River incision into bedrock: Mechanics and relative efficacy of plucking, abrasion and cavitation

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ABSTRACT

Improved formulation of bedrock erosion laws requires knowledge of the actual processes operative at the bed. We present qualitative field evidence from a wide range of settings that the relative efficacy of the various processes of fluvial erosion (e.g., plucking, abrasion, cavitation, solution) is a strong function of substrate lithology, and that joint spacing, fractures, and bedding planes exert the most direct control. The relative importance of the various processes and the nature of the interplay between them are inferred from detailed observations of the morphology of erosional forms on channel bed and banks, and their spatial distributions. We find that plucking dominates wherever rocks are well jointed on a submeter scale. Hydraulic wedging of small clasts into cracks, bashing and abrasion by bedload, and chemical and physical weathering all contribute to the loosening and removal of joint blocks. In more massive rocks, abrasion by suspended sand appears to be rate limiting in the systems studied here. Concentration of erosion on downstream sides of obstacles and tight coupling between fluid-flow patterns and fine-scale morphology of erosion forms testify to the importance of abrasion by suspended-load, rather than bedload, particles. Mechanical analyses indicate that erosion by suspended-load abrasion is considerably more nonlinear in shear stress than erosion by plucking. In addition, a new analysis indicates that cavitation is more likely to occur in natural systems than previously argued. Cavitation must be considered a viable process in many actively incising bedrock channels and may contribute to the fluting and potholing of massive, unjointed rocks that is otherwise attributed to suspended-load abrasion. Direct field evidence of cavitation erosion is, however, lacking. In terms of the well-known shear-stress (or stream-power) erosion law, erosion by plucking is consistent with a slope exponent (*n*) of ~2/3 to 1, whereas erosion by suspended-load abrasion is more consistent with a slope exponent of ~5/3. Given that substrate lithology appears to dictate the dominant erosion process, this finding has important implications for long-term landscape evolution and the models used to study it.

MOTIVATION

Much of the form and dynamics of mountainous landscapes is governed by the processes of bedrock channel erosion. Bedrock channels are characterized by minimal and/or transient alluvial sediment storage and thus occur wherever sediment transport capacity exceeds sediment supply over the long term (Howard et al., 1994). These conditions are commonly met in mountainous and tectonically active landscapes, and bedrock channels are known to dominate steeplands drainage networks (e.g., Wohl, 1993; Montgomery et al., 1996; Hovius et al., 1997). As the three-dimensional structure of drainage networks sets much of the form of terrestrial landscapes, it is clear that a deep appreciation of mountainous landscapes requires knowledge of the controls on bedrock channel morphology. Moreover, bedrock channels play a critical role in the dynamic evolution of mountainous landscapes (Anderson, 1994; Anderson et al., 1994; Howard et al., 1994; Tucker and Slingerland, 1996; Sklar and Dietrich, 1998; Whipple and Tucker, 1999) because they (1) communicate changes in boundary conditions (e.g., climate or tectonics) across landscapes; (2) govern, to first order, system response time to such perturbations; and (3) set the lower boundary condition on hillslopes throughout a landscape. Unfortunately, the processes by which rivers erode bedrock are not sufficiently well known at present to make definitive, quantitative statements about the relation between critical parameters in widely used river incision models and specific sets of field conditions (e.g., climate, lithology, and rock uplift rate).

Robust, quantitative descriptions of the processes of fluvial incision into bedrock and their dependence on climate, lithology, and channel slope stand out as critical unknowns in the study of landscape evolution. In a recent review of current theoretical understanding of bedrock channel evolution, Whipple and Tucker (1999) emphasized the importance of refining knowledge of the processes of erosion active in bedrock channels. Whipple and Tucker demonstrated that the quantitative relationship between erosion rate and channel slope (and thus shear stress or stream power) is a critical unknown in the study of bedrock channel dynamics because it strongly influences (1) fluvial system response timescale; (2) the relative importance of extreme hydrologic events in bedrock channel incision; and (3) the sensitivity of channel morphology (gradient, width) and lowering rate to rock mass quality, climate, and rock uplift rate (defined relative to base level). The last item governs how strongly coupled equilibrium landscape form is to prevailing tectonic, lithologic, and climatic conditions, whereas the response timescale governs how likely equilibrium conditions are to persist in an evolving landscape. Each of these critical aspects of landscape evolution dynamics is thus set by the physics of the processes of bedrock channel incision.

APPROACH AND SCOPE

In this paper, we strive to clarify the mechanics and relative roles of known fluvial bedrock channel erosion processes, the interactions among

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these processes, and the factors that determine process dominance in different field settings. After establishing a qualitative conceptual framework for the competition and interaction among the various processes of bedrock erosion, we outline the apparent rate-limiting physical mechanisms driving each process in an effort to provide a useful framework for future development of a quantitative, physically based analysis of the long-term channel incision problem. Our analysis extends and updates the treatment in Hancock et al. (1998). Here we consider in greater depth (1) first-order scaling of forces involved in fluvial plucking and abrasion, (2) relative contributions of saltating and suspended sediment grains in abrasion, (3) the conditions under which cavitation may occur and contribute to bedrock erosion, and (4) interactions between plucking and abrasion processes in long-term bedrock channel incision. In addition, we place our analysis in the context of the well-known shear-stress (or stream power) erosion law in order to explore the implications of our findings for patterns, rates, and styles of landscape evolution.

Our conceptual model for the competition between and interaction of plucking, abrasion, and cavitation in bedrock river incision is drawn from field study of erosion processes in actively incising bedrock rivers in a wide range of geologic, climatic, and tectonic settings (Table 1). Thus, in addition to the theoretical developments outlined above, we synthesize largely qualitative observations made in diverse field settings including the Indus River, Northwest Himalaya of Pakistan; several rivers draining the western Sierra Nevada, California; two rivers draining the Longmen Shan at the eastern margin of the Tibetan Plateau in China; and the Ukak River in Alaska's Valley of Ten Thousand Smokes. Together these field sites span local river gradients from 0.002 to 0.200, bankfull discharge from 25 to 4000 $m^3 s^{-1}$, and a wide range of substrate rock types, including (1) welded tuff and competent sandstone, (2) low-grade metamorphic rocks (China), and (3) granitic and high-grade metamorphic rocks (Table 1). This comprehensive overview of field observations comes at the expense of detailed quantitative field measurements of channel forms, rock properties, or process rates. Quantitative data presented here are limited to field estimates of average joint (or fracture) spacing characteristic of the studied river reaches, which vary over two orders of magnitude from ~1 m to ~10 m (Table 1). Analyses of selected field sites most suitable to quantitative investigation of processes and process rates are presented elsewhere. Hancock et al. (1998) presented a detailed account of the forms and rates of erosion by suspended-load abrasion from the Indus River field site. Dollenmayer et al. (1997) presented an initial accounting of a detailed study of plucking-dominated erosion on the upper Ukak River in Alaska.

Herein we consider strictly fluvial erosion processes, intentionally omitting discussion of debris-flow scour and glacial erosion. Moreover, we focus primarily on physical erosion (detachment) processes, and only briefly discuss weathering processes that act to prepare the bed for erosion. Erosion by solution, which is unlikely to rival physical erosion processes in most actively incising environments (e.g., Wohl, 1993) is not discussed.

FIELD CONSTRAINTS ON ACTIVE PROCESSES

Processes Contributing to Bedrock Channel Erosion

The suite of processes that potentially contribute significantly to bedrock channel erosion have long been known to include plucking, abrasion (due to both bedload and suspended load), solution, and cavitation (Alexander, 1932; Maxson and Campbell, 1935; Barnes, 1956; Baker, 1974; Shepherd and Schumm, 1974; Foley, 1980; Howard and Kerby, 1983; Miller, 1991; Wohl, 1992, 1993; O'Connor, 1993; Wohl et al., 1994; Zen and Prestegaard, 1994; Wende, 1996). In addition, chemical and physical weathering processes must contribute through weakening and preparation of the bed for erosion (e.g., Wohl, 1993). Indeed, dramatic evidence has recently been cited for the importance of volumetric expansion of biotites during weathering (Stock et al., 1996) and frost shattering in well-jointed, thinly laminated sedimentary rocks (Dollenmayer et al., 1997). What is most lacking in our current understanding of fluvial bedrock erosion processes is knowledge of the relative contribution of these various processes under different conditions. In addition, the physical mechanisms that comprise the processes of plucking, abrasion, and cavitation wear are only just beginning to come to light (Miller, 1991; Wohl et al., 1994; Hancock et al., 1998; Sklar and Dietrich, 1998), and few researchers have considered the interactions among processes that may importantly influence net erosion rates (e.g., Chappell, 1974). Our field observations were motivated by the need to develop a robust conceptual understanding of the mechanics of river incision into bedrock.

