expressed by one of my Ph.D. students at the time, David Merritt, who walked into my office one afternoon for a summary reference on mountain rivers. When I realized that such a reference did not exist, I set out to create one. The inclusion of topics reflected my own belief that rivers need to be examined not solely as physical systems but also as river ecosystems with chemical and biological components that exist in the context of pervasive and long duration human alteration of the environment. As research on topics related to mountain rivers grew dramatically during the past decade, I decided that it was time to write a second edition, and I reorganized the book to reflect my understanding of evolving knowledge.

As with the first edition, this second edition is aimed primarily at an audience already familiar with the basics of river process and form, although the reader with little knowledge of related topics, such as river chemistry, hyporheic zones, or riparian and aquatic ecology, can also gain a quick introductory overview of those topics from this volume. Advanced undergraduates, graduate students, and professional scientists and engineers who possess some general knowledge of river systems will find this volume of use, both for its own sake and to help them build on their existing knowledge of mountain rivers to better understand the unique aspects of these rivers. You can read the book straight through, because each section builds upon the sections that precede it, or use the book as a spot reference to provide a synthesis of current knowledge on specific topics.

The first edition benefited substantially from discussions with, and critical reviews by, Paul Carling (University of Southampton, England), Dan Cenderelli (U.S. Forest Service), Alan Covich (University of Georgia), Janet Curran (U.S. Geological Survey), Jim Finley (Telesto Solutions, Inc.), David Merritt (U.S. Forest Service), and LeRoy Poff (Colorado State University) and AGU reviews by John Costa (U.S. Geological Survey), Avijit Gupta (University of Leeds, England), and Malcolm Newson (University of Newcastle upon Tyne, England). Much of that material is still in this edition, and I thank each of these individuals for their efforts. The second edition has also benefited from discussions with Gordon Grant (U.S. Forest Service), Bob Hilton

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(Durham University), Neils Hovius (Cambridge University), Mark Macklin (University of Aberystwyth), and Grant Meyer (University of New Mexico) and reviews by Jim O'Connor (U.S. Geological Survey) and an anonymous reviewer, as well as the enhanced energy and concentration provided by Whole Foods' organic French roast coffee.

As with the first edition, I would like to dedicate this second edition to my graduate students. They continue to challenge, engage, and surprise me and to provide much of the pleasure that comes from working in fluvial geomorphology.

Ellen Wohl
Colorado State University

1. INTRODUCTION

Rivers shape many of the world's landscapes. In the process of transporting water, sediment, and dissolved chemicals from uplands, rivers redistribute mass across the Earth's surface. Rivers set the pace at which weathering and erosion lower landscapes, and control the gradient of adjacent hillslopes. Fundamentally, rivers organize terrestrial landscapes into drainage basins. As the rivers incise or aggrade in response to changes in baselevel, they create valleys that influence local climate; provide travel corridors for animals and humans; and support aquatic and riparian ecosystems that contain some of the Earth's highest levels of biodiversity.

Scientists have systematically studied rivers for more than two centuries. Among the questions asked have been: How do rivers interact with other variables such as climate, lithology and tectonics that influence landscapes? What governs the spatial distribution of river channels? What factors control the yield of water and sediment from hillslopes to rivers? How do interactions between water and sediment influence channel geometry through time and space?

This volume summarizes contemporary understanding of these and other aspects of rivers, in the context of rivers draining mountainous environments. Although the study of rivers is well-established, investigators typically focused on the lowland rivers along which most people live until the final decades of the 20th century. A substantial increase in the amount of research directed toward mountain rivers during the first decade of the 21st century supports the need for this second edition of *Mountain Rivers*, which was originally published in 2000. Increased attention to rivers in mountainous regions results from several trends within science and the greater society. Among these is the focus on numerically simulating landscape evolution over long timespans, which requires that modelers quantitatively parameterize rates of river incision and rates of crustal uplift in mountainous regions. Another factor driving increased investigation of mountain rivers is attempts to maintain or restore rivers as ecological refuges and as critical components of water supply in mountainous regions, which tend to be less densely populated than adjacent lowlands. Finally, mountain rivers with steep, coarse-grained, poorly-sorted beds, and limited sediment supply are typically poorly described by empirical equations for hydraulics and sediment dynamics developed for rivers with lower gradients, making the study of mountain rivers an intellectual and management challenge.

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1.1. Characteristics of Mountain Rivers

In this volume I define a mountain river as being located within a mountainous region and a mountainous region as having a mean elevation above sea level ≥ 1000 m [*Viviroli et al.*, 2003]. Each of the continents includes at least one major mountainous region (Figure 1.1). (Selected images appear in print. All images are available on the CD-ROM that accompanies the book.) Mountains cover 52% of Asia, 36% of North America, 25% of Europe, 22% of South America, 17% of Australia, and 3% of Africa, as well as substantial areas of islands including Japan, New Guinea, and New Zealand [*Bridges*, 1990]. Mountain rivers are thus widespread. Because of the steep topography of mountainous regions, mountain rivers typically have a gradient ≥ 0.002 m/m along the majority of the channel length [*Jarrett*, 1992], although substantial longitudinal variability of channel geometry is common in mountainous regions as a result of longitudinal variations in rock resistance, glacial history, and hillslope stability. Lower gradient reaches of channel typically occur upstream of glacial end moraines, massive landslide deposits, or beaver dams, for example, but these reaches create relatively short interruptions between the steeper channel segments up- and downstream.

As with lowland rivers, mountain rivers exhibit great variability in hydrologic regime; channel planform; channel gradient, grain size, and bedforms; sediment dynamics; and aquatic and riparian biota, both within individual mountain ranges and among diverse mountainous regions. Mountain rivers, as defined here, include first-order channels less than a meter wide fed by snowmelt draining an alpine meadow (Figure 1.2); wider rivers cutting steep-walled valleys that dense tropical rain forest vegetation cannot stabilize against periodic landslides (Figure 1.3); ephemeral channels incised into bedrock in arid mountains (Figure 1.4); boreal rivers with cutbanks exposing permafrost (Figure 1.5); and big, powerful rivers like the Indus that carry thousands of kilograms of sediment down to the adjacent lowlands each year (Figure 1.6). Perhaps the only consistent characteristic of mountain rivers is their typically steep gradients, although steep gradients tend to correlate with other characteristics, including

- erosionally resistant and hydraulically rough channel boundaries associated with bedrock and coarse clasts;
- highly turbulent flow with numerous longitudinal transitions between sub- and supercritical flow;
- limited supply of sediment of fine gravel and smaller size;
- bedload movement that is highly variable in space and time, with higher thresholds for initiation of motion than many lowland rivers;
- strongly seasonal discharge regime associated with glacial melt, snowmelt, or seasonal rainfall;
- substantial spatial variability in discharge as a result of spatial variability in precipitation and runoff caused by differences in elevation, basin orientation, and land cover;
- large longitudinal variations in channel geometry associated with variations in tectonics, lithology, glacial history, and sediment supply;

- in some cases, lesser temporal variations in channel geometry than lowland rivers because only infrequent floods or debris flows can exceed boundary resistance sufficiently to cause substantial channel change;
- relatively narrow valley bottoms with limited development of floodplains and lateral movements by rivers;
- in the absence of wide valley bottoms and the associated buffering of stream channels from hillslope processes, mountain rivers have the potential for orders-of-magnitude increase in water and sediment yield over a period of a few years following watershed-scale disturbances such as wildfire or timber harvest; and
- longitudinal zonation of aquatic and riparian biota influenced by river characteristics and by elevation as it relates to temperature and precipitation.

