Time constant for equilibration of erosion with tectonic uplift

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ABSTRACT

Following a change in the rate of convergence in a collision zone, the rate of rock uplift also changes. This results in a change in relief, and hence in erosion rate. Thus, over time, the erosion rate tends to adjust to equal the rock-uplift rate. A perturbation analysis of a mountain mass at critical taper suggests that the time constant for this adjustment varies from 10^5 to $>10^8$ yr, depending on initial relief, lithology, and erosion law. Under most conditions, time constants for the adjustment are long enough to preclude attainment of a steady state during a typical orogeny lasting on the order of 10^7 yr. However, if erosion increases exponentially with relief, a close approach to a steady state is possible in the relatively few mountain belts on Earth with exceptional relief and/or weak lithologies.

Keywords: mountains, erosion, response time, tectonic geomorphology.

INTRODUCTION

As the height of a mountain range increases, stream channels and hillslopes become steeper. In addition, there may be an orographic increase in precipitation, a decrease in vegetation, and initiation of glaciation (Molnar and England, 1990; Pinter and Brandon, 1997). All of these factors are likely to increase the erosion rate. Thus, an increase in the rate of rock uplift that leads to an increase in height of a range usually results in a compensating increase in erosion rate. There is, therefore, a tendency toward attainment of a balance or steady state between the rate of rock uplift and the rate of erosion. Hack (1960) was one of the first to espouse this concept.

This steady state is most likely approached asymptotically because forces tending toward the steady state decrease as the steady-state condition is approached. Together with temporal changes in both climate and uplift rate, this means that a steady state is never actually attained. Despite this, it would be useful in studies of mountain geomorphology if we could define the conditions under which a steady state is approached closely. It will be approached more closely if the time scale for adjustment is relatively short compared with that of changes in rates of uplift or erosion (Whipple, 2001). Herein I explore this question.

Pinet and Souriau (1988) provided a simple example of a system that asymptotically approaches a steady state; i.e., that of a landmass with relief, R, defined as the mean height of the landmass above sea level, undergoing denudation at a rate, $\dot{E}$. As erosion progresses, $R \rightarrow 0$. By definition, in the absence of isostatic or tectonic adjustments, $\dot{E} = -\partial R/\partial t$. If the relation between $\dot{E}$ and R is linear, as in $\dot{E} = \text{k}R$, we see that $R = R_0\, e^{-\text{k}t}$, where R_0 is the relief at time $t = 0$. As t becomes large, R decreases, thus decreasing $\dot{E}$. R asymptotically approaches 0, but never truly reaches it; thus, the time required to attain the steady-state value, 0, cannot be used as a measure of the response time of the system. Rather, the accepted measure, and that adopted here, is the time constant, $t_c = 1/\text{k}$.

PREVIOUS WORK

Time constants associated with the equilibration between rates of erosion and uplift have been discussed in various contexts. Using a numerical model of an evolving mountain range, for example, Kooi and Beaumont (1996) calculated the ratio of mass output by erosion to mass input due to tectonic processes and, for a range 50 km across, found that $t_c \approx 8$ m.y. Adding partial isostatic compensation increased this to ~11 m.y. Increasing the width of the range to 500 km increased it to ~130 m.y. Variations with precipitation and erodibility were also discussed. Similarly, Pinet and Souriau (1988) calculated t_c for denudation in the absence of isostatic or tectonic adjustments. In orogenic belts that were younger than 250 Ma, they obtained $t_c \approx 2.5$ m.y., whereas in older belts the value was ~25 m.y. In a reinterpretation of Pinet and Souriau's data, incorporating new data and isostatic compensation, Pazzaglia and Brandon (1996, p. 265) obtained a global average of ~69 m.y.

Taking a different approach, Whipple and Tucker (1999) calculated the time for the profile of a channel to adjust to a tectonic perturbation through migration of either a knickpoint or a wave of erosion from the range front to the tips of headwater channels. This profile adjustment time, t_{pa}, applies only to the bedrock channel. For watersheds in Taiwan with areas of 10^2–10^3 km^2, Whipple (2001) found that t_{pa} was on the order of 1 m.y. However, for 3–20 km^2 basins in the Mendocino triple junction area, Snyder et al. (2000) obtained $t_{pa} \approx 0.1$ m.y. They attributed the short response time to the weakness of the bedrock, an orographic increase in precipitation, and positive feedbacks in the channel erosion process. Isostatic adjustments and lags due to the response of hillslopes to channel incision are not included in these analyses (Whipple and Tucker, 1999, p. 17,668). Isostatic adjustment can lengthen t_c by a factor of 5 to 6 (Baldwin et al., 2003), and lags due to hillslope adjustment are largely responsible for the asymptotic character of the approach to a new steady state. These lags may be shorter in watersheds dominated by steep slopes that fail by sliding.