Morphological Constraints on Process Dominance

In this section we use detailed, qualitative observations of the morphology of erosional forms on channel bed and banks, and their spatial distribution, to infer the relative importance of the various processes and the nature of the interplay between them. We infer that plucking dominates where rock surfaces are predominantly exhumed joint, fracture, and bedding planes (henceforth referred to as joints), and bedload includes locally derived angular blocks. Conversely, we infer that abrasion dominates where surfaces are smooth and polished, where ripples, flutes, and potholes are prominently developed, and where there is a lack of exhumed joint planes. Previous studies have also attributed the development of such erosion features to the action of abrasion by suspended sand (Alexander, 1932; Maxson and Campbell, 1935; Baker, 1974; Wohl, 1992, 1993; Tinkler, 1993; Wohl et al., 1994; Zen and Prestegaard, 1994). In areas where both plucking and abrasion are visibly active, a rough quantitative estimate of the relative rates of these processes can be gained by estimating the volume percent of abrasion wear on recently plucked joint blocks (i.e., for equal contributions of abrasion and plucking, a 50% reduction in joint block size due to abrasion wear is expected). Whether abrasion is accomplished by bedload or suspended load can be inferred from the pattern of erosion on protuberances on the bed and the degree of coupling between erosional forms and fluid flow patterns.

Our field observations unambiguously demonstrate a strong lithologic control of dominant erosion processes and commensurate rates of bed lowering. Over the full range of channel gradient and bankfull discharge conditions examined, wherever rocks are well jointed on a submeter scale, we find that plucking is the dominant process (Table 1). Moreover, this observation holds true for all bedrock types examined, including moderately welded tuff, competent sandstone, low-grade metamorphic rocks, and high-grade metamorphic and granitic rocks (Table 1 and Fig. 1), and is consistent with previous accounts of plucking erosion in well-jointed rocks (Miller, 1991; Tinkler, 1993; Wende, 1996). Intrinsic rock strength appears to be of subordinate importance to the jointing characteristics of rock units as observed in studies of the stability of rock slopes (Selby, 1980). Although our field observations are insufficient to define a quantitative relationship, the threshold joint spacing for erosion by plucking clearly varies with stream discharge and gradient, with maximum block size entrained ranging from 2–3 m on the Indus River to 1–1.5 m on the larger Sierran rivers to ~0.5 m on the smaller Sierran streams (Table 1).

Visual inspection of the degree of abrasion wear on recently plucked blocks and exhumed joint planes indicates that, where active, plucking is easily an order of magnitude more efficient than abrasion (abrasion wear typically has removed well less than 10% of the volume of recently, or nearly, extracted blocks; Fig. 1). Under extreme conditions, rates of erosion by plucking can approach a decimeter per year (Dollenmayer et al., 1997).

For comparison, although the highest known rates of erosion by abrasion reach 4 mm yr^{-1}, these high rates are restricted to the axes of deep flutes, which must migrate across the bed to accomplish overall bed-surface lowering (Hancock et al., 1998). Average erosion rates determined on these fluted surfaces using cosmogenic radio nuclide dating indicate that net lowering rates by abrasion are an order of magnitude less than local maxima in flutes and potholes (Hancock et al., 1998). These quantitative estimates support the primarily morphologic evidence (Fig. 1) that plucking is the rate-limiting process of erosion through jointed bedrock.

In more massive rocks, abrasion by suspended sediment appears to be rate limiting, particularly in the larger, more powerful rivers studied here (drainage area > 500 km^2, gradient > 0.01). Where rocks are unjointed or are jointed at a scale that does not permit entrainment of joint blocks, erosion must proceed by a combination of abrasion by bedload, abrasion by suspended load, and, possibly, cavitation. Our observations consistently show that surfaces cut into massive, poorly jointed rock units are characterized by the following forms associated with abrasional wear: smooth, polished surfaces, erosional flutes and ripples, and potholes ranging from decimeters to meters in diameter (Fig. 2). Although cavitation erosion will be shown later herein to be a plausible mechanism in flows through many steep bedrock channels, and may contribute to pothole and flute erosion, we find no compelling field evidence either for or against a contribution from cavitation erosion. We make the conservative interpretation that erosion by abrasional processes dominates in sections of massive, unjointed rocks.

Our observations of the minimum joint spacing in abrasion-dominated reaches is consistently in accord with the earlier-cited limits on maximum joint block size in plucking-dominated reaches. Minimum joint spacings observed in abrasion-dominated reaches range from 1–1.5 m in the Sierra Nevada drainages studied up to 3–5 m on the Indus River in Pakistan (Table 1). As in the case of plucking in jointed rocks, this observation holds true over the full range of rock types we have studied to date (Fig. 2). Together, these data constitute a preliminary estimate of the critical joint spacing for the process transition from plucking to abrasion as a function of stream size and gradient. The finding that dominant erosion processes vary considerably between field settings, or even locally along a given channel reach, motivates our theoretical analysis of the mechanics of the plucking, abrasion, and cavitation processes. How different are these processes? What are the implications of these differences for long-term landscape evolution?

THEORETICAL FRAMEWORK: THE SHEAR-STRESS EROSION LAW

A widely used basis for modeling bedrock channel evolution is the shear-stress (or stream-power) erosion law:

$$\dot{\varepsilon}_t = KA^m S^n \quad , \tag{1}$$

where $\dot{\varepsilon}_t$ is the total erosion due to all active processes, K is a dimensional coefficient of erosion (primarily encapsulating effects due to lithology, climate, channel width, hydraulics, and sediment load), A is upstream contribution area (proxy for discharge), S is channel gradient, and m and n are positive constants that depend on erosion process, basin hydrology, and channel hydraulic geometry (Howard et al., 1994; Whipple and Tucker, 1999). The shear-stress erosion law (Howard and Kerby, 1983) is built on the reasonable premise that bedrock erosion rate can be approximated as a power-law function of the mean bed shear stress (τ_b):

$$\dot{\varepsilon} \propto \tau_b^a \quad , \tag{2}$$

where a is a constant that depends on the erosion process. In principle, a threshold shear stress below which no erosion occurs should be included in equation 2, but is often omitted in the interest of retaining the simple form of the shear-stress erosion law (equation 1).

Several researchers have recently demonstrated that the area and slope exponents (*m* and *n*) in the shear-stress erosion law exert a profound influence on the rates and styles of landscape evolution (Howard et al., 1994; Moglen and Bras, 1995a; Tucker, 1996; Whipple and Tucker, 1999). Whipple and Tucker (1999) stressed that the mechanics of the dominant erosion process (equation 2) most directly control the values of these critical exponents. This follows because the ratio of area and slope exponents (*m/n*) in equation 1 can be shown to be approximately constant for a broad subset of fluvial erosion processes, and there is a direct relationship between the shear-stress exponent (a) in equation 2 and the slope exponent (*n*) in equation 1. The shear-stress erosion model predicts that the *m/n* ratio for fluvial erosion processes will fall into a narrow range near 0.5 ($0.35 \le m/n \le 0.6$), depending only on the relative rates of increase of channel width with discharge and discharge with drainage area (Whipple and Tucker, 1999). Thus, the *m/n* ratio is notably independent of the mechanics of the various fluvial erosion processes (i.e., exponent a in equation 2). As a note of caution, although the *m/n* ratio is often regarded as synonymous with channel concavity, many factors (*m/n* ratio, lithologic variability, rock uplift patterns, downstream variations in sediment flux, disequilibrium conditions, and process transitions) can influence channel concavity (Sklar and Dietrich, 1998; Whipple and Tucker, 1999). The statement that the *m/n* ratio falls in a narrow range should not be mistaken for an argument that channel concavity is likewise restricted.

Unlike the *m/n* ratio, it is readily shown that the slope exponent *n* is directly dependent on the erosion process (Whipple and Tucker, 1999):

$$n = \frac{2a}{3} \quad . \tag{3}$$

Thus, assuming the *m/n* ratio is approximately constant for the range of fluvial erosion processes of interest here, the mechanics of the dominant erosion process dictate the values of both critical exponents in the shear-stress erosion law.

In the following section we develop simple theoretical expressions for the mechanics of erosion by plucking, abrasion, and cavitation. Our aim is to highlight in particular the differences between these processes in terms of the quantitative relationship between mean bed shear stress (τ_b) and erosion rate (equation 2), as this directly impacts the appropriate form of the shear-stress erosion law (equation 1) commonly used in landscape evolution studies.

