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Climate and the Pace of Erosional Landscape Evolution

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Abstract Earth's climate touches nearly all aspects of landscape evolution, from the breakdown of rock to the delivery of sediment to the oceans. Yet quantifying climate's influence on landscapes is a major challenge, not only because it is difficult to know how landscapes responded to past changes in climate, but also because landscapes are shaped by various processes that respond to climate in different ways. I review the current state of efforts to quantify climate's effects on the rates of the main processes that drive landscape evolution, with a focus on unglaciated landscapes formed by bedrock erosion. Although many uncertainties remain, recent research has clarified how the processes governing hillslopes, bedrock channels, and chemical erosion depend on major climate factors such as precipitation and temperature. A few themes emerge, including the importance of climatically mediated biological processes, the role of variability, and the value of natural experiments for revealing climate's effects.

1. INTRODUCTION

The diversity of Earth's landscapes is intimately tied to climate. From the dunes of wind-swept deserts to wide rivers snaking through jungle, climate is clearly one of the most important factors that shapes Earth's surface, along with volcanism, tectonic deformation, life, and the material properties of Earth's crust. There are both societal and geological reasons to study climate and landscape evolution. We are inherently curious about the landscapes we live on. We depend on them for settlement, water, agriculture, and biological and mineralogical resources. And we are threatened by the hazards they create, from floods and landslides to famine and disease. Understanding how climate shaped the landscapes we see today must be a part of any strategy for managing landscapes in the present and anticipating how they might change under future climates.

Over geological time, the connection between climate and landscapes plays a key role in the Earth system. Consumption of atmospheric CO_2 by chemical weathering of silicates is assumed to regulate Earth's long-term climate (Berner et al. 1983). Changes in topography can alter atmospheric circulation and climate at a continental scale (Ruddiman & Kutzbach 1989; Strecker et al. 2007). Models coupling climate, topography and tectonics show that a strong connection between climate and erosion can in theory have enormous consequences for the evolution of Earth's surface topography and lithosphere, even controlling the size and shape of entire mountain ranges (Beaumont et al. 1992; Willett 1999; Whipple & Meade 2004; Roe et al. 2006; Whipple 2009). Landscapes can also record climate change through mechanisms ranging from signals preserved in river profiles (Whipple & Tucker 1999; Royden & Perron 2013) to sedimentary deposits generated by ancient storms (Cook et al. 2015).

Climate's influence on landscapes is at some level intuitive – anyone can see that rainfall feeds rivers that carve valleys – and in some places the long-term outcomes of such connections are clearly apparent (Figure 1a). But generalizing and quantifying climate's effects on landscape evolution is not as straightforward as it might appear. For example, the notion that wetter average climates make rivers cut faster through rock, which drives some of the coupled models mentioned above, appears to be correct in some cases, but recent research reviewed below predicts a more nuanced relationship. Even effects that can be measured under controlled laboratory conditions, such as the temperature dependence of silicate weathering (Kump et al. 2000) can be difficult to demonstrate in nature.

1.1. Challenges

Efforts to understand how climate drives landscape evolution face fundamental challenges. Climate has varied dramatically over Earth's history, and substantial changes – from Cenozoic cooling to Pleistocene glacial cycles – occurred as the landscapes we see today evolved to their present states (Figure 2). If we are to understand the origins of modern landforms, we must understand how the processes that formed them respond to time-varying climate forcing. Indeed, many studies have asked how erosion rates varied throughout these dramatic climate changes, some with the goal of understanding how the silicate weathering feedback maintains Earth's habitability (Berner et al. 1983). For example, it has alternately been proposed that late Cenozoic cooling (Figure 2) led to dramatic changes in erosion (Molnar & England 1990; Zhang et al. 2001; Molnar 2004), or that increased erosion due to mountain building during this time period caused the cooling (Raymo & Ruddiman 1992).

But the overall influence of climate on landscape evolution cannot be calibrated simply by comparing climate variables with erosion rates. Efforts to do this with techniques that measure erosion rates over various timescales have yielded some positive results, but with an equal measure of ambiguity. Million-year rates of marine sediment deposition (Zhang et al. 2001; Molnar 2004) and million-year exhumation rates from low-temperature thermochronology (Herman et al. 2013; Herman & Champagnac 2016) have been interpreted as evidence that late Cenozoic cooling accelerated average continental erosion rates, but alternate interpretations of the sedimentary record (Willenbring & Jerolmack 2016) and other geochemical measures of million-year weathering rates (Willenbring & von Blanckenburg 2010) show no such trend. Erosion rates from cosmogenic isotopes in minerals (Lal 1988; Bierman & Nichols 2004), which typically average over a few thousand to a few hundred thousand years, have been measured in many locations worldwide. Although these rates do show an influence of climate (Bookhagen & Strecker 2012), they clearly also depend on other variables (Figure 3) (von Blanckenburg 2005; Portenga & Bierman 2011), and some studies find no relationship at all (Burbank et al. 2003; von Blanckenburg 2005). Yields of suspended sediment in rivers, which provide an estimate of erosion spanning days to years averaged over the upstream drainage basin, generally show no clear relationship with mean annual precipitation, except that erosion rates in the driest landscapes increase with precipitation (Langbein & Schumm 1958; Walling & Webb 1983; Riebe et al. 2001a and references therein). However, runoff and temperature do appear to explain a small fraction of the variance in sediment delivery to the oceans by major rivers (Syvitski & Milliman 2007).

Thus, beyond the consensus that erosion is generally slow in very dry landscapes, and despite examples of spatial correlations between precipitation and erosion rates in some locations (e.g.

Reiners et al. 2003), there appears to be no direct, general relationship between climate and erosion rate. One reason is that erosion rates also depend on other factors such as bedrock type and strength, tectonic history, and biological processes. Controlling for some of these factors suggests that climate does influence erosion rates (Portenga & Bierman 2011), but the correlation appears to be weak (Figure 3) and easily obscured (von Blanckenburg 2005). A second reason is that erosion is the net effect of many processes that erode rock and transport sediment, and the blend varies from region to region and through time. A third reason is that those various erosion and transport mechanisms depend on different aspects of climate.

1.2. Scope and Purpose

The purpose of this review is to summarize progress on understanding and quantifying how the rates of erosional processes that drive landscape evolution depend on major climate variables such as precipitation and temperature. Climate touches nearly all landscapes, but I do not attempt to discuss them all. To achieve a useful level of scrutiny, I focus on unglaciated landscapes above sea level that are shaped by the erosion of rock. Glacial processes (Hallet et al. 1996) and coastal landforms (Allan & Komar 2006; FitzGerald et al. 2008) are left to other reviewers.

2. PROCESS LAWS

Most quantitative models of landscape evolution combine conservation of mass equations for rock and sediment with rate laws for individual mass sources, sinks, and transport processes. The conservation equations have the general form

$$\frac{\partial(\rho_r b)}{\partial t} = \rho_r U - \rho_r P - \rho_r E - C_r \tag{1}$$

$$\frac{\partial(\rho_s h)}{\partial t} + \nabla \cdot (\rho_s \mathbf{q}_s) = \frac{\rho_r}{\rho_s} P + \rho_s A_s - C_s\ , \tag{2}$$

where b is the elevation of the bedrock surface relative to a datum, h is the thickness of mobile soil or sediment, ρ_r is the density of rock, ρ_s is the density of soil or sediment, U is the rate of rock uplift, P is the rate of bedrock erosion due to the production of soil or sediment, E is the rate of mechanical bedrock erosion by all other mechanisms, C_r and C_s are the rates of net chemical erosion of rock and soil, $\mathbf{q}_s$ is the net vector flux of sediment or soil, and A_s is the deposition rate of atmospheric aerosols, which is usually negligible in all but the slowest-eroding landscapes and is subsequently ignored here. The elevation of the land surface is $z = b + h$. In this formulation, "bedrock" includes both unweathered rock and immobile saprolite. All terms in Equations 1 and 2 have units of mass per unit horizontal area per time.

Most of the studies reviewed here either constrain the form of the expression used to represent one of the processes on the right-hand side of Equation 1 or 2, or present field or experimental observations that determine how climate influences the parameterizations of these processes. Mathematical descriptions of surface processes are a topic of active research (Dietrich et al. 2003) and are therefore subject to frequent revision or rejection. The expressions I discuss here should not be construed as definitive; rather, they represent the current state of knowledge and provide a useful framework for considering climate's effects on surface processes. Table 1 summarizes the ways in which climate affects the processes and process laws discussed in the following three sections on channels, hillslopes, and chemical erosion.

3. CHANNELS

Channelized flows passing over rock erode their beds, carving valleys that set the base level for the surrounding slopes, and transport away sediment shed by those slopes. Although channels occupy only a small fraction of a landscape's area, channel incision by rivers and debris flows is one of the most important drivers of landscape evolution.

3.1. Bedrock Rivers

Process laws for bedrock river incision have been reviewed elsewhere (Whipple & Tucker 1999; Whipple 2004; Lague 2014), including a brief overview of climate's influence (Whittaker 2012). The most common erosion law, with the most empirical support, is (Howard & Kerby 1983; Howard 1994)

$$E_r = \begin{cases} k_1 f(q_s)(\tau - \tau_c)^a & \tau > \tau_c \\ 0 & \tau \le \tau_c \end{cases}, \tag{3}$$

with

$$\tau = k_2 \left(\frac{Q}{w}\right)^{\alpha} S^{\beta}, \tag{4}$$

where k_1 is a coefficient that depends on rock strength, $f(q_s)$ is a function of the magnitude of sediment flux in the channel, τ is the bed shear stress, τ_c is a threshold shear stress for abrading the bed or plucking fractured blocks from the bed, a is a constant, Q is river discharge, w is river channel width, S is water surface slope, commonly approximated by the bed slope, and k_2, α and β are constants that depend on flow resistance. Channel width w generally scales as a power function of Q (Montgomery & Gran 2001; Wohl & David 2008), which in turn scales as a power function of upstream drainage area A. Using these relationships, and assuming that $f(q_s) = 1$ and that the erosion threshold τ_c has a negligible effect on long-term incision rates, the rate law for bedrock river incision is commonly approximated as

$$E_r = KA^m S^n \,, \tag{5}$$

where K is a coefficient and m and n are constants. This rate law is usually referred to as the "stream power law" because the assumption that erosion rate scales with the rate of energy expenditure by the flow as opposed to shear stress (Equation 3) leads to an equation with the same form as Equation 5 (Seidl & Dietrich 1992; Whipple & Tucker 1999).

The dependence of τ on water discharge suggests that more rain should make rivers erode faster. Field measurements confirm that a higher average precipitation rate does indeed make rivers erode faster in some cases, but theoretical arguments and recent field measurements suggest that both the sign and the strength of the relationship between E_r and precipitation rate might depend on the variability of runoff.

