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FLUVIAL LANDSCAPE RESPONSE TIME: HOW PLAUSIBLE IS STEADY-STATE DENUDATION?

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ABSTRACT. Whether or not steady-state topography and denudation are probable states depends on the timescale of system response to tectonic and climatic perturbations relative to the frequency of those perturbations. This paper presents analytical derivations of algebraic relations for the response time of detachment-limited fluvial bedrock channel systems both to tectonic and climatic perturbations. Detachment-limited fluvial erosion is described by the stream-power incision model, and the derivations are limited to the applicability of that model. All factors likely to influence system response time that are not adequately captured by the stream-power incision model will tend to increase the response time. The calculations presented thus provide minimum estimates of landscape response time and therefore over-predict the probability of attaining and sustaining steady-state topography and denudation. The Central Range of Taiwan is used as a case study to estimate response times in a landscape often argued to be in steady state. Model parameters are fit to modern stream profiles by assuming that the topography represents a quasi-steady-state form. Estimated response times generally range from 0.25 to 2.5 Ma, depending on the non-linearity of the incision rule and the magnitude and type of perturbation. Thus it may be reasonably argued that steady-state topography and denudation are likely to prevail during periods of climatic stability (response time is sufficiently short compared with plate tectonic timescales). However, rapid climatic fluctuation in the Quaternary appears to preclude the attainment of steady-state conditions in modern orogens.

MOTIVATION

Steady-state landforms and denudation rates are the natural attractor state under conditions of invariant rock uplift rates (defined relative to a fixed baselevel), climate, and lithology (Adams, 1985; Hovius, Stark, and Allen, 1997; Howard, 1994; Moglen and Bras, 1995; Ohmori, 2000; Willgoose, Bras, and Rodriguez-Iturbe, 1991). Topographic and denudational steady-state is defined as a delicate balance of erosion and (constant) rock uplift such that a statistically invariant topography and constant denudation rate are maintained. As indicated above, for topographic and denudational steady-state to hold, the distribution of rock types exposed at the surface must remain unchanged (that is, there must be no progressive exhumation of more or less resistant rock types). Thus, the topographic and denudational steady-state discussed in this paper implicitly assumes that the orogen has achieved exhumational steady state (see Willett, Slingerland, and Hovius, this issue, p 455). The competing forces of uplift and erosion naturally tend toward attainment of this balance. For example, as initially low-relief landscapes are uplifted, erosion rates steadily increase over time in response to steepening of river profiles and adjacent hillslopes, further enhanced by orographic precipitation. Eventually erosion rates increase sufficiently to counterbalance the rock uplift rate, and a steady-state landscape is achieved (for example, Adams, 1985). An important exception occurs where erosion is so inefficient or rock uplift rate so rapid that the growth of topography reaches limits imposed by crustal strength (Beaumont and others, 1996; Molnar and Lyon-Caen, 1988). Here I will consider only erosionally-limited topographic development.

Because they are a natural attractor state, steady-state denudation and associated steady-state landforms constitute a powerful concept for the exploration of relationships between the height and relief of mountain ranges and such factors as rock uplift rate, climate, and lithology (for example, Howard, 1994; Moglen and Bras, 1995; Whipple and Tucker, 1999a). Similarly, if steady-state or quasi-steady-state conditions can be demonstrated to hold in particular field areas, morphometric analyses of landforms can be used to test or constrain certain aspects of landscape evolution models (Sklar and Dietrich, 1998; Slingerland, 1999; Slingerland, Willett, and Hovius, 1998; Snyder and others, 2000). Denudational and topographic steady state are often assumed in a wide range of studies: the interpretation of thermochronologic data in terms of exhumation rates (Gallagher, Brown, and Johnson, 1998), the effect of topography on observed cooling rates (ridges versus valleys) (Mancktelow and Grasemann, 1997; Stuwe, White, and Brown, 1994), and morphometric studies of landscape form. However, tectonic and climatic forcing is not constant, and it can be very difficult to demonstrate steady-state conditions. A critical question then is: how probable are steady state landscapes? This question fundamentally concerns landscape response times to a perturbation away from steady state and the frequency of those perturbations.

APPROACH AND SCOPE

Simple algebraic relationships are developed that describe to first order the response time of detachment-limited fluvial landscapes in terms of major controlling variables. Only the response time of bedrock channels is considered. Any lag time associated with hillslope response to channel lowering is implicitly assumed to be negligible. Solutions are given for both tectonic and climatic perturbations (step-function changes) away from an initial steady state. The solution for a tectonic perturbation is generalized from that given in Whipple and Tucker (1999a) to allow for (1) either an increase or decrease in rock uplift rate and (2) a contemporaneous change in climate (for example, changing orographic precipitation). The climatic perturbation solution, presented here for the first time, is valid for both increases and decreases in precipitation (or more precisely increases and decreases in the erosivity of the fluvial system). Using drainages in the northern part of the Central Range of Taiwan as a case study, representative landscape response times are then calculated and compared to typical timescales of tectonic and climatic change.