Mountain rivers tend exhibit high degrees of connectivity. *Landscape connectivity* [*Brierley et al.*, 2006] is high because individual landforms such as hillslopes and stream channels are closely coupled within a drainage basin. *Hydrological connectivity* [*Bracken and Croke*, 2007] is high because water moves rapidly from one landform to another and through the entire drainage basin relative to lowland watersheds with extensive groundwater storage. *Sediment connectivity* [*Fryirs et al.*, 2007] is high because limited storage means that sediment moves relatively rapidly from production sites on hillslopes through the drainage basin. Increasing research emphasis on different forms of connectivity reflects a desire to move beyond small spatial and short temporal scales of investigation in order to focus on emergent properties that evolve from the self-organization inherent in river catchments [*Phillips*, 2003; *McDonnell et al.*, 2007; *Reid et al.*, 2007b; *Ali and Roy*, 2009].

1.2. Advances Since the First Edition

Writing the second edition proved to be a much more time-consuming and expansive process than I had initially expected, but this reflects the dynamic nature of contemporary studies of geomorphology and mountain rivers. Many areas of investigation have expanded dramatically since the late 1990s and the volume of associated literature has grown correspondingly. Dramatic increases in the amount of research in topics such as: the interactions of tectonics, topography, and climate [*Willett et al.*, 2006]; hillslope hydrology and modeling [*Franks et al.*, 2005]; debris flows and associated hazards [*Jakob and Hungr*, 2005]; soil development and hillslope processes [*Heimsath et al.*, 2001; *Roering*, 2004]; hydraulics of steep channels [*Ferguson*, 2007]; braided river process and form [*Sambrook Smith et al.*, 2006]; diverse types of numerical models and associated predictions [*Wilcock and Iverson*, 2003; *Tucker and Hancock*, 2010]; geochronology [*Madsen and Murray*, 2009]; and instrumentation [*Jones et al.*, 2007] have made it challenging to keep track of and synthesize the literature. As a result, I have introduced several new sections to the second edition, substantially expanded other areas, and altered the organization of the volume to reflect changing research emphases within the community.

One broadly applicable change is the increasing emphasis on quantification, numerical modeling, and prediction in studies of the Earth's surface. This is exemplified by *Dietrich et al.*'s [2003] call for increased development and application of *geomorphic transport laws*. "A geomorphic transport law is a mathematical statement derived from a physical principle or mechanism, which expresses the mass flux or erosion caused by one or more processes in a manner that: 1) can be parameterized from field measurements, 2) can be tested in physical models, and 3) can be applied over geomorphically significant spatial and temporal scales" [*Dietrich et al.*, 2003, p. 103]. Geomorphic transport laws have been developed for some processes, including soil production from bedrock and river incision into bedrock, but do not yet exist for many geomorphic processes, including landslides, debris flows, and surface wash. Section 1.4 is designed to highlight the existing geomorphic transport laws relevant to mountain rivers and to provide an overarching conceptual framework for reading the succeeding, more detailed discussions of each of the processes and forms briefly mentioned in section 1.4.

1.3. Purpose and Organization of This Volume

This volume on mountain rivers is intended for the reader who already has a basic understanding of fluvial geomorphology, as developed in texts including *Leopold et al.* [1964], *Schumm* [1977], *Morisawa* [1985], *Richards* [1987], *Easterbrook* [1993], *Ritter et al.* [1995], *Bloom* [1998], *Knighton* [1998], *Bridge* [2003], or *Anderson and Anderson* [2010]. The emphasis of this volume is on channel processes and morphology, but the volume also includes brief reviews of other aspects of mountain rivers. The second chapter focuses on form and process at the scale of drainage basins (10^1-10^6 km^2), starting with interactions among tectonics, climate, and topography, and then reviewing hillslope processes, channel initiation and arrangement in a network, and valley geometry, including changes in process and form during the Quaternary. The third chapter covers process at the channel scale (10^{-2}-10^1 km^2), including hydrology, hydraulics, sediment dynamics, river chemistry, instream wood, and physical disturbances such as floods and debris flows. The fourth chapter examines types of channel morphology characteristic of mountain rivers and the fifth chapter discusses aquatic and riparian communities of mountain rivers. The sixth chapter explores human interactions with mountain rivers.

The diversity of topics addressed in this volume is designed to promote the realization that a mountain river is an integrated physical, chemical and biological system influenced by controls acting across various scales of time and space. The need to move beyond traditional disciplinary boundaries is reflected in the discussion of *Earth system science* starting in the late 20^{th} century. A system is a collection of interdependent parts enclosed within a defined boundary; in this case, the interdependent parts within the boundary of the Earth are the lithosphere, hydrosphere, biosphere, and atmosphere. Emphasis on a systems approach reflects an increasing realization that we cannot effectively respond to global warming, contaminant dispersal, and other contemporary challenges unless we think about natural processes in ways that transcend disciplinary boundaries. The establishment of critical zone observatories in the United States (the *critical zone* is defined as the Earth's outer layer, from the lower atmosphere and vegetation canopy to the soil and groundwater, which sustains living organisms) is also designed to promote integrative study of surface processes and landforms. Rivers provide an obvious mechanism for integrative thinking because a seemingly simple, discrete channelized flow of water in fact reflects influences from high in the atmosphere to deep in the crust and across hemispheres.

This volume is primarily an integration and synthesis of existing knowledge of mountain rivers. Although it is not feasible to cite every published study on all aspects of mountain rivers, the list of references at the end of the volume is unusually long because I wanted to be as inclusive as possible. I have avoided citing abstracts or unpublished theses or dissertations unless these are the only published material relevant to a particular topic and I have mostly avoided citing references that are not in English. Because this volume focuses primarily on physical processes, the discussions and reference lists for river chemistry and for aquatic and riparian ecology are not as complete as those for other topics treated in this volume. Topics of which we have particularly limited knowledge are highlighted throughout this synthesis and the concluding summary emphasizes aspects on which further research is particularly needed.

1.4. A Mountain River Described and Enumerated

Headwater regions encompass substantial spatial and temporal variations in geomorphic processes. The upstream extent of the channel network represents the transition from hillslope to channel processes, and downstream portions of channel networks in steep terrain include the transition from debris flows to fluvial processes, as well as substrate transitions such as bedrock to gravel and gravel to sand [*Sklar and Dietrich*, 1998; *Montgomery*, 1999; *May*, 2007; *Stock and Dietrich*, 2003]. Figure 1.7 presents a schematic overview of the components of mountain rivers discussed in this volume and, where possible, examples of equations developed to quantify these components. These equations are discussed in detail in succeeding portions of the text. Some of the equations are developed from a theoretical basis such as a balance of forces; others are empirical equations that may be of limited usefulness when extrapolated beyond the data from which they were developed. Whether theoretically or empirically based, quantitative statements of geomorphic process and form help to guide and focus continuing research by identifying processes or forms that we cannot yet adequately parameterize or that deviate from existing observations.

