9.28 Bedrock Rivers

KX Whipple and RA DiBiase, Arizona State University, Tempe, AZ, USA
BT Crosby, Idaho State University, Pocatello, ID, USA

Abstract

Bedrock rivers play a critical role in landscape evolution – cutting canyons, creating relief, and driving much of landscape response to changes in climate and tectonics. The bed and banks of bedrock rivers are sporadically rocky but are commonly covered by a thin and fairly continuous layer of alluvium, though bedrock is everywhere close to the surface. Flow hydraulics and channel morphology in bedrock rivers have much in common with coarse-bed alluvial rivers. Interestingly, the width of bedrock channels is similar to alluvial rivers with the same discharge and follows the same scaling with drainage area, suggesting similar controls on channel width despite the difference in substrate strength. Erosion of bedrock in rivers is accomplished by a suite of interacting processes, including abrasion, plucking, cavitation, debris-flow scour, and weathering, and is strongly modulated by river sediment load. Recent models of river incision into bedrock incorporate the dual role of the sediment load as both tools and cover and highlight the importance of discharge variability in the erosion process. Data relating channel steepness, landscape relief, and erosion rate are critical for testing and refining these river incision models, and these increasingly well-established relationships now provide effective tools for diagnosing both spatial and temporal influences of climate and tectonics on landscape evolution. Furthermore, the style and distribution of knickpoints in transient landscapes encodes additional information about climatic and tectonic histories and allows for further discrimination among river incision models that produce similar steady-state forms.

Whipple, K.X., DiBiase, R.A., Crosby, B.T., 2013. Bedrock rivers. In: Shroder, J. (Editor in Chief), Wohl, E. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, vol. 9, Fluvial Geomorphology, pp. 550–573.

 http://dx.doi.org/10.1016/B978-0-12-374739-6.00254-2

9.28.1 Introduction

This chapter emphasizes the unique attributes and special role in landscape evolution of what are generally termed 'bedrock' rivers. River systems that have been described as bedrock rivers in the recent literature span a wide range of characteristics including drainage area, slope, bed state, and bed morphology. Thus, in this section, we include: (1) a clear definition of what is meant by a bedrock river and where these channel types occur; (2) a brief discussion of why bedrock rivers are important in the study of landscape evolution; and (3) a brief guide to how the subject matter in this chapter relates to other chapters in the volume.

9.28.1.1 Definition and Occurrence

The most direct and literal definition of bedrock rivers is that they are rock bound, that is, the bed and banks are largely composed of in-place bedrock. Such channels do occur, but are rare. Most commonly, rock-bound reaches are short and intermittent, comprising only a small fraction of the length of what are often considered bedrock rivers. Rather, it is generally observed that sediment cover in bedrock rivers is thin and patchy – rock outcrop in river bed and banks may be rare, but rock is everywhere close to the surface and may be frequently exposed during flood events or on decadal to centennial timescales (Benda and Dunne, 1997; Howard, 1998). Although most such rivers would perhaps be better termed mixed bedrock–alluvial rivers (Howard, 1998), researchers have settled on the simpler term bedrock river. A primary motivation for this is that what is really special about bedrock rivers is not that the nature of the bed and banks is distinct from alluvial rivers, but because bedrock rivers are commonly actively incising through in-place rock, they play a critical role in landscape evolution.

Accordingly, we adhere to a two-part definition of bedrock rivers. Bedrock rivers may satisfy either or both of the following conditions: (1) the long-term capacity of the river to transport bedload (Q_c) exceeds the long-term supply of bedload (Q_s), resulting in generally sediment-starved conditions, significant rock exposure in bed and banks, and only thin, patchy, and temporary alluvial cover (Montgomery et al., 1996); or (2) the river is, over the long term (millennial to geologic timescales), actively incising through in-place rock. Persistent incision over the long term implies that rock is everywhere near the surface even if a bedrock river has largely alluvial bed and banks, and in many respects has the flow, bed morphology, and sediment transport characteristics of an alluvial channel. Thus, bedrock rivers dominate in areas of net erosion and encompass most mountain rivers (Wohl, 2000a; Wohl and Merritt, 2008) (*see* Chapters 9.1 and 9.27) and also occur in areas with very low coarse sediment supply, even if river incision into rock is negligible or very slow.

9.28.1.2 Importance of Bedrock Rivers

Incision into rock is, simply put, what makes bedrock rivers special and what defines their unique role in landscape evolution. In unglaciated landscapes, bedrock rivers are ultimately responsible for driving landscape response to tectonic uplift, base-level fall, and much of landscape response to climate change. As such, an understanding of bedrock rivers is essential to the study of the potential interactions among climate, tectonics, topography, and erosion (e.g., Whipple, 2004, 2009; Willett, 1999) (*see* Chapter 9.1). Bedrock rivers are responsible for carving canyons and generating topographic relief. The longitudinal profiles of bedrock rivers dictate much of the three-dimensional (3D) architecture of unglaciated mountainous topography (Howard et al., 1994) (**Figure 1**). It is bedrock rivers that communicate signals of tectonics, climate change, and sea-level rise and fall throughout the landscape (Howard et al., 1994; Whipple and Tucker, 1999) and set the magnitude of changes in topographic relief produced by changes in tectonics or climate. Bedrock rivers set the lower boundary condition on all hillslopes within their catchments – dictating the rate of base-level fall experienced by each hillslope. The amount and caliber of sediment in a channel (and variability of both), however, is largely dictated by hillslope inputs and strongly influences not just channel slope, bed state, and morphology, but also the rate of river incision (e.g., Johnson and Whipple, 2007; Sklar and Dietrich, 1998; Sklar and Dietrich, 2004; Turowski et al., 2007). In this way, hillslopes and bedrock channels are strongly coupled and cannot be fully appreciated in isolation.

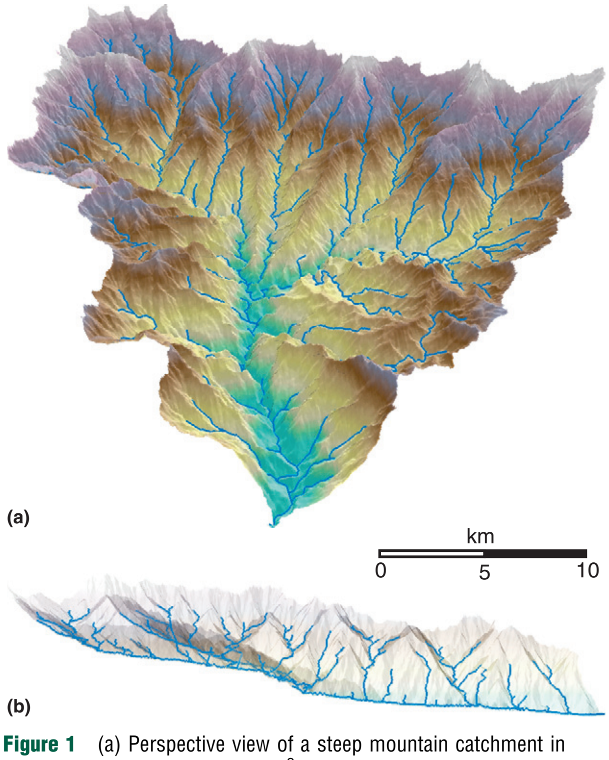


Figure 1 (a) Perspective view of a steep mountain catchment in Taiwan (Liwu catchment, 535 km^2 drainage area, 3650 m relief). Channel segments with upstream drainage area greater than 0.8 km^2 are highlighted in blue. (b) Side-view of this catchment with the topography made transparent to highlight the relation between local relief and the elevation drop on bedrock channels note the knickpoint at basin midpoint (see Section 9.28.4.3) (the extent of the network well described by Flint's Law – eqn [5], Section 9.28.4.1). A 2 $\times$ vertical exaggeration is used in both views.

9.28.1.3 Relation to Other Chapters in Volume 9

As noted above, bedrock rivers have much in common with alluvial rivers, are generally synonymous with mountain rivers (Wohl, 2000a), and encompass many steep, headwater channels (*see* Chapter 9.27). Because bedrock rivers, in fact, generally have thin and patchy to continuous alluvial cover, they are broadly similar to other rivers in terms of reach-scale resistance (*see* Chapter 9.5), bed entrainment (*see* Chapter 9.7), bedload transport (*see* Chapter 9.8), bedforms (*see* Chapter 9.10), and the role of large woody debris (*see* Chapter 9.11). Bedrock rivers commonly exhibit step-pool (*see* Chapter 9.20) or pool-riffle (*see* Chapter 9.21) morphologies (*see* Chapter 9.36 for channel-type classifications) and are similar to other rivers in terms of hydraulic geometry (*see* Chapter 9.18). Strath terraces (*see* Chapter 9.22) are characteristic of many bedrock rivers. Landslides are an important process in the interaction between bedrock rivers and hillslopes (*see* Chapter 9.15). Thus, many aspects of bedrock rivers have been discussed elsewhere in this volume. This chapter focuses exclusively on what is unique about bedrock rivers and their role in long-term landscape evolution and the interactions among climate, tectonics, topography, and erosion (*see* Chapter 9.1).

9.28.2 Flow Hydraulics and Channel Morphology

9.28.2.1 Overview of Flow Hydraulics

This section provides a brief overview of the flow hydraulics of bedrock rivers. As noted in our definition above (Section 9.28.1.1), bedrock rivers encompass not only rock-bound channels, but also what many refer to as mountain rivers – steep (>0.2%) channels characterized by spatially limited floodplains, a direct connection with hillslopes, and commonly flashy hydrographs (Wohl, 2000a; Wohl and Merritt, 2008). Thus, bedrock rivers share much in common with all steepland rivers. We provide a brief synopsis of flow hydraulics here, but more detailed discussions of bedforms and bed morphology (*see* Chapters 9.10, 9.20, and 9.21), flow resistance (*see* Chapter 9.5), and sediment transport (*see* Chapter 9.8) in mountain rivers can be found elsewhere in this volume.

Montgomery and Buffington (1997) categorized mountain channels into seven reach classifications, based primarily on bed forms, and found that in the Pacific Northwest, the different reach morphologies reflect specific combinations of flow conditions and channel geometry. Wohl and Merritt (2008) expanded this field data set to mountain channels worldwide, and evaluated how roughness, slope, width, grain size, and stream power vary as a function of the three most common morphologies in mountain rivers – pool riffle, plane bed, and step pool. Wohl and Merritt found that these channels tend to maximize flow resistance at the reach scale while minimizing variability in energy expenditure between reaches. Together, these studies have recognized predictable relationships between bed morphology and flow hydraulic characteristics that apply to most bedrock rivers (a brief summary is given in **Table 1**). Certain bed morphologies, however, are exclusive to bedrock channels (e.g., potholes, flutes, inner gorges/slots, and large steps/falls) (Richardson and Carling, 2005). In a study of step-pool sequences in Arizona, Wohl (2000b) found that bedrock step-pool reaches serve essentially the same function as their alluvial counterparts; bedrock pool length and depth vary as a function of slope and substrate resistance to minimize inter-reach variability of energy expenditure while maximizing intra-reach energy expenditure (Wohl and Merritt, 2008). Wallforms in slot canyons have also been shown to modulate energy expenditure similar to alluvial bedforms (Wohl et al., 1999). These similarities in form are naturally reflected in similarities in flow resistance.

Frictional resistance to flow by the bed and walls of a channel determines the relationship between velocity and depth, and has implications for both calculating basal shear stress (important for sediment transport and erosion) and quantifying flood hazards. Most work to quantify flow resistance has focused on low-gradient alluvial rivers (*see* Chapters 9.5 and 9.27) and may not translate well to the understanding of bedrock channel flow. Most bedrock channels resemble steep alluvial channels and are typically characterized by large grain size relative to flow depth. Flow resistance in these channels can be dominated by grain resistance (including large woody debris) and spill resistance (from jet or nappe flow), particularly at low flow (*see* Chapters 9.5 and 9.27). Wilcox et al. (2006) evaluated the relative contributions of grain, debris, and spill resistance in step-pool channels in a laboratory flume setting, and found that the individual resistance elements compound each other nonlinearly, making their estimation using physically based models difficult. Nonetheless, Ferguson (2007) showed that laws used for low-gradient alluvial channels can be modified to fit gravel and boulder-bedded channels and calibrated the relationship with field data (*see* Chapter 9.5).

Table 1 Channel morphology in alluvial reaches

Category	*D_{50bm}*	*Gradient*	*Relative roughness (D_{50}/H)*	*Hydraulic roughness*
Cascade	Boulder	5–20%	0.5–1	0.06–0.2
Step-pool	Cobble-boulder	2.5–7.5%	0.3–0.8	0.06–0.2
Plane bed	Gravel-cobble	0.5–4%	0.1–0.8	0.05–0.07
Pool-riffle	Gravel	0.5–3%	<0.3	0.03–0.04

D_{50bm}, median grainsize of the bed material.
Source: Reproduced from Whipple, K.X., 2004. Bedrock rivers and the geomorphology of active orogens. Annual Review of Earth and Planetary Sciences 32, 151–185, with permission from Annual Reviews.

9.28.2.2 Controls on the Width of Bedrock Rivers

Channel width and its variation with water discharge (both at a station and downstream) importantly influence bed shear-stress patterns, and thus play a first-order role in controlling the pattern and tempo of bedrock channel incision. Despite the wide appreciation of this simple statement, few studies have systematically evaluated the factors governing bedrock channel cross-section development. Bedrock channel width must be explicitly defined for river incision models. Given challenges to measuring channel width in mountainous terrain, classical hydraulic geometry relationships among width, discharge, and drainage area (*see* Chapter 9.18) developed for lowland alluvial rivers are substituted for direct field measurements. There is some indication that this approach is appropriate – the widths of bedrock and alluvial channels both appear to scale similarly with drainage area across many orders of magnitude and, in fact, appear to have physically similar widths at comparable drainage areas (**Figure 2**, Montgomery and Gran, 2001; Parker et al., 2007; Whipple, 2004; Wohl and David, 2008; Wohl and Merritt, 2005). This scaling suggests that the factors governing bedrock and alluvial channel width are similar, even though it is widely thought that bedrock channels have essentially fixed boundaries over the timescales of alluvial channel adjustment. Clearly, bedrock rivers are self-formed with channel width dynamically adjusted to flow and sediment transport regimes, and presumably substrate lithology and rock uplift rate.

Although it is convenient to model bedrock channel width as a simple power-law function of drainage area (**Figure 2**), it has been hypothesized that width also depends on uplift rate and serves as an important mode of channel adjustment to base-level change. Recently, there have been several efforts to model the dynamic adjustment of bedrock channel width with uplift rate (and therefore channel slope). Finnegan et al. (2005) and Whittaker et al. (2007a) prescribed a width-to-depth ratio that introduces a power-law slope dependence for width, while Turowski et al. (2009), Wobus et al. (2006b, 2008), and Yanites and Tucker (2010) each modeled the cross-sectional evolution of bedrock channels based on a shear-stress erosion law to arrive at similar relationships. Each of these models predicts a significant narrowing of channels with increasing rock uplift rate, which would notably alter the relationship between channel steepness and rock uplift rate. Despite these consistent theoretical predictions, field observations suggest a more complicated reality.

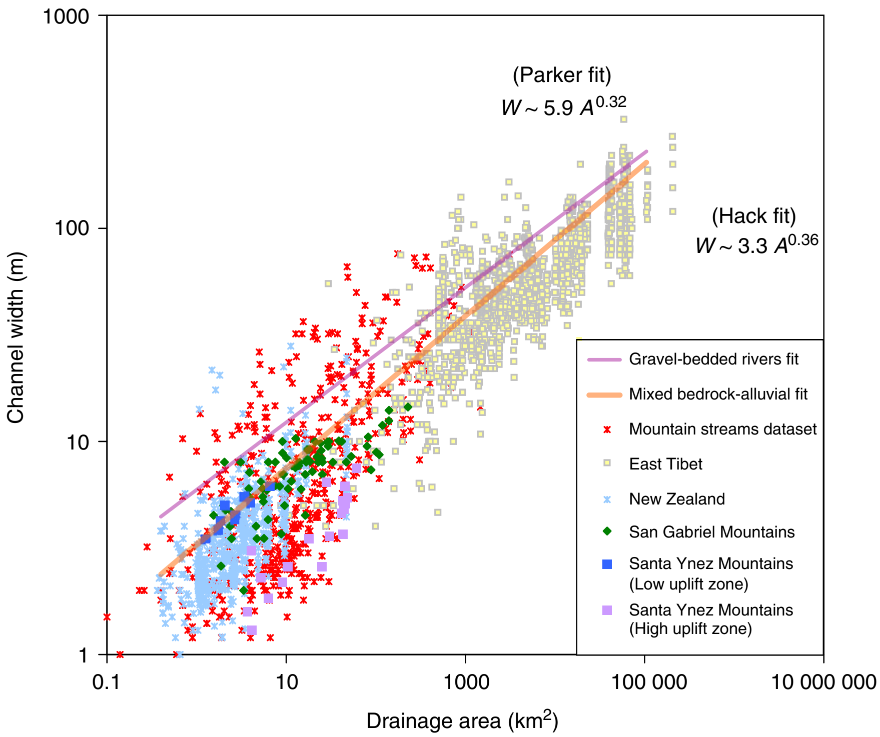


Figure 2 Bedrock channel width as a function of upstream drainage area in graded bedrock rivers. Power-law scaling relations for alluvial gravel-bedded rivers (e.g., Parker et al., 2007) and mixed bedrock–alluvial rivers (Hack, 1957) are shown for comparison. Data includes rivers undergoing a wide range of uplift (and incision) rates from Wohl and Merritts (2005) global compilation, the eastern margin of the Tibetan Plateau (Kirby and Ouimet, 2011), New Zealand (Crosby, 2006), the San Gabriel Mountains (DiBiase et al., 2009), and the Santa Ynez mountains (Duvall et al., 2004). Only the Santa Ynez data show a narrowing of channels in zones of higher rock uplift in graded bedrock rivers. Albeit with considerable scatter, bedrock rivers show the same scaling with drainage area as gravel-bed alluvial rivers (which also show much scatter), with mean channel width slightly narrower in bedrock channels (less than a factor of 2) for the same drainage area. Some of the observed scatter is likely attributable to differences in runoff and flood variability among sites, differences in substrate properties, sediment load, and differences in bed morphology. Reproduced from Wohl, E., Merritt, D.M., 2008. Reach-scale channel geometry of mountain streams. Geomorphology 93(3–4), 168–185.

