

# How do landscapes record tectonics and climate?

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## ABSTRACT

The Earth's surface is shaped by tectonics and climate. This simple statement implies that we should, in principle, be able to use the landscape as an archive of both tectonic rates and of changes to climate regime. To solve this inverse problem, and decipher the geomorphic record effectively, we need a sound understanding of how landscapes respond and erode in response to changes in tectonic or climatic boundary conditions. Rivers have been a major focus of research in this field because they are patently sensitive to tectonic and climatic forcing via their channel gradient and discharge. Theoretical, field, and numerical modeling techniques in the last few years have produced a wealth of insight into the behavior of fluvial landscapes, while the increasing availability of high-resolution topographic models have provided the data sets necessary to address this research challenge across the globe. New work by Miller et al. (2012) in Papua New Guinea highlights the progress we have made in extracting tectonics from topography due to these developments, but also illustrates the problems that still remain. This paper reviews our current knowledge of how fluvial landscapes record tectonics at topographic steady-state and under "transient" conditions, assesses why the climate signal has proven so challenging to interpret, and maps out where we need to go in the future.

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## INTRODUCTION

The Earth's landscape is the time-integrated product of two factors: (1) tectonics, which can create topography and maintain relief through surface and rock uplift (Whipple, 2004; Wobus et al., 2006a; Whittaker et al., 2008; Hartley et al., 2011) and (2) climate, which mediates the erosional processes that wear away upland areas over time (Allen, 2008; Whipple, 2009; Armitage et al., 2011). The interaction of these processes forms, modifies, or destroys geomorphic features on the Earth's surface (Wobus et al., 2006a; Tucker, 2009). This statement has profound implications, because if topography represents a filtered signal of tectonics and climate then we should be able to use the landscape as an archive, or "tape recorder," of these two drivers (Wobus et al., 2006a; Tucker, 2009). If we could decode this archive, we would gain an accessible method of deducing: rates of rock uplift (e.g., Kirby et al., 2003; Cyr et al., 2010); the location and slip rate of active faults (Whittaker et al., 2008; Boulton and Whittaker, 2009); seismic hazard (Kirby et al., 2008); and landscape sensitivity to future climate change (Whipple, 2009; Armitage et al., 2011). "Tectonics from topography" would be particularly powerful in areas where geodetic or geologic data are limited (cf. Wobus et al., 2006a), while a predictive understanding of landscape response to environmental change (e.g., Molnar et al., 2006) remains an outstanding research challenge.

The elements needed to solve this problem are now apparently in place: the widespread availability of digital elevation models (DEMs) covering the globe at high resolution provides a ready source of topographic data that can be analyzed using GIS (geographic information system) software, while the development of sophisticated landscape evolution models (LEMs) has enabled us to investigate the time-dependent evolution of the landscape to tectonic and climate perturbations that cannot be done from field observation alone (Wobus et al., 2006a; Tucker, 2009). Quantitative insights into "rates and dates" from bedrock and detrital thermochronology, cosmogenic nuclide, and optically stimulated luminescence (OSL) methods provide the detailed constraints needed to ground-truth geomorphic estimates of landscape change over time periods of  $10^4$  to  $10^6$  years (e.g., Ouimet et al., 2009; Cyr et al., 2010; DiBiase et al., 2010). However, solving this research question in practice has proved challenging. Which elements of the landscape are most sensitive to changes in tectonic boundary conditions? Over what time scale do landscape features respond to changes in tectonic or environmental boundary conditions? Which topographic metrics record tectonic signals with the greatest fidelity? These questions are difficult because the interaction between climate, tectonics, and landscape is two-way coupled and displays nonlinear dynamics (Wobus et al., 2006a,b; Whipple, 2009; DiBiase and Whipple, 2011). Teasing apart these signals requires thoughtful selection of study locations, where some boundary conditions can be constrained independently, leaving others to be

carefully investigated (Whittaker et al., 2007a,b; Tucker 2009). Moreover, to do this we need to understand how the time-integrated behavior of erosional systems is physically recorded in topography (DiBiase et al., 2010). Rivers have been a major focus here because they are ubiquitous; they are clearly sensitive to tectonic and climate variables in terms of their channel gradient and discharge; they set the boundary conditions for hillslopes and limit relief; and they control the erosional unloading of orogens. The fluvial network therefore acts as the primary agent by which tectono-climatic signals are transmitted to landscape (Whipple, 2004).

