Climate change, flooding in arid environments, and erosion rates

Peter Molnar*
Department of Geological Sciences and Cooperative Institute for Research in Environmental Sciences (CIRES), University of Colorado, Boulder, Colorado 80309, USA

ABSTRACT

Although regional climates throughout much of the world appear to have become more arid in late Cenozoic time, sedimentation rates, and therefore presumably erosion rates, have increased. For sustained erosion of elevated terrain, at least where glaciers are not the major erosive agent, rivers must incise. Therefore bed-load transport by rivers should be a rate-limiting process in erosion. Theories of bed-load transport call for a threshold in either stream power per unit area or shear stress before the bed load can be moved, and most transport should be accomplished during high discharge. The frequency-magnitude distribution of floods shows that the ratio of magnitudes of, for example, 100 yr floods and annual floods is greater in arid than in humid environments. Thus, a shift toward more arid conditions may have increased relative magnitudes of rare floods or, conversely, increased the frequency of large floods. Such a shift, despite a decrease in precipitation and discharge, could have doubled incision rates, particularly in regions already quite arid.

Keywords: floods, erosion, climate change, aridity.

INTRODUCTION

Both global compilations (e.g., Davies et al., 1977; Donnelly, 1982; Hay et al., 1988) and studies of specific regions (e.g., Guillaume and Guillaume, 1982; Hay et al., 1989; Métivier et al., 1999) show increases in sediment accumulation during the past few million years and especially since 2.5–3 Ma. Astronomical revisions to the geomagnetic polarity time scale rule out globally synchronous changes in plate motion (Krijgsman et al., 1999); thus, despite repeated claims of late Cenozoic uplift of mountain ranges throughout the world, global tectonics cannot account for these changes in sedimentation rates (e.g., Molnar and England, 1990; Zhang et al., 2001). Concurrent with that increased sedimentation, global temperatures decreased and Earth became more arid (e.g., Ruddiman et al., 1989; Crowley and North, 1991, p. 196–197). The lack of any other obvious concurrent global process poses the question, How has climate change effected such a change, especially while climates have become more arid? Glaciers seem to erode more rapidly than rivers under otherwise similar conditions (Hallet et al., 1996), but not all regions supplying increased sediment were glaciated (e.g., Zhang et al., 2001). Here I consider how climate change in arid environments, typical of many high, rapidly eroding environments, might have increased erosion.

Not surprisingly, compilations of sediment transported by rivers show that more sediment is carried when rivers are in flood stage than when discharges are low (e.g., Leopold and Maddock, 1953; Leopold et al., 1964). An analysis including the frequency-magnitude distribution of discharge, however, showed that rare large floods carry only a small fraction of the annual sediment transport (Wolman and Miller, 1960). This fact has led to the view that although the rare large storm could temporarily alter the landscape, more frequent (e.g., biannual) storms transport more sediment and, therefore, shape the landscape most (e.g., Wolman and Gerson, 1978).

Most sediment carried by rivers is suspended, but most of that sediment presumably is derived from hill slopes adjacent to rivers. For a river to incise the landscape, it must transport its bed load. Analyses of digital elevation maps suggest that hill slopes reach limiting values (Burbank et al., 1996). Thus, a rate-limiting process for erosion is the rate of bed-load transport. As discussed in the following, most theories of bed-load transport include a threshold below which negligible bed load is moved. Therefore, bed-load transport should depend on the distribution of large floods.

If the rate of valley incision dictates the erosion rate of a region, the most important rate-limiting process should be bedrock erosion. There are the three obvious mechanisms for bedrock erosion: plucking of material by pressure differences within the flow, abrasion by sediment carried in the water column, and cavitation, the reduction in pressure within a flow to below the vapor pressure (e.g., Hancock et al., 1998; Whipple et al., 2000). Among these, only the first moves cobbles and boulders; the other two scour rock surfaces. Moreover, measurements of scouring from the Indus River, which has cut a deep gorge and seems to erode rapidly, suggest much lower rates than have prevailed over geologic time (Hancock et al., 1998). Ron Shreve (1999, personal commun.) pointed out that in many (though not all) regions, bedrock crops out only sparsely in river and stream valleys, so that the rivers are not in contact with bedrock. He suggested that bed-load transport may limit bedrock erosion in such regions. In his view, other processes such as weathering, jointing, and other fracturing (e.g., Miller and Dunne, 1996) loosen material so that rivers can transport it as bed load. As the bed load is moved, particle sizes will diminish, and some can be carried as suspended material, as analyzed in detail by Sklar and Dietrich (1998). Regardless, consistent with Shreve's view, bed-load transport would be the rate-limiting process of incision, and bed-load erosion as a process whereby rivers and their tools remove intact rock from their beds might be a minor process, at least in regions of high relief. It follows that if climate change has increased erosion rates, it has also increased rates of bed-load transport.