Most field observations of channel narrowing in response to rock uplift rate are associated with rivers that cross zones of locally enhanced rock uplift rate (e.g., crossing an anticline) or that are undergoing a transient response to an increase in rock uplift rate (Amos and Burbank, 2007; Harbor, 1998; Lave and Avouac, 2001; Whittaker et al., 2007a, 2007b; Yanites et al., 2010). These observations are broadly consistent with predictions of the models described above. However, studies of well-graded quasi-equilibrium channels that span a wide range of rock uplift rates (together covering 0.1–4 mm a^{-1}) have found no detectable change in channel width (**Figure 2**) (DiBiase et al., 2009; Snyder et al., 2003a), directly contradicting model predictions. An exception is the study by Duvall et al. (2004) (**Figure 2**). Duvall et al. constructed relationships between measured channel width and drainage area across an uplift rate gradient in the Santa Ynez Mountains, California, and found that channels cut into mudstones were systematically narrower (and steeper) in the high uplift (4 mm a^{-1}) zone than in the low uplift (2 mma^{-1}) zone. Interestingly, these mudstone channels were remarkably devoid of coarse bed load – as discussed below, this may provide an important clue to the different behavior of these channels at steady state.

The contrast between well-graded, quasi-equilibrated channels and transient channel response to an increase in rock uplift rate is dramatically illustrated in a study by Whittaker et al. (2007a, 2007b). They studied a set of three bedrock rivers in the Central Apennines, Italy with excellent spatial and temporal constraints on uplift rate. Two of the channels are interpreted to be equilibrated to a spatially variable uplift rate, and for both channels, channel width scales as a simple power-law function of drainage area similar to alluvial channels and relationships seen for well-graded bedrock channels elsewhere (**Figure 2**, Montgomery and Gran, 2001; Whipple, 2004; Wohl and David, 2008). The third river in their study, Rio Torto, is interpreted to be undergoing a transient response to a threefold increase in uplift rate since 1 Ma, and is characterized by a prominent knickpoint downstream of which there is an inner gorge where width remains constant despite a doubling of drainage area (**Figure 3**) (Whittaker et al., 2007a). DiBiase et al. (2009) described a similar contrast in California's San Gabriel Mountains. They reported that whereas well-graded channels have similar widths over a wide range of incision rate (and slope), oversteepened knickzones are consistently much narrower. These observations beg the question as to why well-graded channels and channels undergoing significant downstream increases in rock uplift rate are so different.

The role of the supply of bedload relative to transport capacity is increasingly recognized as an important factor in bedrock channel incision – providing both the tools for abrasion and at times protecting the bed from impacts (Lague, 2010; Sklar and Dietrich, 2004; Turowski et al., 2007). As noted in the definition of bedrock channels, most have a thin, semi-continuous cover of coarse sediment (Howard, 1998). This alluvial cover may act to armor the bed and thus enhance lateral erosion relative to incision, which may influence channel width. Most of the channel-width models described above do not account for this potential role of sediment cover and thus may be missing a key piece of the physics. Bedrock channel experiments conducted by Finnegan et al. (2007) provide important insights into the role of sediment supply in setting channel width. By controlling bedload sediment flux independent of water discharge, Finnegan et al. were able to adjust the ratio of sediment supply (Q_s) to transport capacity (Q_c) and found that this ratio exerts a strong control on channel width. Increasing sediment supply led to alluviation and bank erosion, whereas decreasing sediment supply focused erosion into a narrow inner channel. This co-variation is consistent with field observations of enhanced bedrock exposure (sediment-starved conditions) in transient knickzones where channel width has narrowed in response to accelerated incision (**Figure 3**, DiBiase et al., 2009; Whittaker et al., 2007a). By contrast, well-graded channels in the San Gabriel Mountains show significant alluvial cover regardless of rock uplift rate and channel slope (DiBiase et al., 2009). The lack of coarse bedload in the mudstone channels studied by Duvall et al. (2004) may explain their different response under equilibrium, well-graded conditions (**Figure 2**). Moreover, rivers crossing localized zones of more rapid uplift can be expected to show an increase in sediment transport capacity relative to sediment supply (Whipple and Tucker, 2002) and constitute the bulk of field settings where channel width is known to decrease with increasing channel slope (Amos and Burbank, 2007; Harbor, 1998; Lave and Avouac, 2001; Yanites et al., 2010). These observations support the Finnegan et al. (2007) hypothesis that sediment supply plays a key role in setting bedrock channel width. Models put forward by Turowski et al. (2007) and Yanites and Tucker (2010) have incorporated the influence of alluvial cover in bedrock channel width and discuss how their findings may explain disparate field observations. Despite these important advances, fully disentangling the roles of sediment supply, discharge, slope, and rock strength in setting bedrock channel width remains an important problem, and likely will require carefully designed field and experimental flume studies to guide further theoretical developments.

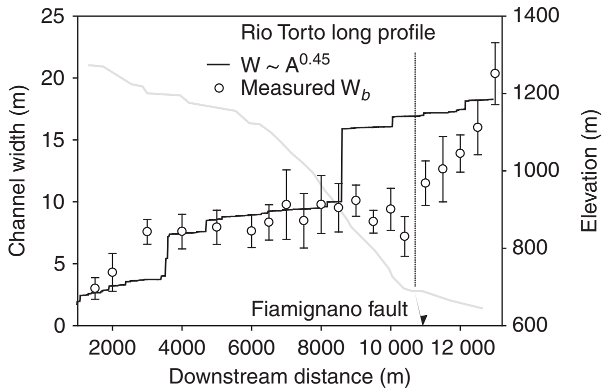


Figure 3 An example of river narrowing in response to a transient increase in channel gradient associated with an acceleration in rock uplift rate relative to base level (Rio Torto). Circles show measured high-flow channel width (error bounds show local variability) in comparison to expected channel width assuming a normal scaling between width and drainage area (solid line). Vertical line indicates mountain front. Reproduced from Figure 11c in Whittaker, A.C., Cowie, P.A., Attal, M., Tucker, G.E., Roberts, G.P., 2007b. Contrasting transient and steady-state rivers crossing active normal faults: new field observations from the Central Apennines, Italy. Basin Research 19(4), 529–556, with permission from Wiley.

9.28.3 Erosion Processes and Bedforms

Understanding the linkages among climate, tectonics, and landscape evolution requires that we can predict the rate of river incision into rock under a given set of conditions (e.g., drainage area, channel slope, climate, sediment delivery rate, and rock properties). Such predictions must be informed by mechanistic understanding of the array of processes involved. Bedrock exposed from time to time in the bed and banks of a channel can be: (1) abraded by sediment moving either as bedload (hopping and rolling along the bed) or as suspended load carried in the water column (*see* Chapters 9.8 and 9.9) (**Figures 4(c)** and **4(d)**); (2) plucked from the bed by hydraulic forces (**Figures 4(a)** and **4(b)**); (3) damaged by the violent collapse of cavitation bubbles; (4) weakened by weathering processes; or (5) scoured by passing debris flows (probably a combination of abrasion and plucking). The efficacy of each of these processes is affected by both flow conditions (discharge, velocity, and bed shear stress) and the amount and size distribution of sediment in the system. Interactions among various processes are almost certainly important to the overall incision rate, but have been little studied. River incision models tend to either lump all processes together or isolate one process or another for detailed analysis. In order to motivate the discussion of river incision models (Section 9.28.4) and to highlight some limits to the representation of actual processes in current models, we briefly review each of these processes and their potential interactions here.

9.28.3.1 Abrasion

Fluvial abrasion is responsible for most of the aesthetically appealing erosional bedforms that grace rock outcrops in the bed and banks of rivers including flutes, scallops, and potholes (Richardson and Carling, 2005) (**Figure 4(c)**). At a finer scale, abrasion acts to smooth or even polish rock surfaces by breaking off sharp edges, corners, and protuberances. In this sense, and indeed mechanistically as well, fluvial abrasion has much in common with sand blasting. Erosion of the bed is caused by energetic collisions between traveling sediment grains and exposed bedrock surfaces. The simplest way to think of this is that with every impact some damage is done to both the impacting grain (familiar from studies of downstream fining) (e.g., Parker, 1991) and the rock bed. The number of impacts, and thus the erosion rate, must scale with the sediment flux and the percentage of the bed where rock is exposed. Thus, there will be a tools effect (a greater flux of sediment in transport means a greater number of collisions) and a cover effect (a greater flux of sediment in transport conversely means a lesser fraction of these collisions will be with exposed bedrock) (Chatanantavet and Parker, 2009; Sklar and Dietrich, 1998; Sklar and Dietrich, 2004; Turowski et al., 2007; see also Section 9.28.3.6).

Abrasion, particularly abrasion by bedload, is the most well-studied incision process, owing largely to the tractability of experimental study. Several experimental flume studies have confirmed the theoretical expectation first articulated by Sklar and Dietrich (1998, 2004) that erosion rate scales with the

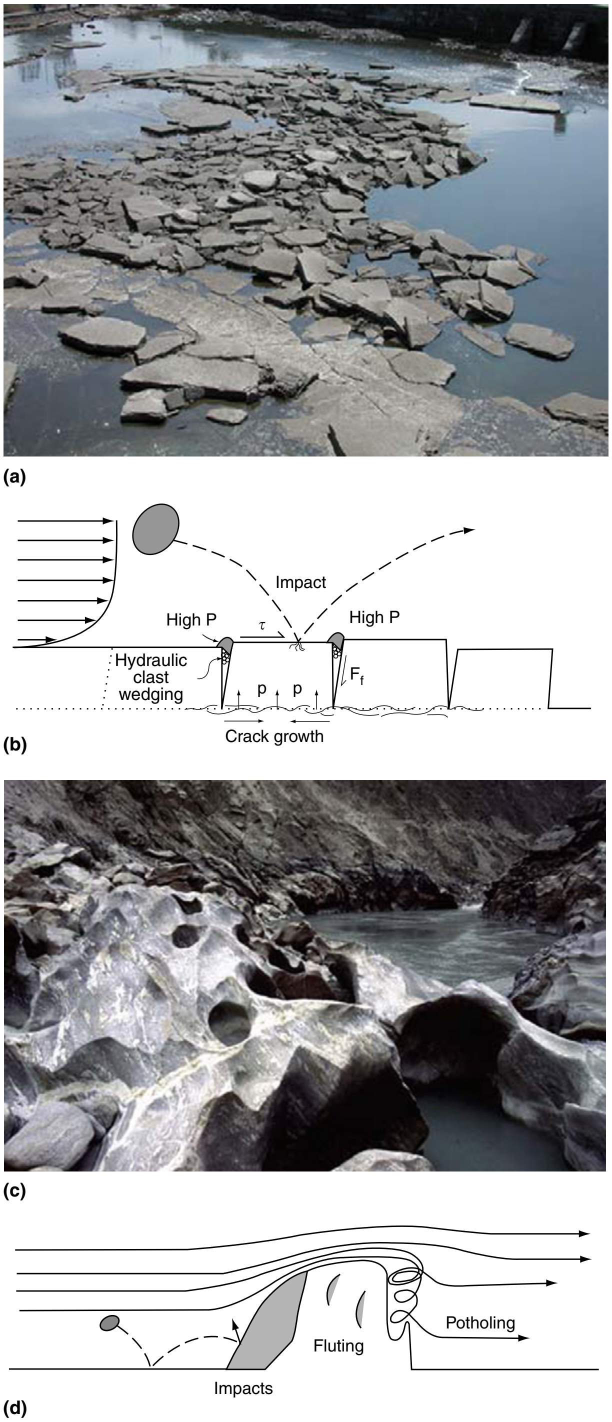


Figure 4 River-incision processes. (a) Photo of imbricated rock slabs plucked from the bed of the Mississippi river at St. Anthony Falls, MN. Pair of tire tracks from large trucks (middle left to lower right corner) for scale. (b) Schematic of plucking process. (c) Photo of abrasion flutes and potholes carved in gneiss, Indus River, Pakistan. (d) Schematic of abrasion process. Modified from Whipple, K.X., Hancock, G.S., Anderson, R.S., 2000. River incision into bedrock: mechanics and relative efficacy of plucking, abrasion, and cavitation. Geological Society of America Bulletin 112(3), 490–503, with permission from The Society.

product of sediment flux per unit channel width and the fraction of exposed bedrock in the channel (Chatanantavet and Parker, 2009; Sklar and Dietrich, 1998; Sklar and Dietrich, 2004; Turowski et al., 2007).

Beyond this fundamental and important result, however, much uncertainty remains regarding how flow conditions influence erosion rate for a given sediment flux, even where the fraction of rock exposure is held fixed. Theory suggests that, all else held equal, bedload abrasion rate should decrease with increasing bed shear stress (Sklar and Dietrich, 2004) because the number of bed impacts decreases with increasing bed shear stress resulting in a net decrease in erosion rate. In theory, the role of water discharge and channel slope should be encapsulated in this predicted dependence on bed shear stress. Experimental observations, however, show that bedload abrasion rates increase with increasing channel slope (e.g., Johnson and Whipple, 2007; Wohl and Ikeda, 1997) and are independent of water discharge when all other variables are held constant (Johnson and Whipple, 2010). Further experimental work on the roles of particle saltation trajectories and evolving bed roughness in erosion by abrasion will be required to bring theory in line with observations.

The efficacy of abrasion by suspended load is also a matter of debate. Several researchers have argued that many finely sculpted erosional bedforms etched into rocky river beds, walls, and large rarely mobile boulders (flutes, spindles, scallops, even potholes), as well as observed spatial patterns of erosion, testify to vigorous erosion by suspended load (Hancock et al., 1998; Hartshorn et al., 2002; Springer et al., 2006; Whipple et al., 2000). These authors appeal to flow vortices and macroturbulence shed by bed steps and rocky protrusions to bring suspended sediment in contact with the bed (**Figure 4(d)**). Conversely, Sklar and Dietrich (2004) argued that where erosion of the channel bed is at issue, sediment in suspension cannot contribute measurably due to lack of contact – it may ornament rocky protrusions and banks, but not quantitatively contribute to incision. Lamb et al. (2008) augmented the original Sklar and Dietrich (2004) analysis and demonstrated some contribution by suspended load, but still found it minor compared to bedload abrasion. These theoretical arguments have been framed, however, in the context of normal flow over a planar bed lacking significant vortices. Further flume experimentation is needed to quantitatively resolve the relative roles of abrasion by bedload and suspended load.

9.28.3.2 Plucking

Plucking of blocks from the bed and banks is a simple, efficient, but understudied process. Observations indicate that wherever rocks are fractured at a scale that makes plucking by hydraulic forces possible, plucking is the dominant erosion process. Recently plucked blocks typically are only slightly rounded and sculpted by abrasion, leading to the interpretation that plucking can be an order of magnitude more efficient than abrasion (Hancock et al., 1998; Whipple et al., 2000) (**Figure 4(a)**). Similarly, reaches of rivers cut in highly fractured bedrock are commonly somewhat less steep and fully covered in a litter of plucked blocks and bedload clasts, suggesting that it is the transport of bedload and plucked blocks that limits the rate of river incision, as has been argued for weak substrates in general (Sklar and Dietrich, 2004; van der Beek and Bishop, 2003; Whipple and Tucker, 2002). Conversely, reaches carved into massive rock are more likely to exhibit much exposed rock, which is commonly sculpted by abrasional bedforms, suggesting that in these sections it is the abrasion process that limits the rate of river incision.

