THE INFLUENCE OF LITHOLOGY
ON THE FORMS OF HILLSLOPES AND DRAINAGE BASINS
IN SOIL MANTLED LANDSCAPES

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DOCTOR OF PHILOSOPHY

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Abstract

Much of the Earth's surface is blanketed by a mantle of soil and weathered rock. In these settings, topographic forms are typically smooth, with convex hillslopes grading subtly into the concave valley networks that define drainage basins. In 1877, G.K. Gilbert recognized that this smooth topography masked the underlying variability in geologic structure and proclaimed this the 'Law of Divides.' Gilbert reasoned that in settings where rock weathered to soil faster than that soil could be eroded, topographic forms would be dictated by the processes transporting that soil rather than the variations in strength of the underlying rocks. There are exceptions to every rule, and this dissertation focuses on exceptions to Gilbert's 'Law of Divides,' situations where lithology influences the form of soil-mantled landscapes. I begin at hillslopes, demonstrating that if soil transport is sensitive to soil thickness, which in turn depends on underlying rock properties, variations in hillslope gradient can arise at geologic contacts, a phenomenon observed in some landscapes. Following this, I focus on the heads of valley networks and ultimately entire drainage basins. In the Gabilan Mesa, in the central Coast Ranges of California, properties of the weathered zone vary with slope aspect, placing rocks that favor headward channel growth close to the surface on south facing slopes. As a result, the preferential expansion of channel heads on south-facing slopes has led to a basin-scale topographic asymmetry. I conclude by measuring asymmetry in drainage basins in a number of other landscapes, and demonstrate that asymmetry in the structure of drainage

networks is most pronounced in rock types that may be especially vulnerable to erosion. Hence, even at the largest scales, soil-mantled topography may encode some signature of the properties of underlying rocks.

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Chapter 1

Introduction

The surface of the Earth is shaped by the protracted competition between the tectonic construction of relief and the destruction of that relief by surface processes. Given adequate time, the prevailing wisdom is that the forms of landscapes should adjust to reach a dynamic equilibrium where the rate of removal of material from the surface is uniform (Gilbert, 1877). This is generally accomplished by slope adjustments, which are thought to control the intensity of erosion or mass transport within rivers and on hillslopes (Gilbert, 1877, 1909; Exner, 1920; Culling, 1960; Howard and Kerby, 1983). However, the same processes producing relief and denuding mountains have given rise to a diversity of different rock types. Some of these rocks are more resistant to surface processes than others, and early work proposed that the construction of steeper landforms within more resistant geologic units would yield a balance in the rate of erosion within less resistant rocks (Gilbert, 1877). Gilbert (1877) referred to this as the “Law of Structure,” where the form of landscapes were dictated by the variations in the resistance of those rock units exposed at the surface.

Across much of the Earth’s surface rock is buried beneath a mantle of soil. The weathering and break down of rock to form soil controls the initial chemistry and physical properties of soils, determines the rate at which life-supporting nutrients are

added to the soil (e.g., Walker and Syers, 1976; Vitousek et al., 2003; Hilley et al., 2010, and references therein), and at the scale of the planet may help regulate global climate (e.g., Berner et al., 1983). Gilbert (1877) considered these landscapes to be instead dictated by the “Law of Divides”. Given that weathering can produce a mantle of decomposed rock faster than it is removed by erosion, Gilbert (1877) reasoned that topographic forms would reflect the slope required to transport this weathered material rather than the strength of underlying rocks. This dissertation focuses on apparent exceptions to Gilbert’s Law of Divides: instances where the strength of rocks beneath a soil mantle influences the form of evolving landscapes.

To assess the possible connections between varying lithology and the form of soil mantled landscapes, I employ a mix of analyses of digital representations of topography, field observations, and simulations of topographic evolution. In this thesis, I seek to demonstrate a link between substrate and topographic form, and the analysis of digital topography provides the means to quantify topographic forms at scales ranging from individual hillslopes, to catchments, to entire mountain ranges. By simulating the evolution of topography we can make testable predictions for topographic analyses that are rooted in the processes we attempt to describe in simplified models of surface processes. Field observations provide the primary data necessary to both inspire and assess the hypotheses I test in this dissertation.

Much of this dissertation focuses on the Gabilan Mesa, a landscape that has been used as a type example of soil mantled landscapes (e.g., Dietrich and Perron, 2006; Roering et al., 2007; Perron et al., 2009). The Gabilan Mesa is nestled between the Salinas Valley and the San Andreas Fault in the California Coast Ranges. In contrast to the “dark and brooding – unfriendly and dangerous” Santa Lucia Mountains west of the Salinas Valley, author John Steinbeck described the Gabilans as:

*... light gay mountains full of sun and loveliness and a kind of invitation,
so that you wanted to climb into their warm foothills almost as you want*

to climb into the lap of a beloved mother. (Steinbeck, *East of Eden*)

These “beckoning mountains with a brown grass love” seemed the ideal place to focus a dissertation. Much of the Gabilan Mesa is developed in sub-horizontal Miocene and younger marine and non-marine sediments (Galehouse, 1967; Dohrenwend, 1974; Durham, 1974). As a result of what is likely syn- to post-depositional tectonic deformation (Galehouse, 1967), the stratigraphy within the Gabilan Mesa and low-relief surface that defines the mesa dip slightly to the West-South West. This tilt appears to have preconditioned the landscape to form drainage basins that flow generally parallel to this dip direction (Dohrenwend, 1974, 1978).

Each of the chapters in this dissertation is motivated by an observation of the geomorphology of the Gabilan Mesa. Chapter 2 focuses on undulations in hillslope gradient that relate to underlying stratigraphy, chapter 3 on the preferential exposure of bedrock in south flowing valley heads, and chapter 4 on differences in the size and number of tributaries flowing north and south. While I was unaware of these features when I first set out to the Gabilan Mesa, they are not subtle features to a trained eye and had not gone unnoticed. Decades before I began this dissertation, Dohrenwend (1974) had recorded these observations and more generally noted that there were unique attributes to the morphology of the Gabilan Mesa where it was underlain by particular rock types, somewhat at odds with the “Law of Divides” (Gilbert, 1877). In the work of Dohrenwend (1974), the features of the topography that motivate chapters 2 and 3 of this dissertation received little more than a mention, in this dissertation I unite these observations within the existing framework for the evolution of soil mantled landscapes. Asymmetry in the structure of drainage basins in the Gabilan Mesa was a central contribution of Dohrenwend’s (Dohrenwend, 1974, 1978), in chapters 3 and 4 I expand on the work of Dohrenwend and our results support and refute various conclusions he had about the development of drainage basin asymmetry.

In Chapter 2, I begin my investigation at the smallest scale landscape element: hillslopes. Hillslopes occupy the regions of drainage basins upslope of where water flowing over the surface concentrates into channels and forms valleys. Gilbert (1909) argued that these regions would inevitably reach a steady-state, such that every point along the hillslope was capable of transporting the flux introduced to it from upslope. Gilbert (1909) reasoned that a dependence of flux on topographic slope drove hillslopes to construct their characteristic convex forms to achieve this steady state. Over a century later, Gilbert's general hypothesis holds. Culling (1960) first formalized this conceptual model as a linear diffusion model of hillslope evolution. While later efforts recognized that in many cases soil flux and slope are more likely non-linearly related (Roering et al., 1999), studies have shown that topographic slope and measured sediment flux do increase downslope in the manner Gilbert (1909) predicted (McKean et al., 1993). In this model the coefficient relating slope and flux is typically constant throughout a region, and therefore variations in underlying lithology cannot influence the form of the evolving hillslope surface.