3.1.1. Effect of Average Precipitation To first order, and neglecting losses to groundwater and evapotranspiration, conservation of water volume requires that discharge Q is proportional to the precipitation rate multiplied by the drainage area A. Precipitation rate should therefore have a strong influence on the coefficient of the function relating Q to A, which is subsumed within K. This implies that K is not a constant, but should vary with precipitation rate. However, the difficulty of independently measuring K while controlling for other important factors such as bedrock erodibility has made this clear expectation challenging to test. Ferrier et al. (2013a) compared bedrock rivers on the Hawaiian island of Kauai, which has uniform basaltic lithology and a trade wind-driven orographic rainfall gradient that spans roughly 70% of the range of mean annual precipitation rates on Earth over just 25 km (Figure 1a). They found that K for Kauai's rivers increases as a power function of upstream-averaged mean annual precipitation (Figure 1b). They also found a linear correlation between long-term river incision rates and stream power (n = 1 in Equation 5), computed with a measured dependence of Q on upstream-averaged precipitation, consistent with studies of basin-averaged erosion rates spanning precipitation gradients (Bookhagen & Strecker 2012). Their analysis shows that, in at least some cases, wetter climates make rivers erode bedrock more efficiently.

3.1.2. Effects of Thresholds and Discharge Variability It is likely that some bedrock rivers only erode their beds during large floods. Most rivers eroding bedrock contain sediment that partly or completely covers their beds during low flow conditions, and this sediment must be mobilized before the underlying rock can erode. Moreover, the underlying rock usually has some strength that small flows may be too weak to overcome. Both effects should create a threshold for bedrock erosion by rivers (Snyder et al. 2003). As expressed in Equation 3, the erosion threshold τ_c has two effects: it suppresses erosion by small flows that do not generate enough shear stress to overcome the threshold, and it reduces the erosion accomplished by flows that do exceed the threshold. The fact that some landscapes show simple correlations between average precipitation and river incision rates despite these effects (Bookhagen & Strecker 2012; Ferrier et al. 2013a) indicates that Equation 5, which neglects the threshold, can be a useful approximation. In other landscapes, however, storms of varying magnitude and frequency may interact with the nonlinear dependence of incision rate on Q expressed in Equations 3 and 4 to complicate the relationship between precipitation rate and river incision.

The case of rivers that build their own channels by depositing sediment offers a useful analogy articulated by Wolman and Miller (1960): small flows are frequent but move little sediment; large flows are rare but move much sediment; and the cumulative effect on the channel is the integrated product of the flood frequency distribution and the function relating sediment transport to flow magnitude. A number of studies have applied the same concept to bedrock river incision. Tucker & Bras (2000) and Tucker (2004) developed a model in which shear stress- or stream power-dependent river incision is driven by a stochastic distribution of floods, and demonstrated that the effect of variable precipitation can be more important than the mean. Snyder et al. (2003) calibrated and applied this modeling framework to a landscape in the California Coast Ranges that they had previously studied in the context of Equation 5 and found that the effect of variable discharge can explain river longitudinal profiles over a range of uplift rates and precipitation regimes better than the simplified model. The long-term bedrock incision rate of a river that experiences a probability density distribution of discharges $p(Q)$ can be written as (Lague et al. 2005; DiBiase & Whipple 2011)

$$E_r = \int_0^{Q_{max}} I(Q, \bar{Q})\, p(Q) dQ \,, \tag{6}$$

in which Q is taken to be the discharge over a time interval comparable to flood duration, typically the daily discharge; $\bar{Q}$ is a reference discharge, typically the mean daily discharge; $Q_{\max}$ is the maximum Q over the time period of interest; and I is a function that gives the instantaneous incision rate. In most studies using this formulation, I takes on a form similar to the rate law in Equations 3 and 4, with modifications to account for variations in channel width with varying Q that introduce the dependence on $\bar{Q}$ (Lague et al. 2005; DiBiase & Whipple 2011). One of the main predictions of this probabilistic model is that rivers should not only incise more efficiently in wetter climates, they should also incise more efficiently in climates with more variable runoff (Tucker 2004; Lague et al. 2005).

Rainfall events typically follow an exponential distribution (Eagleson 1978; Tucker & Bras 2000; Tucker 2004), but discharge records for rivers indicate that larger events follow a power-law tail (Davy & Crave 2000; Molnar et al. 2006; Malamud & Turcotte 2006), suggesting that $p(Q)$ is better approximated by an inverse gamma distribution (Figure 4a) (Davy & Crave 2000; Lague et al. 2005). The exponent describing the power-law tail, $-(2+k)$ in Lague's (2005) terminology, can be interpreted as a measure of climate variability, with smaller k corresponding to a heavier tail of large floods and therefore more variability (Turcotte & Greene 1993; Lague et al. 2005; Molnar et al. 2006). Molnar et al. (2006) showed that k is typically between 0.01 and 5 for a set of stream gauge records from across the US, and that flows tend to be less variable (larger k) as annual runoff increases.

DiBiase & Whipple (2011) combined this observed correlation between mean runoff and discharge variability with the probabilistic discharge model (Equation 6), calibrated it to basin-averaged erosion rates in a study area in the San Gabriel Mountains, California, and showed how increases in average runoff could trade off with less frequent extreme floods to determine long-term bedrock incision rates. They found that discharge variability and an erosion threshold can make the long-term channel incision rate less sensitive to average runoff, with the blunting of the relationship more pronounced when the threshold is higher, or when discharge variability decreases more sharply with increasing average runoff (Figure 4b). Their results also show that steeper rivers – generally, rivers with a higher ratio of uplift rate U to erosion coefficient K – are more likely to erode faster with increasing average runoff, suggesting that average climate may still influence river incision rates in the most tectonically active landscapes despite the blunting of the erosion-runoff relationship.

Do these results imply that bedrock river incision is generally insensitive to changes in runoff? Lague (2014) used a compilation of data from rivers in tectonically active environments to argue that the erosion threshold's influence on long-term erosion is important. It is plausible that there are scenarios in which increased mean discharge under wetter conditions would be offset by less variable discharge, with only a minor net effect on long-term channel incision rate. But the stochastic discharge model and runoff-variability scaling have been calibrated in too few places to give a general picture of bedrock rivers' sensitivity to average runoff, partly because it is very difficult to measure erosion thresholds. A complete model of climate's effects on river incision must also be able to relate precipitation and runoff to the long-term distribution of river floods via drainage basin hydrology and ecology (Lague 2014). Efforts to explore this connection have found that mean annual precipitation is a surprisingly useful proxy for flood magnitude and frequency in basins where large floods dominate river incision (Rossi et al. 2016).

3.1.3. Other Climatic Effects on River Incision Three other potential climatic effects on bedrock river incision deserve mention. First, nearly all current models depend on empirical expressions relating channel width w to discharge Q, which were originally recognized in alluvial channels (Leopold & Maddock 1953) and subsequently found to apply to bedrock channels (Montgomery & Gran 2001; Wohl & David 2008). Although there is evidence that these expressions can capture width variations across landscapes with spatially variable precipitation (Craddock et al. 2007), they must typically be calibrated for each site, and the lack of a mechanistic channel width law could obscure important climatic effects, including the relative importance of average flows and discharge variability discussed in Section 3.1. This is an active area of research (e.g. Finnegan et al. 2005; Finnegan et al. 2007; Nelson & Seminara

2011), but it is too soon to assess how channel width mechanics might influence the connection between climate and bedrock river incision.

Second, bedrock river incision can depend on the river's sediment load: grains carried by the flow act as tools that impact and abrade the bed, but when grains cover the bed they armor it against erosion. At the simplest level, these competing effects, which have been subjected to much recent study (e.g. Sklar & Dietrich 2001; Turowski et al. 2007), suggest that the function $f(q_s)$ may have a maximum at an intermediate value of q_s. River discharge influences a river's capacity to transport its sediment load, and this influences the balance of the tool and cover effects (and in the extreme case of cover, whether the river erodes bedrock at all). However, the different topographic outcomes of river incision models that incorporate sediment flux are most apparent during the transient response of rivers to changes in forcing (e.g. Gasparini et al. 2007; Attal et al. 2011), and efforts to constrain climatic influences on sediment flux-dependent river incision with field measurements are in their early stages. The efficiency of bedrock river incision may also depend indirectly on climate via the size distribution of grains shed by hillslopes (Sklar et al. 2016; see Section 4.1).

Third, the rate coefficient for bedrock river incision, k_1 or K, depends on rock erodibility. As I discuss in more detail in Section 5, there is abundant evidence that climate influences chemical weathering rates, and recent field measurements suggest that chemical weathering can affect rates of river incision in certain settings (Han et al. 2014; Murphy et al. 2016).

3.2. Debris Flows

In many steep landscapes, the uppermost reaches of the channel network are incised by brief, rapid flows of grain-fluid mixtures known as debris flows (Stock & Dietrich 2003; Stock &

Dietrich 2006). As with bedrock rivers, this introduces the challenge of modeling the effective long-term rate of an episodic process governed by short-term dynamics. Despite progress in understanding the complicated mechanics and rheology of debris flows (Iverson 1997), the challenging task of formulating a long-term debris flow erosion law remains an active area of research.

Field observations indicate that debris flows erode mainly through collisions of large particles with the bed, suggesting that the erosion caused by a debris flow should scale with the bulk collisional normal stresses of the large grains (commonly boulders) in the flow (Stock & Dietrich 2006). Stock & Dietrich (2006) expressed this hypothesis as an event-based erosion law:

$$E_d = K_d f_d L \left[\rho \nu {D_p}^2 \left(\frac{u}{h_d}\right)^{\gamma} \cos\theta\right]^{\eta}, \tag{7}$$

where E_d is debris flow erosion rate, K_d is a coefficient that depends on rock properties and the distribution of flow impact stresses, f_d is debris flow occurrence frequency, L is the length of debris flow that erodes bedrock, ρ is particle density, ν is the volume fraction of solids, D_p is an effective particle diameter, u is debris flow surface speed, h_d is debris flow depth, u/h_d is an approximation for shear rate, θ is bed slope angle, and γ and η are constants. They also generalized this expression into a long-term erosion law that accounts for the observations that debris flows typically originate at the heads of valleys as shallow landslides or raveling events and scour bedrock as they travel down networks of converging valleys, stopping once they reach a slope sufficiently gentle that their internal shear slows and they no longer behave like a fluid.

Hsu et al. (2008) studied debris flow incision through experiments in which miniature granular slurries in a cylindrical drum rotating on a horizontal axis passed repeatedly over a panel of synthetic, erodible rock. Their results are broadly consistent with the basic form of

Equation 7. Field measurements of grain impact forces and bedrock erosion during debris flows (McCoy et al. 2013) and larger laboratory experiments (Hsu et al. 2014) support the idea that large grains colliding with the bed cause most of the erosion and underscore the importance of grain size, which may be a strong function of climate (see Section 4.1).