The Whipple and Tucker (1999) models approximate conditions in a horst or on the uplifted flank of a tilted block-faulted range. They are less applicable to mountain ranges in convergent settings where significant vertical strain occurs far from the range boundaries (e.g., see Dahlen and Suppe, 1988; Beaumont et al., 1994).

ANALYSIS

Consider a mountain range of width W formed at a convergent margin (Fig. 1). The range has a bulk density ρ_c. It is underlain by lithospheric mantle of density $\rho_{\ell m}$ and is "floating" in an asthenosphere of density ρ_a. Consider a control volume consisting of a vertical slice of unit thickness in a direction parallel to the axis of the range and extending across its width (Fig. 1). The slice extends from the base of the crust to and above the land surface. Mass (rock) may enter or leave this control volume through the ends, sides, and bottom, either by viscoplastic deformation or as magma. For convenience, we express this flux in terms of a volume of rock per unit time and denote it by Q. Erosion off the top occurs at a spatially averaged rate, $\dot{E}$. Both the width of the control volume, W, and the mean height of the slice of crustal rock in it, $\bar{H}$, may increase, remain constant, or decrease, depending on whether Q is greater than, equal to, or less than $\dot{E}W$. $\dot{E}$ depends upon, among other things, the mean height of the range above sea level, referred to here as the relief, R. Owing to isostasy, R is proportional to $\bar{H}$.

Our objective is to study the time scale over which W and $\bar{H}$, and hence $\dot{E}$, adjust to balance Q, and thus achieve an erosional steady state. In such a system, conservation of mass is expressed by:

$$Q - \dot{E}W = W\frac{\partial \bar{H}}{\partial t} + \bar{H}\frac{\partial W}{\partial t}. \qquad (1)$$

Any difference between Q and $\dot{E}W$ is accommodated by thickening of the crust in the orogen at a rate $\partial \bar{H}/\partial t$, by widening it at a rate $\partial W/\partial t$, or both.

We assume that the shape of the range approximates a wedge with a critical taper such that the Mohr-Coulomb strength is reached at every point in the wedge (Dahlen, 1984). The critical taper is the angle, φ, between the basal décollement and a plane extending from base level to

