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BEDROCK RIVERS AND THE GEOMORPHOLOGY OF ACTIVE OROGENS

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■ **Abstract** Bedrock rivers set much of the relief structure of active orogens and dictate rates and patterns of denudation. Quantitative understanding of the role of climate-driven denudation in the evolution of unglaciated orogens depends first and foremost on knowledge of fluvial erosion processes and the factors that control incision rate. The results of intense research in the past decade are reviewed here, with the aim of highlighting remaining unknowns and suggesting fruitful avenues for further research. This review considers in turn (*a*) the occurrence and morphology of bedrock channels and their relation to tectonic setting; (*b*) the physical processes of fluvial incision into rock; and (*c*) models of river incision, their implications, and the field and laboratory data needed to test, refine, and extend them.

INTRODUCTION

One of the most interesting geoscience discoveries in the past 20 years is arguably the recognition that there exists a dynamic coupling between climate-driven erosion and tectonics. Numerical simulations using coupled tectonic and surface process models (e.g., Koons 1989, Beaumont et al. 1992, Willett 1999) have clearly demonstrated that the geodynamics of active orogens is significantly influenced by the surface boundary conditions. Although erosion does not build mountains, it does fundamentally influence the size, shape, and evolution of active orogens: For a given tectonic setting (mass influx) and set of material properties, climate-controlled erosional efficiency can be shown to determine orogen height, width, and crustal thickness (Dahlen & Suppe 1988, Whipple & Meade 2004). An implication of these models and data is that at orogen scale, rock uplift rates and patterns are strongly influenced by erosional efficiency and its spatial distribution.

Although river channels occupy a very small percentage of the land surface, river incision rates ultimately control regional denudation rates and patterns because rivers set the boundary condition for hillslope erosion. Channels also dictate much of the topographic form of mountainous landscapes: The drainage network

Figure 1 Perspective view of Haulien River basin (447 km^2 drainage area, 3580 m relief), Central Range, Taiwan. Channel network with minimum supporting first-order drainage area of 0.8 km^2 shown ($A > A_{cr}$, see Equation 2). Eighty to 90% of topographic relief is on the channel network (Table 2).

defines the planview texture of the landscape, and channel longitudinal profiles set much of the relief structure of unglaciated mountain ranges (Figure 1). Moreover, as a consequence of slope stability thresholds, hillslope form becomes insensitive to further increases in tectonic forcing at fairly low uplift rates (Strahler 1950, Schmidt & Montgomery 1995, Burbank et al. 1996, Densmore et al. 1998)—in areas of rapid erosion and tectonic uplift ($>\sim$0.2 mm $year^{-1}$), the longitudinal profiles of bedrock (or mixed bedrock-alluvial) channels are more sensitive indicators of uplift rate than other morphological properties. In addition, changes in base level (tectonic/eustatic/drainage reorganization) and climate are transmitted through landscapes along channels; the rate at which these signals are conveyed sets, to first order, the timescale of landscape response to perturbation (Whipple & Tucker 1999, Whipple 2001). Thus, the controls on rates of river incision into bedrock largely dictate the relationships among climate, lithology, tectonics, and topography.

In the context of orogen evolution and the coupling between climate and tectonics, tectonic geomorphology has recently been in an exciting phase of initial exploration and discovery with considerable progress in field studies, laboratory experimentation, theory, and numerical simulation. However, much of the analysis of the interactions among climate, erosion, and tectonics has been accomplished with generic rule sets. Much has been gained from these generic analyses. We now know well what we need to know better and why, as summarized here. There is a clear need to develop and test process-specific models of river incision into bedrock to augment and complement more generic analyses and models—it is time for the field to move into a new stage of development.

Here, I review current understanding of the complex phenomenon of river incision into rock and the resulting topographic signature of tectonics, lithology, and climate. I first review what we know from direct observation concerning (*a*) the definition and occurrence of bedrock channels, (*b*) the morphology of channels (bed morphology, width, longitudinal profile form) and their relation to tectonic setting, and (*c*) the physical processes of fluvial incision into rock and factors that dictate the relative efficacy of these processes. Then I turn to theoretical developments, their implications, and tests of competing models against laboratory and field data. This review is restricted to fluvial processes. Glaciers, glacial erosion, and glaciated landforms are not discussed.

DEFINITION AND OCCURRENCE

Bedrock channels lack a continuous cover of alluvial sediments, even at low flow, and exist only where transport capacity (Q_c) exceeds bedload sediment flux (Q_s) over the long term ($Q_s/Q_c < 1$) (Gilbert 1877, Howard 1980, Howard et al. 1994, Montgomery et al. 1996). However, channels with long stretches of bare bedrock along the bed and banks are rare. Short-term pulses of rapid sediment delivery from hillslopes (owing to nearby landslides, recent climate or land use change, fires, etc.) may produce temporary sediment fills (Benda & Dunne 1997, Densmore et al. 1998), or thin, patchy mantles of alluvium may persist despite active incision through rock (Figure 2*B*,*C*) (Howard 1980, Howard et al. 1994). Although most might be better termed mixed bedrock-alluvial channels, the simple descriptor "bedrock channel" is retained here in keeping with recent usage, and because bedrock is typically very close to the surface even if a thin alluvial cover is present (Howard 1980, Howard et al. 1994, Montgomery et al. 1996, Wohl & Ikeda 1998, Massong & Montgomery 2000, Pazzaglia & Brandon 2001, Wohl & Merritt 2001). Following Montgomery et al. (1996) and Massong & Montgomery (2000), I use "bedrock reach" to denote a local absence of even a thin alluvial cover and "bedrock channel" to denote a channel that has frequent exposures of rock along its bed and banks, including occasional bedrock reaches. A temporally and spatially continuous blanket of transportable sediment on both the bed and banks, on the other hand, characterizes alluvial channels. Contrary to classical definitions, bedrock channels are self-formed. Bed and banks are not composed of transportable sediment but are erodible. Flow, sediment flux, substrate properties, and base-level conditions dictate self-adjusted combinations of channel gradient, width, and bed morphology in bedrock channels (e.g., Wohl & Ikeda 1998, Wohl et al. 1999, Wohl & Merritt 2001).

Bedrock channels occur mainly, but not exclusively, in actively incising portions of landscapes and where channels are cut into resistant rock units, most often in actively uplifting areas. Bedrock channels are not restricted to headwater regions and can extend to large drainage areas depending on river network geometry and tectonic setting. Little observational data exists, however, to determine how the degree of rock exposure in channel bed and banks varies as a function of drainage area, rock uplift rate, rock properties, and climate in systems undisturbed by recent land use change.

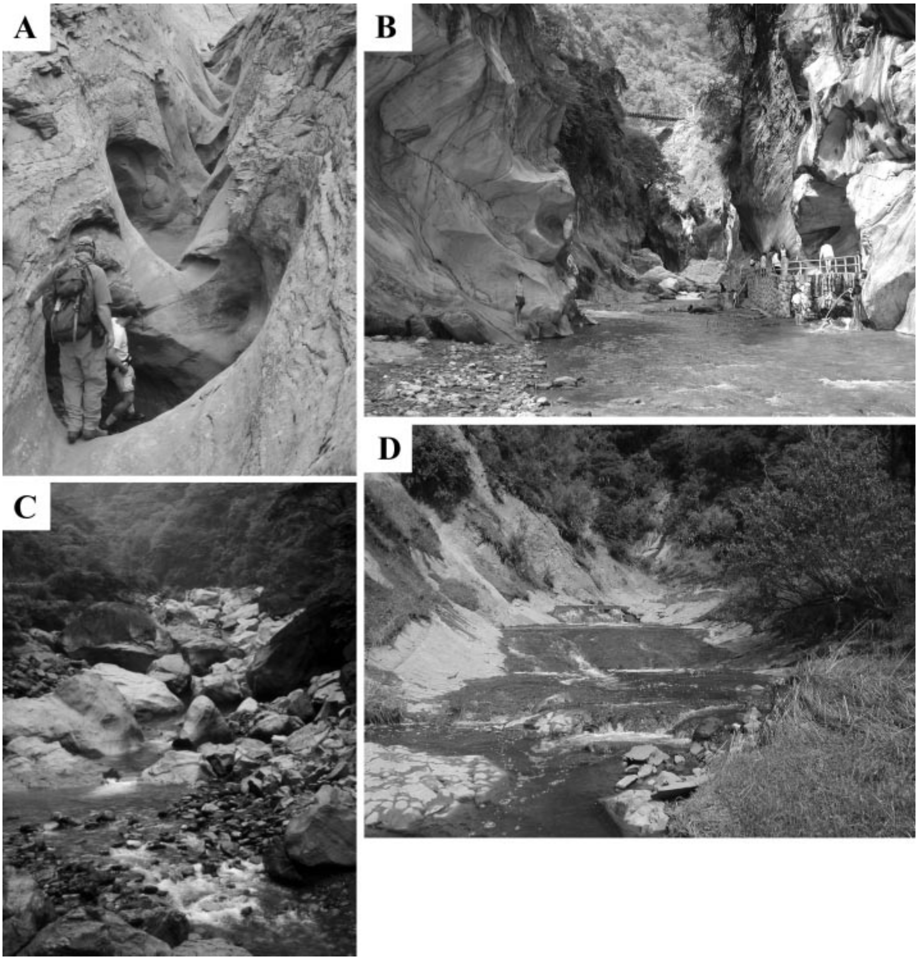

Figure 2 (*A*) Chain of potholes in Navajo Sandstone, Utah. (*B*) Gravel-bedded incised bedrock river, Taroko National Park, Taiwan. Note 1–2 m diameter flutes high on rock bank (*left*) indicating active suspended load abrasion. (*C*) Landslide-derived boulder lag, Longmenshan, China. (*D*) Stepped plane-bed bedrock channel in mudstone, Waipaoa catchment, North Island, New Zealand.

MORPHOLOGY AND TECTONIC SIGNATURE

Many analyses of river longitudinal profiles, and most landscape evolution models, either explicitly or implicitly make the simplifying assumption that channel gradient is the only degree of freedom in river response to climate, tectonics, and lithology. Naturally, the reality of steeplands river channels is much more complex. There are a multitude of ways in which a river may respond to external forcings in addition to adjustments to channel gradient: Channel width, sinuosity, extent of alluvial cover, bed material grain size, bed morphology, and hydraulic roughness are all potentially important variables. In this section, I briefly review what is known about each aspect of river morphology to highlight the potential importance of each and to highlight the many unknowns in river response to tectonic and climatic forcing that require further field and laboratory study.

Channel Bed Morphology

Montgomery & Buffington (1997) developed a classification scheme that defines four categories of bed morphology in alluvial reaches of steepland rivers, each associated with a characteristic median grainsize of bed material (D_{50bm}), gradient (S), relative roughness (ratio of D_{50bm} to flow depth, H), and hydraulic roughness (Manning's N): cascade, step-pool, plane-bed, and pool-riffle (Table 1). Montgomery & Buffington (1997) emphasize that cascade and step-pool morphologies are associated with low sediment flux (Q_s) to transport capacity (Q_c) ratios (and therefore low D_{50bl}/D_{50bm} ratios, where D_{50bl} is bedload median size), whereas plane-bed and pool-riffle morphologies are associated with high Q_s/Q_c ratios. They suggest that plane-bed morphology marks the onset of approximately transport-limited conditions ($Q_s/Q_c \sim 1$).

Wohl & Merritt (2001) subdivide bedrock reaches into stepped, plane-bed, inner-channel, and undulating wall types. Stepped channels are marked by small knickpoints, chute/pothole pairs, flutes and longitudinal grooves, tend to be steep, and form in either resistant or heterogeneous substrates (Wohl & Ikeda 1998, Wohl & Merritt 2001). Small knickpoints (bedrock steps) are often formed on more resistant ledges, particularly in strongly stratified rocks (Miller 1991, Wohl et al. 1994, Snyder et al. 2003a). However, steeper bedrock reaches tend to be associated with more frequent knickpoints or chute/pothole pairs (step-pool equivalents), sometimes with the appearance of cyclic steps in homogeneous rocks. Plane-bed bedrock channel reaches tend to be relatively low gradient and are associated with weak sedimentary rocks (especially where susceptible to wet/dry weathering and slaking) (Wohl & Ikeda 1998, Pazzaglia & Brandon 2001). Inner-channels and channels with undulating walls are often inferred to have formed by the coalescence

TABLE 1 Channel morphology in alluvial reaches

Category	**D_{50bm}**	**Gradient**	**Relative roughness (D_{50}/H)**	**Manning's N**
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

of potholes (Figure 2*A*) and longitudinal grooves in resistant rock types (Wohl et al. 1999, Whipple et al. 2000a, Wohl & Merritt 2001).

