

Long Profile Development of Bedrock Channels: Interaction of Weathering, Mass Wasting, Bed Erosion, and Sediment Transport

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Erosion of bedrock channels seldom involves solely hydraulic processes such as plucking, abrasion, and solution. Weathering, mass wasting and burial by sediment cover modulate the rate of bedrock erosion. In headwater channels weathering generally must reduce rock strength to the point that entrainment may occur by hydraulic processes or rapid mass wasting. Simple quantitative models are introduced that demonstrate how erosion rates can depend upon both weathering rate and stress applied by moving fluids and debris. Rockfalls and avalanches can trigger additional failures in partially weathered bedrock on lower parts of alpine bedrock slopes before they would fail solely by weathering; this generates an economy of scale that results in development of spur and gully landforms. Streambed weathering also enhances bed erosion by water and debris flow in headwater bedrock channels within moderate relief landscapes. In large bedrock channels erosion rates are controlled both directly and indirectly by the throughflowing sediment. Abrasion by bedload and suspended load is often the dominant process. The rock beds of many streams are mantled partially or shallowly by alluvium. Two primary issues are unresolved about long-term evolution of these mixed bedrock-alluvial channels: 1) how and when the bedrock is eroded and 2) whether the gradient is determined by the necessity to transport the alluvium or to erode the bed. A semi-quantitative model suggests that bed erosion occurs due to exposure during extreme floods, at the base of migrating bedforms, and during periods of low sediment influx. Erosion rates in rapidly downcutting bedrock channel reaches are often regulated by influx of boulders that partially or wholly mantle the bed. These locally contributed boulders are primarily derived from steepening of sideslopes and tributaries due to the rapid incision.

INTRODUCTION

Present understanding of the processes and evolution of bedrock channels lags significantly behind that for alluvial channels. Little is known about the distribution of bedrock channels, process models are primitive and incomplete, quantitative field measurements are rare and difficult to

make, and the erosional history evolution of such channels is largely uncertain. This lack of quantitative characterization of erosional processes in bedrock channels is unfortunate, because such channels are widespread in high-relief terrain and the pace of long-term erosion and the overall relief is often governed by channel bed erodibility [e.g. *Burbank et al.*, 1996]. Realistic modeling of the interactions between tectonic deformation, landform development and erosion, and sedimentation processes will require better characterization of bedrock channel erosion. The same is true for prediction of the effects of short term climatic and land use changes upon channel morphology and sediment transport.

This paper explores three related topics. The first relates to the difficulty in making *a-priori* predictions about the distribution of channel types due to the interplay of controlling factors. The second issue is that erosion of bedrock channels seldom can be characterized solely as a relationship between applied fluid force and the rate of channel bed incision. Rather, erosion most often involves interplay between weathering, mass-wasting, sediment transport and fluid motion. This interplay is particularly important for the common mixed bedrock-alluvial channels that constitute the third topic.

These issues are first explored for headwater channels in which weathering, mass-wasting, and fluvial erosion interact. Further downstream in the stream network, weathering becomes less important, but bed erosion still involves the interaction of sediment transport, local mass wasting, and hydraulics.

HEADWATER BEDROCK CHANNELS

Channels and hollows forming the headward tips of the drainage network are commonly bedrock floored, at least episodically. These low-order channels differ from large bedrock streams in that the erosional process involves a mixture of weathering and rapid mass wasting in addition to fluvial erosion. Furthermore, because of the convergent topography of headwater hollows, colluvial infilling competes with erosion [e.g., Dietrich and Dunne, 1978; Dietrich *et al.*, 1982; Montgomery and Dietrich, 1994; Dietrich *et al.*, 1995]. Because the mixture of processes varies between landscapes, few generalizations are possible.

The simplest headwater bedrock channels occur when the main concentrative erosional process is fluvial erosion rather than rapid mass movement. In order to sustain a topographic hollow the long-term fluvial erosion along the hollow axis must be sufficient to erode colluvial infilling from superjacent slopes as well as the bedrock in the channel bed. In Howard's [1994a] drainage basin model, it is envisioned that during each simulation timestep runoff first erodes colluvial infilling and then the underlying bedrock. On convex and straight hillslopes, however, the colluvial flux is sufficient to prevent permanent channel development, although ephemeral rills may occur. In badland landscapes, the cycles of colluvial infilling and fluvial scour may follow simple seasonal patterns (Figure 1), with attendant growth and retreat of the fluvial network [Schumm, 1956; Howard and Kerby, 1983]. More typically, epicycles of infilling and scour may occur over much longer timescales [e.g., Hack and Goodlett, 1960; Dietrich and Dunne, 1978; Reneau *et al.*, 1989].

Interaction of Weathering and Fluvial Scour

The mechanism of fluvial erosion in headwater bedrock channels has received little study. Howard and Kerby [1983] and Howard [1994a] proposed that the rate of bedrock erosion is proportional to the shear stress exerted by runoff, with an implicit assumption that the bedrock can be directly scoured by runoff. In badland landscapes developed on weak sediments or saprolite, this assumption may be appropriate. Howard and Kerby [1983] showed that the spatial pattern of erosion rates in badlands on Coastal Plain sediments in Virginia is consistent with erosion being proportional to the shear stress exerted by runoff. In most other rock types, however, the bedrock must be partially weathered prior to fluvial erosion. Howard [1994b] presented a conceptual model of how weathering and detachment might interact in headwater channels. Assume that the flow characteristically removes weathered layers of thickness δd (e.g., weathered shale chips or exfoliation sheets) and that the weathering at that depth decreases the shearing resistance C at a negative exponential rate from the initial cohesion C_i to a minimum value C_f

$$C = C_f + (C_i - C_f) e^{-\lambda \delta t}, \quad (1)$$

where δt is the elapsed time since weathering has begun and λ is a characteristic weathering rate that might depend upon wetting duration, bedrock or regolith permeability, and the substrate physico-chemical properties. Detachment of the weathered layer occurs when $\tau \geq C$, which occurs after a time

$$\delta t = \frac{1}{\lambda} \ln \left[\frac{(C_i - C_f)}{(\tau - C_f)} \right]. \quad (2)$$

Assuming that weathering begins anew when a layer is stripped by the flow and that τ_e is the effective shear stress, then the average erosion rate would be given by

$$\frac{\partial z}{\partial t} = - \frac{\partial d}{\delta t} = - \delta d \lambda / \ln \left[\frac{(C_i - C_f)}{(\tau_e - C_f)} \right]. \quad (3)$$

Although (3) suggests that the rate of erosion would increase with layer thickness, δd , the intrinsic weathering rate should decrease with depth beneath the surface. For example, if

$$\lambda \propto \frac{1}{\delta d^k}, \quad (4)$$



Figure 1. Exposed shale bedrock in badlands in an abandoned claypit, Shenandoah Valley near Vesuvius, Virginia. Sunlight glints on shale surface. Badland slopes are underlain by a thin weathered shale regolith. Picture taken in early fall. Mass wasting from adjacent slopes mantles the bedrock floor during the winter season.

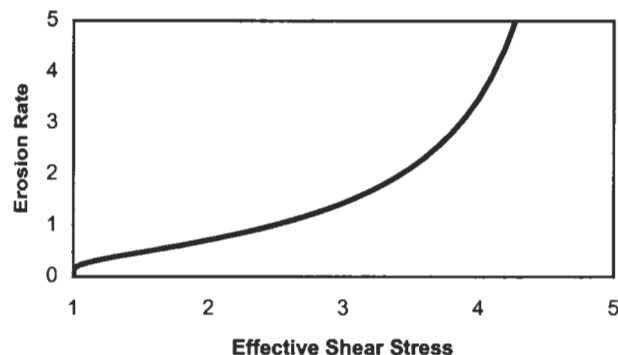


Figure 2. Relationship between effective shear stress and long-term erosion rate in a model of combined weathering and shear stress detachment in headwater channels. For this diagram $C_f=5$, $C_i=1$, and $\lambda=1$.

then the erosion rate decreases with layer thickness for $k > 1$.

If $\tau_e < C_f$ no erosion occurs; if $\tau_e > C_f$ there is a minimum erosion rate of about $0.2\delta\lambda$, and as τ_e approaches (or exceeds) C_i the erosion rate becomes infinite (Figure 2). For the more interesting case of $C_i > \tau_e > C_f$ the erosion rate increases with τ_e , and nearly linearly so if $C_i \gg \tau_e > C_f$. Although this model is simplistic (for example, it does not account for gravitational stresses on the weathered material) and lacks supporting field or laboratory measurements, it shows how erosion in headwater bedrock channels can involve interaction between weathering and fluvial detachment.

The potential weathering rate of bedrock exposed in headwater channels may be greater or less than that for bedrock on adjacent slopes (either exposed or with a regolith cover). Flow in such channels is likely to be ephemeral, so that if the rock is susceptible to physical weathering by wetting and drying or freeze-thaw, weathering potential may be high [Stock *et al.*, 1996]. On the other hand, if the weathering on slopes is primarily chemical, the lack of a soil cover and the prevalence of exfiltrating flow may restrict the weathering potential of bedrock exposed in channels.

