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Geomorphic constraints on landscape sensitivity to climate in tectonically active areas

Mitch D'Arcy *, Alexander C. Whittaker

Department of Earth Science and Engineering, Imperial College London, SW7 2AZ, United Kingdom

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ABSTRACT

The geomorphology of fluvial landscapes is known to record information about uplift rate, spatial patterns of faulting, and tectonic history. Data is far less available when addressing the sensitivity of common geomorphological metrics, such as channel steepness, to climatic boundary conditions. We test the relationship between channel steepness and precipitation rate by measuring a large number of channels in different mountainous areas. These regions exhibit a tenfold variation in precipitation rate between them (~100–1000 mm y^{-1}) but have similar uplift rates, allowing the tectonic variable to be controlled. By accounting for the orographic coupling of rainfall with uplifted topography, we find that channel steepness is significantly suppressed by higher precipitation rates in a measurable way that conforms to simple stream power erosion laws and empirical constraints on their parameters. We demonstrate this using modern and estimated glacial precipitation rates; and climate emerges as an important, quantifiable control on channel geometry. These findings help to explain why highly variable measurements of channel steepness are reported from different locations and provide important empirical constraints on how climate shapes tectonically active landscapes.

1. Introduction

1.1. Background

The geomorphology of the landscape represents the balance between processes creating and destroying topographic relief. Over geologic time, topography is produced by tectonic forces through the time-integrated effects of surface and rock uplift, subsidence, and crustal deformation (e.g., Whipple, 2004; Wobus et al., 2006; Allen, 2008; Whittaker and Boulton, 2012). The resulting relief is significantly modified by erosion: a process that is strongly controlled by climate. The form and evolution of a landscape is therefore determined by several fundamental controls, including tectonics, erosion, climate, lithology, and pre-existing geomorphology (Cowie et al., 2008; DiBiase and Whipple, 2011; Kirby and Whipple, 2012; Whittaker, 2012).

In principle, inverting the landscape for these boundary conditions in a quantitative way should be possible using theoretical, empirical, or modelling approaches (e.g., Kirby et al., 2003; Wobus et al., 2006; Attal et al., 2008; Tucker, 2009; Miller et al., 2012). This would enable us to convert geomorphological measurements into empirical information about uplift rates, spatial patterns of faulting, and landscape sensitivity to future climate change, among other useful insights (Whittaker, 2012). Deciphering the underlying equations that govern geomorphic form raises the prospect of using the landscape as a rich information archive about how tectonics, climate, and surface processes have evolved through time (Wobus et al., 2006; Allen, 2008; Whittaker et al., 2008; Kirby and Whipple, 2012). To achieve this goal, we need reliable measurements of landscape form that can be compared in time and space. A good example is the longitudinal geometry of river channels (e.g., Tucker and Whipple, 2002). Rivers are widespread, easily measured, and known to be patently sensitive to their tectonic and climatic boundary conditions (Whipple and Tucker, 1999; Snyder et al., 2000; Whittaker et al., 2008; DiBiase and Whipple, 2011; Whittaker, 2012). Stream power erosion laws describing channel geometry have been successfully applied to a range of tectonically active fluvial landscapes, and the way rivers transmit tectonic signals to the landscape has already been partially quantified (e.g., Tucker and Whipple, 2002; Whipple and Tucker, 2002; Crosby and Whipple, 2006; Whittaker et al., 2007a,b, 2008; DiBiase et al., 2010; Miller et al., 2012; amongst many). However, much less is known about how climate controls landscape form, and this remains a major challenge for geomorphologists (Wobus et al., 2010; Champagnac et al., 2012; Whittaker, 2012). The aim of this paper is to introduce climate into a common stream power erosion law and to test how well this approach fits real landscapes.