Although in some rivers it is clear that steeper headwater channel segments are associated with less frequent and shorter alluvial reaches interspersed with stepped bedrock reaches (Pazzaglia & Brandon 2001, Snyder et al. 2003a), the response of channel bed morphology to increased rock uplift rates is not well known, and available field data show no indication of a universal mode of response. For instance, in both the Clearwater River studied by Pazzaglia & Brandon (2001) and the small coastal streams of the King Range studied by Snyder et al. (2000, 2003a), higher uplift rates are associated with greater exposure of bedrock and more frequent bedrock steps, but in each case there are factors that complicate direct interpretation of the observed correlation. In the Clearwater, incision rates increase upstream (Pazzaglia & Brandon 2001), such that greater exposure of bedrock is also correlated with channel gradient and inversely correlated with drainage area, and the causative factor is unclear. In the coastal streams of the King Range, the recent land use history of timber harvest in the low uplift zone has delivered massive quantities of sediment to the channels, confounding interpretation of differences in bed morphology between the low and high uplift rate zones (Snyder et al. 2003a). Furthermore, observations elsewhere do not corroborate a simple positive correlation between percent bedrock exposure and rock uplift rate. Preliminary observations in the San Gabriel Mountains indicate that armoring of channel beds with boulders delivered by landslides and debris flows is prevalent in both the high and low uplift zones defined by Blythe et al. (2000). Similarly, transverse rivers that cut through the high Himalayas in Nepal and the steep rivers draining the east flank of Taiwan's Central Range have only sparse exposure of bedrock in channel beds, despite high incision rates (Figure 2*B*). Reaches where the channel bed is blanketed by large landslide-derived blocks are common in rapidly eroding landscapes (Figure 2*C*). Conversely, slowly incising rivers cut in resistant rock units in the Appalachians, on the Colorado Plateau, and on the Brazilian craton, for example, often have considerable bedrock exposure in the bed and banks (Figure 2*A*). Collection of comprehensive data on percent bedrock exposure in a wide range of field settings is required before the controls on this important variable in river response to climatic and tectonic forcing can be unraveled.

Channel Width

The controls on bedrock channel width remain a major unsolved problem. The bedrock channel width problem is closely linked to lateral bank erosion rates and the strath terrace formation problem, and is probably strongly correlated to the degree and frequency of alluviation of the channel bed (Pazzaglia et al. 1998, Pazzaglia & Brandon 2001, Stark & Stark 2001, Hancock & Anderson 2002). Empirical studies have recently looked at the influences of upstream drainage area, lithology, and rock uplift rate on bedrock channel width, though little theoretical

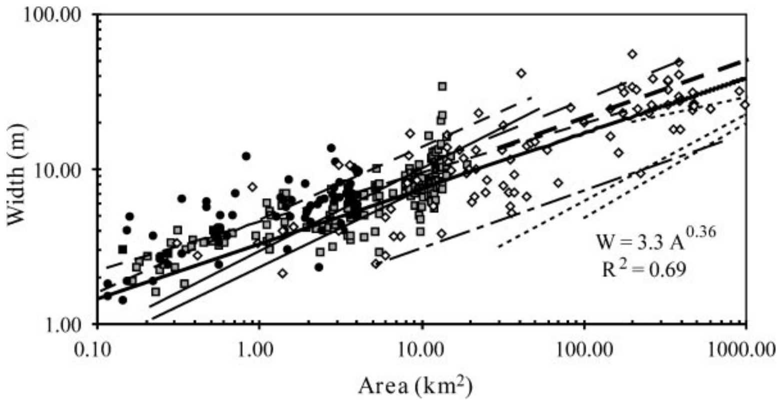

Figure 3 Bedrock channel bankfull width data. Open diamonds: All data from Hack (1957). Thick solid line and equation are regression fit to these data. Solid black circles: King Range, CA, high uplift rate zone (Snyder et al. 2003a). Gray squares: King Range, CA, low uplift rate zone (Snyder et al. 2003a). Thin solid lines: Fits to Oregon Coast Range data (Montgomery & Gran 2001). Thin, short dashed lines: Fit to Olympic Mountains data (Montgomery & Gran 2001). Thin, long dashed line: Fit to Olympic Mountains data (Tomkin et al. 2003). Dash-dot line: Fit to Sierra Nevada data (Montgomery & Gran 2001). Dotted lines: Fits to unpublished data from the Longmenshan, China (E. Kirby, written communication, 2003). Thick dashed line: Fit to gravel-bedded alluvial channels (Parker et al. 2003). Trends from last two data sets extend to considerably greater drainage areas than shown.

work has been done (Suzuki 1982). Here I present and discuss a compilation of available data, including a comparison to better-known gravel bedded alluvial channels (e.g., Parker 1978, Parker et al. 2003).

Although locally variable, bedrock channel width (W) varies systematically with drainage area (A, proxy for water and sediment discharge) in a manner similar to that observed in alluvial channels (Figure 3) (Hack 1957, Montgomery & Gran 2001, Snyder et al. 2003a, Tomkin et al. 2003, van der Beek & Bishop 2003):

$$W \propto A^{0.3-0.5}. \tag{1}$$

Montgomery & Gran (2001) found that in channels cut into weak rocks there was no measurable difference in channel width between alternating bedrock and alluviated reaches, suggesting a fundamental role of bedload flux in setting channel width in these systems (e.g., Parker 1978, Parker et al 2003). Indeed, the width of most bedrock channels (shown in Figure 3) is approximately that expected for gravel-bed alluvial channels with the same drainage area, strongly suggesting that gravel-bed channel width and bedrock channel width are controlled by the same physics, at least in these settings. However, in one case in the Sierra Nevada, Montgomery & Gran (2001) found that channel width was considerably narrower where cut through a highly resistant rock unit (Figure 3). In addition, recent debris-flow scour widened channels at drainage areas of less than 1 km^2, although only an uncertain fraction of this apparent widening was attributable to actual bedrock bank erosion (Montgomery & Gran 2001).

Although the notion that channel narrowing in response to increased incision rate is intuitively appealing, available data suggests this occurs in some settings but not in others. With the exception of the Sierra Nevada data, most of the bedrock channels for which data is plotted on Figure 3 show remarkably similar widths as a function of drainage area (and compared to gravel-bed alluvial rivers), despite considerable variation in climate, lithology, and incision rate. For example, Snyder et al. (2003a) measured channel widths in incising bedrock channels similar in all ways except rock uplift rate. Surprisingly, they found no measurable difference in channel width between areas being uplifted at $\sim$0.5 and 3–4 mm year^{-1} (Figure 3). However, one might argue in this landscape that the channels of the high uplift zone were effectively narrower given the orographically enhanced precipitation and flood discharges in those channels (Snyder et al. 2003a). Similarly, preliminary measurements in the San Gabriel Mountains, CA (not shown), indicate, at best, a subtle narrowing in the high uplift rate zone defined by Blythe et al. (2000). In addition, data on the Clearwater River, WA, follows a normal increase in channel width with drainage area ($W \sim A^{0.42}$; Figure 3), despite a strong along-stream gradient in rock uplift rate (Pazzaglia & Brandon 2001, Tomkin et al. 2003).

Most published examples of river-width adjustments to uplift rate involve either alluvial rivers (e.g., Harbor 1998) or transitions between alluviated, nonincising segments and rapidly incising segments of bedrock rivers that record mostly a change in floodplain width rather than bankfull channel width (e.g., Lave & Avouac 2000, 2001). However, new data from a few localities appear to show a strong correlation between rock uplift rate and channel width (Figure 3). Data from the Longmenshan in Sichuan Province, China (E. Kirby, unpublished data on channels studied by Kirby et al. 2003) record an apparent narrowing of channel width in response to increased rock uplift rate on large bedrock rivers: $W \sim A^{0.2}$ along rivers with rock uplift rate increasing downstream, and $W \sim A^{0.6}$ along rivers with rock uplift rate decreasing downstream (Figure 3). In addition, unpublished data from the Santa Ynes Mountains near Santa Barbara, CA, show a strong relative narrowing in the high uplift zone (Duvall et al. 2003; A. Duvall, E. Kirby & D. Burbank, "Tectonic and Lithologic Controls on Bedrock Channel Profiles and Processes in Coastal California," manuscript in review at the *Journal of Geophysical Research—Earth Surface*). Finally, although partly reflecting a transition from a sandy, alluviated bed to a bedrock-floored reach, Montgomery et al. (2002) have described another example: The width of the Tsangpo River narrows dramatically as it crosses a knickpoint into a zone of high rock uplift. Why river width sometimes narrows in response to increased rock uplift rates and sometimes does not is an important, unanswered question.

Channel Longitudinal Profile Form

Where not marked by abrupt knickpoints (discrete steps in either channel elevation or gradient), longitudinal profiles of bedrock rivers are often smoothly concave-up and reasonably well described by Flint's law relating local channel gradient to upstream drainage area:

$$S = k_s A^{-\theta}, \tag{2}$$

where k_s is known as the steepness index and θ as the concavity index (Figure 4). The slope-area scaling in Equation 2, however, only holds downstream of a critical

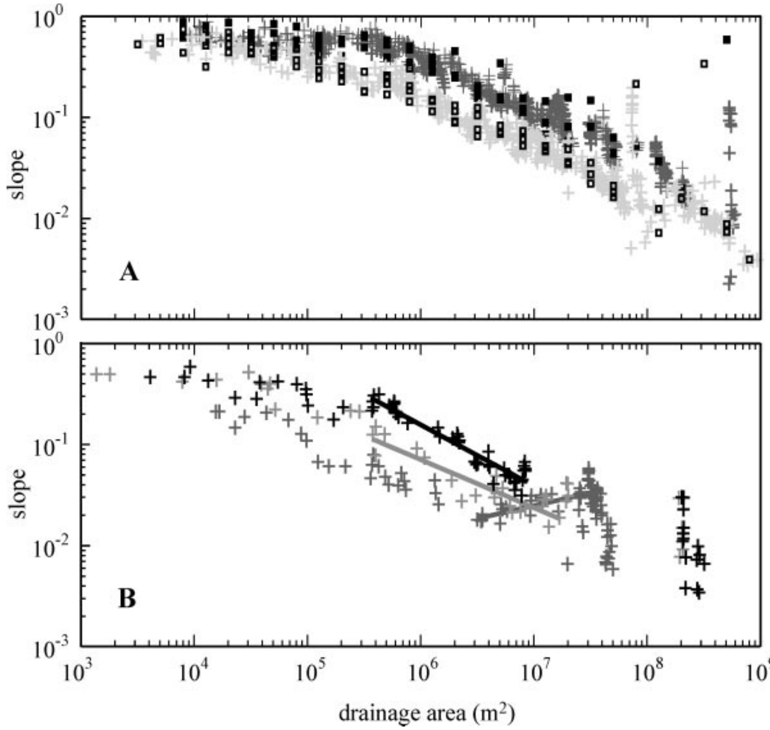

Figure 4 Channel profile slope-area data. (*A*) San Gabriel Mountains data. High uplift rate zone: dark gray crosses (*solid squares* are log-bin average slopes). Low uplift rate zone: light gray crosses (*open squares* are log-bin average slopes). Data shows uniform steepening by a factor of ~3 in the fluvial regime within the high uplift rate zone. (*B*) Siwalik Hills, Nepal data. Uniform high uplift rate (14 mm year^{-1}): black crosses. Uniform moderate uplift rate (7 mm year^{-1}): light gray crosses. Transverse drainage crossing from low to high uplift rate: medium gray crosses. Regression lines help illustrate differences in channel concavity index in fluvial reaches within the area of active uplift.

drainage area, A_{cr} (e.g., Montgomery & Foufoula-Georgiou 1993, Sklar & Dietrich 1998, Whipple & Tucker 1999). Also, it is important to recognize that within the fluvial slope-area scaling regime, knickpoints may separate channel reaches with distinct steepness and concavity indices, depending on the spatial distribution of substrate properties, spatial and temporal rock uplift and climatic patterns, and transitions from bedrock to alluvial channel types (e.g., Whipple & Tucker 1999, Kirby & Whipple 2001, Kirby et al. 2003, VanLaningham 2003, Wobus et al. 2003).