However in the Gabilan Mesa, undulations in the slope of some hillslopes parallel underlying variations in stratigraphy, suggesting that underlying lithology *is* influencing the form of these landscapes. In this chapter I unite field observations and measurements with analysis of gridded topography and numerical models to test a new relation between slope and soil flux that additionally depends on soil thickness. This new formulation is able to predict situations like that seen in the Gabilan Mesa, where lithology can influence hillslope forms, and the perhaps more common situation where contacts between rock units are masked by smooth landscape surfaces.

Chapter 3 poses that the strength profile of the critical zone, as dictated by pedogenesis and the weathering of rock, can leave an imprint on the form of drainage basins. Over landscape-forming timescales, the transition from hillslopes to channels is controlled by the relative efficiency of soil transport and channel incision (Perron

et al., 2009). This transition remains stable in the face of slight fluctuations in erosion rate given that increases in erosion rate are likely to expose rocks that are stronger than their weathered products, slowing any further increase in erosion rate.

What would happen if the opposite were true, that increases in erosion rate exposed highly vulnerable rocks, causing an unstable increase in erosion rate? Rocks rich in swelling clays often slake, a process that pervasively fractures these rocks when they are exposed to oscillations in moisture content. The Pancho Rico Formation is highly susceptible to slaking and underlies a large portion of the Gabilan Mesa. Using laboratory tests of tensile strength, I demonstrate that Gabilan Mesa soils have a greater tensile strength than rocks of the Pancho Rico Formation after that rock has been exposed to a single wetting and drying cycle. Exposure of vulnerable rock may then represent an instability in landscape development, where increases in erosion rate that expose bedrock trigger further increases in erosion rates. Within the Gabilan Mesa, the Pancho Rico Formation is primarily exposed in incisional channel features we refer to as arroyos. Using topographic analysis and geomorphic mapping, I demonstrate that the morphology of these features and their position in the landscape is consistent with arroyos expanding headward at the expense of soil mantled hillslopes, as would be expected given unstable erosion. Arroyos are exclusively hosted within south-flowing catchments, and I reason that this is the consequence of a thinner mantle of soil observed on these slopes allowing arroyos to be triggered more easily. This asymmetric triggering of headward channel expansion appears to drive profound topographic asymmetry, a property observable at the catchment scale. Taken together, these observations suggest that the inversion of strength that results from the weathering and production of soils on weak rocks ultimately influences the morphology of entire drainage basins.

The final chapter expands my observations of topographic asymmetry from Chapter 3 to the scale of multiple neighboring drainage basins. Observations from

Chapter 3 suggest that the development of strong asymmetry in the Gabilan Mesa is driven by the unstable growth of south-flowing channels that incise into rocks that present little resistance to erosion. Dohrenwend (1978) observed that in the region surrounding the Gabilan Mesa, the strongest asymmetry in valley networks was developed in relatively weak lithologies. However, some degree of topographic asymmetry has been observed across the northern and southern hemispheres in a great variety of rock types (Poulos et al., 2012).

I compare the morphometry of neighboring drainage basins developed in similar climates but contrasting rock types. Past efforts have utilized the contrast in the mean slope of gridded topographic data on opposing slope to quantify asymmetry (Poulos et al., 2012), however I highlight that while this metric does capture variability in topographic asymmetry, it under-represents the extent of asymmetry seen in the shape of drainage basins. The most asymmetric landscapes show a strong preference for the formation of south-flowing valleys. While the configuration of drainage networks displays non-random components in a variety of landscapes surveyed here, it is only in sites in weak rocks where a strong north-south asymmetry in the density and length of low-order tributaries is observed. Past studies have focused on the role of contrasts in vegetation with aspect in driving asymmetry in channel networks (Istanbulluoglu et al., 2008; McGuire et al., 2014; Yetemen et al., 2015) and these certainly drive differences in landscape evolution. However, using remotely-sensed proxies for vegetation productivity, I demonstrate that all the surveyed sites are formed in climates that favor contrasts in vegetation with aspect. This suggests, as Dohrenwend (1978) proposed, that the development of strong asymmetry is sensitive to local rock type, perhaps due to a mechanism like the one posed in Chapter 3.

Chapter 2

Lithologic controls on the form of soil mantled hillslopes*

2.1 Abstract

Slopes in steady-state soil-mantled landscapes tend to increase downslope in a way that balances local transport capacity with the sediment supplied from progressively larger source areas. Most formulations of sediment transport due to hillslope processes scale transport rate with local slope, which produces convex-up forms that are independent of the properties of the underlying lithologies. In this study, we document soil-mantled hillslopes that show variations in slope that mimic the underlying stratigraphy. We present stratigraphic and soil-thickness measurements, topographic analyses, and numerical models to demonstrate that variations in rock type can impact the forms of these soil mantled hillslopes if hillslope transport rates scale with local slope and soil thickness. This demonstrates that hillslope forms in soil-mantled landscapes can be influenced by the underlying lithology through a coupling between

*Reproduced with permission from Johnstone, S.A. and G.E. Hilley (2015), Lithologic control on the form of soil-mantled hillslopes, *Geology*, V. 43(1), 83-86, doi:10.1130/G36052.1.

the processes that produce soil from rock and those that transport this soil downslope.

2.2 Introduction

This study investigates how the geologic substrate impacts landscape forms on soil-mantled hillslopes. Gilbert (1877) proposed that when weathering can keep pace with rates of erosion a mantle of soil buffers surface morphologies from the structure of the subsurface geology. Gilbert (1909) then reasoned that convex forms characteristic of soil-mantled hillslopes balance the mass of mobile material produced along them with that transported by processes whose rates increase with slope. This model of slope-dependent hillslope transport was later formalized as a hillslope diffusion rule (Geomorphic Transport Law, GTL, *sensu* Dietrich et al., 2003, Culling, 1960). Previous studies have also proposed that transport rates may depend on slope and soil thickness (Ahnert, 1976; Braun et al., 2001; Anderson, 2002; Roering, 2008; Furbish et al., 2009) and some empirical evidence supports this view (Heimsath et al., 2005; West et al., 2014). Soil thickness may reflect some properties of underlying rocks and rates of erosion (Ahnert, 1976), because rates of soil production are expected and observed to depend on soil thickness and lithology (Heimsath et al., 2005). As a result, variations in underlying rocks may impact soil thickness, and in turn, transport rates. Therefore, changes in transport rates for a given slope that may accompany lithologic contacts should be expressed in the form of soil-mantled hillslopes. If this mechanism is in operation, the proposition of Gilbert (1877) that a soil mantle shields topographic forms from underlying lithologic variations needs to be revisited.

Hillslopes in the northern Gabilan Mesa (GM), in the central California Coast Ranges, display the characteristic convex-up morphology predicted by Gilbert (1909) and existing GTLs. However, superimposed on largely convex hillslopes are regular undulations in slope termed ‘shadow beds’: subhorizontal, soil-mantled features that

traverse hillslopes and valleys (Dohrenwend, 1974) (Figure 2.1). Individual shadow beds seem to correlate with the underlying stratigraphy of the Pancho Rico Formation, suggesting that geologic structure is impacting hillslope forms in this soil-mantled landscape.