Building on *Schumm*'s [1977] zonation of a fluvial system into three basic zones of production, transfer and deposition, Figure 1.7 organizes mountain rivers into three primarily spatial zones, each of which is dominated by a distinct suite of geomorphic processes and landforms. The *colluvial-fluvial transition* area occupies the uppermost portion of the drainage basin, where sediment produced from bedrock weathering is

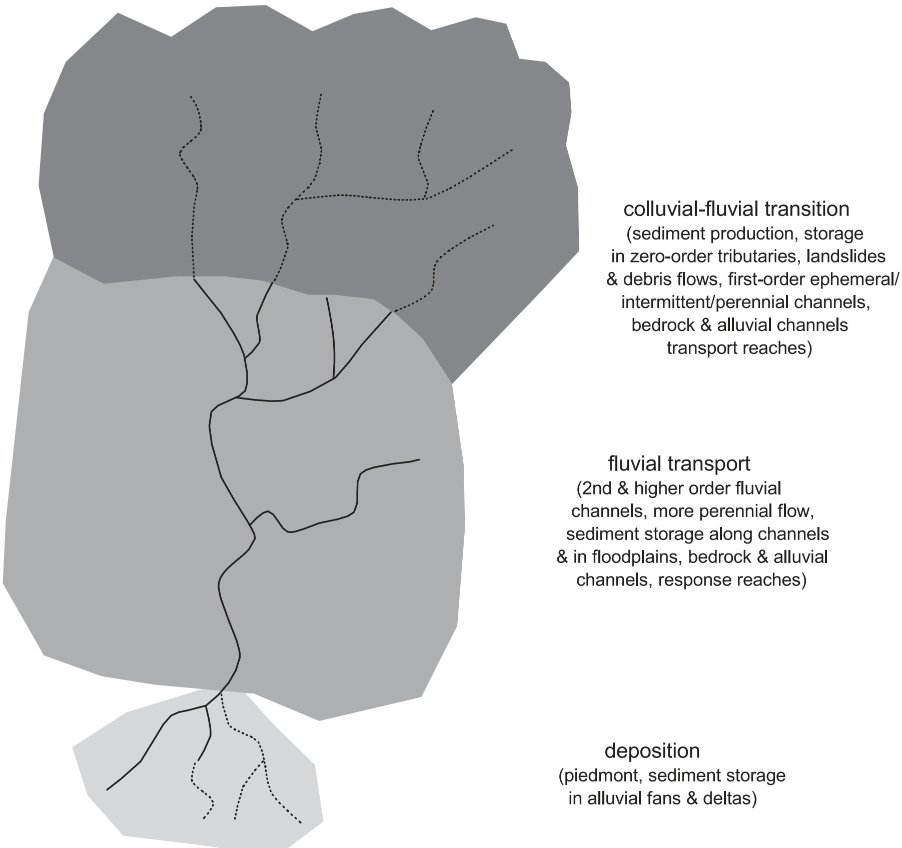

Figure 1.7 Highly stylized illustration of the three primary zones of a mountain drainage basin, followed by some of the equations used to describe process and form in each of those three zones. Variables used in each equation are defined in subsequent portions of the text.

moved downslope into channels by mass movements such as debris flows and landslides, and where fluvial channels begin. Channels in the *fluvial transport zone* in the middle section of the basin typically have progressively less direct hillslope influences as wider valley bottoms and floodplains buffer materials coming from hillslopes by creating at least temporary storage zones. Lower gradients, less lateral confinement, and/or lower velocity and discharge facilitate deposition along channels in the *depositional zone*, which is typically beyond the mountain front but may also occur in locally wider valleys.

This downstream zonation of mountain drainage basins reflects progressive downstream trends in discharge, gradient, grain size and other stream characteristics that numerous investigators have documented across a range of mountain drainage basins. Other variables that do not show progressive downstream trends also characterize mountain drainage basins; hydraulic resistance and magnitude of bedload transport, for example, do not necessarily change progressively downstream. Most variables show both progressive downstream trends and dominantly local (10^1-10^3 m) variation, depending on the spatial scale under consideration: Gradient and grain size both decrease downstream at the scale of a larger mountain watershed, but can exhibit local reversals as a result of spatial and temporal variation in driving factors such as lithology, tectonic uplift or hillslope stability and associated sediment inputs (Figure 1.8).

The local, and potentially longitudinally discontinuous, values of some parameters support the concept of *geomorphic process domains*. Spatial variability in geomorphic processes governs temporal patterns of disturbances that influence ecosystem structure and dynamics [*Montgomery*, 1999]. Mass transfer in the uppermost portions of hillslopes might be dominated by avalanches and rockfall, for example, whereas debris flows exert a greater influence in the middle portions of the catchment, and fluvial processes dominate the lower portions.

One way to conceptualize mountain river form and process is within the framework of driving forces versus substrate resistance. Channel configuration at any point along the drainage network fundamentally reflects the ratio of hydraulic driving forces to substrate resistance. *Hydraulic driving forces* reflect the movement of a volume of water from higher to lower elevation and thus incorporate discharge and channel gradient. The potential energy converted to kinetic energy via the downstream flow of water can be expended on overcoming external frictional resistance, internal frictional resistance, and sediment transport; the expenditure of energy thus incorporates channel configuration, sediment supply, and the erodibility of the channel boundaries. The ratio of driving forces and *substrate resistance* varies temporally as tectonic uplift alters landscape relief or storms passing over the watershed or land use alter water and sediment yield to the channel. The ratio also varies spatially as progressively greater contributing area increases discharge in the channel or as the channel flows from glaciated to unglaciated portions of the catchment. Some forms of spatial variation, such as downstream increase in discharge, are well documented from a range of field settings and are best described as linear or exponential functions. Some forms of spatial variation, such as the magnitude of external frictional resistance, may show analogous downstream trends, but lack extensive field documentation. Other forms of spatial variation, such as bank resistance created by riparian vegetation, are not adequately described by linear or exponential functions and appear to predominantly reflect local controls that do not vary progressively downstream. Figure 1.9 lists channel forms and processes and what is known about their downstream trends in mountain rivers. Although limited work to date suggests that hydraulic driving force as reflected in stream power peaks in the upper third to middle part of the basin [*Knighton*, 1999], substrate resistance is so spatially variable in mountain drainage basins that it precludes generalizations. It may thus be more useful to apply the ratio of driving force to substrate resistance at the local scale rather than at the basin scale.

Parameter	Downstream Trend	Documentation
discharge (Q)	exponential increase	strong
gradient (S)	exponential decrease	strong
valley geometry	highly variable	limited
sediment supply	highly variable	limited
external resistance (f)	declines downstream	limited
total stream power	peaks at mid-basin	limited
suspended sediment	highly variable	limited
bedload transport	highly variable	moderate
bedforms	progressive change with S	strong
sinuosity	highly variable	limited
channel lateral mobility	highly variable	limited
bank resistance from riparian vegetation	highly variable	limited
instream wood	highly variable	limited

Figure 1.9 Downstream trends in selected parameters for mountain rivers and relative documentation (with progressively less documentation from strong through moderate to limited) of these trends based on field data from diverse settings.

Each of the very broad parameter categories outlined in Figure 1.9 is explored in greater detail in subsequent sections of this book, but Figure 1.9 provides a quick overview of our relative understanding of diverse patterns in mountain drainage basins. This figure also indicates how much work remains to be done.