The analysis is subject to several important limitations. First, only detachment-limited fluvial bedrock channel erosion as described by the stream-power incision model is considered. Transitions to glacial erosion and their impact on landscape morphology are not discussed. The potential role of sediment flux in modulating river incision rates and landscape response times is not explicitly treated, although to first order this effect can be folded into a dynamic adjustment of fluvial erosivity commensurate with changes in denudation rate. Further, the analytical solutions presented are predicated on the assumption that migrating knickpoints (defined here as abrupt changes in channel gradient rather than elevation) are maintained as abrupt, discrete features, and therefore no information is conveyed above the migrating knickpoint. Any rounding of knickpoints due to diffusive processes (see Rosenbloom and Anderson, 1994) or less-than-instantaneous changes in boundary conditions will, under some circumstances, result in communication of base-level information ahead of knickpoints, ultimately resulting in considerably longer response times. Thus, because diffusion of knickpoints may occur in nature (Rosenbloom and Anderson, 1994), because tectonic and climatic changes are not instantaneous, and because hillslope response times are often not negligible (Fernandes and Dietrich, 1997), the solutions presented here are minimum estimates of actual response time and will over-estimate the probability that steady-state conditions will be attained in nature. Finally, only block uplifts with spatially uniform uplift rate are considered here. Extension to more complex scenarios can readily be achieved in numerical simulations.

THEORY: RESPONSE TIMESCALE OF DETACHMENT-LIMITED FLUVIAL SYSTEMS

Incision of rivers into bedrock under detachment-limited conditions is often described by the stream-power (or similar shear-stress) model (Howard, 1994; Howard and Kerby, 1983; Howard, Seidl, and Dietrich, 1994; Moglen and Bras, 1995; Rosenbloom and Anderson, 1994; Snyder and others, 2000; Stock and Montgomery, 1999; Tucker and Slingerland, 1996; Whipple, Kirby, and Brocklehurst, 1999). Where river incision is described by the stream-power model, a channel profile evolution equation can be written as:

$$\frac{dz}{dt} = U - KA^mS^n \tag{1}$$

where dz/dt is the rate of change of channel bed elevation, U the rock uplift rate relative to baselevel, A upstream drainage area (a proxy for discharge), S stream gradient, K a dimensional coefficient of erosion reflecting both rock strength and the erosivity of the fluvial system, and m and n are positive constants that reflect the mechanics of the erosion process, basin hydrology, and channel geometry (Howard, Seidl, and Dietrich, 1994; Whipple and Tucker, 1999a). The slope exponent (n) has been argued to depend on the dominant erosion process and to vary between ~2/3 and ~5/3 (Whipple, Hancock, and Anderson, 2000). Because Whipple and Tucker (1999a) have stressed the importance of the slope exponent (n) in determining landscape response time, in the analysis of the Taiwanese landscape I will allow n to take values from 2/3 to 2. Note that depending on the relationship between sediment flux and the coefficient of erosion (see Sklar and Dietrich, 1998; Slingerland, Willett, and Hennessey, 1997; Slingerland, Willett, and Hovius, 1998), K may be a function of both space and time. This potentially important complication is only considered here in a highly simplified manner.

At steady state river incision rate perfectly counterbalances rock uplift rate such that elevations along the channel profile are unchanging in time ($dz/dt = 0$). Under this condition, eq (1) can be solved for the steady-state channel-head elevation ($z(x_c)_{ss}$) (Whipple and Tucker, 1999a):

$$z(x_c)_{ss} = z(L) + \beta U^{1/n}K^{-1/n} \tag{2A}$$

$$\beta = k_a^{-m/n}\left(1 - \frac{hm}{n}\right)^{-1}(L^{1-hm/n} - x_c^{1-hm/n});\ \frac{hm}{n} \neq 1 \tag{2B}$$

$$\beta = k_a^{-m/n} \ln\,(L/x_c);\ \frac{hm}{n} = 1 \tag{2C}$$

$$A = k_ax^h \tag{2D}$$

where x_c is the distance from the divide to the channel head (Montgomery and Dietrich, 1992), L is total streamwise channel length (from the divide), and k_a and h describe the relation between downstream position and upstream drainage area (h is the inverse of the Hack exponent) (Hack, 1957). Because the ratio m/n is determined by the downstream rates of increase in discharge and channel width (Whipple and Tucker, 1999a), β in eq (2A) is entirely a geometric term and will not vary significantly in response either to tectonic or climatic perturbations. Expected variations in