Relating climate to bulk debris flow properties like those in Equation 7, and testing whether rate laws based on bulk flow properties can predict long-term erosion rates, is an outstanding challenge. It seems likely that, like K in Equation 5, K_d increases with chemical weathering of the bed; that f_d scales with landslide frequency, discussed in the context of climate in Section 4.4.1; and that climatic effects on drainage density (the total length of channels per unit area of the landscape) influence f_d by controlling the number of upstream source areas (Perron et al. 2009, 2012; Chadwick et al. 2012). Climate-influenced soil, vegetation and landslide characteristics (see Section 4) may affect L, ν, D_p, u and h_d, but understanding the specific effects will require more field and experimental studies of debris flow incision.

4. HILLSLOPES

Hillslopes occupy most of a landscape's area, and therefore most bedrock erosion occurs on hillslopes. I divide this review of climate's effects on hillslope processes into soil production from bedrock and the downslope transport of that soil by creep, overland flow, and mass movement (principally landslides). Chemical erosion is treated separately in Section 5.

4.1. Soil Production

Numerous mechanisms contribute to the mechanical breakup of bedrock that produces soil (defined here as a layer of mobile material atop structurally intact rock), including animal and insect activity, plant rooting, and the growth of mineral or ice crystals. It has long been suggested that as soil thickens, the underlying bedrock-soil interface is less affected by these near-surface mechanisms, and the rate of soil production slows (Gilbert 1877). Heimsath et al. (1997, and subsequent studies) confirmed this with geochemical measurements of sub-soil bedrock erosion rates, and found empirical support for a negative exponential dependence of soil production on thickness,

$$P(h) = P_0 e^{-h/h^*}, \tag{8}$$

where P_0 is the soil production rate when $h = 0$ (that is, on bare bedrock) and h^* is the e-folding depth of the soil production rate. Alternatively, Gilbert (1877) proposed that bare bedrock should erode more slowly due to the lack of biological disturbance and the absence of porous soil to retain reactive water. This idea is more difficult to test, but there is empirical support in some sites for soil production rates peaking at non-zero soil depth (Heimsath et al. 2009). The transition between bare bedrock and soil-mantled topography is a subject of active research (Heimsath et al. 2012). Less is known about the mechanical processes that erode bare bedrock hillslopes, and I focus here on soil-mantled slopes.

Measurements of soil production rates around the world are beginning to reveal climate's effects. Many initial studies were in relatively temperate landscapes, making it difficult to discern climate-related trends, but by examining sites in a range of climate zones across Australia, Heimsath et al. (2010) found that soil production rates (P) generally increased with mean annual precipitation (MAP) and decreased with mean annual temperature (MAT). Others have proposed related empirical soil production functions that depend explicitly on MAP and MAT (Pelletier & Rasmussen 2009; Norton et al. 2014).

Studies of very dry landscapes have made the climate dependence of soil production even more apparent. Fieldwork by Owen et al. (2011) in the Atacama Desert, combined with a compilation of soil production rates from the literature, confirmed that P increases with MAP. Additional consideration of soil thickness variations in their dataset suggests that P_0 increases as a power law of MAP, with an exponent less than 1 (Figure 5a). Based on their Atacama measurements, Owen et al. (2011) suggested that P may be independent of h in very dry sites, suggesting that h^* may decrease with MAP, but currently available data are insufficient to resolve such a trend.

The size distribution of rock fragments in soil affects the mechanics of soil production and transport on hillslopes as well as sediment transport and bedrock incision in rivers. There have been few attempts to characterize or model how the grain size delivered from hillslopes to channels varies systematically among landscapes, let alone how it depends on climate. But efforts to develop a process law for soil grain size are underway. Sklar et al. (2016) present a model of soil grain size evolution and suggest that the balance between mechanical and chemical erosion on hillslopes, a topic discussed below in Section 5, should be one of the main factors governing the evolution of grain size as soils traverse hillslopes. Field studies have shown that grain size varies with elevation, a surrogate for temperature, in high-relief alpine watersheds, with coarser size distributions found in higher, colder locations (Riebe et al. 2015). In a sequence of soils spanning a steep orographic rainfall gradient on the Island of Hawaii, rock fragments coarser than sand are much less abundant in soils that receive more rainfall (Chadwick et al.

2003; Marshall & Sklar 2012). Continued efforts will help generalize these relationships, but it seems likely that climate has predictable effects on soil grain size.

4.2. Soil Creep

Soil is transported down hillslopes by a combination of gravity and mechanical disturbance mechanisms that include abiotic phenomena, such as raindrop impacts, and biological phenomena, such as animal burrowing, root growth and the toppling of trees. Much evidence (see references in Perron 2011) supports a transport law with the form

$$\mathbf{q}_s = -D\nabla z \, , \tag{9}$$

where D is a coefficient with the same units as a diffusivity, length^2/time. In some landscapes, the soil flux also appears to depend on soil thickness, probably because the disturbance mechanisms are most active close to the surface. This has led some to suggest an alternate form of the transport law (Heimsath et al. 2005), $\mathbf{q}_s = -K_c h \nabla z$, where K_c has units of length/time. On very steep slopes, shallow landslides and raveling of grains can also move soil, with the rate of transport by these mechanisms increasing as the slope approaches a critical steepness S_c that depends on local conditions. In such landscapes, measurements support a transport law that depends nonlinearly on slope (Roering et al. 1999), $\mathbf{q}_s = -D\nabla z/[1-(|\nabla z|/S_c)^2]$. However, the coefficient D in the numerator has the same dimensions as in the linear transport law, and topographic outcomes of the two laws on gentle slopes well below S_c are very similar.

Many studies have estimated D with various techniques, making it possible to assemble compilations that span a wide range of climates, so I focus on this parameter. There are relatively few estimates of K_c and S_c, although some likely climatic effects on these alternate forms of the transport law are apparent (Table 1). The most extensive compilation of D estimates to date is that of Richardson (2015), which includes an earlier compilation by Hurst et al. (2013) and the measurements of Callaghan (2012) and Owen (2009). No single variable explains all the variability of D, but the data do reveal a clear dependence of D on MAP (Figure 5b).

Over the broad range of moisture in Figure 5, differences in P_0 and D probably reflect the extent to which life can gain a foothold (Gilbert 1877; Gabet et al. 2003; Dietrich & Perron 2006). Organisms have basic liquid water requirements, and some key measures of biological activity, such as primary productivity, scale with average precipitation (Field et al. 1998). In very arid landscapes where life is sparse, it does not appear to contribute substantially to the damage, disaggregation or displacement of rock (Owen et al. 2011), and the remaining mechanisms that produce and transport soil, such as salt weathering (Owen et al. 2011) and rainsplash (Dunne et al. 2010) can be slow or infrequent. Among landscapes with more abundant life, ecological differences may be important. Callaghan (2012) noted that differences in vegetation type correspond to transitions in the magnitude of D among her sites in Chile. Richardson's (2015) compilation offers some support for this idea: among landscapes with similar MAP, there are significant differences in mean D between landscapes with very different vegetation types (Figure 5b).

Is the biologically mediated influence of climate on soil production and transport monotonic, or does it plateau once life gains a strong foothold? The trends in Figures 5a and 5b would not be as apparent if only the data with MAP > ~0.5 m/yr were considered, an observation replicated in some studies of single regions (Callaghan 2012). Owen (2009) notes that there may be a steep drop-off in biological productivity for MAP < ~0.1 m (Lieth 1973). Under these dry conditions, the efficiency of soil production and transport may be limited by the type and amount of biological activity; but under conditions wet enough to support abundant life, variables other than moisture may dictate P_0 and D (Owen et al. 2011). Further increases in moisture may change a landscape's ecology without causing more efficient production and transport of soil, either because niches for bioturbating organisms are equally occupied in landscapes with intermediate and high moisture, or because additional biological mechanisms that impede production and transport, such as reinforcement by plant roots, become more important (Owen et al. 2011; Richardson 2015).

4.3. Overland Flow

Hillslope soils that receive rainfall or snowmelt that exceeds their infiltration capacity can experience sheetflow that entrains soil particles (Horton 1945). Many short-term models of overland flow erosion have been developed, especially in the domain of soil conservation (Renard et al. 1997). These largely empirical models do incorporate climate factors, but no widely applicable long-term process laws have been developed (see discussion in Dietrich et al. 2003). Despite the lack of a process law, some insights about climate and overland flow from short-term studies probably apply to long-term erosion as well. Erosional efficiency likely peaks in dry climates with large, rare storms (Evans et al. 2000), especially in conditions dry enough to render soils hydrophobic. Vegetation can inhibit erosion by enhancing infiltration, adding roughness, and binding soil with roots (Prosser et al. 1995; Istanbulluoglu & Bras 2005). Wetter average conditions can bring soils closer to saturation, promoting overland flow during subsequent storms; but wetter conditions also favor vegetation growth (Prosser et al. 1995). Field studies of overland flow in a geomorphological context have generally focused on short-term behavior (e.g. Prosser & Dietrich 1995), but studies of multi-decadal effects of overland flow on topography are now inspiring the development of longer-term process models (e.g. Geng et al. 2015).

4.4. Mass Movement

Numerous other mechanisms transport soil and rock downslope in steep landscapes. I focus on the most common mechanisms relevant to long-term landscape evolution. Because many mass movement processes are episodic, theoretical treatments have generally focused on event-based models. These models are a useful starting point for developing long-term process laws, and they suggest some important climate controls.

4.4.1. Shallow Landslides Shallow landslides occur when the downslope gravitational forces on a layer of soil or rock exceed the resistive forces from friction and cohesion at the base and margins of the layer. The failure surface commonly parallels the land surface and usually is located at the interface between bedrock and mobile soil. Rainfall and snowmelt are common triggers of shallow landslides because they increase soil pore water pressure, reducing the effective normal force, and therefore also the frictional force, on the failure surface. Landscape-scale models, which attempt to calculate the stability of all the hillslopes in an area, typically capture this effect by expressing landslide susceptibility in terms of the intensity of infiltrating precipitation relative to the soil's ability to convey subsurface flow (Montgomery & Dietrich 1994).

The fact that landslides occur at a particular time during or after storms implies that the time dependence of precipitation is also important. In general, landslides are more likely to occur if rainfall is more intense, prolonged, and frequent. Many studies have defined empirical thresholds of rainfall intensity, duration and frequency for landslide triggering in different regions (e.g. Caine 1980; Guzzetti et al. 2008). These observations can help constrain how rainfall characteristics influence the triggering mechanism in more mechanistic models, as well as providing a basis for probabilistic models of landslide occurrence over longer time intervals (Moon et al. 2011).