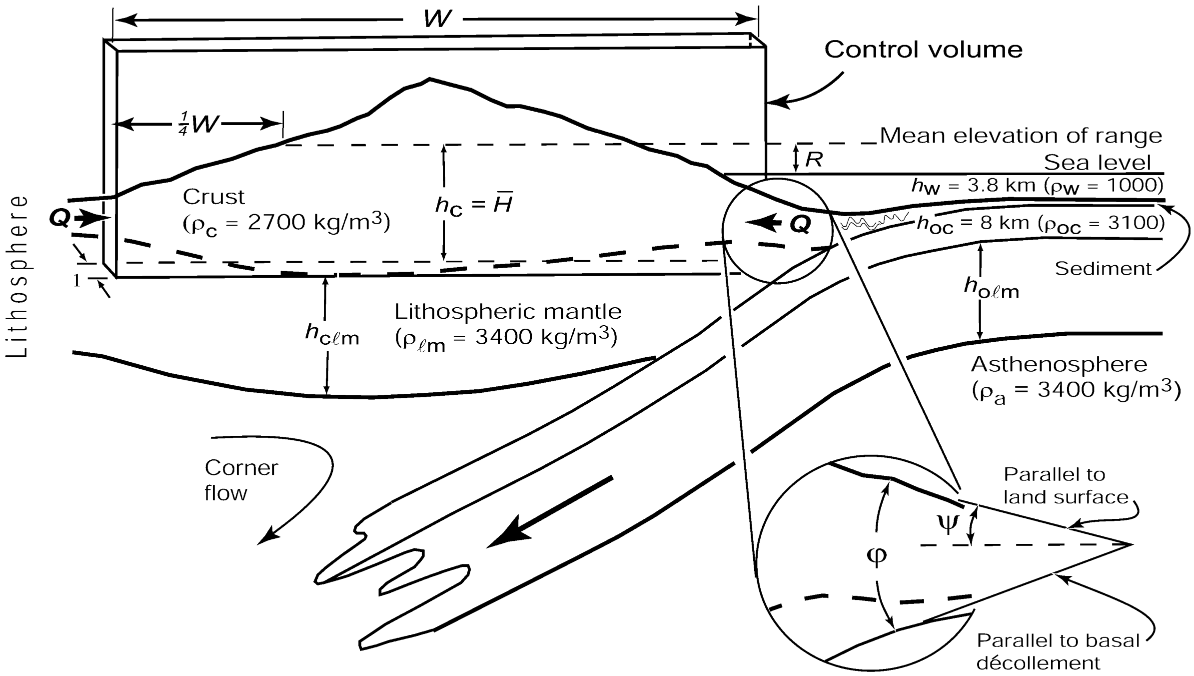

Figure 1. Sketch of mountain range and definitions of some variables used in analysis. Mean ocean depth and mean thickness of oceanic crust are from Skinner and Porter (1995, p. 25, 370) (ρ_w, ρ_{oc} in kg/m³).

the range crest (Fig. 1, inset). The latter plane makes an angle, ψ, with the horizontal; ψ is termed the critical surface slope. Valleys are cut below this plane and ridge crests rise above it. This is a reasonable approximation to the shape of mountain ranges like those of Taiwan (Dahlen and Suppe, 1988).

In such a mountain mass, the spatial distribution of erosion is an independent variable dictated by climate and lithology, and vertical rock velocities at the surface adjust so that they are everywhere equal to the erosion rate. For example, suppose that at some point in the range, streams are able to remove *N* tons of sediment annually. The value of *N* is determined by the stream discharge and gradient. The latter would be controlled by ψ. If hillslopes at this point in the range are not steep enough to deliver *N* tons of sediment to the stream, the stream will downcut, increasing the steepness of the hillslopes. The resulting decrease in stream gradient downstream will temporarily reduce the local mean slope of the range below ψ. This will increase the flux of rock to this point, thus elevating both the trunk channel and the ridge crests and restoring the critical surface slope. Now, however, hillslopes are steeper and thus able to supply the sediment load the stream is capable of carrying. This is a straightforward application of Strahler's Law (Hooke, 2000).

Let us now express Q, $\bar{H}$, $\dot{E}$, and W in terms of their steady-state values (subscript 0) and small deviations from the steady state (subscript 1); thus, $Q = Q_0 + Q_1$, $\bar{H} = \bar{H}_0 + \bar{H}_1$, $\dot{E} = \dot{E}_0 + \dot{E}_1$, and $W = W_0 + W_1$. Substituting these values into equation 1, noting that derivatives of steady-state values with respect to time are 0 by definition, and subtracting the steady-state relation from the resulting equation, ignoring second-order terms, yields:

$$Q_1 - \dot{E}_0 W_1 - \dot{E}_1 W_0 = (W_0 + W_1)\frac{\partial \bar{H}_1}{\partial t} + (\bar{H}_0 + \bar{H}_1)\frac{\partial W_1}{\partial t}. \tag{2}$$

To proceed, we need a relation between $\bar{H}$ and W. As the range is isostatically compensated, we approach this through Airy isostasy, thus:

$$R = h_c\left(1 - \frac{\rho_c}{\rho_a}\right) + (h_{c\ell m} - h_{o\ell m})\left(1 - \frac{\rho_{\ell m}}{\rho_a}\right) - h_w\left(1 - \frac{\rho_w}{\rho_a}\right) - h_{oc}\left(1 - \frac{\rho_{oc}}{\rho_a}\right) \approx \beta h_c + \delta \Delta h_{\ell m} - \gamma, \tag{3}$$

where ρ is density, h is thickness, and the subscripts c, a, o, ℓm, and w refer to crust, asthenosphere, ocean, lithospheric mantle, and water, respectively (Fig. 1). The thin layer of sediment on the ocean floor has been ignored. With the densities and thicknesses shown in Figure 1, $\beta = 0.21$, $\delta = 0$, and $\gamma = 3.4$ km. (Although $\delta = 0$ with the density structure shown, I retain this term for completeness. If it becomes significant, the analysis is much more complicated.) Equating h_c with $\bar{H}$ (Fig. 1) then yields:

$$\bar{H} = \frac{1}{\beta}(R + \gamma), \tag{4}$$

and in terms of the perturbed quantities:

$$R_1 = \beta \bar{H}_1. \tag{5}$$

Note that in an isostatically compensated symmetrical wedge at critical taper, the mean surface elevation (R) is reached at a point one-half of the way from the edge to the apex (Fig. 1), so we have:

$$W = \frac{4R}{\tan\psi} = \frac{4R}{\beta \tan\varphi} = \frac{4R}{\beta\zeta}, \tag{6}$$

where $\zeta = \tan\phi$. From equation 6, $W_1 = 4R_1/\beta\zeta$. Making appropriate substitutions in equation 2 yields:

$$\frac{dR_1}{dt} = \frac{\frac{1}{4}\beta^2\zeta Q_1 - \beta(\dot{E}_0 R_1 + \dot{E}_1 R_0)}{2(R_0 + R_1) + \gamma}. \tag{7}$$

Our goal is to use empirical studies to relate $\dot{E}$ to R and then solve equation 7 for $R_1(t)$. We discuss the empirical studies next.

Erosion Laws

In developing an erosion law for use in the present analysis, we are constrained by the fact that empirical data usually consist of measured spatial differences in sediment yield at an instant in time; in a perturbation analysis, we are dealing with small, temporal changes from an existing state. We use the measured spatial differences to estimate how a temporal change in a particular independent variable is likely to influence a dependent variable. To be specific, we are interested in how a change in R of a watershed will change $\dot{E}$ in that watershed. Thus, in our case, lithology does not change, and changes in climate are small and reasonably well prescribed. In contrast, in the empirical data spatial variations in lithology and climate are likely to be both significant and poorly prescribed, and they may mask or obscure the effects with which we are primarily concerned. In addition, empirical data may include watersheds that are in different stages of adjustment in response to previous perturbations or that have been affected by human activities (K. Whipple, 2002, personal commun.). For all these reasons, we use empirical data only as a guide to the form and approximate magnitude of the effects in which we are interested.

I have argued, theoretically, that $\dot{E}$ should be proportional to the fraction, $\wp$, of the area of a watershed over which hillslopes equal or exceed the angle of repose (Hooke, 2000). Intuitively, it is clear that $\wp$ must increase with the mean slope, $\bar{\alpha}$. If hillslope magnitudes in the watershed are normally distributed, the increase in $\wp$ with $\bar{\alpha}$ is nonlinear (Hooke, 2000). This is consistent with empirical data of Montgomery and Brandon (2002), who also showed that $\bar{\alpha}$ was correlated with the local relief between valleys and ridge crests, R_z. With the exception of plateau areas, R_z is arguably proportional to R as defined herein (Fig. 1). Thus, based on a combination of intuitive physical, theoretical, and empirical grounds, we may expect that $\dot{E}$ is correlated with R. This is particularly true in the present problem, as we are dealing with a perturbation to a particular system.

In tectonically active terrains, particularly those underlain by relatively weak lithologies, hillslope declivity may be limited by the angle of repose. In such situations, $\dot{E}$ becomes independent of $\bar{\alpha}$ (Burbank et al., 1996); it is, instead, related to the frequency of landsliding, and hence to the ability of streams to remove slide debris (Hooke, 2000; Hovius et al., 2000). The latter is controlled by climate and by the slopes of stream channels, which in turn is related to R. Hence, once again we expect $\dot{E}$ to be positively correlated with R.

Various empirical studies have sought to quantify this correlation. Ahnert (1970) and Pinet and Souriau (1988) favored linear relations, whereas Schumm and Hadley (1961) and Summerfield and Hulton (1994) presented data supporting exponential ones. In studies requiring a relation between $\dot{E}$ and R for other purposes, linear erosion laws are commonly used (e.g., Sleep, 1971; Dahlen and Suppe, 1988; Pinot and Souriau, 1988; Pazzaglia and Brandon, 1996; Stüwe and Barr, 1998). However, there are cogent physical reasons for favoring a nonlinear law. Principal among these is the increase, with relief, in the number of angle-of-repose hillslopes that act as sediment sources (Hooke, 2000). Changes in orographic precipitation, vegetation, and glaciation that normally accompany increased relief may also play a role.

For these reasons, we expect that within a given mountain range, temporal changes in R will result in nonlinear changes in $\dot{E}$. However, we need to start with the simpler linear model and use that to examine the nonlinear case.