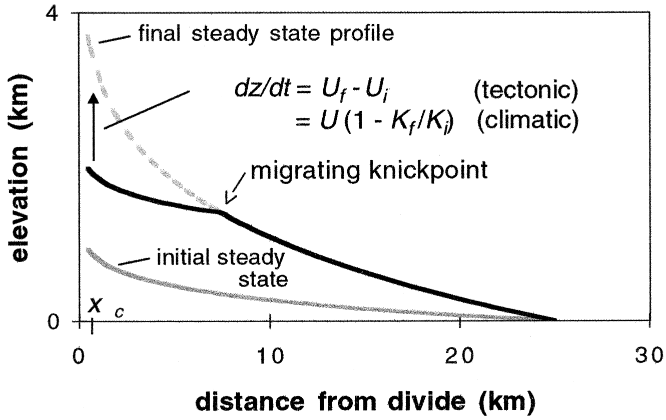

Fig. 1. Definition sketch for derivation of channel response time. A "rising state" transient is shown such as might follow either an increase in rock uplift rate or a decrease in the coefficient of erosion. A migrating knickpoint separates an upstream segment that has been uplifted but not yet steepened from a downstream segment that has been steepened to the new steady state gradient. Because the gradient at the channel head remains unchanged until the knickpoint has swept through the entire system, the rate of channel head elevation change is constant as indicated for either tectonic or climatic perturbations. Figure 1 is modified from Whipple and Tucker (1999a).

drainage density and thus x_c (Montgomery and Dietrich, 1992), associated with tectonic or climatic change (Howard, 1997; Rinaldo and others, 1995; Tucker and Bras, 1998) may involve dramatic short-term fluctuations in sediment yield (for example, Tucker and Slingerland, 1997) but will generally have only minor influences on β.

As first suggested by Whipple and Tucker (1999a), response time (T) of the detachment-limited channel system is given by the change in channel-head elevation between initial and final steady states divided by the time-integrated rate of channel head elevation change:

$$T = \frac{z_f(x_c)_{ss} - z_i(x_c)_{ss}}{\frac{1}{T}\int_0^T \left(\frac{dz}{dt}\right)_{x_c} d\tau} \tag{3}$$

where subscripts f and i refer to final and initial conditions, respectively, and τ is a dummy integration variable for time. An instantaneous step-function change in either tectonic or climatic conditions will initiate an upstream propagating knickpoint – an abrupt change in channel gradient from the initial steady-state gradient upstream of the knickpoint and the final steady-state gradient downstream (Whipple and Tucker, 1999a). Assuming no rounding of knickpoints (and therefore no change in the shape of the migrating waveform), no information propagates upstream of the migrating knickpoint, and the rate of channel-head elevation change ($(dz/dt)x_c$) is constant if uplift is constant (fig. 1). This follows because the gradient at the channel head does not change until the knickpoint has swept through the entire channel system. The final steady-state form is achieved at this instant.

The rate of channel-head elevation change in general depends on the transient imbalance between uplift and erosion at the channel head. Before the perturbation the erosion rate is everywhere equal to the initial rock uplift rate (U_i), and the transient rate of channel head elevation change can be readily defined (fig. 1):

$$\left(\frac{dz}{dt}\right)_{x_c} = U_i(f_U - 1) \quad \text{(tectonic)} \tag{4A}$$

$$\left(\frac{dz}{dt}\right)_{x_c} = U(1 - f_K) \quad \text{(climatic)} \tag{4B}$$

$$\left(\frac{dz}{dt}\right)_{x_c} = U_i(f_U - f_K) \quad \text{(both tectonic and climatic)} \tag{4C}$$

where f_U and f_K are the fractional change in uplift and erosion coefficient, respectively, and are defined as

$$f_U = \frac{U_f}{U_i};\ f_K = \frac{K_f}{K_i} \tag{5}$$

Both f_U and f_K are by definition greater than unity for an increase in uplift rate and an increase in the coefficient of erosion, respectively. By substituting eqs (2A) and (4) into (3) as appropriate, response times for systems subjected to step-function changes in uplift rate (U), climate (K), or both can be derived.

Tectonic forcing.—Fluvial response time to a step-function change in uplift rate with no change in erosion coefficient was derived by Whipple and Tucker (1999a) for the case of an increase in uplift rate. A generalized form of this solution can be written to include either increases or decreases in rock uplift rate (substitute eqs (2A) and (4A) into (3)):

$$T_U = \beta K^{-1/n} U_i^{1/n-1}(f_U^{1/n} - 1)(f_U - 1)^{-1};\ f_U \neq 1 \tag{6}$$

Note that for the commonly assumed linear case ($n = 1$) system response time is independent of both the initial uplift rate and the fractional change in uplift rate. Surprisingly, eq (6) indicates that for $n > 1$ response time decreases for both greater initial uplift rates and greater fractional increases in uplift rate. The converse is true for $n < 1$. If starting from the same initial steady-state profile, response time is in all cases greater for lower values of the slope exponent (n). Graphical examples are given in the following section for the specific case of the Central Range of Taiwan.