FLOOD, OR DISCHARGE, FREQUENCY

Following Turcotte and Greene (1993), suppose that flood frequency obeys a cumulative frequency-magnitude equation of the form

*E-mail: molnar@terra.colorado.edu.

TABLE 1. VALUES OF α

River	α*
Arroyo Seco, California	1.10
Llamo, Texas	1.39
Mora, New Mexico	1.59
Humboldt, Nevada	1.62
Elkhorn, Nevada	1.85
Oconee, Georgia	1.85
Little Missouri, Wyoming	1.92
Mississippi, Minnesota	2.13
Westfield, Massachusetts	2.30
Wenatchee, Washington	3.23

Note: Data from Turcotte and Greene (1993).
*α is the dimensionless exponent defined in equation 1.

$$N(F) = CF^{-\alpha}, \tag{1}$$

where F is the magnitude of a flood at its peak, and N is the number of floods with magnitudes greater than F per unit time. N has units of time^{-1}, C has units of (flood size)$^{\alpha}$ time^{-1}. Turcotte and Greene (1993) showed that flood statistics compiled by Benson (1968) for several regions fit such a distribution law. They obtained values of the exponent α as small as 1.1 and as large as 3.1 (Table 1). Obviously, small α means that large floods are more frequent than for large α, when compared to the same frequency of small floods. (Note that for hurricanes, $\alpha = -1$, which implies that hurricanes of all sizes between the minimum storm designated a hurricane and the maximum are equally common [Emanuel, 2000].) Because peak discharge provides a sensible measure of flood magnitude, we can rewrite equation 2 by substituting in discharge, Q for F.

Necessary for estimating the amount of bed-load transport is the frequency or probability distribution for floods, $n(Q)$, where $n(Q)dQ$ is the probability of a flood with a magnitude in the range between Q and $Q + dQ$. Using equation 1,

$$n(Q) = -\frac{dN}{dQ} = -\alpha C Q^{-(1+\alpha)} = Q_{\mathrm{ref}}^{\alpha} Q^{-(1+\alpha)}, \tag{2}$$

where the constant of integration is $C = Q_{\mathrm{ref}}^{\alpha}/\alpha$, and Q_{ref} is a reference discharge. Thus, we assume that discharge also obeys a power-law frequency-magnitude distribution when a river is in a flood state. This amounts to treating durations of all floods as the same, but it seems likely that large floods will last longer than small ones, which would make them even more effective as erosive agents than I have assumed. The analysis would be jeopardized only if the integrated discharge of individual small floods exceeded that of large ones, which seems unlikely enough to allow us to ignore it here.

Because there must be an upper limit to discharge in a flood, if only because there is a finite amount of water on Earth, $n(Q)$ as given by equation 3 must be bounded at the top by a maximum discharge, $Q_{\max}$. There also must be a minimum discharge for a flood to be sensibly considered a flood. For $Q \leq Q_{\min}$, flow would be just part of an uneventful background.

DEPENDENCE OF EROSION ON DISCHARGE

Two aspects of a river contribute to erosion and sediment transport. First, a shear exerted by the flow on the bed load must exceed the frictional resisting stress for movement of sediment. Thus, there must be a threshold below which the basal shear cannot move the bed load. Second, clearly the larger the discharge, the greater the stream power, and the more sediment a river can transport. Because shear stress and discharge (and stream power) are related to one another, this has led to formulae of the form (e.g., Knighton, 1984)

$$Q_s \propto W(\tau_b - \tau_c)^{\beta}, \qquad \tau_b \geq \tau_c, \tag{3}$$

$$Q_s \propto W(\omega - \omega_c)^{\beta}, \qquad \omega \geq \omega_c, \tag{4}$$

where Q_s is the volume rate of sediment transport, W is the width of the stream, τ_b is the basal shear stress, τ_c is a critical (or threshold) basal shear stress for sediment to be moved, $\omega = \rho g QS/W$ is stream power per unit area, ρ is density, g is gravity, S is the stream gradient, ω_c is the critical stream power per unit width necessary to transport sediment, and β is an empirical exponent.