A new paper by Miller et al. (2012, this issue), which addresses the geomorphic evidence for uplift in the Woodlark Rift area, eastern Papua New Guinea, provides a good opportunity to assess how far we have come in using the landscape as an archive of tectonics and climate. Miller et al., in an area where geologic data are sparse, derive the spatial pattern and history of rock and surface uplift from stream profile analyses on the Papuan mainland and surrounding islands. They demonstrate that channel-long profiles have well-defined knickpoints that record a transient erosional response to an increase in the rate of uplift during the Quaternary, the magnitude of which increases east to west across the study area. These results are exciting because they demonstrate that sophisticated insights into the relative magnitude and timing of surface and rock uplift can now be derived from geomorphic analyses, even where existing geologic/geodetic constraints are poor. The results are also provocative, because the authors argue that the effect is

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the result of dynamic topography driven by buoyant, asthenospheric upwelling beneath thinned crust. Such findings highlight the role of landscape in directly recording the large-scale coupling of surface, crustal, and mantle processes (Whipple, 2009; Hartley et al., 2011).

Nevertheless, this study illustrates some of the problems we face in extracting tectonics from topography, because the authors struggle to convert topographic metrics (however calculated) explicitly to absolute rates of rock or surface uplift. Consequently, without independent data (e.g., on rock exhumation rates) most quantitative geomorphic studies still produce qualitative tectonic insights (cf. DiBiase and Whipple, 2011). Moreover, the extent to which climate should be taken into account remains contentious: many geomorphologists have implicitly hoped that time-averaging of this signal over  $10^5$ – $10^7$  years would result in the climate control on erosion being damped relative to the long-term tectonic driver (Kirby et al., 2003; Whittaker et al., 2007a,b). This could be wishful thinking depending on the extent to which erosional systems on the Earth's surface are buffered or are sensitive to climate variability over a range of realistic amplitudes and frequencies (Armitage et al., 2011).

### TECTONICS FROM TOPOGRAPHY: TOPOGRAPHIC STEADY-STATE

Most work to date in extracting tectonics from topography in fluvial landscapes can be framed, explicitly or implicitly, with reference to the stream power erosion “law,” in reality a set of closely related equations that treat fluvial erosion rates as power law functions of both upstream drainage area,  $A$  (a proxy for catchment discharge), and channel gradient,  $S$  (e.g. Tucker and Whipple, 2002; Whipple and Tucker, 2002; Crosby and Whipple, 2006; Whittaker et al., 2008). Stream power erosion laws typically have the form

$$\frac{dZ}{dt} = U - f(Q_s)K \frac{A^m}{W} S^n, \quad (1)$$

where the rate of change of elevation,  $Z$ , with time,  $t$ , depends on the imposed uplift rate,  $U$ ,  $m$ , and  $n$  are positive exponents that describe the relative dependency of erosion rates on  $A$  and  $S$ ,  $W$  is channel width,  $f(Q_s)$  is a term that describes sediment supply effects (cf. Sklar and Dietrich, 2004) and  $K$  is an erosional efficiency parameter controlled by factors such as lithology.  $W$  may be described as a power law function of  $A$ , in which case its effect can be subsumed into  $m$  and  $K$  (e.g., Attal et al., 2008, 2011), and  $f(Q_s)$  can be (and is often) taken to be equal to 1 for detachment-limited (i.e., bedrock) rivers (cf. Cowie et al., 2008). For catchments

in topographic steady-state (Fig. 1A), where the rate of rock uplift,  $U$ , equals the rate of incision,

$$\frac{dZ}{dt}, \text{ we obtain } S = k_s A^{-\left(\frac{m}{n}\right)}, \quad (2)$$