Climatic forcing.—Fluvial response time to a step-function change in the coefficient of erosion—meant to represent system response to a change in climate (for example, a change in precipitation or the sediment flux delivered to channels from hillslopes)—is given by substituting eqs (2A) and (4B) into (3):

$$T_K = \beta K_i^{-1/n} U^{1/n-1}(f_K^{-1/n} - 1)(1 - f_K)^{-1};\ f_K \neq 1 \tag{7}$$

This expression has a form similar to eq (6) but exhibits a notably different behavior. Whereas lower values of n still give the longest response times, for all values of n response time is predicted to decrease monotonically with the fractional change in K for $f_K > 1$ and increase monotonically with the fractional change in K for $f_K < 1$.

Combined climatic and tectonic forcing.—Slightly more complex scenarios than described by eqs (6) and (7) can also be handled analytically. A change in rock uplift rate, for instance, may be accompanied by a commensurate change in the coefficient of erosion. Such a change in K might be expected from a change in orographic precipitation, a change in the sediment flux carried by the river (Sklar and Dietrich, 1998), or an adjustment of channel width. Indeed, Snyder and others (2000) have found evidence for such a dynamic change in K in coastal streams in northern California that are responding to a recent increase in rock uplift rate.

Two idealized scenarios can be envisioned: (1) the coefficient of erosion changes everywhere along the profile (as might be expected for a change in orographic precipitation, here assumed for simplicity to occur at the instant of the change in rock uplift rate), or (2) the coefficient of erosion only changes downstream of the migrating knickpoint (as might be the case if a change in channel width or sediment flux is important). In the first case, system response time to contemporaneous tectonic and climatic perturbations is given by substituting eqs (2A) and (4C) into (3):

$$T_{U,K} = \beta K_i^{-1/n} U_i^{1/n-1} (f_K^{-1/n} f_U^{1/n} - 1)(f_U - f_K)^{-1};\ f_U \neq f_K \tag{8}$$

In the second case, system response time to a step-function change in rock uplift rate and an adjustment of the fluvial erosivity downstream of the migrating knickpoint is given by substituting eqs (2A) and (4A) into (3) and noting that $z_f(x_c)_{ss}$ (only) is affected by the change in K:

$$T_{U,K} = \beta K_i^{-1/n} U_i^{1/n-1} (f_K^{-1/n} f_U^{1/n} - 1)(f_U - 1)^{-1};\ f_U \neq 1 \tag{9}$$

Eqs (8) and (9) describe the reduction in predicted response time associated with an increase in erosivity in concert with an increase in rock uplift rate. Again, graphical examples are given in the following section for the specific case of the Central Range of Taiwan. In eq (9), predicted response time goes to zero if $f_K = f_U$. This unrealistic result derives from the implicit assumption that there is no time lag between the passing of the migrating knickpoint and the ensuing increase in fluvial erosivity (for example, in the case of increasing sediment flux, the response of all side tributaries and hillslopes is assumed to be instantaneous).

In all cases (eqs 6-9), it is interesting to note that the dependence of response time on system size (L) is considerably less than linear (see eqs 2B and 2C). Although response time does increase monotonically with basin size, the dependence is weaker than one might have intuited and depends directly on the hm/n ratio that importantly influences the concavity of streamwise longitudinal profiles (see eq 2 and Whipple and Tucker, 1999a).

Limitations.—As mentioned earlier, derivation of eqs (6-9) is predicated on the assumption that no rounding of knickpoints (here defined as abrupt changes in channel gradient) accompanies their upstream migration. If a migrating knickpoint is rounded such that there is a continuous change in channel gradient, then progressive changes in the waveform (Weissel and Seidl, 1998; Royden and others, 2000; Royden, Clark, and Whipple, 2000) will in some cases cause progressive rounding and dissipation of the knickpoint such that changes in channel gradient occur upstream of the otherwise expected knickpoint position, resulting in temporal variation in the rate of channel-head elevation change. When this occurs, generally much longer response times result. For the linear case ($n = 1$), no such effects occur, and eqs (6-9) are always valid. For sub-linear erosion processes ($n < 1$), progressive rounding of knickpoints and the propagation of signals ahead of knickpoints occur in "rising state" transients: an increase in uplift rate or a decrease in the coefficient of erosion. Conversely, for supra-linear erosion processes ($n > 1$), these effects are expected in "declining state" transients: a decrease in uplift rate or an increase in the coefficient of erosion. In these cases, a precise response time is difficult to define as the new steady-state condition will be only asymptotically approached. Response time is considerably longer in all such cases than predicted by eqs (6-9). These behaviors are seen in finite-difference numerical solutions (that is, in standard landscape evolution models) due to the rounding of knickpoints caused by numerical diffusion (Baldwin and Whipple, 1999) and are borne out in newly developed analytical solutions to the channel profile evolution equation (Royden and others, 2000; Royden, Clark, and Whipple, 2000). Thus, the response times calculated here are minimum estimates.