We present field surveys of soil thickness, topographic analyses, and stratigraphic measurements that establish the connection between bedrock geology and the morphology of a soil-mantled hillslope. Using numerical simulations of landscapes and modeling field data, we demonstrate that shadow bedding is consistent with the quasi-equilibrium form of hillslopes traversing rocks with varying resistances to disaggregation given a soil-thickness-dependent transport rule. We propose a modified rule for hillslope soil flux that casts flux as the integral of a soil velocity profile that decays exponentially with depth. This velocity profile shows similar dependence on depth as the activity of a number of processes thought to be responsible for soil transport.

2.3 Methods

2.3.1 Theory

Vertically measured soil thickness, $H_v[L]$, evolves according to the local input of soil by production, which depends on the surface normal soil thickness, $H[L]$, and the divergence of the flux of soil per unit width, $q_s[L^2t^{-1}]$ (e.g., Dietrich et al., 2003):

$$\frac{dH_v}{dt} = \frac{\rho_{br}}{\rho_s} \frac{W_0}{\cos(\theta)} e^{-\alpha H} - \nabla \cdot q_s \quad (2.1)$$

The first term on the right hand side describes the empirically calibrated rate of soil production. $W_0[Lt^{-1}]$ is the bedrock lowering rate at zero soil thickness, $\alpha[L^{-1}]$ is a constant describing the decay in production rate with increasing soil thickness,

ρ_{br} and $\rho_s[ML^{-3}]$ are the densities of bedrock and soil, and θ is the hillslope angle. Theoretical expectations Ahnert (1976) and empirical evidence (Heimsath et al., 2005) suggest that lithology controls rates of soil production through W_0 , while α is fairly constant across different climates and rock types, with a value of $\sim 2 m^{-1}$. Lithology-dependent variations in W_0 require different soil thicknesses to achieve equivalent soil production rates over different lithologies. Adding rock uplift, $U[Lt^{-1}]$, and solving for the evolution of surface elevations, $z[L]$, gives:

$$\frac{dz}{dt} = U + \left(\frac{\rho_{br}}{\rho_s} - 1 \right) \frac{W_0}{\cos(\theta)} e^{-\alpha H} - \nabla \cdot q_s \quad (2.2)$$

We assume that downslope soil velocities reach a maximum at the surface and decay exponentially with depth throughout the entirety of the soil column. This is similar to previously developed expressions (Kirkby, 1967; Anderson, 2002; Roering, 2008; Furbish et al., 2009), but ignores the possibility of an inactive region in deep soil columns:

$$V(h) = V_0 e^{-h/d_c} \quad (2.3)$$

The surface normal depth at a point in the soil column is $h[L]$, $d_c[L]$ is the scaling depth of the velocity profile, and $V_0[Lt^{-1}]$ is the surface velocity. Integrating Equation 2.3 over H yields the flux per unit width:

$$q_s = V_0 d_c (1 - e^{-H/d_c}) \quad (2.4)$$

As d_c increases relative to H , a greater fraction of the soil column is present in the high velocity portion of the profile and flux becomes increasingly sensitive to variations in thickness. If $d_c \gg H$, Equation 2.4 approaches a plug flow condition, $q_s = V_0 H$. By normalizing the flux in Equation 2.4 as $q_s^* = q_s/(V_0 H)$, such that $q_s^* = D^*(1 - e^{-1/D^*})$, we capture the sensitivity of flux to soil thickness with the value

$D^* = d_c/H$. As D^* increases, variations in soil thickness increasingly impact the flux at a given slope. We incorporate Equation 2.4 into simulations of landscape evolution by allowing V_0 to vary as a function of slope and a constant $k[Lt^{-1}]$:

$$V_0 = -k\nabla z \quad (2.5)$$

for the case of thickness-dependent linear diffusion or:

$$V_0 = \frac{-k\nabla z}{1 - \left(\frac{|\nabla z|}{S_c}\right)^2} \quad (2.6)$$

for thickness-dependent nonlinear diffusive transport (Roering, 2008). Here, S_c is a threshold slope, with values commonly ~ 1 (Roering et al., 1999).

2.3.2 Field location and data

We test this theory with observations from an area of the Gabilan Mesa contained within the Pliocene shallow marine Pancho Rico Formation (Durham and Addicot, 1965). The Pancho Rico Formation comprises beds of mudstone with scattered very fine sand; fine sandstone with scattered coarse sand; and fossiliferous, pebbly conglomerate. Stratigraphy is only exposed at the heads of recently incised arroyos, where two stratigraphic columns were measured (Figure 2.1).

Soil thickness was measured in soil pits and by driving a rod down to the soil-bedrock contact. We averaged thicknesses that were measured within one LiDAR-DEM pixel (1 m), which amounts to 115 measurements from soil pits and 112 collected with the rod (Figure 2.1). The soil-bedrock contact was reliably identified with the rod in some soils, but the rod occasionally penetrated easily disaggregated bedrock. Thus, we base our interpretations and modeling of data only on measurements collected in soil pits. Vertically measured thicknesses were converted to surface-normal

measurements using local slope.

To compare the above theory with field data, we assume topographic steady state. Given this, hillslopes should be adjusted to transport the flux received from upslope (Gilbert, 1909), which should scale linearly with the upslope area on the landscape. This allows us to compare data from different hillslope positions according to the drainage area per contour length, a/b . We calculate a using the D_{inf} algorithm (Tarboton, 1997; Perron, 2010), and use the 1 m pixel dimension as b . Slope magnitude is calculated using a second-order, finite-difference kernel in the cardinal directions.

We focus on nonlinear relations between slope and flux, as hillslopes tend to become more planar downslope, with slopes below ~ 0.5 (Figures 2.1 & 2.2 A). A new GTL prediction of flux is presented in each of the three rows of plots shown in Figure 2.2. These relationships show a nonlinear function of slope (Figure 2.2 A) (Roering et al., 1999); the product of soil thickness and a nonlinear function of slope (representing transport due to plug flow; Figure 2.2 C) (Heimsath et al., 2005); and the result of combining Equations 2.4 and 2.6 (Figure 2.2 E) (e.g., Roering, 2008). Plotted against a/b are the components of each GTL that should vary smoothly as a function of a/b . Modeled values for these parameters, shown as dashed lines, were computed by least-squares minimization between observed data and a prediction based on the relevant GTL and the steady state assumption (e.g., Roering et al., 1999, justification in supporting material). For GTLs that depend on soil thickness, we use the mean soil thickness from our observations to develop this single prediction. In the right column of plots we test if deviations from the modelled predictions correlate with soil thickness.

2.3.3 Simulating landscapes

To illustrate the variable role lithology may play in shaping landscapes as a function of D^* we numerically integrate Equations 2.1 and 2.2 through the application of Equations 2.4 and 2.5 from a flat initial topography in 1D. The model domain was subjected to uniform uplift except at the boundaries, whose elevations were held fixed. To approximate a situation such as that in the GM, we assume that lithology controls W_0 and let this vary with the amount of exhumation that has occurred at a point. Here, modeled topography is incising through laterally continuous oscillations between 2.5 m thick “recessive” and “resistant” beds; we specify these as having values of W_0 50% greater or less than a mean value, respectively. The cyclic values of W_0 are advected upward by U , causing the model to evolve to a state in which relief oscillates around a mean value. To ensure initial topographic conditions were eliminated, we ran the model for 4×10^6 model years, a total uplift of $>10\times$ the total basin relief.