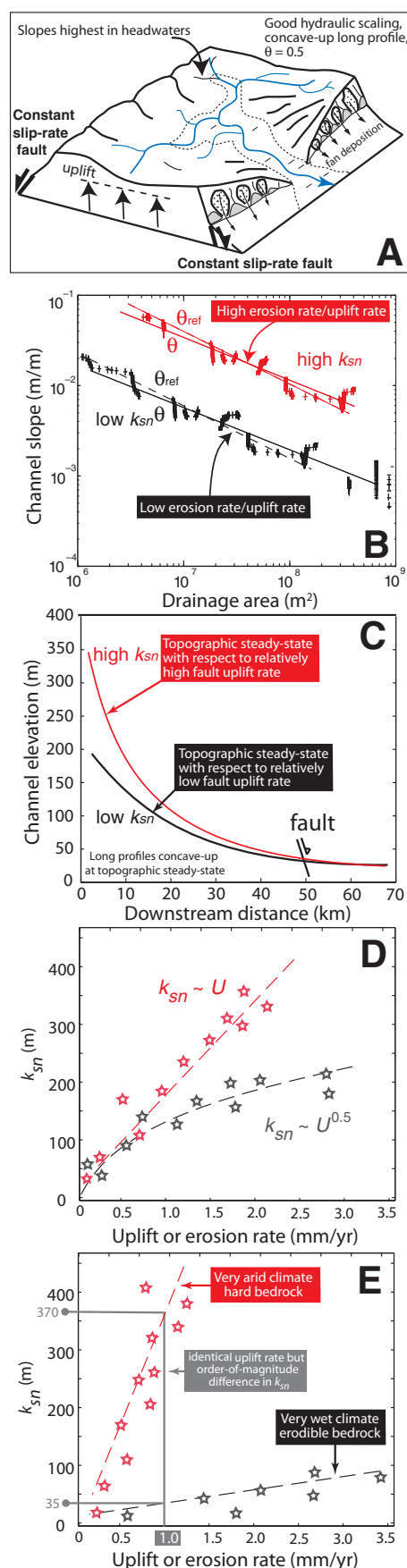
where the pre-factor,  $k_s = \left(\frac{U}{K'}\right)^{\frac{1}{n}}$  is the steepness index of the channel that depends on uplift rate and  $K'$  (which embeds all other relevant factors from Eq. 1).  $m/n$  is the concavity (usually given the symbol  $\theta$ ).  $k_s$  and  $\theta$  can be readily estimated from log-log plots of  $S$  and  $A$  extracted from DEMs, but as the steepness index and concavity necessarily co-vary, normalized steepness,  $k_{sn}$  is typically calculated using a reference concavity for the study region (i.e.,  $0.45 < \theta < 0.5$ , which determines the units of  $k_{sn}$ ) (Wobus et al., 2006a; Ouimet et al., 2009; DiBiase et al., 2010) (Fig. 1B).

Consequently, most researchers attempting to extract tectonics from topography have focused on trends in normalized steepness index between or along rivers, which are not explained by factors such as lithology (Fig. 1C). Note that at topographic steady-state, for the case of active uniform fault uplift, river long profiles are concave up but differ in terms of their steepness (Fig. 1C).  $k_{sn}$  is much more useful than simply considering mean catchment hill-slopes, because the latter are insensitive to uplift rates once the appropriate threshold is passed (Ouimet et al., 2009). Normalized steepness indices have shown a good correlation with independent measures of uplift rate, erosion rate, or base level fall in a wide range of areas from Tibet (e.g., Kirby et al., 2003; Harkins et al., 2007, Ouimet et al., 2009) to the Italian Apennines (Whittaker et al., 2008; Cyr et al., 2010) and they are now justifiably and routinely used in many geomorphic studies.