CASE STUDY: CENTRAL RANGE OF TAIWAN

A case study of the Central Range of Taiwan is used here to assess whether natural systems are likely to achieve and sustain a condition of steady-state denudation. The Central Range of Taiwan has long been upheld as an example of a steady-state orogen in which erosion balances rock uplift (Hovius and others, 2000; Teng, 1990). A combination of modern sediment yields, offshore sedimentation rates, marine terrace uplift rates, and thermochronologic data have been used to argue that long-term rates of rock uplift are balanced by erosion (Li, 1976; Liew, Hsieh, and Lai, 1990; Liu, 1982; Wang and Burnett, 1990). Whether or not steady-state landscapes are a probable state, however, depends precisely on the relative timescales of catchment response and tectonic and/or climatic forcing. Only where response time is short compared with the timescale of external forcing can steady-state conditions be expected to prevail.

In order to explore this critical issue, I have computed representative catchment response times for the Central Range of Taiwan. In Taiwan denudation is dominantly by landsliding (Hovius and others, 2000), and it is reasonable to assume that hillslope response is geologically instantaneous. Thus, the fluvial response times given by eqs (6-9) should be representative of landscape response time. To constrain these calculations, the various model parameters in eqs (6-9) can be fit to Taiwanese stream profiles by making the assumption that the modern landscape approximates a steady-state form. This calibration process only serves to provide ballpark values of key parameters to allow calculation of response times representative of a narrow, rapidly uplifting mountain range eroded by fluvial processes. Potential deviations from the assumed steady, uniform conditions (Slingerland, 1999) are unlikely to impact significantly the predicted system response times.

The trunk stream of the typical east-side drainage basin in the northern part of the Central Range of Taiwan is 20 to 50km long (L), with a drainage area of $10^8 - 10^9$ m^2 (fig. 2A; table 1). The critical drainage area at which fluvial processes become dominant is roughly estimated at 10^5 m^2 based on the break in scaling in the slope-area relationship (fig. 2A inset; Montgomery and Foufoula-Georgiou, 1993; Snyder and others, 2000). Regression of drainage area against downstream distance yields well-defined Hack's law relationships ($k_a \sim 1$; $h \sim 1.9$) (fig. 2B; table 1). Using Hack's law, the critical drainage area estimated above gives $x_c = 450$ m.

Longitudinal profiles of bedrock streams are often observed to follow a power-law relationship between channel gradient and drainage area (for example, Flint, 1974; Tarboton, Bras, and Rodriguez-Iturbe, 1989):

$$S = k_s A^{-\theta} \tag{10}$$

Such power-law scaling is well defined for the Taiwanese river profiles analyzed (fig. 2A). At steady state where the coefficient of erosion and rock uplift rate are spatially uniform, channel steepness index (k_s) and concavity (θ) are related to model parameters according to the relations (Moglen and Bras, 1995; Sklar and Dietrich, 1998):

$$k_s = (U/K)^{1/n} \tag{11A}$$

$$\theta = m/n \tag{11B}$$

Thus m/n is given by the slope of the gradient-area relationship in a log-log plot, and for any given value of the slope exponent (n) an appropriate coefficient of erosion (K) can be calculated from the intercept ($\log k_s$) of a regression of log gradient against log drainage area, provided the rock uplift rate is known (fig. 2A; table 1).

Power-law regression of channel gradient against drainage area for the trunk streams of two typical drainage basins yields mean estimates of steepness index (k_s) and concavity (θ) of 126 and 0.43, respectively (fig. 2A; table 1). Owing to the considerable