In tectonically active mountain ranges, 70%–90% of topographic relief occurs in the regime where Equation 2 is a reasonable approximation to channel profile form ($A > A_{cr}$), as documented in Table 2 and illustrated in Figure 1. Although often termed the fluvial relief (e.g., Whipple et al. 1999, Whipple & Tucker 1999), this is strictly the relief within the fluvial slope-area scaling regime. Stock & Dietrich (2003) have cautioned that debris flows may importantly influence channel profile form out to considerably larger drainage areas ($A \leq 10$ km^2) and lower gradients ($S \geq 0.03$–10) than suggested by the apparent kink in the slope-area relation ($A_{cr} \sim 1$ km^2, typically). At the very least, debris flows appear to play a fundamental role in setting A_{cr} and limiting channel slopes above this point, thereby importantly influencing total landscape relief. In addition, it is possible that the transitional debris-flow-to-fluvial section has a different sensitivity to uplift, climate, or lithology than does the rest of the profile. If the latter were true, however, one would expect systematic changes in channel concavity with these variables in a manner that has not yet been documented, as discussed below. Still, given the present uncertainty, landscape evolution models (e.g., Beaumont et al. 1992,

TABLE 2 Fluvial relief statistics in active orogens[a]

Field area	**Critical drainage area[b]** A_c, 10^5 m^2	**Average colluvial slope** S_c	**% Fluvial relief** R_f/R_t	**Sample size** N
King Range, California (high uplift rate)	0.59 ± 0.20	0.54 ± 0.11	79 ± 7	14
King Range, California (low uplift rate)	0.72 ± 0.24	0.36 ± 0.05	80 ± 5	7
Central Range, Taiwan (high uplift rate)	0.84 ± 0.72	0.36 ± 0.11	85 ± 5	17
San Gabriel Mountains, California (high uplift rate)	0.6–8.9	0.54 ± 0.15	71 ± 7	5
San Gabriel Mountains, California (low uplift rate)	0.25 ± 0.17	0.36 ± 0.12	78 ± 6	7

[a]All uncertainties indicate 1-sigma error bars.

[b]A_c defined by break in slope-area scaling in longitudinal profile data (only).

Tucker & Bras 1998) and studies such as Snyder et al. (2000), Kirby & Whipple (2001), and Roe et al. (2002) that use fluvial process rules to address river profile response at drainage areas less than $\sim$10 km^2 must exercise caution and should admit the possibility that the debris-flow processes not captured in fluvial models could importantly influence the analysis in as yet incompletely understood ways.

The concavity index determined over the full bedrock channel length (from A_{cr} to either the river mouth, a major confluence, or to the point where fully alluvial conditions prevail, depending on the study) varies widely from 0.3–1.2 (e.g., Tarboton et al. 1989, Sklar & Dietrich 1998, Snyder et al. 2000, Kirby & Whipple 2001, Brocklehurst & Whipple 2002, Tucker & Whipple 2002, Kirby et al. 2003, VanLaningham 2003; L.M. Schoenbohm, K.X. Whipple, B.C. Burchfiel & L. Chen, "Geomorphic Constraints on Surface Uplift, Exhumation, and Plateau Growth in the Red River Region, Yunnan Province, China," manuscript in review at *Geophysical Society of America Bulletin*) and can be either negative or extreme ($>$1) over short reaches. Fortunately, some general controls on channel concavity can be identified. Low concavities ($<$0.4) are associated either with short, steep drainages importantly influenced by debris flows (e.g., eastern Sierra Nevada) (Brocklehurst & Whipple 2002) or with downstream increases in either incision rate or rock strength, commonly associated with knickpoints (e.g., Kirby & Whipple 2001, Kirby et al. 2003). Moderate concavities (0.4–0.7) are associated with actively uplifting bedrock channels in homogeneous substrates experiencing uniform (or close to uniform) rock uplift (e.g., Sub, Lesser, and High Himalaya in Nepal; Longmenshan, China; East Central Range, Taiwan; San Gabriel Mountains, CA; King Range, CA; see Figure 4). High concavities (0.7–1.0) are associated with downstream decreases in rock uplift rate or rock strength (Kirby & Whipple 2001, Kirby et al. 2003); downstream transitions to fully alluvial conditions (e.g., Clearwater, WA; Oregon Coast Range; Southern East Central Range, Taiwan) and disequilibrium conditions resulting from a temporal decline in rock uplift rate. Extreme concavities (negative or $>$1) are associated with abrupt knickpoints owing either to pronounced along-stream changes in substrate properties (e.g., VanLaningham 2003) or to spatial or temporal differences in rock uplift rate (e.g., Figure 5) (L.M. Schoenbohm, K. Whipple, B.C. Burchfiel & L. Chen, manuscript in review), including transitions from incisional to depositional conditions.

Channel steepness index is known to be a function of (at least) rock uplift rate, lithology, and climate (Snyder et al. 2000, Kirby & Whipple 2001, Duvall et al. 2003, Kirby et al. 2003, Wobus et al. 2003; A. Duvall, E. Kirby & D. Burbank, manuscript in review). Where one can compare undisturbed river profiles in field areas with independently known high and low uplift rates and similar climate and lithology, channel profile data is consistently composed of subparallel slope-area arrays with different intercepts, indicating little relation between the concavity index and rock uplift rate. Figure 4 shows examples from the San Gabriel Mountains and the Siwalik Hills, Nepal. Data from these and other landscapes (the King Range, CA; the Nepal Himalaya; eastern margin of the Tibetan Plateau; North Island of New Zealand; southern Appalachian Mountains; the Ural Mountains;

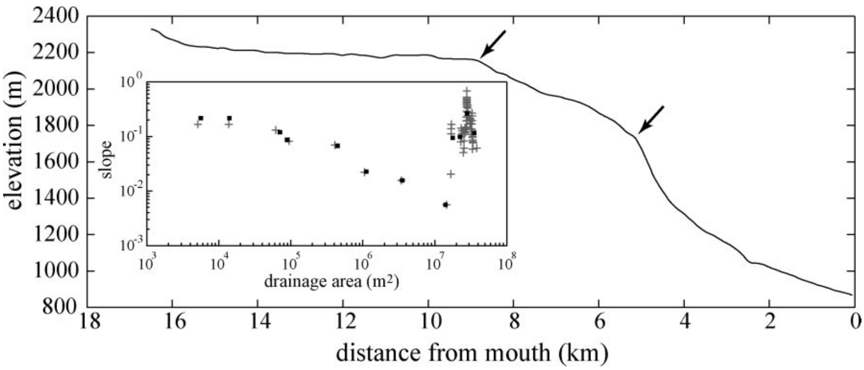

Figure 5 Example of transient river profile, Ailao Shan, Yunnan, China. Arrows indicate major knickpoints. Slope-area data shown in inset. Note extreme concavity index below second knickpoint (rapid slope decrease at A $>$ 30 km^2).

and the Central Range of Taiwan) indicate that the channel steepness index normalized (k_{sn}) to a reference concavity index (θ_{ref}) varies over only slightly more than one order of magnitude ($k_{sn} = 20$–600 for $\theta_{ref} = 0.45$), suggesting a highly damped system despite a clear, consistent, and readily measurable tectonic influence. However, it is important to note that strong lithologic contrasts can produce differences in channel steepness index comparable to those associated with large gradients in rock uplift rate (Snyder et al. 2000, Duvall et al. 2003, Stock & Dietrich 2003, van der Beek & Bishop 2003; A. Duvall, E. Kirby & D. Burbank, manuscript in review). Given that rock uplift rate and lithology strongly influence channel steepness index values, it is also apparent that systematic along-stream uplift rate gradients or changes in rock properties will affect channel concavity (Figure 4*B*) (Whipple & Tucker 1999, Kirby & Whipple 2001, Duvall et al. 2003).

In stream-profile analysis, it is important to separate morphological indices (steepness k_s, concavity θ) from model parameters. There is a well-established tradition of directly interpreting these indices, especially the concavity index θ, in terms of the stream power river incision model (e.g., Tarboton et al. 1989, Seidl & Dietrich 1992). Unfortunately this can be misleading as a range of models is consistent with the form of Equation 2, both at and away from steady state (Whipple & Tucker 2002, Sklar 2003), and many factors influence the quantitative relationship between environmental conditions (lithology, climate, rock uplift rate, sediment flux, and grain-size) and k_s and θ, as is discussed in more detail below. Incision model parameters should be predicted by theory and constrained by direct measurements (e.g., Howard & Kerby 1983, Whipple & Tucker 1999, Whipple et al. 2000a, Sklar & Dietrich 2001, Dietrich et al. 2003, Sklar 2003). However, slope-area analysis of channel profiles can be a powerful qualitative tool even without a complete understanding of fluvial incision processes; slope-area analysis is not tied to the assumptions and shortcomings of any particular river incision model.

FLUVIAL INCISION PROCESSES

The processes of river incision into bedrock include plucking, macroabrasion (chipping and block fracture driven by sediment impacts), wear (incremental, grain-by-grain abrasion), chemical and physical weathering, and possibly cavitation (Baker 1974, Shepherd & Schumm 1974, Foley 1980, Howard & Kerby 1983, Miller 1991, Wohl 1993, Wohl et al. 1994, Baker & Kale 1998, Hancock et al. 1998, Sklar & Dietrich 1998, Wohl & Ikeda 1998, Whipple et al. 2000a, Hartshorn et al. 2002). I will use the term abrasion to include both wear and macroabrasion and to include work done by both bedload and suspended load particles. Debris-flow scour also contributes in some settings and may be the dominant incision process at small drainage areas in steepland drainage networks (Seidl & Dietrich 1992, Montgomery & Foufoula-Georgiou 1993, Howard et al. 1994, Stock & Dietrich 2003).

The degree to which physical and chemical weathering facilitates or favors certain erosional mechanisms is not well known, but clearly depends on both substrate and environmental conditions (Wohl 1993, Wohl et al. 1994, Whipple et al. 2000b). Substrate rock mass quality (e.g., Selby 1980) controls the capacity for fluids to penetrate and interact with the rock. Substrate chemistry controls its susceptibility to the volume expansion of clays and the dissolution of carbonates and other cements. Environmental conditions in the channel set the rate at which these substrate-dependent processes occur. These environmental conditions include incision rate, stream chemistry, flow and sediment cover variability (e.g., wet-dry cycles), and temperature variability (freeze-thaw cycles). In some environments, weathering clearly contributes significantly to erosion rates and patterns, even where incision rates are quite high (e.g., Whipple et al. 2000b). Channel width and the depth of some potholes and other erosional bedforms may well depend at least in part on the action of concentrated weathering processes.

The mechanics and relative efficacy of the various physical erosion processes have been discussed by Hancock et al. (1998), Sklar & Dietrich (1998), Wohl & Ikeda (1998) and Whipple et al. (2000a), and are briefly reviewed below. These processes all include critical thresholds, suggesting that most work may be done by large floods (Baker 1974, Wohl 1993, Wohl et al. 1994, Baker & Kale 1998, Tucker & Bras 2000), which implies that extreme events, such as glacier outburst floods, and ice- and landslide-dam-break floods may play a key role in landscape evolution and long-term denudation rates. However, lower thresholds, higher rainfall, and steeper, narrower channels allow a greater percentage of floods to contribute importantly to stream incision (Snyder et al. 2003b, Tucker 2004). Indeed, Hartshorn et al. (2002) have presented evidence that relatively common floods may accomplish most of the erosion in the rapidly incising rivers of the east flank of Taiwan's Central Range. In addition, each of the various processes of river incision is modulated by the river's sediment load (flux, size distribution, hardness), which plays a dual role: providing tools for abrasion (and potentially driving the production of joint blocks), but protecting the bed when bedload is abundant (e.g., Shepherd

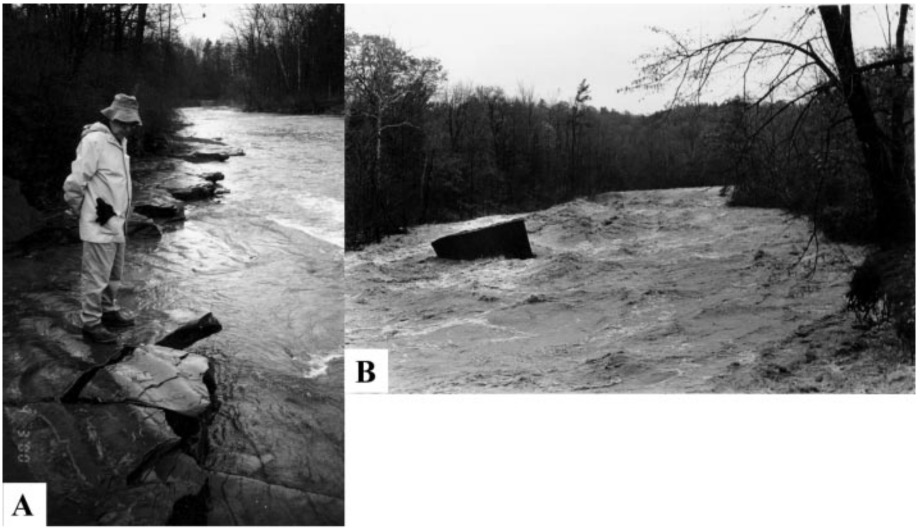

Figure 6 (*A*) Geomorphologist examines large joint blocks of turbidite sands along bank of Fall Creek, Ithaca, NY. (*B*) Large, plucked joint block ($\sim 4 \times 2 \times 0.3$ m) tumbling through this reach in October, 1981 flood (photo by R. Palmer, courtesy of A. Bloom). First published by Snyder et al. (2003b) and reproduced/modified by permission of the American Geophysical Union.