Ultimately, plucking occurs where drag and lift forces are sufficient to extract blocks from the channel bed, but a variety of processes contribute to the expansion, weakening, and opening of fractures around blocks (freeze/thaw, heating/cooling, weathering, bedload impacts, and pressure fluctuations in cracks) (**Figure 4**; Hancock et al., 1998; Whipple et al., 2000). Thus, a significant threshold shear stress for the onset of plucking is expected and shares much in common with the problem of initial motion of bedload (*see* Chapter 9.7). Despite its prevalence in natural settings and qualitative indications that it is generally the dominant process, incision by plucking has been little studied in the laboratory, is explicit in only one river incision model (Chatanantavet and Parker, 2009), and is generally ignored in applications of this model (and derivatives of it) (e.g., Chatanantavet and Parker, 2009; Crosby et al., 2007; Gasparini et al., 2007).

9.28.3.3 Cavitation and Corrosion

Cavitation here refers to damage to rock surfaces caused by the violent implosion of small bubbles (Barnes, 1956). Corrosion refers to a suite of weathering and dissolution processes that can weaken rock fabric, joints, and even remove mass. Both may contribute in most settings primarily by making rock surfaces more susceptible to abrasion and plucking. Corrosion certainly occurs through weathering of silicate rocks and dissolution of carbonate rocks. In many actively incising bedrock rivers, it has commonly been surmised that physical erosion processes are dominant and that corrosion may have insufficient time to contribute significantly. This perception has not, however, been quantitatively evaluated. In addition, it is clear that weathering is an important control on bed-lowering rates and channel width where the substrate (e.g., mudstone) is highly susceptible to wetting and drying cycles (Montgomery, 2004; Stock and Dietrich, 2006). Given these observations and the fact that researchers are recognizing a greater contribution of weathering processes on hillslopes and in soils than previously appreciated even where erosion rates are quite high (e.g., Chamberlain et al., 2007; Dixon et al., 2009; Ferrier and Kirchner, 2008; Riebe et al., 2004), the role of corrosion in river incision into bedrock merits greater attention. Cavitation damage is known to have played a key role in the destruction of some concrete spillways (Arndt, 1981), but whether and how much cavitation contributes to river incision in natural settings remains unclear. No direct evidence of a significant role has been found, although theoretical considerations suggest that hydrodynamic conditions in many bedrock river reaches are conducive to cavitation damage (Whipple et al., 2000).

9.28.3.4 Debris Flow Scour

Episodic debris flows are a common occurrence in many steep, headwater bedrock channels. Channels steeper than $\sim 10\%$ and with contributing areas $< 1\text{–}10\ \text{km}^2$ are commonly traversed by debris flows (e.g., Montgomery and Foufoula-Georgiou, 1993; Stock and Dietrich, 2003). Where channels are steep and debris flows fast moving, channels normally covered in a thin layer of alluvium and colluvium are commonly observed to be swept clean to bedrock by the passage of debris flows. The abrasive power of fast-moving debris flows may be great and there is ample direct evidence of both abrasional wear and plucking of blocks by debris flows (Hsu et al., 2008; Stock and Dietrich, 2003). In addition, it is plausible that corrosion under a thin blanket of alluvium and colluvium between debris-flow scour events helps facilitate significant bed lowering in a positive feedback where the occasional scouring makes available fresh rock that can be more rapidly weathered (Stock and Dietrich, 2006). Further work is needed to fully resolve the controls on the efficacy of debris-flow scour and also the interactions and relative contributions of fluvial and debris-flow erosion processes as a function of position in the landscape.

9.28.3.5 Process Interactions

As mentioned above, at any given point along a bedrock river, all erosion mechanisms may be at work, and their interactions may be just as important to the overall rate of incision as is the physics of each process independently. One example is the potential for corrosion to weaken the substrate in between major flood or debris-flow scour events. Similarly, it seems clear that the morphology of the river bed influences the efficacy of all processes described above and in turn that this morphology is set by the combination of active processes and their interactions (Johnson and Whipple, 2007; Johnson and Whipple, 2010). Interactions of abrasion and plucking evident in the field make a case in point. First, plucking of large blocks can generate macro-roughness elements that shed vortices that both influence the pattern of bedload abrasion and allow suspended load abrasion to occur (**Figure 4**, Whipple et al., 2000). Second, in areas of particularly massive rocks, it has been observed that abrasional fluting and pothole formation proceeds until so little is left of the original rock mass that it becomes susceptible to fracturing and plucking (Whipple et al., 2000). These process interactions have not been explicitly represented in any model and have not yet been studied experimentally.

9.28.3.6 Models of River Incision into Bedrock

A variety of river incision models, ranging from phenomenological to process specific, have been proposed (*see* Chapter 9.34). The most well known is the stream-power river incision model (or family of models) (e.g., Howard et al., 1994; Howard and Kerby, 1983; Tucker and Whipple, 2002). The family of stream-power models is not process specific. In other words, these models do not treat the physics of individual incision processes or their interactions, but rather attempt to lump all processes together as an overall dependence of river incision rate on local bed shear stress and metrics of rock strength and flood frequency. Rationalizations have been given for parameter values more consistent with erosion by plucking, bedload abrasion, and suspended load abrasion (e.g., Whipple et al., 2000) – hence, the family of models. Despite these limitations, the stream-power family of models has proved to be a useful tool for investigation. Moreover, various process-specific models can be cast as variants of the stream-power model, shedding light on the similarities and differences between models. This approach is taken here.

9.28.3.6.1 Essentials

In the stream-power family of models, bed erosion rate varies as a power function of mean bed shear stress which must exceed a critical threshold of bedload motion or rock detachment, whichever is greater (Howard and Kerby, 1983; Whipple and Tucker, 1999), and can be written as

$$E = k_e f(q_s)\left[1 - \left(\tau_c{}^a/\tau_b{}^a\right)\right]\tau_b{}^a \qquad [1]$$

where k_e is a function of substrate properties (large values for weak, easily eroded materials), $f(q_s)$ describes the influence of tools and cover briefly mentioned in Section 9.28.3.1 (often neglected), τ_b is the mean bed shear stress, τ_c is the threshold shear stress, and the exponent a depends on the mechanics of erosion (Howard and Kerby, 1983; Whipple et al., 2000). As unit stream power scales with $\tau_b{}^{3/2}$ under normal flow conditions, a unit-stream-power model only differs from a shear-stress model in the effective value of the exponent a (e.g., Whipple and Tucker, 1999); both are described by eqn [1]. The term in brackets vanishes if the threshold stress is negligible for the floods of interest, as is often assumed.

In most models, including more sophisticated process-specific models (e.g., Sklar and Dietrich, 2004), it is assumed that flow is approximately steady and uniform (the normal flow assumption) such that shear stress can be approximately described in terms of water discharge (Q), bankfull width (W), and bed slope (S) (e.g., Howard et al., 1994; Tucker, 2004):

$$\tau_b = k_t (Q/W)^{\alpha} S^{\beta} \qquad [2]$$

where k_t, α, and β are set by a flow resistance relation (e.g., Manning's equation; *see* Chapter 9.5). Employing eqn [2] and empirical relations among bankfull discharge, drainage area (A), and bankfull channel width (see Section 9.28.2.2 and Chapter 9.18), a generalized form of the stream-power family of models can be written as (Whipple, 2004)

$$E = K_r K_c K_{\tau_{cr}} f(q_s) A^m S^n \qquad [3]$$

where K_r is set by k_e, k_t, and channel width, K_c is set by climatic conditions, $K_{\tau cr}$ is a threshold term ($0 \le K_{\tau cr} \le 1$) equivalent to the bracketed term in eqn [1], and exponents m and n are set by the exponents a (eqn [1]), α, and β (eqn [2]) and the relationships among Q, A, and W (e.g., Howard and Kerby, 1983; Whipple and Tucker, 1999). The controls on channel width were discussed in Section 9.28.2.2 (**Figures 2** and **3**). With some approximations, all published river incision models can be cast in the form of eqn [3] (e.g., Gasparini et al., 2007; Whipple, 2004). It is important to recognize that exponents m and n are not free parameters as is often surmised. Rather, m and n take on specific values that reflect the mechanics of the dominant river-incision process and the m/n ratio is constrained by typical empirical relationships among Q, A, and W to be around 0.5 (Whipple and Tucker, 1999). Exponents m and n may be zero (no net dependence on shear stress) (Chatanantavet and Parker, 2009) or negative (erosion rate decreasing with shear stress, all else held equal) (Sklar and Dietrich, 2004). The definitions of K_r, K_c, $K_{\tau cr}$, $f(q_s)$, m, and n relevant to most published river incision models are given in **Table 2** (modified from Whipple, 2004). Despite the many differences, all models predict monotonic increases in steady-state channel gradient with increasing rock uplift rate (even where exponents m and n are zero or negative).

9.28.3.6.2 Bed cover and tools

Gilbert (1877) recognized that sediment plays a dual role in river incision: (1) providing the tools to wear the bed by abrasion; and (2) shielding the bed from impacts and from hydrodynamic plucking where sediment is sufficiently abundant. Sklar and Dietrich (1998, 2004) postulated that the tools effect could be captured with a linear dependence on sediment flux per unit width (q_s) and the cover effect by a linear dependence on the ratio of sediment flux to transport capacity (q_s/q_c):

$$f(q_s) = q_s[1 - (q_s/q_c)] \qquad [4]$$

Turowski et al. (2007) suggested a more complex alternative that has some intuitive appeal. However, as yet, experimental data do not indicate that a more complex relationship is needed to explain observations (Chatanantavet and Parker, 2008; Finnegan et al., 2007; Johnson and Whipple, 2010; Sklar and Dietrich, 2001). Both the Sklar and Dietrich (1998, 2004) formulation and that of Turowski et al. (2007) imply that there is an optimum bedload sediment flux for erosion by abrasion: too little and erosion is inhibited by a lack of tools, too much and erosion is inhibited by bed cover. Field and experimental data strongly support the need to account for the effects of bed cover (Chatanantavet and Parker, 2008; Cowie et al., 2008; Finnegan et al., 2007; Johnson et al., 2009; Johnson and Whipple, 2007; Johnson and Whipple, 2010; Valla et al., 2010; van der Beek and Bishop, 2003), regardless of whether abrasion is the dominant erosion mechanism. Interestingly, theoretical considerations suggest that variability in daily flows and sediment supply may be the dominant control on the long-term influence of bed cover (Lague, 2010). Experimental data for erosion by abrasion alone thus far support the linear dependence on q_s (the tools effect) (Chatanantavet and Parker, 2008; Johnson and Whipple, 2010; Sklar and Dietrich, 2001). Whereas some field observations support the need to incorporate a tools effect (e.g., Cowie et al., 2008; Crosby et al., 2007; Wobus et al., 2006a), no field data yet provide a clear test of the proposed linear dependence on sediment flux (all else held equal), which might be expected to take a different form if plucking were important (e.g., Whipple et al., 2000).

9.28.3.6.3 Erosion thresholds and flood frequency

Much as recent papers have driven home the importance of the sediment flux relative to transport capacity (eqns [3] and [4]), recent analyses have highlighted the critical role of the erosion threshold term (τ_c in eqn [1] and $K_{\tau c}$ in eqn [3]) in the relationships among climate, topography (channel steepness), and tectonics (Lague et al., 2005; Snyder et al., 2003b; Snyder et al., 2003c; Tucker, 2004). A central conclusion of these analyses is that the influence of an erosion threshold cannot be properly evaluated unless the probability distribution of floods is considered – adding a threshold term to a standard effective discharge model captures little of the predicted influence of the threshold. This follows because, at its most fundamental, the threshold term determines what portion of the full distribution of river flows actually contributes to erosion. Naturally, a greater percentage of floods will generate shear stresses in excess of the erosion threshold in steeper channels (for the same drainage area), resulting in more efficient erosion (**Figure 5**). This effect is expected regardless of the dominant incision process and whether or not the bed is covered by a thin layer of alluvium.

The finding that the threshold term has a greater influence at low channel steepness (and low erosion rate or tectonic uplift rate) imparts a strong nonlinearity to the relationship between channel steepness and erosion rate at steady state (Lague et al., 2005; Snyder et al., 2003b; Snyder et al., 2003c; Tucker, 2004). Interestingly, the nature of this nonlinearity depends on the magnitude of the threshold, the rock uplift rate, the climate (both mean and variability of rainfall/runoff), and whether large floods follow an exponential (Tucker, 2004) or power-law (Lague et al., 2005) probability distribution. The most important difference in the exponential and power-law models is in the relationship among climate, channel steepness, and erosion rate, particularly at relatively low channel steepness and erosion rate where the threshold term is dominant. The two models are also somewhat differently sensitive to climatic variability. These effects and, to some degree, the differences between these two models have important implications for the strength of the coupling between climate and tectonics (e.g., Whipple, 2009) (*see* Chapter 9.1). In addition, these models make important predictions of the relative roles of mean precipitation (or runoff) and its variability (Lague et al., 2005; Molnar, 2001; Molnar et al., 2006; Tucker, 2004). Both models should be tested against field data, and the implications of their differences should be more fully explored.

9.28.4 River Profiles and Landscape Relief

As noted earlier, the longitudinal profiles of bedrock rivers dictate much of the 3D architecture of unglaciated mountainous topography (Howard et al., 1994). In large drainage basins (>50 km^2), 80% or more of topographic relief is set by the elevation drop along bedrock rivers, with hillslopes and colluvial channels contributing the remainder (Whipple, 2004) (**Figure 1**). This is why bedrock channel incision models are so crucial to exploring the linkages among climate, lithology, tectonics, and topography; the controls on longitudinal profile form effectively set the overall relief structure of the topography. In the sections below, we first consider the form of steady-state (or graded, Mackin, 1948) river profiles, including the influence of lithology, climate, and tectonics on

Table 2 Equation sets for the stream power family of models

Shared internal relations			*Shared exponents*	
Hydrology[a]	*Hydraulic geometry*[b]	*Conservation of momentum*	*Area*	*Slope*
$Q = K_q A^c$	$W = K_w Q^b$ $W/W_b = (Q/Q_b)^s$	$\tau_b = k_t (Q/W)^\alpha S^\beta$; $k_t = \rho g^\alpha C_f^{\alpha/2}$ $\alpha = \frac{2}{3}$; $\beta = \frac{2}{3}$ (Chezy); $\alpha = \frac{3}{5}$; $\beta = \frac{7}{10}$ (Manning)	$m = \alpha a c(1-b)$	$n = \beta a$

Relations for Erosional Efficiency $(K = K_r K_c K_{\tau_{cr}} f(q_s))$

Model	K_r	K_c	$K_{\tau_{cr}}$	$f(q_s)$
Shear stress	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = k_q^{\alpha a(1-b)}$	$K_{\tau_{cr}} = 1$	$f(q_s) = 1$; $Q_s/Q_c < 1$ $f(q_s) = 0$; $Q_s/Q_c \geq 1$
Linear decline	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = k_q^{\alpha a(1-b)}$	$K_{\tau_{cr}} = 1$	$f(q_s) = 1 - Q_s/Q_c$
Parabolic	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = k_q^{\alpha a(1-b)}$	$K_{\tau_{cr}} = 1$	$f(q_s) = 1 - 4(Q_s/Q_c - 1/2)^2$
Scour depth[c]	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = k_q^{\alpha a(1-b)}$	$K_{\tau_{cr}} = 1$	$f(q_s) = 1$; $Q_s/Q_c < 1$ $f(q_s) = \exp(-h/L)$; $Q_s/Q_c \geq 1$
Saltation-abrasion[d]	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = k_q^{\alpha a(1-b)}$	N/A $\tau_{cr} > 0$; eqn 3	$f(q_s) = (Q_s/W)(1 - Q_s/Q_c)$
Stochastic-threshold[e]	$K_R = k_e k_w^{-\alpha a} k_t^{\ a}$	$K_C = \left(\frac{Tr}{Tr + Tb}\right) P^{\gamma_b} R_b^{-\varepsilon_b} \exp\left(\frac{-I}{P}\right) \Gamma(\gamma_b + 1)$	$K_{\tau_{cr}} = \frac{\Gamma\left(\gamma_b + 1, \frac{R_c}{P}\right) - \left(\frac{R_c}{P}\right)^{\gamma_b} \exp\left(\frac{-R_c}{P}\right)}{\Gamma(\gamma_b + 1)}$	$f(q_s) = 1$; $Q_s/Q_c < 1$ $f(q_s) = 0$; $Q_s/Q_c \geq 1$

[a] $Q_b = R_b A$.
[b] $W = k_w Q^b$ for bankfull flow in stochastic model.
[c] $L \propto Q_c/W$.
[d] $a = -0.88$.
[e] $\gamma_b = \alpha a(1-s)$, $\varepsilon_b = \alpha a(b-s)$.
Source: Reproduced from Whipple, K.X., 2004. Bedrock rivers and the geomorphology of active orogens. Annual Review of Earth and Planetary Sciences 32, 151–185, with permission from Annual Reviews

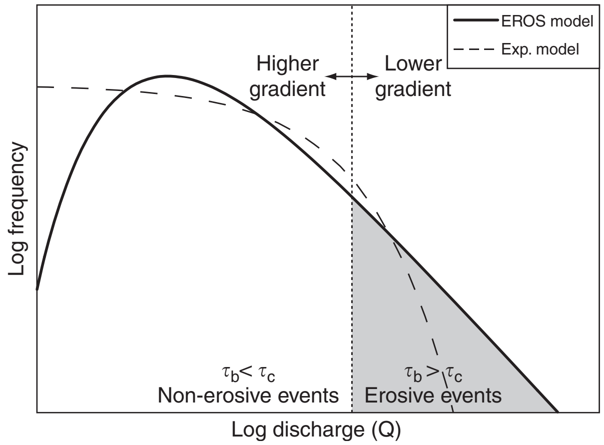


Figure 5 Illustration of the role of a critical threshold for coarse bedload mobility and/or bedrock incision and the probability distribution of flood magnitudes. Log–log plot of the frequency of each flow event as a function of mean daily flow showing both an exponential discharge distribution (Tucker, 2004) and the EROS modified power-law distribution (Lague et al., 2005). The shaded region shows flow events in the EROS model in which the critical shear stress is exceeded – only these events can do any geomorphic work. For the exponential model, only flow events in the shaded region and under the dashed line could do geomorphic work. As illustrated, the fraction of flows that can do geomorphic work increases with increasing channel slope, all else held equal.

both channel steepness and concavity. Next, we discuss (1) the relationship between channel profile form and topographic relief measured at different length scales, and (2) the relations between these topographic metrics and erosion rate in quasi-steady-state landscapes. In the last section, we discuss the controls on transient river response – how channel profiles record landscape response to significant changes in climate, base level, and tectonics, including the formation and migration of river knickpoints.