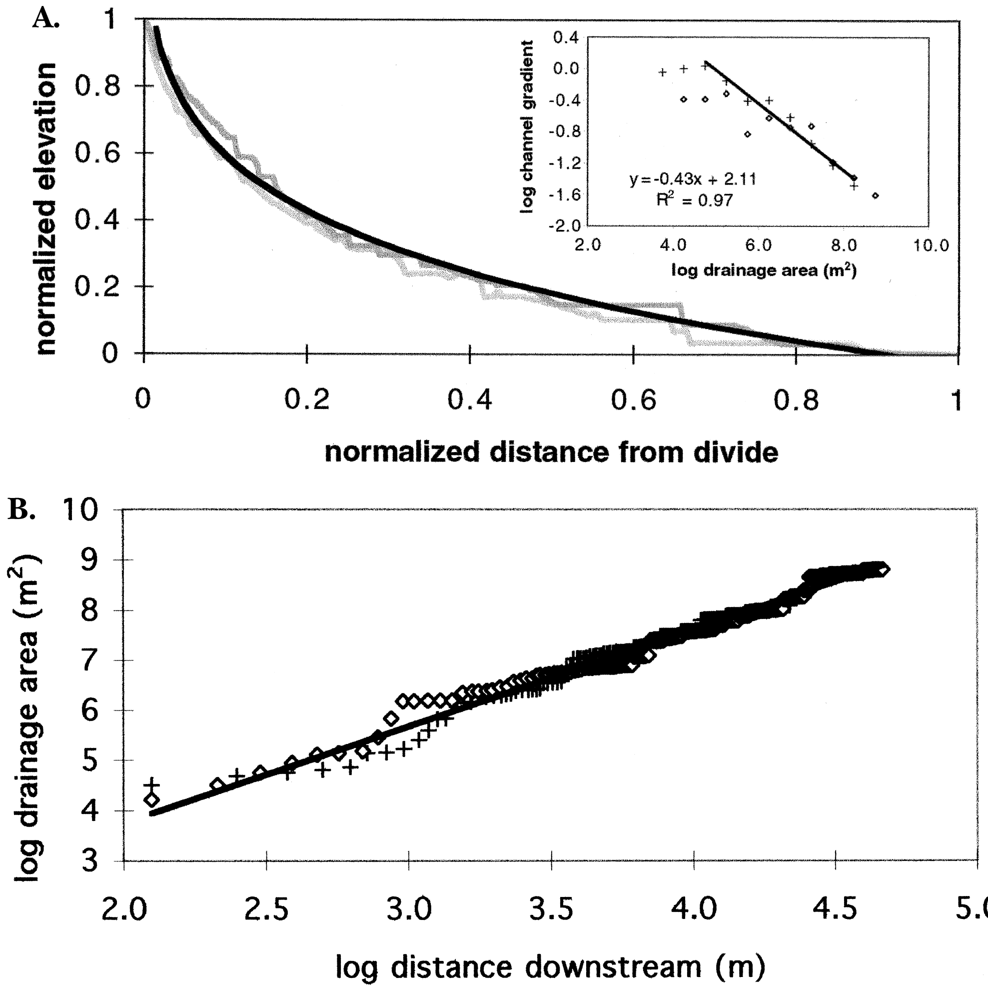

Fig. 2(A) Normalized longitudinal stream profiles from the northern part of the Central Range of Taiwan (dark gray: profile A; light gray: profile B) shown in comparison to the best-fit steady-state profile (black; fit to profile A). Jagged, stair-step nature of the profiles is an artifact of the 90-m resolution DEM. Inset shows a log-log plot of channel gradient against drainage area (m^2) (diamonds: profile A; crosses: profile B). (B) Log-log plot of drainage area (m^2) versus downstream distance (m) (diamonds: profile A; crosses: profile B). Fits to slope-area and area-distance relations are given in table 1.

TABLE 1

Model parameters

Drainage	L [km]	x_c [m]	k_a [$m^{0.1}$]	h	m/n*	k_s [$m^{0.84}$]	K (n = 2/3)** [$m^{0.72}a^{-1}$]	K (n = 1)** [$m^{0.16}a^{-1}$]	K (n = 2)** [$m^{-0.68}a^{-1}$]
Profile A	46.6	450	0.77	1.93	0.42	123	2.03E-04	4.10E-05	3.36E-07
Profile B	21.5	450	1.05	1.89	0.43	130	1.97E-04	3.90E-05	3.04E-07

* For assumed steady, uniform conditions $\theta = m/n$.
** K calculated from eq (11) assuming $U = 0.005\ ma^{-1}$.

scatter in pixel-by-pixel channel gradients derived from digital topographic data, channel gradients were binned in equal increments (0.5) of the logarithm of drainage area (m^2) and averaged prior to regression (Tarboton, Bras, and Rodriguez-Iturbe, 1989). Estimates of remaining critical model parameters (K, m/n) are determined by assuming – for the purpose of response time calculations – that (1) these channel profiles are in a quasi-steady state, (2) the rate of rock uplift and the coefficient of erosion are spatially uniform, and (3) the rate of rock uplift is adequately constrained at 5 mma^{-1} (Hovius and others, 2000; Liew, Hsieh, and Lai, 1990; Teng, 1990; Wang and Burnett, 1990). These values can then be used in eqs (6-9) to compute approximate channel system response times.

The results of these calculations are presented in figure 3. All calculations are based on parameters fit to profile A (table 1). Response time in all cases is on the order of 1 Ma (roughly 0.25 – 2.5 Ma for the range of conditions considered). Response times for the shorter channel (profile B, $L = 21.5$ km) are approx 30 percent less than those for profile A ($L = 46.6$ km), reflecting the relatively weak dependence on basin size (eq 2B and C). Figure 3A shows the predicted response time to a step-function change in rock uplift rate alone (eq 6) as a function of the fractional change in uplift rate (f_U). For a doubling of rock uplift rate ($f_U = 2$), predicted response times are 1.18, 0.64, and 0.27 Ma for $n = 2/3$, $n = 1$, and $n = 2$ respectively. Figure 3B shows the reduction (for $f_K > 1$) in response time from these values should the coefficient of erosion vary in concert with rock uplift rate due to a narrowing in channel width, an increase in sediment flux, or an increase in orographic precipitation (eq 9). Finally, figure 3C shows predicted response times to a step function change in climate (more precisely the coefficient of erosion), with response times ranging from 0.25 to 2.5 Ma for $0.5 \leq f_K \leq 2$ (eq 7).