& Schumm 1974; Foley 1980; Wohl & Ikeda 1997; Hancock et al. 1998; Sklar & Dietrich 1998, 2001; Whipple et al. 2000a; Whipple & Tucker 2002).

Most rock units are fractured and/or bedded at length scales of 1 m or less, and plucking processes are probably significant in most natural environments (Whipple et al. 2000a, Hartshorn et al. 2002, Snyder et al. 2003a). Erosion by plucking (Figure 6) requires both production of loose joint blocks and subsequent entrainment and transport of the blocks (Hancock et al. 1998, Whipple et al. 2000a). Either step may be rate limiting, depending on substrate properties, flow hydraulics, sediment load (flux and grain-size distribution), and incision rate. The critical flow conditions for entrainment of large joint blocks have been studied in the laboratory by Reinius (1986) and Coleman et al. (2003). Snyder et al. (2003b) provide one field-based estimate of a critical shear stress for block entrainment. Annandale (1995) presents a model for erosion limited by joint-block entrainment. However, because plucking entails the interaction of a complex suite of processes, including physical and chemical weathering, sand-wedging (Hancock et al. 1998), macroabrasion-driven rock fracture and crack propagation, as well as hydrodynamic block extraction, a quantitative mechanistic description of erosion by plucking has proven difficult to derive (Whipple et al. 2000a).

Where rocks are either not fractured at a scale that permits erosion by plucking and macroabrasion or are inherently weak at the grain-scale, erosion by wear is dominant. Both suspended load and bedload particles contribute to wear (Wohl 1993; Wohl et al.1994, 1999; Hancock et al. 1998; Sklar & Dietrich 1998, 2001; Whipple et al. 2000a; Hartshorn et al. 2002). Under these conditions, channel bed and banks are marked by smoothly polished rock surfaces sculpted by flutes and potholes associated with swirling vortex flows shed off topographic irregularities of all scales (Figure 2*A*,*B*) (Baker 1974, Wohl 1993, Hancock et al. 1998, Wohl et al. 1999, Whipple et al. 2000a). It is likely that abrasion in flutes and potholes is accomplished by the fine end of bedload and the coarse end of suspended load (maximum effective tool diameter $<$10% feature diameter is a reasonable order-of-magnitude estimate). The strongest vortices are shed in zones of flow separation on the downstream side of bed protuberances and bedrock steps. These vortices localize and concentrate abrasion, and further act to maximize grain-bed contact time as particles grind along margins of flutes and potholes in relatively continuous contact with the bed, thereby greatly enhancing the efficiency of incision by abrasion. Once flutes and potholes begin to form, a strong positive feedback develops because the developing microtopography of the erosional form enhances and stabilizes the vortex structure, further strengthening the localized attack of abrasion by both suspended and bedload particles. As a result, most observable abrasion wear occurs on the downstream side of flow obstructions and in the immediate vicinity of bedrock steps (Figure 2*A*,*B*). Mechanistic description of river incision by abrasion cannot be divorced from the irregular geometry of the riverbed and the associated macroturbulence (Baker 1974, Wohl 1993, Wohl & Ikeda 1997, Hancock et al.1998, Wohl et al. 1999, Whipple et al. 2000a).

Process Dominance and the Scaling-Up Problem

Scaling analyses of the various erosion processes indicate that they are governed by markedly different physics (Hancock et al. 1998, Whipple et al. 2000a). As these differences in the physics of erosion have far-reaching implications for landscape evolution and the coupling between climate, tectonics, and topography (e.g., Howard et al. 1994, Whipple & Tucker 1999, Whipple & Meade 2004), it is important to study the controls on process dominance as a function of climate state, substrate properties, sediment load, incision rate, and channel slope. However, in any given river system, the relative importance of individual processes likely varies in both space and time in response to substrate heterogeneities, knickpoint formation and migration, land use and longer-term climate change, downstream changes in the size and quantity of bedload, and the stochastic nature of storms and sediment delivery to channels.

Field observations indicate that where active, plucking of joint blocks is by far the most efficient erosional process (Whipple et al. 2000a,b, Snyder et al. 2003a). However, the relative roles of extraction of joint blocks (plucking plus macroabrasion) and wear integrated over space and time are unresolved. Whether plucking is active clearly depends on substrate properties, flow conditions, sediment load, and the relative rates of river incision and loose joint block production (Howard

1998, Whipple et al. 2000a). However, it is not clear where to draw the distinction between macroabrasion and plucking given that impacts by coarse bedload may be a key player in joint block production (Whipple et al. 2000a, Sklar & Dietrich 2001). For instance, in the small coastal streams studied by Snyder et al. (2000, 2003a), erosional morphologies indicate removal of joint blocks as the dominant incision mechanism, but, as confirmed by simple field experimentation, impacts by coarse bedload particles are clearly capable of dislodging joint blocks in these highly fractured rocks. Similarly, field observations indicate that the decimeter-scale block removal observed by Hartshorn et al. (2002) is driven by bedload impacts on well-jointed parts of the quartzite rib they studied.

Where abrasion appears to dominate, the relative contributions of bedload and suspended load to wear are uncertain, but they appear to vary as a function of drainage area, sediment load, and substrate properties. For instance, the Indus River where it crosses the western Himalaya in northwestern Pakistan is a big, powerful river cutting through hard, poorly jointed rock units with a relatively low bedload flux, and suspended-load abrasion and pothole erosion appear to be the dominant processes (Hancock et al. 1998). Channels incised into the resistant sandstones of the Colorado Plateau in Utah carry much sand in suspension and only a minimal amount of bedload, and yet steep reaches are marked by chains of polished chutes and potholes, often in narrow canyons with undulating walls (Figure 2*A*) (Wohl et al. 1999). It is unclear, however, whether these potholes represent vigorous abrasion by suspended load or an effective concentration of bedload owing to long residence times in potholes and more continuous contact with the bed. Similarly, suspended-load abrasion is clearly an important contributor (at least commensurate with bedload abrasion rates) in the steep channels draining the eastern slope of the Taiwan Central Range (Figure 2*B*) (Hartshorn et al. 2002). Conversely, large rivers crossing the western foothills of Taiwan's Central Range carry a considerable flux of coarse bedload as they cut through weak mudstones, and bedload abrasion and macroabrasion appear to be dominant (L. Sklar, personal communication, 2003). The plane-bed geometry of these channels ensures that particles carried in suspension will have little interaction with the bed. Researchers generally agree that bedload abrasion appears dominant in this type of system and concur that the combination of weak rock and planar beds inhibits pothole formation (Figure 2*D*) (Pazzaglia et al. 1998, Sklar & Dietrich 1998, Wohl & Ikeda 1998, Pazzaglia & Brandon 2001, Sklar & Dietrich 2001). Although marked by frequent small bedrock steps, the highly fractured rocks of the King Range drainages similarly show no evidence of an important role of suspended-load abrasion (Snyder et al. 2003a).

Perhaps more important than the question of transport mode (suspended load versus bedload abrasion) is the recognition that in most environments, abrasion wear is predominantly accomplished by particles vigorously spun in vortices (Figure 2*A*,*B*), as described above. Unraveling the exact nature of the relationship among erosion rate and climate state, substrate properties, channel slope, and sediment load (flux, hardness, and grain-size distribution) depends on a quantitative understanding of the different mechanics of plucking, macroabrasion, and wear, and how the above factors determine the relative contributions of each process.

Knickpoint Migration

Knickpoint migration has long been recognized as a primary mechanism of river response to sudden base-level fall. As sudden base-level fall can be triggered by tectonic upheaval, climatic change, sea-level fall, or river capture, knickpoint migration is an important aspect of channel incision and landscape evolution in a wide range of geologic, tectonic, and climatic settings. Indeed, in some environments, knickpoint propagation appears to be the dominant mode of channel lowering (e.g., Seidl et al. 1994, 1997, Wohl et al. 1994, Weissel & Seidl 1998, Stock & Montgomery 1999). Knickpoint propagation dominates incision where either the threshold stress required to initiate erosion is sufficiently large that it is only exceeded on steep knickpoints (e.g., Howard 1998) or where bed erosion is largely inhibited by a veneer of cobbles and boulders except on steep, bare-rock knickpoints (e.g., Seidl et al. 1994, 1997, Howard 1998). The fact that river incision is often episodic and associated with knickpoint formation and propagation, rather than a continuous process of channel bed lowering, presents a formidable complication to efforts to test and/or calibrate river incision models (see discussion and example given by Stock & Montgomery 1999); long-term incision rate may depend more on knickpoint gradient, the processes of knickpoint retreat, and the frequency of knickpoint formation than on the modern-day channel gradient. Field and laboratory studies (Holland & Pickup 1976, Gardner 1983, Seidl et al. 1994, 1997, Wohl et al. 1994, Weissel & Seidl 1998) also strongly suggest that the dominant processes, migration rates, and the form of knickpoints may be highly dependent on the mechanical properties of substrate lithology. Systematic study is needed to fully discover the interconnections among substrate lithology, sediment flux, knickpoint form (and its temporal evolution), and knickpoint migration rate.

RIVER INCISION MODELS

The ideal model of river incision would include physically based representation of all processes, including all intrinsic thresholds, with a minimum of parameters that are each directly measurable in the field or laboratory (Dietrich et al. 2003). Relative process dominance ought to emerge naturally as a function of conditions (channel slope, incision rate, sediment load characteristics, substrate properties, discharge, etc.). Internal relations would allow dynamic adjustment of channel width, percent bedrock exposure, hydraulic roughness, and bed material grain-size distribution (including the influence of the influx of large landslide-derived boulders). Finally, the ideal model would also have to deal with the scaling jump from local flow conditions (shear stress, velocity, flow accelerations, turbulence intensity, vorticity) to reach-averaged conditions useful to large-scale landscape evolution modeling and would have to handle the stochastic nature of floods and sediment supply.

Most published river incision models, including all those used in landscape evolution models, use generic formulations that describe the complex suite of interacting processes as a simple bulk relationship between mean bed shear stress or unit stream power and incision rate. Despite their simplicity, generic models can still usefully guide our intuition about which aspects of model formulation are critical to landscape evolution and how field data can be used to discriminate among models. Although generic models with tunable parameters are able to encompass a wide range of plausible behaviors (e.g., Whipple & Tucker 2002, Tomkin et al. 2003), it can be difficult to ascribe physical meaning to the values of model parameters that cannot be determined through measurement of material properties in the field or the laboratory (e.g., Dietrich et al. 2003).