Erosion by Debris Flows and Avalanches

In high-relief landscapes bedrock erosion in headwater channels may primarily occur by energetic mass movement. In mountainous areas of the Appalachian Mountains (Figure 3, Figure 4) and in the Pacific Coast Ranges, debris flows episodically flush accumulated colluvium from hollows. Some progress has been made to develop quantitative models of such slope failures [e.g., Montgomery and Dietrich, 1994; Dietrich *et al.*, 1995; Benda and Dunne, 1997a]. On the other hand, little attention has been directed towards the role of debris flows eroding the underlying

bedrock. In intervals between debris flows colluvium refills the hollows. In most cases the unweathered bedrock is sufficiently massive that debris flows would be ineffective in deepening hollows and low-order channels without accompanying weathering, suggesting that weathering and scour by debris flows interact much as scour and weathering in the channels discussed above. The relative roles of weathering by physical processes when bedrock is exposed following debris flows versus chemical weathering beneath colluvium is uncertain, and may vary in different rock types and climates. Some evidence suggests a different erosion rate law characterizes debris-flow dominated headwater bedrock channels than downstream fluvial channels, because there is a kink in the area-gradient relationship such that debris flow channels are less concave than fluvial channels [Seidl and Dietrich, 1992; Montgomery and Foufoula-Georgiou, 1993].

On mountainous slopes debris avalanches and rockfalls may also be a concentrative process, eroding steep bedrock chutes on headwall slopes. Examples include arctic and alpine mountain slopes (Figure 5) [Matthes, 1938; Blackwelder, 1942; Rapp, 1960a,b; Akerman, 1984; Rudberg, 1986; Luckman, 1977, 1978], canyon walls on Mars (Figure 6) [Sharp and Malin, 1975; Blasius *et al.*, 1977; Lucchitta, 1978], and the pali landscapes of tropical mountains (Figure 7). The main distinction between these and the mountain slopes discussed above is that energetic mass movement occurs over the whole landscape and not just the hollows. In arctic and alpine terrain dry rockfalls, debris avalanches, and snow avalanches appear to be capable of rock erosion [Matthes, 1938; Blackwelder, 1942; Rapp, 1960a,b; Peev, 1966; Gardiner, 1970, 1983; Luckman, 1977, 1978; Hewitt, 1972; Corner, 1980; O'Loughlin and Pearce, 1982; Ackroyd, 1987].

Again, weathering and erosion by rapid mass wasting probably interact to erode headwall chutes. The upper slopes of such landscapes are organized into steep, primitive basins ('spur-and-gully' topography) with divides at the scarp crest and along the crests of spurs extending down the slope (Figures 5 and 6). Topographic profiles from the divides at scarp or spur crests are concave, with the upper portions being very steep (45-90°) and bedrock floored, giving way abruptly downslope to talus at the angle of repose (30-45°, depending upon talus angularity and grain size range). Two classes of models might explain the development of these basin forms. One possibility is that stress-strain-failure relationships in near surface rocks coupled with topography, spatially variable rock resistance or fracture patterns, and surface-directed weathering processes might develop spur and gully forms independent of direct involvement of mass-wasting processes. Although not specifically applied to spur-and-gully forms, a number of investigators have proffered such arguments for development of crenellated forms of alpine relief [Whalley, 1984, and references

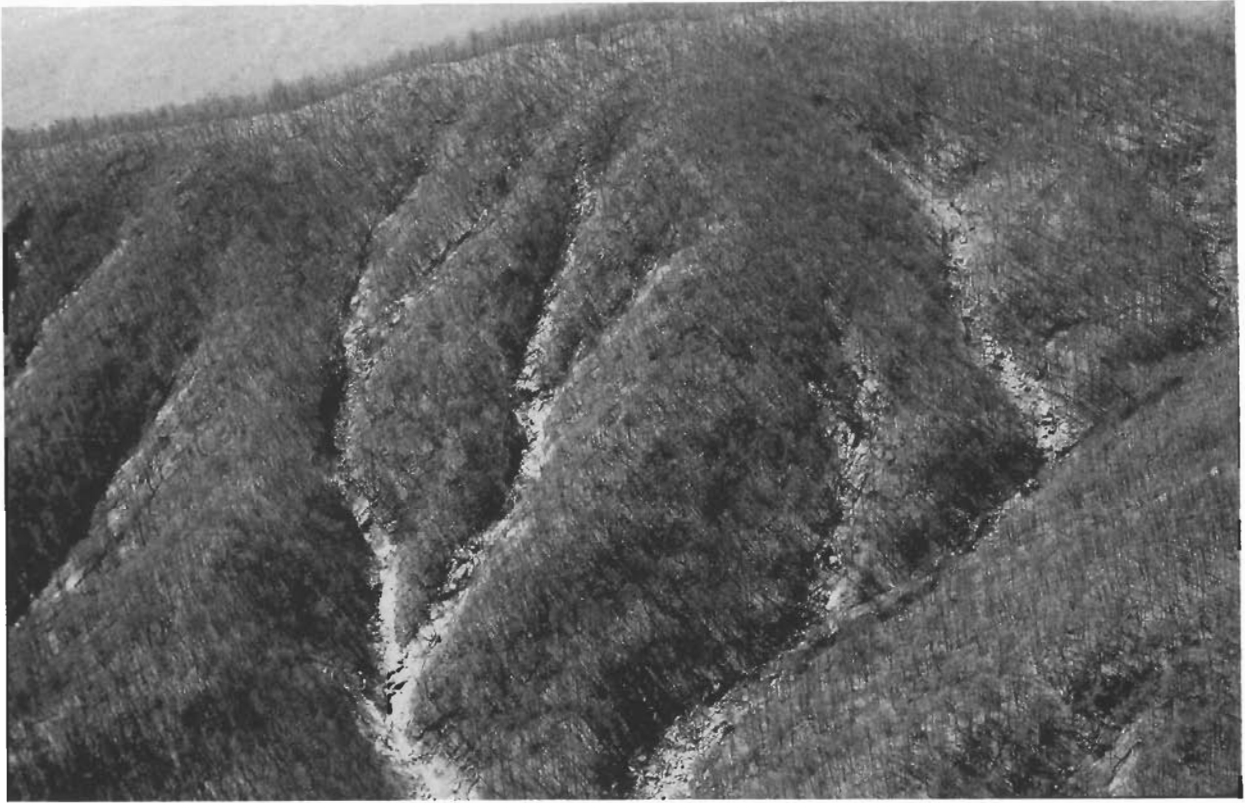


Figure 3. Avalanche scars in hollows on Kirtley Mountain, Madison County, Virginia resulting from more than 600 mm of rain in 8 hours in June, 1995. As opposed to the landscapes of Figures 5 and 6, debris avalanche scour is primarily limited to the hollows and low order channels.

therein]. The above-cited studies suggest, however, that the rockfall and avalanche processes are erosive, such that the basins develop due to economy of scale in the erosive processes similar to that responsible for creation of fluvial drainage basins, although structural influences complicate the resulting pattern.

Howard [1990] modeled the development of a mountain slope in profile from a combination of weathering and erosion and deposition by rapid mass movement. In this model, the basic driving process is assumed to be physical weathering (e.g., frost wedging, progressive failure, etc.) extending inwards from the rock surface. Rock shearing is modeled by Coulomb failure with a linear relationship between maximum shearing strength, τ_f , and normal stress, σ_f , on the failure plane:

$$\tau_f = c + \sigma_f \tan \Phi, \quad (5)$$

where c is cohesion and Φ is the angle of internal friction.

Weathering slowly reduces cohesion through a characteristic thickness, d , of the exposed rock while Φ remains constant:

$$c = c_0 e^{-\alpha(t-t_0)}, \quad (6)$$

where c_0 is the initial cohesion at time t_0 and α is a characteristic rate of weathering. This temporal change in cohesion differs from (1) in that the cohesion eventually drops to zero. The cohesion in this case is envisioned to be due to coherent bedrock between fractures, the extent of which diminishes as physical weathering extends and connects fractures. Individual failures are assumed thin ($d \ll H$, the overall relief) so that a potential failure plane parallel to an infinite slope can be assumed. Under these conditions failure occurs if:

$$1 \geq \frac{c + \rho g d \cos \theta \tan \Phi}{\rho g d \sin \theta + \tau_s}, \quad (7)$$



Figure 4. Debris avalanche scar in Nelson County, Virginia resulting from more than 600 mm. of precipitation in a few hours in August, 1969. Most of the colluvial cover was stripped, exposing bedrock that had undergone varying degrees of saprolitic weathering.



Figure 5. Steep bedrock slopes in the Alaska Range, Alaska eroded into spur-and-gully forms by avalanching from exposed bedrock and scree accumulation at the slope base. Note the crudely dendritic avalanche chutes eroded into the bedrock exposures. The initial steep relief was created by glacial erosion of the slope base.