2.4 Results

2.4.1 Soil thickness and topography

For a given a/b , slopes are higher above thin soils (Figure 2.2 A & B). This is seen as a negative bias in the residuals as a function of soil thickness (Figure 2.2 B). Bias increases for the case that a soil thickness, slope product relates soil thickness to transport rate at a given slope (Figure 2.2 C & D). The velocity-profile-based prediction does not eliminate the noise present in the data (Figure 2.2 E). However, with a d_c of 12 cm, the segregation of points at a particular value of a/b based on thickness (e.g., Figure 2.2 A & B) is no longer present. This is reflected in the residuals, which are generally centered near zero for all but the thinnest soils (Figure

2.2 F).

2.4.2 Modeling

In Figure 2.3 A, we plot q^* as a function of D^* to reference each of the simulation results shown. Variations in H arise from the variations in W_0 in these simulations. As a result there are steady-state values of D^* for resistant (low D^*) and recessive (high D^*) layers; bars in Figure 2.3 B show the range of these values within each simulation. We scale k in Equation 2.5 so that flux remains relatively constant for a given slope and soil thickness, which reduces fluctuations in relief as d_c varies.

Soil thickness varies in all the simulations, but small values of D^* (Figure 2.3 i) decouple soil production and transport in a way that prevents variations in soil thickness (and hence lithology) from being expressed in the modeled topography. For larger values of D^* , balancing upslope sediment supply requires that V_0 (and thus slope) must change to compensate the changes in H that arise at lithologic boundaries. D^* values of ~ 1 (Figure 2.3 ii) produce weak correspondence between slope and soil thickness, an effect that becomes progressively more pronounced as D^* increases (Figure 2.3 iii & iv).

2.5 Discussion

Beds in the Pancho Rico stratigraphy can be traced from exposures in arroyos to laterally continuous undulations in slope on soil mantled hillslopes (Figure 2.1). Soils tend to be thinner in steep portions of these shadow beds than in low slope sections (Figure 2.2 A & B). A linear relationship between soil thickness and transport overestimates transport through thick soils, suggesting transport does not occur at equivalent rates throughout the soil column (Figure 2.2 C & D). A prediction of flux based on Equation 2.3 accounts for variations in transport with depth in the soil

column, and produces a balanced estimate of sediment flux along the length of the hillslope when d_c is 12 cm (Figure 2.2 E). This prediction results in some bias in the residuals of the thinnest soils toward negative values (Figure 2.2 F), perhaps related to an underestimation of high slopes on gridded data or the fact that the proposed velocity profile is an oversimplification (Kirkby, 1967; Lewis, 1976). What remains to be demonstrated is that W_0 varies with the lithologies present and that the proposed velocity profile (Equation 2.3) matches velocity profiles from the GM.

Numerical modeling (Figure 2.3 ii-iv) demonstrates that landscape forms influenced by subsurface lithology can occur if rates of soil transport, q_s , depend on soil thickness and slope (Equation 2.4 - 2.6) and if different lithologies require different soil thicknesses to achieve equivalent soil production rates (Ahnert, 1976; Heimsath et al., 2005). Shadow beds form in simulations of hillslopes crossing lithologic contacts, represented with values of W_0 , when soil is transported with the proposed velocity profile as long as H is small relative to d_c (Figure 2.3 iii and iv).

Soil transport is commonly related to the disturbance of soils yielding a net-downslope flux (Davis, 1892; Culling, 1960; Roering et al., 1999). The velocity profile we model could alternatively reflect a decrease in the frequency of soil-disturbing events with depth. We use the term ‘activity’ to refer to some measure of the soil disturbance accomplished by a range of processes. Ground-squirrel burrows and drying cracks in soils suggest that burrowing and shrink-swell processes are two important agents in soil disturbance in the GM. Data on the depth dependence of these and other soil disturbing processes suggest that their activity commonly declines with depth (Figure 2.3 A). This suggests that the upper portions of the soil consist of rapidly mobilized material, while deeper portions may be disturbed less frequently. In thin soils the high-activity zone may include the entire mobile layer (e.g., high D^*), generating a tight coupling between soil production and landscape form. In thick soils, transport may be decoupled from deep portions of the soil column and

landscape forms may be less sensitive to soil production processes.

How common is lithologic control on hillslope form? Our simulations of 1D landscapes evolving by linear, depth-dependent transport indicate that the sensitivity of soil-mantled landscapes to thickness occurs when $q_s^* > \sim 0.5 - 0.7$ (Figure 2.3 B), which occurs when $D^* > \sim 1 - \sim 2$. Given a hillslope where soil thickness has adjusted so that soil production rates match rock uplift rates, D^* can be calculated as:

$$D^* = \frac{d_c}{H} = \frac{-\alpha d_c}{\ln(\frac{U \cos \theta}{W_0})} \quad (2.7)$$

In the case that α^{-1} and d_c are similar and θ is small, the equilibrium D^* only depends on U and W_0 . In this situation, values of D^* will exceed ~ 1 , and fluxes will be sensitive to soil thickness, when the ratio of U to W_0 is greater than $\sim e^{-1}$ (~ 0.4). For a value of W_0 around ~ 0.1 mm/yr (Dietrich et al., 2003), and assuming W_0 and U are uncorrelated, soil mantled landscapes where uplift (or mean erosion rates) are greater than 0.04 mm/yr are likely to be sensitive to variations in soil thickness (and therefore lithology). This suggests that a broad range of soil-mantled hillslopes may encode some signature of the underlying rock type. However, the dominance of nonlinear hillslope transport (Roering et al., 1999) in steep terrains may allow small changes in slope to compensate for potentially large variations in W_0 . Ultimately, nonlinear hillslope transport may limit features such as shadow beds to landscapes with large variations in W_0 and in which base-level lowering is rapid enough to produce thin soils, but slow enough (and with hillslopes short enough) to maintain slopes well below their critical values.

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2.7 Supporting material

2.7.1 Assumption of steady state soil flux

For a 1-dimensional, non-linear thickness independent transport rule, topographic steady state predicts that:

$$q = k \frac{S}{(1 - (S/S_c)^2)} = U \frac{\rho_{br}}{\rho_s} x. \quad (2.8)$$

Where S [] is the topographic slope, x [L] is the distance from the divide, and all other parameters are defined in the main text. Similar expressions can be derived to calculate slope for linear and thickness-dependent transport rules as well. Non-linear equations (e.g., Equation 2.6 and 2.8) can be solved iteratively for S to predict how slope varies as a function of x , while equilibrium slopes can be calculated directly for linear equations (e.g., Equation 2.5). To develop a continuous prediction for slope along our hillslopes (as shown in Figure 2.3), we minimize the sum of the squared errors between the equilibrium assumption (defined above) and topographic data. In two dimensions we use the drainage area per contour length, a/b , in place of x (Roering et al., 1999), where the contour length is specified as the grid cell width (1 m). This accounts for convergence and divergence of flux due to planform curvature. Drainage area was calculated using the D_{inf} algorithm (Tarboton, 1997; Perron, 2010).

Many of the hilltops in the Gabilan Mesa are characterized by a low-relief surface defined by the Paso Robles formation, suggesting that, on the whole, this landscape is not at topographic equilibrium and that this approach may not be valid. To assess the quality of our assumption of steady state, we present results of simulations of one dimensional evolving topography. With these we demonstrate that while the hillslope studied is not likely to be in steady state, the current slopes are still likely to vary with

distance in a similar manner to those predicted by the steady-state approximation.