Equations like 3 and 4 seem to derive from arguments and equations of Bagnold (1973, 1977, 1980). Bagnold (1973) assumed that bed load transport could be written in terms of the product of a shear stress at the bottom necessary to move the bed load and a speed of flow near the bottom. Such a product has the form of a power per unit area. Thus, he assumed that the bed-load transport rate should depend on the stream power linearly, but not proportionally, because only a fraction of the stream power per unit width of a river could transport the bed load; most of the stream power per unit width would be used merely to maintain flow of the water and suspended sediment.

Although Bagnold (1973) presented data that supported his theory, he promptly modified it. Using mostly laboratory data, but some field observations, Bagnold (1977) showed that plots of $\log Q_s$ versus $\log(\omega - \omega_c)$ defined straight lines with a slope of 3/2, corresponding to

$$\frac{Q_s}{W} \propto (\omega - \omega_c)^{3/2}. \tag{5}$$

The proportionality constant, which Bagnold (1980) modified again, depends on the ratio of water depth to particle size at relatively low, but not high, discharge. I ignore it here. With the predilection of hydrologists and geomorphologists for power-law relationships, this seems to have evolved into equations 3 or 4. Subsequently, others have derived relationships between bed-load transport and shear stress or stream power from a similar logical basis (see Table 1 of Bridge and Dominic, 1984). Bridge and Dominic (1984) derived a form:

$$\frac{Q_s}{W} \propto (\tau - \tau_c)(u^* - u^*_{\mathrm{crit}}). \tag{6}$$

Essentially all theories of bed-load transport assume that there is a threshold in shear stress, stream power, or mean flow speed below which no erosion will occur. In considering bed-load transport and stream incision, this assumption makes the definition of Q_{ref} in equation 2 obvious: $Q_{\mathrm{ref}} = Q_c$, where Q_c is the threshold discharge below which bed load is not transported.

EFFECT OF FLOOD FREQUENCY ON CALCULATED BED-LOAD TRANSPORT

My interest is not in deriving an equation for predicting erosion or sediment transport for a myriad of conditions, but merely to examine how changes in the frequency-magnitude distribution of discharge, which appears to depend on climate, might affect erosion rates. I seek an understanding of how the annual bed-load transport rate $\overline{Q}_s$ depends on the frequency of floods. When equation 2 is used to characterize the frequency of large discharges above a threshold and expressions such as equations 3–6 to describe how bed-load transport Q_s depends on discharge, the long-term average annual bed-load transport rate becomes

$$\bar{Q}_s = \int_{Q_c}^{Q_{max}} n(Q)Q_s(Q)\, dQ. \tag{7}$$

With $n(Q)$ defined by equation 2, some simplicity is gained by rendering equation 7 dimensionless using $Q' = Q/Q_c$, $Q'_s = Q_s/Q_c$, $\bar{Q}'_s = \bar{Q}_s/Q_c$, and $Q'_{max} = Q_{max}/Q_c$:

$$\bar{Q}'_s = \int_1^{Q'_{max}} Q'^{-(1-\alpha)} Q'_s(Q')\, dQ'. \tag{8}$$

Let's first consider a particularly simple form of equation 3, which Knighton (1984, p. 73) called the Schoklitsch type:

$$Q_s \propto Q - Q_c. \tag{9}$$

Nondimensionalizing this, inserting it into equation 8, and integrating gives

$$\bar{Q}'_s \propto \left[\frac{1 - Q'^{1-\alpha}_{max}}{\alpha - 1} - \frac{1 - Q'^{\alpha}_{max}}{\alpha}\right]. \tag{10}$$

The form of equation 10 illustrates the key features of all of the forms assumed for Q_s. First, the value of $\bar{Q}'_s$ depends strongly on α, especially as α approaches 1, $\bar{Q}'_s$ decreasing as α increases (Fig. 1A). Second and perhaps more important, $\bar{Q}'_s$ is virtually independent of Q'_{max} for $\alpha > 1.5$. For large α, big floods are not important, whereas for small α, they are. Figure 1 shows $\bar{Q}'_s$ for two values of Q'_{max}, 100 and 1000. Benson's (1968) 50-year compilation showed values of $Q'_{max} \approx 20$; extrapolation of equation 2 or 3 to thousands of years makes $Q'_{max} \approx 100$ virtually certain and $Q'_{max} \approx 1000$ plausible for the Holocene period.