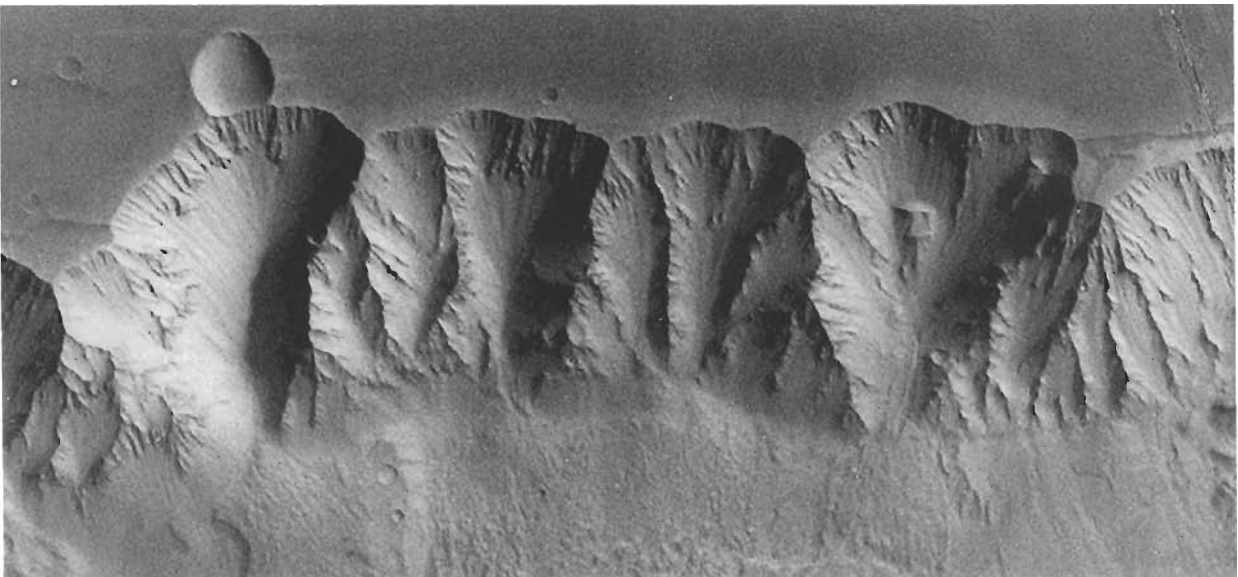


Figure 6. Spur and gully landforms dissecting the north wall of Ophir Chasma, Mars. The image is approximately 60 km across. The drop from the flat upland at the top of the picture to the base of the scarp is approximately 1.6 km. The bottom of the chasma is partially mantled by debris from large landslides, which have helped to create the large alcoves. Subsequent to the landslides, weathering and mass wasting have created the spur and gully terrain (part of Viking image 911A12).

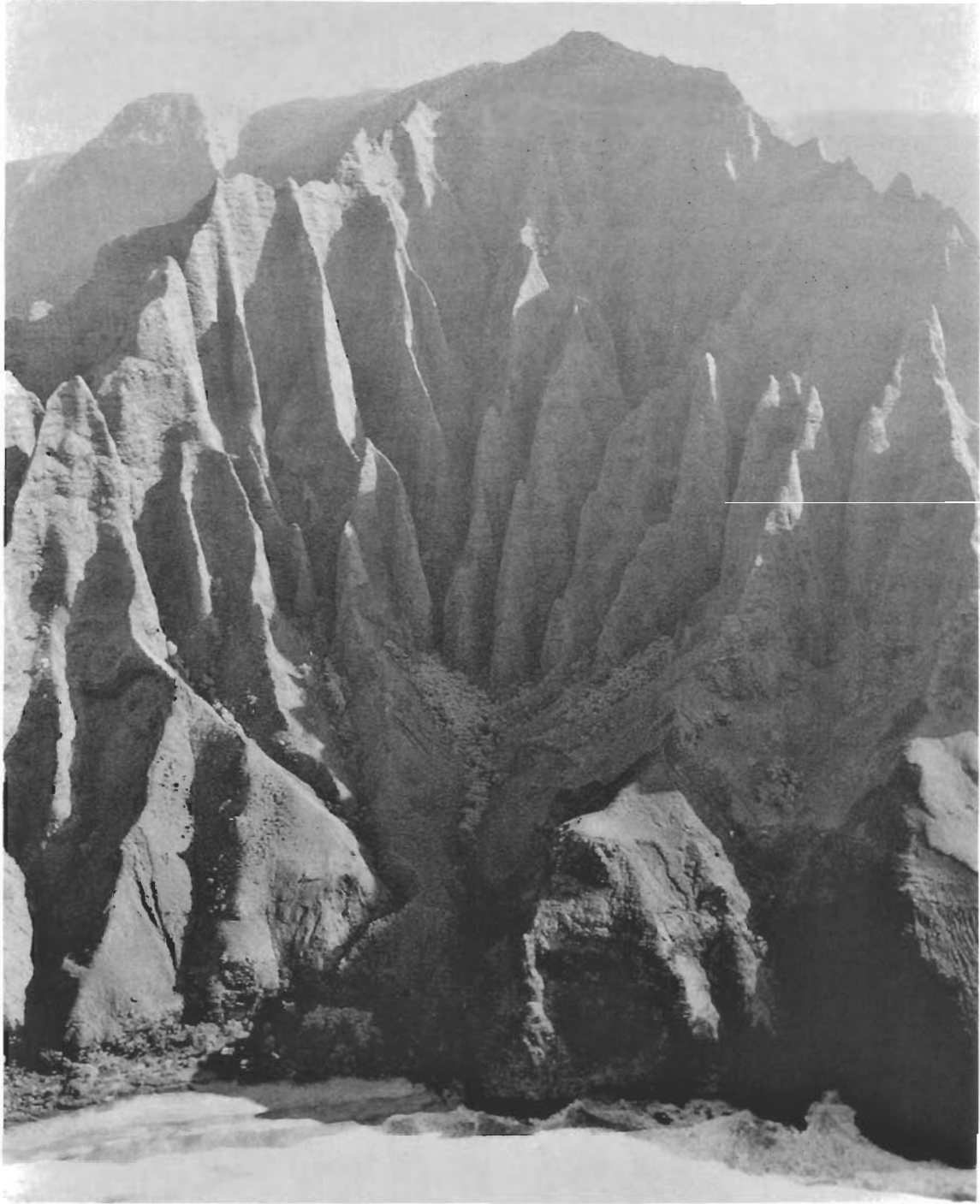


Figure 7. Steep landscape along the Napali Coast, Kauai Island, Hawaii. Note the nearly vertical slopes with knife-edged divides and steep chutes.

where θ is the local slope gradient and τ_s is a surface shear exerted by mass-wasting debris shed from higher on the slope. If a section of scarp becomes unstable due to decreasing cohesion, c , the debris shed from the slope is

routed downslope. Models of snow and rock avalanche motion [Perla *et al.*, 1980; Dent and Lang, 1980, 1983; Martinelli *et al.*, 1980; Pariseau, 1980; Lang and Dent, 1982; McClung and Schaerer, 1983; Schiewiller and

Hunter, 1983; Cannon and Savage, 1988; McEwen and Malin, 1990] suggest the following form for flow resistance:

$$\tau_r = \rho g d \cos \theta \tan \mu + \rho C_l V^2, \quad (8)$$

where μ is a coefficient of sliding friction, V is mean velocity, and C_l is a coefficient of "turbulent" friction. A theoretical basis for C_l is not firmly established and may represent air drag, internal frictional dissipation, and "plowing" of surface material [Perla et al., 1980]. Some models [Perla et al., 1980; Martinelli et al., 1980; McEwen and Malin, 1990] suggest an additional "laminar" frictional term proportional to velocity. Empirical estimates in snow avalanching suggest $\tan \mu \approx 0.27$ and $\xi \equiv g/C_l \approx 1500 \text{ m/s}^2$ [Perla et al., 1980; Dent and Lang, 1980; Martinelli, 1980; Buser and Frutiger, 1980; McClung and Schaerer, 1983]. In rock avalanches, air drag at the avalanche surface is generally unimportant so that shear at the avalanche-bedrock interface, τ_s , is equated with flow resistance, τ_r . Change in flow momentum equals the difference between downslope gravitational force and flow resistance, such that:

$$\frac{\partial(\rho d V)}{\partial t} = \rho g d \sin \theta - \rho g d \cos \theta \tan \mu - \rho C_l V^2. \quad (9)$$

For simplicity, the avalanche thickness is assumed to equal that of the failed layer and both avalanche depth and density are assumed to remain constant during motion. Because

$$\frac{\partial V}{\partial t} = \frac{\partial V}{\partial s} \frac{ds}{dt} = V \frac{\partial V}{\partial s} = \frac{1}{2} \frac{\partial V^2}{\partial s}, \quad (10)$$

where s is distance along the flow path, then

$$\frac{\partial V^2}{\partial s} = 2g \left(\sin \theta - \tan \mu \cos \theta - \frac{V^2}{\xi d} \right). \quad (11)$$

Eroded material is deposited where V decreases to zero, generally on the talus slope.

The above assumptions are incorporated into a profile (2-D) finite-difference simulation model. Initial conditions (Figure 8) are a mountain front, or scarp, of height $H=5000$ m extending above a flat valley floor with constant initial gradient $\theta_0=70^\circ$, and randomly assigned values of c_0 (values are assumed to be lognormally distributed with a specified mean and variance). This scaling was selected to model spur-and-gully development on the 2-10 km high structural scarps of Valles Marineris on Mars (Figure 6). The values of

c_0 are chosen to assure initial slope stability (values of simulation parameters are given in the figure caption for Figure 8). A vertical, rather than the normal horizontal grid (100 m increments), is utilized because of the steep gradients, so that the rate of horizontal retreat of the slope is modeled. Erosion is directed perpendicular to the surface, so that the horizontal erosion rate, $\partial x/\partial t$, equals the normally-directed erosion rate, $\partial n/\partial t$, divided by the sine of the slope angle. The simulation model progressed through iterations, with weathering gradually reducing the strength of the surface layer as indicated above. Once an avalanche occurs on a given segment of slope, flow of that plug of material is routed downslope and the factor of safety is determined for each slope segment traversed by the flow. In general, the value of τ_s is sufficiently large that a number of downslope segments also fail. Debris from the additional failed segments is also routed downslope (for simplicity it is assumed that each plug moves independently, although observations suggest an almost simultaneous movement of all portions of the slope involved in an individual avalanche). Once the surface skin of thickness d is removed from the slope, weathering of the underlying layer begins, with a value for c_0 assigned randomly as discussed above. In the simulations, provision is made for lack of flow contact and weathering exposure in overhangs. Furthermore, only the component of flow momentum in the direction of the continuing flow is assumed to be preserved when the avalanche changes direction, e.g.,

$$V_d^2 = V_u^2 \cos^2 \psi, \quad (12)$$

where V_d is the flow velocity below a bend of angle ψ , and V_u is the flow velocity entering the bend.