The so-called stream-power family of models is based on the postulate that river incision rate depends on a power of mean bed shear stress and that this can be approximately described by relations for steady uniform flow (see Table 3) (Howard & Kerby 1983, Whipple & Tucker 1999):

$$E = k_e f(q_s)\left(\tau_b^a - \tau_c^a\right) \tag{3a}$$

$$E = k_e f(q_s)\left[1 - (\tau_c/\tau_b)^a\right]\tau_b^a \tag{3b}$$

$$\tau_b = k_t (Q/W)^{\alpha} S^{\beta}, \tag{3c}$$

where $f(q_s)$ denotes the influence of sediment load (flux, grain-size distribution) on incision rate, τ_b is the mean bed shear stress, τ_c is a threshold shear stress for incision, the exponent a is expected to vary with the dominant incision process (Hancock et al. 1998, Whipple et al. 2000a), and Q is water discharge. Note that under conditions of steady uniform flow, unit stream power scales with $\tau_b{}^{3/2}$, such that there is no meaningful difference between a model based on shear stress and one based on unit stream power (Whipple & Tucker 1999). By substituting well-known empirical relations among channel bankfull width (W), bankfull discharge (Q), and drainage area (A) (see Table 3) into Equations 3b and 3c and rearranging, a generalized form of the stream-power family of models can be written (Tucker & Bras 2000, Whipple & Tucker 2002, Snyder et al. 2003b):

$$E = K_r K_c K_{\tau_{cr}} f(q_s) A^m S^n, \tag{4}$$

where K_r represents erosional resistance (lithology, hydraulic roughness, channel width), K_c represents climatic conditions, and $K_{\tau cr}$ is a threshold term ($0 \leq K_{\tau cr} \leq 1$) set by the fraction of storm events that exceed the critical shear stress for incision (see Tucker & Bras 2000, Tucker 2004). Note that $K_{\tau cr}$ is the bracketed term in Equation 3b. Table 3 lists the definitions of K_r, K_c, $K_{\tau_{cr}}$, and $f(q_s)$ relevant to published variants of the stream-power model. The importance of the threshold term and its relation to the stochastic distribution of floods are discussed below. Tomkin et al. (2003), van der Beek et al. (2003), and Sklar (2003) each offer different,

TABLE 3 Equation sets for the stream power family of models

Shared internal relations			Shared exponents	
Hydrology[1]	**Hydraulic geometry**[2]	**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[3]	$K_R = k_e k_w^{-\alpha a} k_t^a$	$K_C = k_q^{\alpha a(1-b)}$	$K_{\tau_{cr}} = 1$	$f(q_s) = 1; Q_s/Q_c < 1$ $f(q_s) = \exp(-h/L); Q_s/Q_c \geq 1$
Saltation-abrasion[4]	$K_R = k_e k_w^{-\alpha a} k_t^a$	$K_C = k_q^{\alpha a(1-b)}$	N/A $\tau_{cr} > 0$; eqn 3	$f(q_s) = (Q_s/W)(1 - Q_s/Q_c)$
Stochastic-threshold[5]	$K_R = k_e k_w^{-\alpha a} k_t^a$	$K_C = \left(\frac{Tr}{Tr + Tb}\right) P^{\gamma_b} R_b^{-\varepsilon_b} \times \exp\left(\frac{-I}{P}\right) \Gamma(\gamma_b + 1)$	$K_{\tau_{cr}} = \frac{\Gamma\left(\gamma_b + 1, \frac{R_c}{P}\right) - \left(\frac{R_c}{P}\right)^{\gamma_b} \exp\left(\frac{-R_c}{P}\right)}{\Gamma(\gamma_b + 1)}$	$f(q_s) = 1; Q_s/Q_c < 1$ $f(q_s) = 0; Q_s/Q_c \geq 1$

[1] $Q_b = R_b A$.

[2] $W = k_w Q^b$ for bankfull flow in stochastic model.

[3] $WQL_c \propto$.

[4] $a = -0.88$.

[5] $\gamma_b = \alpha a(1 - s)$, $\varepsilon_b = \alpha a(b - s)$.

but analogous, formulations of generic river incision models. The generic forms proposed by Whipple & Tucker (2002) (see Table 3) encompass the undercapacity model (Beaumont et al. 1992) as a special case and capture some essentials of the saltation-abrasion model (Sklar & Dietrich 1998, Sklar 2003), although the important roles of grain-size and bedload saltation length in their model are represented only as a general power-law relationship between erosion rate and mean bed shear stress (through the *m* and *n* exponents; see Table 3). Another explicit limitation to the Whipple & Tucker (2002) analysis is the assumption that thresholds of sediment motion and bedrock erosion are negligible for the floods of interest. This will greatly change equilibrium slopes at low uplift rate (Sklar 2003, Snyder et al. 2003b, Tucker 2004).

No analysis has yet implemented both the stochastic/threshold ($K_{\tau cr}$) and sediment flux [$f(q_s)$] influences working in concert, much less handled multiple grain sizes and the stochastic supply of sediment to the channel. Such a model, at least in generic form, however, is within reach and may reveal many important insights about landscape dynamics and the interaction of climate and tectonics in orogen evolution. Still, as discussed above, there are many degrees of freedom in channel response to climate, lithology, and tectonics that have not yet been explicitly treated in either river incision or landscape evolution models. Stark & Stark (2001) present a first effort to parameterize all factors that influence channel sediment conductance and bedrock incision into a lumped channelization term. They propose a channelization evolution equation, complete with both positive and negative feedbacks. However, as described above, the field and laboratory data to defend or test the form of the proposed evolution equation for the channelization parameter do not yet exist.

Detailed models of the physics of individual incision processes (e.g., Foley 1980; Sklar & Dietrich 1998, 2001; Sklar 2003) can be very instructive regarding fundamental aspects of the river incision problem and also as a guide to incorporating the process in question in a more complete model. Indeed, the saltation-abrasion model of Sklar & Dietrich (Sklar & Dietrich 1998, 2001; Sklar 2003) and similar work by Slingerland et al. (1997) together constitute one of the most important contributions to the study of bedrock channels in the past decade. It is time for generic models to give way to more detailed process-based models. This advance is required if bedrock channel incision models are to become truly portable, predictive tools. However, models specific to a single process cannot hope to fully capture the richer dynamics of bedrock channels that involve the interaction of a suite of processes. In detail, the relationships among climate, tectonics and topography will depend on these richer dynamics, as discussed by Whipple & Meade (2004).

The most fully developed process-specific model is that of Sklar & Dietrich (1998, 2001) and Sklar (2003) for abrasion by saltating bedload alone. In their model, three multiplicative terms dictate abrasion by saltating bedload: volume removed per impact, impacts per unit time, and the fraction of bedrock exposed at the bed. Sklar (2003) combines these terms to write

$$E = \frac{Q_s w_{si}^2}{2WL_s\varepsilon_v}\left(1 - \frac{Q_s}{Q_c}\right), \tag{5}$$

where w_{si} is the vertical component of velocity on impact, L_s is the saltation hop length (a function of grain size and flow conditions), and ε_v is a measure of rock resistance to abrasion. The bracketed term is the bed-cover term. Grain size is explicitly modeled (it influences Q_c, w_{si}, and L_s) and has important effects, but as yet only a single size has been considered. The rolling and sliding components of bedload transport are ignored. Sklar (2003) uses empirical data on particle saltation to write w_{si} and L_s in terms of the mean bed Shields stress:

$$E = \left(\frac{\mathcal{R}_b g}{25\varepsilon_v}\left(\frac{\tau_*}{\tau_{*c}} - 1\right)^{-\frac{1}{2}}\right)\frac{Q_s}{W}\left(1 - \frac{Q_s}{Q_c}\right) \tag{6}$$

for $1 < \tau_*/\tau_{*c} < 10$ (i.e., below the threshold for incipient suspension of the sediment), where $\mathcal{R}_b$ denotes the submerged specific gravity of sediment, g is the gravitational acceleration, τ_* is the Shields stress, and τ_{*c} is the critical Shields stress for initiation of particle motion. The negative exponent on excess Shields stress reflects the fact that saltation length increases much more rapidly with τ_* than does the vertical component of particle impact velocity (Sklar 2003).

Sklar (2003) presents a comprehensive discussion of the formulation of this model, laboratory tests to calibrate material coefficients, comparison to previous models, and implications for lithologic, climatic, and tectonic controls on channel gradient and landscape relief. As noted by Sklar (2003), the most important limitations of this model for analysis of the interplay of climate, topography, and tectonics are (*a*) the restriction to one sediment grain size, (*b*) the restriction to erosion by saltating bedload alone, (*c*) the assumption of a planar bed geometry, (*d*) the simplistic formulation of the cover term, and (*e*) the incomplete treatment of the magnitude and frequency of stream flow and sediment supply events.

Published Tests of River Incision Models

Most tests of river incision models have been simple confirmations that a given model is capable of reproducing observed longitudinal profile form—the approximate power-law relation between local gradient and upstream drainage area (Equation 2) (Figure 4). However, it has long been known that a wide range of models can reproduce this relation (Howard 1980, Willgoose et al. 1991, Whipple & Tucker 2002), indicating that steady-state landscape morphology and incision rate patterns cannot be used to discriminate among models. One may interpret the Tomkin et al. (2003) analysis of incision rate patterns on the Clearwater River, Olympic Mountains, WA (assumed on the basis of available data to be at steady state) as also confirming this: All models tested fit the data equally, with nearly identical, uncorrelated residuals between observed and best-fit predicted incision rates in all cases. A partial exception to this rule exists where multiple realizations of steady-state morphologies (representing a range of drainage areas, uplift rates, and uplift rate gradients) are available for study, as in the Siwalik Hills of Nepal, for example (Figure 4) (Lave & Avouac 2000, Kirby & Whipple 2001).

Some models, however, do fail this most basic test, including the undercapacity model (Beaumont et al. 1992) and all hybrid models with $f(q_s) = (Q_s/W)(1-Q_s/Q_c)$, including the saltation-abrasion model (Sklar 2003). These models produce curving log slope–log area arrays in all cases (Whipple & Tucker 2002, Sklar 2003). This observation suggests that the formulation of these models is not complete. However, something as simple as hydraulic roughness variation with drainage area, or the intrabasin distribution of orographic rainfall (Roe et al. 2002), neither of which is included in any of these models at present, could rectify this mismatch. Furthermore, in some cases, the predicted curvature is subtle enough that it would be hard to use as a diagnostic criterion given the typical scatter in slope-area data (Figure 4).

Only a few studies have provided field calibrations of river incision model parameters (Howard & Kerby 1983; Seidl & Dietrich 1992; Rosenbloom & Anderson 1994; Seidl et al. 1994; Stock & Montgomery 1999; Snyder et al. 2000, 2003a,b; Whipple et al. 2000b; Kirby & Whipple 2001; Lave & Avouac 2001; Tomkin et al. 2003; van der Beek & Bishop 2003). Most tests assume that the incision process is steady and that channel gradient is not a function of time. Obviously, if incision is episodic and driven by knickpoint migration, this assumption is invalid, as nicely demonstrated by Stock & Montgomery (1999). Where knickpoint migration dominates, an analysis comparing incision rates to modern stream gradients can be rendered meaningless. This creates a major challenge, as knickpoint migration may not always leave a decipherable record—only field areas with exquisitely preserved histories of channel incision will be useful for quantitative testing of river incision models.

Although each of the above studies tests whether a given model accommodates field observations, only Tomkin et al. (2003) and van der Beek & Bishop (2003) attempt to test the relative merits of several alternate river incision models. The analysis by van der Beek & Bishop (2003) exploits transient river profile evolution and yields greater differentiation among models. The standard stream-power and linear undercapacity models (Beaumont et al. 1992) are most successful in fitting the observed patterns of river incision in this field setting. Best-fit parameters for the stream-power model are the same within error as those found earlier by Stock & Montgomery (1999).

A key contribution of theoretical analyses of both process-specific and generic models is the identification of testable hypotheses that can be used to evaluate critically model applicability and discriminate between successful and unsuccessful models. Further, these analyses illuminate what types of data are needed in various settings. Important examples include model predictions discussed by Sklar & Dietrich (1998, 2001), Whipple & Tucker (2002), Gasparini et al. (2002), and

Sklar (2003). The most important of these involve transient river response to known changes in the rate of base-level fall and are reviewed below. Much of this understanding is only just coming to light and has not yet been exploited widely in the field, although van der Beek & Bishop (2003) provide an important example.

DISCUSSION

Just as bedrock, mixed, and alluvial channels can be defined according to bed morphology (Howard et al. 1994), channel types can be classified on the basis of what factors dictate the channel incision rate. The incision rate in transport-limited systems (usually associated with alluvial channels) is by definition set by the downstream divergence of sediment transport capacity. The stream may incise into bedrock and still be transport limited, so long as it is the divergence of the transport capacity that limits the rate of incision (Howard 1980, Howard & Kerby 1983). Conversely, the incision rate in detachment-limited systems is by definition determined by the stream's ability to erode the bed, usually by a combination of abrasion and plucking (e.g., Baker 1974, Hancock et al. 1998, Wohl & Ikeda 1998, Whipple et al. 2000a). As most bedrock channels have mixed bedrock-alluvial bed morphologies, and sediment flux clearly plays an important role in the dominant processes of river incision into rock, the dynamics of bedrock channel behavior is likely to be some intermediate hybrid of both transport-limited and detachment-limited conditions, as reflected by incision models with a strong dependence on sediment load. Variable bedrock exposure in response to environmental conditions, frequent small knickpoints, and the occurrence of hanging tributaries in rapidly incising environments indicate bedrock rivers are more complex than transport-limited alluvial rivers, particularly in their response to perturbation. Although it is often assumed that smaller drainage areas, stronger rocks, and higher rock uplift rates lead to increasingly detachment-limited behavior, it remains unclear at present whether this is the case.

As Whipple & Tucker (2002) point out, the two clear diagnostics of transport-limited versus detachment-limited behavior are channel response to abrupt along-stream changes in substrate lithology or rock uplift rate and channel response to a sudden temporal increase in rock uplift rate. Only detachment-limited channels will exhibit abrupt knickpoints; transport-limited systems respond to integrated upstream conditions rather than local conditions. Numerous examples of abrupt knickpoints associated with lithologic and structural boundaries as well as recent increases in rock uplift have been documented (e.g., Figure 5). Thus, although it is plausible that bedrock channels approach a transport-limited condition at steady state (Sklar & Dietrich 1998, 2001; Whipple & Tucker 2002; Sklar 2003), it seems clear that the transient evolution of channel profiles is dictated by the processes of river incision into solid rock.