To compare our simulated landscapes to the real landscape of the Gabilan Mesa, we define a dimensionless number that compares the minimum amount of exhumation ($'min|E|'$ in Figure 2.5) that has occurred on a hillslope to the current relief on the hillslope. For a simulation evolving due to fixed base levels and uplift of an initially flat topography this is expressed as:

$$E^* = \frac{Ut}{Z_{crest}} \quad (2.9)$$

Where $t[t]$ is the total time that has elapsed in the simulation such that $U[Lt^{-1}]$ is the rate of uplift or base level lowering and $Z_{crest}[L]$ is the elevation at the crest of the hill. Due to the preservation of the Paso Robles surface in the Gabilan Mesa we can also define an equivalent quantity for our field site:

$$E^* = \frac{R_{surf}}{R_{hs}} \quad (2.10)$$

Where $R[L]$ is the relief and the subscripts *surf* and *hs* denote the relict land surface and hillslope, respectively. We define relief as the difference in elevation between the base of hillslopes and the Paso Robles surface and hill crest, respectively. The development of some subtle topography on this surface indicates that even where best preserved this surface has likely been lowered slightly. The crest of the hillslope studied here is approximately 20 - 30 m below this low-relief surface and has ~ 100 m of relief, suggesting it has an E^* value of 1.2.

In Figure 2.5 we present comparisons between the equilibrium slope and actual slope for values of E^* of 1.1, 1.2, 1.5, and ~ 4 for linear (Figure 2.5 B) and non-linear (Figure 2.5 C) simulations of evolving hillslope profiles. Equilibrium slopes are plotted on the y-axis, slopes at different stages of E^* are shown on the x-axis of each plot, and points are color coded for different E^* values. Dashed lines are contours of

current slopes being 100%, 80%, and 60% of equilibrium slopes.

In the range of E^* values similar to those of the study area, slopes are only 60 - 80% of steady state slopes. However, they are relatively consistent about this value (particularly in the non-linear case) along the length of the hillslope. That is, when using slope to solve for flux in these transient settings, the degree of misestimation of flux is similar along the length of the hillslope. In Figure 2.2, we are not trying to calibrate erosion rates or fluxes specifically; rather we are simply trying to use the current topography to model slopes as a continuous function of topographic position. For this reason we feel justified that the modeled slopes shown in Figure 2.2 provide a reasonable, continuous prediction for slopes along the hillslope from which we can measure residuals. Additionally, because the fluxes calculated from these modeled fits are not likely to be true steady-state fluxes, the parameters calibrated by minimizing the error between actual and predicted slopes are irrelevant and therefore not reported.

2.7.2 Colluvial hollows

An interesting feature of shadow bedding is that it appears to be crosscut by soil accumulations in colluvial hollows. That is, the oscillations in slope that define shadow bedding are truncated in areas of convergent hillslope topography. Figures 2.1 and 2.7 A and B highlight an example of a colluvial hollow cross cutting shadow bedding in a photograph and slope map of the same hillslope. Beneath the yellow star (and on the north end of Figure 2.7 B) is an area of planform concavity on an otherwise convex to planar hillslope. Shadow beds can be traced continuously along this hillslope until encountering this concavity, where they are terminated and reappear again on the opposite side at an equivalent topographic position.

In the main text we argue that shadow bedding arises such that variations in

slope balance variations in soil thickness to yield a uniform flux given a thickness-dependent soil flux rule. An end-member alternative to this equilibrium hypothesis is a purely transient one. Rather than reflecting thickness-dependent transport, perhaps shadow bedding is a remnant feature that has yet to equilibrate to the expected smooth, convex topographic forms characteristic of soil mantled hillslopes. Specifically, hillslopes in the Wildhorse Valley may still be responding to some episode in the past that thinned or stripped soils and produced a stepped bedrock or thinly soil-veiled topography. We argue here that the observation of soil accumulations in colluvial hollows that crosscut shadow bedding precludes this hypothesis. Specifically, that diffusion-like behavior of purely slope dependent soil transport (Culling, 1960, 1963) will act to smooth high-frequency topography before accumulating soil preferentially in colluvial hollows. To test this we present results of simulations from a two-dimensional hillslope model identical to the one presented in the main text (Figure 2.7 C).

We evolve these simulations from an initially stepped topography for ~ 15 ka model years given a thickness-independent transport rule (Culling, 1960, 1963) and the integrated soil velocity profile presented in Equations 2.4 and 2.5 of the main text. The initially stepped topography consists of slopes that oscillate between different lithologies. A uniform, thin mantle of soil (5 cm) is applied to the model, so that fluxes are non-zero at the start of the simulation. We represent different lithologies as having values of W_0 which vary 50% about a mean value. We introduce plan form curvature into the topography to represent colluvial hollows. The base of the model (y-Distance = 0) is fixed at 0 elevation; all other boundaries are periodic. We do not uplift the model or continually lower base level, as this would ultimately have the effect of removing the planform curvature - which is the feature we are fundamentally interested in observing.

In the thickness-independent model (Figure 2.7 i-iii), the initially stepped topography quickly begins to diffuse, and by 10 ka remnants of the initially stepped topography are present, but strongly subdued. By 15 ka there is no record of the initially stepped topography in the hollows or on the convex hillslopes. By contrast, in the thickness-dependent model (Figure 2.7 iv - vi), the initially stepped topography does not diffuse as quickly (as soil thicknesses and therefore fluxes are low), preserving the initially stepped topography in both the simulated hollows and convex portions of hillslopes. However by 15 ka, accumulations of soil in hollows has resulted in differential preservation of the formerly stepped topography. While subdued steps are still present on the convex portions of the slope, they are absent from the hollows. This occurs because accumulations of soil thickness push the hollows to low values of D^* , causing flux to be relatively insensitive to fluctuations in depth (Figure 2.3 B & i). While no results are shown here, when performing identical simulations with a plug-flow flux rule (e.g., $q = kH\nabla z$), the lithologic undulations are equally preserved in both the convex and concave portions of the landscape. Unlike in the integrated velocity profile (Figures 2.7 iv-vi), with the plug-flow rule variations in the thickness of soils require variations in slope to balance fluxes, regardless of soil thickness.

These simulations support our hypothesis of thickness-dependent soil flux, and in particular a transport rule that accounts for the decay in downslope soil velocities with depth in the soil column. Even in the case that hillslopes are equilibrating from a past event that produced stepped, bedrock topography, the observation that colluvial hollows cross cut shadow bedding requires a thickness-dependent transport rule.