1.4.1. North St. Vrain Creek, Colorado, USA

I use the specific example of North St. Vrain Creek in the Colorado Front Range, USA to further illustrate how individual parameters vary downstream or locally. I chose this watershed because it is one of the least altered by land uses in the region and because I have done much of my own research there. North St. Vrain Creek represents neither an exceptionally well-studied watershed nor a little known one; it falls somewhere between these extremes and in this respect represents many other mountainous drainages.

North St. Vrain Creek drains eastward from the Continental Divide (4050 m elevation) onto the Great Plains (1945 m elevation at the base of the mountains) and eventually joins the South Platte River (Figure 1.10A). The portion of the catchment within the mountains includes 250 km^2 of steep terrain underlain by Precambrian-age granites, gneiss, and schist [*Tweto*, 1979]. The Front Range has been relatively tectonically quiescent since the early Tertiary [*Crowley et al.*, 2002; *Anderson et al.*, 2006b]. Pleistocene valley glaciers extended down to approximately 2500 m elevation [*Madole et al.*, 1998]. Narrow, glaciated spines form the range crests at 4000 m elevation, below which lie widespread surfaces of low relief at 2300-3000 m elevation. Fluvial canyons are deeply incised into these low-relief surfaces [*Anderson et al.*,

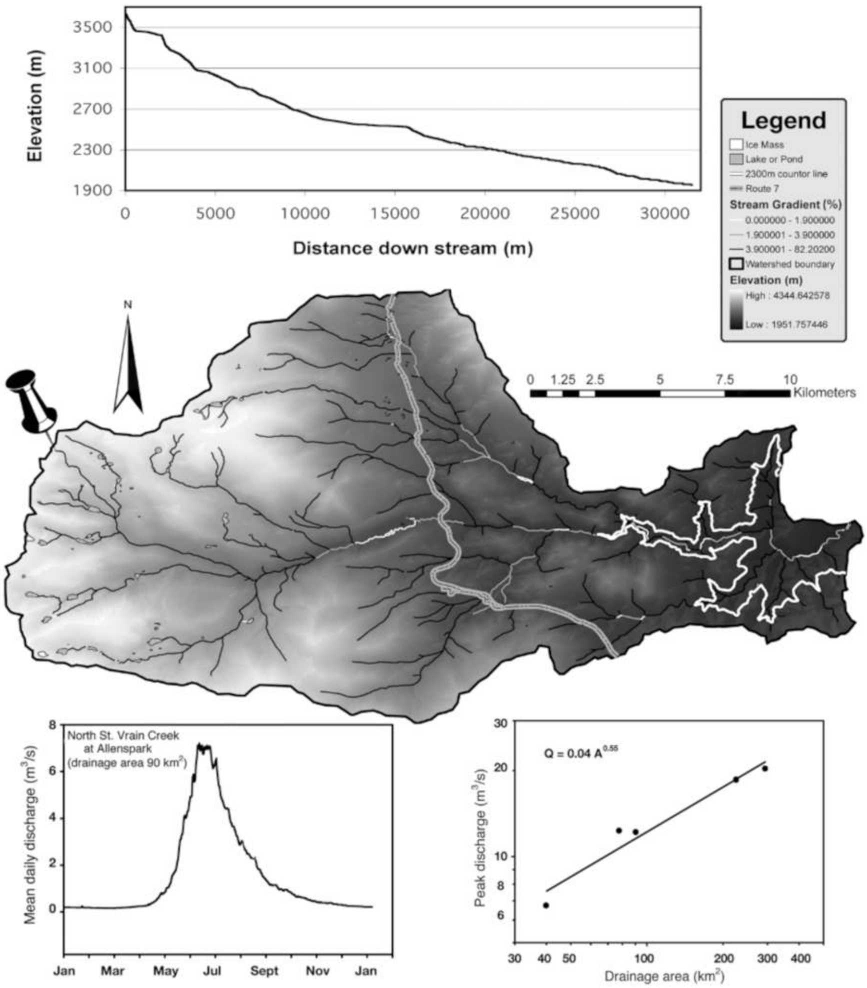

Figure 1.10 NSV map.

2006b]. Most bedrock outcrops in the region are densely jointed, and joint spacing and valley geometry correlate with the location of shear zones of Precambrian and Laramide age [*Abbott*, 1976]; wider, lower gradient portions of fluvial valleys typically correspond to more closely spaced joints and the location of shear zones [*Ehlen and Wohl*, 2002]. Variations in joint density, glacial history, and other large-scale controls create pronounced downstream variations in valley and channel geometry.

Snowmelt runoff dominates the annual hydrograph at all elevations within the catchment, producing a sustained May-June peak. On average, 85% of the annual flow

Parameter	Downstream Trend and Notes
discharge (Q)	exponential increase; as illustrated in Figure 1.10(A), peak annual discharge varies with drainage area with an exponent of 0.55, based on stream gage records covering multiple years at five sites with unregulated flow in and near the North St. Vrain catchment
gradient (S)	exponential decrease; as illustrated in Figure 1.10(A), longitudinal variation in stream gradient is readily obtained from 10-m DEM coverage of the catchment; spatial variation reflects primarily Pleistocene glacial history
valley geometry	highly variable; valley geometry can be directly estimated from 10-m DEMs via metrics such as connectedness (lateral distance between channel and base of valley wall) and entrenchment (ratio of channel width to valley width) or indirectly estimated from stream gradient on 10-m DEMs; Figure 1.10(A) illustrates spatial variation in stream gradient within the catchment, and the steepest gradient segments correspond to relatively deep, narrow valleys(< 50 m wide valley bottom), the moderate gradient segments to valleys of intermediate width and depth, and the lowest gradient segments to glacial troughs and broad valleys (> 50 m wide valley bottom) with meadows and wetlands
sediment supply	highly variable; little documentation in the North St. Vrain catchment, but volume and frequency likely vary with valley geometry, with coarser grained sediment episodically entering channels in mass movements along steep, narrow valley segments
external resistance (f)	declines downstream; limited documentation indicates that, as *S* decreases downstream, *f* also decreases (Wohl et al., 2004)
total stream power	peaks at mid-basin, as predicted by Knighton (1999), although values of stream power display a substantial amount of scatter rather than following smoothly ascending or descending trends (Wohl et al., 2004)
suspended sediment	highly variable; limited documentation indicates that suspended sediment increases during the annual snowmelt peak flow and following disturbances such as wildfire or debris flows
bedload transport	highly variable; limited documentation indicates increasing bedload transport in slightly finer grained channel segments downstream
bedforms	progressive change with *S*; spatial distribution of cascade, step-pool, plane-bed, and pool-riffle segments correlates well with *S* and can thus be predicted using 10-m DEM data (Wohl et al., 2004, 2007)
sinuosity	highly variable; like valley geometry, this correlates with *S* and can thus be indirectly estimated from 10-m DEM data, with high gradient corresponding to straight channels and lower stream gradients corresponding to greater sinuosity
channel lateral mobility	highly variable; as with sinuosity, this correlates with stream gradient and can be indirectly estimated from 10-m DEM data; steeper channels have lower lateral mobility than channels of lower gradient
bank resistance from riparian vegetation	highly variable; type of riparian vegetation varies with elevation and with valley geometry and can thus be indirectly estimated from 10-m DEM data; lower gradient channel segments flowing through relatively wide valleys are more likely to have dense herbaceous vegetation and willow (*Salix*) communities in relatively wide bands along the channel, whereas steep channel segments have limited riparian communities dominated by coniferous trees (Polvi, 2009)
instream wood	highly variable; limited documentation (e.g., Wohl and Jaeger, 2009; Wohl and Cadol, in press) indicates that higher wood loads and more frequent channel-spanning jams correspond to lower gradient stream segments

Figure 1.10 (continued)

occurs between May and September. Elevations below 2300 m also experience flash floods caused by summer convective storms. Rivers above this elevation have unit discharges of ~ 1 $m^3/s/km^2$, whereas rivers below 2300 m can have unit discharges of 40 $m^3/s/km^2$ [*Jarrett*, 1989]. Climate in the Front Range varies with elevation. Mean annual temperature varies from 1°C at the highest elevations to 11°C at the base of the range. Mean annual precipitation decreases from approximately 100 cm at the highest elevations to 36 cm at the mountain front, and the percentage of precipitation falling as snow also decreases with elevation.