Importance of Transport/Incision Thresholds and the Stochastic Distribution of Floods

Mobilization of coarse bedload is a prerequisite for erosion of the channel bed, regardless of the dominant incision process. Bedload sediment transport is well known to require shear stresses above a threshold of motion (e.g., Graf 1977). Although the existence of this threshold and the likely existence of another higher threshold for incision into rock has been known since the earliest studies on bedrock channels (Baker 1974, Howard 1980, Howard & Kerby 1983), until recently, most models have dropped the threshold term, often for the convenience of working with analytical solutions (exceptions include Howard 1994; Densmore et al. 1998; Tucker & Bras 1998, 2000; Tucker et al. 2001). The argument was that the floods of interest (the geomorphically dominant discharge) greatly exceeded the threshold shear stress (Howard & Kerby 1983). For some problems, such as the controls on steady-state channel concavity in areas of uniform rock properties and uplift rate, this assumption is inconsequential for detachment-limited models, as pointed out by Howard & Kerby (1983). However, a threshold shear stress does significantly influence the concavity of transport-limited rivers (Tucker & Bras 1998, Gasparini et al. 2004, Tucker 2004). Moreover, a series of papers investigating the role of the stochastic distribution of flood magnitudes, durations, and frequencies (Tucker & Bras 2000, Snyder et al. 2003b, Tucker 2004) has demonstrated that for many problems of interest, the influence of the threshold term is of first-order significance and cannot be divorced from the full stochastic distribution of floods. In the stochastic-threshold model of Tucker & Bras (2000) and Tucker (2004), the combined influence of thresholds and a stochastic distribution of storms is captured in the $K_{\tau_{cr}}$ term of Equation 4 (see Table 3).

The presence of a threshold that is exceeded by only a fraction of flood events results in a nonlinear relation between channel gradient and rock uplift rate that approximates a power-law relation ($S \sim U^p$) with an exponent less than unity (Snyder et al. 2003b, Tucker 2004) such that channel gradient is most sensitive to tectonics at relatively low uplift rates (Figure 7). Snyder et al. (2003b) demonstrated with a field example that this effect fundamentally changes the relationship between steady-state channel gradient and tectonic rock uplift rate even where the critical shear stress to initiate bedrock incision is on the order of the threshold of motion for coarse bedload. Consequently, thresholds of motion and incision must be incorporated into bedrock incision models and cannot be accurately modeled without considering the integrated effect of the stochastic distribution of floods. Several laboratory studies have examined critical flow conditions for entrainment of large joint blocks (e.g., Reinius 1986, Coleman et al. 2003), as has at least one field study (Snyder et al. 2003b). Although macroabrasion likely requires exceedence of a critical threshold in either impact kinetic energy or exceedence of a critical amount of damage accumulation and fracture propagation (e.g., Miller & Dunne 1996), experimental data indicates that thresholds for abrasional wear are negligible (Sklar & Dietrich 2001, Sklar 2003).

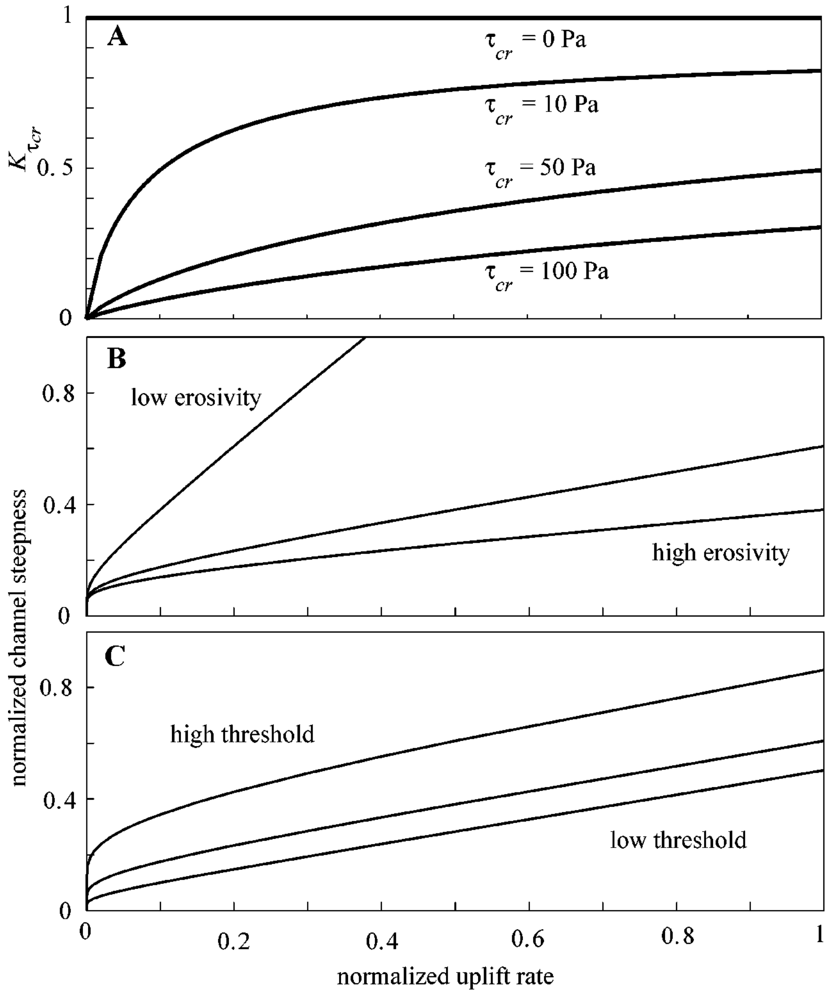

Figure 7 Influence of a critical threshold and a stochastic distribution of floods. (*A*) Threshold term in generalized stream-power model as a function of rock uplift rate for different values of the threshold. (*B*) Steady-state channel steepness (k_s in Equation 2, normalized to a reference value) as a function of rock uplift rate (normalized to a reference value) for different intrinsic erodibility of rock. (*C*) Steady-state normalized channel steepness as a function of normalized rock uplift rate for different thresholds. Modified from figures 3 and 5 in Snyder et al. (2003b), reproduced/modified by permission of the American Geophysical Union.

Steady-State Gradient and Uplift Rate

Topographic steady state is achieved wherever the long-term erosion rate balances the rock uplift rate and, consequently, the topography is statistically invariant over the long term. At steady state, all rivers must (*a*) incise the bed at a rate matching the local rock uplift rate and (*b*) transport all the sediment supplied to them from upstream. Therefore, as rock uplift rate increases, channel gradient must at a minimum increase sufficiently to carry the greater bedload sediment flux. Complications include possible dynamic adjustments of channel width, input sediment grain-size distribution (likely a function of the active hillslope erosion processes and the erosion rate), hydraulic roughness, and the degree of alluvial cover, as well as the possibility that standard transport and hydraulic relations break down when roughness becomes extreme ($D_{50bm}/H \sim 1$). The various bedrock incision models summarized in Table 3 predict varying degrees to which steady-state channel gradient increases beyond the transport slope as a function of rock erodibility, the fraction of bedload-sized material delivered to channels, and rock uplift rate. Where the steady-state gradient predicted by one incision process (e.g., saltation abrasion) exceeds that predicted by another (e.g., potholing or plucking), the process requiring the lower gradient to match the rock uplift rate would become the dominant process. In addition, two climate-related factors will dampen the gradient-uplift relationship relative to that predicted by simple models: (*a*) an increasing fraction of floods will exceed critical thresholds of transport and incision and contribute to incision on steeper channels (Figure 7) (Snyder et al. 2003b, Tucker 2004) and (*b*) enhanced orographic precipitation will further increase the frequency and duration of threshold-exceeding floods (e.g., Roe et al. 2002).

Given the above, it is not surprising that observed channel steepness indices in the fluvial zone ($A > A_{cr}$) normalized (k_{sn}) to a reference concavity index (θ_{ref}) vary systematically with rock uplift rate, nor is it surprising that the full range of observed values only spans slightly more than one order of magnitude ($20 \leq k_{sn} \leq 600$ for $\theta_{ref} = 0.45$). Although published estimates of the coefficient of erosion in the stream-power incision model (Equation 4) (see Table 3) vary over four orders of magnitude (Stock & Montgomery 1999), this is at odds with the observed range of channel gradients and probably reflects a combination of (*a*) the different controls on relatively short-term incision rates and equilibrium channel gradient and (*b*) the incomplete formulation of the stream-power model. For instance, the very low values of the coefficient of erosion found for channels in SE Australia by Stock & Montgomery (1999) and van der Beek & Bishop (2003) probably reflect an important influence of a critical threshold for bedload transport and/or bedrock incision in these low-gradient and slowly incising systems. Also, the highest estimates of the coefficient of erosion all come from channels etched into very weak lithologies (Howard & Kerby 1983, Stock & Montgomery 1999, Kirby & Whipple 2001) that form only a thin surficial layer that would be stripped off during the early stages of mountain building.

Sediment-Flux-Dependent River Incision

The importance of the bedload cover effect has been known qualitatively for over a century (Gilbert 1877); however, until the publication of Sklar & Dietrich's (1998) paper it was present in landscape evolution models only in one of three ways: (*a*) an assumption that channel incision is transport-limited, which implies the cover effect is dominant (e.g., Willgoose et al. 1991); (*b*) a caveat that the treatment applies only to detachment-limited conditions, implying the cover effect is negligible (e.g., Moglen & Bras 1995); or (*c*) a step-function threshold between detachment-limited (no cover effect) and transport-limited (complete cover) conditions when sediment supply increases to match the sediment carrying capacity (e.g., Howard 1980, 1994; Tucker & Slingerland 1996; Densmore et al. 1998). Since the publication of Sklar & Dietrich (1998), and because of the accumulating evidence that strath formation occurs at times of increased sediment flux (e.g., Pazzaglia et al. 1998, Pazzaglia & Brandon 2001, Hancock & Anderson 2002), the importance of the cover effect has become well established. However, as indicated in Table 3, considerable debate remains regarding the nature of the tools effect and the quantitative nature of the cover effect, i.e., the shape of the $f(q_s)$ function and the factors that control it remain uncertain. The problem is complex because to properly capture landscape dynamics, the $f(q_s)$ function (or equivalent) must capture the integrated effect of a stochastic distribution of floods and sediment supply events, acting on a bed of mixed grain sizes, and influencing a suite of competing and interacting incision processes. In addition, decadal to millennial climatic fluctuations can cause temporal variations in the amount of alluvial cover on timescales that are short compared to the river profile response time (e.g., Densmore et al. 1998, Whipple & Tucker 1999, Whipple 2001).

Experiments to date, while instructive, treat only steady flows, mono-sized sediment, and a single erosion process (Wohl & Ikeda 1997, Sklar & Dietrich 2001, Sklar 2003). The shape of the tools/cover function discovered in such experiments does not necessarily place strong constraints on conditions in natural environments. Potentially important unanswered questions include the following: (*a*) Does incision become inhibited by an active cloud of saltating bedload particles when sediment flux is still well below capacity, as in the saltation-abrasion model (Sklar & Dietrich 1998, 2001; Sklar 2003)? (*b*) Does incision instead become inhibited only when a sediment deposit thicker than flood scour depth is present, as suggested by Howard (1998) and Hancock & Anderson (2002)? (*c*) To what degree do the stochastic variability of flood magnitudes and sediment supply play a role, and how should this problem be formulated (for a preliminary treatment see Tucker et al. 2001)? (*d*) How does concentration of even minimal amounts of bedload in potholes and vortices influence the relation between the flux of tools and incision efficiency (see Wohl & Ikeda 1997)? (*e*) How does the competition and interaction among wear (by both suspended load and bedload), macroabrasion, and plucking influence the overall dependence of incision rate on sediment supply? Although it seems clear that a sediment tools/cover effect must be included in a successful river incision model, given current understanding, the range of plausible $f(q_s)$ functions includes almost any curve with a rising and falling limb. Field and laboratory data are required to narrow the range of plausible models.