Another challenge in modeling climate's effect on landslides is determining the importance of average or seasonal precipitation. Soils saturated by antecedent precipitation can fail during smaller storms, and are also more susceptible to other triggering mechanisms such as earthquakes. Landscape-scale landslide models that incorporate both background saturation and infiltration driven by rainfall time series (Iverson 2000) are usually applied over the timescales of individual storms (Salciarini et al. 2006), but are now beginning to be applied over longer time intervals relevant to landscape evolution (Rosso et al. 2006; Bellugi et al. 2015).

One of the largest challenges of calibrating precipitation's influence on long-term landslide occurrence is the unknown legacy of past landslides. This is one reason why landscape-scale models tend to over-predict landslides: slopes that fail are less likely to experience a repeat failure until soil thickness has recovered, but it is difficult to locate past failures, even in the field, or to know when they occurred. Montgomery et al. (2000) mapped landslides that occurred during large storms over a 10-year interval in a 0.43 km^2 area of the Oregon Coast Range, USA. They found that nearly half of the landslides occurred during a storm with intermediate rainfall intensity, whereas a subsequent storm that delivered the most intense rainfall on record in the region triggered no more landslides than some of the smallest storms in their dataset. Although this counterintuitive result is partly a consequence of human activities – timber harvesting shortly before the smaller storms made some parts of the study area more susceptible to failure during those storms – it nonetheless demonstrates the complicating influence of past landslides on the relationship between climate and slope failure.

Montgomery et al. (2000) also speculated that the time between the most recent timber harvest and the largest storm may have been long enough to allow substantial regrowth of vegetation roots that strengthened the soil. The effective cohesion supplied by vegetation, particularly the deep roots of trees, is another well-documented control on landslide occurrence that depends on climate (Schmidt et al. 2001; Sidle & Ochiai 2006). Vegetation can also reduce soil saturation by increasing hydraulic conductivity, making soils drain faster, and through transpiration, which removes water from the soil and delivers it to the atmosphere. The influence of precipitation on soil production and creep (Figure 5), which refill landslide scars with soil, is another reason to expect that landslides can occur more frequently in landscapes with wetter average climates, even if the distribution of large storms is the same.

Storm characteristics can influence landslide size and location in addition to frequency of occurrence. Bellugi et al. (2015) used a model that predicts the boundaries of individual landslides to simulate the response to the sequence of storms studied by Montgomery et al. (2000). They showed that more intense rainfall generally triggers larger landslides that occur farther downslope, consistent with the observations.

4.4.2. Creeping Landslides and Earthflows Shallow landslides typically move rapidly, and can completely evacuate their scars as they transform into debris flows. Other mass movement processes involve more gradual, yet still substantial, displacements that also depend on climate. Creeping landslides have been observed to move faster during or after rainy periods (Hilley et al. 2004) and slower during droughts (Bennett et al. 2016), with deeper slides showing more sensitivity to longer-term climate (Bennett et al. 2016). These measurements suggest that pore water pressure governs the climatic influence. Efforts to develop long-term process laws for creeping landslides are off to an encouraging start (Booth et al. 2013), and more field and geodetic observations will help incorporate climate into these models.

4.4.3. Raveling Although connections between climate and dry raveling of loose debris might not be obvious, the effects of fire on raveling illustrate how climate's influence on mass movement extends beyond precipitation. After a fire destroys vegetation, reducing surface roughness and root cohesion that can hold soil particles in place, raveling rates can increase tremendously (Roering & Gerber 2005), especially when fire is followed by a wet season (Anderson 1959; Lavé & Burbank 2004). The multiple roles of vegetation in fire-prone landscapes are beginning to be incorporated into process models that predict raveling rates (Gabet 2003; Roering & Gerber 2005; Lamb et al. 2011).

5. CHEMICAL EROSION

Chemical weathering is the alteration of minerals in rock or soil through chemical reactions, which usually involve reactants in air or water. Chemical erosion is net mass loss due to the removal of reaction products, usually in solution. However, the term chemical weathering is often used to encompass both mineral alteration and mass loss, with the assumption that the dissolved products are indeed transported away. At the scale of major river basins, mechanical erosion typically dominates, but chemical erosion can account for a substantial fraction of total erosion, even dominating in some slowly eroding basins (Summerfield & Hulton 1994; Milliman & Farnsworth 2011). In particular domains within a landscape, such as in soils, the typical contribution of chemical erosion can be even higher (Dixon et al. 2009). Chemical weathering may also affect the rates of other processes by altering bedrock erodibility (Han et al. 2014; Murphy et al. 2016).

In addition to the consequences for landscape evolution, chemical weathering is important because silicate weathering consumes atmospheric CO_2, a greenhouse gas. The resulting negative feedback, in which warmer temperatures accelerate silicate weathering, reducing the small atmospheric reservoir of CO_2 and weakening the greenhouse effect, is thought to be a principal regulator of Earth's long-term surface temperature (Berner et al. 1983). This is the basis for the suggestion that tectonically accelerated erosion could lead to global cooling (Raymo & Ruddiman 1992). Chemical weathering also releases essential nutrients from rocks into the biosphere.

One of the most important chemical weathering mechanisms is rock dissolution by acids, usually in the form of carbonic acid (dissolved CO_2) in rainwater or organic acids from plants. The other essential ingredients are a supply of fresh, reactive minerals and enough water to dissolve them in. The controls on chemical erosion rates are complicated, but through much research the following picture has emerged. Higher temperatures accelerate chemical erosion, but only if there is a supply of fresh minerals. In slowly eroding landscapes, or locally in soils, the reduced availability of fresh minerals can slow chemical erosion. Both the temperature ("kinetic") effect and the mineral supply effect are modulated by the rate of water flow through watersheds, which governs how far weathering reactions proceed towards equilibrium. Climate influences this balance in several ways.

5.1. Temperature

Laboratory experiments have confirmed that silicate weathering proceeds faster at warmer temperatures (Kump et al. 2000). Studies of chemical erosion rates in a single rock type spanning a range of climates suggest that this kinetic effect also occurs in nature. Riebe et al. (2004b) showed that chemical weathering rates in a granitic alpine environment decrease with altitude even faster than would be expected from the temperature effect, perhaps due to the additional influence of sparser vegetation and increased snow cover. Basalt has been the focus of numerous studies due to its relatively minor structural and compositional variations, the global distribution of basalts across a wide range of temperature and runoff, and the abundance of young basalt containing fresh minerals. Most of these studies have measured the concentration of dissolved bicarbonate, a product of silicate dissolution, in river water to obtain a spatially averaged measure of chemical erosion. These field measurements are consistent with a temperature dependence of chemical erosion (Dessert et al. 2003). The largest and most geographically extensive compilation to date (Li et al. 2016) finds support for increasing chemical erosion rates with increasing mean annual temperature (MAT) (Figure 6a), but the correlation is strongest among inactive (dormant) volcanic fields, probably because active fields have complicating hydrothermal effects and because young basalt is very reactive (Rad et al. 2007). MAT is a useful climate variable in the context of mineral weathering because it is a good proxy for soil temperature (Ferrier et al. 2012), although it is not perfect: reaction kinetics depend exponentially on temperature, so warm-season weathering may dominate, and chemical weathering may be arrested by a lack of liquid water in soils that freeze (Li et al. 2016).

5.2. Mineral Supply

In landscapes where fresh minerals are less abundant, the supply of minerals, and not temperature, can limit chemical erosion rates. The major control on mineral supply in most landscapes is mechanical erosion, which removes more weathered soil and rock from the surface and exposes the fresher material underneath. This idea is consistent with the observation that short-term chemical and mechanical erosion rates in large river basins are positively correlated (Gaillardet et al. 1999). Techniques for independently and simultaneously measuring longer-term mechanical and chemical erosion rates in soils using cosmogenic nuclides and elemental mass balance (Riebe et al. 2003) have revealed a similar correlation (Figure 6b) (Riebe et al. 2001b; Riebe et al. 2004a; Dixon et al. 2009), and compilations of cosmogenic nuclide erosion rates suggest that the correlation for river basins also applies over longer timescales (Figure 6b) (West et al. 2005). In steady-state topography ($\partial z/\partial t = 0$), the erosion rate equals the rock uplift rate, suggesting that tectonics can influence chemical erosion rates by regulating the mineral supply (Riebe et al. 2001b).

In a given landscape, what determines whether the chemical erosion rate is limited by the supply of fresh minerals due to mechanical erosion, or by temperature-dependent reaction kinetics? Models of coupled mechanical and chemical erosion in soils (Ferrier & Kirchner 2008) predict that the relationship between the two rates has a humped form: chemical erosion rate should increase with mechanical erosion rate in slowly eroding landscapes with thick soils where minerals have time to weather extensively once they reach the surface, and decrease with mechanical erosion rate in fast-eroding landscapes with thin soils where grains spend little time weathering before being transported away. Dixon et al. (2012) measured such a trend along a series of sites with varying mechanical erosion rates in the San Gabriel Mountains, California. The compilation of river solute fluxes by West et al. (2005) shows a strong correlation with mechanical erosion rate at slow erosion rates, but chemical erosion rates fall below this trend at fast mechanical erosion rates (Figure 6b). Interestingly, the positive correlation of chemical and mechanical rates extends to faster mechanical erosion rates in soils than in basins (Figure 6b) (West et al. 2005; Dixon et al. 2009), including some of the fastest erosion rates in the world (Larsen et al. 2014). This could be due to the different assumptions and biases involved in estimating chemical erosion rates from river and soil chemistry, the much longer averaging time of soil weathering rates (West et al. 2005), or effects beyond the supply and kinetic limitations.

5.3. Runoff

The effect of water on chemical erosion rates is conspicuously absent from the preceding discussion. Water can accelerate chemical weathering, both because it increases the wetted surface area of mineral grains where weathering reactions occur, and because more water flowing through the weathering zone dilutes the reaction products, keeping weathering reactions farther from equilibrium. Many studies of chemical erosion and climate use runoff as an independent variable rather than precipitation, partly because runoff excludes evapotranspiration, and partly because chemical erosion fluxes are commonly measured from river water.