DISCUSSION

Response times estimated here for the actively uplifting Central Range of Taiwan imply that attainment of steady-state denudation is plausible during periods of climatic stability but unlikely during periods characterized by climatic instability, such as the Quaternary. Particularly in an orogen subjected to rapid erosion, it seems reasonable to expect that tectonic convergence rates and thus rock uplift rate may be relatively steady over intervals considerably longer (> 5 Ma) than estimated response times. The Central Range of Taiwan, the Southern Alps of New Zealand, and the Himalaya all show evidence of long-term persistence of tectonic rock uplift (Hodges, 2000; Teng, 1990; Walcott, 1998). Rapid erosion is required to ensure that the topography is erosionally limited and that the principal structures accommodating uplift and exhumation are persistent and stable, such that the orogen does not steadily grow in width by a forward stepping of thrust faults, for instance (Beaumont and others, 1996; Willett, 1999). Thus topographic and denudational steady state may reasonably be expected in rapidly eroded landscapes in periods of climatic stability.

Steady-state conditions (with respect to tectonic perturbation) are even more likely where feedbacks associated with orographic precipitation or internal adjustments (channel width, sediment flux) dampen landscape response to tectonic forcing and reduce system response time (eqs 8 and 9; fig. 3B). The King Range on the northern California coast serves as a case in point. In this field setting Snyder and others (2000) have presented evidence that the trunk streams of small coastal drainages (4-20 km^2) have already adjusted to an ~8-fold increase in rock uplift rate that occurred only about 100,000 yrs ago (Merritts and Bull, 1989). Response times for these coastal drainages are short not only because of their small size compared to the drainages of the Central Range of Taiwan but also because of a strong apparent enhancement of fluvial erosivity (K) in the zone of high uplift rates. Snyder and others suggest that the apparent increase in K likely reflects some combination of orographic