MECHANICS OF FLUVIAL EROSION OF BEDROCK

Plucking

Erosion by plucking requires both the production of loose joint blocks—which generally involves weathering, crack propagation, and rock fracture—and the subsequent entrainment and transport of loosened joint blocks. For the case of plucking by glaciers (also called quarrying), Hallet (1996) argued that the production of joint blocks associated with crack propagation is rate limiting and then derived a simple expression for quarrying rate as a function of glaciological variables. Hallet cited the dearth of large rock fragments on bedrock surfaces recently exposed by glacial retreat as evidence that englacial entrainment and transport of blocks is much more efficient than block fracture. The rate limiting process in fluvial erosion by plucking is less clear, though the balance of evidence indicates that entrainment rather than rock fracture may be the limiting step, at least for larger joint blocks. Certainly, fluvial transport of large blocks is considerably less efficient than englacial transport. Actively plucked channel reaches are often littered by significant accumulations of occasionally mobile joint blocks (Fig. 3). Accordingly, the "mixed" bedrock–alluvial channel type discussed by Howard (1996) is quite common in rivers incising through well-jointed rocks. As discussed by Howard (1996), it is likely that both the erosion of new joint blocks and the transport of the resulting lag deposit are important controls on the net rate of bedrock incision by plucking. Which is rate limiting at a given time and place may depend on downstream and hydrologic boundary conditions that influence the efficiency of boulder transport. Because the processes of particle transport in rivers are relatively well known (Graf, 1977), we focus instead in the following paragraphs on the suite of processes that contribute to the production and initial extraction or entrainment of joint blocks.

The production of transportable joint blocks on river beds has many similarities to the subglacial case considered by Hallet (1996), although the processes contributing to the development of differential stresses sufficient to propagate fractures and loosen joint blocks are qualitatively and quantitatively different. Field observations and considerations of the physics of fluid flow and sediment transport indicate that at least four processes contribute to the fracture and loosening of joint blocks: (1) chemical and physical weathering along joints (e.g., frost shattering of exposed surfaces in winter), (2) hydraulic wedging of sand, pebbles, and fine gravel into progressively opening cracks (see Fig. 8 in Hancock et al., 1998), (3) vertical and lateral crack propagation induced by high instantaneous differential stresses associated with impacts of large saltating clasts, and, possibly, (4) crack propagation induced by flexing of the bed associated with instantaneous pressure fluctuations in intense turbulent flows (Fig. 4).

Unfortunately, the complex suite of processes involved and the extreme sensitivity of crack propagation rates to the instantaneous differential stresses developed within the rock (see discussion by Hallet, 1996) preclude at present the development of a definite, quantitative statement of the relationships among fluid-flow variables (e.g., flow depth, mean velocity, mean bed shear stress), bedload sediment flux, and the rate of joint-block fracture and loosening. Sklar and Dietrich (1998) offered one approach, based on the engineering "wear" literature, which probably is best applied to weak or finely jointed (decimeter scale) rock types where the dominant erosion process is more directly akin to abrasion than plucking of large joint blocks, although their model may also effectively capture the mechanics of plucking erosion where block fracture due to bedload impacts is rate limiting. For more competent rocks, joint-block fracture and loosening rate ($\dot{L}$) is likely independent of bedload flux, again unless crack propagation induced by impacts of large saltating grains (3 above) is the dominant mechanism. In this case, we might expect a relationship between bedload flux (q_s) and block loosening rate ($\dot{L}$). In all other cases we expect no strong, direct dependence of block loosening rate on sediment flux. Because several researchers have argued for an important relationship between bedload flux and erosion rate (for abrasive wear, Foley, 1980; Sklar and Dietrich, 1998), we derive an estimate of the maximum possible contribution of bedload sediment flux to block loosening rate by using a relationship for transport-limited bedload flux and assuming a direct dependence of block loosening rate on bedload flux.

Transport-limited bedload flux is reasonably well described by the Meyer-Peter-Mueller transport equation (Meyer-Peter and Mueller, 1948), which can be written in general terms as

$$qs \propto (\tau_b - \tau_c)^{3/2}, \tag{4}$$

where τ_c is the critical shear stress to entrain coarse bedload. Assuming a simple power-law relation between bedload flux and block loosening rate, we may write

$$\dot{L} \propto q_s^p \propto \left(\tau_b - \tau_c\right)^{3p/2}, \tag{5}$$

where p is an unknown positive constant probably close to unity. Protection of the bed by a partial bed cover may dampen the dependence on sediment flux (Sklar and Dietrich, 1998). Also, the size of transported grains will likely play an important role, similar to the role of grain size in the Sklar and Dietrich (1998) model. Where rare impacts by very large grains capable of generating sufficient differential stresses at crack tips to fracture joint blocks are rate limiting, block loosening rate may depend mostly on the critical shear stress to move large blocks (similar in form to equation 7 below). Joint-block loosening driven by other processes is expected to be less nonlinear in shear stress. Thus the formulation of equation 5 serves as an end-member case for comparison with block extraction processes and abrasional processes.

Once a set of open joints has formed around a block, allowing communication of fluid pressure under the block, the block must be extracted from its position in the bed before normal processes of particle transport take over. For a joint block wedged in between adjacent blocks (Fig. 4), entrainment is difficult but has been observed to occur in the laboratory (Reinius, 1986) and in the field (Wende, 1996; Dollenmayer et al., 1997). Forces resisting entrainment include the normal component of block buoyant weight, friction on the lateral ($F_{fy}|_l$), upstream ($F_{fy}|_u$) and downstream ($F_{fy}|_d$) block edges, and the instantaneous pressure force averaged across the upper surface of the block ($\bar{p}'_s$), where the overbar denotes a spatial and not a temporal average. The only forces available to lift the block out of place are the instantaneous pressure force averaged across the base of the block ($\bar{p}'_s$) and, possibly, drag forces associated with through flow upwelling at the downstream margin (F_{Dz}). Note that the lateral frictional forces include the effects of rotation of the block as it is lifted up into the flow, the crushing or displacement of the granular material wedged into cracks, and the component of block weight normal to the sidewalls. Ignoring the possible effects of through flow, the condition for entrainment for a block of thickness h, width w, length l, and density ρ_s can be written

$$\bar{p}'_b - \bar{p}'_s \geq g\left(\rho_s - \rho\right)h + \left(F_{fx}\Big|_u + F_{fx}\Big|_d\right)\frac{h}{l} + 2\left(F_{fy}\Big|_l\right)\frac{h}{w}. \tag{6}$$

Equation 6 implies that this type of floating entrainment is most effective for small block thickness-to-length ratios, consistent with laboratory observations (Reinius, 1986) and inferences from field relations between large rock slabs and the holes from which they were extracted (see Fig. 1b; Wende, 1996; Dollenmayer et al., 1997). The recent treatment by Hancock et al. (1998), which intended to derive a minimum threshold for block entrainment, neglected side-wall frictional effects and thus did not acknowledge the importance of block shape (i.e., h/l and h/w).

It is well known that mean dynamic pressures scale with mean shear stress (e.g., Graf, 1977). Assuming that the magnitude of instantaneous pressure fluctuations also scale with mean shear stress, we can infer that block initial block extraction rate ($\dot{E}$) scales with mean shear stress beyond a critical value (τ_e):

$$\dot{E} \propto \tau_b - \tau_e. \tag{7}$$

In cases where the downstream neighbor has already been removed, entrainment of joint blocks is facilitated in several ways: (1) a large pressure drop develops at the downstream end of the block; (2) the block is now free to slide out of the pocket (Hancock et al., 1998); (3) frictional forces associated with block rotation and the displacement of granular material wedged into cracks are somewhat alleviated; and (4) drag forces on the block increase rapidly as the block is lifted and tilted off the bed (Fig. 4). Note that both the induced pressure drop and the drag forces on a tilted block also scale with mean basal shear stress (τ_b).

Thus, although a satisfying mechanistic description of erosion by plucking proves difficult to obtain because of the complex suite of processes that are operative, a simplified analysis indicates that the rate of erosion by plucking should be only linear to weakly nonlinear in shear stress beyond a critical threshold (equations 5 and 7).

Abrasion by Suspended Load

Unjointed, cohesive rock types are eroded principally by incremental wear processes, probably dominated by abrasion by sediment grains transported in the flow. Foley (1980) and recently Sklar and Dietrich (1998) presented quantitative formulations of abrasion wear erosion for the case of abrasion by saltating bedload particles. Here, we outline the mechanics of abrasion erosion by suspended-load particles. Later, in the Discussion section, we address the issue of whether, and under what circumstances, suspended load may play the dominant role in erosion of massive, relatively unjointed rocks.

Erosion by abrasion has been studied in both fluvial (Alexander, 1932; Foley, 1980; Sklar and Dietrich, 1997, 1998; Slingerland et al., 1997) and eolian environments (Sharp, 1964; Whitney, 1978; Suzuki and Takahashi, 1981; Greeley and Iversen, 1985; Anderson, 1986). In principle, eolian and fluvial abrasion are fundamentally the same, with greater fluid density and viscosity in the fluvial case being the only significant differences. We follow the analysis of eolian abrasion wear given by Anderson (1986) in our discussion of abrasion by suspended sediment.