9.28.4.1 Longitudinal River Profiles – Steady-State Forms

In our usage, steady state refers to a condition in which the channel profile has fully adjusted to the climatic, lithologic, and tectonic conditions imposed upon it. These factors need not be uniform in space, but must be invariant over sufficiently long time to allow the river to adjust its longitudinal profile to the prevailing conditions. Climate and tectonic uplift both vary over a wide range of timescales, so a perfect steady state can probably never be achieved. However, oscillatory fluctuations with alternating periods of aggradation and incision (*see* Chapters 9.22 and 9.15) generally do not significantly affect the overall form of the river profile (e.g., Snyder et al., 2002); perturbations that are short compared to the response time of the river profile can influence short-term incision rates, but by definition do not persist long enough to much affect the river profile (Whipple, 2001). Thus, steady state as used here refers to a long-term ($>\sim 100$ ka) condition in which average incision rate balances the average rock uplift rate relative to base level, such that the longitudinal profile varies only slightly through time.

There is an extensive literature on the form of river profiles. First recognized by Hack (1957), many steady-state or graded river profiles are well described as a power–law relationship between local channel gradient (S) and upstream drainage area (A) that has become known as Flint's law (Flint, 1974):

$$S = k_s A^{-\theta} \qquad [5]$$

where k_s is known as the channel steepness index and θ as the concavity index (**Figure 6**). The channel steepness index is a generalized and more reliable version of Hack's (1973) gradient index (see also Goldrick and Bishop, 2007). Given Flint's law, the relationship between drainage area and distance downstream often described with Hack's law (Hack, 1957) strongly influences the rate of change in channel gradient with distance downstream, which of course defines the concavity of river profiles. Hence, the term 'concavity index' is used for the exponent in eqn [5] to make the distinction between the rate of change of slope with drainage area and with distance downstream clear.

Two important caveats apply to the expectation that river profiles generally conform to eqn [5]. First, Flint's law only applies downstream of a critical drainage area, A_{cr} (**Figure 6**), that generally ranges between 0.1 and 5 km^2 (Montgomery and Foufoula-Georgiou, 1993; Stock and Dietrich, 2003; Wobus et al., 2006c). Second, either abrupt along-stream differences or abrupt temporal changes in tectonics, climate, or exposed lithology can cause segmentation of a river profile into sections that are each usually well described by eqn [5], but with differing steepneses (k_s), concavity indices (θ), or both (e.g., Harkins et al., 2007; Kirby and Whipple, 2001; Wobus et al., 2006c). In addition, local variability in channel steepness as recorded in digital elevation data (DEM) can add considerable noise to slope-area plots and may reflect variations in rock strength, landslide-related knickpoints (*see* Chapter 9.15), or inaccuracies in the DEM. Analyses of river profiles should be undertaken only with these complexities in mind.

9.28.4.1.1 Controls on channel concavity

A range of perspectives exists on the variability in channel concavity index. Simply measuring channel concavity index from A_{cr} to the outlet on a large number of rivers will yield considerable variability (measured this way, observed concavity indices range from 0.3–1.2 or more) (e.g., Tucker and Whipple, 2002; Whipple, 2004 and references therein). However, restricting analyses to river systems that are arguably approximately in steady state based on independent data reduces this variability greatly. Further, recognizing the segmented nature of many river profiles, and restricting analyses to channel segments that appear to have relatively uniform lithology, climate, and rock uplift rate along stream (especially avoiding the transition from erosional to depositional conditions that commonly occurs at mountain fronts) reduces the variability of observed channel concavity indices to a narrow range between 0.4 and 0.7 (Whipple, 2004).

Observations of approximately steady-state channel profiles over a range of tectonic, climatic, and lithologic conditions (so long as these are uniform along measured river segments) reveal that the channel concavity index is not systematically controlled by these variables – invariance with uplift rate is well illustrated in the San Gabriel Mountains (**Figure 6**) and has been documented in many landscapes (e.g., Wobus et al., 2006c). River incision models (see Section 9.28.3.6 above) predict that the concavity index is most strongly controlled by the relative rates of downstream increase in river discharge and channel width (Tucker, 2004; Whipple and Tucker, 1999). The concavity index can be modified also by systematic downstream changes in: (1) rock uplift rate (Kirby and Whipple, 2001); (2) substrate properties (Moglen and Bras, 1995); (3) runoff (Craddock et al., 2007; Roe et al., 2002; Zaprowski et al., 2005); or (4) the frequency and extent of rock exposure in the channel bed (Sklar and Dietrich, 2004; Sklar and Dietrich, 2006). With the caveat that rainfall and runoff are commonly spatially variable due to orographic effects, as a good rule of thumb under steady-state conditions, the channel concavity index can be considered independent of climate, lithology, and tectonics (Wobus et al., 2006c).

9.28.4.1.2 Controls on channel steepness

Given that the concavity index appears to be largely independent of climate, lithology, and tectonics under steady-state conditions, the channel steepness index becomes a very useful metric of landform response to these controlling factors. However, small variations and uncertainties of the concavity index can greatly influence the value of k_s found from linear regression of log S versus log A (Sklar and Dietrich, 1998). The simple solution is to do regressions with an imposed reference concavity (θ_{ref}) such that channel steepness values can be directly compared:

$$S = k_{\mathrm{sn}} A^{-\theta_{\mathrm{ref}}} \qquad [6]$$

where k_{sn} is termed the normalized channel steepness index (Wobus et al., 2006c), effectively a measure of channel slope that has been corrected for the expected dependence of local slope on drainage area.

The normalized channel steepness index can be expected to vary with rock uplift rate (relative to base level), lithology, and climate. The observed global range of channel steepness (averaged over 3–50 km distance in graded channels) is 20–600 for $\theta_{\mathrm{ref}} = 0.45$ (Whipple, 2004). Numerous studies have documented that the channel steepness index of graded channels increases monotonically with either rock uplift relative to base level or erosion rate (DiBiase et al., 2010; Duvall et al., 2004; Harkins et al., 2007; Kirby and Whipple, 2001; Kirby et al., 2003; Kobor and Roering, 2004; Lague and Davy, 2003; Ouimet et al., 2009; Safran et al., 2005; Snyder et al., 2000; Wobus et al., 2006c) (**Figure 7**). Lithology, in some cases, clearly influences the channel steepness index with harder, less fractured rock associated with steeper channels (e.g., Duvall et al., 2004), but in other cases lithology appears to have no measurable influence (DiBiase et al., 2010; Kirby et al., 2003; Ouimet et al., 2009). Rock properties can influence channel slope directly through control of the efficiency of abrasion and plucking, or indirectly through control of the size and abundance of coarse debris (both immobile and bedload). Conversely, where channels are largely blanketed by a thin layer of alluvium, rock susceptibility to abrasion or plucking will have little influence on steady-state channel slope (Sklar and Dietrich, 2006), perhaps explaining why in some landscapes lithologic variability does not appear to influence channel steepness (Johnson et al., 2009; Kirby et al., 2003; van der Beek and Bishop, 2003).

Interestingly, there are few data to support the expectation that wetter and stormier climates should be associated with lower channel steepness, all else held equal (e.g., Aalto et al., 2006). Or, in other words, that wetter and stormier climates should be associated with higher erosion rates for the same channel steepness. This expectation is embedded in all channel incision models (e.g., Howard, 1994; Lague et al., 2005; Tucker, 1996) and hence in all models of the coupling between climate and tectonics (e.g., Whipple, 2009; Willett, 1999). However, as yet no published data directly and conclusively demonstrate, much less quantify, the expected influence of climate on steady-state channel steepness index.

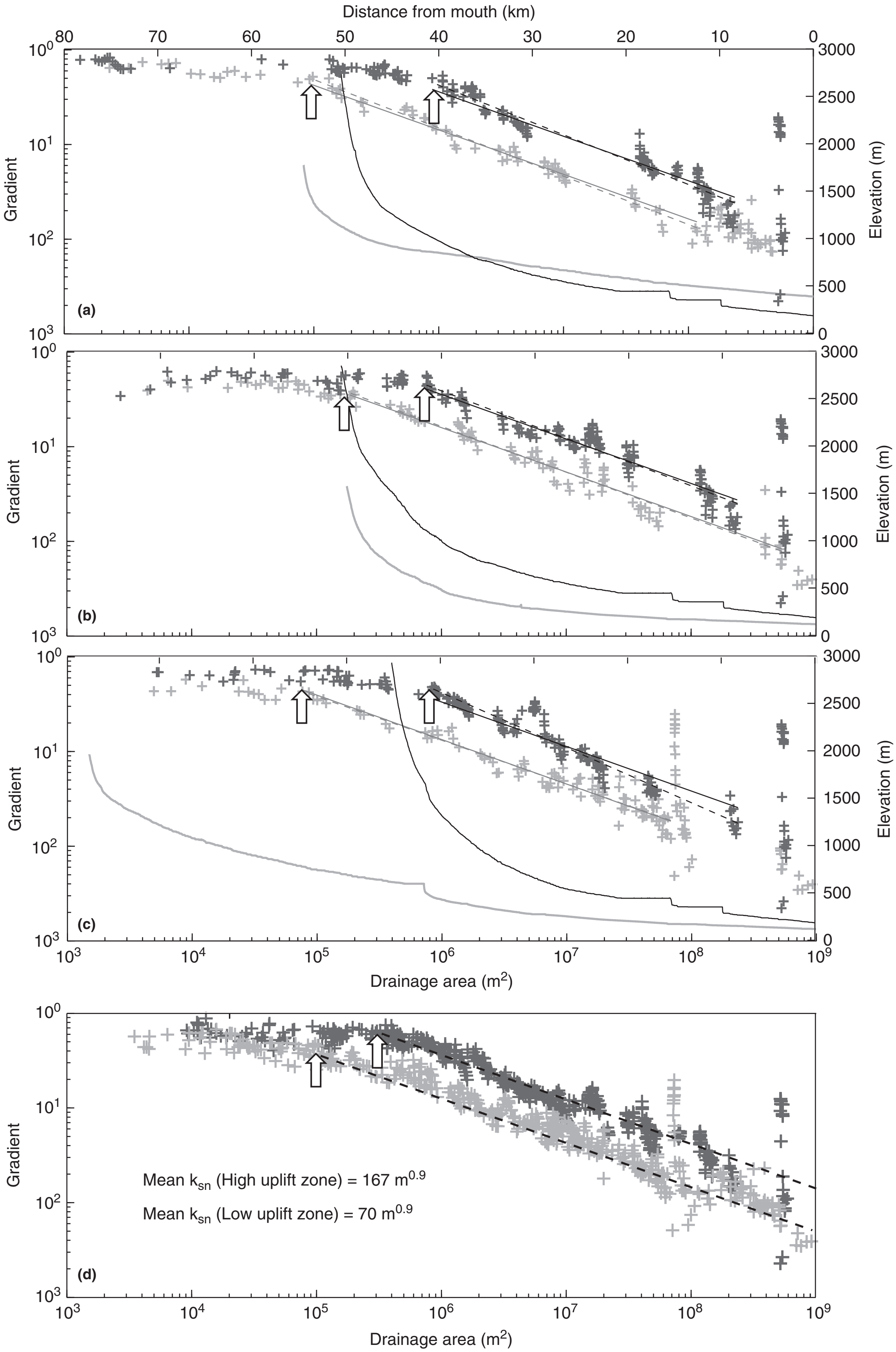
9.28.4.2 Implications for Landscape Relief at Steady State

Bedrock channels carve deep canyons and define the relief structure of mountainous regions. Specifically, it is the longitudinal profiles of bedrock channels that define much of the relief, as dictated by the controls on the channel steepness index discussed above (**Figure 1**). But how do various measures of relief relate to the channel steepness index? Relief (the elevation range within a specified area) is by definition a scale-dependent metric of landscape form. In the next section, we review the relationship between channel steepness and local relief measured at various length scales as documented in a recent study. In the following section, we then summarize the available data on the relation between erosion rate and either channel steepness or local relief at steady state.

9.28.4.2.1 Scales of relief and relation to channel steepness

DiBiase et al. (2010) recently exploited a west–east gradient in relief and mean elevation across the San Gabriel Mountains in southern California to study the relationships among hillslope gradient, channel steepness, and local relief at a range of scales. Although generally discussed as synonymous with hillslope relief, local relief is generally measured over scales of 1–10 km (Ahnert, 1970; Finnegan et al., 2008; Montgomery and Brandon, 2002) and logically must reflect some combination of hillslope gradient and length, tributary steepness and length, trunk stream steepness and length, and basin shape. Which of these factors most strongly influence local relief undoubtedly varies as a function of the scale of observation.