2.8 Figures and tables

ID	Publication			Maximum Value	
Trees		Measurement ID	Species		
1	Danjon et al., 1999	Oak 1	Sessile oak <i>Quercus petraea</i>	2.81E+03	cm ³
2		Oak 2	Sessile oak, <i>Quercus petraea</i>	6.59E+02	cm ³
3		Oak 3	Sessile oak, <i>Quercus petraea</i>	2.59E+02	cm ³
4		Pine (30 trees)	Mean of 30x Maritime pine, <i>Pinus pinaster</i>	1.91E+02	cm ³
5	Danjon et al., 2008	Tree D	White oak, <i>Quercus alba</i>	1.11E+03	cm ²
6		Tree S	White oak, <i>Quercus alba</i>	2.83E+02	cm ²
7	Roering et al., 2010	Pine	Douglas fir, <i>Pseudotsuga menziesii</i>	5.30E-03	Root-soil area ratio
Burrowers		Measurement type			
8	Grinnell and Dixon, 1918	Burrow Area (from 3 sketches)	Ground squirrel, <i>Otospermophilus beecheyi</i>	6.21E+03	cm ²
9	Kolb, 1984	Burrow Area (from sketch)	European rabbit, <i>Oryctolagus cuniculus</i>	7.23E+02	cm ²
10	Miller, 1957	Burrow Volume	Pocket gopher, <i>Thomomys umbrinus</i>	5.23E+05	cm ³
11	Yensen et al., 1991	Burrow Area (from sketch)	Ground squirrel, <i>Spermophilus brunneus</i>	9.12E+02	cm ²
Soil Creep		Measurement ID	Measurement interval		
12	Fleming and Johnson, 1974	S11	1.17 yrs	3.76E+00	cm
13		S12	1.18 yrs	1.67E+00	cm
14		S18	1.14 yrs	1.02E+00	cm
15		S19	0.89 yrs	9.14E-01	cm
16		S20	1.14 yrs	1.02E+00	cm
17	Kirkby, 1967	Bluff	1-2 yrs	1.33E-01	cm yr ⁻¹
18		Peat	1-2 yrs	1.10E-01	cm yr ⁻¹
19		Rock	1-2 yrs	3.13E-01	cm yr ⁻¹
20		Till	1-2 yrs	1.54E-01	cm yr ⁻¹
21	Lewis, 1976	Marimonte	5.4 yrs	2.50E+01	cm
22		Infiernillo	4.4 yrs	1.70E+01	cm
23		Coama	5.4 yrs	2.54E+01	cm
24		LaPaguera	4.4 yrs	1.50E+01	cm
25	Moeyersons, 1987	Piedra	4.4 yrs	1.34E+01	cm
26			4 yrs	1.50E+00	cm

Table 2.1: Summary of data compiled and shown in Figure 2.3 B. ID corresponds to the legend shown in Figure 2.3 B. Measurement ID refers to the sample or location number in the publication. Maximum value gives the units and value used to normalize each data set. When appropriate, we specify either the species studied or the duration of the measurement interval over which creep rates or distances were measured.

Figure 2.1: Field photo and shaded slope map illustrate shadow bedding. Map shows locations of stratigraphic sections, locations of thickness measurements, and a prediction for the location of an easily correlated bed in the stratigraphic section. This prediction is the intersection of the topography with a plane fit to points surveyed along the base of that bed and is truncated at a small normal fault in the central arroyo. Slope map produced from 1m ALSM derived DEM collected and processed by NCALM. Additional field photos and stratigraphic column are available in the supplemental material.

Figure 2.2: GTL predictions plotted against a/b (A,C,E) and the residuals between models fits based on these predictions and data (B,D,F). Residuals are reported as the difference between observed and predicted slopes, S_o and S_p , or observed and predicted functions of slopes and thickness, f_o and f_p , and normalized to the mean observed value. Large circles are color-coded by soil thickness, small dots are points measured with the rod and were not used in any analysis. Dashed lines in B, D, and F show a linear regression with 95% confidence intervals. Spearman's rank correlation coefficient, r_s , and a p-value estimate for the no-correlation hypothesis are also reported in each plot.

Figure 2.3: A) Dimensionless flux as a function of D^* . Inset plots i-iv show hillslope simulations with flux calculated by Equation 2.4 and 2.5, black line is the land surface, gray line is the soil bedrock contact. Black bars show the D^* values for the range of thicknesses that arise over resistant (R) and recessive (r) layers. B) Variability in the activity of agents thought responsible for soil disturbance with depth. All values are normalized to the maximum reported value of each. Normalizing values and type of measurements are given in the supplement1. Heavy dashed line is a velocity profile (Equation 2.3) with a d_c of 33 cm. Data from: 1-4 Danjon et al. (1999); 5-6 Danjon et al. (2008); Roering et al. (2010); 8 Grinnell and Dixon (1918); 9 Kolb (1985); 10 Miller (1957); 11 Yensen et al. (1991); 12-16 Fleming and Johnson (1975); 17-20 Kirkby (1967); 21-25 Lewis (1976); 26 Moeyersons (1988).

Figure 2.4: Additional field photos documenting shadow bedding from the Northern Gabilan Mesa. A) Shadow bedded hillslope with cross cutting colluvial hollows south of Topo Valley Road. B) Shadow bedding is seen on hillslopes in the foreground and background of this South Facing Photo. Pancho Rico Creek forms the valley in the midground of the photo. In the foreground, shadow bedding is partially highlighted by variations in vegetation. C) Aerial photograph highlighting that shadow bedding can be traced for long distances along hillslopes and across valleys. Photo courtesy of Juan C. Fernandez Diaz

Figure 2.5: Transient slope evolution of linear and non-linear hillslopes. A) Depiction of terms used to compare real and modeled landscapes. Solid line depicts hillslope, dashed line shows the total uplift that has accrued or, alternatively, an incised landscape surface. E^* depends on the degree to which hillslopes have evolved beyond their initial condition and may be defined in the field utilizing an incised but relatively continuous geomorphic surface (the “mesa” of the Gabilan Mesa). The two presented equations for E^* may be used to calculate this value in different circumstances. In the field, direct measurements of the hillslope relief, R_{hs} , and the incised surface relief, R_{surf} , can be made. In simulations, the equivalent quantities are the elevation of the hillslope crest, Z_{crest} , and the product of the uplift rate, u , and the model time, t . Relief is defined relative to the elevation at the base of the hillslope. The crest of the hillslope studied here is approximately 20 - 30 m below the low-relief mesa surface and has 100m of relief, suggesting it has an E^* value of ~ 1.2 . B and C) Plots of the steady-state slope, S_{ss} , plotted against the slope at a specified simulation time, S_i , for different values of E^* . In B this is shown for a system evolving according to a linear transport rule, while in C the simulation was governed by a non-linear transport rule, as indicated by the titles of each plot. Dashed lines are contours of current slopes being 100%, 80% and 60% of equilibrium slopes.

Figure 2.6: Composite stratigraphic section from two sections measured at nearly equivalent stratigraphic positions in the Pancho Rico Formation. Sections were correlated based on the resistant, bioclastic conglomerate unit 16 m from the base of the displayed section. The expected location of this conglomerate in the landscape is shown in Figure 2.1, and was computed by least-squares fitting of a plane to the surveyed basal contact of this bed. Dotted outlines of stratigraphy indicate beds that could not be accessed directly in one section and were partly soil mantled in the other. Measurements of thickness in these top beds were made from photographs, theodolite, GPS surveys and grain-size profiles were partly inferred from weathering profiles and based on limited outcrop access. Grain sizes are reported for the matrix, scale is labelled as mud (md), very-fine lower and upper sand (vfL and vfU), fine lower and fine upper sand (fL and fU), and medium lower sand (mL). Shell and rare pebble clasts are seen in beds near the middle of the section. In the two thin beds between 14 and 16 meters, long axis of clasts are typically 1-2 cm in diameter. In the thicker bed, fossil clasts are typically ~5 - 10 cm and in some cases even larger.

Figure 2.7: Field observations and simulation results of interactions between colluvial hollows and shadow bedding. A) Photograph of a hillslope highlighting shadow bedding and a colluvial hollow. B) Slope map computed from 1m ALSM derived DEM collected and processed by NCALM. C) Results of simulations of evolution of stepped topography. Image at far left shows the slope of initial topography for simulations. Two columns show the evolution of slopes when soil transport is depth-independent (i-iii) and when applying the proposed transport law (iv-vi) with the corresponding flux rules labeled at the top of each column.