Runoff does appear to influence chemical erosion rates, but the empirical evidence is somewhat equivocal. The compilation of Li et al. (2016) is consistent with a dependence of basalt chemical erosion rates on annual runoff; however, they caution that this correlation could be an artifact of higher average runoff in active, faster-weathering volcanic fields, which tend to be located near the ocean, and covariation due to the fact that the weathering rate is calculated from runoff. Studies of silicate weathering rates in other environments show correlations with runoff (White & Blum 1995), but scatter in the data suggests that other variables are important. Several investigators have developed models that isolate the effects of runoff, temperature and mechanical erosion on chemical erosion rates, and showed that runoff's influence is significant

(Gaillardet et al. 1999; Riebe et al. 2004a; West et al. 2005). But it is important to note that field studies of climate and chemical erosion that control for lithology do not universally find support for simple dependences on mineral supply and climate variables (e.g. Ferrier et al. 2012). Part of the explanation may be that climate has a nonlinear effect on soil chemistry that depends on thresholds of moisture availability (Chadwick et al. 2003; Dixon et al. 2016).

A second possibility is that topography and watershed hydrology influence the balance between mineral supply and kinetic effects on chemical erosion rates. Maher & Chamberlain (2014) proposed a potentially unifying framework in which weathering intensity depends on the supply of fresh minerals (and therefore mechanical erosion rate) and temperature, and also on the progress of chemical weathering reactions towards equilibrium as water travels through watersheds, which depends on runoff and hydrological flow path lengths set by topography. For a given rate of runoff, in their framework, chemical erosion is fastest when water is in contact for long enough that it becomes saturated in dissolved minerals, and slowest when the water is in contact for so little time that little dissolution can occur. They framed this balance in terms of a fluid travel time through the landscape, $T_f = L_f \phi / q_r$, and an equilibrium time for chemical weathering reactions, $T_{\mathrm{eq}} = C_{\mathrm{eq}}/(R_{n,\max} f_w)$, where L_f is the reactive flow path length, ϕ is porosity, q_r is runoff, C_{eq} is the maximum solute concentration, $R_{n,\max}$ is the maximum weathering reaction rate of mineral species n, and f_w is a function of mechanical erosion rate and other factors that scales the reaction rate according to the supply of fresh minerals, with $f_w = 1$ for unweathered rock or soil. The ratio T_f / T_{eq} defines the Damköhler number of the advection-reaction system within a landscape. Factoring out runoff, Maher & Chamberlain (2014) obtained a dimensional coefficient,

$$Dw = \frac{L_f \phi R_{n,max}\, f_w}{C_{eq}}, \tag{10}$$

that characterizes the progress of weathering reactions in analytical solutions of the advection-reaction equation for the system. Dw is an efficiency factor for chemical erosion that scales positively with temperature (via $R_{n,\max}$), mechanical erosion rate (via f_w), and topographic relief or valley spacing (via L_f). For a given runoff, larger Dw means a higher solute concentration and therefore a faster chemical erosion rate, but only up to the maximum possible concentration (Figure 6c). Comparing this model with a global compilation of chemical and mechanical erosion rates in major river basins shows that the data fall below this "thermodynamic limit" (Figure 6c), but that faster-eroding basins have higher Dw, making their chemical erosion rates more sensitive to runoff. The variability of chemical erosion rates measured in fast-eroding landscapes (Figure 6b) may reflect this heightened sensitivity to runoff.

This framework has not yet been widely tested with field measurements, however, and it is not clear whether it can explain finer-scale observations that have been made in many landscapes. Estimates of chemical erosion from short-term river chemistry can also be biased, and may neglect groundwater solute fluxes that bypass surface drainage systems (Schopka & Derry 2012).

5.4. Vegetation

Other reviews have documented the mechanisms by which vegetation influences chemical weathering (Berner et al. 2003), but the effects of vegetation on chemical erosion over the timescales of landscape evolution have not been quantified systematically, despite important first steps (Doughty et al. 2014). The observation that mineralogy influences vegetation types (Hahm et al. 2014) raises the possibility of feedbacks between chemical erosion and vegetation.

6. CONSEQUENCES FOR TOPOGRAPHY

Although this review is mainly concerned with climate's effects on process rates, illustrating a few of the major observed and hypothesized effects of climate on topography helps put the preceding discussion in context.

6.1. Topographic Signatures of Climate

Climate variations can leave a measurable signature in the shapes of landforms. The quantitative evidence for climatic control of process rates reviewed above illuminates the origins of some of these characteristic topographic forms. For example, Stark (2010) found that bedrock rivers in the Western Pacific region are more sinuous where rainfall rates are more variable, as measured by the frequency of tropical cyclone strikes, suggesting that variability of Q may be important for bank erosion in bedrock rivers as well as vertical incision. Gentler hillslope angles in mountainous regions with heavier precipitation may be a consequence of increased landslide susceptibility (Gabet et al. 2004).

Climate can alter the balance of river and hillslope processes that sets a landscape's drainage density (Tucker & Slingerland 1997) and the size of hillslopes. Perron et al. (2009) showed that the spacing of first-order river valleys (the smallest valleys, with no tributaries) in soil-mantled landscapes is proportional to a characteristic horizontal length scale given by $(D/K)^{1/(2m+1)}$. They estimated this quantity for several landscapes and noted that landscapes with a smaller value of the length scale, and therefore more closely spaced ridges and valleys, tended to have drier climates. They speculated that this trend could be a consequence of less permeable soils producing more runoff in drier regions, which would increase K, and less biologically driven soil disturbance in drier regions, which would reduce D. Chadwick et al. (2013) tested this idea by examining drainage density and process differences across a rainfall gradient with uniform granitic lithology and uniformly slow erosion rates in South Africa. They measured a clear decrease in drainage density with increasing rainfall (Figure 7), and their field observations suggested that differences in runoff and biotic effects were largely responsible. They also found that chemical weathering intensity increased with precipitation.

One of the most dramatic topographic illustrations of climate-sensitive landscape evolution is the widespread occurrence of aspect-dependent slope asymmetry. It has been noted for many years that slopes facing Earth's equator have different microclimates than slopes that face the poles, and that these microclimates can create systematic differences in steepness, with pole-facing slopes at middle latitudes typically being steeper than equator-facing slopes (Figure 8) (Poulos et al. 2012; see review in Richardson 2015). This observation suggests that solar radiation has an indirect but strong effect on long-term erosion rates, and researchers have begun to measure and model how this connection occurs through climatic effects on individual landscape processes (Burnett et al. 2008; Istanbulluoglu et al. 2008; Perron & Hamon 2012; West et al. 2014; Richardson 2015; Aldred et al. 2016).

6.2. Feedbacks Between Climate and Landscape Evolution

6.2.1. Tectonics, Climate and Erosion As noted in Section 1, models predict that climate-erosion couplings can have an enormous influence on spatial patterns of exhumation, which in turn can alter the size and shape of entire mountain ranges (Beaumont et al. 1992; Willett 1999; Whipple & Meade 2004; Roe et al. 2006; Whipple 2009). These models typically add the influence of climate by making erosion processes more or less efficient in proportion to spatial or temporal variations in climate; for example, making rivers erode rock faster where orographic rainfall is highest. Studies reviewed here are quantifying and refining these connections, providing the basis for a new generation of coupled models.

6.2.2. Topographic Perturbation of Climate The response of climate to landscape evolution may extend well beyond orographic rainfall. Mountain ranges have been argued to shape the hydroclimate of entire regions (Strecker et al. 2007), generate monsoons (though the importance of this effect in certain regions is controversial (Molnar et al. 2010)), and even drive global cooling by deflecting jet streams (Ruddiman & Kutzbach 1989). The role of major mountain building events in perturbing the silicate weathering feedback (Raymo & Ruddiman 1992) is still debated.

7. RESEARCH NEEDS AND DIRECTIONS

7.1. Research Needs

Of the many research needs highlighted in this review, several stand out. (1) The most pressing need is for landscape process models that explicitly include climate mechanisms. The empirical insights afforded by previous work offer a good starting point. A major challenge for some processes, which has inspired some of the work reviewed here, will be relating parameters in process laws to characteristics of long-term climate when those processes are actually responding to the full time series of weather. (2) A recurring theme in this review is the role of life in mediating climate's influence on landscape evolution. Efforts to quantify this role are still at an early stage (Istanbulluoglu & Bras 2005; Dietrich & Perron 2006; Jeffery et al. 2014). (3) The landforms we see today are the topographic legacy of past climates (Figure 2) (Godard et al.

2013). Comparisons of long-term process rates with present-day climate patterns can be illuminating, but they can also be misleading. Well-chosen natural experiments can reveal how landscapes consisting of landforms with different response times react to geological climate change. (4) When it comes to landscapes' response to past climates, we are data-poor. Understanding how sedimentary archives do (Armitage et al. 2011) or do not (Jerolmack & Paola 2010) record these responses is a daunting but essential endeavor (Hajek & Straub 2017). (5) We will probably soon discover how landscapes respond to anthropogenic climate change, and any means of anticipating the consequences will be valuable. This urgent need, which has been reviewed elsewhere (Pelletier et al. 2015), is a key extension of research on long-term landscape evolution.

7.2. Research Directions

The preceding review of current knowledge and needs suggests a few research themes that could lead to especially useful insights.

7.2.1. Natural Experiments Some of the clearest signals reviewed here come from landscapes with uniform bedrock and tectonic history but climates that vary in ways that remain stable through geologic time. Altitudinal transects (e.g. Riebe et al. 2004b) and orographic rainfall patterns (e.g. Chadwick et al. 2003; Ferrier et al. 2013a,b; Murphy et al. 2016) have already attracted much attention. Volcanic islands have several advantages over continents (Jefferson et al. 2014), including the ability to estimate the age and topography of an initial surface, and research on islands appears to be accelerating. As an added incentive, weathering of young basalt may be an important CO_2 sink in the climate-weathering feedback (Dessert et al. 2003).

7.2.2. Energy Budgets Most of the erosional work that shapes landscapes is ultimately derived from gravitational potential energy and energy from the sun. The pathways through which solar energy shapes landscapes are both abiotic (for example, precipitation patterns depend on evaporation and temperature) and biotic (for example, bioturbation by animals and plants depends on photosynthesis). Energy constraints are at the root of many climatic and biological differences among landscapes, but tracing the flow of energy through landscapes and quantifying the consequences for landscape evolution is very challenging. Nonetheless, a few studies have begun to relate energy budgets to landscape evolution (e.g. Yoo et al. 2005; Phillips 2009; Pelletier & Rasmussen 2009). Aspect-dependent slope asymmetry (Figure 8), one of the topographic signatures mentioned in Section 6.1, is a natural experiment that may hold a key to understanding landscape-scale energy budgets.

7.2.3. Rate-Limiting Mechanisms A change in the rate-limiting erosion or transport mechanism at different rates of tectonic or climatic forcing appeared as a theme in every major section of this review: biotic soil production and transport, mineral supply and chemical erosion, and floods and river incision. Efforts to measure such transitions in the field will help clarify climatic process controls.

7.2.4. Climate Extremes Studying how extreme events and climate states impact process rates should help reveal how landscape evolution has varied under past climates and lay bare the flaws in current process laws. Models of landscape evolution stand to gain from recent progress on the thermodynamics (O'Gorman & Schneider 2009) and records (Cook et al. 2015) of climate extremes.