Following work by Scattergood and Routbort (1983), Greeley and Iversen (1985), and Suzuki and Takahashi (1981), Anderson (1986) showed, where threshold impact velocity and threshold particle size are small, that the erosion rate due to impact abrasion ($\dot{\varepsilon}_t$) can be approximated as

$$\dot{\varepsilon}_a = \frac{S_a q_{ke}}{p_r}, \tag{8}$$

where S_a is the susceptibility of the substrate to abrasion, q_{ke} is the normal flux of kinetic energy to a rock surface per unit time, and ρ_r is the rock density. The flux of kinetic energy due to suspended grains of size class r in the fluid (Anderson, 1986) is

$$q_{ke} = \frac{1}{2} p_s C_{vr} U^3, \tag{9}$$

where ρ*s* is sediment density, C_{vr} is the volumetric concentration of sediment of size r, and U is the fluid velocity. Because suspended sediment concentration (C_{vr}) scales with fluid velocity squared (Anderson, 1986) in a transport-limited setting, the kinetic energy flux, and thus abrasion rate, scales with flow velocity to the fifth power. Given that standard hydraulic resistance relations show that mean bed shear stress scales with velocity squared (Graf, 1977), we may write

$$\dot{\varepsilon}_t \propto q_{ke} \propto U^5 \propto \tau_b^{5/2}. \tag{10}$$

Note that at high flood stages in bedrock channels suspended sediment concentration may be supply limited, so equation 10 represents an upper bound on the sensitivity of abrasion erosion to shear stress.

Anderson's (1986) treatment also provides insight into the spatial distribution of erosion by suspended-load abrasion. The actual flux of kinetic energy delivered to the rock surface is a function of both the flux of kinetic energy due to suspended particles in the fluid and a "capture efficiency," which describes the relative decoupling of particles from the flow required for particles to impact the rock surface. Capture efficiency is determined by streamline curvature, fluid density, fluid viscosity, and particle mass. The largest particles have considerable inertia and must strike obstructions on the upstream side. Intermediate size particles follow gently curved streamlines upstream of obstructions, but impact on the lee side of obstructions where shedding vortices produce sharply curved streamlines. The smallest particles follow streamlines faithfully and never impact the rock surface (Fig. 5). The greater density and viscosity of water and their effect on capture efficiency are responsible for the most significant difference between fluvial and eolian erosion: whereas eolian abrasion is focused on upstream sides of obstructions where sand and granule-sized particles impact, fluvial abrasion is focused on the lee of obstructions or negative steps (knickpoints) and in potholes where suspended sand and granules spun in vortices energetically impact the rock surface (Figs. 2 and 6).

Cavitation

Cavitation-induced erosion of dam spillways, tunnels, and turbines is well known in the engineering literature and can be responsible for extremely rapid rates of incision into concrete flow structures (see Fig. 1 in Arndt, 1981). Since the pioneering work by Barnes (1956), the possible role of cavitation in the erosion of bedrock channels has been considered sporadically in the literature (Baker, 1974; Ball, 1976; Wohl, 1992; O'Connor, 1993) but remains elusive. In fact, considerations of cavitation erosion in the geological literature, including a recent paper by the authors (Hancock et al., 1998), have thus far been limited to direct applications of the work of Barnes (1956), despite the fact that cavitation inception and cavitation damage have been the focus of much attention in the field of hydraulic engineering. Much of the body of knowledge of the cavitation phenomenon has been developed since the time of Barnes's work (see reviews by Hammitt, 1980; Arndt, 1981; Arndt and Maines, 1994). Here we briefly review recent advances in the understanding of cavitation in turbulent shear flows with the aim of constraining the plausibility or implausibility of cavitation erosion as a mechanism for enhancing erosion of flutes and potholes in natural channels.

Cavitation is the formation of vapor and air bubbles in water, which occurs when local fluid pressure drops below the vapor pressure of dissolved air (e.g., Arndt, 1981). Cavitation damage occurs when bubbles are advected into regions of higher pressure and caused to implode in the vicinity of the water-rock interface. The onset of cavitation is usefully cast in terms of the cavitation inception index (σ), which is defined as the ratio of the difference between hydrostatic pressure (p_o) and vapor pressure (p_v) to the free-stream dynamic pressure, as follows:

$$\sigma = \frac{P_o - P_v}{\frac{1}{2}\rho U_o^2}, \tag{11}$$

where *r* is fluid density and *Uo* is the free-stream velocity. In principle, cavitation should occur whenever the cavitation inception index falls below unity (as assumed by Barnes, 1956). However, the critical cavitation inception index (σ_c) is importantly influenced by flow and environmental conditions (Arndt, 1981). Of particular interest to the problem of river incision into bedrock are observed dependencies of σ_c on flow Reynolds number, the concentration of fine suspended sediment, and the degree of flow aeration, including both the concentration of air bubbles and the dissolved air content.

Turbulent shear flows exhibit instabilities that result in the formation of both spanwise and streamwise vortices, particularly in shear layers developed downstream of obstructions or negative steps (e.g., O'Hern, 1990). At high flow Reynolds numbers (10^5–10^6), cavitation is observed to occur at values of the cavitation index (σ) as high as 3.6 and commonly in the range of 2–3 (Arndt, 1981; O Hern, 1990). O Hern (1990), Arndt and Maines (1994), and others have demonstrated that cavitation bubbles develop preferentially in streamwise vortices of the type associated with erosional flutes (see Fig. 2 in O'Hern, 1990) because local pressures in the cores of vortices can drop well below mean dynamic pressures considered in the definition of the cavitation index (equation 11). For example, in studies of rapid erosion of dam spillways, cavitation damage is often clearly associated with the formation of horseshoe vortices around flow obstructions (see Fig. 1 in Arndt and Maines, 1994).

O'Hern (1990) reports instantaneous pressure fluctuations on the order of ±300% of the mean dynamic pressure, consistent with the observation of cavitation inception at σ = 2–3 in his study. Further, cavitation inception depends not only on the requisite drop in dynamic pressure but also on the presence of a sufficient concentration of bubble nucleation sites of appropriate size (Arndt, 1981). Nucleation points are usually either small air bubbles or fine suspended sediment particles. The concentration of small air bubbles increases with dissolved air content, and O'Hern (1990) demonstrated a strong positive correlation between σ_c and dissolved air content, which varied between 3 and 15 ppm (saturation) in his experiments. Thus, owing to the natural aeration of waters in turbulent open channel flow, the presence of significant concentrations of fine sediment particles, and the development of powerful vortices shed from irregularities in bedrock channels, cavitation is in fact considerably more likely to occur than surmised by Barnes (1956).

There is, however, a dual role of flow aeration that must be considered, as pointed out by Barnes (1956) and recently by Hancock et al. (1998). First and foremost, venting of pressures through communication with the flow surface in areas of frothing white water in shallow flows is likely to strongly inhibit cavitation. Second, although the presence of air bubbles may induce cavitation at values of σ greater than unity, the air actually cushions the collapse of cavitation bubbles, thus inhibiting vigorous cavitation erosion. Thus, in settings where cavitation might not otherwise occur, flow aeration (air bubbles) can induce cavitation and produce some cavitation damage. However, in settings where vigorous cavitation is already occurring, aeration can substantially reduce the amount of cavitation damage. This latter effect has been demonstrated experimentally (Arndt et al., 1993) and has been employed as a mitigation strategy in engineering structures at high risk of severe cavitation erosion.

Following Baker (1974) and O'Connor (1993), we find it useful to consider the conditions for cavitation erosion in terms of the cross-sectionally averaged critical flow velocity (U_c). It is commonly observed that critical or near-critical flows (Froude number = 1) are often attained during high flows in steep bedrock channels (Tinkler, 1996; Tinkler and Wohl, 1996). Critical flows are defined by a Froude number of unity, such that critical velocity (U_c) is given by

$$U_c = \sqrt{gd}\ , \tag{12}$$

where g denotes gravitational acceleration and d flow depth. We calculate the flow velocities necessary for the inception of cavitation in vortex cores relative to critical flow conditions as a function of flow depth. Based on the findings of Arndt (1981) and O Hern (1990), we arbitrarily set the condition for possible cavitation as σ = 4 and that for likely cavitation as σ = 2 (Fig. 7). For flow depths in excess of 10 m, cavitation is possible at local velocities equal to the cross-sectional average velocity for critical flow (Fig. 7a). Because local free-stream velocity over and around protuberances may greatly exceed the cross-sectional average, we also calculate the velocity excess required for cavitation as a function of flow depth (Fig. 7b). For convenience, the local velocity excess in Figure 7b is defined as the ratio of local velocity to the critical cross-sectionally averaged velocity (U_l/U_c). Using equation 12, U_c can be calculated readily for any flow of interest, and the local velocity required to induce cavitation (U_l) can be determined from Figure 7b.