To examine equation 8, using equation 5 (Bagnold, 1977), first assume that $\omega_c = \rho g Q_c S/W$. Because this analysis considers the effect of different climates on $\bar{Q}'_s$ at a site, we may assume that S is a constant. This assumption for ω_c also minimizes the dependence of Q_s on W in equation 5. So, let's ignore both the coefficients, ρ, g, and S, and any dependence of W on Q. Numerical integration of equation 11 with equation 8, nondimensionalized as before, relates $\bar{Q}'_s$ to Q' (Fig. 1B). As for the Schoklitsch form, the dimensionless erosion rate depends strongly on α and increasingly so as α approaches 1. In addition, more than for the Schoklitsch form, the erosion rate depends on the magnitude of the largest discharge Q'_{max}.

Using a standard relationship between stream power and discharge, and assuming that the width obeyed a relationship like $W \propto Q^n$, Tucker and Slingerland (1997) assigned $n = 1/2$ and converted equation 6 into

$$Q_s \propto Q^{1/2}(Q^{1/3} - Q_c^{1/3})(Q^{1/6} - Q_c^{1/6}). \tag{11}$$

Again, the result of nondimensionalizing equation 11, inserting it into equation 8, and integrating depends strongly on α and decreases as α increases (Fig. 1C). Unlike Bagnold's form (Fig. 1B), the dimensionless erosion rate is nearly independent of Q'_{max} for values greater than 100.