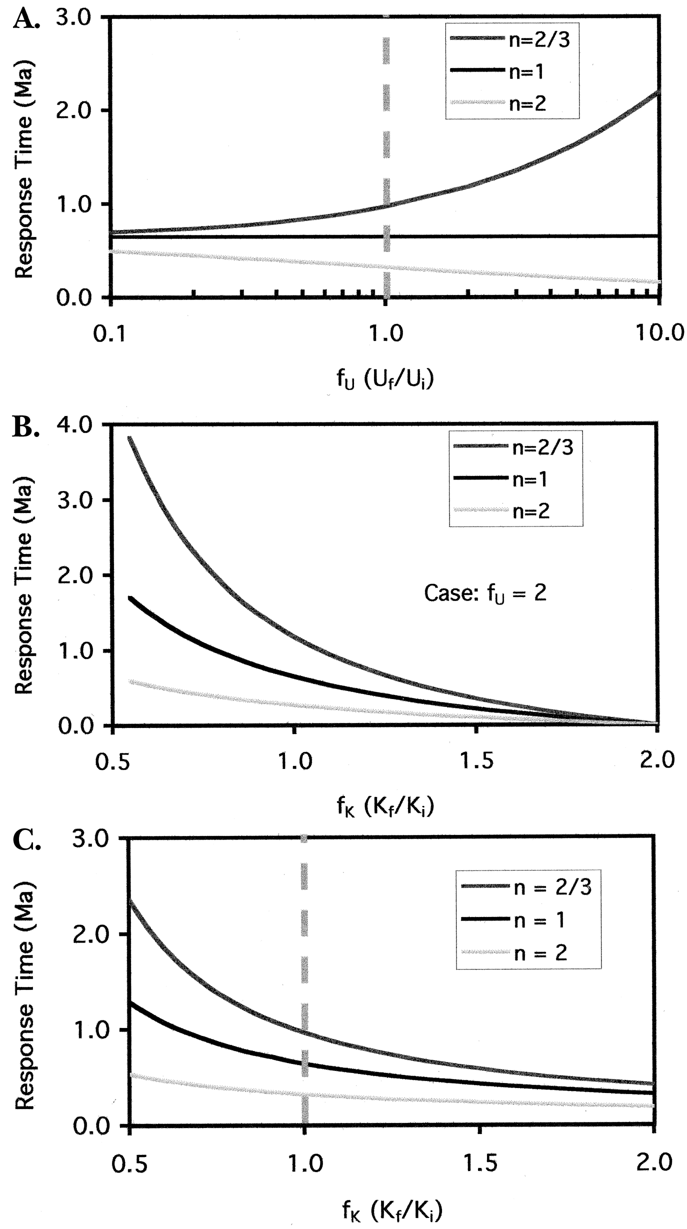

Fig. 3. Predicted fluvial response timescales. All parameters used are listed in table 1. (A) Response time to a step function change in rock uplift rate (relative to erosional baselevel) (eq 6). Note log scale on abscissa. Heavy dashed line indicates that response times are not defined for $f_U = 1$. Note that response time is the same for any magnitude of tectonic perturbation if $n = 1$. Interestingly, response times are predicted to *decrease* with greater fractional increases in rock uplift rate for the case $n = 2$. (B) Modulation of tectonic response time for a contemporaneous change in the coefficient of erosion (K) as a function of the fractional change in K (f_K) computed for the case $f_U = 2$ (see 3a) (eq 9). Note that response time *increases* for $f_K < 1$. (C) Response time to a step function change in erosivity (K) as a function of the fractional change in K (f_K) (eq7). Heavy dashed line indicates that response times are not defined for $f_K = 1$. Unlike the case for tectonic perturbations, response time to climate changes decreases monotonically with fractional increases in K for all values of n.

enhancement of precipitation, an increase in sediment flux, a decrease in channel width, and an increase in the frequency of debris flows. Response times predicted for this landscape are on the order of 100 ka, consistent with the observation that the channels appear to have adjusted to the recent increase in rock uplift rate. However, this does not necessarily imply that the King Range is an example of a landscape in topographic and denudational steady state (Snyder and others, 2000). This follows because hillslope response has lagged behind channel response and because Quaternary climate change has affected these basins on a 10 ka timescale.

Steady-state topography and steady-state denudation (that is, erosion in balance with rock uplift rate) are less likely to be attained and sustained during periods of climatic instability. The Quaternary, for instance, has been marked by dramatic climatic fluctuations on a timescale of 40 to 100 ka or less (Imbrie and others, 1984). Thus the timescale of recent climatic forcing is considerably shorter than the system response times determined above (fig. 3C). Possible deviations from ideal steady-state profile forms in the Taiwanese rivers (Slingerland, 1999) may be an indicator of the expected disequilibrium, although several other interpretations (for example, spatial variations in uplift rate or lithologic erodibility) are equally plausible. Indeed, given that the timescale of climatic forcing is so short compared with system response time, it may reasonably be expected that rapid climatic fluctuations would have negligible impact on landscape form despite significant short-term fluctuations in denudation rate of the type that have been observed in the field (Bull, 1991) and in landscape evolution models (Tucker and Slingerland, 1997). This follows because in such cases there will be insufficient time to alter significantly channel profiles and associated catchment relief (Whipple, Kirby, and Brocklehurst, 1999; Whipple and Tucker, 1999a) before the climate returns to its previous state. However, there are many interesting complications and further study of landscape response to oscillatory changes in climate are clearly warranted.

As discussed above, a significant limitation of the present work is the strict application to purely detachment-limited channels. If channels are transport- rather than detachment-limited, or if a transition from one channel type to another occurs in response to changing slope and sediment flux, system response times to both tectonic and climatic perturbations are significantly longer than that predicted by eqs (6 - 9) (Baldwin and Whipple, 1999; Whipple and Tucker, 1999b). A manuscript currently in review (Baldwin, Whipple, and Tucker, submitted to Journal of Geophysical Research) will address the impact of a transition from detachment-limited to transport-limited erosion on the timescale of post-orogenic topographic decay. The present interpretation that steady state conditions are unlikely in the Quaternary climatic regime, however, is conservative in that the calculations presented above provide a minimum estimate of system response time and thus over-estimate the likelihood that steady-state conditions can be sustained. Conversely, considerably longer periods of tectonic stability may be required to attain steady-state topography where channels are transport-limited.

CONCLUSIONS

Calculations of the timescale of channel response to tectonic perturbation show that steady-state topography and denudation are plausible under stable climatic conditions – the timescale of tectonic perturbation is sufficiently long compared to landscape response time. Quaternary climate fluctuation, however, is too rapid to allow for true steady-state conditions to hold for any modern landscape. Despite this disequilibrium in denudation rate, quasi-steady state landscape forms may persist in modern landscapes as Quaternary climate fluctuations have been so rapid that significant morphologic adjustment may not have time to occur on the drainage-basin scale during any given climatic oscillation. On the other hand, the Quaternary period has been sufficiently long that actively uplifting and eroding modern landscapes such as the Central Range of Taiwan have likely been able to adjust to mean Quaternary climatic conditions. Thus it is not surprising that quasi-steady-state channel profiles characterize the mountainous landscape of Taiwan, but it is surprising that modern suspended sediment records appear to yield erosion rate estimates that balance estimates of the long-term rock uplift rate (Hovius and others, 2000; Li, 1976; Liew, Hsieh, and Lai, 1990; Liu, 1982).

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