Linear Model

Equation 7 is expressed in terms of R, so we start with the linear relation:

$$\dot{E}_0 = a + bR_0, \tag{8}$$

or in terms of the perturbations, $\dot{E}_1 = bR_1$. Substituting these relations in equation 7 yields a relation for dR_1/dt that may be integrated to:

$$R_1 = \frac{\beta^2 s Q_1}{4\lambda}(1 - e^{-2R_1/\lambda t_c} e^{-t/t_c}), \tag{9}$$

where

$$\lambda = \beta(a + 2bR_0) \quad \text{and} \tag{10}$$

$$t_c = \frac{1}{\lambda}\left(2R_0 + \gamma - \frac{\beta^2 \zeta Q_1}{2\lambda}\right). \tag{11}$$

Because $\dot{E} = 0$ when $R = 0$, a is clearly 0 in this case. However, we retain it for use later. Because R_1 appears on both sides of equation 9, an iterative solution is necessary. In general, however, $\lambda t_c \gg 2R_1$, so $e^{-2R_1/\lambda t_c} \cong 1.0$.

To obtain numerical results, we need a value for b. In a reinterpretation of Ahnert's (1970) data, Montgomery and Brandon (2002) found that $\dot{E} \cong b_A R_z$ where $b_A = 2 \times 10^{-7}$ yr^{-1}. For purposes of illustration, let's assume $R = \varepsilon R_z$, where ε is expected to be greater than, but of order unity. Here I use $\varepsilon = 2$ for purposes of illustration. Thus, $b \approx 1 \times 10^{-7}$ yr^{-1}.

As an example, consider a mountain range with a mean elevation, R, of 2 km. $\dot{E}_0$ is then 0.2 mm/yr (equation 8). If $\psi = 3°$ and $\varphi = 9°$, as in Taiwan (Dahlen and Suppe, 1988), the orogen will have a subaerial width of ~150 km (equation 6). In the steady state, the erosion, $\dot{E}_0 W_0 =$ 30 m^3/m.y., is balanced by an equal flux, Q_0, into the control volume. Let us choose a perturbation in Q_0 of 2% or 0.6 m^3/m.y. Using $dQ_0 = d\dot{E}W = \dot{E}dW + Wd\dot{E}$ and equations 6 and 8, one can straightforwardly calculate that W and R must increase by ~1500 m and ~20 m, respectively, to absorb this perturbation. However as noted, the approach to the new steady state is asymptotic (Fig. 2). In this example, $t_c \cong 89$ m.y.

Because the increase in R is so small, the increase in $\dot{E}$ is only 0.002 mm/yr. Despite this, 50% of the perturbation in Q is absorbed by this increase in erosion rate ($W_0 d\dot{E}$). The remaining 50% is absorbed by the increase in erosion resulting from the increase in width of the orogen ($\dot{E}_0 dW$). Both $W_0 d\dot{E}$ and $\dot{E}_0 dW$ increase linearly with R, and their ratio remains constant.

Equation 11 may be used to calculate t_c for other values of R (Fig. 3), remembering that Q_0, and hence Q_1, varies with R. Note also that t_c is dependent on W through equation 6. This effect may be amplified in detachment-limited landscapes where the ability of streams to erode their beds increases with discharge downstream and their ability to entrain material is not limited by sediment already in transport.

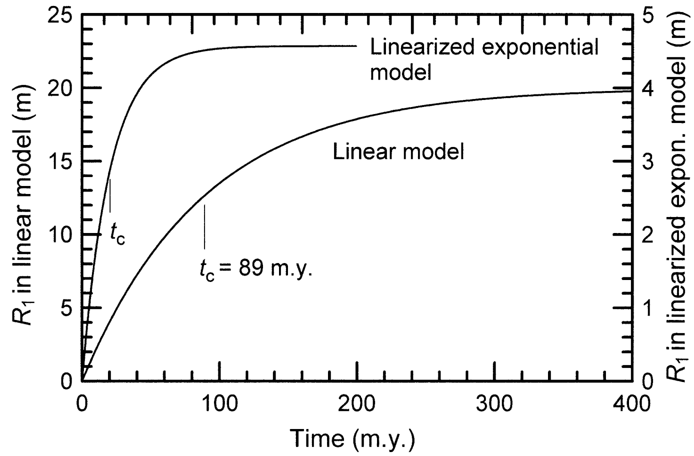

Figure 2. Adjustment of mountain topographic system to perturbation, Q_1, of 2% assuming linear erosion law with constants based on Ahnert (1970) and linearized exponential erosion law with constants based on Summerfield and Hulton (1994). Solution of equation 9 for R_1 for any given time, t, was accomplished by using value of R_1 from previous time step on right side.