Vegetation communities also vary with elevation, from alpine tundra above 3400 m, through subalpine spruce-fir forest, and montane pine forest below 2700 m [*Veblen and Donnegan*, 2005]. Wildfire and insect outbreaks are the most important forest disturbances in terms of extent, severity, and frequency. Three general types of historic fire regimes present in the catchment are: (i) infrequent, high-severity fires that kill all canopy trees over areas of hundreds to thousands of hectares and recur at intervals greater than 100 years in the subalpine zone; (ii) a complex pattern of low- and high-severity fires that burn areas of approximately 100 ha and recur at intervals of 40 to 100 years in the middle and upper montane zone; and (iii) frequent, low-severity firest that burn mainly the ground surface over areas of approximately 100 ha at intervals of 5-30 years in the lower montane zone [*Veblen and Donnegan*, 2005].

Beaver were trapped along the channels of the watershed starting in the early 19th century; the creek is named for French fur trapper Ceran St. Vrain. Although beaver have gradually recolonized the watershed, their populations are smaller than prior to trapping [*Wohl*, 2001]. The watershed is bisected by a two-lane highway; portions of the catchment upstream are largely in Rocky Mountain National Park and the mountainous portion downstream is largely in the Roosevelt National Forest. Flow in the creek is regulated starting at the base of the mountains. The information summarized in Figure 1.10 is drawn primarily from *Thompson et al.* [1996, 1999], *Wohl et al.* [2004], *Flores et al.* [2006], *Polvi* [2009], *David et al.* [2010], and *Wohl and Cadol* [in press]; with the exception of *David et al.* [2010], which is based on data collected in nearby drainages, these studies were conducted within the North St. Vrain catchment. Figure 1.10B reiterates Figure 1.9 with respect to the North St. Vrain catchment.

My research on North St. Vrain Creek and other mountainous catchments around the world has led me to conceptualize form and process in mountain rivers as illustrated in Figure 1.11. In this figure reach-scale gradient assumes primary importance. Gradient at channel lengths of 10^1-10^3 m can be a quasi-independent variable when the river does not have sufficient energy to create a smoothly concave longitudinal profile as a result of longitudinal variations in uplift rate, rock resistance, glacial history, sediment supply, or other parameters that influence gradient. Many other parameters correlate directly with reach-scale gradient (the solid arrows in Figure 1.11) and indirectly via intermediary parameters (the dashed arrows in Figure 1.11). Channel reaches of lower gradient, for example, correlate with wider valley bottoms or lower levels of connectedness (average distance from the channel edge to the valley edge) and higher values of entrenchment (ratio of valley width to channel width) [*Polvi*, 2009]. Wider valley

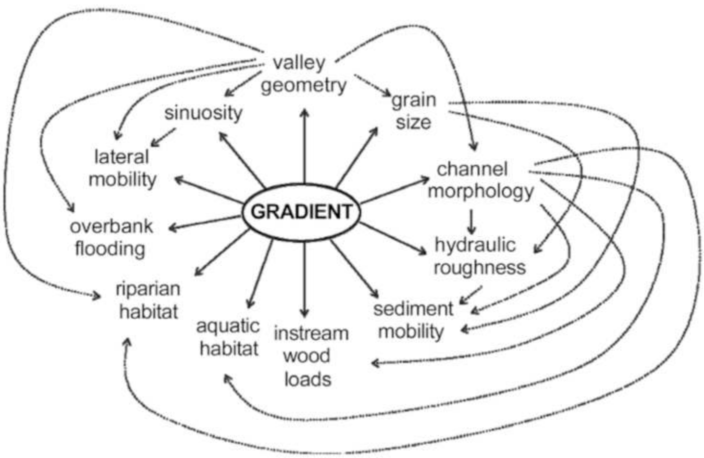

Figure 1.11 Schematic illustration of the correlations among variables along a mountain river. Reach-scale gradient assumes primary importance in this diagram because so many other variables directly or indirectly correlate with gradient, and because gradient is readily obtained from topographic data such as digital elevation models.

bottoms in turn correlate with greater sinuosity, lateral channel mobility and overbank flooding, riparian habitat associated with greater inundation and higher water tables, finer grain sizes in the streambed, and channel morphology such as pool-riffle or dune-ripple [*Montgomery and Buffington*, 1997; *Wohl et al.*, 2007; *Polvi*, 2009]. These channel morphologies associated with lower gradient have lower levels of hydraulic resistance [Darcy-Weisbach *f* or Manning's *n* coefficients; *Wohl et al.*, 2004; *David et al.*, 2010], greater sediment mobility, larger instream wood loads [*Morris et al.*, 2010; *Wohl and Cadol*, 2010], and greater pool volume than high-gradient channels.

Most mountainous catchments around the world now have some level of topographic data available, allowing reach-scale gradient to be quantified at varying degrees of spatial resolution. The correlations between reach-scale gradient and a wide variety of other parameters thus provides an entry point for understanding at least relative spatial variations in multiple parameters within a catchment. Field calibration of these relations can of course improve the ability to specify the degree of variation within variables such as grain size or instream wood load with respect to gradient.

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2. MOUNTAIN DRAINAGE BASINS

This chapter begins with a brief overview of regional-scale interactions among climate, tectonics, and erosion as these influence the form and process of mountain drainage basins. Section 2 discusses hillslope process and form, including the production of sediment from bedrock and the distribution of that sediment along hillslopes through mass movements and diffusive sediment transport. Section 3 returns to climate in the context of understanding the types of precipitation that influence mountainous catchments, the spatial and temporal heterogeneities in the distribution of precipitation across high-relief terrain, and the surface and subsurface paths that water follows as it moves down hillslopes and into channels. An overview of the processes that influence how and where channels initiate (section 4) is followed by a review of different morphometric basin parameters and how these interact with the movement of water to influence hydrographs (section 5). Section 6 covers characteristics of valley morphology in mountain drainages. Sections 7 and 8 address longitudinal profiles, with an emphasis on processes and modeling of bedrock channel segments. Section 9 covers paleo-longitudinal profiles as preserved in terraces, and section 10 addresses alluvial fans, which can occur throughout mountainous drainages wherever wider valley bottoms and lower gradients facilitate persistent deposition. The common organizing framework of this chapter is that of the larger-scale variables and processes that influence the geomorphology of a drainage basin and how these interact with water falling as precipitation and then moving down hillslopes, with bedrock weathering and sediment production and transfer, and with tectonics. It is useful to begin with a brief review of how mountainous topography originates.