The calculations given by Baker (1974) and O'Connor (1993) differ from ours in that they follow Barnes (1956) in assuming σ_c = 1 and a local velocity excess (U_l/U_c) of 1.5–2.4. Similarly, Hancock et al. (1998) considered cavitation "possible" at σ = 1 and U_l/U_c = 3, and cavitation "likely" at σ = 1 and U_l/U_c = 2. Our calculations indicate that, for U_l/U_c = 2, cavitation is possible for flow depths greater than 2 m and likely for flow depths greater than 4 m. For comparison, annual peak flows on the Indus River average 15 m deep with peak velocities of approximately 10 m/s through the study reach. From Figure 7b, at 15 m flow depth, a velocity excess of 1.3 is required for the likely onset of cavitation. From equation 12, annual peak flows in the Indus have cross-sectional velocities that approach critical ($U = 0.8U_c$). Thus, so long as local velocities around flow obstructions exceed the mean velocity by more than 70%, cavitation may well be induced down the cores of vortices during annual peak flows on the Indus River, contrary to the conclusions drawn by Hancock et al. (1998).

Cavitation erosion must be considered a plausible candidate in the erosion of flutes and potholes, which we conservatively attribute to abrasion by suspended particles given the lack of any definitive evidence that cavitation-induced erosion is occurring. Although cavitation damage leaves distinctive pitted textures on eroded metallic surfaces (e.g., turbine blades; Arndt, 1981), there is no known distinctive signature of cavitation-damaged rock surfaces. In addition, if cavitation is indeed contributing to flute and pothole erosion, it certainly acts in concert with active abrasion, which may eradicate any sign of cavitation damage. However, it is possible that the high temperatures and impact pressures associated with the implosion of cavitation bubbles (Arndt, 1981) may leave a distinctive chemical/physical signature on stressed mineral crystals. This may prove a fruitful area of research in future studies of cavitation erosion in natural systems.

If active as an erosion process, the rate of cavitation erosion ($\dot{\varepsilon}_c$) may be characterized as highly nonlinear in flow velocity and depth, requiring a threshold velocity (U_{cav}) that increases with depth (Fig. 7) and increasing rapidly as cavitation proceeds from initial inception to increasingly full development (Arndt, 1981):

$$\dot{\varepsilon}_c \propto \left(U - U_{cav}\right)^q \propto \tau_b^{q/2}, \tag{13}$$

where U_{cav} is a function of flow depth, fine sediment concentration, atmospheric pressure, dissolved air content, and flow Reynolds number, and values of q up to 7 have been reported (Murai et al., 1997). In addition, it is clear that the rate of cavitation-induced wear depends on the type of cavitation (sheet vs. vortex core), cavity size, cavity pressure differential, and the frequency of cavity formation and collapse. However, despite a vast and rapidly growing literature, at present we are unable to provide a more robust scaling of cavitation wear than that given in equation 13.

DISCUSSION

Relative Roles of Suspended and Bed-Load in Abrasion Wear

Several researchers have argued that erosion rates in bedrock channels should scale with sediment flux rates (Foley, 1980; Sklar and Dietrich, 1997, 1998; Slingerland et al., 1997). The importance of sediment flux is corroborated by our observations and analyses of both plucking and abrasion erosion mechanisms. However, it seems unlikely that any one of the models proposed to date is applicable to the full range of field settings studied here. Indeed, in all the river systems we have investigated, plucking-dominated reaches where bedload flux may play a critical role commonly alternate with abrasion-dominated reaches where the importance of bedload flux is less clear. Both morphological field evidence and process-rate data (presented in Hancock et al., 1998) argue that abrasion by suspended-load particles is dominant in a wide range of field settings (Table 1) in reaches where widely spaced joints and fractures inhibit erosion by plucking, particularly in the larger, more powerful river systems we have studied. Conversely, in the small coastal streams (drainage area < 20 km^2) studied by Snyder et al. (2000), erosion by bedload-driven plucking and bedload abrasion appear to be the dominant processes.

The interpretation that bedload abrasion is less important in the larger river systems studied here is supported by several lines of evidence. Concentration of erosion on downstream sides of obstacles is indicative of abrasion by suspended-load particles (sand in the systems studied here) and is documented both by the spatial distribution of flutes and potholes (Figs. 2 and 6) and by process-rate measurements. Whereas we have documented surface lowering rates of up to 4 mm yr^{-1} in the cores of large flutes and potholes on the Indus River, both erosion-hole monitoring and cosmogenic radionuclide dating of upstream-facing surfaces have recorded negligible lowering rates (Hancock et al., 1998). Some upstream-facing surfaces are in fact lightly varnished, while the downstream side shows signs of rapid erosion (Fig. 6a). In addition, tight coupling between fluid-flow patterns and fine-scale morphology of erosion forms (Figs. 2 and 6) rule out an important component of abrasion by coarse bedload on the abrasion-dominated surfaces we have examined in the field. Finally, we find a dearth of crescentic fractures, percussion marks, and other signs of significant damage associated with impacts by large bedload particles. We do not argue that this type of impact damage does not occur, only that the preponderance of morphological evidence on the bed and banks of the powerful rivers we have studied indicates that this process is secondary to vigorous fluting and potholing.

A criticism of our evidence that abrasion wear is dominantly accomplished by suspended sediment in these systems may be raised: owing to limitations of accessibility, many of the bedrock surfaces we have examined are generally above the bed and away from the thalweg, where most coarse bedload transport probably occurs. However, we have observed the same erosional forms and patterns on every bedrock surface cut in unjointed rocks that we have examined, including many that do lie directly on the channel bed (see Fig. 8). Moreover, there are a number of factors that together may enhance abrasion by suspended load and retard abrasion by bedload, as outlined below.

As argued by Sklar and Dietrich (1997, 1998), much of the kinetic energy associated with vigorous bedload transport can be dissipated in collisions between one saltating grain and another, and between saltating grains and a basal carpet of sliding and rolling clasts. In addition, in most fluvial systems bedload sediment accounts for no more than 5%–10% of the total sediment flux. Although one might argue that much of the suspended sediment flux passes through the system without much interaction with the channel boundaries, channel reaches where joint block plucking is inhibited typically develop significant and stable topographic irregularities, which generate intense vortices that bring the suspended load into contact with the bed (Fig. 5). These vortices in fact focus abrasion damage on specific areas of the bed, resulting in the initial development of flutes and potholes. 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 suspended particles. Finally, it is plausible that the inception of cavitation bubbles down the cores of vortices contributes to the focusing of erosion in flutes and potholes, as has been argued by some previous investigators (Baker, 1974; Wohl, 1992; O'Connor, 1993). If cavitation indeed occurs in natural systems, the likely onset of cavitation within vortices during high-velocity flow may help explain the apparent dominance of fluting and potholing over abrasion wear by large, vigorously saltating particles.

Thus we conclude that impacts due to large saltating clasts probably contribute importantly to the "plucking" process (as observed by Snyder et al., 2000), but are ineffective on massive, unjointed rocks. A likely exception is expected in rock types that are weak at the grain scale, such that abrasion wear by saltating particles will outpace the loosening and extraction of joint blocks. Although an argument about whether "bashing" of the bed by bedload particles should be considered as part of the "plucking" or "abrasion" process may seem purely semantic, the issue is, in fact, important. This follows because erosion rates due to plucking (whether driven by bedload impacts or not) and abrasive wear scale quite differently with channel gradient and water discharge as demonstrated in the preceding section. Such differences in scaling have important implications for the rates and styles of landscape evolution.

Interactions Between Processes

The predominant mechanisms of bedrock channel incision, plucking, and fluting/potholing—associated with abrasion by suspended sediment and possibly cavitation—interact both locally and at the reach scale. Local interactions play a direct role in the lowering of the bed over the short term. Joint spacing and the openness of individual joints is highly variable in most rock units. Efficient plucking of joint blocks from zones of more locally fractured rock produces the rough bed and bank microtopography required to initiate the flow separation and vortex formation that drives the abrasion process (Fig. 5). In turn, abrasion and pothole formation can wear away massive rock units until the walls between adjacent potholes become thin enough to be fractured and carried away by plucking (Fig. 2).

At the reach scale, this interaction of plucking and abrasional processes appears to be responsible for the formation of discontinuous strath terraces that are commonly preserved along bedrock channels (see Fig. 3b and Wohl, 1992, 1993). Reach-scale (tens of meters) variations in joint spacing set up a spatial variation in process dominance. Well-jointed plucked reaches erode more rapidly than adjacent massive reaches that must be eroded by slower abrasional processes. Whereas the downstream ends of massive rock ribs tend to form prominent knickpoints (1–10 m high), erosion of upstream jointed reaches is limited by the local base level of the slowly eroding rock rib (Figs. 3 and 9). Upstream jointed sections tend to become buried in eroded joint blocks that cannot be transported away on the reduced channel gradient. Massive rock ribs are slowly beveled into a wide strath until the downstream knickpoint develops to such a point that sufficient focusing of turbulent energy into vortices shed from the topographic step induce concentrated, intense abrasion and pothole formation. Once begun, this process progressively cuts a narrow inner gorge through the beveled rock rib, producing a discontinuous strath of the type seen in experiments (Shepherd and Schumm, 1974) and in the field (Baker, 1974; Wohl, 1992, 1993; O'Connor, 1993). Discontinuous straths produced in this manner naturally converge with the channel bed in the upstream direction.