1.2. Previous work

As the longitudinal form of rivers responds to tectonic and erosional driving forces, it can capture the balance between processes creating and destroying topography (e.g., Whipple and Tucker, 1999; Kirby and Whipple, 2012). At a basic level, this balance can be described

* Corresponding author. Tel.: +44 20 7594 7491; fax: +44 20 7594 7444.
E-mail address: mitchall.darcy06@imperial.ac.uk (M. D'Arcy).

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geometrically. A starting point is a steady-state equation where the rate of change of elevation, dz / dt, is zero if the local rock uplift rate, U, is equal to the local erosion rate, E (e.g., Whipple and Tucker, 2002):

$$\frac{dz}{dt} = U - E. \tag{1}$$

The erosion rate in turn depends on other factors; and for fluvial systems it is often expressed as a stream power erosion law (Howard, 1994; Whipple and Tucker, 1999):

$$E = KA^m S^n \tag{2}$$

where K is a dimensionless erodibility coefficient that encapsulates lithology, climate, and transport processes; A is the drainage area raised to the power of an empirical constant m; and S is the local channel slope raised to the power of an empirical constant n (Hack, 1957; Seidl and Dietrich, 1992; Whipple and Tucker, 1999; Tucker and Whipple, 2002). Drainage area is typically taken as a proxy for discharge (Wobus et al., 2006). Given that discharge must also be a function of precipitation rate, P, averaged over a geomorphically relevant period of time, this can be incorporated into a stream power erosion law that has been used to describe varied landscapes (e.g., Whittaker et al., 2008; Whittaker and Boulton, 2012):

$$\frac{dz}{dt} = U - K(PA)^m S^n. \tag{3}$$

Assuming that uplift rate is counterbalanced by an equal erosion rate, i.e., the landscape is in 'topographic steady state', dz / dt in Eq. (3) is zero and the right-hand side can be rearranged to find the local slope S:

$$S = \left(\frac{U}{P^m K}\right)^{\frac{1}{n}} \cdot A^{-\frac{m}{n}}. \tag{4}$$

The exponent m/n is often represented by θ, the concavity index (Whipple and Tucker, 1999). The constant of proportionality between local slope and drainage area in Eq. (4) is termed the 'channel steepness index', k_s, i.e.,

$$k_s = \left(\frac{U}{P^m K}\right)^{\frac{1}{n}}. \tag{5}$$

Channel steepness is explicitly sensitive to rock uplift rate, U, and to precipitation rate, P (Wobus et al., 2006; Kirby and Whipple, 2012). Measuring k_s from a log–log plot of channel slope against upstream catchment area is simple, and it is usually normalised to a reference concavity index, θ_{ref} (Kirby and Whipple, 2012; Whittaker, 2012). Its units depend on the choice of this (dimensionless) concavity, e.g., $m^{0.9}$ for $\theta_{ref} = 0.45$. Normalisation facilitates k_{sn} comparisons between different rivers, locations, and studies (Kirby and Whipple, 2012), and if individual measurements of θ do not deviate widely from the local mean value, the effect of normalisation on k_s primarily reflects the scatter of the log–log plot (Snyder et al., 2000).

A significant quantity of research in the last 10 years has investigated the links between uplift and k_{sn} where climate (precipitation) gradients have either been neglected or assumed to be unimportant relative to the tectonic driver (c.f. Kirby et al., 2003; Boulton and Whittaker, 2009; Bookhagen and Strecker, 2012) These studies have successfully revealed a positive correlation between U and k_{sn}, albeit with some complexities in postulated functional form (e.g., Snyder et al., 2000; Kirby and Whipple, 2001; Ouimet et al., 2009; DiBiase et al., 2010). In some areas this correlation appears to be strongly linear (e.g., Lague and Davy, 2003), in others it appears sublinear (e.g., DiBiase et al., 2010), while in some places it has apparent linearity until a threshold uplift rate above which significant nonlinearity is observed. For example, in northern and southern Italy, Cyr et al. (2010) observed a threshold of nonlinearity at ~1 mm y^{-1} mean catchment uplift rate. More data is needed to shed further light on this relationship; we provide some in this study. Nevertheless, stream profile analysis is often compatible with independent erosion rate measurements (Ouimet et al., 2009; DiBiase et al., 2010) and has been used both to identify transient responses to tectonic change and to deduce rock uplift rates across a landscape (Kirby et al., 2003; Harkins et al., 2007; Whittaker et al., 2007b, 2008). Other work has shown that indices such as k_{sn} can capture fine spatial details in uplift rate. For example, in eastern Idaho, USA, Densmore et al. (2007) observed that channels near active normal fault tips are characterised by gentler slopes, while those nearer to the faults' along-strike centres are steeper and less concave.