One simulation was run in which there was no instability due to surface stress ($\tau_s=0$). In this case (Figure 8A), the valley wall retreats by parallel motion while the foot of the slope is covered by a growing talus mound, and the bedrock-talus contact develops the convex upward profile that is characteristic of scarps undergoing cliff retreat and talus accumulation [Lehmann, 1933; Bakker and Le Heux, 1952; Scheidegger, 1991, pp. 130-134]. The slightly irregular profile of the retreating upper slope is due to the random assignment of values of initial cohesion, c_0 . Under these conditions, there is no enhanced downslope erosion and spur-and-gully topography would be unlikely to form in the 3-D case.

In a second simulation (Figure 8B) instability due to surface stress by moving debris was permitted, with $\tau_s=\tau_r$. The important feature of this case is the valley wall steepening through time due to greater chances for instability at the base of the slope due to weight of the avalanche block plus the V^2 dependence of τ_s . That is, for a

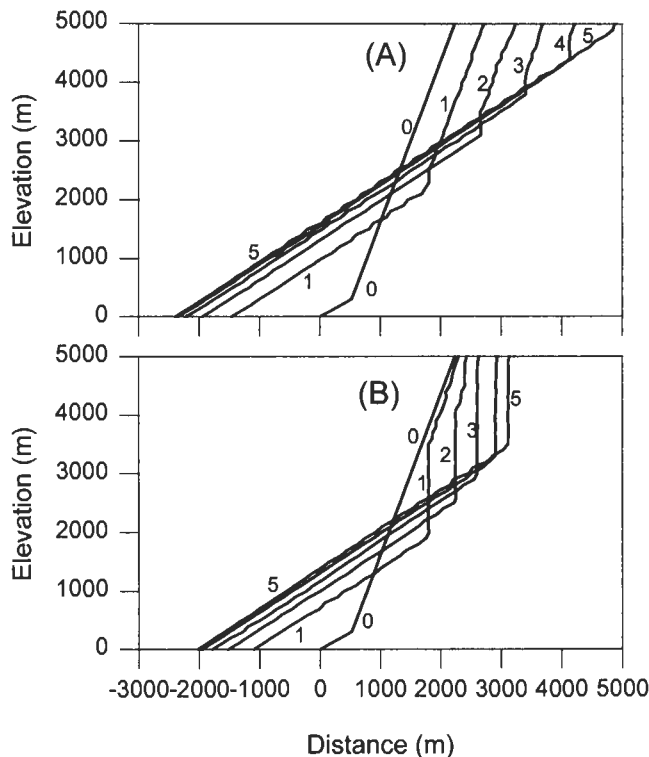


Figure 8. Profile models of slope erosion by the combined effects of weathering and avalanche scour with scree accumulation at the slope base. Numbers indicate successive profiles. (a) Erosion by weathering only. (b) Combined erosion by weathering and avalanche detachment. Note that the bedrock profile steepens. Parameters for the simulation are $d=10$ m, $g=3.9$ m s⁻² (appropriate for Mars), $c_0=6.125 \times 10^5$ kg m⁻¹ s⁻², and $\rho=2500$ kg m⁻³. For rock $\Phi=45^\circ$ and for talus $\Phi=35^\circ$. For avalanche motion on rock $\xi=500$ m² s⁻¹ and $\mu=15^\circ$, and on talus $\xi=10$ m² s⁻¹ and $\mu=30^\circ$. The weathering time units are arbitrary, with α in (1) being 0.03 with time measured in iterations, and the simulation is continued over 500,000 iterations.

constant gradient, the potential rate of bedrock erosion increases with distance downslope. This is an advective process having an economy of scale that is analogous to development of channels by fluvial erosion (e.g., Howard [1994a]), and can result in development and deepening of flow chutes. In three dimensions, the scale economy would be magnified by the potential for avalanche flow convergence into established chutes. Full simulation of chute development would require an areal model with a surface-conforming grid to permit treatment of vertical or overhanging slopes.

The pinnacled slopes of the Hawaiian Napali Coast (Figure 7) are probably eroded primarily by debris avalanches. Wentworth [1943] and White [1949] describe

the 'tipping bucket' cycle of weathering and rapid mass wasting that characterize steep mountain slopes on Hawaii. The rapid basal erosion by streams and the greater mobility of debris avalanches in the wet environment preclude much scree accumulation at the foot of the slopes.

In steep landscapes with vegetated, regolith-covered slopes, episodic landslides in hollows trigger wet debris flows that travel through the headwater tips of the channel network [e.g., Dietrich and Dunne, 1978; Montgomery and Dietrich, 1994; Dietrich et al., 1995; Benda and Dunne, 1997a]. The bedrock flooring these debris flow channels is eroded by a combination of weathering and debris flow detachment in a manner similar to the mountain slopes described above [Seidl and Dietrich, 1992; Montgomery and Fofoula-Georgiou, 1993]. The wet debris flows that occur in such channels have a wide range of composition and water content, so that no single model of rheology and motion can cover all cases. A variety of rheology models have been utilized, including Coulomb friction (the first term in (8)), empirical velocity-dependent friction (the second term in (8)), grain flow mechanics [Takahashi, 1991; Savage and Hutter, 1991], Bingham fluids [Johnson, 1970; Whipple, 1997], and non-linear fluids (e.g., Chen [1988]). Flow routing methods include Eulerian center-of-mass routing (as in the avalanche model presented above), kinematic wave routing [Weir, 1982; Hunt, 1994; Huang and Garcia, 1997], and Lagrangian routing [e.g., Iverson, 1997b; Hungr, 1995; Rickenmann and Koch, 1997]. A promising approach is outlined by Iverson [1997a,b], which combines Lagrangian routing with depth-averaged equations, a Coulomb rheology incorporating effects of pore water pressure on reducing effective normal stresses on the bed, and Rankine earth pressure theory (active stresses in extending parts of the flow and passive stresses in the compressional region at the front of the flow).

Headwater Channels: Conclusions

Modeling of erosion by debris flows in headwater channels requires rate laws for bedrock weathering and debris flow detachment in addition to flow routing models. Such erosion models are, unfortunately, only in a speculative state of development, as illustrated in the previous discussion. Systematic field observations will be required to elucidate erosion mechanisms and rates.

In summary, the rate of erosion of headwater bedrock channels is controlled by interplay between scour by water and debris flows, infilling by colluvium, and weathering processes. Because erosion by runoff or debris flow/avalanche strips partially weathered bedrock from slopes before they would fail by a combination of weathering and gravity alone, these energetic flows are concentrative erosional agents that create hollows and low-order channels.

DOWNSTREAM BEDROCK CHANNELS

The discussion in this section focuses on larger bedrock and mixed bedrock-alluvial channels in which weathering processes and scour by debris flows are quantitatively subordinate to fluvial erosion. Several important issues are discussed: 1) What factors determine whether channels are bedrock, wholly alluvial, or a mixture; 2) What processes are responsible for erosion of bedrock channel beds; and 3) Can erosion rates in bedrock channels be quantitatively modeled?

The Distribution of Bedrock Channels

Bedrock channels, which lack an appreciable cover of alluvium, occur when stream flow has excess transporting capacity, compared to supply rate, for all size ranges supplied from upstream and from local slope erosion. Channel incision into bedrock occurs when the supply of sediment to the channel cannot keep it continuously mantled with an alluvial cover, usually due either to steep gradients or to meager sediment supply. Thus, bedrock channels are favored by one or more of the following factors: high relative relief, high uplift rates and steep slopes, rapid local upwarping or faulting, resistant bedrock, and low sediment yields. Because of scouring and plucking that occur during high flow stages, channels with a thin alluvial cover can erode the underlying bedrock while maintaining an alluvial cover during low flow conditions [Howard and Kerby, 1983]. The bedrock erosional capacity of alluvial channels is limited, so that if downstream erosion rates exceed this capacity, local gradients steepen and bedrock becomes exposed [Merritts and Vincent, 1989]. This may occur in particularly resistant rock, as a result of differential uplift along a river profile, or as a result of relative land-sea elevation changes (such as the Fall Line in the Appalachian Mid-Atlantic region [Reed, 1981; Hack, 1982]).