Nevertheless, converting  $k_{sn}$  values to quantitative estimates of tectonic rates has proved difficult for two reasons: first, the form of the relationship between  $k_{sn}$  and uplift or erosion



**Figure 1. (A) Cartoon showing a river eroding an uplifting fault-bounded horst, which has achieved topographic steady-state with the rock uplift rate in the footwall. (B) Theoretical channel slope-drainage area plots for the river in (A). If the fault uplifts at a higher rate, the normalized steepness index  $k_{sn}$  of the channel is greater (i.e.,  $S$  is larger for identical  $A$  than at low uplift/erosion rates).  $\theta$  is the best-fit concavity (i.e., gradient) of the  $S$ - $A$  plot, and  $\theta_{ref}$  is the regional reference concavity (typically 0.45–0.5) used to derive  $k_{sn}$ . (C) Representative long profiles implied for the high and low  $k_{sn}$  rivers shown in (B). (D) Theoretical relationships between  $k_{sn}$  and uplift or erosion rate vary from linear (red) to sub-linear (black), depending on erosion dynamics. (E)  $k_{sn}$  against uplift or erosion rate in areas with different climate or rock strengths.**



rate is not fully constrained (Fig. 1D). Initially, workers assumed a linear relationship between  $k_{sn}$  and  $U$ , consistent with  $n = 1$  in a stream power erosion law (Eq. 1) (Kirby et al., 2003; Wobus et al., 2006a). However, some studies have obtained a better match to erosion rate data with  $k_{sn} \sim U^{0.5}$  (e.g., Ouimet et al., 2009; DiBiase et al., 2010). This disparity in functional form makes a big difference in how estimates of  $k_{sn}$ , pinned to measurements of erosion rate (e.g., Ouimet et al., 2009; DiBiase et al., 2010) or fault throw (e.g., Whittaker et al., 2008; Boulton and Whittaker, 2009), are converted to tectonic rates across a study region (Fig. 1D). Fundamentally, the uncertainties in this functional relationship reflect uncertainties in the long-term erosional dynamics of upland channels that Equation 1 attempts to characterize (Wobus et al., 2006b; DiBiase and Whipple, 2011).

Second, channels in topographic steady-state can have identical  $k_{sn}$  values for significantly different uplift rates (Fig. 1E). Normalized  $k_{sn}$  values in Turkey and the Italian Apennines are typically 100–200 m (Whittaker et al., 2008; Boulton and Whittaker, 2009) for fault uplift rates of <0.4 mm/yr. However, in the Siwalik Hills of the Himalaya,  $k_{sn} \ll 100 \text{ m}^{0.9}$  can be found for channels eroding twenty times faster, at rates up to a centimeter a year (Ouimet et al., 2009). This is no surprise, as  $k_{sn}$  implicitly and necessarily conflates a range of parameters, such as rock strength and climate regime that obviously vary between study areas. But it does underline that there is presently no context-independent means to convert  $k_{sn}$  to tectonic rates without other geologic data (DiBiase and Whipple, 2011). The message is not that normalized steepness indices aren't useful (categorically untrue), but rather that the challenge for geomorphologists in the field is to apply these metrics effectively as a function of the geologic data available.