Bedrock rivers in the western Sierra Nevada are replete with examples of alternating stretches of mixed bedrock-alluvial conditions (presumably eroded by plucking) and short, beveled straths cut by narrow inner gorges (e.g., Fig. 3). These straths indicate that both plucking and abrasional processes are important in these rivers, and that the three-dimensional distribution of intact rock ribs may limit both bedrock channel lowering rate and equilibrium slope, as illustrated schematically in Figure 9. How to account for this sort of spatial variability and interaction of processes in reach-scale erosion laws useful for modeling landscape evolution is an important and unanswered question.

Shear-Stress Erosion Law and Implications for Landscape Evolution Dynamics

Since its introduction some 15 years ago (Howard and Kerby, 1983), the shear-stress (or stream-power) erosion law (equation 1) has been used extensively in landscape evolution modeling studies (e.g., Seidl and Dietrich, 1992; Anderson, 1994; Howard, 1994; Howard et al., 1994; Rosenbloom and Anderson, 1994; Seidl et al., 1994; Humphrey and Heller, 1995; Moglen and Bras, 1995a, 1995b; Tucker and Slingerland, 1996, 1997). In most such modeling studies to date the slope exponent (n) has been assumed to be either 2/3 (the so-called "shear-stress" erosion rule) or 1 (the so-called "stream-power" erosion rule). Although these choices were appropriate given the level of knowledge of the operative processes of erosion at the time, Whipple and Tucker (1999) recently showed that the behavior of landscape evolution models is critically dependent on the choice of this parameter (especially so given the argument that the m/n ratio falls into a narrow range near 0.5). Our analysis indicates that erosion by plucking is most consistent with values of n near unity. Erosion by suspended-load (sand) abrasion is more nonlinear, consistent with $n = 5/3$. In addition, cavitation erosion, if active, exhibits a high initiation threshold and scales with shear stress to a high power, suggesting values of n as high as 7/3. Erosion by bedload abrasion (or plucking where rock fracture driven by bedload impacts is rate limiting) is probably best described by the Sklar and Dietrich (1998) model. Importantly, our extensive, if qualitative, field observations indicate that the dominant erosion process, and thus the effective value of the critical exponent n, is strongly dependent on the characteristic joint spacing of the substrate rock units (Table 1).

Whipple and Tucker (1999) emphasize that the relationship between erosion rate and slope (exponent n in equation 1) importantly influences (1) the sensitivity of equilibrium channel slope (and thus equilibrium range crest elevation) to uplift rate, rock mass quality, and precipitation; (2) the pattern of transient response to a change in rock uplift rate or climate; (3) the timescale of response to such a change; and (4) the sensitivity of erosion rate to extreme hydrologic events. Thus, the above findings that the dominant process is a sensitive function of lithology (mostly dependent on jointing or fracture density), and that the rates of erosion by plucking, abrasion, and cavitation each scale differently with the mean bed shear stress imply that the rates and styles of landscape evolution may differ significantly in both space and time. As inferred by Stock (1996) and Stock and Montgomery (1999), there may be, in addition to the type of short-term variations in process dominance implied in Figure 9, dramatic changes in the dynamics of landscape evolution during an orogenic cycle as weak, fractured sedimentary cover rocks are stripped away and the resistant and more massive crystalline core of the range is exposed.

Besides grappling with this range of plausible erosion-law exponents, handling the scaling jump from local flow conditions (velocity, shear stress, vorticity) to reach-averaged conditions useful to large-scale landscape evolution modeling looms as a significant challenge. Field observations and analysis of the processes involved both attest to the sensitivity of abrasion rate to local flow conditions. The generation of intense vortices that plays the central role in erosion by abrasional processes is entirely dependent on local topographic irregularities and convective accelerations that are explicitly not considered in the shear stress or stream power erosion law. Indeed, Dick et al. (1997) and Hancock et al. (1998) have documented short-term erosion rates that are highly correlative with local slope and yet show almost no correlation to reach-averaged slope. Relations must be sought between the frequency and size of topographic irregularities (local drops) and reach-averaged slope.

CONCLUSIONS

Field investigation of the morphology and distribution of erosional forms on the bed and banks of bedrock channels in a wide variety of geologic, climatic, and tectonic settings indicates that plucking, bashing by bedload, abrasion by suspended load, and possibly cavitation all contribute importantly to river incision into bedrock. Which erosion process is rate limiting in a given place at a given time is dictated largely by the joint or fracture density in the rock. The marked efficiency of the plucking process ensures that wherever rocks are pervasively jointed at a submeter scale, plucking will be the dominant erosion process. Both field observations and theoretical predictions suggest that the critical joint spacing for the transition from abrasional to plucking processes increases with channel gradient and bankfull discharge (Table 1). Impacts by saltating bedload, wedging of small pebbles into existing cracks, and flexing of the bed due to pressure fluctuations act in concert to fracture rock and open preexisting joint and/or bedding planes. Rapid pressure fluctuations in turbulent flow apparently play a key role in the initial extraction and entrainment of large joint blocks. The common occurrence of mixed bedrock-alluvial channels, particularly in reaches with pervasively fractured rock, suggest that plucking may be such an efficient erosion mechanism that channel floors become armored with large joint blocks and tend toward a transport-limited rather than detachment-limited condition.

Where rocks are massive or joint sets are so widely spaced that plucking is inhibited given the prevailing channel slope and discharge conditions, some combination of abrasion by bedload, abrasion by suspended load, and cavitation is responsible for bed lowering. 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—an observation that holds true for the full range of cohesive, unjointed rock types we have investigated (from gneiss to moderately welded tuff). The strongest vortices are shed in zones of flow separation on the downstream side of bed protuberances (or small knickpoints). These vortices drive localized abrasional processes, possibly including wear due to cavitation damage (Barnes, 1956; Arndt, 1981). As a result, almost all observable abrasion wear occurs on the downstream side of flow obstructions (or negative steps). This spatial distribution of abrasion wear rules out any significant contribution by saltating bedload particles, which must impact the upstream side of any obstructions in the flow (Fig. 5). The greater flux of suspended sediment, the localization and concentration of erosive action by stabilized vortices, strong damping of bedload abrasion by clast-clast collisions (Sklar and Dietrich, 1997, 1998), and possibly a contribution due to cavitation damage are put forward as plausible explanations for the observed spatial pattern of abrasion wear and the implied dominance of suspended-load abrasion. Bedload abrasion may be more important in less powerful river systems (Sklar and Dietrich, 1998; Snyder et al., 2000) and may contribute importantly to the plucking process.

Scaling analyses of the various erosion processes indicate that erosion by plucking and erosion by combined abrasional and cavitation processes are governed by markedly different physics and cannot be described by a single universal bedrock channel erosion law. The different scalings between erosion rate and mean bed shear stress (or stream power) have far-reaching consequences in terms of the rates and styles of landscape evolution (e.g., Howard et al., 1994; Whipple and Tucker, 1999). However, several key issues remain to be resolved before the findings presented here can be confidently translated in terms of long-term and large-scale dynamics of bedrock channel systems. First and foremost, most rock units exhibit much heterogeneity in terms of joint spacing and rock mass quality over scales of tens to hundreds of meters. The parameters that determine whether the rate is rate limiting at which the occasional massive rock rib can be worn through by abrasional processes have yet to be elucidated. Directly related to this is the unresolved issue of how to appropriately scale abrasional processes, which are sensitive to local flow conditions that drive vortex generation, in terms of reach-averaged quantities useful for large-scale modeling. One potentially useful avenue of research will be field studies designed to constrain the long-term relationship between reach-averaged slope and erosion rate (e.g., Stock, 1996; Snyder and Whipple, 1998; Stock and Montgomery, 1999; Snyder et al., 2000), which could be interpreted in terms of the long-term, spatially averaged rate-limiting process(es). In addition, further work is need to resolve the issues of the relative contributions of bedload and suspended-load abrasion and of the relative contribution of cavitation damage to the fluting and potholing characteristic of suspended-load abrasion.