Despite these general tendencies, attempts at *a-priori* prediction of the nature of the channel bed (bedrock, gravel, sand) based solely upon basin relief and channel gradient are likely to be erroneous, as is illustrated by the following simple analysis. Consider a reach in a river system that is subject to a constant rate of base level lowering at its lower end. If the channel is wholly alluvial, the rate of bed lowering is governed by the divergence of sediment transport:

$$\frac{\partial z}{\partial t} = -\frac{\partial q_s}{\partial x}, \quad (13)$$

where q_s is the volumetric rate of sediment transport and x is the downstream direction. Figure 9 shows how the gradient of an alluvial channel changes as the rate of lowering of the lower end of the reach is varied (a typical bedload transport

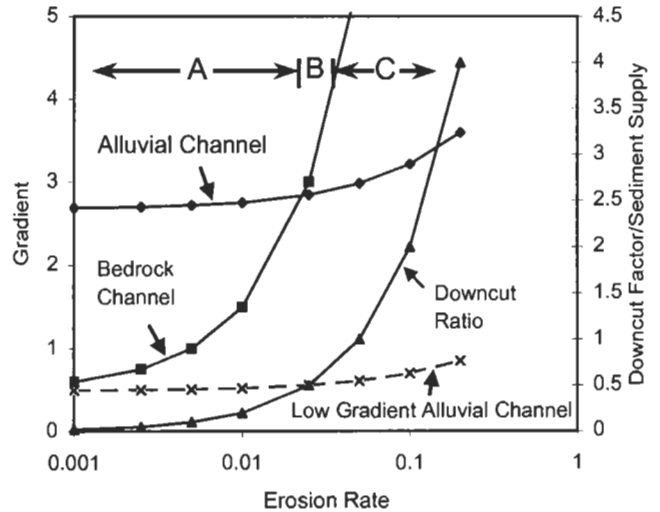


Figure 9. Relationship between channel gradient and erosion rate for bedrock and alluvial channels. The "Downtcut Ratio" for an alluvial channel is defined as $(E L) / q_s$, where E is the erosion rate, L is the stream reach length, and q_s is the volumetric rate of bedload supply from upstream. In region 'A' the required alluvial channel gradient is much greater than for a bedrock channel, in region 'B' the gradients are commensurate, and in 'C' the required bedrock gradient is much greater than for an alluvial channel. The dashed curve shows alluvial channel gradients for a basin with a meager bedload supply; such a river system would be bedrock throughout. The gradient and erosion rate scales are arbitrary.

formula is assumed, see Howard [1994a, p. 2265-6]). The rate of supply of alluvium from upstream is taken to be fixed and independent of short-term or reach-length variations in main channel erosion rate. If the rate of bed lowering is close to zero (left side of Figure 9) the channel gradient is simply that required to transport sediment supplied from upstream. As the rate of erosion increases, the gradient must steepen to transport both sediment supplied from upstream and that from local bed lowering. Until the rate of erosion reaches very high values the bed steepening is very modest, which is why it is often assumed that channel gradients are in equilibrium with sediment supply from upstream and unaffected by erosion rate [Mackin, 1948]. On the other hand, if bedrock channel erosion rates depend upon shear stress or stream power (see analysis below), channel gradients must steepen appreciably to accommodate greater rates of downcutting (Figure 9 and (20)). For this analysis, the assumption is made that the gradient required for erosion of a bare bedrock channel at low rates of downcutting is less than that required to transport sediment supplied from upstream. As a result, the curves for required gradient for alluvial and bedrock channels should cross at a critical

erosion rate, such that for low rates of erosion gradient control by sediment transport should dominate and for high rates the bedrock erosion should be controlling. For low rates of erosion (Region A in Figure 9) the required gradient for bedrock erosion is much lower than that for sediment transport. Bedrock erosion during infrequent intervals of deep scour of the alluvial bed might suffice to keep pace with base level lowering, and the bed would be alluvial. For high rates of erosion (Region C in Figure 9), the steep required gradient would discourage deposition of bed sediment, even during waning flow stages, producing the commonly observed steep, "clean" bedrock channels. Only in a narrow range of erosion rates (Region B) would required gradients for bedrock erosion and alluvial transport be commensurate. For such reaches a partial alluvial mantling might be expected -- the mixed alluvial-bedrock channels discussed below. This simple analysis thus suggests that, within a basin of uniform sediment yield, steep channel gradients should be associated with bedrock channels, whereas low-gradient channels should favor alluvial beds.

These expectations for channel bed type are often invalid. The South Fork Eel River (Figure 10a) has a high-gradient canyon reach that is mantled with coarse boulders (Figure 11) whereas a low gradient reach upstream is largely exposed bedrock (Figure 12). The steep canyon reaches of the Colorado River expose bedrock only in deep scour holes and rapids are floored by boulders contributed by side-canyon debris flows [Howard and Dolan, 1981; Howard *et al.*, 1994; Grams and Schmidt, 1997]. Finally, a short, steep canyon reach of the Maury River in Virginia (Figures 10b and 13) is largely mantled by boulders. The common thread for these examples, considered more fully below, is the importance of locally contributed coarse debris.

The central Coastal Ranges of Oregon near Coos Bay have been subject to uplift rates of about 0.1 mm/yr, producing steep relief. Steep headwater tributaries are bedrock-floored. However, stream profiles are strongly concave, and the low-gradient downstream reaches of rivers such as the Smith River and the Umpqua River might be expected to be alluvial, whereas they generally remain bedrock-floored. This is probably because the ease of weathering and comminution of the Tye Sandstone bedrock produces little gravel bedload, and the sand is transported largely in suspension (although some caution is in order because lumbering earlier in the century may have removed large woody debris and sediment from channels due to 'splash damming' -- large short-lived floods produced by creation and intentional breaching of temporary dams). Thus the alluvial channel gradients required for the range of local erosion rates appears to be lower than that for bedrock erosion (e.g., the dashed line in Figure 9). Coastal ranges to the north and south of this region in more indurated bedrock support steeper gravel rivers. These examples illustrate the

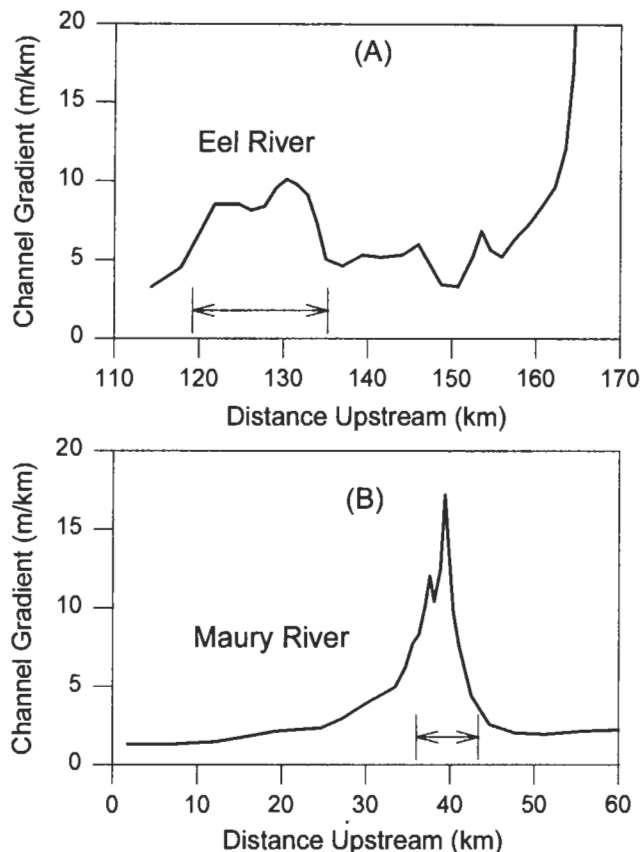


Figure 10. Channel gradient of (a) the South Fork Eel River, California and (b) the Maury River, Virginia. Distances measured from the mouth of the Eel River, but from an arbitrary datum for the Maury River. Segments marked with arrows are canyon sections.

need to consider the properties of sediment supplied from slope erosion, both regionally and locally, in addition to relief and uplift rate as determinants of channel gradient and bed type.

Finally, the above analysis suggests that few channels should exhibit beds transitional between full alluvial and bare bedrock. The discussion of mixed bedrock-alluvial channels below demonstrates that this expectation is also incorrect.

Quantifying Erosional Processes in Downstream Bedrock Channels

In streams with bedrock beds, the critical concern is the rate of bed erosion. Erosion may occur by several mechanisms, including hydraulic plucking [Miller, 1991; Wende, 1997; Whipple *et al.*, 1997; Dollenmayer and Whipple, 1997], cavitation [Barnes, 1956; Matthes, 1947],

abrasion by sediment [King, 1927; Alexander, 1932; Maxson and Campbell, 1935; Maxson, 1940; Foley, 1980; Sharpe and Shaw, 1989; Howard et al., 1994; Slingerland et al., 1997; Ellis et al., 1997; Whipple et al., 1997; Tinkler, 1997; Sklar and Dietrich, 1996, 1998], solution [King, 1927], and weathering [Stock et al., 1996]. General reviews are given by Baker [1978], Baker and Pickup [1987], Baker [1988], Tinkler and Wohl [1998], Dick et al., [1998], and Wohl [1998]. The relative importance of these processes depends upon rock type, channel hydraulics, water chemistry, sediment type and load, and climate. Thus there is no simple, universal law of bed erosion, and due to the general slowness of bed erosion in resistant rocks, few process observations have been made.

Modeling of erosion in bedrock channels to date has primarily been based upon the assumption that erosion rate $\partial z/\partial t$ is a function of some measure of flow intensity, \mathcal{G} :

$$\frac{\partial z}{\partial t} = -K_i (\mathcal{G} - \mathcal{G}_c)^\zeta, \quad (14)$$

where \mathcal{G}_c is a critical flow intensity that must be exceeded for erosion to occur, K_i is an erodibility that depends upon bedrock properties, and ζ is an exponent (most models have assumed that ζ is unity). The most common assumed measures of flow intensity are the bed shear stress, τ , and the stream power per unit area of channel bed, $\omega = \tau V$, where V is mean flow velocity. Simple equations of continuity, flow resistance, and downstream hydraulic geometry are usually also assumed [e.g. Howard, 1994a].

$$\tau = \gamma R S, \quad (15)$$

$$V = K_n R^{2/3} S^{1/2} / N, \quad (16)$$

$$Q = K_p R W V, \quad (17)$$

$$Q = K_a A^e, \quad (18)$$

$$W = K_w Q^b, \quad (19)$$

where γ is the unit weight of water, R is hydraulic radius, S is channel gradient, V is mean velocity, N is Manning's resistance coefficient, Q is an effective discharge, A is drainage area, and K_n , K_p , K_a , K_w are coefficients. The coefficients and exponents are generally assumed temporally and spatially invariant. These, when substituted into (14)

allow the erosion rate to be expressed as a function of drainage area and local gradient:

$$\frac{\partial z}{\partial t} = -K_i (K_z A^g S^h - \mathcal{G}_c)^\zeta, \quad (20)$$

where the various coefficients are incorporated into K_z . The exponents have the values $g=0.6e(1-b)$ and $h=0.7$ for $\mathcal{G}=\tau$, whereas $g=e(1-b)$ and $h=1.0$ for $\mathcal{G}=\omega$.