To evaluate the controls on relief at different scales, DiBiase et al. (2010) measured the local relief (elevation range) within circular windows with radii ranging from 0.1 to 5 km centered on every pixel in a 10-m-resolution DEM. They reported that at


Distance from mouth (km)
80 70 60 50 40 30 20 10 0
Gradient
Elevation (m)
(a)
(b)
(c)
(d)
Drainage area (m^2)
Mean k_{sn} (High uplift zone) = 167 m$^{0.9}$
Mean k_{sn} (Low uplift zone) = 70 m$^{0.9}$

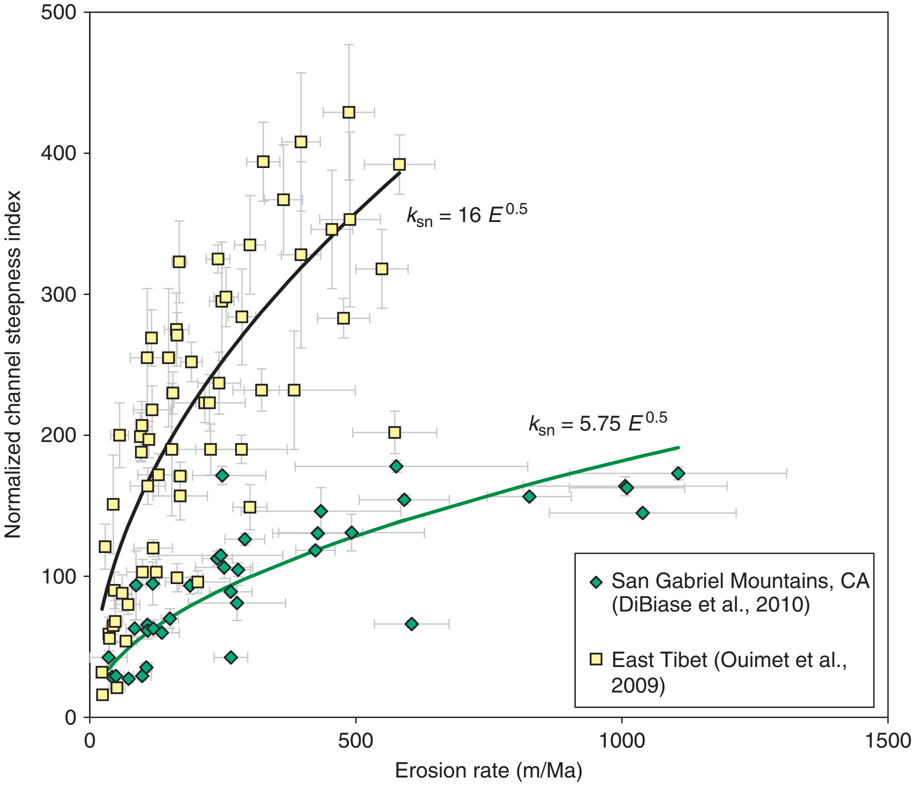


Figure 7 The relationship between the normalized channel steepness index (k_{sn}) and erosion rate (E) as measured by detrital cosmogenic radioisotopes. Data for catchments with well-graded (smooth concave-up river profiles well described by Flint's law, eqn [5]) river profiles in two landscapes. Yellow squares are from the eastern margin of the Tibetan Plateau (Reproduced from Ouimet, W.B., Whipple, K.X., Granger, D.E., 2009. Beyond threshold hills lopes: channel adjustment to base-level fall in tectonically active mountain ranges. Geology 37(7), 579–582.), mean annual precipitation $\sim$0.3 m yr^{-1}. Green diamonds are from the San Gabriel Mountains, mean annual precipitation $\sim$0.75 m yr^{-1}. Reproduced from DiBiase, R.A., Whipple, K.X., Heimsath, A.M., Ouimet, W.B., 2010. Landscape form and millennial erosion rates in the San Gabriel Mountains, CA. Earth and Planetary Science Letters 289(1–2), 134–144.

small scales (radii of $\leq$0.1 km), local relief is essentially an alternate measure of mean hillslope gradient and carries no additional information content. At intermediate scales (radii of 0.25–1.0 km), hillslope gradient, hillslope length, and channel steepness each contribute significantly to local relief. Finally, at large scales (radii of 1.0–2.5 km), local relief is essentially a measure of relief on the channel network, which is directly correlated to the normalized channel steepness index (e.g., Whipple and Tucker, 1999) (**Figure 8**). An important implication is that at the scale over which local relief is generally measured, this metric is effectively a rough proxy for average channel steepness within the measurement window, and is largely independent of hillslope gradient or hillslope relief. Although it has been argued that local relief at the 1–5 km scale saturates at threshold values as erosion rates become extreme (Montgomery and Brandon, 2002), there has been no clear demonstration of such a threshold in any given setting.

9.28.4.2.2 Channel steepness, local relief, and erosion rate

Several recent studies have documented monotonic, positive correlations between erosion rate as measured by detrital cosmogenic radio nuclides in river sands (Bierman and Nichols, 2004; Granger et al., 1996) and normalized channel steepness index (Cyr et al., 2008; DiBiase et al., 2010; Harkins et al., 2007; Ouimet et al., 2009; Safran et al., 2005). Together, these data span much of the known global range of

Figure 6 Example channel profiles exhibiting Flint's Law slope-area scaling and illustrating that although channel steepness (k_{sn}) varies strongly with rock uplift rate (U), the channel concavity index (θ) remains constant. Arrows indicate the critical drainage area (A_{cr}) upstream of which Flint's Law is no longer a good descriptor of channel profile form. Panels (a)–(c) each show pairs of channels from the San Gabriel Mountains (SGM), one channel from the slowly uplifting western SGM and one from the rapidly uplifting eastern SGM. In each panel, channel profiles (elevation vs. distance from mouth) – right and top axes (light gray) – are shown with solid lines (dark gray) and slope versus area – left and bottom axes, log scales – are shown with crosses. Panel (d) is a composite slope-area plot of all six channels, illustrating the factor of $\sim$2 difference in channel steepness index between low and high uplift zones and the constant concavity index ($\theta\sim$0.45) across the range. Modified from Wobus, C.W., Whipple, K.X., Kirby, E., et al., 2006c. Tectonics from topography: procedures, promise, and pitfalls. Geological Society of America Special Paper 398, 55–74, with permission from GSA.

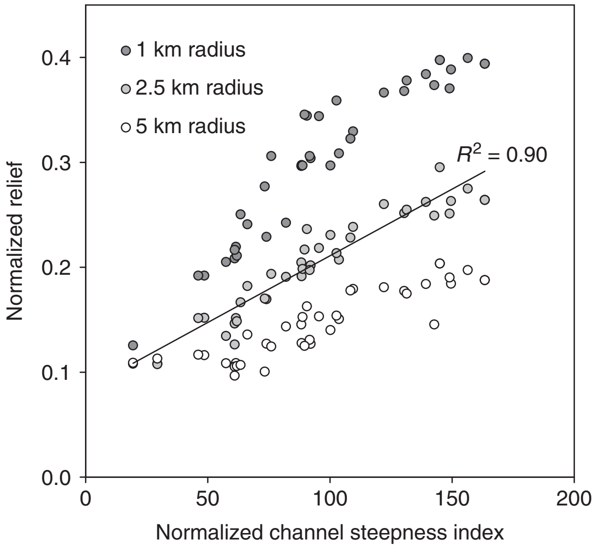


Figure 8 Relationship between local relief and the normalized channel steepness index in the San Gabriel Mountains, CA. Relief is normalized by the length scale over which relief is measured (radius of a circular search window). In this landscape, mean local relief measured within 2.5 km search radii is linearly correlated to the mean channel steepness in the catchment. Over shorter length scales, normalized relief approaches a threshold at high channel steepness values – reflecting an influence of hillslope gradient (and thus hillslope relief) on local relief. Over longer length scales, the linear relationship between normalized relief and channel steepness again breaks down because the search window begins to exceed catchment dimensions. Modified from DiBiase, R.A., Whipple, K.X., Heimsath, A.M., Ouimet, W.B., 2010. Landscape form and millennial erosion rates in the San Gabriel Mountains, CA. Earth and Planetary Science Letters 289(1–2), 134–144, with permission from Wiley.

normalized steepness index in graded or approximately steady-state channel profiles (see Whipple, 2004). Where study catchments were carefully selected to include only well-graded channel profiles and the data span a wide range of erosion rates, the observed relationship is nonlinear with an approximately power-law form ($k_{sn} \sim E^{1/2}$) (DiBiase et al., 2010; Ouimet et al., 2009) (**Figure 7**). As discussed by DiBiase et al. (2010), the form of these relationships is broadly consistent with predictions of the stochastic-threshold river incision models of Tucker (2004) and Lague et al. (2005). The growing body of data on the relationship between channel steepness and erosion rate has promise for testing the predictive capability of competing river incision models under steady-state conditions. However, as discussed in the context of the controls on channel width, it is important to recognize that the transient evolution of river profiles in response to changing tectonic, base level, or climatic conditions may well be quite distinct from expected behavior at steady state and is likely to provide the most stringent test of the predictive capability of river incision models (e.g., Valla et al., 2010; Whipple and Tucker, 2002).

9.28.4.3 Longitudinal River Profiles – Transient Evolution

Above, we employed the theoretical construct of an equilibrium longitudinal river profile to suggest that the form of a bedrock river can be adjusted to the combined, steady influence of tectonics, lithology, and climate. As we show in this section, this equilibrium construct also provides a foundation for interpretation of channel profiles away from equilibrium – channels that are undergoing a transient response to a change in climatic or tectonic conditions. This understanding of how channel profiles evolve – how different types of knickpoints form and migrate through river networks – is used in Section 9.28.5 to discuss how information about the history of rock uplift can be extracted from channel profiles. Although we focus on the transient adjustment of river profiles and do not explicitly discuss hillslope response, it is important to remember that these adjustments are buffered by two-way feedbacks between channels and hillslopes. For example, channel incision may undercut and destabilize a hillslope, but the increased sediment supply from the resultant landslide may armor the bed and slow channel incision for a time (*see* Chapter 9.15). Negative feedbacks such as this can prolong transient conditions in bedrock rivers.

We describe river response to external forcing as transient because channels (and the landscapes they dissect) experience a finite period of adjustment (the response time) following a change in boundary conditions. Depending on the erosional efficiency of the system, the nature of the perturbation, and system size, the period of transience can persist in the coupled channel/hillslope/orogen system from 10^4–10^6 years (Baldwin et al., 2003; Whipple, 2001). The perturbation that triggers the transient response can generally be attributed to either a change in relative base level or a change in climate. Relative base level can be defined locally by a confluence with a higher order stream, an active structural boundary, or regionally by sea level. Changes in climate modify the timing and magnitude of fluxes of water and coarse sediment through streams, altering the efficiency of river incision. Either can cause changes in river profile form.

The spatial extent of the forcing can be well distributed across a landscape (e.g., a change in temperature or precipitation) or it can be localized along sharp boundaries (e.g., block uplift along a fault or sea-level fall along a coastline with steep bathymetry). The forcing can also be distributed across a gradient or in a nonuniform manner (e.g., anticlinal folding, uplift of a rotating block, or isostatic rebound). The duration of the forcing can be characterized as discrete, persistent, or cyclic. Discrete forcings such as stream capture or fault rupture temporarily perturb the system, generating a pulse-like deviation from its initial state. Persistent forcings, such as the development of a new tectonic stress regime or fault system or a change to a new climate state, drive the system toward a new steady state that can be distinctly different from the initial state (e.g., Bonnet and Crave, 2003; Tucker and Whipple, 2002). When the forcing is cyclic, the system response depends on the period of the forcing relative to the system response time. If the period is longer than system response time, the system oscillates between different stable states; if the forcing period is less than the system response time, the system remains in a constant state of disequilibrium. However, where the forcing cycle is much shorter than the system response time, the channel profile will approximate a steady-state profile adjusted to the mean forcing (Snyder et al., 2002; Whipple, 2001).

9.28.4.3.1 Transient river profile evolution by knickpoint retreat

Following a disturbance, a signal of adjustment propagates upstream toward the headwaters (Howard, 1994). As the signal propagates up the network, it defines a boundary between downstream regions that are adjusted or adjusting to the new forcing and upstream relict regions that retain topographic characteristics adjusted to the initial or background forcing (Berlin and Anderson, 2007; Crosby and Whipple, 2006; Reinhardt et al., 2007; Schoenbohm et al., 2004). In transient bedrock rivers, we define this discrete, mobile boundary between the adjusting and relict regions as a knickpoint – the form of the knickpoint that develops depends on the nature of the forcing perturbation and on the mechanics of river incision.

Following Haviv et al. (2010), we distinguish two end-member knickpoint morphologies: vertical-step knickpoints and slope-break knickpoints (**Figure 9**). Both types of knickpoint can be either mobile or anchored in place (**Figure 9**) and occur in transient, quasi-steady-state, and steady-state landscapes, as is discussed in Section 9.28.4.3.2. Both vertical-step and slope-break knickpoints are marked by a distinctive change in channel gradient. Vertical-step knickpoints are defined by a local, discrete increase in channel gradient that can range in height from a few to a few hundred meters and in extent from an abrupt vertical waterfall with an undercut plunge pool to a sequence of cascades and are readily recognized as spikes in slope-distance or slope-area plots (e.g., Goldrick and Bishop, 2007). Slope-break knickpoints, on the other hand, are defined by a persistent, longitudinally extensive change in channel gradient. In other words, slope-break knickpoints separate channel reaches with different channel steepness (k_s) values and are readily recognized on both longitudinal profiles and slope-area plots (e.g., Wobus et al., 2006c). Under some circumstances, vertical-step knickpoints (or waterfalls) develop immediately below slope-break knickpoints at the upstream end of the steepened reach as a consequence of feedbacks in the river incision process or as a

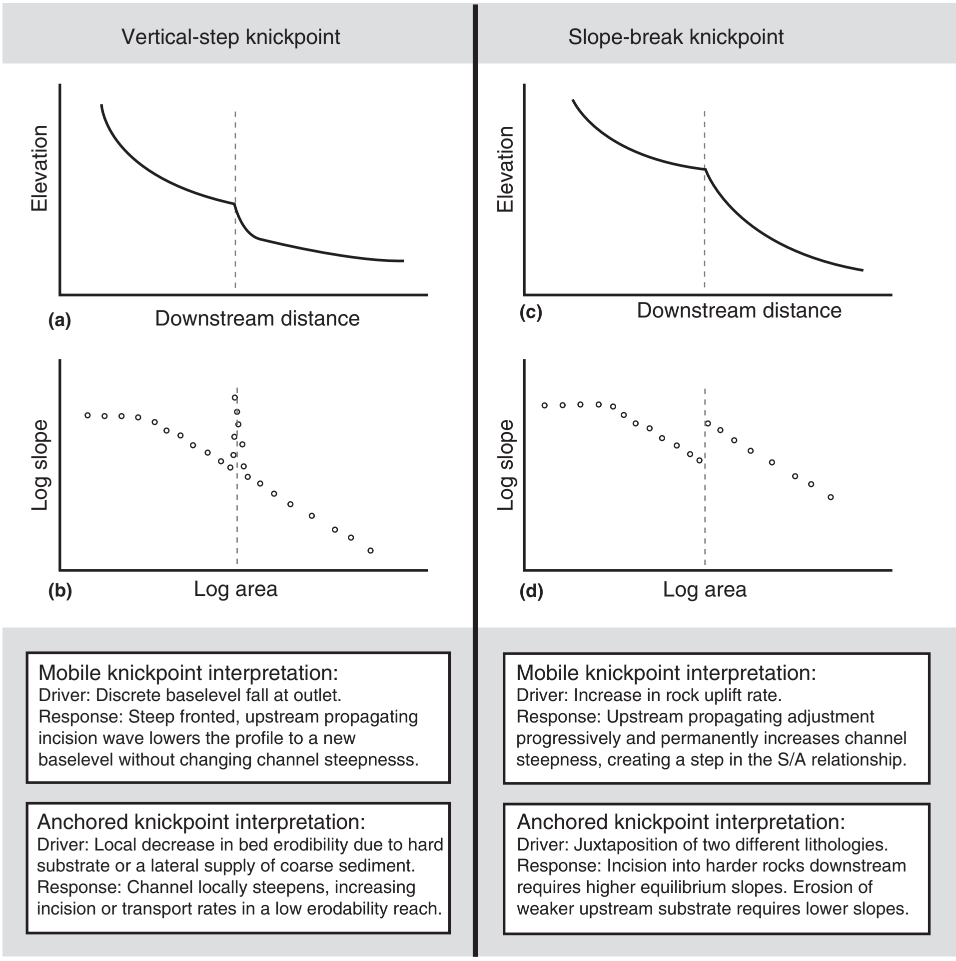


Figure 9 The term 'knickpoint' is commonly used to describe two distinct landform elements, causing some confusion. We follow Haviv et al. (2010) in suggesting a refined terminology to distinguish between vertical-step knickpoints (panels (a) and (b)) and slope-break knickpoints (panels (c) and (d)), here illustrated both in terms of channel profile form (panels (a) and (c)) and slope-area relations (panels (b) and (d)). Further, as discussed in the text, both types of knickpoint may be either anchored or mobile (migrating upstream) landforms. The lower panels give nonunique examples of circumstances that produce both anchored and mobile knickpoints of both vertical-step and slope-break forms (see text for details).

result of spatial variation in substrate properties (e.g., Berlin and Anderson, 2007; Haviv et al., 2006) (**Figure 10**).

Mobile vertical-step knickpoints in transient landscapes form in response to discrete, finite drops in relative base level, as might be caused by stream capture, sea-level fall (but see Snyder et al., 2002), or a pulse of rock uplift. Once formed, mobile vertical-step knickpoints migrate upstream because the steeper local gradient creates greater stresses and potential for abrasion, plucking, and undercutting of bedrock (see Section 9.28.3). As the knickpoint propagates upstream through the channel network, it lowers the base level at tributary junctions, thus forming new knickpoints in each tributary (e.g., Bishop et al., 2005; Crosby and Whipple, 2006; Goldrick and Bishop, 2007).

Vertical-step knickpoint retreat rates are suggested to be a function of water and sediment flux (Berlin and Anderson, 2007; Bishop et al., 2005; Crosby and Whipple, 2006; Crosby et al., 2007; Lamb et al., 2007; Reinhardt et al., 2007) as well as knickpoint morphology (Berlin and Anderson, 2009; Haviv et al., 2010; Haviv et al., 2006) and substrate character (Lamb and Dietrich, 2009). The evolution of vertical-step knickpoint form depends on the rates of erosion at the base, the face, and the lip of knickpoint (Gardner, 1983; Haviv et al., 2010). If rates are highest at the lip, the slope of the step will decay through time. If the rates are highest at the face and at the base, then the knickpoint will experience parallel retreat where the form of the step does not change through time. Because of slow rates of retreat, knickpoint mobility is commonly inferred using downstream evidence of recent retreat (Frankel et al., 2007). Corroborating evidence comes from strath terrace elevations that correlate with height of the knickpoint lip, but definitive confirmation depends on patterns of incision rate data (e.g., Reusser et al., 2004).