Mountainous regions are produced by four general types of deformation; folding, volcanism, fault block uplift, and vertical uplift [*Press and Siever*, 1986]. Folded mountains result from lateral compression, usually at the convergent boundary between two tectonic plates. Examples include the Appalachian Mountains of the eastern United States, the Alps of southern France, the Urals at the boundary between Europe and Asia, and the Transantarctic Mountains. The topography of folded mountains may be controlled by differential weathering of the lithologies exposed by uplift, with more resistant lithologies forming steeper slopes.

Volcanic mountains generally form at a divergent or convergent plate boundary, or at an intraplate hot spot such as the Hawaiian Islands. Examples of volcanic mountain

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ranges include the highlands of New Guinea, the North Island of New Zealand, the Japan Alps, and the Cascade Range of the northwestern United States. The geologic controls on volcanic islands are a function of the style of eruption and chemical composition of the lava.

Fault block uplift tends to produce mountain ranges with one very steep side parallel to the fault and a gentler side that does not have an active fault, as in the Teton Range of Wyoming, USA. Mountains produced by vertical uplift have faults parallel to both sides of the range, as in the Front Range of Colorado, USA.

Most of the world's major mountain belts include folded, faulted, and volcanic regions, as well as igneous plutons. The Himalaya mountain ranges, for example, include the folded and thrust-faulted zone of the Siwalik hills at the southern margin, and thrust-faulted and intruded rocks in the Middle Himalaya and the High Himalaya [*Bridges*, 1990]. The Andes Mountains of western South America include high volcanic peaks, folded belts, igneous intrusions, and extensive faults [*Bridges*, 1990].

2.1. Mountain Rivers and Tectonics

Early work on mountain rivers and tectonics emphasized the effect of mountainous topography and active uplift on drainage networks, noting that channels do not always follow existing slopes. Working in the western United States, *Powell* [1875, 1876] described both antecedent drainage networks in which pre-existing channels had maintained their spatial arrangement while the underlying landmass was deformed and uplifted, and superimposed channels which had incised downward to a buried structure. Either scenario could result in a river flowing through or across a mountain range (transverse drainage), rather than channels draining from the crest of the range downward to the neighboring lowlands (Figure 2.1). Thus, the drainage network was not a consequence of present topography. More recent studies discuss transverse drainage associated with the Coastal Range of Taiwan [*Lundberg and Dorsey*, 1990], isostatic uplift of the Apuseni Mountains, Hungary [*Thamo-Bozso and Kercsmar*, 2002], and the Betic Cordillera of southeastern Spain [*Stokes and Mather*, 2003]. *Humphrey and Konrad* [2000] argue that river sediment flux and tectonic uplift rate are the most important variables in determining whether a river will incise through or divert around an evolving bedrock uplift. Flume experiments testing the four general mechanisms proposed for transverse drainage (antecedence, superimposition, overflow, piracy) supported the ability of all of the mechanisms except superimposition to produce transverse drainage [*Douglass and Schmeeckle*, 2007]. Investigators continue to use drainage pattern, as well as terrace and channel geometry, as indicators of tectonic effects on rivers [*Schumm et al.*, 2000].

For cratonic or passive-margin settings, *Young* [1989] argues that although the alignment and form of individual valleys may reflect surface variations in lithology and structure, deeper crustal features control drainage patterns at the sub-continental scale. These deeper crustal features may only be discernible using remote-sensing technology to detect patterns such as Bouguer gravity anomalies [*Young*, 1989]. *Brookfield*

[1998] describes the importance of tectonic history in creating three regionally distinctive patterns among the major river systems of Asia. Differential compression and right-lateral shear produced highlands from which rivers of the Helmand-Farah system drain into arid depressions. Differential shear and clockwise rotation between the compressing Tibetan plateau and Southeast Asia produced large sigmoidal bends in widely separated rivers including the Chang Jiang, Mekong, and Salween. Southward thrusting and massive frontal erosion of the Himalaya caused progressive truncation of rivers including the Tsangpo, Indus, and Sutlej on the plateau [*Brookfield*, 1998]. Drawing on flume and field studies, *Schumm et al.* [2000] summarize how different alluvial channel morphologies respond to various types of tectonic deformation, with channel response partly governed by proximity to the axis of deformation. Bedrock rivers can show analogous spatially variable responses; following the 1999 Chi-Chi earthquake in Taiwan, river incision intensified near the fault scarp, whereas landslides induced by the earthquake mantle the river bed with sediment and impede bedrock incision in reaches distal to the fault [*Yanites et al.*, 2010a].

Recent work tends to emphasize more complex interactions in which rivers influence, as well as respond to, mountainous topography. Investigators have proposed for decades that arrangement and incision of valley networks can affect mountain relief and elevation. *Corbel* [1959] represents an early study indicating that rivers remove up to five times more sediment per unit area from mountain basins than from lowland basins. This type of comparison emphasizes the role of mountain rivers as conveyors of sediment from upland regions. Starting in the 1990s, investigators took this insight to the next level and proposed that the pattern of river incision can affect crustal structure in mountain belts by changing the distribution of stress in the crust [*Hoffman and Grotzinger*, 1993; *Beaumont and Quinlan*, 1994].

The effect of interactions among tectonic forces, climate and erosive processes in shaping mountainous topography has been the subject of much recent attention [*Koons*, 2009]. *Raymo et al.* [1988] and *Raymo and Ruddiman* [1992] propose that accelerated tectonic uplift increase weathering. Subsequent field studies support this assertion [*Carey et al.*, 2006]. Because chemical weathering is an important sink for CO_2, the removal of CO_2 from the atmosphere could have produced lower temperatures during the past 40 million years, facilitating glacial cycles. Glacial erosion may then have accelerated uplift and mountain building as removal of mass facilitated isostatic uplift [*Molnar and England*, 1990; *Hallet et al.*, 1996]. This is the so-called *glacial buzzsaw effect*; glacial erosion rapidly removes mass raised above the altitude of the local glacier equilibrium line [*Brozovic et al.*, 1997; *Whipple et al.*, 1999; *Montgomery et al.*, 2001; *Brocklehurst and Whipple*, 2002; *Mitchell and Montgomery*, 2006; *Naylor and Gabet*, 2007]. Evidence for the glacial buzzsaw effect comes primarily from field studies in areas with large glaciers. Investigations in regions with smaller alpine glaciers [*Foster et al.*, 2008], as well as tests using numerical models [*Tomkin and Braun*, 2002; *Tomkin*, 2007], suggest that the hypothesized effect is complicated by other factors such as whether the glacier base is frozen, how fracture density influences the pace of glacial erosion [*Dühnforth et al.*, 2010], and that there is a minimum magnitude of glacier erosion below which insufficient rock mass is removed to isostatically raise summit elevations [*Foster et al.*, 2010]. *Alley et al.* [2003] note that the long profiles of beds of highly erosive glaciers tend towards steady-state angles related to the overlying ice-air surface slopes, beyond which additional subglacial deepening depends on non-glacial processes. This suggests a limit to the erosion conceptualized in the glacial buzzsaw effect. Quantification of glacial erosion patterns using cosmogenic radionuclides [e.g., *Li et al.*, 2005] has substantially enhanced the ability to test hypotheses such as the glacial buzzsaw.