7.3. Concluding Statement

Abundant evidence that climate sets the pace of erosional landscape evolution implies that many widely applied rate laws for erosion and sediment transport, which do not explicitly include climate, are incomplete. Progress on quantifying climate's influence on the rates of individual processes is bringing us closer to understanding the net effect of global changes in climate on erosion and landscape evolution.

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TABLE

Table 1. A summary of climate's role in rate laws for surface processes in erosional landscapes.

Process	Process law[a]	Climate effects[b]
Bedrock river incision	$E_r = k_1 f(q_s)(\tau - \tau_c)^a$ with $\tau = k_2 \left(\frac{Q}{w}\right)^{\alpha} S^{\beta}$, usually approximated as $E_r = KA^m S^n$	• Q generally increases with MAP, making river incision more efficient (higher K) in wetter climates under some conditions • Q is generally less variable where runoff is higher, making river incision less sensitive to MAP, or even less efficient in wetter climates; higher τ_c enhances this effect • w scales as a positive power law of Q; other climate effects on w unclear due to lack of a well-tested theory for w • q_s depends on size and abundance of grains shed by climate-dependent soil production and transport on hillslopes • k_1 increases with chemical weathering of river bed (see below)
Debris flows	$E_d = K_d f_d L \left[\rho \nu {D_p}^2 \left(\frac{u}{h_d}\right)^{\gamma} \cos\theta\right]^{\eta}$	• K_d increases with chemical weathering of bed (see below) • f_d scales with landslide frequency; higher drainage density in drier climates may increase f_d by creating more upstream source areas • Climate-influenced soil, vegetation and landslide characteristics may affect L, ν, D_p, u and h_d, but specific effects unclear
Soil production	$P(h) = P_0 e^{-h/h^*}$	• P_0 increases with MAP, especially across the transition from dry, lifeless landscapes to humid landscapes with abundant biological activity • Some evidence that P depends less on h in very dry landscapes, suggesting h^* may decrease with MAP, but insufficient data to test
Soil creep	$\mathbf{q}_s = -D\nabla z$ or $\mathbf{q}_s = -K_c h \nabla z$ or $\mathbf{q}_s = \frac{-D\nabla z}{1-(\lvert\nabla z\rvert/S_c)^2}$	• D and K_c increase with MAP • h is an outcome of soil production (see above) • S_c is generally higher where climate favors deep-rooting vegetation that binds soil, and is partly determined by landslide thresholds (see below)
Overland flow	No long-term process law, but many short-term expressions proposed	• Efficiency likely peaks in dry climates with large, rare storms • Climate-influenced vegetation can inhibit erosion by enhancing infiltration, adding roughness, and binding soil with roots
Landslides	No long-term process law, but many mechanistic models of slope instability	• Intense, prolonged or frequent runoff triggers landslides • Wetter seasonal conditions make it easier for storms or earthquakes to trigger landslides • Climate affects vegetation, which adds root cohesion that inhibits landslides; vegetation can also reduce soil saturation through transpiration and by enhancing soil infiltration capacity and conductivity • Deeper soils are generally more prone to failure (see soil production)

Chemical erosion	No long-term process law, but important empirical constraints and first steps toward a long-term theory	• Chemical erosion rate increases with annual runoff and MAT, but also with mechanical erosion rate, which controls the supply rate of fresh minerals • Climate-influenced vegetation enhances chemical weathering • Hydrological processes in watersheds govern the balance of climatic and mineral supply effects

[a] Symbol definitions: E, erosion rate; k_1, coefficient that depends on rock strength; $f(q_s)$, a function of fluvial sediment flux; τ, bed shear stress; τ_c, critical shear stress for erosion; a, constant; Q, river discharge; w, river channel width; S, water surface or channel slope; k_2, α and β, constants that depend on flow resistance; K, coefficient that depends on rock strength, precipitation and channel geometry; A, upstream drainage area; m and n, constants; E_d, debris flow erosion rate; K_d, coefficient that depends on rock properties and distribution of flow impact stresses; f_d, debris flow occurrence frequency; L, length of debris flow that erodes bedrock; ρ, particle density; ν, solid volume fraction; D_p, effective particle diameter; u, debris flow surface speed; h_d, debris flow depth; θ, bed slope angle; γ and η, constants; P, soil production rate from bedrock; P_0, soil production rate from bare bedrock; h, soil thickness; h^*, constant; $\mathbf{q}_s$, vector soil flux; D and K_c, soil transport coefficients; z, elevation; S_c, critical slope gradient.

[b] Abbreviations: MAP, mean annual precipitation; MAT, mean annual temperature.

FIGURE CAPTIONS

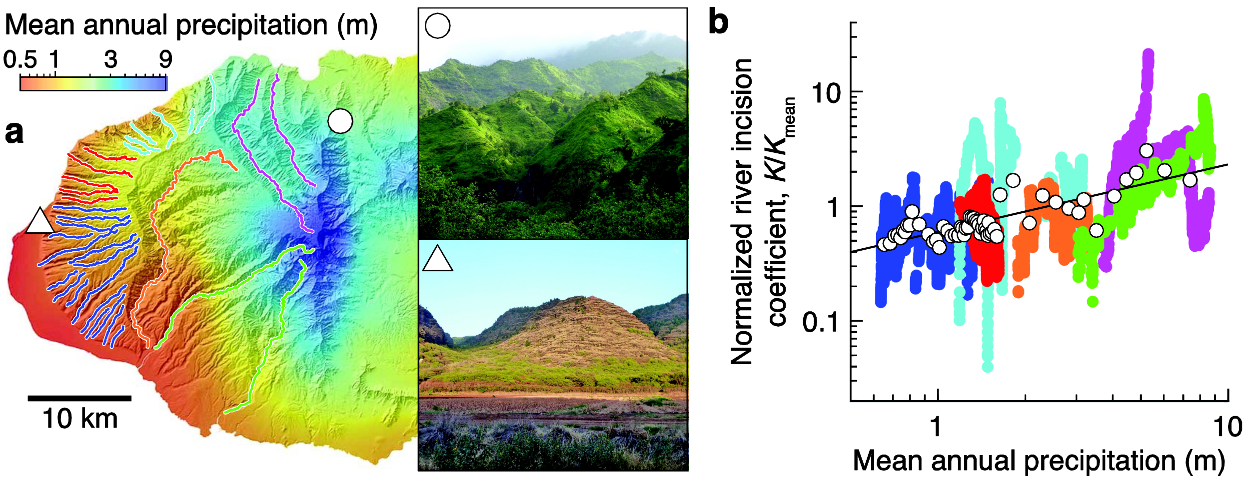

Figure 1: A landscape that bears a clear imprint of climate. (a) Shaded relief map of the Hawaiian island of Kauai with colors indicating mean annual precipitation. River canyons that drain the wetter parts of the island are deeper and wider than canyons that drain the drier parts. Inset images show landscapes on the wet and dry sides of the island. (b) Dependence of the rate coefficient for bedrock river incision (Equation 5) on upstream-averaged mean annual precipitation on Kauai. Point colors correspond to the rivers marked on the shaded relief map in (a). White points are logarithmic means in non-overlapping bins containing equal numbers of points. Black line is a power law with an exponent of 0.59 fit to the data. The value of the river incision coefficient *K* at each point has been normalized by the mean value of all points. Modified from Ferrier et al. (2013a).

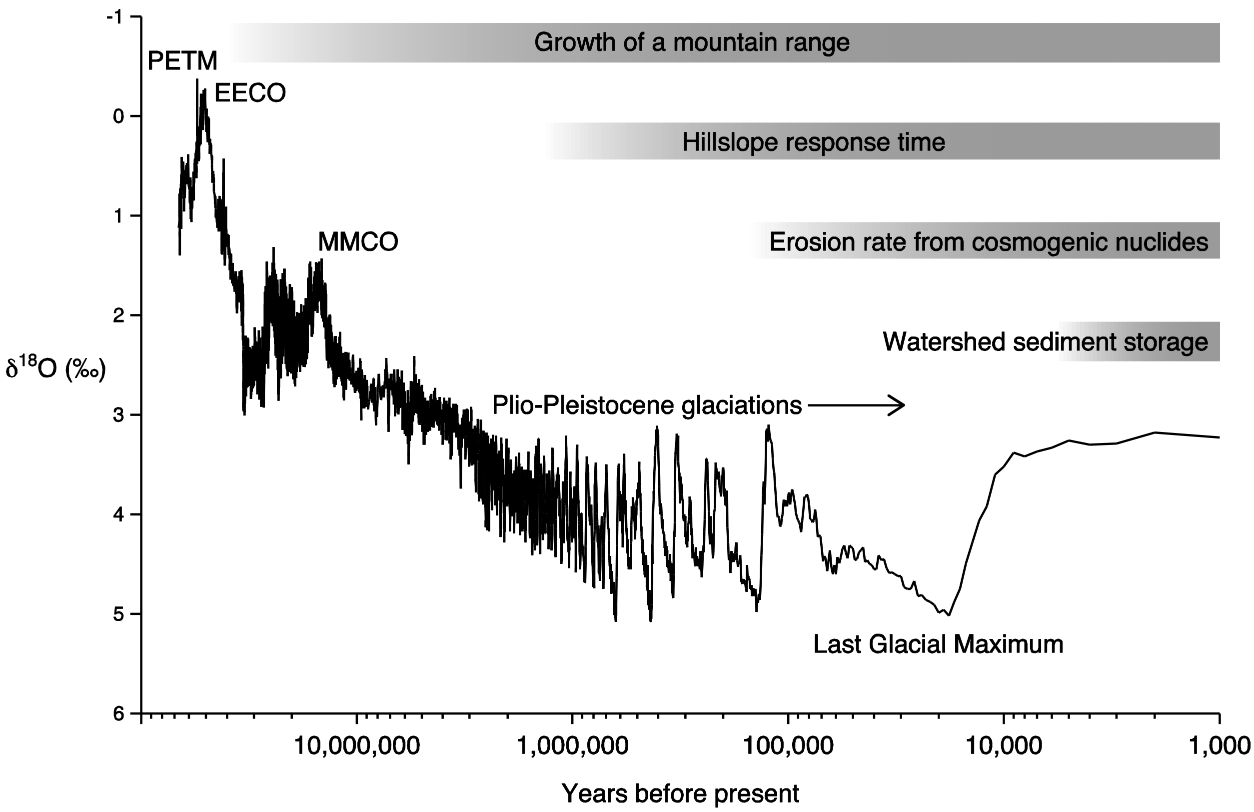

Figure 2: Relative timescales of geological climate change and landscape evolution. Marine oxygen isotope record, a proxy for global temperature and ice volume, from approximately 70 Ma to the present. Left edges of shaded bars indicate approximate timespans of several phenomena relevant to landscape evolution. Oxygen isotope data are from the benthic stack compilations of Zachos et al. (2008) prior to 5 Ma and Lisiecki & Raymo (2005) between 5 Ma and the present. Abbreviations for climate events: PETM, Paleocene-Eocene Thermal Maximum; EECO, Early Eocene Climatic Optimum; MMCO, Mid-Miocene Climatic Optimum.