In contrast to vertical-step knickpoints, mobile slope-break knickpoints develop in response to a persistent change in boundary conditions such as an increase or decrease in the rate of relative base-level fall (commonly set by rock uplift rate) or a change in climate that enhances or reduces the efficiency of river incision. As a consequence of this different mode of formation, it is the river profile upstream of slope-break knickpoints that undergoes the most rapid change relative to base level – the imbalance between river incision and the rate of relative base-level fall (or rock uplift) is greatest upstream of the knickpoint. It is this imbalance upstream of the knickpoint that drives landscape evolution and upstream migration of the slope-break knickpoint (e.g., Whipple and Tucker, 1999). Although mobile slope-break knickpoints involving a downstream increase in gradient (and thus incision rate) are most readily recognized and most likely preserved as discrete knickpoints, mobile slope-break knickpoints involving a downstream decrease in gradient also occur and behave similarly (e.g., Baldwin et al., 2003; Hilley and Arrowsmith, 2008; Whipple, 2001).

The migration of either type of mobile knickpoint through a river network can modify multiple attributes of a bedrock channel including gradient, width, erosion processes, bed morphology, and bed cover characteristics such as grain size and percent bedrock exposure (Sklar and Dietrich, 2006; Turowski et al., 2007; Whittaker et al., 2007a, 2007b). During the transmission of the transient signal, narrowing and

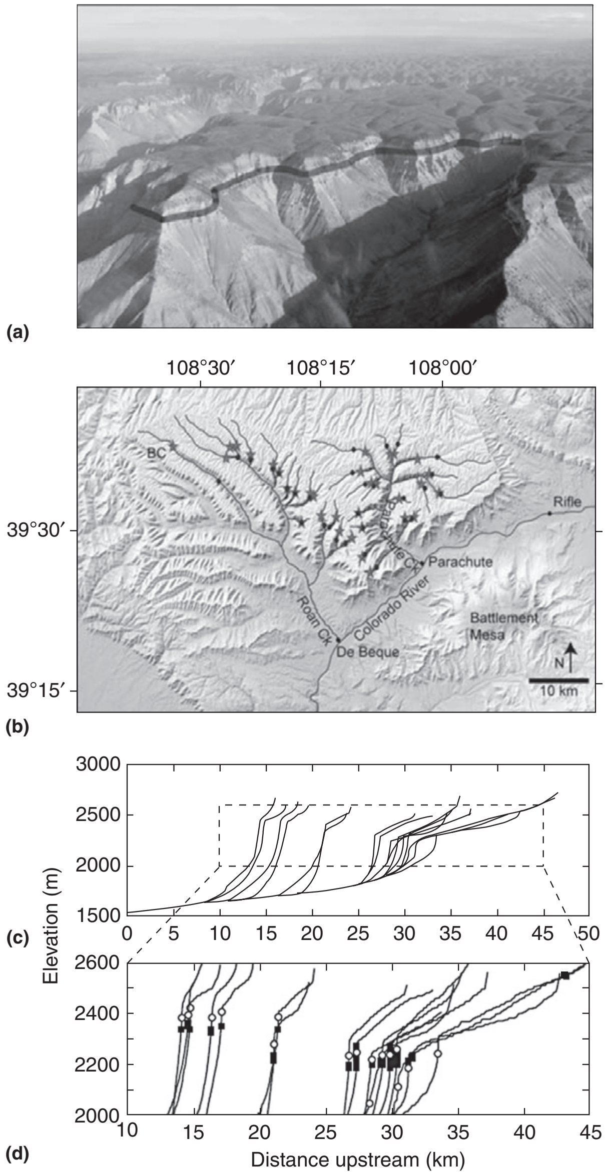


Figure 10 Example of transient landscape responding to an increase in rock uplift rate relative to base level (Roan Plateau, base level set by the Colorado River). (a) Aerial view of Parachute Creek catchment, view to the NW; grey band indicates location of the Mahogany oil-shale zone. (b) Shaded relief image of the Roan Plateau from 10 m DEM. Stars indicate slope-break knickpoint positions, dots indicate knickpoint positions predicted by a simple model. (c) Profiles of mainstem and tributaries to Parachute Creek showing dramatic increase in channel slope downstream of the slope-break knickpoints. (d) Zoomed view of detail of upper part of stream profiles including the slope-break knickpoints with open circles marking knickpoint positions and solid squares marking the top of a resistant unit that influences knickpoint position and form. In this landscape, waterfalls (vertical-step knickpoints) have formed immediately downstream of slope-break knickpoints in many channels. Modified from Berlin, M.M., Anderson, R.S., 2007. Modeling of knickpoint retreat on the Roan Plateau, western Colorado. Journal of Geophysical Research – Earth Surface 112, F03S06, with permission from AGU.

increased bed exposure can accompany the gradient response, but these adjustments are characteristically restricted to channel segments in the direct vicinity of the knickpoint (see also Section 9.28.2.2). As discussed below, the pattern of river transient response is most diagnostic of different river incision models. Even where more than one model can explain the variation in landscape form at steady state (such as in **Figure 7**), these models are likely to make divergent predictions of channel profile evolution during periods of transient adjustment (e.g., Crosby et al., 2007; Gasparini et al., 2007; Howard et al., 1994; Whipple and Tucker, 2002). Moreover, the spatial distribution of transient slope-break knickpoints in the landscape is essential to interpretation of tectonic and climatic histories encoded in landforms (addressed in Section 9.28.5).

9.28.4.3.2 Knickpoints in steady landscapes

Care must be taken when interpreting the origin of steps or slope-breaks in longitudinal profiles because these forms occur in both transient and equilibrium landscapes. As highlighted in **Figure 9**, both knickpoint types can be either mobile – moving freely through the drainage network – or anchored in space by lithologic, structural, or drainage network boundaries. Although mobile knickpoints are a key aspect of all transient landscapes and anchored knickpoints are generally associated with steady- or quasi-steady-state landscapes, there is no unique correspondence between knickpoint activities and transient landscapes. For example, migrating transient knickpoints of either form can, under certain circumstances, trigger the formation of hanging tributaries (Crosby and Whipple, 2006; Crosby et al., 2007; Goode and Burbank, 2009; Wobus et al., 2006a), effectively leaving some knickpoints anchored to tributary junctions for extended periods of time. Fluvial hanging valleys are at once one of the most dramatic examples of disequilibrium landforms that clearly record a change in climatic or tectonic forcing and yet are effectively immobile, anchored in place at tributary junctions.

The position and form of more commonplace anchored knickpoints may reflect either spatial patterns of rock uplift or substrate properties (**Figure 9**). For example, both vertical-step and slope-break knickpoints can form where a river traverses a change in substrate resistance; if the contact between stronger and weaker rock is subvertical, the knickpoint will be both anchored to the outcropping of stronger rock and effectively immobile. If the resistant rock outcrop is laterally extensive, a slope-break knickpoint forms, otherwise an isolated steep reach more akin to a vertical-step knickpoint develops (**Figure 9**). For example, vertical-step knickpoints or localized knickzones can develop on, and will be anchored to, isolated patches of resistant rock (such as a granitic body or a dike intruded into weak sedimentary rocks). Where beds are nonvertical, knickpoints anchored to resistant units will naturally migrate upstream or downstream at a rate dictated by the slope of the river, the dip of the contact between stronger and weaker rock, and the rate of incision. For example, in bedrock streams with horizontally stratified rock, mobile vertical-step knickpoints may be the dominant erosion process as layers of rock are successively plucked away or undermined (Miller, 1991), even under long-term steady-state conditions.

As suggested above, immobile slope-break knickpoints are generally anchored to active structures (see **Figure 11**). For example, if a river flows from a headwater region with high (or low) rock uplift rate across a fault into an area with a lower (or higher) rock uplift rate, the channel gradient will decrease (or increase) abruptly on the downstream side of the fault, forming a slope-break knickpoint (e.g., Wobus et al., 2003). This local concavity (or convexity) in the channel profile persists in space and time as long as the fault location and juxtaposed uplift rates remain constant (**Figures 9** and **11(c)**). Should tectonic activity on such a structure cease, the previously anchored knickpoint would be released to sweep upstream as the landscape enters a period of transient adjustment to the change in tectonic forcing.

9.28.5 Tectonic Interpretation of River Profiles

Relations between steady-state channel steepness, local relief, and rock uplift rate relative to base level (Sections 9.28.4.1 and 9.28.4.2) and expected river response to changes in rock uplift rate relative to base level (Section 9.28.4.3) can be effectively used to diagnose both spatial and temporal patterns in rock uplift rate (e.g., Wobus et al., 2006c). The methods outlined below, however, must be applied with caution and are best considered reconnaissance tools to gauge relative rates of rock uplift. A core reason for this limitation is that the controls on the quantitative relationship between channel steepness and rock uplift rate (**Figure 7**) are complex and not yet fully understood and are likely distinct under steady-state and various transient conditions (e.g., climate change vs. tectonic change or increasing vs. decreasing rock uplift rate) (see also Whipple, 2004). In addition, a great many factors not related to rock uplift can cause local perturbations to river profiles (variation in rock properties, landslides or other sources of enhanced sediment delivery, and DEM artifacts); river profiles are thus most reliably interpretable at long wavelength (several kilometers or more) (Wobus et al., 2006c). Moreover, care must be taken to differentiate among lithologic effects (e.g., Duvall et al., 2004), climatic effects (e.g., Craddock et al., 2007; Roe et al., 2002), and both spatial and temporal tectonics effects (e.g., Whittaker et al., 2007a; Wobus et al., 2006c).

Spatial variability in channel steepness within a landscape may reflect either spatial differences in rock properties, abundance and size of bedload, climate (e.g., orographic rainfall patterns), and spatial variations in rock uplift rate, or may reflect temporal changes in climate or tectonics. Spatial patterns in rock properties, sediment characteristics, and climate can usually be directly measured to assess whether these correlate with spatial patterns in channel steepness. Under many circumstances, spatial and temporal changes in tectonics can be anticipated to have distinct spatial signatures, and thus can often be distinguished as well (**Figure 11**).

Spatial changes in rock uplift rate may be abrupt (differential uplift across an active fault) or may be gradational (folding of an anticline, rotation of a fault block). In either

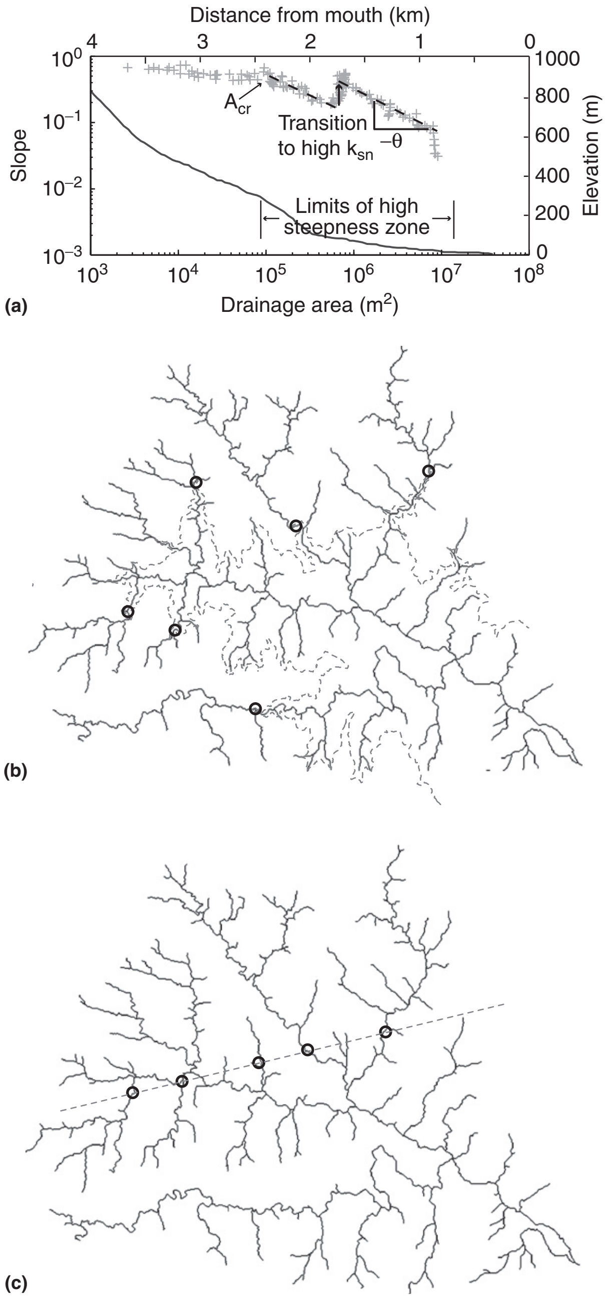


Figure 11 Illustration of contrasting expected patterns of channel steepness index and slope-break knickpoint locations (shown by circles) for spatial and temporal changes in rock uplift rate. (a) Typical slope-area diagram for a channel profile with a distinct slope-break knickpoint. The spatial distribution across the landscape of the high and low channel steepness zones, and the slope-break knickpoints separating them, can help distinguish spatial from temporal differences in rock uplift rate as illustrated in panels (b) and (c). (b) Temporal change in rock uplift rate causes slope-break knickpoints to propagate through the landscape at a constant vertical rate, so knickpoints are expected to lie on a contour – see also **Figure 10**. (c) Abrupt spatial differences in rock uplift rate occur on faults or sharp folds and thus usually cut across contours following structural trends. Modified from Wobus, C.W., Whipple, K.X., Kirby, E., et al., 2006c. Tectonics from topography: procedures, promise, and pitfalls. Geological Society of America Special Paper 398, 55–74, with permission from GSA.

case, the observed changes in channel steepness should have simple plan-view patterns such as slope-break knickpoints aligned along fault traces or smooth regional patterns of the channel steepness index (e.g., gradually increasing toward a fault trace) (**Figure 11**). But even in the case of abrupt changes in rock uplift rate, the channel steepness response may be somewhat diffuse. For example, in cases with a downstream decrease in rock uplift rate, it is normal to expect channel steepness to gradually decline downstream of the fault, defining a high-concavity channel segment that may extend for several kilometers (Whipple and Tucker, 2002). Moreover, corroborating evidence is required to be confident that a spatial pattern in channel steepness reflects a spatial pattern in rock uplift rate. Typically, this corroborating evidence would be differences in channel incision or catchment-averaged erosion rate (whether measured over decadal timescales or over millions of years) and/or independent evidence of active deformation.

Temporal changes in rock uplift rate, however, also produce spatial patterns in channel steepness and erosion rate (**Figures 10** and **11**). As discussed earlier, persistent changes in rock uplift rate generate upstream-migrating slope-break knickpoints that mark the boundary between adjusted or adjusting channel segments downstream (adjusting channel geometry, including gradient, bed state, and incision rate to the new tectonic conditions) and relict channel segments upstream (with channel characteristics and incision rate that still reflect the previous tectonic conditions). Thus, patterns of channel steepness and incision rate will be strongly correlated in a landscape responding to a change in tectonics as well. Fortunately, upstream-migrating slope-break knickpoints sweep through the landscape at predictable rates and can be expected to form easily recognizable patterns (Attal et al., 2008; Berlin and Anderson, 2007; Bishop et al., 2005; Crosby and Whipple, 2006; Goldrick and Bishop, 2007; Wobus et al., 2006c) (**Figure 11**).

Following Rosenbloom and Anderson (1994), Whipple and Tucker (1999) showed that the stream-power incision model (Section 9.28.3.6, eqn [3]) predicts a strong drainage-area dependence on the plan-view knickpoint migration rate. More importantly, however, Niemann et al. (2001) used geometrical arguments to demonstrate that slope-break knickpoints migrate at a constant vertical rate for any channel incision rule that yields steady-state channel profiles well described by Flints Law (Section 9.28.4, eqn [5]) – where the channel steepness index varies with rock uplift rate but the concavity index does not (consistent with observations, **Figure 6**) – and transient response is characterized by slope-break knickpoints that separate channel segments above in equilibrium with prior conditions and channel segments below in approximate equilibrium with current conditions (see also Wobus et al., 2006c). This means that a suite of slope-break knickpoints that represent a transient landscape response to a persistent change in rock uplift rate should lie at a consistent elevation above base level (i.e., the elevation of the main stem river where it crosses the fault). Within a given drainage basin, this means that slope-break knickpoints should lie at approximately the same topographic contour (**Figures 10** and **11**). Such a pattern has recently been recognized in a number of different landscapes (Berlin and Anderson, 2007;

Clark et al., 2005; Cook et al., 2009; Reinhardt et al., 2007; Schoenbohm et al., 2004; Wobus et al., 2006c). Deviation from this idealized channel profile response will result in a dispersion of knickpoint elevations. For example, under extreme circumstances, this expected pattern of response breaks down and fluvial hanging valleys form instead of upstream-migrating slope-break knickpoints (Crosby et al., 2007).