Hales and Roering [2009] propose a frost buzzsaw. Noting that rockfall controls erosion in the eastern Southern Alps of New Zealand, and that frost cracking is the primary rockfall mechanism, they correlate climate and elevational controls on frost cracking intensity with the elevation of the highest peaks and suggest that the height of these peaks is limited by a frost buzzsaw.

A further complication of the climate-tectonics-erosion interactions is that formation of large ice sheets results in substantial, albedo-induced cooling of the Earth's atmosphere. *Kuhle* [2007] describes a scenario in which development of a Tibetan ice sheet occurred as the Tibetan Plateau was lifted above snowline. Albedo-induced cooling from the ice sheet disrupted the summer monsoon circulation and facilitated the global depression of snowline and development of other ice sheets. Glacial-isostatic lowering of Tibet caused melting of the ice sheet during a period of positive radiation anomalies, which triggered an interglacial period. Glacial-isostatic rebound then lifted the plateau above snowline, triggering the next glacial period [*Kuhle*, 2007]. These alternating episodes of glacial advance and retreat influenced river dynamics by changing the supply of meltwater and sediment [*Rahaman et al.*, 2009] and in other ways: Steep fluvial knickpoints formed at the southeastern margin of the Tibetan plateau should erode rapidly back into the plateau, but *Korup and Montgomery* [2008] propose that the plateau edge has been preserved because numerous moraine dams on major rivers impede bedrock river incision.

Apart from large-scale glacial erosion, various field and modeling studies indicate that at smaller, regional scales, spatial gradients in the climate forcing that drives erosion can influence the development of geologic structures [*Hoffman and Grotzinger*, 1993; *Willett et al.*, 1993, 2001; *Avouac and Burov*, 1996; *Horton*, 1999; *Willett*, 1999]. This is expressed in the *tectonic aneurysm model* [*Zeitler et al.*, 2001] (Figure 2.2) in which local rheological variations arise in a deforming orogen as a result of deep and rapid incision. The crust weakens as the strong upper crust is locally stripped from above by erosion. The local geotherm is then steepened from below by a focused rapid uplift of hot rock. If efficient erosion continues, a positive feedback keeps material flowing into this weakened zone, which maintains local elevation and relief [*Koons et al.*, 2002; *A. L. Booth et al.*, 2009; *A. M. Booth et al.*, 2009]. These ideas led to numerous studies of the interactions among uplift, river incision, and climate fluctuations, and the extent and magnitude of glacial versus nonglacial erosion [*Harbor and Warburton*, 1993; *Burbank et al.*, 1996; *Hallet et al.*, 1996; *Whipple and*

Tucker, 1999; *Galy and France-Lanord*, 2001; *Lavé and Avouac*, 2001; *Dadson et al.*, 2003; *Snyder et al.*, 2003; *Korup et al.*, 2005; *Anderson et al.*, 2006a; *Barros et al.*, 2006; *Schaller and Ehlers*, 2006; *Smith*, 2006; *Anders et al.*, 2010; *Binnie et al.*, 2010; *Pelletier et al.*, 2010]. The result of this work is consensus that climate, erosion, and tectonics are strongly coupled through large-scale feedback systems [*Montgomery*, 2004a]. Gradients in climate [*Bookhagen and Burbank*, 2010] and tectonic forcing influence erosional intensity, which governs the development of topography, which in turn influences climate and tectonics. These interactions can be expressed in steady-state longitudinal river profiles along which different degrees of curvature reflect orographically-induced variations in precipitation [*Roe et al.*, 2002].

Several other studies also address the effect of valley incision on mountain topography. Modeling the effect of isostatically compensated valley incision on the elevation of mountain peaks, *Montgomery* [1994b] finds that this compensation could account for at most 5-10% of the present elevation of mountain peaks in the central Sierra Nevada of California, USA and the Tibetan Plateau and *Mitchell et al.* [2009] estimate that it adds <25% of height to peaks in the Cascade Range of Washington, USA. Such compensation could account for 20-30% of the present elevation of peaks in the Himalaya, however. *Montgomery and Stolar* [2006] propose that Himalayan river anticlines (major Himalayan rivers flow parallel to and down the axis of anticlines oriented transverse to the primary structural grain of the range) are the consequences of focused rock uplift in response to significantly larger net erosion along major rivers than in surrounding regions. Even in areas with less rapidly changing baselevel, such as the highlands of eastern Australia, the headward erosion of river gorges is the most important process denuding these highlands during the last 30 million years [*Nott et al.*, 1996]. Stream erosion of new drainage basins in extensional mountain ranges of the southwestern United States exceeds hillslope retreat, leading to elevation of summit plateaus [*Harbor*, 1997].

Recent developments in geochronology facilitate estimation of regional erosion rates in mountains. Techniques include use of cosmogenic ^{10}Be in sediment carried by streams (because minerals at depth are shielded from cosmic rays, ^{10}Be concentration when minerals reach the surface indirectly records their exhumation rate) [*Kirchner et al.*, 2001] and low-temperature thermochronologic data in which spatial patterns of mineral cooling ages are related to the rates at which buried rocks move toward the surface [*Safran*, 2003; *Schildgen et al.*, 2009]. Nuclides such as ^{10}Be or ^{26}Al are produced when secondary cosmic rays interact with the uppermost layer of the Earth's surface. The nuclides are produced within a characteristic depth scale of about 1 m, so that measured concentrations in rock exposures record erosion rates at that point and concentrations in sediments record an integrated denudation history while material passed through this depth interval [*Bierman and Nichols*, 2004]. Depending on the denudation rate, the resulting integration time scales are 10^3-10^5 years, providing a long-term estimate of denudation [*von Blanckenburg*, 2005]. Interpretation of erosional history from cosmogenic isotope ages requires some knowledge of geomorphic processes: Using a numerical simulation of cosmogenic nuclide production and distribution, *Niemi et al.* [2005] find that larger catchment areas must be sampled to accurately evaluate long-term erosion rates as the frequency of landsliding increases, and that sediment sampling is more appropriate than sampling bedrock surfaces in regions dominated by mass movement processes.

Luminescence dating is also applied to hillslope sediments. Luminescence techniques utilize the ability of some natural crystalline materials to store energy released by background radioactive decay over long periods. The stored energy can be released by stimulation by heat (thermoluminescence) or light (optically stimulated luminescence). *Fuchs and Lang* [2009] review luminescence dating of hillslope deposits and *Madsen and Murray* [2009] review optically stimulated luminescence dating of sediments <1,000 years in age.

Remote sensing imagery, digital elevation models, and geomorphometry – the quantitative description and analysis of geometric-topologic characteristics of the landscape – have been key to quantifying parameters such as relief, glacial and fluvial dissection, and hillslope and valley geometry [*Bishop et al.*, 2002, 2003; *Misukoshi and Aniya*, 2002; *Montgomery*, 2004a] and to modeling feedbacks among tectonic forcing, erosion, isostatic rebound, and rock exhumation [*Montgomery*, 2001b]. *Bishop et al.* [2004] review remote-sensing techniques and *Rasemann et al.* [2004] review geomorphometric variables and analysis in mountain environments using GIS. *Hengl and Reuter* [2009] provide a comprehensive overview of geomorphometry.