Chapter 3

Soil development over mud-rich rocks produces landscape-scale erosional instabilities*

3.1 Abstract

In soil-mantled landscapes the length of hillslopes is set by the relative efficiency of hillslope transport processes and incision by channelized flows. In cases where rapid channel incision exposes less weathered, more resistant rocks, incision should slow, creating a damping feedback that prevents run-away channel expansion. We present observations from the northern Gabilan Mesa (central Coast Ranges, CA), which is underlain by mud-rich rocks that may provide less resistance to erosion than the soils produced from them. This strength inversion arises because unconfined, mud-rich units of the Pancho Rico Fm. undergo a process called slaking, in which pervasive

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fractures, spaced centimeters apart, develop when subjected to wet-dry cycles. In contrast, consecutive wetting and drying has produced a cohesive regolith with higher tensile strength than the slaked bedrock. Exposure of the relatively weak bedrock may create a positive feedback, where the acceleration of erosion within the bedrock undermines the cohesive regolith and allows channels to expand headward at the expense of hillslopes. Geomorphic mapping reveals numerous channelized features that have incised deeply into bedrock, and cross-cutting relationships and topographic analyses indicate that these features have consumed regions of the landscape previously occupied by hillslopes. However, these features are only present in south flowing drainages, where soil is thinnest and therefore bedrock is most likely to be accessed by abrupt erosion. Preferential headward expansion of these south flowing channels has established a profound topographic asymmetry, where south flowing tributaries occupy $\sim 75\%$ of drainage basin width. These results suggest that processes operating at the scale of the critical zone, which may be unique to certain rock types, conspire to produce an instability that shapes the morphology of drainage basins.

3.2 Introduction

Weathered rock products derived from soil mantled landscapes supply sediment to rivers and low-lying depositional areas (Gilbert, 1877), provide nutrients to soil through chemical weathering and landscape lowering (Walker and Syers, 1976; Vitousek et al., 2003), and therefore act as a medium for agriculture. It is usually assumed that the physical and chemical disaggregation of the underlying rock creates a soil whose physical strength is less than that of the parent bedrock (e.g., Howard, 1994), which facilitates the downslope transport of this now-mobile material (Gilbert, 1877; Davis, 1892). In this situation, rapid removal of the soil by channel incision exposes strong bedrock, which in turn slows the rate of incision. The damping effect of exposing

strong bedrock prevents the unstable growth of channel networks. As a result, even in the face of small perturbations in erosion rate, a steady-state might ultimately be achieved in which landscape relief and transitions between hillslopes and channels remain relatively invariant with time (Smith and Bretherton, 1972; Perron et al., 2009).

Although this view of soil-mantled landscape development appears sensible in many natural landscapes (Perron et al., 2009; Chadwick et al., 2013), past research on soils, and recent research on processes of bedrock disaggregation challenge the view that the bedrock underlying a landscape is always more resistant to incision than the soils formed from it. In some arid environments soils may be stabilized by silica or carbonate cements that have been precipitated during soil formation (e.g., Machette, 1985), which can entomb landscape forms as erosion is retarded (Alonso-Zarza et al., 1998; Eppes et al., 2002). Likewise, bedrock lithologies that contain swelling clays can undergo profound disaggregation upon several wetting and drying cycles through the process of slaking (Montgomery, 2004; Stock et al., 2005; Johnson and Finnegan, 2015). It is plausible, then, that the confluence of both factors - the weakening of particular lithologies by slaking and the hardening of soils by pedogenesis - might create soils that are more resistant to incision than bedrock. In such circumstances, erosion that exposes the weaker, underlying bedrock would accelerate incision that could undermine the cohesive soil mantle, allowing channels to expand their extent at the expense of hillslopes and reconfigure once stable drainage networks.

For a landscape where the soil mantle offered greater erosional resistance than the bedrock, unstable erosion through this bedrock would occur when incision rates in excess of soil-production rates (defined here as the rate at which bedrock is transformed to a cohesive soil mantle) were sustained long enough to strip soils and access bedrock. In arid and semi-arid climates, variability in solar radiation as a function of slope aspect can drive hydrologic and ecologic variations (Istanbulluoglu et al., 2008;

Gutiérrez-Jurado et al., 2013). If soil production is sensitive to these changes in the local water balance, the triggering of unstable erosion may have some relation to solar insolation, and hence hillslope aspect.

Using observations from physical samples and the landscape itself, we argue that parts of the Gabilan Mesa, central Coast Ranges, California, represent a case where the soil mantle is more resistant to erosion than the bedrock from which it formed. At the scale of individual samples, laboratory experiments demonstrate that soils here have higher tensile strength than the parent bedrock, once this rock has undergone wetting and drying cycles. Soils tend to be relatively thin on south-facing slopes, and in south-flowing catchments we document the presence of incisional features that expose bedrock in their upstream sections. The boundaries of these incisional features cross-cut smooth, convex topography and topographic analysis suggests that the channels hosting these features extend upslope into regions of the landscape otherwise occupied by soil-mantled hillslopes. Thus it appears that these bedrock-incising features consume hillslopes through headward migration. We propose that this is the manifestation of the unstable erosion through bedrock expected when incision through a relatively resistant surface encounters readily erodible material. Conversely, soils on north-facing slopes are more than three times as thick as those observed on south-facing slopes, and north-flowing tributaries do not contain these same bedrock-incising channels.

These aspect-dependent differences in process appear to produce corresponding differences in watershed-scale landscape morphology. At this scale, we highlight (as others have; Dohrenwend, 1978) that south-facing watersheds are longer than those that face north, creating a landscape-scale asymmetry in basin geometry. This is consistent with the idea that incisional instabilities are preferentially triggered on south facing slopes, where soils are thin. Once the incisional instabilities are generated within south-facing basins, their headward growth may, in part, lengthen these basins

to produce the observed asymmetry. These observations and inferences suggest that, in some cases, local processes of rock weathering and soil genesis may leave a strong imprint on the geometric configuration of soil-mantled landscapes.

3.3 Geologic setting

We focus on the northern Gabilan Mesa (central California; Fig. 3.1), which is bordered to the east by the San Andreas Fault and the west by the Salinas Valley. The summits of the Gabilan Mesa are defined by a low relief surface sloping to the south-east at $\sim 3^\circ$ (Fig. 3.1). This surface is mapped as the slightly lowered depositional surface of the Paso Robles Fm., a Pliocene to (?) Pleistocene non-marine deposit extensive throughout the central California Coast Ranges (Galehouse, 1967; Dohrenwend, 1974; Durham, 1974). Provenance analysis and paleocurrent reconstructions of the Paso Robles Fm. indicate that the transport depositing this sediment at the northern limit of Fig. 1 was largely directed toward the west (Dohrenwend, 1974) but transitioned to being south to south-east directed starting near the southern limit of Fig. 3.1 (Galehouse, 1967). Hence in some locations the slope of the Gabilan Mesa surface may partly reflect the original angle of deposition.

The Gabilan Mesa drainage basins depicted in Fig. 3.1 that have incised beneath the Paso Robles Fm. are hosted within the Pliocene Pancho Rico Fm., excepting limited exposures of Salinian basement that immediately underlie the Paso Robles Fm. at the north end of Fig. 3.1, and a few limited outcrops deeper in the section to the south. The Pancho Rico Fm. contains sandstones, mudstones, and conglomerates. Sandstones and mudstones are typically massive and light yellow to white in color. Sandstones are fine- to very fine-grained with abundant clay, and mudstones often contain abundant diatomite. Conglomerates are often fossiliferous and exist as decimeter- to meter- scale beds within sandstones and mudstones (Durham and

Addicot, 1965).