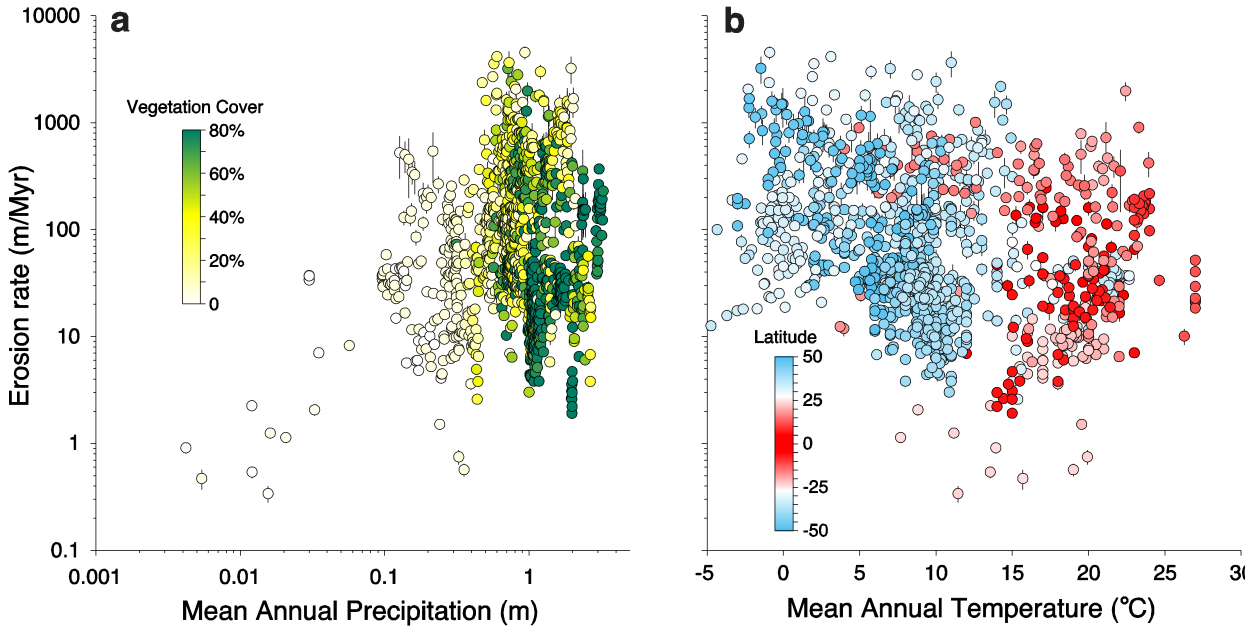

Figure 3: Global compilation of erosion rates from cosmogenic nuclides compared with mean annual precipitation and temperature. Points are colored according to drainage basin vegetation cover in (a) and average drainage basin latitude in (b). Climate data, land-cover data, and drainage basin-averaged erosion rates and uncertainties based on concentrations of cosmogenic ^{10}Be in river sediment are from the compilation of Portenga & Bierman (2011).

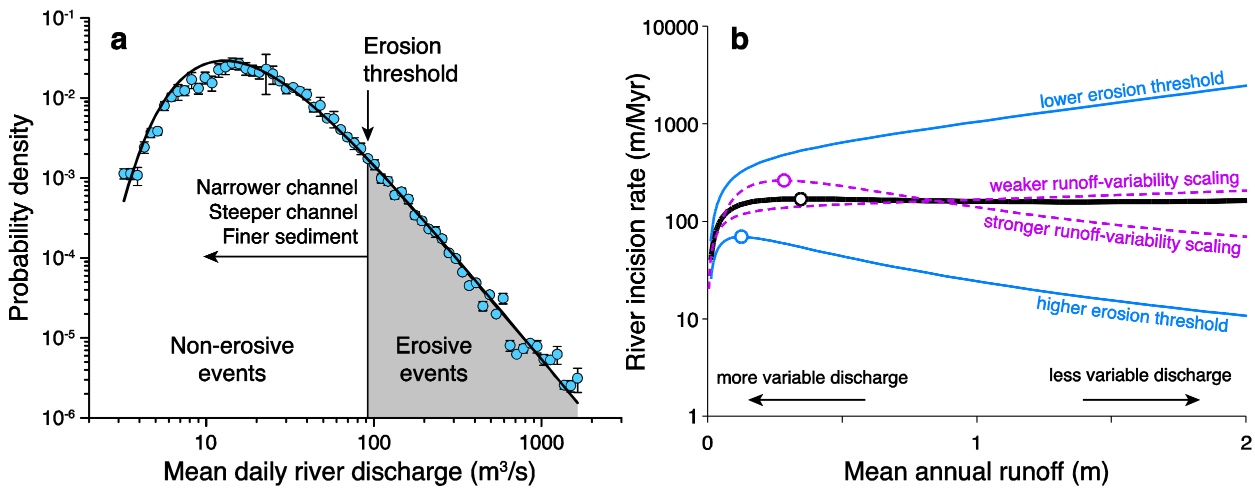

Figure 4. Effects of erosion thresholds and variable discharge on bedrock river incision. (a) Probability density function of mean daily discharge for the Hoping River, Taiwan, fit with an inverse gamma distribution (solid line). Shaded region indicates flows that exceed the estimated erosion threshold, which depends on channel width, steepness, and sediment grain size. The tail of the distribution corresponding to large floods is very close to a power law for which a more negative exponent (steeper slope in the figure) corresponds to less variable discharge. In many regions, less variable discharge correlates with higher runoff (Molnar et al. 2006). Modified from Lague (2005). Used with permission. (b) Calculation illustrating how long-term bedrock river incision rate scales with average runoff when runoff and variability are correlated. Black line is a base case calibrated for the San Gabriel Mountains, California, USA, for a channel steepness index of 75 (DiBiase & Whipple 2011). Open symbol marks the runoff that produces the fastest incision rate. Solid blue lines show the effect of raising or lowering the critical shear stress for channel incision by a factor of 2. Dashed magenta lines show the effect of strengthening or weakening the relationship between discharge variability and runoff. Either effect can change the sign of the relationship between long-term incision rate and runoff for all but the driest scenarios

and can transform the relationship into a monotonic trend with no maximum. Modified from DiBiase & Whipple (2011). Used with permission.

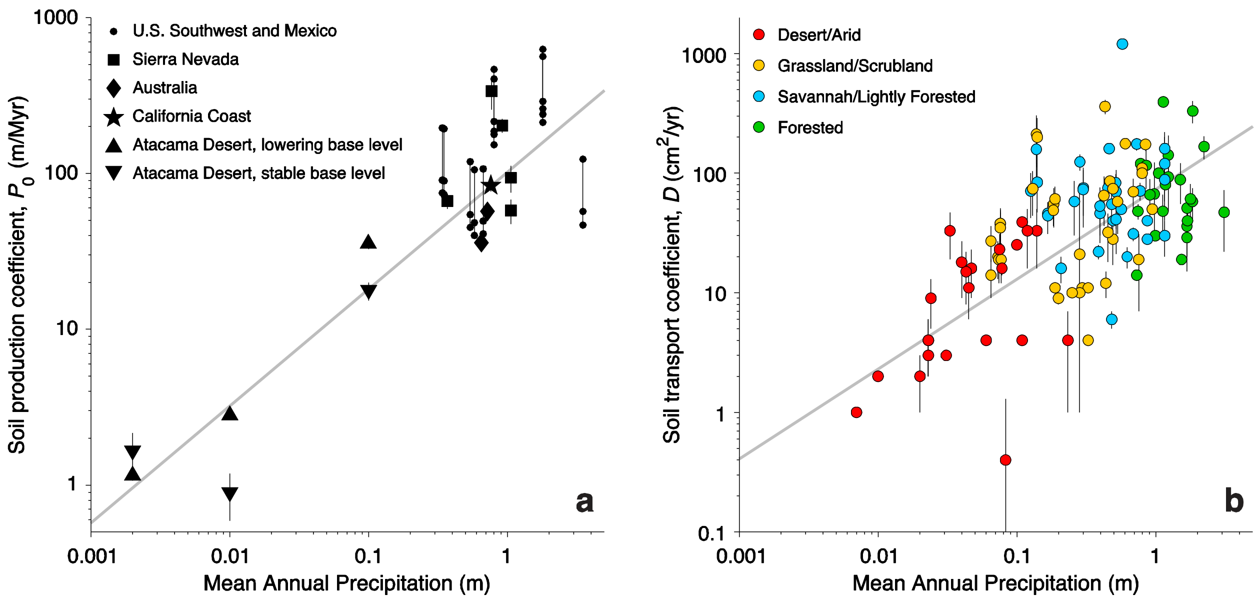

Figure 5. Rate coefficients for soil production and transport compared with mean annual precipitation. (a) Values of the soil production coefficient P_0 in Equation 8 on granitic hillslopes, estimated from the data and compilation of Owen et al. (2011), compared with mean annual precipitation, also from Owen et al. (2011). Estimates of P_0 and h^* were obtained by linear regression of log(P) against h. In cases where the fractional standard error of the h^* estimate exceeded 50%, P_0 was estimated by assuming $h^* = 0.52$ m, the mean value for sites with well-constrained estimates of h^*. Uncertainties are one standard error of the mean for the estimated regression parameters, and should be interpreted as minimum estimates of uncertainty. Uncertainties for small circles represent ranges of P_0 estimates for the measurements of Riebe et al. (2004b) (who reported that $h < 0.6$ m) for $h = 0$, 0.1 m and 0.5 m and $h^* = 0.52$ m. Gray line is a power law with an exponent of 0.75 included as a visual guide (not a fit to data). See Supplemental Table 1 for data. (b) Values of the soil transport coefficient D in the linear or nonlinear slope-dependent soil transport laws compared with mean annual precipitation. Points are colored according to the category of vegetation present at each site. Data are from the compilation of Richardson (2015), which includes the results of Callaghan (2012), the

compilation of Hurst (2013), additional measurements from the literature, and new estimates. Uncertainties are one standard error of the mean where it could be calculated from the original data; otherwise, uncertainties are as originally reported and include standard deviations and estimated ranges. Gray line is a power law with an exponent of 0.75 included as a visual guide (not a fit to data).