Fluvial hanging valleys have been defined as tributaries with vertical-step or oversteepened slope-break knickpoints at or near their confluence with the mainstem river and form when mainstem incision outpaces the tributary response (Crosby et al., 2007; Goode and Burbank, 2009; Wobus et al., 2006a). They occur most commonly in small tributaries to large mainstem rivers (with drainage area an order of magnitude or more greater than the tributary) in landscapes where mainstem incision rates have recently greatly accelerated. Crosby et al. (2007) used a sediment-flux-dependent river incision model (Section 9.28.3.6, eqns [3] and [4]) to explore the conditions under which fluvial hanging valleys may form. They showed that where channel gradients in tributaries increase beyond a threshold, the efficiency of river incision by bedload abrasion decreases with increasing slope (Sklar and Dietrich, 2004), triggering a positive feedback where the knickpoint becomes taller and steeper, eventually forming a hanging valley. Continued retreat of the oversteepend hanging-valley knickpoint slows and apparently relies on slower mass-wasting processes (Crosby and Whipple, 2006; Haviv et al., 2010; Weissel and Seidl, 1998). Thus, the formation of hanging valleys delays landscape response, extending the duration of transience, and disrupts expected patterns of transient landscape evolution. However, fluvial hanging valleys do appear to be a reliable indicator of a rapid pulse of accelerated river incision.

As a practical matter, analyses of transient longitudinal profiles aimed at recovering a record of the tectonic forcing focus on channel segments sufficiently upstream and downstream of the slope-break knickpoint. This is done to avoid local oversteepenings associated with poor DEM resolution in narrow canyons (e.g., Wobus et al., 2006c), the leaking of signals across knickpoints (Berlin and Anderson, 2009; Haviv et al., 2006, 2010), the covariation of gradient with width and bed cover (e.g., Sklar and Dietrich, 2006; Turowski et al., 2009), and the sporadic development of vertical-step knickpoints and/or hanging valleys in the wake of slope-break knickpoints (Crosby et al., 2007). This approach ensures that the segments best represent the gradients in the relict and adjusting portions of the landscape, and thus come closest to recording channel adjustment to the initial and final tectonic conditions.

9.28.6 Concluding Remarks

Bedrock channels, defined as either channels exposing a significant fraction of rock in their bed and bank or channels that are actively incising through rock, play a critical role in landscape evolution and in the interaction between climate and tectonics. Bedrock channels define much of the relief structure of mountain ranges (**Figure 1**). They convey signals of climate change and tectonic forcing through the landscape, setting landscape response time. The incision rate on bedrock channels sets the lower boundary condition (local base level) for all hillslopes. Together with controls on channel slope, the link between channel incision and hillslope response dictates the relationship between topography and erosion rate, or between topography and rock uplift rate at steady state (**Figure 7**). Although there are some unique flow hydraulic and sediment transport characteristics of rock-bound reaches, most bedrock channels have at least a thin, patchy alluvial mantle and thus have much in common with alluvial channels, particularly those occurring in steeplands (e.g., Wohl and David, 2008; Wohl and Merritt, 2008). Perhaps most interesting, the width-area scaling of bedrock channels is indistinguishable from gravel-bedded alluvial channels, emphasizing that bedrock channels are indeed self formed and hinting that the role of bank strength in controlling channel width may be much weaker than generally thought (**Figure 2**). Details of flow, channel morphology, bed-state, and sediment transport are covered in detail in other chapters of this volume (as noted in Section 9.28.1). Consequently, this chapter focuses on the role of bedrock channels in landscape evolution, highlighting controls on channel form, landscape relief, and the relation between landscape form and rock uplift rate both at steady state and during transient response to a change in tectonics. This knowledge is used to develop some general guidelines for drawing tectonic interpretations from landscape form (**Figure 11**).

Taken as a whole, this review of bedrock channel process and form has highlighted a number of critical issues that merit further study. In regard to the signature of tectonic conditions and history preserved in landscape form and the strength of coupling between climate and tectonics, the essential questions include: (1) what factors control the relationship between channel steepness (or local relief, see **Figure 8**) and erosion rate? (2) how variable is this relationship with differences in substrate lithology or climate? (3) how different is this relationship at steady state compared to, during transient adjustment, a change in climate or tectonics? and (4) does the nature of the perturbation (e.g., climate change, increase in rock uplift rate, and decrease in rock uplift rate) make a difference? The lack of quantitative empirical knowledge of the relations among climate variables (such as mean annual precipitation, temperature, runoff, seasonality, and storminess) and erosional efficiency is surprising. Answering these questions requires refined knowledge of the suite of processes that contribute to river incision into bedrock and the factors that determine which process(es) is/are dominant. For instance, we must resolve how important are the details of the influence of sediment load on the efficiency of river incision (form of the tools and cover terms, role of grain size, relative importance of suspended load and bedload). In addition, we must resolve what dictates the nonlinearity of the relation between channel steepness and erosion rate (**Figure 7**) – what are the relative influences of: (1) the mechanics of the incision process, (2) channel narrowing at high incision rates, (3) the diminishing importance of a threshold shear stress in steeper channels, (4) the probability distribution of flood discharges (e.g., exponential vs. power-law), and (5) greater frequency of debris flows in areas with higher erosion rates and steeper channels?

Exploration of the implications of various river incision models through numerical simulation, laboratory experimentation, and field studies of transient landscape response to tectonic and climatic perturbation each have strong potential for advancing understanding and resolving these critical questions.

References

Aalto, R., Dunne, T., Guyot, J.L., 2006. Geomorphic controls on Andean denudation rates. Journal of Geology 114, 85–99.
Ahnert, F., 1970. Functional relationships between denudation, relief, and uplift in large mid-latitude drainage basins. American Journal of Science 268, 243–263.
Amos, C.B., Burbank, D.W., 2007. Channel width response to differential uplift. Journal of Geophysical Research 112, F02010, doi:10.1029/2006JF000672.
Arndt, R.E.A., 1981. Cavitation in fluid machinery and hydraulic structures. Annual Reviews of Fluid Mechanics 13, 273–328.
Attal, M., Tucker, G.E., Whittaker, A.C., Cowie, P.A., Roberts, G.P., 2008. Modeling fluvial incision and transient landscape evolution. Influence of dynamic channel adjustment. Journal of Geophysical Research – Earth Surface 113(F3), F03013.
Baldwin, J.A., Whipple, K.W., Tucker, G.E., 2003. Implications of the shear stress river incision model for the timescale of postorogenic decay of topography. Journal of Geophysical Research 108(B3), http://dx.doi.org/10.1029/2001JB000550.
Barnes, H.L., 1956. Cavitation as a geological agent. American Journal of Science 254, 493–505.
Benda, L., Dunne, T., 1997. Stochastic forcing of sediment routing and storage in channel networks. Water Resources Research 33(12), 2865–2880.
Berlin, M.M., Anderson, R.S., 2007. Modeling of knickpoint retreat on the Roan Plateau, western Colorado. Journal of Geophysical Research – Earth Surface 112, F03S06.
Berlin, M.M., Anderson, R.S., 2009. Steepened channels upstream of knickpoints: controls on relict landscape response. Journal of Geophysical Research – Earth Surface 114, F03018.
Bierman, P.R., Nichols, K.K., 2004. Rock to sediment – slope to sea with 10Be – rates of landscape change. Annual Review of Earth and Planetary Sciences 32, 215–235.
Bishop, P., Hoey, T.B., Jansen, J.D., Artza, I.L., 2005. Knickpoint recession rate and catchment area: the case of uplifted rivers in Eastern Scotland. Earth Surface Processes and Landforms 30(6), 767–778.
Bonnet, S., Crave, A., 2003. Landscape response to climate change: insights from experimental modeling and implications for tectonics versus climatic uplift of topography. Geology 31(2), 123–126.
Chamberlain, C.P., Hilley, G.E., Hren, M.T., Porder, S., Jacobson, A.D., 2007. Steady state chemical weathering in an eroding landscape. Geochimica et Cosmochimica Acta 71(15), A158.
Chatanantavet, P., Parker, G., 2008. Experimental study of bedrock channel alluviation under varied sediment supply and hydraulic conditions. Water Resources Research 44(12), http://dx.doi.org/10.1029/2007WR006581.
Chatanantavet, P., Parker, G., 2009. Physically based modeling of bedrock incision by abrasion, plucking, and macroabrasion. Journal of Geophysical Research – Earth Surface 114, F04018.
Clark, M.K., Maheo, G., Saleeby, J., Farley, K.A., 2005. The non-equilibrium landscape of the southern Sierra Nevada, California. GSA Today 15(9), 4–10.
Cook, K.L., Whipple, K., Heimsath, A., Hanks, T.C., 2009. Rapid incision of the Colorado River in Glen Canyon – insights from channel profiles, local incision rates, and modeling of lithologic controls. Earth Surface Processes and Landforms 34(7), 994–1010.
Cowie, P.A., Whittaker, A.C., Attal, M., Roberts, G., Tucker, G.E., Ganas, A., 2008. New constraints on sediment-flux-dependent river incision: Implications for extracting tectonic signals from river profiles. Geology 36(7), 535–538.
Craddock, W.H., Burbank, D.W., Bookhagen, B., Gabet, E.J., 2007. Bedrock channel geometry along an orographic rainfall gradient in the upper Marsyandi River valley in central Nepal. Journal of Geophysical Research – Earth Surface 112, F03007.
Crosby, B.T., 2006. The transient response of bedrock river networks to sudden base level fall. PhD dissertation, MIT, Cambridge, MA, 144 pp.
Crosby, B.T., Whipple, K.X., 2006. Knickpoint initiation and distribution within fluvial networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand. Geomorphology 82, 16–38.
Crosby, B.T., Whipple, K.X., Gasparini, N.M., Wobus, C.W., 2007. Formation of fluvial hanging valleys: theory and simulation. Journal of Geophysical Research – Earth Surface 112, F03S10.
Cyr, A.J., Olivetti, V., Granger, D., Molin, P., Faccenna, C., 2008. Comparing the spatial variability of cosmogenic 10Be erosion rates and channel steepness to Quaternary Uplift rates in northern and southern Italy. Geological Society of America, Abstracts with Programs 40(6), 431.
DiBiase, R., Whipple, K., Heimsath, A., 2009. Channel incision and the role of sediment supply in the San Gabriel Mountains, California. Eos Transactions of the American Geophysical Union 90, EP54A–02.
DiBiase, R.A., Whipple, K.X., Heimsath, A.M., Ouimet, W.B., 2010. Landscape form and millennial erosion rates in the San Gabriel Mountains, CA. Earth and Planetary Science Letters 289(1–2), 134–144.
Dixon, J.L., Heimsath, A.M., Amundson, R., 2009. The critical role of climate and saprolite weathering in landscape evolution. Earth Surface Processes and Landforms 34(11), 1507–1521.
Duvall, A., Kirby, E., Burbank, D., 2004. Tectonic and lithologic controls on bedrock channel profiles and processes in coastal California. Journal of Geophysical Research 109, F03002. http://dx.doi.org/10.1029/2003JF000086.
Ferguson, R., 2007. Flow resistance equations for gravel- and boulder-bed streams. Water Resources Research 43, W05427.
Ferrier, K.L., Kirchner, J.W., 2008. Effects of physical erosion on chemical denudation rates: a numerical modeling study of soil-mantled hillslopes. Earth and Planetary Science Letters 272(3–4), 591–599.
Finnegan, N.J., Hallet, B., Montgomery, D.R., Zeitler, P.K., Stone, J.O., Anders, A.M., Yuping, L., 2008. Coupling of rock uplift and river incision in the Namche Barwa-Gyala Peri massif, Tibet. Geological Society of America Bulletin 120(1–2), 142–155.
Finnegan, N.J., Roe, G.H., Montgomery, D.R., Hallet, B., 2005. Controls on the channel width of rivers: implications for modeling fluvial incision of bedrock. Geology (Boulder) 33(3), 229–232.
Finnegan, N.J., Sklar, L.S., Fuller, T.K., 2007. Interplay of sediment supply, river incision, and channel morphology revealed by the transient evolution of an experimental bedrock channel. Journal of Geophysical Research – Earth Surface 112, F03S11.
Flint, J.J., 1974. Stream gradient as a function of order, magnitude, and discharge. Water Resources Research 10, 969–973.
Frankel, K.L., Pazzaglia, F.J., Vaughn, J.D., 2007. Knickpoint evolution in a vertically bedded substrate, upstream-dipping terraces, and Atlantic slope bedrock channels. Geological Society of America Bulletin 119(3–4), 476–486.
Gardner, T.W., 1983. Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material. Geological Society of America Bulletin 94(5), 664–672.
Gasparini, N.M., Whipple, K.X., Bras, R.L., 2007. Predictions of steady state and transient landscape morphology using sediment-flux-dependent river incision models. Journal of Geophysical Research – Earth Surface 112, F03S09.
Gilbert, G.K., 1877. Geology of the Henry Mountains (Utah), Washington, D.C., United States Government Printing Office, U.S. Geographical and Geological Survey of the Rocky Mountain region, 160 pp.
Goldrick, G., Bishop, P., 2007. Regional analysis of bedrock stream long profiles: evaluation of Hack's SL form, and formulation and assessment of an alternative (the DS form). Earth Surface Processes and Landforms 32(5), 649–671.
Goode, J.K., Burbank, D.W., 2009. Numerical study of degradation of fluvial hanging valleys due to climate change. Journal of Geophysical Research – Earth Surface 114, F01017.
Granger, D.E., Kirchner, J.W., Finkel, R., 1996. Spatially averaged long-term erosion rates measured from *In Situ*-produced Cosmogenic Nuclides in alluvial sediment. Journal of Geology 104(3), 249–257.
Hack, J.T., 1957. Studies of longitudinal stream profiles in Virginia and Maryland. U.S. Geological Survey Professional Paper 294-B, 97.
Hack, J.T., 1973. Stream-profile analysis and stream-gradient index. Journal of Research of the U. S. Geological Survey 1(4), 421–429.
Hancock, G.S., Anderson, R.S., Whipple, K.X., 1998. Beyond power: bedrock river incision process and form. In: Tinkler, K., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels, Geophysical Monograph Series. AGU Press, Washington, DC, Volume 107, pp. 35–60.
Harbor, D.J., 1998. Dynamic equilibrium between an active uplift and the Sevier River, Utah. Journal of Geology 106(2), 181–198.
Harkins, N., Kirby, E., Heimsath, A., Robinson, R., Reiser, U., 2007. Transient fluvial incision in the headwaters of the Yellow River, northeastern Tibet, China. Journal of Geophysical Research – Earth Surface 112, F03S04.
Hartshorn, K., Hovius, N., Dade, W.B., Slingerland, R., 2002. Climate-driven bedrock incision in an active mountain belt. Science 297, 2036–2038.