Transient Response: The Underexploited, Critical Test

Despite the importance of transient response as the only truly discriminating test, Howard & Kerby (1983), Stock & Montgomery (1999), Whipple et al. (2000b), and van der Beek & Bishop (2003) are the only published studies to date that effectively exploit transient river profile adjustments to test and/or calibrate models. Model predictions of transient river profile response to sudden changes in base level (or the rate of relative base-level fall) can be diagnostic, even when multiple models successfully predict the relationship between steady-state steepness and concavity as a function of rock uplift rate. For instance, as discussed by Whipple & Tucker (2002), transport- and detachment-limited models exhibit diffusive and kinematic-wave behaviors, respectively (Figure 8). As illustrated in Figure 8*A*, kinematic-wave behavior is characterized by an abrupt jump in channel gradient, from the initial to the final steady-state value, that sweeps upstream as a discrete knickpoint (abrupt change in gradient). As illustrated in Figure 8*B*, transport-limited systems do not retain knickpoints and transients are characterized by quasi-synchronous adjustments of channel gradient throughout the drainage network; the change in channel gradient sweeps upstream in a diffusive manner, only asymptotically approaching the final steady-state value. Thus, disequilibrium conditions in transport-limited systems can be recognized only by subtle differences in concavity index (θ is reduced if uplift rates have increased). Detachment-limited systems exhibit a kinematic-wave response associated with upstream-migrating knickpoints (either abrupt changes in bed elevation or gradient) and disequilibrium conditions are readily recognized (Figure 8) (Whipple & Tucker 2002).

Hybrid models—those in which sediment load and sediment transport capacity play important roles—exhibit combinations of both kinematic-wave and diffusive responses, depending on the magnitude and direction of the perturbation to the rate of base-level fall (Figure 8) (Gasparini et al. 2002, Whipple & Tucker 2002). Many interesting and complex transient responses are possible and will differ depending on model details. The fact that hillslope, and therefore sediment flux, response must lag behind channel response—which must propagate upstream from the basin mouth or the point of relative base-level fall—is the key to the complex responses that are predicted.

As pointed out by Whipple & Tucker (2002), the lagged response in sediment flux means that hybrid channels will tend toward transport-limited conditions during declining-state transients and toward detachment-limited conditions during rising-state transients, with the implication that there will be a certain hysteresis in landscape evolution: largely detachment-limited, kinematic-wave behavior in response to increasing uplift rate, and diffusive, transport-limited behavior in response to decreasing uplift rate. This suggests that whereas deviations from

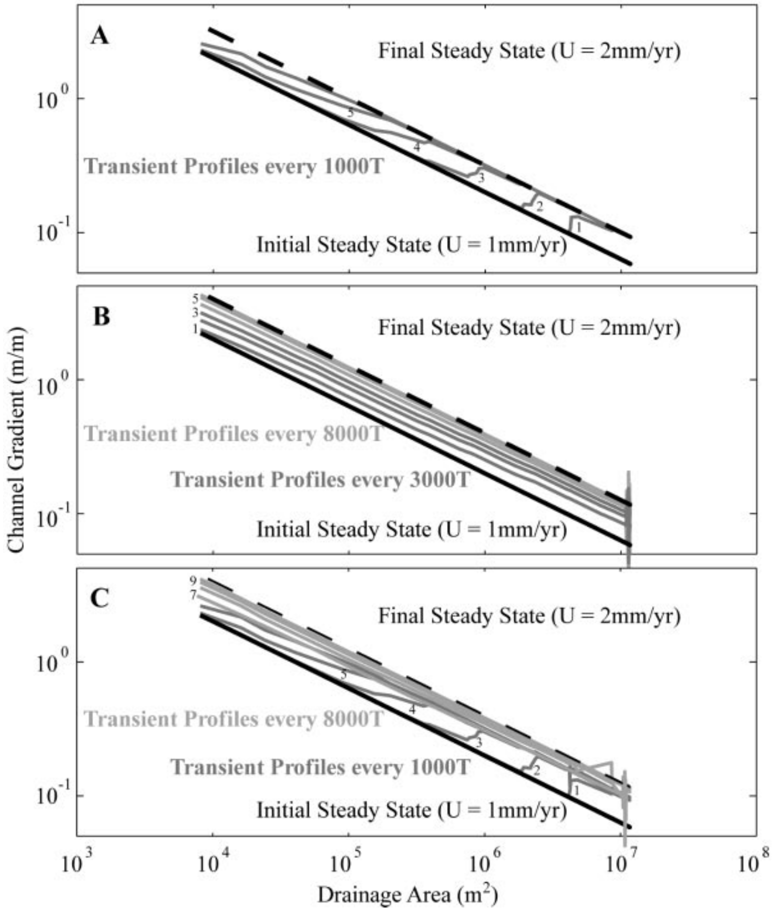

Figure 8 Transient profile evolution in response to a sudden increase in rock uplift rate (slope-area space). Initial steady state (*solid black*), final steady state (*dashed black*), and transient time steps (numbered sequentially; *dark gray*: every 1000 timesteps; *light gray*: every 8000 timesteps) are shown in each case. Corresponding channel profiles (plots of elevation versus distance) are shown by Whipple & Tucker (2002). (*A*) Detachment-limited case: kinematic wave response. Note that the smearing of the initially abrupt slope break as it migrates upstream is an artifact of numerical diffusion. (*B*) Transport-limited case: diffusive response. (*C*) Hybrid case: initial kinematic wave followed by diffusive response. (Modified from Whipple & Tucker (2002), figures 7b, 8b, and 10d; reproduced/modified by permission of the American Geophysical Union.

steady-state forms due to recent increases in the rate of base-level fall may be readily recognizable in analyses of stream profiles, the characteristic forms identified by Willgoose (1994) for declining-state equilibrium in transport-limited systems may confound any such analysis of decreases in base-level fall. This is consistent with available observational data to date: Whereas knickpoints associated with an increase in channel gradient are common (e.g., Figure 5), there are no known cases of knickpoints consistent with a sudden decrease in channel gradient such as might be associated with a deceleration in the rate of base-level fall. However, at some level, this type of hysteresis should be common to all hybrid models; it is only a good test as to whether end-member transport- and detachment-limited models should be discarded.

In detail, the initial wave-like response expected from hybrid models will differ significantly as a function of how sensitive the model is to changes in mean bed shear stresses and the alluvial cover term. This follows because the initial response to an increase in the rate of base-level fall will always be a steepening of the channel without a commensurate increase in sediment flux (Gasparini et al. 2002). If the forcing is strong enough—which depends on model sensitivity (models with lower values of the exponent n in Equation 4 or the equivalent being more sensitive)—the lower reach of the channel will oversteepen relative to its eventual steady-state gradient. Indeed, the transient channel gradient will grow without bound in the most sensitive models (Sklar & Dietrich 1998, Sklar 2003), producing waterfalls and/or hanging tributaries. Channel gradients, however, will rapidly decline to less-than-steady-state values downstream, in the wake of the migrating knickpoints, producing locally high concavity indices (Gasparini et al. 2002), as has been noted in a few field examples (e.g., Figure 5) (VanLaningham 2003, L.M. Schoenbohm, K. Whipple, B.C. Burchfiel & L. Chen, manuscript in review). Although transient responses of even generic hybrid river incision models have not yet been systematically explored, one may anticipate that channel responses to various magnitudes of transient perturbation promise to be the most exacting of tests for any river incision model. However, there are very likely to be complications associated with transient adjustments in channel width, bed roughness, and possibly the dominant erosional mechanisms. Comprehensive, quantitative field data on channels caught “in the act” of responding to recent increases in base-level fall are greatly needed.

CONCLUSIONS AND RESEARCH NEEDS

A primary conclusion of this review is that although much progress has been made, fundamental field and laboratory data on many aspects of the bedrock river problem are sorely lacking. However, it is now clear what data are needed in various types of settings, as detailed above. Collection of these data should be directed toward development of a new generation of river incision models that include physically based representation of the full suite of fluvial erosion processes and their interactions, such that relative process dominance emerges naturally as a function of conditions (channel slope, incision rate, sediment load characteristics, substrate properties, flood frequencies, etc.). Internal relations that allow dynamic adjustment of channel width, degree of bedrock exposure, hydraulic roughness, and bed material grain-size distribution need to be developed and tested against field data. An objective and quantitative in situ measure of rock erodibility over appropriate length scales is badly needed. In addition, the scaling jump from local flow conditions (shear stress, velocity, flow accelerations, turbulence intensity, vorticity) to reach-averaged conditions useful to large-scale landscape evolution modeling must be bridged. Finally, predicting long-term incision rates relevant to landscape evolution requires integration over thousands of years of discrete flood and sediment supply events. Careful laboratory experimentation, instrumentation of rapidly evolving field sites, and exploitation of well-constrained natural experiments in river profile evolution promise to be the most fruitful avenues of pursuit.

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LITERATURE CITED

Annandale GW. 1995. Erodibility. *J. Hydraul. Res.* 33:471–94

Baker V. 1974. Erosional forms and processes from the catastrophic Pleistocene Missoula floods in eastern Washington. In *Fluvial Geomorphology*, ed. M Morisawa, pp. 123–48. London: Allen Unwin

Baker VR, Kale VS. 1998. The role of extreme floods in shaping bedrock channels. See Tinkler & Wohl 1998, pp. 153–65

Beaumont C, Fullsack P, Hamilton J. 1992. Erosional control of active compressional orogens. In *Thrust Tectonics*, ed. KR McClay, pp. 1–18. New York: Chapman Hall

Benda L, Dunne T. 1997. Stochastic forcing of sediment routing and storage in channel networks. *Water Resour. Res.* 33:2865–80

Blythe AE, Burbank DW, Farley KA, Fielding EJ. 2000. Structural and topographic evolution of the central Transverse Ranges, California, from apatite fission-track, (U-Th)/He and digital elevation model analyses. *Basin Res.* 12:97–114

Brocklehurst S, Whipple K. 2002. Glacial erosion and relief production in the Eastern Sierra Nevada, California. *Geomorphology* 42:1–24

Burbank DW, Leland J, Fielding E, Anderson RS, Brozovic N, et al. 1996. Bedrock incision, rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature* 379:505–10

Coleman SE, Melville BW, Gore L. 2003. Fluvial entrainment of protruding fractured rock. *J. Hydraul. Eng.* 129:872–84

Dahlen FA, Suppe J. 1988. Mechanics, growth, and erosion of mountain belts. In *Processes in Continental Lithospheric Deformation*, ed. SPJ Clark, BC Burchfiel, J Suppe, pp. 161–78. Denver, CO: Geol. Soc. Am.

Densmore AL, Ellis MA, Anderson RS. 1998. Landsliding and the evolution of normal-fault-bounded mountains. *J. Geophys. Res. B Solid Earth Planets* 103:15203–19

Dietrich WE, Bellugi D, Heimsath AM, Roering JJ, Sklar L, Stock JD. 2003. Geomorphic transport laws for predicting the form evolution of landscapes. In *Prediction in Geomorphology*, ed. PR Wilcock, R Iverson, pp. 103–32. Washington, DC: Am. Geophys. Union

Duvall A, Kirby E, Burbank D. 2003. *Bedrock channel response to rock strength and rock uplift rate in the Santa-Ynes Mountains, California*. Annu. Meet. *Geol. Soc. Am.,* Seattle. *Abstr. Prog.*

Foley MG. 1980. Bed-rock incision by streams. *Geol. Soc. Am. Bull. Part II* 91:2189–213

Gardner TW. 1983. Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material. *Geol. Soc. Am. Bull.* 94:664–72

Gasparini N, Bras RL, Whipple K. 2002. Examining the effects of a sediment-flux-dependent river incision rule on the evolution of river networks. *Eos Trans. Am. Geophys. Union* 83:F534

Gasparini NM, Tucker GE, Bras RL. 2004. Network-scale dynamics of grain-size sorting: implications for downstream fining, stream-profile concavity, and drainage basin morphology. *Earth Surf. Process. Landf.* In press

Gilbert GK. 1877. *Geology of the Henry Mountains (Utah)*. Washington, DC: U.S. Gov. Print. Off. 160 pp.

Graf WH. 1977. *Hydraulics of Sediment Transport*. New York: McGraw-Hill. 513 pp.

Hack JT. 1957. Studies of longitudinal stream profiles in Virginia Maryland. *U.S. Geol. Surv. Prof. Pap.* 294–B:97

Hancock GS, Anderson RS. 2002. Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. *Geol. Soc. Am. Bull.* 114:1131–42

Hancock GS, Anderson RS, Whipple KX. 1998. Beyond power: bedrock river incision process and form. See Tinkler & Wohl 1998, pp. 35–60

Harbor DJ. 1998. Dynamic equilibrium between an active uplift and the Sevier River, Utah. *J. Geol.* 106:181–98

Harbor DJ, Schumm SA, Harvey MD. 1994. Tectonic control of the Indus River in Sindh, Pakistan. In *The Variability of Large Rivers*, ed. SA Schumm, BR Winkley, pp. 161–75. New York: Am. Soc. Civ. Eng.