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TABLE 1. FIELD SITE CHARACTERISTICS AND CRITICAL JOINT-SPACING DATA

Location	Lithology	Drainage area (km^2)	Bankfull discharge ($m^3 s^{-1}$)	Average gradient	Joint spacing*	
					Plucked reaches (m)	Abraded reaches (m)
Sierra Nevada, California						
Indian Creek	Granitic; metamorphic	1914	110	0.009	0.1–0.5	1–1.5
Middle Fork Stanislaus River	Granitic	123	25	0.019	N.D.	2–4
North Fork Yuba River	Granitic; metamorphic	647	200	0.018	0.5–1	1.5
Mokelumne River	Granitic	1409	170	0.008	0.3–1.5	1.5–3.0
Spanish Creek	Metamorphic	502	60	0.013	0.5–1	N.D.
South Fork American River	Granitic	190	70	0.033	0.3–0.5	1.5
Valley of Ten Thousand Smokes, Alaska						
Upper Ukak River	Sandstone; welded tuff	300	>100	0.018–0.086	0.1–1	N.D.
Northwest Himalaya, Pakistan						
Indus River sites	Metamorphic; granitic	100 000	4000	0.01	≤2	>3
Kachura	Schist			0.007	N.D.	≥5
Baghicha	Granite			0.012	N.D.	≥5
Mendi	Granite			0.002	1–2	N.D.
Ganji†	Amphibolite			0.2	N.D.	≥5
Subsar	Gneiss			0.02	2 -3	N.D.
Shengus†	Gneiss			0.02	1–2	N.D.
Burumdoir	Gneiss			0.008	N.D.	≥5
Unnamed†	Metasedimentary			N.D.	1–2	N.D.
Hanochal	Schist			0.023	3§	>3
Longmen Shan, China						
Min Jiang	Granitic; metasedimentary	1000–10 000	N.D.	0.010–0.050	≤1–2	>2#
Hei Shui He	Metasedimentary	1000–10 000	N.D.	0.01	≤1–2	>2#

Note: N.D.—no data.

*Reported values are maximum and minimum observed joint spacings in plucking-dominated and abrasion-dominated reaches, respectively.

†Indus River sites not described in Hancock et al (1998).

§Hanochal site was abrasion dominated over all, but areas with joint spacing as low as 3 m showed some signs of plucking activity.

#Rivers draining the Longmen Shan tap mostly well-jointed rocks, and abrasion-dominated sites are rare. Data are from reconnaissance transects along river courses.

Figure 1. Photographs of plucking-dominated reaches cut into well-jointed rock units. (A) Yuba River cutting through jointed (0.5–1 m spacing), high-grade metamorphic rocks of the western Sierra Nevada (see Table 1). Exposed surfaces are smooth, slightly rounded, and polished. Occasional small flutes and potholes were noted in the field; however, rock surfaces are predominantly exhumed joint, fracture, and bedding planes, with only minor volume loss attributable to abrasion. (B) Banks aligned along fracture planes and a partially extracted joint block testify to the efficiency of plucking along this reach of the upper Ukak River, Alaska, which is incised into competent but well-jointed sandstone (Table 1). (C) Imbricated joint blocks in the high-flow channel of the Ukak River. Note that intense frost shattering has pulverized one bedding-plane thickness of the upper surface of these exposed blocks.

Figure 2. Photographs of flutes, ripples, and potholes carved into massive, cohesive rocks. (A) Rhythmic fine flutes and ripples carved into very hard, fine-grained, high-grade metamorphic rocks on the banks of the Indus River near Nanga Parbat, Pakistan (Ganji site, Table 1). Field notebook measures 12 × 19 cm for scale. Annual peak flows of the Indus inundate this site with some 6 m of water. Several large, coalesced potholes have carved deep (2–4 m) potholes immediately downstream of the pictured rock surface. (B) Large coalescing potholes have carved out huge volumes of massive gneiss at this site on the Indus River (Burumdoir site, Table 1). The large potholes at the right foreground are annually inundated by 12–15 m of sediment-laden water. (C) View looking straight down the axis of a small pothole (0.4 m diameter) carved on the flanks of a large pothole (4–5 m diameter) carved into coarse-grained granite on the banks of the Indus River (Baghicha site, Table 1). Standing water in the center of the master pothole can be seen at top center. The "yin-yang" structure (a streamlined double pit) at the base of the pothole is typical of many active potholes, testifies to a tight coupling between fluid flow and erosion patterns, and is inconsistent with the familiar "grindstone" hypothesis of pothole erosion. (D) Potholes, half-potholes, and large flutes carved into unjointed, moderately welded tuff by the River Lethe (tributary to the Ukak) in Alaska's Valley of Ten Thousand Smokes.

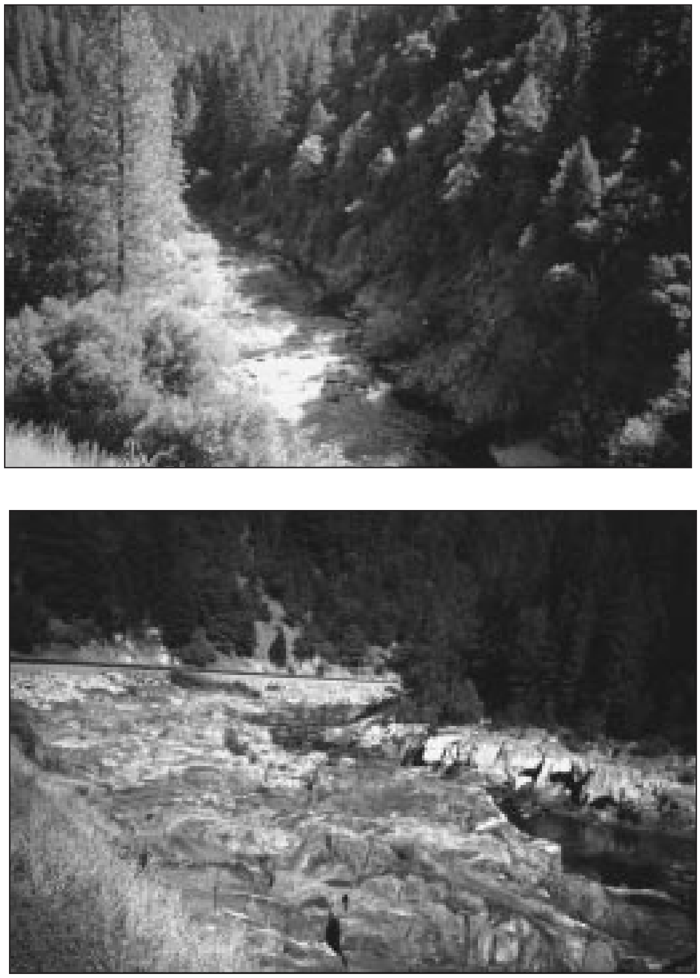


Figure 3. Alteration of plucking-dominated and abrasion-dominated reaches. (A) Boulder cascade in a mixed bedrock-alluvial stretch along Indian Creek in California's Sierra Nevada. Exposed bedrock is visible in the shade, on the far bank. The predominantly metamorphic rocks along this reach are very well jointed, and we infer that erosion is by plucking (Table 1). This reach is arguably transport, rather than detachment, limited. (B) Just upstream of the reach in A, the river has encountered a more coherent, massive rock rib (joint spacing 1–1.5 m, Table 1). A short (entire length visible in photo) strath terrace has formed where the river apparently hung up on this more resistant rock rib. To the right in the photograph an inner channel is visible. Here, and at several other sites along this and other Sierran rivers, the incision of the inner channel appears to have been either accomplished or initiated by pothole erosion, which is germinated at the knickpoint at the downstream end of the strath.

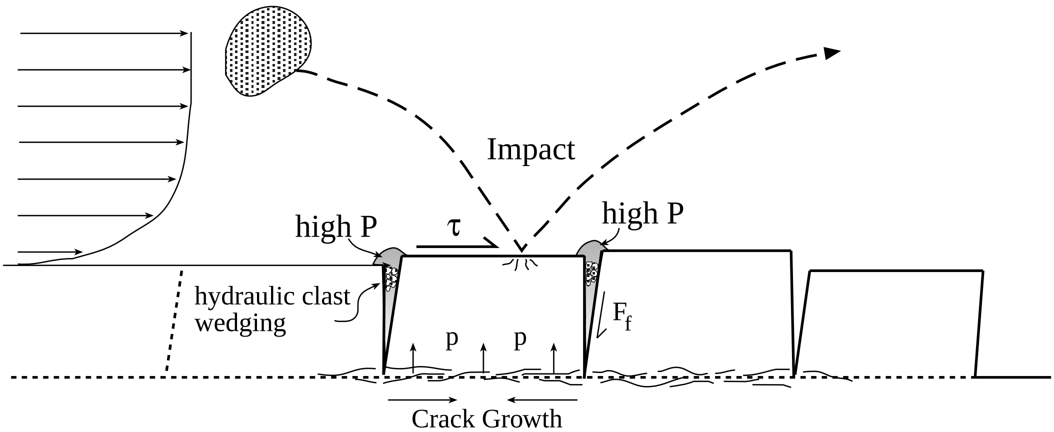


Figure 4. Schematic illustration of the processes and forces contributing to erosion by plucking. Impacts by large saltating grains produce some direct abrasion damage but contribute most importantly to the generation of stresses that drive the crack propagation necessary to loosen joint blocks. Hydraulic clast wedging works to further open cracks. Surface drag forces and differential pressures across the block act to lift loosened blocks. Where the downstream neighbor of a block has previously been removed, both rotation and sliding become possible, and extraction is greatly facilitated.