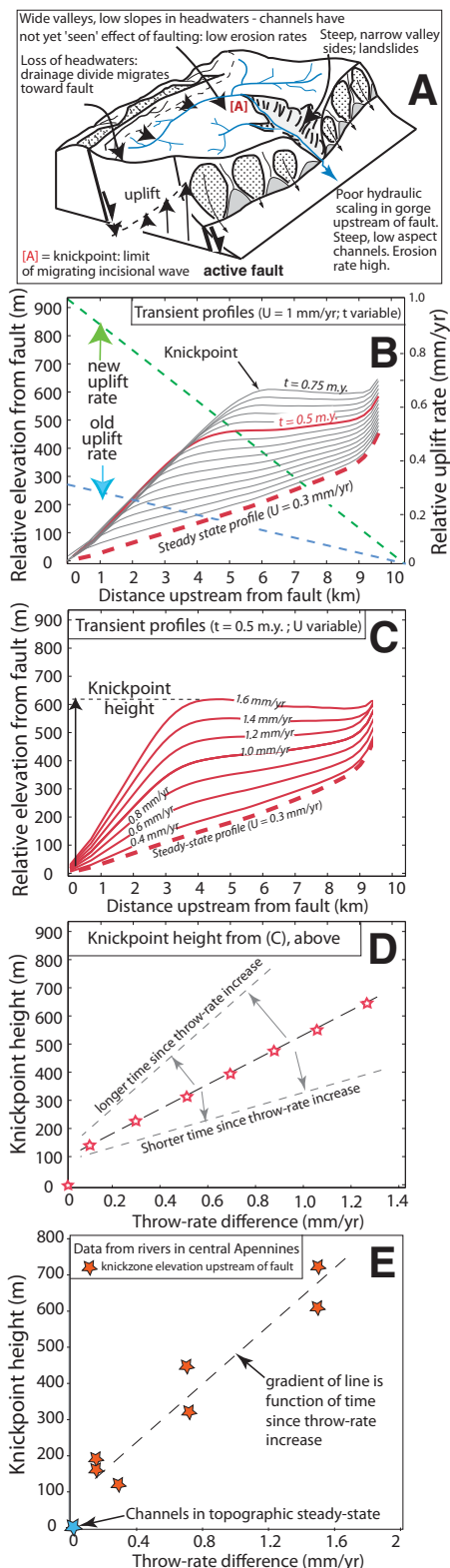
## TRANSIENT LANDSCAPES

The use of steepness indices as a proxy for uplift rates was initially rooted in notions of topographic steady-state (Eq. 2). However, studies suggest that landscapes may take several million years to respond to changes in tectonic rates (Whittaker et al., 2008; Armitage et al., 2011). This transient, time-dependent behavior can result in landscape morphologies that do not resemble fluvial catchments in topographic steady-state (Tucker, 2009). Theoretical, observational, and modeling studies have shown that when detachment-limited “bedrock” channels are perturbed by a change in relative base level, a steep transient convex reach develops as the channel adjusts its form to incise and keep pace with the new boundary conditions (Tucker and Whip-

ple, 2002; Crosby and Whipple, 2006; Whittaker et al., 2007a) (Fig. 2A and B). A knickpoint separates an incised downstream part of the catchment that has adjusted to the perturbation, such as slip on a fault, from the upstream catchment that is yet to respond (Crosby and Whipple, 2006; Whittaker et al., 2007b). Migration of the knickpoint creates a wave of incision that progressively transmits the signal of boundary condition change to the whole catchment (Tucker and Whipple, 2002; Harkins et al., 2007; Whittaker et al., 2010). In contrast, sediment-dominated transport-limited rivers display a diffusive style of behavior in response to boundary condition change (Tucker and Whipple, 2002; Whipple and Tucker, 2002).

Rivers undergoing a transient response to tectonics make  $k_{sn}$  indices more challenging to interpret (Wobus et al., 2006a,b) because the assumption that fluvial erosion rates balance rock uplift rates (Eq. 1) is not met. However, transient responses also embed tectonic information in other ways. Theory and numerical modeling suggests that the vertical rate of transient knickpoint propagation through a landscape should be independent of catchment size and discharge, but should relate to the relative uplift rate perturbation experienced by the channel, for example, as it crosses a fault (Crosby and Whipple, 2006; Wobus et al., 2006b). The elevation of transient knickpoints in the landscape therefore grows over time, e.g. since a change in fault throw rate (Fig. 2B), and should be larger, after a given time, for greater fault slip rates (Fig. 2C) (Wobus et al., 2006a; Whittaker et al., 2008; Attal et al., 2008). A plot of knickpoint height versus relative uplift rate should form a straight line if the tectonic perturbation were synchronous across the area, and the gradient of the line would be related to the