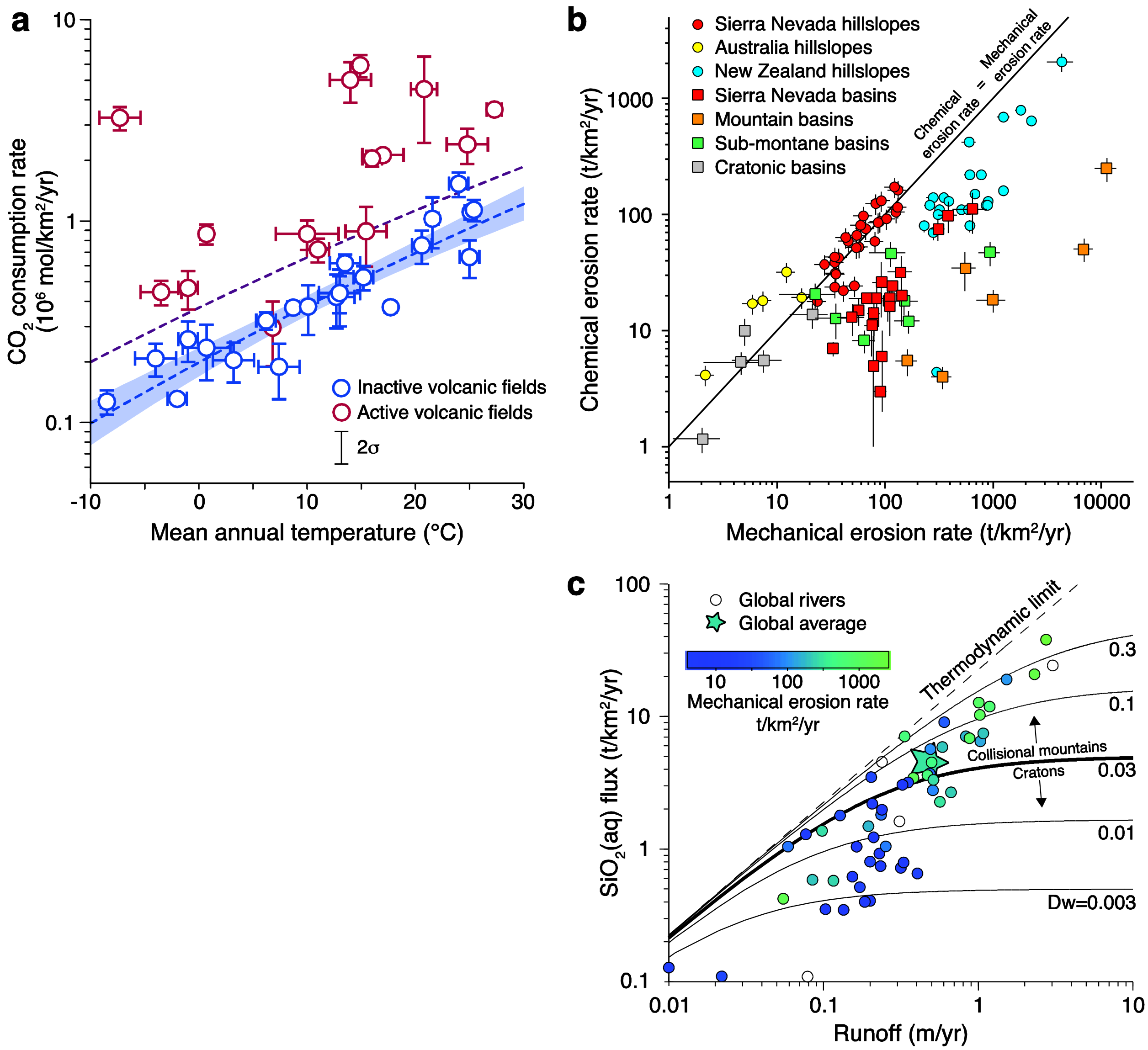

Figure 6. Climatic effects on chemical erosion. (a) Plot of CO_2 consumption rate, a measure of chemical erosion rate, against mean annual temperature for a global compilation of active and inactive basaltic volcanic fields, modified from Li et al. (2016). Blue line with shaded uncertainty bounds is a fit to inactive fields; purple line is a fit to all data. Used under a Creative Commons license: https://creativecommons.org/licenses/by/4.0/. (b) Plot of chemical erosion rate against mechanical erosion rate for hillslopes and river basins, based on compilations by Dixon et al. (2009) and West et al. (2005) and data from Riebe et al. (2004b) and Larsen et al.

(2014). Inspired by Dixon et al. (2009). Black line corresponds to equal mechanical and chemical erosion rates. See Supplemental Table 2 for data. (c) Plot of dissolved silica flux, a measure of chemical erosion rate, against annual runoff for a global compilation of major rivers by Gaillardet et al. (1999). Points are colored according to mechanical erosion rate, which is inferred from suspended sediment load. Solid lines are predictions of the model of Maher & Chamberlain (2014) for different values of the Damköhler coefficient, Dw, which measures the relative influences of chemical weathering reaction rates, mineral supply rates and water flow rates on chemical erosion. The thermodynamic limit corresponds to the maximum possible concentration of dissolved silica. Modified from Maher & Chamberlain (2014). Used with permission.

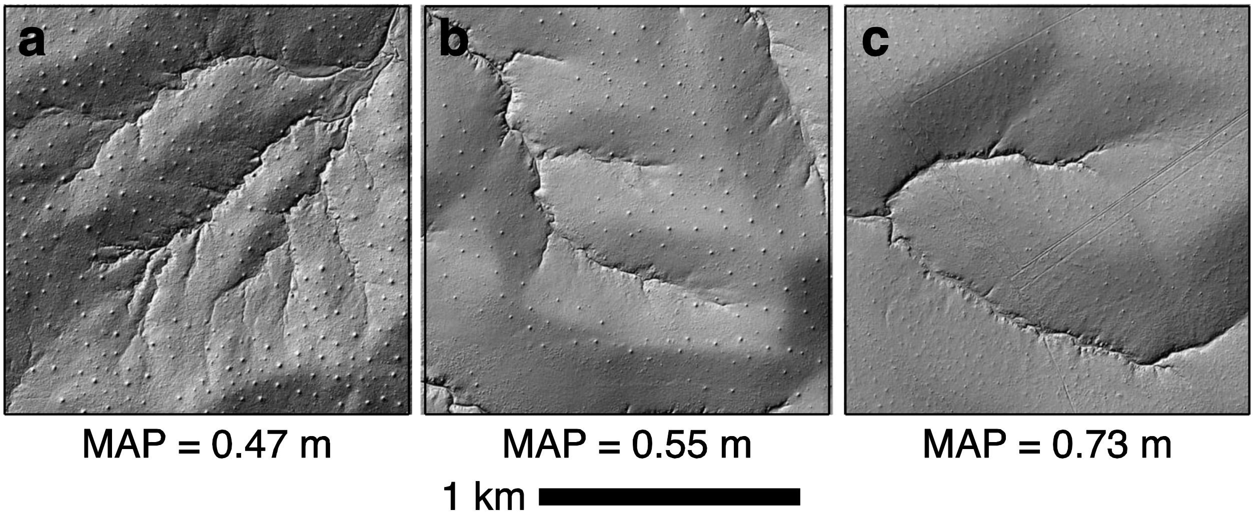

Figure 7. Lower drainage density and longer hillslopes in wetter sites. Shaded relief maps of three landscapes in Kruger National Park, South Africa, with similar erosion rates and granitic bedrock but differences in mean annual precipitation (MAP) that have likely persisted for millions of years. Chadwick et al. (2013) show that the transition from concave-down hillslopes to concave-up valleys occurs at a larger drainage area in wetter sites because changes in soil hydrologic properties, chemical weathering, soil transport mechanisms and bioturbation increase the strength of hillslope weathering and soil transport relative to fluvial incision. Small bumps on hillslopes are termite mounds. Modified from Chadwick et al. (2013). Used with permission.

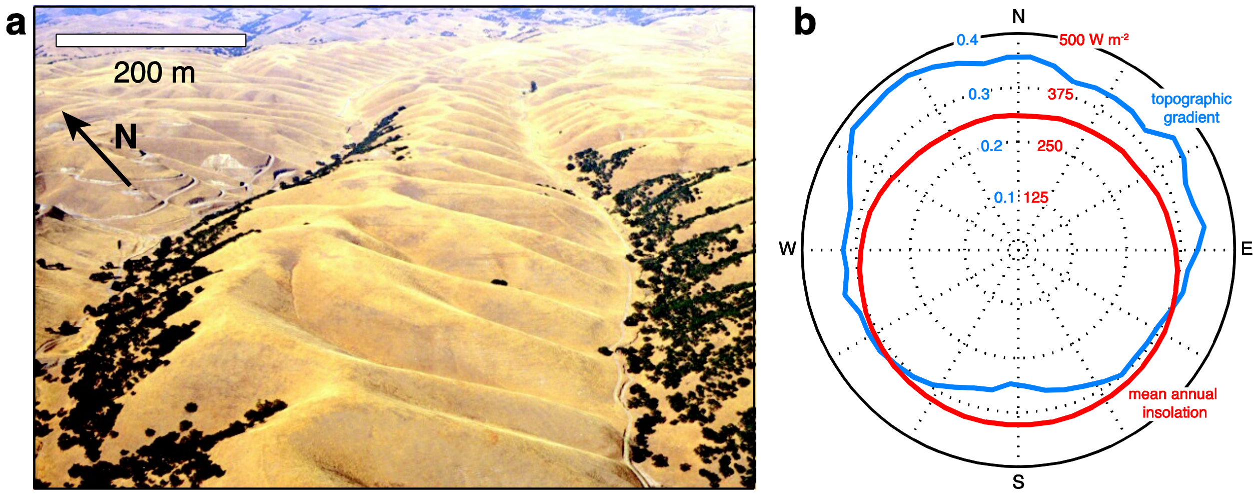

Figure 8. Effects of slope aspect on landscape evolution. (a) Aerial photograph of an asymmetric ridge in Gabilan Mesa near Bradley, California. North-facing slopes, which have a distinct microclimate because they receive less sunlight, are steeper, more vegetated and have shallower valleys than south-facing slopes. (b) Plot of the steepness of the topographic gradient (blue) and the mean annual solar radiation (red) as a function of slope aspect, the azimuth of the topographic gradient vector, for the area shown in (a). On average, north-facing slopes in the area analyzed are 42% steeper and receive 23% less insolation than south-facing slopes. Calculations were performed on a 1 m/pixel topographic map derived from airborne laser altimetry. Solar radiation for cloud-free conditions was estimated using the approach of Fu & Rich (1999), which accounts for direct and diffuse radiation, viewshed, and time-dependent sun position.