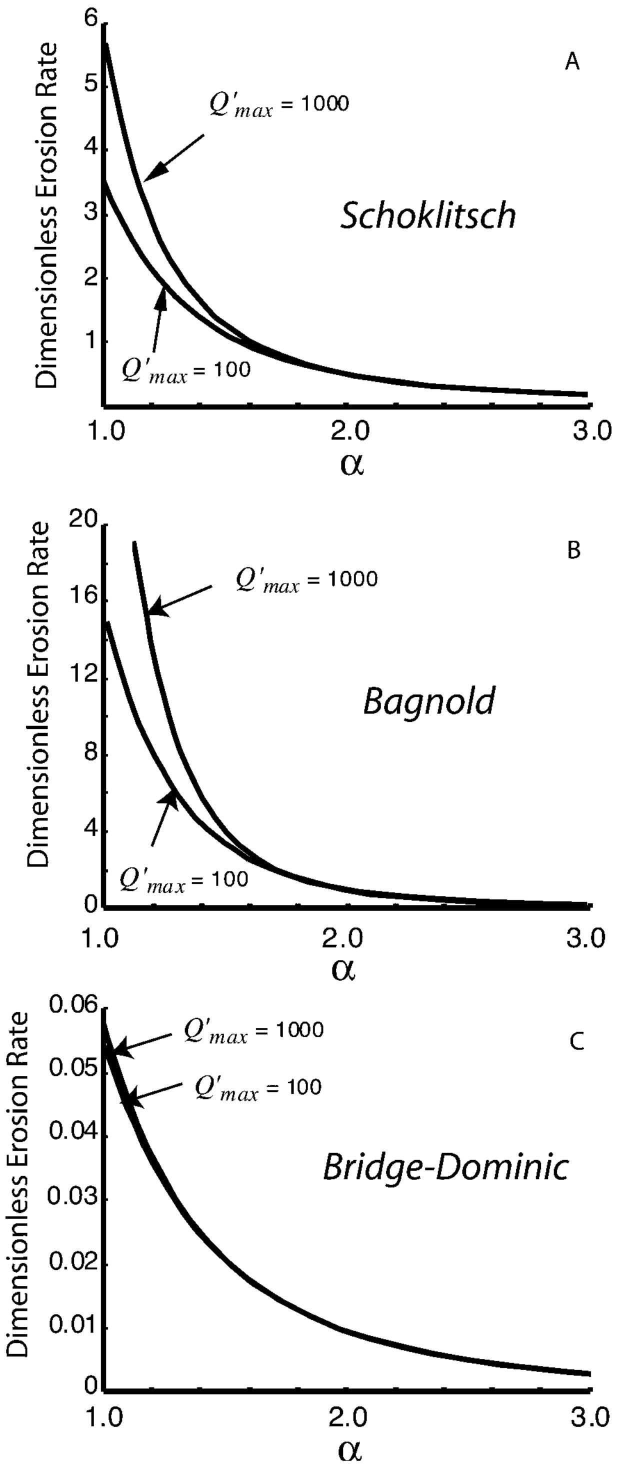

Figure 1. Plots of dimensionless annual bed-load transport, $\bar{Q}'_s$, for three laws relating bed-load transport to stream power or basal shear stress, and for two different ratios of maximum flood discharge to threshold discharge, Q'_{max}, necessary to move bed load. A: Schoklitsch form (Knighton, 1984), equation 9 here. B: Bagnold form (1977), equation 5 here. C: Bridge and Dominic (1984) form, equation 6 here.

EFFECT OF CLIMATE CHANGE ON BED-LOAD TRANSPORT RATES

The analysis described herein highlights the importance of the frequency-magnitude distribution of discharge during floods on bed-load transport. As α decreases, bed-load transport and river incision increase. Turcotte and Greene (1993) showed that the value of α correlated qualitatively with aridity, more arid environments being characterized by smaller values of α. A similar relationship can be seen in other studies, if they examined the probability distribution of discharge (or flood magnitude) differently. Pitlick (1994) showed that for drainage basins in different climatic zones but similar in physiography to the western United States, the ratios of magnitudes of the 100 yr to the 10 yr floods can vary from as little as two to as much as ten. The more arid regions undergo the larger range of flood magnitudes. He found a similar pattern for separate hydrologic regions of the upper Mississippi basin (Pitlick, 1997); larger ratios of magnitudes of rare floods to annual floods characterize the more arid regions.

These correlations suggest that the increased aridity associated with global climate change over the past few million years would lead to a different flood frequency distribution: more large-magnitude floods per small-magnitude flood. Such a change would, in turn, lead to increased bed-load transport, which would accelerate incision. If hillsides maintained steep slopes (Burbank et al., 1996) as valley floors deepened, then not just incision but also erosion and sedimentation would increase. Thus, a climate change toward increased aridity could accelerate erosion rates despite the decreased discharge in rivers.

ACKNOWLEDGMENTS

This work grew out of attempts to understand work by Greg Tucker and his encouragement to pursue a slightly different path. I thank also Ron Shreve for opening my eyes to the possibility that bedrock erosion sensu stricto is a rare process; Bernard Hallet for making meetings with Shreve frequent and stimulating; John Pitlick for guidance; and N. Hovius and C.S. Riebe for thorough and constructive reviews. Riebe awakened me to assumptions that I had made implicitly.

REFERENCES CITED

Bagnold, R.A., 1973, The nature of saltation and of bed-load transport in water: Royal Society of London Proceedings, ser. A, v. 332, p. 473–504.

Bagnold, R.A., 1977, Bed-load transport by natural rivers: Water Resources Research, v. 13, p. 303–312.

Bagnold, R.A., 1980, An empirical correlation of bed-load transport in flumes and natural rivers: Royal Society of London Proceedings, ser. A, v. 372, p. 453–473.

Benson, M.A., 1968, Uniform flood-frequency estimating methods for federal agencies: Water Resources Research, v. 4, p. 891–908.

Bridge, J.S., and Dominic, D.F., 1984, Bed load grain velocities and sediment transport rates: Water Resources Research, v. 20, p. 476–490.

Burbank, D.W., Leland, J., Fielding, E., Anderson, R.S., Brozovic, N., Reid, M.R., and Duncan, C., 1996, Bedrock incision, rock uplift, and threshold hillslopes in the northwestern Himalayas: Nature, v. 379, p. 505–510.

Crowley, T.J., and North, G.R., 1991, Paleoclimatology: Oxford, UK, Oxford University Press, 339 p.

Davies, T.A., Hay, W.W., Southam, J.R., and Worsley, T.R., 1977, Estimates of Cenozoic oceanic sedimentation rates: Science, v. 197, p. 53–55.

Donnelly, T.W., 1982, Worldwide continental denudation and climatic deterioration during the late Tertiary: Evidence from deep-sea sediments: Geology, v. 10, p. 451–454.