time since the transient wave of incision started to propagate (Fig. 2D). Whittaker et al. (2008) demonstrated that rivers crossing active faults in the Italian Apennines had knickpoints (Fig. 3A and B) that reflected a mid-Pleistocene (0.8 Ma)



**Figure 2.** (A) Cartoon showing a transient river response to a slip rate change on an active fault. (B) Modeled long profile evolution for a river crossing a fault (located at 0 km) that undergoes an increase in uplift rate (from Whittaker et al., 2008). Initial profile (red dashes) is in steady-state with the “old” throw rate (0.3 mm/yr at fault—blue dashes). Throw rate is instantly increased to 1 mm/yr at fault (green dashes). Grey lines show transient profiles in 0.05 m.y. time-steps. Red line shows profile at  $t = 0.5 \text{ m.y.}$ , identical to “1 mm/yr” long profile in (C). The profile downstream of the knickpoint is re-equilibrated with respect to the new uplift field. (C) Long profiles evolved to 0.5 m.y. for a range of uplift rates, using identical starting conditions to (B). Arrow illustrates knickpoint height for a throw rate of 1.6 mm/yr. (D) Knickpoint height at 0.5 m.y. as function of the throw rates in (C). Arrows show how distribution of knickpoint heights would vary for shorter/longer model run times. (E) Field data for transient rivers in the Apennines show knickpoint height scales with the known throw-rate increase at 0.8 Ma (Whittaker et al., 2008).

increase in fault throw rate and validated that the vertical height of knickpoints upstream of faults scaled directly with fault displacement rate along strike (Fig. 3C and D). Knickpoint height plotted against throw-rate increase for faults across the Apennines forms a linear trend, consistent with a synchronous change in fault uplift rate (Fig. 2E). The fluvial response time was >1 m.y., based on the observation that rivers, crossing faults that had moved at a constant rate for 3 m.y., had reached topographic steady-state. Miller et al. (2012) also use knickpoints to deduce the magnitude of incision related to Quaternary faulting and rock uplift in Papua New Guinea. They show that  $k_{sn}$  values increase in the steepened “transient” reach of the channels as the degree of incision (and the integrated rock uplift) becomes greater. The authors do not have firm constraints on the timing of uplift, but it has taken place since the Pliocene, supporting the idea that transient landscapes may be used as tape recorders of tectonics over million-year time scales. This kind of approach is particularly powerful if the location and relative rates of active faulting can be converted to moment magnitude ( $M_w$ ) or peak ground motion values in an earthquake using additional geologic data (Kirby et al., 2008; Boulton and Whittaker, 2009). For example, Boulton and Whittaker (2009) combined structural measurements of fault throw and segment length in the Hatay Graben, Turkey, with geomorphic measurements of knickpoint height and numerical calculations from fault interaction theory to deduce both the throw rate on the graben-bounding faults and the maximum  $M_w$  expected for a quake rupturing up to half the fault length.

Nevertheless, caution is required because landscape responses are also controlled by flu-