Physical experiments also provide insight into interactions among uplift and erosion at spatial scales from a single channel segment to entire watersheds [*Schumm et al.*, 1987; *Ouchi*, 2004]. *Lague et al.* [2003] use physical experiments to investigate landscape response to uplift and erosion and find that topography always reaches a steady state, with a mean elevation linearly dependent on uplift rate. Their steady-state surfaces exhibit a well-defined slope-area power law with a constant exponent of -0.12, a result consistent with a stream power erosion model (equation 2.31) that includes a non-negligible threshold for particle detachment.

Because bedrock channel incision can exert such an important control on hillslope stability and regional rates of uplift and erosion, many investigators have used river morphology to interpret the scale, magnitude, and timing of rock uplift, for which other evidence is often limited. Although measures such as mountain front sinuosity can be used [*Pérez-Peña et al.*, 2010], river morphology across drainage basins or regions is typically characterized in terms of gradient and longitudinal profile, which are readily obtained from digital elevation models (DEMs) [*Snyder et al.*, 2000; *Duvall et al.*, 2004; *Font et al.*, 2010]. Because longitudinal profile irregularities can reflect downstream variations in lithology and erodibility [*Valla et al.*, 2010] and glacial history [*Hobley et al.*, 2010], as well as rock uplift, profiles must be carefully interpreted within a geologic-geomorphic context. In addition to longitudinal profile, longitudinal variations in the width of bedrock channels indicate differential uplift [*Whittaker et al.*, 2007a, 2007b; *Attal et al.*, 2008; *Yanites et al.*, 2010b]. Numerical derivations of scaling relations for bedrock channel width, w, drainage area, A, gradient, S, and discharge, Q, have been derived from: flow resistance equations and mass conservation principles, producing $w \sim Q^{0.38}$ and $w \sim S^{-0.2}$ [*Finnegan et al.*, 2005]; assumptions that erosion rate scales with local shear stress, which results in $w \sim Q^{0.4}$ and $w \sim S^{-0.2}$ [*Wobus et al.*, 2006a]; and minimization of potential energy, for which $w \sim A^{0.5}$ [*Turowski et al.*, 2007]. Limited field investigations of downstream hydraulic geometry in bedrock channels tend to follow the lead of *Montgomery and Gran* [2001] in substituting A for Q, although this introduces uncertainties associated with hydroclimatic variation along a channel or drainage basin [*Flores et al.*, 2006]. The exponent in field-based $w \sim A^b$ relations has varied from 0.32 [*Montgomery and Gran*, 2001] to 0.55 [*van der Beek and Bishop*, 2003]. Compiling a large field data set from many regions, *Wohl and David* [2008] propose that scaling relations are relatively consistent among bedrock and alluvial channels, such that $w \sim A^{0.3}$ and $w \sim Q^{0.5}$, although bedrock channels tend to be consistently narrower than alluvial channels for a given drainage area. Because changes in rock erodibility or uplift rate can alter downstream scaling relations [*Wohl and Merritt*, 2001; *Cowie et al.*, 2006; *Jansen*, 2006], unexpected deviations in channel width from *w-A* or *w-Q* relations can be used to infer uplift.

Field data and numerical simulations indicate that adjustments to width and gradient as a result of increasing substrate resistance or uplift are typically tightly coupled [*Whipple*, 2004; *Stark*, 2006]. Holding substrate erodibility constant, gradient increases and width declines on rivers with higher uplift rates in southern California, USA [*Duvall et al.*, 2004]. Using physical experiments, *Turoswki et al.* [2006] demonstrate that, as uplift rate increases, channel width and cross-sectional area decrease and velocity increases approximately linearly. *Whittaker et al.* [2007] show that traditional hydraulic scaling laws break down along bedrock channels crossing an active fault in the central Italian Apennines; channel widths become decoupled from drainage area upstream of the fault and values of unit stream power are approximately four times those predicted by scaling relations. Similarly, *Amos and Burbank* [2007] find that small rivers crossing growing folds in New Zealand respond with channel narrowing up to some threshold of differential uplift, beyond which channel gradient also increases.

Regional rates of uplift can be compared to regional rates of denudation as an index of the efficiency of mountain hillslope and channel processes. Mountainous topography results from the imbalance between uplift caused by tectonics and denudation by tectonic (extensional faulting) or surface (glacial, hillslope, and fluvial) processes [*Burbank et al.*, 1996]. Early estimates of regional denudation rates came primarily from sediment yields averaged over decades or longer. *In-situ* produced cosmogenic nuclides are now widely used to infer denudation rates [*Vance et al.*, 2003; *Schaller et al.*, 2004].

Rates of both uplift and denudation can have substantial spatial and temporal variability. *Leopold et al.* [1964] use the 629,520 km^2 basin of the Colorado River in the southwestern U.S. as an example of spatial variability in denudation rate, as estimated from suspended sediment load expressed in centimeters derived from the drainage basin per unit of time. Denudation rates range from approximately 0.4 to 17 cm/ky and show fairly strong correlation with climate [*Leopold et al.*, 1964].

Oguchi [1996b] compares Holocene and contemporary denudation rates for a series of river basins in central Japan and finds that contemporary rates are up to three times higher than Holocene rates. Despite this variability, regional rates of the type listed in Table 2.1 may still be useful indicators of relative efficiency of weathering and erosion in various regions. (All tables are available on the CD-ROM that accompanies the book.) Both climate and relief strongly influence denudation rate.

Published rates of bedrock channel incision vary from 5 to 10,000 mm/ky, with the highest rates occurring in regions of tectonic uplift [*Wohl et al.*, 1994a; *Wohl*, 1998]. Most of these channel incision rates are long-term (Quaternary) averages for third-order or higher channels, but they indicate that tectonic uplift corresponds with increased transport ability and channel incision in mountainous regions, regardless of climate or lithology.

In summary, work within the past decade demonstrates that mountain rivers do not simply respond to tectonically controlled gradient; rather, spatially and temporally variable interactions among uplift, climate, and fluvial erosion allow rivers to both respond to and influence uplift, elevation, relief, and the distribution of mass across a landscape. These interactions are exemplified by the tectonic aneurysm model in which deep and rapid incision alters crustal properties such that a positive feedback develops and maintains local elevation and relief. Spatial variations in the gradient and width of bedrock rivers can reflect spatial variations in tectonics. Geochronological advances that facilitate quantification of uplift and denudation rates, mapping and modeling of river longitudinal profiles, and numerical and physical models of diverse landscape processes, all enhance our understanding of the interactions between mountain rivers and tectonics. However, as *Tucker* [2009] notes, there remains a pressing need to identify natural experiments in landscape evolution in which only one element varies significantly and for which the driving forces, initial conditions and/or boundary conditions are well constrained.

2.2. Hillslopes

Schumm [1977] divides the fluvial system into an upstream zone that serves as the primary sediment source for a drainage basin, a middle transfer zone, and a downstream zone that is primarily depositional, analogous to Figure 1.7. Mountain rivers occupy the upstream sediment-source zone of a drainage basin, and primarily reflect the controls of climate, geology, and land use as these influence water and sediment yield to the channel, and channel-boundary resistance. Geology is here taken to include lithology, structure, and tectonic regime. These characteristics will, in combination with climate, determine rate and manner of weathering, and thus slope morphology and processes of water and sediment movement.

This section explores form and process on hillslopes in some detail because hillslopes exert such strong influences on form and process in mountain rivers. The first subsection discusses how tectonics, lithology and climate influence weathering and erosion on slopes. This leads, in the second subsection, to the concepts of *steady-*