Several approaches can be used to estimate the values of the coefficients in (20) from field data. All of these assume $\mathcal{G}_c=0$, so that three parameters $K_z=K_i K_z^\zeta$, $m=b\zeta$, and $n=h\zeta$ must be estimated. The most direct approach is to collect data on net channel bed erosion over a known period of time for a range of values for contributing drainage area and channel gradient, ideally for a single bedrock type. Howard and Kerby [1983] estimated K_z , m , and n by regression analysis of 10 years of erosion in badland channels in Coastal Plain sediments in Virginia, finding $m=0.45$ and $n=0.7$, consistent with linear dependency of erosion rate upon shear stress. Similar methods have been used for larger channels in more indurated bedrock, with varying results. Seidl et al. [1994] found $m=n=1$ for channels incised into a volcanic shield on the island of Kauai. Stock and Montgomery [1998] analyze erosion data from several different rivers with known prior profiles and find a considerable variation in estimates of K_z , as might be expected for differences in rock type and climate. The estimated values of m and n also varied widely, however, with m ranging from 0 to 0.5 and n from 0 to 2. This might reflect variations in bedrock erosion processes amongst locations, but parameter estimation may also be compromised by 1) limited ranges of drainage area and gradient in the target streams, 2) downstream variations in lithology, 3) uncertainties in estimation of the initial profile, 4) presence of alluvial reaches along the stream profile (erosion rates will be lower where a protective alluvial cover is present), and 5) uncertain or irregular relationship between discharge and contributing area (18). Even where insufficient information on erosional history is available to estimate all parameters, it is sometimes possible to estimate the ratio of m to n . Seidl and Dietrich [1992] reasoned that tributaries and mainstem streams should be lowering at the same rate near their junctions, so that measurements of A and S in both streams allows calculation of m/n . Similarly, if geologic evidence suggests that a drainage basin has been undergoing a constant long-term erosion rate (that is, the topography is in steady state), then the ratio of m/n can be estimated by regressing channel gradient on drainage area:

$$S = -\left(\frac{\partial z}{\partial t} / K_r\right)^{1/n} A^{m/n}. \quad (21)$$

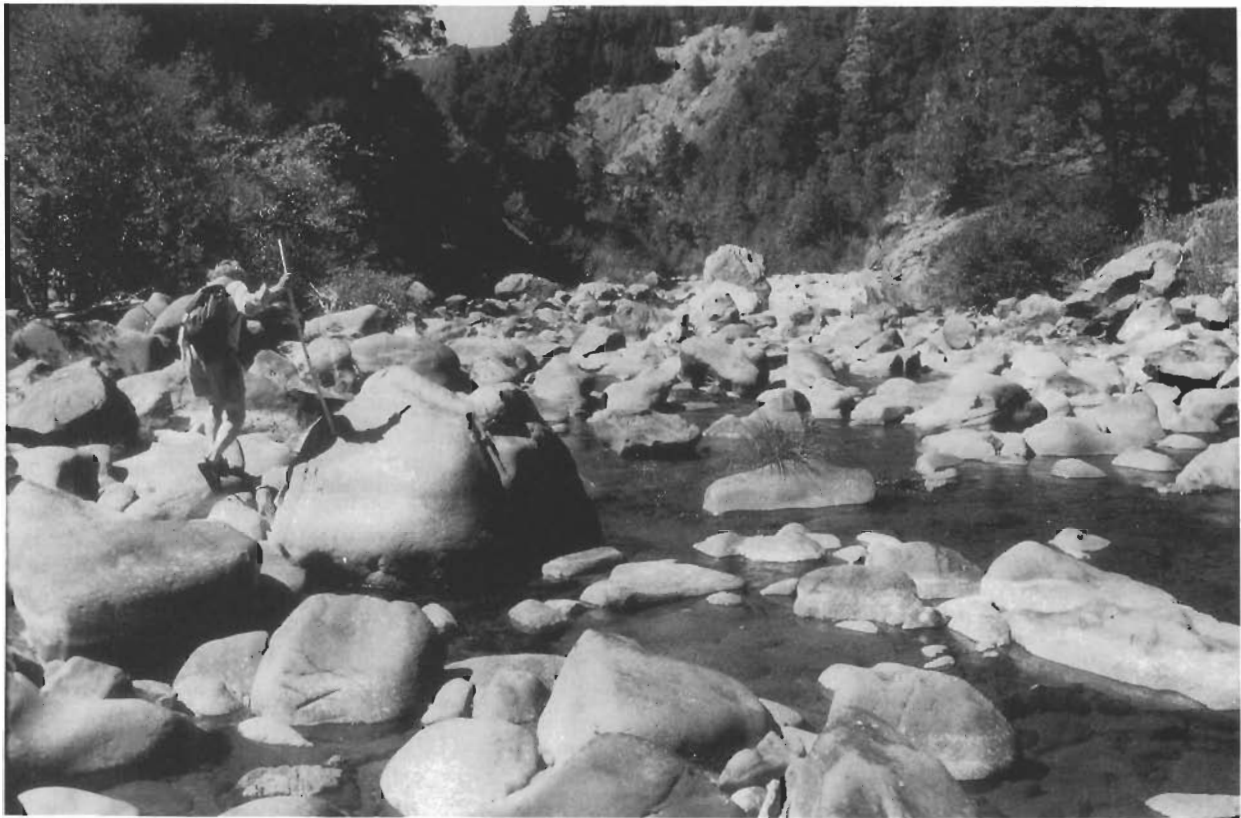


Figure 11. Channel bed of the South Fork Eel River at about km. 131 in Figure 10a. The bed is dominated by boulders derived by mass wasting from canyon walls, the channel bed, and local tributaries. Bedrock exposures occur locally in scour holes. The sandstone boulders are strongly rounded and fluted by suspended load abrasion except for the tops of the largest monoliths.

The assumption of a simple bedrock erosion rate law, such as (20), has been motivated by the desire to model long term landscape evolution in a variety of geologic and climatologic settings. It is uncertain at present how reasonable these assumptions will turn out to be as the study of bedrock channels progresses. It is probable that, even if such equations remain viable, no universal values of the exponents m and n will emerge because of a wide diversity of processes eroding bedrock channels.

The gradient of some bedrock channels may be determined by the threshold of detachment, \mathcal{Q}_c , in (20). In thin-bedded or well-fractured bedrock, hydraulic plucking may dominate bedrock erosion, such that there is a well defined flow intensity \mathcal{Q}_c for plucking. Erosion would progress rapidly until gradients dropped such that only the

largest floods could detach bedrock slabs. From (20) the gradient would be:

$$S \approx \left(\frac{\mathcal{Q}_c}{K_z A^k} \right)^{1/h}. \quad (22)$$

This situation would have a close analogy to threshold gravel bed channels [Howard, 1980; Howard *et al.*, 1994]. Operation of a bedrock channel system close to threshold conditions would also occur if the exponent ζ in (20) were greater than unity, because gradient would be only a slight function of erosion rate.

The temptation to use simplified models of bedrock erosion such as (14) and (20) is great given the paucity of



Figure 12. Channel bed of the South Fork Eel River at about km. 142 in Figure 10a. The bed is predominantly exposed sandstone, with thin gravel mantling in some of the low points.



Figure 13. The Maury River in the Goshen Pass (approximately km. 40 in Figure 10b). The bed is dominated by large blocks of Tuscarora Sandstone delivered by mass-wasting from adjacent canyon slopes. Bedrock ledges are exposed in the rapids in the upper left of the picture and locally elsewhere in deeper portions of the channel.

quantitative observations and the need for erosional rate laws in regional models of uplift, denudation, and sedimentation. The multiplicity of processes involved in bedrock channel erosion suggests caution. Some of this variability can be accounted for in appropriate choices of intrinsic bedrock erodibility (K , in (14)). The most glaring omission in (14) is the lack of explicit treatment of the role of sediment load in bed erosion. Several new models incorporate abrasion explicitly [Sklar *et al.*, 1996; Sklar and Dietrich, 1998; Slingerland *et al.*, 1997; Ellis *et al.*, 1997; Dick *et al.*, 1998]. If sediment load is low and sediment contributions are areally uniform, abrasion rate laws may converge to a form similar to (14) or (20).

Mixed Bedrock-Alluvial Channels

A surprising number of streambeds expose bedrock locally during normal low flows (say 5% to 60% of total bed area) while elsewhere the alluvial cover is no more than 2-3 meters thick – these are the mixed bedrock-alluvial channels

discussed here (simplified to mixed channels for this discussion) [Miller, 1991; Seidl and Dietrich, 1992; Wohl, 1992, 1993, Howard *et al.* 1994]. As pointed out by Brush [1961] and Howard *et al.* [1994], many of the streams in the Appalachian Mountain region could be so classified. Flume experiments of erosion of weak “bedrock” by through-flowing sediment commonly exhibit alternation of bedrock exposures in narrow sections and alluvium mantle in divergent flow [Wohl and Ikeda, 1997]. The simple but flawed analysis considered above (Figure 9) suggested that such mixed channels should be uncommon.