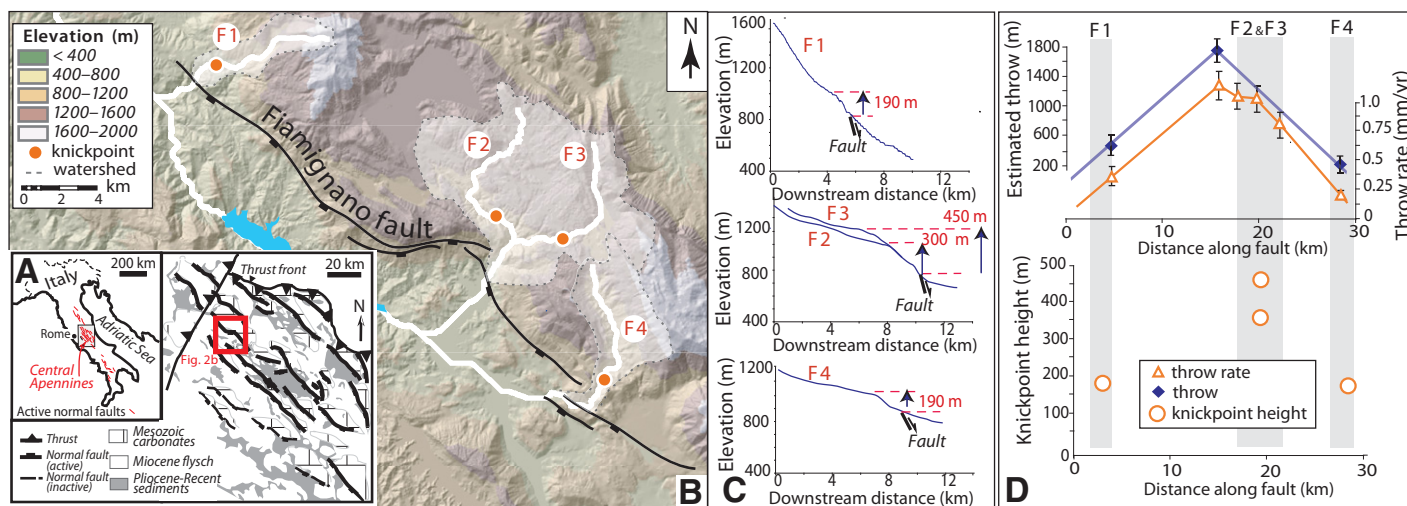
vial erosion dynamics. Cowie et al. (2008) investigated a river in Greece that was responding to a tectonic perturbation similar in timing and magnitude to those described in the Apennines, above. In Greece, however, there was no knickpoint upstream of the fault, because the abundant sediment supply from poorly consolidated conglomerates in the upper catchment boosted the ability of the river to cut across the footwall, without having to steepen to incise. This led to a significantly more diffusive landscape response such that the base-level change was transmitted to the whole catchment in less than 500 k.y. Modeling studies using sediment-dependent fluvial incision laws (cf. Sklar and Dietrich, 2004) have clearly demonstrated a more complex transient response to tectonics than for simple “bedrock” rivers, meaning that channel slopes may steepen and then relax depending on the timing and quantity of sediment supply delivered to the system (Gasparini et al., 2006). Additionally, dynamic channel adjustment may allow rivers to narrow and incise, rather than to steepen during a transient response to tectonic perturbation (Whittaker et al., 2007a; Attal et al., 2008). These complexities make it harder to form simple links between long profile convexity heights, normalized steepness and rock uplift magnitude and rate, without the use of LEMs calibrated using an appropriate erosion “law.”

**WHERE'S THE CLIMATE?**

So far we haven't concentrated on the impact of climate on landscape evolution. This is not because the question isn't interesting (it is!) but because it has proven a difficult problem to address satisfactorily (Tucker, 2004; S6lyom and