Emanuel, K., 2000, A statistical analysis of tropical cyclone intensity: Monthly Weather Review, v. 128, p. 1139–1152.

Guillaume, A., and Guillaume, S., 1982, L'érosion dans les Alpes au Plio-Quaternaire et au Miocène: Eclogae Geologicae Helvetiae, v. 75, p. 247–268.

Hallet, B., Hunter, L., and Bogen, J., 1996, Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications: Global and Planetary Change, v. 12, p. 213–235.

Hancock, G.S., Anderson, R.S., and Whipple, K.X., 1998, Beyond power: Bedrock river incision process and form, *in* Tinkler, K., and Wohl, E.E., eds., Rivers over rock: Fluvial processes in bedrock channels: American Geophysical Union Geophysical Monograph 107, p. 35–60.

Hay, W.W., Sloan, J.L., and Wold, C.N., 1988, Mass/age distribution and composition of sediments on the ocean floor and the global rate of sediment subduction: Journal of Geophysical Research, v. 93, p. 14 933–14 940.

Hay, W.W., Shaw, C.A., and Wold, C.N., 1989, Mass-balanced paleogeographic reconstructions: Geologische Rundschau, v. 78, p. 207–242.

Knighton, D., 1984, Fluvial process and form: London, Edward Arnold, 218 p.

Krijgsman, W., Hilgen, F.J., Raffi, I., Sierro, F.J., and Wilson, D.S., 1999, Chronology, causes and progression of the Messinian salinity crisis: Nature, v. 400, p. 652–655.

Leopold, L.B., and Maddock, T., 1953, The hydraulic geometry of stream channels and some physiographic implications: U.S. Geological Survey Professional Paper 252, 57 p.

Leopold, L.B., Wolman, M.G., and Miller, J.P., 1964, Fluvial processes in geomorphology: San Francisco, W.H. Freeman and Company, 522 p.

Métivier, F., Gaudemer, Y., Tapponnier, P., and Klein, M., 1999, Mass accumulation rates in Asia during the Cenozoic: Geophysical Journal International, v. 137, p. 280–318.

Miller, D.J., and Dunne, T., 1996, Topographic perturbations of regional stresses and consequent bedrock fracturing: Journal of Geophysical Research, v. 101, p. 25 523–25 536.

Molnar, P., and England, P., 1990, Late Cenozoic uplift of mountain ranges and global climatic change: Chicken or egg?: Nature, v. 346, p. 29–34.

Pitlick, J., 1994, Relation between peak flows, precipitation, and physiography for five mountainous regions in the western USA: Journal of Hydrology, v. 158, p. 219–240.

Pitlick, J., 1997, A regional perspective of the hydrology of the 1993 Mississippi River Basin flood: Association of American Geographers Annals, v. 87, p. 135–151.

Ruddiman, W.F., Prell, W.L., and Raymo, M.E., 1989, Late Cenozoic uplift in southern Asia and the American West: Rationale for general circulation modeling experiments: Journal of Geophysical Research, v. 94, p. 18 379–18 391.

Sklar, L., and Dietrich, W.E., 1998, River longitudinal profiles and bedrock incision models: Stream power and the influence of sediment supply, *in* Tinkler, K., and Wohl, E.E., eds., Rivers over rock: Fluvial processes in bedrock channels: American Geophysical Union Geophysical Monograph 107, p. 237–260.

Tucker, G.E., and Slingerland, R., 1997, Drainage responses to climate change: Water Resources Research, v. 33, p. 2031–2047.

Turcotte, D.L., and Greene, L., 1993, A scale-invariant approach to flood-frequency analysis: Stochastic Hydrology and Hydraulics, v. 7, p. 33–40.

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

Wolman, M.G., and Gerson, R., 1978, Relative scales of time and effectiveness of climate in watershed geomorphology: Earth Surface Processes and Landforms, v. 3, p. 189–208.

Wolman, M.G., and Miller, J.P., 1960, Magnitude and frequency of forces in geomorphic processes: Journal of Geology, v. 68, p. 54–74.

Zhang Peizhen, Molnar, P., and Downs, W.R., 2001, Increased sedimentation rates and grain sizes 2–4 Myr ago due to the influence of climate change on erosion rates: Nature, v. 410, p. 891–897.

Manuscript received April 3, 2001
Revised manuscript received June 15, 2001
Manuscript accepted July 9, 2001

Printed in USA