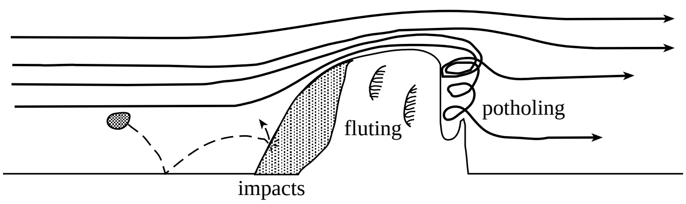


Figure 5. Schematic illustration of the processes contributing to erosion by abrasion. Bedload and the coarsest suspended-load grains are strongly decoupled from the fluid flow and impact the upstream faces of protuberances and obstructions (shaded region). In this zone, surfaces are observed to be smooth and polished, but little abrasion damage appears to occur (see data in Hancock et al., 1998). Fine-scaled flutes and ripples adorn the flanks of massive protuberances where flow separation induces tight stream-wise vortices (see Figs. 2A and 6A). Large, often coalescing potholes characterize the lee side of obstructions, protuberances, and knickpoints (see Figs. 2B and 6B). The complete obliteration of massive, very hard rocks in these potholed zones testifies to the awesome erosive power of the intense vortices shed in the lee of obstructions (see Fig. 2B).

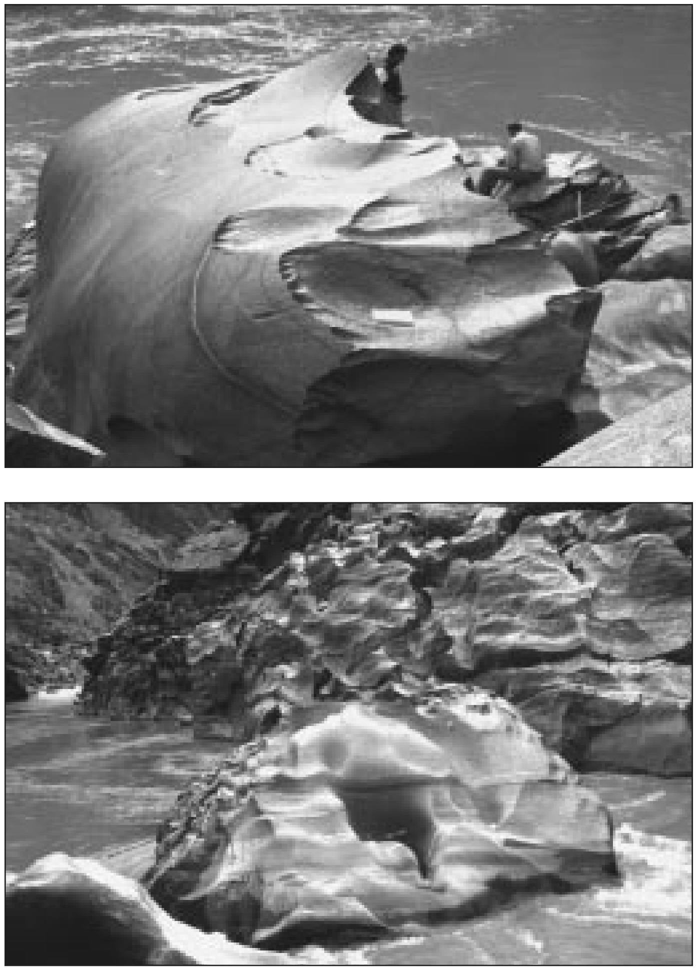
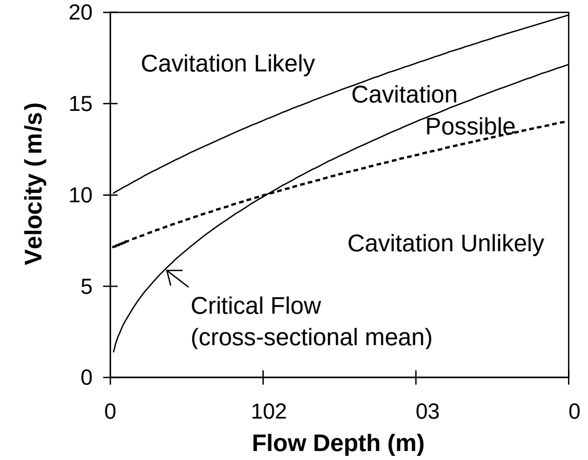


Figure 6. Spatial distribution of flutes and potholes. (A) The boulder that inspired the illustration in Figure 5, with Anderson and Hancock in the background (Mendi site, Table 1). Note lack of obvious damage and subtle varnish coating on the exposed upstream side (this boulder is completely inundated in annual peak flows). Although not obvious from this angle, the downstream face of the boulder is adorned by a series of deep, vertical potholes. Potentially a considerable volume of the originally rounded glacial boulder has been pulverized by potholing. (B) View from downstream of an analogous boulder lying in the thalweg of the Indus River, seen at the extreme low flow of mid-spring (Burumdoir site, Table 1). A prominent deep pothole has drilled into the downstream face, surrounded by a number of incipient potholes. At high flow this boulder is covered by 12–15 m of water. The upstream face shows much less wear, but is adorned by fine-scaled flutes, as in A.

Figure 7. Conditions for cavitation inception. All calculations for atmospheric pressure (sea level) and water temperature of 10 °C. (A) Critical velocities for cavitation as a function of flow depth are calculated for $\sigma_c = 4$ (possible cavitation dashed line) and $\sigma_c = 2$ (likely cavitation thin solid line). Shown for comparison is the condition for critical flow as a function of flow depth (heavy solid line). (B) Required local velocity excess (U_l/U_c) is shown as a function of flow depth for $\sigma_c = 4$ (possible cavitation dashed line) and $\sigma_c = 2$ (likely cavitation solid line).



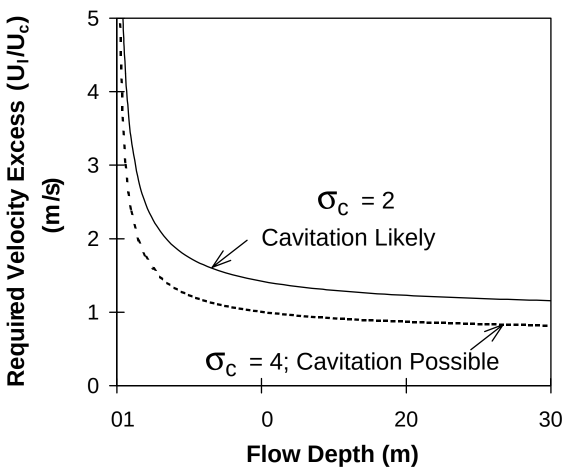


Figure 8. Joint-block plucking and pothole formation at the channel bed. (A) Indus River at low flow. Fins of plucking-resistant rock protrude between eroded joint/fracture planes. Surfaces are smooth, polished, and adorned with meter-scale potholes. (B) The bed of a dry, high-flow channel of the Ukak River, Alaska, is comprised of exhumed joint planes and a series of shallow (1 m), wide (3–5 m) potholes.

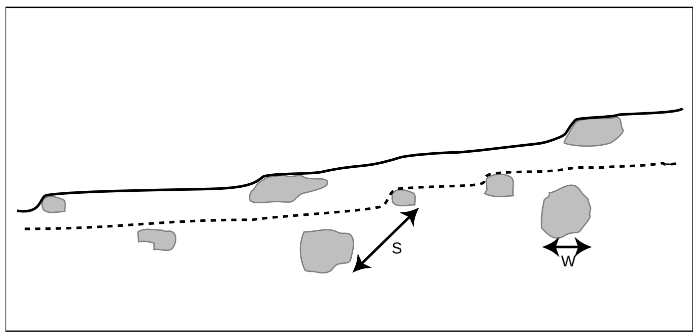


Figure 9. Schematic illustration of the possible influence of scattered resistant rock ribs on the temporal evolution of a bedrock channel. A random distribution of massive rock ribs (gray) with characteristic length-scale *W* and spacing *S* is shown with an initial channel profile (solid line) and a future profile (dashed). Active strath formation and knickpoint migration is occurring where the channel is draped over rock ribs. The channel lowering rate, channel slope, and whether abrasional or plucking processes dominate over the long term probably depend on *W/S*, the rock rib size to spacing ratio.