Tucker, 2004; Molnar et al., 2006; DiBiase and Whipple, 2011). However, we know that climate differences, such as runoff magnitude and variability, in conjunction with erosion thresholds and vegetation effects, can significantly influence the rates and patterns of landscape evolution and hence topographic metrics used as markers for tectonics (Tucker, 2004; Molnar et al., 2006; Attal et al., 2011). Wobus et al. (2010) argued that climate change can produce transient fluvial landscapes that differ characteristically from the effect of a tectonic perturbation: model results suggested an increase in the relative amount of water delivered to a river system produced a downstream migrating wave of incision and channel gradient relaxation, which contrasted sharply with the upstream wave of incision expected for a relative base-level fall. DiBiase and Whipple (2011) neatly demonstrated that differences in the time-integrated distribution of discharge, including the tail of large events, controlled the functional relationship between  $k_{sn}$  and erosion rate. This relationship is strongly sub-linear in climate regimes where the discharge-related erosion rate is small compared to an erosion threshold, but approaches linearity when the erosion rate dwarfs the threshold term. Trade-offs between the mean runoff and runoff variability thus exert a significant control on erosional efficiency and hence on  $k_{sn}$  with changing erosion rate (Fig. 1D).

These effects should not be ignored, but they do raise the question of “knowability”: if landscape response to climate change can only be properly understood in the context of complete time-series of discharge or storm intensity, information that is essentially unknowable over geologic time periods, is this question, on a



**Figure 3. (A) Location and geological map of active normal faults in the Central Apennines. (B) The Fiamignano fault and study rivers F1–F4; transient knickpoints related to a throw-rate increase at 0.8 Ma are shown as red dots. (C) Long profiles of F1–F4, showing knickpoint elevation measured from the fault. (D) Throw and throw rate along strike of the fault (from Whittaker et al., 2008). Knickpoint height mirrors the throw and throw-rate distribution.**

certain level, unanswerable (Molnar et al., 2006; Allen, 2008; Armitage et al., 2011)? Squaring this circle requires us to test good model output with well-designed case studies where independent constraints on differing paleo-climates are as good as possible, so we can balance critically our knowledge of which climate factors matter with the level of data that is realistically obtainable. A key unknown is the extent to which landscapes are buffered against high frequency climate change, from 100-k.y. glacial cycles to shorter-term variation (Armitage et al., 2011), time periods much smaller than documented fluvial responses to tectonic perturbations, above. If topography forms a non-unique or difficult-to-decode record of past climate, it is likely that the sedimentary record, if and where complete, forms the best archive of landscape response to past climate (Allen, 2008; Whittaker et al., 2010; Armitage et al., 2011).

## THE FUTURE?

We cannot, yet, take a DEM, run a suite of algorithms to extract a set of topographic metrics, and instantly derive quantitative information on tectonic rates, in the absence of other geologic constraints. In fact, tectonics from topography in such a crude form is likely to be a false prospectus, while climate from topography is still in relative infancy (Whipple, 2009). But we are increasingly able to derive geomorphic constraints on tectonic rates from poorly known study areas because we have high resolution DEMs, we understand far better which bits of the landscape are sensitive to tectono-climatic forcing, and we better understand how steepness indices and knickpoints functionally record landscape response to tectonics and climate in steady-state and transient settings (Whittaker et al., 2008; DiBiase and Whipple, 2011). Field and dating studies have provided a great deal of information to calibrate LEM output and quantify landscape response times, allowing us to investigate the time-dependent landscape evolution for a range of boundary conditions (Tucker, 2009). Moreover, new topographic metrics are being developed that are complementary to methods based on river long profiles. For example, work by Hurst et al. (2012) suggests the use of topographic curvature on hilltops as a robust metric of erosion rates where soil production rates are high. Curvature would therefore provide a good proxy for local erosion (and rock uplift) rate, even if the whole landscape had not reached topographic steady-state, if hillslopes respond rapidly to transient fluvial downcutting (Hurst et al., 2012).

Further success will undoubtedly come from detailed assimilation of field and model-

ing approaches using well-constrained study sites where we can calibrate and quantify time-integrated landscape response to tectonics with respect to the full parameter range of climate, lithology, and erosion dynamics. If this parameter space were mapped in sufficient detail, the findings could be applied to study areas where ground conditions were similar but tectonic rates were unknown. Such an approach would enable us to convert the topographic metrics derived by Miller et al. (2012) explicitly to quantitative estimates of tectonic rates.

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