Other workers have found that channel geometry is also sensitive to erosion rate, the other half of Eq. (1) (e.g., Finnegan et al., 2005; Whittaker et al., 2007a; Ouimet et al., 2009; DiBiase et al., 2010). Of particular relevance is a review of six different regional studies by Kirby and Whipple (2012) and references therein, which demonstrates a logarithmic (sublinear) scaling between normalised channel steepness and measured erosion rates for catchments in tectonically active ranges. In addition, channel steepness depends on a set of 'erodibility' variables including bedrock lithology and climate. This is likely why a wide range of k_s values have been reported from different locations when tectonics alone is considered to be the primary variable (Whittaker, 2012). One example that highlights this problem is a comparison of Boulton and Whittaker (2009) and Snyder et al. (2000). Boulton and Whittaker (2009) studied channels in southern Turkey where uplift rate has been well-constrained at 0.45 mm y^{-1}, and using $\theta_{ref} = 0.5$ they measure k_{sn} averaging 230 m and up to a maximum of 485 m. In contrast, Snyder et al. (2000) measured channels in northern California with uplift rates of 0.5 mm y^{-1} and k_{sn} values < 60 $m^{0.86}$ (the authors use a slightly lower reference concavity, $\theta_{ref} = 0.43$). This large difference in k_{sn}, despite equivalent uplift rates, cannot be explained by small differences in θ_{ref}. Both areas drain mixed sedimentary bedrock, so it is also unlikely to reflect lithological differences alone. The discrepancy may be partially climatic in origin: while the Turkish catchments receive ~800–900 mm y^{-1} rainfall, the California catchments receive significantly more at ~1500–1800 mm y^{-1} (Hijmans et al., 2005).

That a higher precipitation rate would correlate with reduced k_s values (uplift rate being constant) is clear within Eqs. (4) and (5). However, little empirical work has directly tested the relationship between channel steepness and precipitation rate. A global investigation by Champagnac et al. (2012) used digital elevation models (DEMs) to examine 69 mountain ranges and compared their gross geomorphology with tectonic and climatic variables. From a broad perspective, they found that the size and shape of mountainous landscapes is partly controlled by latitudinal variations in climate. This invites more detailed work examining the direct effects of climate on longitudinal channel form. Recently, Bookhagen and Strecker (2012) found that channels in the Argentinian Andes tend to have higher slopes where the precipitation rate is lower. They also found cosmogenic nuclide erosion rates decreasing by an order of magnitude in response to local aridification, following a wetter period 25–40 ka ago. Their data suggest that even in a tectonically active area, rainfall rate is an important control on river long profiles and that this control can be quantified. However to date, no precipitation signal has been explicitly extracted from k_{sn} data gathered in disparate study areas, suggesting that there is some way to go before we have a complete understanding of how channel steepness records climatic and tectonic boundary conditions together.

However, theoretical studies have already emphasised the influence of precipitation rate on channel form (Sólyom and Tucker, 2004; Wobus et al., 2006, 2010; DiBiase and Whipple, 2011). For instance, DiBiase and Whipple (2011) used a numerical model calibrated with erosion rate constraints from the San Gabriel Mountains in California to demonstrate that the relationship between erosion rate and channel steepness is controlled by river discharge and, furthermore, can be strongly nonlinearized