Exponential Model

As noted, Summerfield and Hulton's (1994) data can be interpreted as supporting an exponential increase in $\dot{E}$ with their measure of topographic relief, R_S, thus: $\dot{E} = m_S e^{p_S R_S}$. We wish to cast this in the form:

$$\dot{E} = me^{pR}. \tag{12}$$

Assuming once again that $R = 2R_S$ and evaluating m_S and p_S from Summerfield and Hulton's plot, we find $m = 7.59 \times 10^{-6}$ m/yr and $p = 0.0019$ m^{-1}.

Use of equation 12 in equation 7 results in a relation that is not readily integrated. However, for a given $\dot{E}_0$ and small perturbations, we

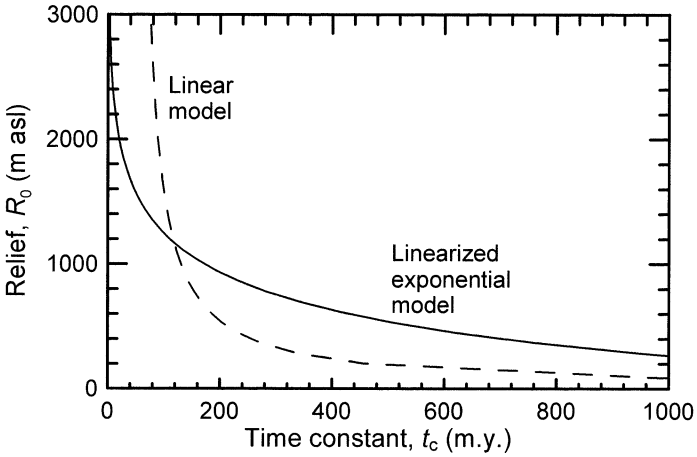

Figure 3. Variation in time constant, t_c, with relief for linear (dashed) and linearized exponential (solid) erosion laws with Q_1 = 2% and constants based on Ahnert (1970) and Summerfield and Hulton (1994), respectively. Result is insensitive to changes in Q_1 (asl—above sea level).

can approximate the exponential relation by a linear one, $\dot{E} = a_e + b_e R$. The slope, b_e, of the linear relation is set equal to the slope of the exponential one at the value of $\dot{E}$ in question: $b_e = mpe^{pR}$. Thus, $a_e = \dot{E} - b_e R$ and λ then may be calculated from equation 10, substituting a_e for a, and b_e for b. Finally, t_c is obtained from equation 11. For our mountain range with $R = 2$ km, t_c is ~20 m.y. (Fig. 2). Figure 3 shows how t_c varies with R_o.

DISCUSSION

Because the time scale for orogenies such as that currently maintaining the height of the Himalayas is of order 10^7 yr, time constants $\leq \sim 10^6$ yr are necessary to maintain a rough balance between rates of erosion and rock uplift. With an exponential erosion law and constants derived from data of Summerfield and Hulton (1994), such time constants appear likely only in areas with relief in excess of 3000 m; with a linear law and constants based on Ahnert (1970), they are not found (Fig. 3). With the longer time constants in areas of more moderate relief that seem likely from this analysis, a significant phase lag will develop between the times of maximum rates of rock uplift and erosion.

Both the Summerfield and Hulton and the Ahnert data sets are from humid temperate areas and relatively inactive or nonorogenic regions. In more erosive climatic zones or in more active orogenic belts, potentially with less well indurated rocks with higher fracture densities, b or p may be higher, and hence t_c lower. This is consistent with both intuition and numerical modeling experiments (e.g., Kooi and Beaumont, 1996). The role of climate is complicated by the fact that changes in precipitation may be offset by changes in vegetation (Langbein and Schumm, 1958). The most important changes in climate are likely to be those that affect the ability of streams to undercut slopes at the angle of repose and to carry away material sloughed from these slopes (Hooke, 2000).

A reasonable conclusion from these calculations is that if erosion rates increase nonlinearly with relief, as seems likely, an erosional steady state may be approached reasonably closely in the relatively few mountainous areas of Earth with mean elevations above 3000 m, and in somewhat lower ranges composed of rocks prone to erosion. In lower areas or areas of more resistant rock, changes in erosion rate are likely to lag changes in rock uplift rate sufficiently to preclude a close approach to a steady state. The lag will increase as relief decreases. These conclusions hold over a range of b and p that is probably sufficient to cover most natural situations.

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