Hartshorn K, Hovius N, Dade WB, Slingerland R. 2002. Climate-driven bedrock incision in an active mountain belt. *Science* 297:2036–38

Holland WN, Pickup G. 1976. Flume study of knickpoint development in stratified sediment. *Geol. Soc. Am. Bull.* 87:76–82

Howard AD. 1980. Thresholds in river regimes. In *Thresholds in Geomorphology*, ed. DR Coates, JD Vitek, pp. 227–58. Boston: Allen Unwin

Howard AD. 1994. A detachment-limited model of drainage basin evolution. *Water Resour. Res.* 30:2261–85

Howard AD. 1998. Long profile development of bedrock channels: interaction of weathering, mass wasting, bed erosion, and sediment transport. See Tinkler & Wohl, pp. 297–319

Howard AD, Kerby G. 1983. Channel changes in badlands. *Geol. Soc. Am. Bull.* 94:739–52

Howard AD, Seidl MA, Dietrich WE. 1994. Modeling fluvial erosion on regional to continental scales. *J. Geophys. Res.* 99:13971–86

Kirby E, Whipple K. 2001. Quantifying differential rock-uplift rates via stream profile analysis. *Geology* 29:415–18

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. *J. Geophys Res.* 108:doi: 10.1029/2001JB000861

Koons PO. 1989. The topographic evolution of collisional mountain belts: a numerical look at the Southern Alps, New Zealand. *Am. J. Sci.* 289:1041–69

Lave J, Avouac J-P. 2000. Active folding of fluvial terraces across the Siwalik Hills, Himalayas of central Nepal. *J. Geophys. Res.* 105:5735–70

Lave J, Avouac J-P. 2001. Fluvial incision and tectonic uplift across the Himalayas of central Nepal. *J. Geophys. Res.* 106:26561–91

Massong TM, Montgomery DR. 2000. Influence of sediment supply, lithology, and wood debris on the distribution of bedrock and alluvial channels. *Geol. Soc. Am. Bull.* 112:591–99

Miller JR. 1991. The influence of bedrock geology on knickpoint development and channel-bed degradation along downcutting streams in south-central Indiana. *J. Geol.* 99:591–605

Miller DJ, Dunne T. 1996. Topographic perturbations of regional stresses and consequent bedrock fracturing. *J. Geophys. Res. B Solid Earth Planets* 101:25,523–36

Moglen GE, Bras RL. 1995. The effect of spatial heterogeneities on geomorphic expression in a model of basin evolution. *Water Resour. Res.* 31:2613–23

Montgomery DR, Abbe TB, Buffington JM, Peterson NP, Schmidt KM, Stock JD. 1996. Distribution of bedrock alluvial channels in forested mountain drainage basins. *Nature* 381:587–89

Montgomery DR, Buffington JM. 1997. Channel-reach morphology in mountain drainage basins. *Geol. Soc. Am. Bull.* 109: 596–611

Montgomery DR, Finnegan N, Anders A, Hallet B. 2002. Downstream adjustment of channel width to spatial gradients in rates of rock uplift at Namche Barwa. *Abstr. Progr. Geol. Soc. Am.* 34:241

Montgomery DR, Foufoula-Georgiou E. 1993. Channel network representation using digital elevation models. *Water Resour. Res.* 29:1178–91

Montgomery DR, Gran KB. 2001. Downstream variations in the width of bedrock channels. *Water Resour. Res.* 37:1841–46

Parker G. 1978. Self-formed straight rivers with equilibrium banks mobile bed. Part II. The gravel river. *J. Fluid Mech.* 89:127–48

Parker G, Toro-Escobar C, Ramey M, Beck S. 2003. Effect of floodwater extraction on mountain stream morphology. *J. Hydraul. Eng.* 129:885–95

Pazzaglia FJ, Brandon MT. 2001. A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State. *Am. J. Sci.* 301:385–431

Pazzaglia FJ, Gardner TW, Merritts DJ. 1998. Bedrock fluvial incision and longitudinal profile development over geologic time scales determined by fluvial terraces. See Tinkler & Wohl 1998, pp. 207–35

Reinius E. 1986. Rock Erosion. *Water Power Dam Constr.* 38: 43–48

Roe GH, Montgomery DR, Hallet B. 2002. Effects of orographic precipitation variations on the concavity of steady-state river profiles. *Geology* 30:143–46

Rosenbloom NA, Anderson RS. 1994. Hillslope and channel evolution in a marine terraced landscape, Santa Cruz, California. *J. Geophys. Res.* 99:14013–29

Schmidt KM, Montgomery DR. 1995. Limits to relief. *Science* 270:617–20

Selby MJ. 1980. A rock mass strength classification for geomorphic purposes with tests from Antarctica and New Zealand. *Z. Geomorphol.* 24:31–51

Seidl M, Dietrich WE, Kirchner JW. 1994. Longitudinal profile development into bedrock: an analysis of Hawaiian channels. *J. Geol.* 102:457–74

Seidl MA, Dietrich WE. 1992. The problem of channel erosion into bedrock. *Catena Suppl.* 23:101–24

Seidl MA, Finkel RC, Caffee MW, Hudson GB, Dietrich WE. 1997. Cosmogenic isotope analyses applied to river longitudinal profile evolution: problems and interpretations. *Earth Surf. Process. Landf.* 22:195–209

Shepherd RG, Schumm SA. 1974. Experimental study of river incision. *Geol. Soc. Am. Bull.* 85:257–68

Sklar L. 2003. *The influence of grain size, sediment supply, and rock strength on rates of river incision into bedrock*. PhD thesis. Univ. Calif., Berkeley. 343 pp.

Sklar L, Dietrich WE. 1998. River longitudinal profiles and bedrock incision models: stream power and the influence of sediment supply. See Tinkler & Wohl 1998, pp. 237–60

Sklar LS, Dietrich WE. 2001. Sediment and rock strength controls on river incision into bedrock. *Geology* 29:1087–90

Slingerland R, Willett SD, Hennessey HL. 1997. A new fluvial bedrock erosion model based on the work-energy principle. *Eos Trans. Am. Geophys. Union* 78:F299 (Fall Meet. Suppl.)

Snyder N, Whipple K, Tucker G, Merritts D. 2000. Landscape response to tectonic forcing: digital elevation model analysis of stream profiles in the Mendocino triple junction region, northern California. *Geol. Soc. Am. Bull.* 112:1250–63

Snyder NP, Whipple KX, Tucker GE, Merritts DM. 2003a. Channel response to tectonic forcing: analysis of stream morphology and hydrology in the Mendocino triple junction region, northern California. *Geomorphology* 53:97–127

Snyder NP, Whipple KX, Tucker GE, Merritts DM. 2003b. Importance of a stochastic distribution of floods and erosion thresholds in the bedrock river incision problem. *J. Geophys. Res.* 108:doi: 10.1029/2001JB001655

Stark CP, Stark GJ. 2001. A channelization model of landscape evolution. *Am. J. Sci.* 301:486–512

Stock JD, Dietrich WE. 2003. Valley incision by debris flows: evidence of a topographic signature. *Water Resour. Res.* 39:doi: 10.1029/2001WR001057

Stock JD, Montgomery DR. 1999. Geologic constraints on bedrock river incision using the stream power law. *J. Geophys. Res.* 104:4983–93

Strahler AN. 1950. Equilibrium theory of erosional slopes approached by frequency distribution analysis. *Am. J. Sci.* 248:673–96

Suzuki T. 1982. Rate of lateral planation by Iwaki River, Japan. *Trans. Jpn. Geomorphol. Union* 3:1–24

Tarboton DG, Bras RL, Rodriguez-Iturbe I. 1989. Scaling and elevation in river networks. *Water Resour. Res.* 25:2037–51

Tinkler K, Wohl EE. 1998. *Rivers Over Rock: Fluvial Processes in Bedrock Channels*. Washington, DC: Am. Geophys. Union

Tomkin JH, Brandon MT, Pazzaglia FJ, Barbour JR, Willett SD. 2003. Quantitative testing of bedrock incision models for the Clearwater River, NW Washington State. *J. Geophy. Res.* 108:2308; doi: 10.1029/2001JB000862

Tucker G. 2004. Drainage basin sensitivity to tectonic climatic forcing: implications of a stochastic model for the role of entrainment and erosion thresholds. *Earth Surf. Process. Landf.* doi: 10.1002/esp.1020

Tucker GE, Bras RL. 1998. Hillslope processes, drainage density, and landscape morphology. *Water Resour. Res.* 34:2751–64

Tucker GE, Bras RL. 2000. A stochastic approach to modeling the role of rainfall variability in drainage basin evolution. *Water Resour. Res.* 36:1953–64

Tucker GE, Lancaster ST, Gasparini NM, Bras RL. 2001. The channel-hillslope integrated landscape development model (CHILD). In *Landscape Erosion Evolution and Modeling*, ed. RS Harmon, WWI Doe, pp. 349–88. New York: Kluwer Academic/Plenum Publ.

Tucker GE, Slingerland R. 1996. Predicting sediment flux from fold thrust belts. *Basin Res.* 8:329–49

Tucker GE, Whipple KX. 2002. Topographic outcomes predicted by stream erosion models: sensitivity analysis and intermodel comparison. *J. Geophys. Res.* 107:doi: 10.1029/2001JB000162

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. *J. Geophys. Res.* 108:2309; doi: 10.1029/2002JB002125

VanLaningham S. 2003. *The role of rock resistance and rock uplift on topographic relief and river longitudinal profiles in the coastal mountains of Oregon and a landscape scale test for steady-state conditions*. MS thesis. Oregon State Univ., Corvallis. 200 pp.

Weissel JK, Seidl MA. 1998. Inland propagation of erosional escarpments and river profile evolution across the southeastern Australian passive continental margin. See Tinkler & Wohl 1998, pp. 189–206

Whipple K. 2001. Fluvial landscape response time: how plausible is steady state denudation? *Am. J. Sci.* 301:313–25

Whipple K, Kirby E, Brocklehurst S. 1999. Geomorphic limits to climatically induced increases in topographic relief. *Nature* 401:39–43

Whipple K, Meade BJ. 2004. Controls on the strength of coupling among climate, erosion, and deformation in two-sided, frictional orogenic wedges. *J. Geophys. Res.* In press

Whipple K, Tucker G. 2002. Implications of sediment-flux dependent river incision models for landscape evolution. *J. Geophys. Res.* 107:doi: 10.1029/2000JB000044

Whipple KX, Hancock GS, Anderson RS. 2000a. River incision into bedrock: mechanics and relative efficacy of plucking, abrasion, and cavitation. *Geol. Soc. Am. Bull.* 112:490–503

Whipple KX, Snyder NP, Dollenmayer K. 2000b. Rates and processes of bedrock incision by the Upper Ukak River since the 1912 Novarupta ash flow in the Valley of Ten Thousand Smokes, Alaska. *Geology* 28:835–38

Whipple KX, Tucker GE. 1999. Dynamics of the stream-power river incision model: implications for height limits of mountain ranges, landscape response timescales, and research needs. *J. Geophys. Res.* 104:17661–74

Willett SD. 1999. Orogeny orography: the effects of erosion on the structure of mountain belts. *J. Geophys. Res.* 104:28957–81

Willgoose G. 1994. A physical explanation for an observed area-slope-elevation relationship for catchments with declining relief. *Water Resour. Res.* 30:151–59

Willgoose G, Bras RL, Rodriguez-Iturbe I. 1991. A coupled channel network growth and hillslope evolution model. 1. Theory. *Water Resour. Res.* 27:1671–84

Wobus CW, Hodges K, Whipple K. 2003. Has focused denudation at the Himalayan topographic front sustained active thrusting near the Main Central Thrust? *Geology* 31:861–64

Wohl EE. 1993. Bedrock channel incision along Picanniny Creek, Australia. *J. Geol.* 101:749–61

Wohl EE, Greenbaum N, Schick AP, Baker VR. 1994. Controls on bedrock channel incision along Nahal Paran, Israel. *Earth Surf. Process. Landf.* 19:1–13

Wohl EE, Ikeda H. 1997. Experimental simulation of channel incision into a cohesive substrate at varying gradients. *Geology* 25:295–98

Wohl EE, Ikeda H. 1998. Patterns of bedrock channel erosion on the Boso Peninsula, Japan. *J. Geol.* 106:331–45

Wohl EE, Merritt DM. 2001. Bedrock channel morphology. *Geol. Soc. Am. Bull.* 113:1205–12

Wohl EE, Thompson DM, Miller AJ. 1999. Canyons with undulating walls. *Geol. Soc. Am. Bull.* 111:949–59