Haviv, I., Enzel, Y., Whipple, K.X., Zilberman, E., Matmon, A., Stone, J., Fifield, K.L., 2010. Evolution of vertical knickpoints (waterfalls) with resistant caprock: insights from numerical modeling. Journal of Geophysical Research – Earth Surface 115, F03028.
Haviv, I., Enzel, Y., Whipple, K.X., Zilberman, E., Stone, J., Matmon, A., Fifield, L.K., 2006. Amplified erosion above waterfalls and oversteepened bedrock reaches. Journal of Geophysical Research – Earth Surface 111, F04004.
Hilley, G.E., Arrowsmith, J.R., 2008. Geomorphic response to uplift along the Dragon's Back pressure ridge, Carrizo Plain, California. Geology (Boulder) 36(5), 367–370.
Howard, A., 1994. A detachment-limited model of drainage basin evolution. Water Resources Research 30(7), 2261–2285.
Howard, A.D., 1998. Long profile development of bedrock channels: interaction of weathering, mass wasting, bed erosion, and sediment transport. In: Tinkler, K., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels, Geophysical Monograph Series, AGU Press, Washington, DC, volume 107, pp. 297–319.
Howard, A.D., Dietrich, W.E., Seidl, M.A., 1994. Modeling fluvial erosion on regional to continental scales. Journal of Geophysical Research – Solid Earth 99(B7), 13971–13986.
Howard, A.D., Kerby, G., 1983. Channel changes in badlands. Geological Society of America Bulletin 94, 739–752.
Hsu, L., Dietrich, W.E., Sklar, L.S., 2008. Experimental study of bedrock erosion by granular flows. Journal of Geophysical Research – Earth Surface 113, F02017.
Johnson, J.P., Whipple, K.X., 2007. Feedbacks between erosion and sediment transport in experimental bedrock channels. Earth Surface Processes and Landforms 32, 1048–1062.
Johnson, J.P., Whipple, K.X., 2010. Evaluating the controls of shear stress, sediment supply, alluvial cover, and channel morphology on experimental bedrock incision rate. Journal of Geophysical Research – Earth Surface 115, F02018.
Johnson, J.P., Whipple, K., Sklar, L., Hanks, T.C., 2009. Transport slopes, sediment cover, and bedrock channel incision in the Henry Mountains, Utah, USA. Journal of Geophysical Research 114, F02014.
Kirby, E., Ouimet, W.B., 2011. Tectonic Geomorphology along the Eastern Margin of Tibet: Insights into the Pattern and Processes or Active Deformation Adjacent to the Sichuan Basin. Geological Society, London, Special Publications, vol. 353, pp. 165–188, doi: 10.1144/SP353.9.
Kirby, E., Whipple, K., 2001. Quantifying differential rock-uplift rates via stream profile analysis. Geology 29, 415–418.
Kirby, E., Whipple, K., Tang, W., Chen, Z., 2003. Distribution of active rock uplift along the eastern margin of the Tibetan Plateau: inferences from bedrock channel longitudinal profiles. Journal of Geophysical Research 108(B4), 2217. http://dx.doi.org/10.1029/2001JB000861.
Kobor, J.S., Roering, J.J., 2004. Systematic variation of bedrock channel gradients in the central Oregon Coast Range: implications for rock uplift and shallow landsliding. Geomorphology 62(3–4), 239–256.
Lague, D., 2010. Reduction of long-term bedrock incision efficiency by short-term alluvial cover intermittency. Journal of Geophysical Research – Earth Surface 115, F02011.
Lague, D., Davy, P., 2003. Constraints on the long-term colluvial erosion law by analyzing slope-area relationships at various tectonic uplift rates in the Siwalik Hills (Nepal). Journal of Geophysical Research 108(B2), 2129. http://dx.doi.org/10.1029/2002JB001893.
Lague, D., Hovius, N., Davy, P., 2005. Discharge, discharge variability, and the bedrock channel profile. Journal of Geophysical Research 110(F04006), http://dx.doi.org/10.1029/2004JF000259.
Lamb, M.P., Dietrich, W.E., 2009. The persistence of waterfalls in fractured rock. Geological Society of America Bulletin 121(7–8), 1123–1134.
Lamb, M.P., Dietrich, W.E., Sklar, L.S., 2008. A model for fluvial bedrock incision by impacting suspended and bed load sediment. Journal of Geophysical Research – Earth Surface 113, F03025.
Lamb, M.P., Howard, A.D., Dietrich, W.E., Perron, J.T., 2007. Formation of amphitheater-headed valleys by waterfall erosion after large-scale slumping on Hawai 'i. GSA Bulletin 119(7/8), 805–822.
Lave, J., Avouac, J.-P., 2001. Fluvial incision and tectonic uplift across the Himalayas of central Nepal. Journal of Geophysical Research 106(B11), 26,561–526,591.
Mackin, J.H., 1948. Concept of the graded river. Geological Society of America Bulletin 101, 1373–1388.
Miller, J.R., 1991. The influence of bedrock geology on knickpoint development and channel-bed degradation along downcutting streams in south-central Indiana. Journal of Geology 99, 591–605.
Moglen, G.E., Bras, R.L., 1995. The effect of spatial heterogeneities on geomorphic expression in a model of basin evolution. Water Resources Research 31(10), 2613–2623.
Molnar, P., 2001. Climate change, flooding in arid environments, and erosion rates. Geology 29(12), 1071–1074.
Molnar, P., Anderson, R.S., Kier, G., Rose, J., 2006. Relationships among probability distributions of stream discharges in floods, climate, bed load transport, and river incision. Journal of Geophysical Research, 111. http://dx.doi.org/10.1029/2005JF000310.
Montgomery, D.R., 2004. Observations on the role of lithology in strath terrace formation and bedrock channel width. American Journal of Science 304(5), 454–476.
Montgomery, D.R., Abbe, T.B., Buffington, J.M., Peterson, N.P., Schmidt, K.M., Stock, J.D., 1996. Distribution of bedrock and alluvial channels in forested mountain drainage basins. Nature 381, 587–589.
Montgomery, D.R., Brandon, M.T., 2002. Topographic controls on erosion rates in tectonically active mountain ranges. Earth and Planetary Science Letters 201(3–4), 481–489.
Montgomery, D.R., Buffington, J.M., 1997. Channel-reach morphology in mountain drainage basins. Geological Society of America, Bulletin 109(5), 596–611.
Montgomery, D.R., Foufoula-Georgiou, E., 1993. Channel network representation using digital elevation models. Water Resources Research 29, 1178–1191.
Montgomery, D.R., Gran, K.B., 2001. Downstream variations in the width of bedrock channels. Water Resources Research 37(6), 1841–1846.
Niemann, J.D., Gasparini, N.M., Tucker, G.E., Bras, R.L., 2001. A quantitative evaluation of Playfair's law and its use in testing long-term stream erosion models. Earth Surface Processes and Landforms 26(12), 1317–1332.
Ouimet, W.B., Whipple, K.X., Granger, D.E., 2009. Beyond threshold hillslopes: channel adjustment to base-level fall in tectonically active mountain ranges. Geology 37(7), 579–582.
Parker, G., 1991. Selective Sorting and Abrasion of River Grave 1. Theory. Journal of Hydraulic Engineering – ASCE 117(2), 131–149.
Parker, G., Wilcock, P.R., Paola, C., Dietrich, W.E., Pitlick, J., 2007. Physical basis for quasi-universal relations describing bankfull hydraulic geometry of single-thread gravel bed rivers. Journal of Geophysical Research – Earth Surface 112, F04005.
Reinhardt, L.J., Bishop, P., Hoey, T.B., Dempster, T.J., Sanderson, D.C.W., 2007. Quantification of the transient response to base-level fall in a small mountain catchment: Sierra Nevada, southern Spain. Journal of Geophysical Research – Earth Surface 112, F03S05.
Reusser, L.J., Bierman, P.R., Pavich, M.J., Zen, E.A., Larsen, J., Finkel, R., 2004. Rapid late Pleistocene incision of Atlantic passive-margin river gorges. Science 305(5683), 499–502.
Richardson, K., Carling, P.A., 2005. A Typology of sculpted forms in open bedrock channels. Geological Society America Special Paper 392, 109.
Riebe, C.S., Kirchner, J.W., Finkel, R.C., 2004. Erosional and climatic effects on long-term chemical weathering rates in granitic landscapes spanning diverse climate regimes. Earth and Planetary Science Letters 224, 547–562.
Roe, G.H., Montgomery, D.R., Hallet, B., 2002. Effects of orographic precipitation variations on the concavity of steady-state river profiles. Geology (Boulder) 30(2), 143–146.
Rosenbloom, N.A., Anderson, R.S., 1994. Hillslope and channel evolution in a marine terraced landscape, Santa Cruz, California. Journal of Geophysical Research 99(B7), 14,013–014,029.
Safran, E.B., Bierman, P.R., Aalto, R., Dunne, T., Whipple, K., Caffee, M., 2005. Erosion rates driven by channel network incision in the Bolivian Andes. Earth Surface Processes and Landforms 30, 1007–1024.
Schoenbohm, L.M., Whipple, K.X., Burchfiel, B.C., Chen, L., 2004. Geomorphic constraints on surface uplift, exhumation, and plateau growth in the Red River region, Yunnan Province, China. Geological Society of America Bulletin 116(7–8), 895–909.
Sklar, L., Dietrich, W.E., 1998. River longitudinal profiles and bedrock incision models: Stream power and the influence of sediment supply. In: Tinkler, K.J., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels, Geophysical Monograph Series. AGU Press, Washington, DC, volume 107, pp. 237–260.
Sklar, L., Dietrich, W.E., 2001. Sediment and rock strength controls on river incision into bedrock. Geology 29(12), 1087–1090.
Sklar, L., Dietrich, W.E., 2004. A mechanistic model for river incision into bedrock by saltating bed load. Water Resources Research 40, W06301. http://dx.doi.org/10.1029/2003WR002496.
Sklar, L.S., Dietrich, W.E., 2006. The role of sediment in controlling steady-state bedrock channel slope: implications of the saltation-abrasion incision model. Geomorphology 82(1–2), 58–83.

Snyder, N., Whipple, K., Tucker, G., Merritts, D.J, 2000. Landscape response to tectonic forcing: digital elevation model analysis of stream profiles in the Mendocino triple junction region, northern California. Geological Society of America, Bulletin 112(8), 1250–1263.

Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2002. Interactions between onshore bedrock-channel incision and nearshore wave-base erosion forced by eustacy and tectonics. Basin Research 14, 105–127.

Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2003a. Channel response to tectonic forcing: field analysis of stream morphology and hydrology in the Mendocino triple junction region, northern California. Geomorphology 53, 97–127.

Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2003b. Importance of a stochastic distribution of floods and erosion thresholds in the bedrock river incision problem. Journal of Geophysical Research 108(B2), 2117.

Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2003c. Correction to "Importance of a stochastic distribution of floods and erosion thresholds in the bedrock river incision problem". Journal of Geophysical Research 108(B8), 2388. http://dx.doi.org/10.1029/2001JB002649.

Springer, G.S., Tooth, S., Wohl, E.E., 2006. Theoretical modeling of stream potholes based upon empirical observations from the Orange River, Republic of South Africa. Geomorphology 82(1–2), 160–176.

Stock, J., Dietrich, W.E., 2003. Valley incision by debris flows: evidence of a topographic signature. Water Resources Research 39(4), 1089.

Stock, J.D., Dietrich, W.E., 2006. Erosion of steepland valleys by debris flows. Geological Society of America Bulletin 118(9), 1125–1148.

Tucker, G., 2004. Drainage basin sensitivity to tectonic and climatic forcing: implications of a stochastic model for the role of entrainment and erosion thresholds. Earth Surface Processes and Landforms 29, 185–205.

Tucker, G.E., 1996. Modeling the large-scale interaction of climate, tectonics and topography. PhD thesis, Pennsylvania State University, Earth System Science Center , University Park, PA.

Tucker, G.E., Whipple, K.X., 2002. Topographic outcomes predicted by stream erosion models: sensitivity analysis and intermodel comparison. Journal of Geophysical Research 107(B9), http://dx.doi.org/10.1029/2001JB000162.

Turowski, J.M., Lague, D., Hovius, N., 2007. Cover effect in bedrock abrasion: a new derivation and its implications for the modeling of bedrock channel morphology. Journal of Geophysical Research – Earth Surface 112, F04006.

Turowski, J.M., Lague, D., Hovius, N., 2009. Response of bedrock channel width to tectonic forcing: insights from a numerical model, theoretical considerations, and comparison with field data. Journal of Geophysical Research – Earth Surface 114, F03016.

Valla, P.G., van der Beek, P.A., Lague, D., 2010. Fluvial incision into bedrock: insights from morphometric analysis and numerical modeling of gorges incising glacial hanging valleys (Western Alps, France). Journal of Geophysical Research – Earth Surface 115, F02010.

van der Beek, P., Bishop, P., 2003. Cenozoic river profile development in the upper Lachlan catchment (SE Australia) as a test of quantitative fluvial incision models. Journal of Geophysical Research 108(B6), 2309. http://dx.doi.org/10.1029/2002JB002125.

Weissel, J.K., Seidl, M.A., 1998. Inland propagation of erosional escarpments and river profile evolution across the southeastern Australian passive continental margin. In: Tinkler, K., Wohl, E.E. (Eds.), Rivers Over Rock: Fluvial Processes in Bedrock Channels, Geophysical Monograph Series, AGU Press, Washington, DC, volume 107, pp. 189–206.

Whipple, K., 2001. Fluvial landscape response time: how plausible is steady state denudation? American Journal of Science 301, 313–325.

Whipple, K., 2009. The influence of climate on the tectonic evolution of mountain belts. Nature Geoscience 2, 97–104. http://dx.doi.org/10.1038/ngeo1413.

Whipple, K., Tucker, G., 2002. Implications of sediment-flux dependent river incision models for landscape evolution. Journal of Geophysical Research 107(B2), http://dx.doi.org/10.1029/2000JB000044.

Whipple, K.X., 2004. Bedrock rivers and the geomorphology of active orogens. Annual Review of Earth and Planetary Sciences 32, 151–185.

Whipple, K.X., Hancock, G.S., Anderson, R.S., 2000. River incision into bedrock: mechanics and relative efficacy of plucking, abrasion, and cavitation. Geological Society of America Bulletin 112(3), 490–503.

Whipple, K.X., Tucker, G.E., 1999. Dynamics of the stream-power river incision model: implications for height limits of mountain ranges, landscape response timescales, and research needs. Journal of Geophysical Research 104(B8), 17661–17674.

Whittaker, A.C., Cowie, P.A., Attal, M., Tucker, G.E., Roberts, G.P., 2007a. Bedrock channel adjustment to tectonic forcing: implications for predicting river incision rates. Geology 35(2), 103–106.

Whittaker, A.C., Cowie, P.A., Attal, M., Tucker, G.E., Roberts, G.P., 2007b. Contrasting transient and steady-state rivers crossing active normal faults: new field observations from the Central Apennines, Italy. Basin Research 19(4), 529–556.

Wilcox, A.C., Nelson, J.M., Wohl, E.E., 2006. Flow resistance dynamics in step-pool channels: 2. Partitioning between grain, spill, and woody debris resistance. Water Resources Research 42, W05419.

Willett, S.D., 1999. Orogeny and orography: the effects of erosion on the structure of mountain belts. Journal of Geophysical Research 104(B12), 28,957–928,981.

Wobus, C.W., Crosby, B.T., Whipple, K.X., 2006a. Hanging valleys in fluvial systems: controls on occurrence and implications for landscape evolution. Journal of Geophysical Research – Earth Surface 111, F02017.

Wobus, C.W., Hodges, K.V., Whipple, K.X., 2003. Has focused denudation sustained active thrusting at the Himalayan topographic front? Geology 31(10), 861–864.

Wobus, C.W., Kean, J.W., Tucker, G.E., Anderson, R.S., 2008. Modeling the evolution of channel shape: balancing computational efficiency with hydraulic fidelity. Journal of Geophysical Research – Earth Surface 113, F02004.

Wobus, C.W., Tucker, G.E., Anderson, R.S., 2006b. Self-formed bedrock channels. Geophysical Research Letters 33, L20402.

Wobus, C.W., Whipple, K.X., Kirby, E., et al., 2006c. Tectonics from topography: procedures, promise, and pitfalls. Geological Society of America Special Paper 398, 55–74.

Wohl, E., 2000a. Mountain Rivers. American Geophysical Union, Washington, DC, 320 pp.

Wohl, E., 2000b. Substrate influences on step-pool sequences in the Christopher Creek drainage, Arizona. Journal of Geology 108(1), 121–129.

Wohl, E., David, G.C.L., 2008. Consistency of scaling relations among bedrock and alluvial channels. Journal of Geophysical Research – Earth Surface 113, F04013.

Wohl, E., Ikeda, H., 1997. Experimental simulation of channel incision into a cohesive substrate at varying gradients. Geology 25(4), 295–298.

Wohl, E., Merritt, D., 2005. Prediction of mountain stream morphology. Water Resources Research 41, W08419.

Wohl, E., Merritt, D.M., 2008. Reach-scale channel geometry of mountain streams. Geomorphology 93(3–4), 168–185.

Wohl, E.E., Thompson, D.M., Miller, A.J., 1999. Canyons with undulating walls. Geological Society of America Bulletin 111(7), 949–959.

Yanites, B.J., Tucker, G.E., 2010. Controls and limits on bedrock channel geometry. Journal of Geophysical Research – Earth Surface 115, F04019.

Yanites, B.J., Tucker, G.E., Mueller, K.J., Chen, Y.G., Wilcox, T., Huang, S.Y., Shi, K.W., 2010. Incision and channel morphology across active structures along the Peikang River, central Taiwan: implications for the importance of channel width. Geological Society of America Bulletin 122(7–8), 1192–1208.

Zaprowski, B.J., Pazzaglia, F.J., Evenson, E.B., 2005. Climatic influences on profile concavity and river incision. Journal of Geophysical Research – Earth Surface 110, F03004.

Biographical Sketch

Kelin Whipple received his PhD from the University of Washington in 1994 and is currently a professor at Arizona State University. His primary research focus has been on long-term landscape evolution in response to changes in climatic conditions and tectonic forcing.

Roman DiBiase is currently a graduate student at the Arizona State University. His PhD work involves quantifying channel and hillslope processes in the San Gabriel Mountains, CA.

Ben Crosby received his PhD from MIT in 2006 and is currently an assistant professor at Idaho State University. Though his PhD focused on knickpoints and long-term landscape response to base-level fall, he now applies that same conceptual framework to evaluate how smaller magnitude signals of contemporary climate change propagate through Arctic and high relief landscapes.