What, then explains the frequent occurrence of mixed channels? Howard *et al.*, [1994] propose two scenarios of temporal change that could result in (geologically) short-term coexistence of alluvial and bedrock channels. The first case occurs when alteration in sediment load and discharge occasioned by climatic or land use change causes the channel to undergo transition between bedrock and alluvial cover (in either direction). A mixed channel might persist for some time during the transition. The second case occurs

when sudden drop of baselevel causes dissection of a former alluvial channel system. Most of the subsequent erosion occurs by migration of a steep bedrock knickpoint. Channel sections well upstream from the knickpoint experience modest steepening and incision, however, as observed in experiments by *Gardner* [1983] and simulations by *Howard et al.*, [1994]. These sections upstream from the knickpoint might be mixed bed.

A reliance on evolutionary scenarios to explain the widespread occurrence of mixed channels seems *ad hoc*. The remainder of this discussion will focus on the possibility that such channels are either an equilibrium form or one adjusted to short-term oscillations in sediment supply. If mixed channels are temporally persistent as the river system downcuts, then the most crucial question is how bedrock erosion can occur when the bedrock is largely mantled. *Gilbert* [1880] suggested that the most important mechanism of bed erosion in bedrock channels is scour by sediment in transport. When the quantity of bedload is small, erosion rate should be proportional to the quantity of sediment in transport. But *Gilbert* pointed out that when the rate of sediment supply is large, grains interfere with each other and begin to mask the bed, so that the rate of erosion reaches a maximum and presumably goes to zero as the bed becomes 100% covered by alluvium. This inhibition of abrasion by large sediment load has been observed in studies of industrial slurry transport. Recent models of abrasional bedrock erosion by *Sklar et al.* [1996], *Sklar and Dietrich* [1998] and *Slingerland et al.* [1997] highlight the non-linear relationship between abrasion rates and quantity of sediment in transport. Not only is there reduced abrasion during high transport rates, but a rapid and sudden transition from exposed bedrock to nearly complete alluvial cover is favored by higher frictional dissipation in grain-to-grain collisions on the bed than in grain-to-bedrock collisions [*Howard*, 1980].

Several circumstances can explain the widespread occurrence and temporal persistence of mixed channels. One case is where the bedrock exposures are particularly resistant requiring the development of local rapids or falls for erosion to keep pace with the overall rate of stream lowering. The short knickpoints described by *Miller* [1991] in sedimentary rock may be an example, where development of knickpoints permits quarrying or undermining of resistant beds. However, irregular alternating bedrock and alluvial sections are often found even when the bedrock is massive and apparently homogeneous [*Wohl*, 1992, 1993].

A second possible mechanism permitting or requiring mixed channels is bedrock erosion primarily through migration of local knickpoints or waterfalls separating alluvial reaches. Because an alluvial cover inhibits bedrock corrosion and weathering, erosion of bedrock might only occur in steep sections where high flow velocities maintain a largely sediment-free bed. In order to maintain a stable, migrating knickpoint, the potential erosion rate of the bed

for a given channel gradient must be greater at the base of the knickpoint than at the crest. Otherwise, the knickpoint will gradually disappear by diffusional flattening, as is implicit in bedrock erosional models such as (14). Two mechanisms can produce concentrated basal attack. One is exposure of weak units or bedding planes at the base of the knickpoint, permitting undermining, as in the streams in sedimentary sequences described by *Miller* [1991] and the famous Niagara Falls. The other is development of locally supercritical flow over the crest of the knickpoint, which accelerates bedload particles and induces high turbulence at a hydraulic jump at the base of the knickpoint. The presumably stable and migrating knickpoints described by *Seidl and Dietrich* [1992], *Wohl* [1992, 1993], and *Dick et al.* [1997] may be examples. The height of knickpoints can be determined by bed thickness in sedimentary rocks. Knickpoints developed by rapid base level lowering of a master stream are influenced by the depth of downcutting. In addition, knickpoints might be a stable feature of a relatively constant rate of downcutting in mixed channel systems, even in homogeneous bedrock. The height of such knickpoints might be conditioned by the spatial scale required by the flow to develop a supercritical transition-hydraulic jump pair or, possibly, an integrated vortex system. Characteristic of knickpoints in many cases is a downstream transition from broad longitudinal grooves on a nearly flat exposed bed to incised inner channel [*Wohl*, 1992, 1993]. If hydraulic controls determine the height, ΔZ , of knickpoints, then, if a master stream is eroding at a rate $E = -\partial z/\partial t$, a knickpoint on a tributary will form after a time $\Delta T = E\Delta Z$. Assuming that knickpoints migrate upstream with a constant velocity, $V_k = \partial x/\partial t$, the linear density of knickpoints on the tributary, $n/\Delta X$, will equal $E/(V_k \Delta Z)$. For sufficiently high rates of base level lowering, all alluvial cover will be stripped, and a totally bedrock reach will occur.

The remaining scenarios for mixed channels involve temporal alternation of exposed and mantled bedrock. Migrating bedforms such as dunes, bars, and sediment waves may provide local exposure of bedrock that permits continuing erosion. Spatial consistency of bed erosion rates would be enforced by gradual lagging of bedrock scour and eventual bedrock exposure in bed areas that would otherwise favor relatively permanent alluvial cover. In headwater channels in forested watersheds large woody debris serves to trap sediment [*Keller et al.*, 1995; *Montgomery et al.*, 1995; *Abbe and Montgomery*, 1996; *Montgomery and Buffington*, 1997], and in some cases creates short, temporary alluvial reaches in otherwise bedrock channel [*Montgomery et al.*, 1996].

The final explanation for mixed channels involves episodic sediment delivery to the channel system. In many high-relief areas, sediment is contributed primarily by debris flows during intense precipitation events whose recurrence interval is multi-decadal to millennial [e.g., *Williams and*

Guy, 1973; Dietrich and Dunne, 1978; Dietrich et al., 1982; Benda, 1990; Benda and Dunne, 1997a]. The interaction of sediment supply and bedrock erosion in this case will be illustrated by a simple spreadsheet-based conceptual model in which all sediment delivered to the valley bottom occurs by infrequent debris flows associated with large floods. Relative to the overall simulation timescale these delivery events are essentially instantaneous. In between events, this sediment is reworked by moderate floods, causing bed abrasion. The moderate floods are represented as a continuum process, implying that they are very frequent compared to intervals between the debris flow delivery events. For the purposes of discussion only, the dependency of bedrock corrosion by these moderate floods upon the thickness of sediment cover will be modeled by the relationship:

$$\frac{\partial z}{\partial t} = -K_e T_s \left[1 - e^{-\eta(T_c - T_s)} \right], \quad (23)$$

where the coefficient K_e depends upon bedrock and sediment mechanical properties as well as flow intensity, T_s is the average thickness of sediment cover over the bedrock, T_c is a critical alluvial thickness beyond which $\partial z/\partial t = 0$, and η is sufficiently large that the exponential term becomes important only when T_s approaches T_c . This is a humped relationship going to zero at both T_c and T_s . The sediment thickness, T_s , is averaged over the whole bed (or valley bottom), including exposed bedrock and bar deposits. The model implies that the bed sediment load during moderate floods is derived from the valley bottom deposits and is therefore an increasing function of T_s . Sediment delivery by debris flows occurs during rare, high-intensity events (Figure 14a). Between these rare storms are the numerous moderate floods (not explicitly shown in Figure 14a). These moderate floods gradually entrain, transport, and comminute the sediment introduced by the major floods, so that the average sediment cover diminishes exponentially between high-intensity storms (Figure 14b). This stochastic modeling of sediment supply, gradual removal, and episodic bed exposure is similar to Benda and Dunne [1997b]. Channel aggradation following large storms and gradual removal by smaller storms has been noted in many streams [e.g., Benda, 1990; Madej and Ozaki, 1996]. The rate of bedrock scour by moderate floods is assumed to follow (23), with $T_c=0.2$ (all units in Figure 14 are arbitrary). Consequently, the rate of bedrock erosion is limited by both too great a sediment mantling ($T_s \geq T_c$) and too little sediment supply ($T_s=0$) (Figure 14c). In addition, the major sediment-producing floods are assumed to have high scour potential, at least up to the point that they bury the bed (T_c is assumed to be 0.4 for these large floods, reflecting greater scour potential of major floods). Note that the last major flood

occurs so soon after the previous flood that the bed is still sediment-covered, resulting in no additional bed erosion.

This last scenario differs from the climatic-change explanation for mixed channels in that no systematic environmental change is envisioned. All of the "steady-state" explanations for mixed channels suggest that bedrock is episodically exposed. This permits bedrock erosion despite a sediment thickness that, if spread over the bed and averaged through time, might be thick enough to prevent bed erosion. The episodic exposure can result from exposure in the troughs of migrating bedforms, in migrating knickpoints, or as a result of episodic addition and removal of sediment from the channel bed and valley floor (Figure 14). Channels with such episodic exposure are operating in region (B) as well as the portion of region (A) close to region (B) in Figure 9.

Gradient Control in Mixed Channels

For long-term erosional modeling it is important to be able to predict the relationship between channel gradient and erosion rate. In alluvial channels the use of equation (13) plus a bedload transport formula will allow prediction of channel gradient if the size distribution and rate of sediment influx can be estimated. For pure bedrock channels a process approach such as (14) may be appropriate for predicting long profile evolution. However, in mixed channels it is not clear whether the gradient is determined primarily by divergence of sediment transport or by the necessity for bed erosion. The actual gradient may even be greater than for a pure bedrock or alluvial channel because of the necessity to accomplish both bed scour and transport [Howard and Kerby, 1983]. The issue is further complicated by the feedback between rate of incision and the quantity and size of sediment supplied.

Kodama and Nakamura [1996] discuss one such feedback. In a canyon section of the Ojika River, local additions of nearly immobile coarse boulders from tributaries is associated with steepening of the mainstem gradient. Flume experiments [Kodama and Nakamura 1996] show that steeper gradients are required to transport the same amount of bedload through boulder-strewn reaches than through boulder-free reaches. If bedrock erosion were primarily related to the quantity of bedload in transport, steeper gradients would be required in bouldery reaches to assure continuity of sediment transport and equality in bed erosion rates as in boulder-free reaches. The presence of boulders does not necessarily reduce bed erosion rates, however, because obstructions on the bed increase turbulent intensity and may create systematic vorticity that can locally enhance bed scour by suspended load [Sharpe and Shaw, 1989; Tinkler, 1997].

When rates of bedrock river incision are high, negative feedback in the form of increased sediment delivery can

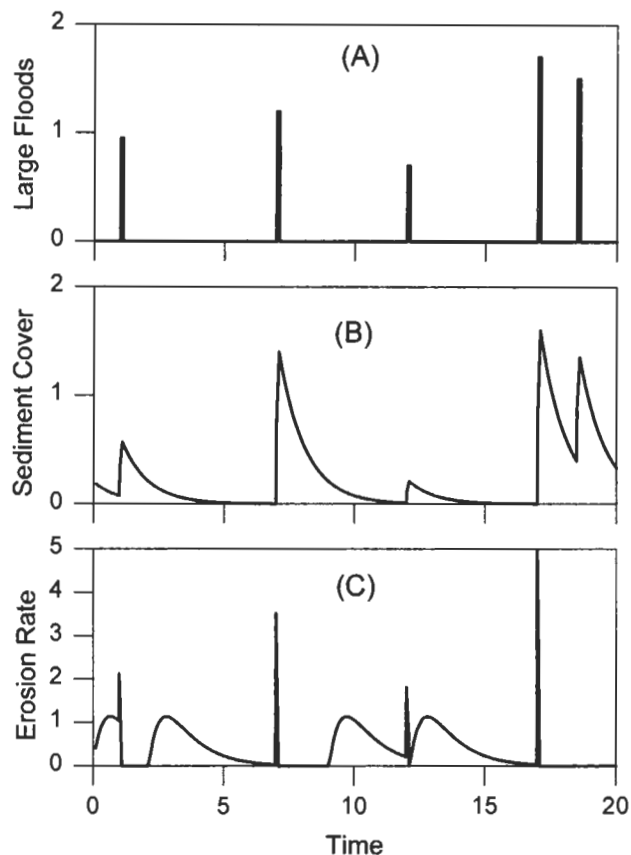


Figure 14. Conceptual model of episodic sediment supply, sediment cover, and erosion of bedrock channel bed. Scales are arbitrary. See text for explanation.

limit the rate of incision and even control the channel gradient. Although long-continued rapid incision results in generation of steep relief throughout the drainage basin, and thus high sediment loads, the immediate effect of accelerated incision is to cause canyon reaches with steep slopes and steep tributaries. This in turn triggers delivery of coarse boulders that may partially or wholly mantle the channel bed. This *locally contributed* debris often controls gradient and more severely limits incision rates than *through-flowing* alluvium derived from the entire drainage basin [Howard and Dolan, 1981; Seidl and Dietrich, 1992]. The Ojika River, cited above, appears to be an example. A classic example is the Colorado River in the Grand Canyon. As discussed by Howard and Dolan [1981], Kieffer [1985, 1990], Webb *et al.* [1989], and Howard *et al.* [1994], debris flows from canyon walls delivered through short, steep tributaries create fan-like deposits containing large boulders that are reworked into the steep rapids that account for most of the elevation fall through the Grand Canyon. A similar

control of channel gradient by debris-flow fans occurs in other canyon reaches of the Colorado River and its tributaries [Grams and Schmidt, 1997]. The continued production of coarse debris from canyon walls has been sufficient to balance the weathering and comminution of boulders in the debris fans such that the Colorado River within the canyon has downcut very little during the Pleistocene, at least in the western portion [Lucchitta, 1990]. The overall gradient of the river is controlled by the long-term balance between rate of debris flow delivery and rate of reworking of the debris fans by the Colorado River.

Another example is the canyon reach of the Maury River at Goshen Pass, Virginia, where the river cuts through the resistant Tuscarora sandstone. Above and well below the canyon reach the alluvial bed is coarse gravel, but in the canyon mass-wasted sandstone monoliths 2-5 m in diameter cover 20 to 100 percent of the bed (Figure 13). The gradient of the river steepens dramatically through this reach (Figure 10b). It remains an open question whether the gradient is controlled primarily by the necessity to erode the bedrock or by the need to transport and comminute the locally contributed alluvium. This influx of coarse sediment affects the channel gradient for about 10 km downstream from the canyon. The degree of bedrock exposure in this reach may have varied throughout the Quaternary due to climate changes, particularly as they affect mass wasting from the canyon walls.

A final example is the South Fork Eel River, California (Figure 10a). Between kilometers 74 and 85 the river has an atypically steep gradient and high relief, steep slopes adjacent to the channel. The steep reach has originated by an uncertain combination of higher bedrock resistance or rapid incision due to tectonic deformation or stream capture. A largely bedrock-floored channel with patches of fine gravel is present above the canyon (Figure 12). This bedrock reach probably undergoes episodic burial by sediment contributed by debris flows from tributaries during major floods [W.E. Dietrich, personal communication] and is probably an exemplar of the time varying rates of bedrock erosion shown in Figure 14. At present, bed erosion may be limited by the small quantity of bedload (the last major flood occurred in 1964). In the canyon reach erosion is limited by the opposite circumstance – a nearly complete cover of boulders derived from superjacent slopes and steep tributaries (Figure 11). Steep relief generated by earlier rapid incision has permitted rapid mass wasting and avalanche delivery of boulders to the valley bottom, largely burying the channel bed. These boulders exhibit the streamlined upstream faces, sharp downstream-pointed edges and concave potholing on the downstream faces that indicate that suspended load abrasion rather than impacts with other large boulders dominates their comminution. Abrasion by relative movement of bed particles without net transport (abrasion-in-place) may also occur

[Schumm and Stevens, 1973]. Deep scour pools downstream from constrictions and large bar forms indicate that all but the largest boulders can be moved locally by the largest floods, but it is uncertain whether there has been sufficient supply of boulders to force the overall gradient to a threshold condition.

DISCUSSION

Erosion of bedrock channels seldom involves just hydraulic detachment. In headwater channels some weathering must precede erosion, and rapid mass wasting is often the mechanism for mobilizing weathered bedrock. In larger channels a variable mix of hydraulic plucking, cavitation, abrasion by bedload and suspended load, and weathering is involved. Consequently there can never be a rate law representing a single process of bedrock channel erosion that applies universally [Seidl and Dietrich, 1992; Howard *et al.*, 1994]. Nonetheless, predictive models must be developed for various classes of dominant processes; this and other papers in this volume present initial analyses for certain types of bedrock channels. A number of unresolved and complicating issues remain unaddressed in these models, however.

One of these issues is how rock beds are eroded when an appreciable sediment cover is present. The simplest hypothesis assumes that erosion ceases when a single grain thickness is present over bedrock or if sediment transport reaches capacity. However, bedform migration, episodic scour and fill, and alluvial bed suspension during extreme floods may permit long-term erosion even with an appreciable bed cover under normal conditions. Perhaps gradual reductions in predicted erosion rate as sediment cover increases (e.g., (20)) might suffice in erosion models.

Modeling of long-term profile development in bedrock channels must account for the sediment flux through the channel. This is true not only because abrasion by bedload and sediment load may be the dominant erosional process, but also because locally contributed coarse sediment often partially or wholly mantles the bedrock channel, reducing erosion rates. Long-term rates of bedrock channel erosion are often regulated by delivery of coarse boulders to the valley bottom due to steep relief created by past rapid incision. Because these boulders must be comminuted in place before further transport by processes similar to those eroding the bedrock proper, channels partially or totally mantled by locally contributed coarse sediment can be viewed as a special type of bedrock channel.

The timescale over which bedrock channels have eroded is much longer than that for alluvial channel grading. Therefore, present processes and bed characteristics are not necessarily representative of those pertaining during development of the channel profile. In mixed alluvial-bedrock channels the fractional coverage by alluvium may vary

temporally. Inheritance of channel profiles, valley form, and slope and channel sediment from different past climates may be important. Most erosion may occur during extreme floods that have not occurred during the period of observation. Most rivers have been severely modified by recent land-use changes: deforestation leading to increased sediment delivery by mass wasting or overland flow, diminishment of large woody debris in channels, elimination of beavers, gravel mining, construction of flood levees, channelization, *inter alia*. In extreme cases, the "natural" morphology of the channel may be unknowable.

Despite these difficulties, field investigations, laboratory experiments, and theoretical model development will gradually improve our understanding of the evolution of bedrock channels.

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