3.4 Geomorphologic background

The nearby town of King City, Ca receives ~ 300 mm of precipitation a year, most of which falls in the winter. In the summer average temperatures vary between 12°C and 30°C , cooling in the winter to between 4°C and 18°C , although the Pleistocene climate of much of California may have been cooler and wetter (Johnson, 1977). South-facing slopes are currently dominated by grasslands, while north-facing slopes host oak trees (*Quercus wislizenii*) (Fig. 3.2 C).

The inclined mesa surface appears to have preconditioned the landscape to form a series of west-southwest draining catchments, parallel to the dip direction of the surface (Fig. 3.1, Dohrenwend, 1978). Soils mantle most of the study region, although resistant units within the Paso Robles Fm. are present near the margins of the mesa surface, and channels often expose the Pancho Rico Fm.. The floors of large valleys within the Gabilan Mesa are composed of multiple generations of inset fluvial surfaces (Dohrenwend, 1974). The chronology and stratigraphy of at least one of these surfaces suggests that their formation is related to variations in sediment delivery to channels during times of known climate change (García and Mahan, 2009). Other work on these surfaces concluded that the Pancho Rico Fm. is susceptible to lateral erosion and strath formation because of its lack of resistance, and in particular its rapid weathering due to slaking (Garcia, 2006).

One of the unique characteristics of the Gabilan Mesa is profound north-south topographic asymmetry (Dohrenwend, 1974; Kane, 1978; Poulos et al., 2012). North-facing hillslopes are significantly shorter and steeper than their south-facing counterparts. South-facing slopes also show a much larger degree of drainage dissection.

Aspect-dependent differences in landscape morphology are not uncommon, and barring tectonic or geologic control are typically attributed to variations in the relative intensity of different geomorphic processes across the microclimatic gradients that arise on opposing slopes due to variations in incoming solar radiation (Bass, 1929; Istanbulluoglu et al., 2008; McGuire et al., 2014; West et al., 2014; Yetemen et al., 2015). Early work was divided on the cause of topographic asymmetry in the Gabilan Mesa. In one model, a slight increase in the efficiency of channel incision on south-facing slopes produced a larger flux of sediment from south-flowing tributaries that drove lateral incision of trunk channels into north-facing slopes (Dohrenwend, 1978). Other research simply favored contrasts in the efficiency of erosion by overland flow (Kane, 1978), and recent developments suggest multiple mechanisms may be at play (Richardson, 2015).

Many low-order tributaries in the Gabilan Mesa are characterized by breaks in slope that separate smooth, convex to planar hillslopes or broad concave valleys mantled by soil from planar to concave bedrock-floored or scree-mantled slopes (Fig. 3.2 B). These breaks in slope were termed knicklines by García and Mahan (2009). These authors found that the heads of knicklines often extend into the hillslope domain, where bedrock is commonly exposed in near-vertical cliffs. Based on both their cross-sectional and longitudinal morphologies, García and Mahan (2009) interpreted these features to be largely sculpted by debris flows. We refer to these bedrock-floored or scree-mantled incisional features as arroyos.

3.5 Methods

Measurements were aimed at assessing whether soils in the Gabilan Mesa offer greater resistance to erosion than the bedrock they mantle and if this configuration has shaped the morphology of this landscape. We first measure physical properties of the soils and

rocks to infer their erodibility when exposed and document variability in the extent of soil cover. Next we use geomorphic maps and topographic data to determine whether bedrock-incising channels extend valley topography farther upslope than is observed in soil-mantled channels, as would be expected for the case of a runaway instability produced by easily eroded bedrock overlain by an erosion-resistant soil. Finally, we analyze tributary basins that flow south and north, where features interpreted to reflect unstable erosion are and are not present, respectively, to link the distribution of erosional processes to the geometry of the landscape.

3.5.1 Characterizing the extent and physical properties of surface materials

We use brazilian splitting tests to measure the tensile strength of soil and rock (Vutukuri and Lama, 1974; Sklar and Dietrich, 2001), and use this to infer whether soils in the Gabilan Mesa may offer greater resistance to erosion than the underlying Pancho Rico Fm.. Samples of Pancho Rico Fm. and cohesive blocks of soil were collected from exposed bedrock faces at the heads of arroyos and adjacent hillslopes. Eighteen ~ 44 mm diameter Pancho Rico Fm. cores were exposed to a single wetting (submerged in ~ 30 mm of water for 3 hours) and drying (air dried at room temperature for 24 hours) cycle to assess the effect of oscillations in moisture content on strength. We used fracture spacing to measure the impact of slaking on collected cores - this was measured by dividing the maximum number of fractures intersecting a line defining the diameter of the core by this line's length. The remaining 14 cores of Pancho Rico Fm. were tested without being subjected to wetting and drying. We also measured the strength of 15 cores from top-soil blocks. The readout on the Brazilian testing apparatus measured load with 0.01 kN precision, however at low loads the readout frequently fluctuated ± 0.02 kN. Therefore, we assign a detection limit of 0.03 kN.

For the size of the cores analyzed in this study this corresponds to a tensile strength of 0.02 MPa.

We documented differences in the particle size distributions of soils using samples from prominent horizons identified in 8 soil pits collected on convex portions of hillslopes: 3 pits with 13 samples from north facing slopes and 5 pits with 14 samples from south facing slopes. Attributes of soils were assigned to the full depth range, measured vertically from the land surface down, of the horizon from which they were sampled. Soil analyses were performed by the Utah State University Analytical Lab.

We measured soil thickness within 168 soil pits in the Gabilan Mesa, 152 of which are from the Wildhorse Valley (this includes thicknesses reported in Johnstone and Hilley, 2015) to identify variations in the extent of cover above bedrock. The base of the soil column was identified as the top of an abrupt transition to fractured and/or altered bedrock. Soil thickness was measured vertically in the field and converted to a surface-normal thickness using the local topographic slope measured with digital elevation data.

3.5.2 Relating geomorphic process distribution to landscape form

Geomorphic mapping

We document the distribution of geomorphic domains at two scales, first at the scale of individual arroyos (e.g. Fig. 3.2 A) and then at the drainage basin scale. Detailed geomorphic maps focused on the morphology below the breaks in slope bounding arroyos. These maps were constructed in the field at a scale of 1:1,200 in all but the case of the largest arroyo, which was mapped at 1:3,000 (Fig. 3.3). Detailed mapping locations were selected to span a range of arroyo sizes and apparent stages of development.

Below the breaks in slope that bound arroyos (i.e., the knicklines of Garca and Mahan, García and Mahan, 2009), we categorized the surface as falling within one of four regimes in detailed geomorphic maps: bare bedrock; bedrock covered by thin, patchy mantles of scree; well established and continuous scree accumulations; and thick scree aprons with well-established vegetation (Fig. 4). At the outlets of arroyos, we identified fans as features of positive relief extending into mainstem valleys with positive fanning dips. Inset into, or on top of scree slopes, we identified cones of debris significantly steeper than fans, which we identified as debris fans. Debris fans were often found at the bases of rills that dissect the scree slopes. At the margins of arroyos, where incision has exposed the soil-bedrock contact, we identified pedogenic carbonates at the soil-bedrock contact and assigned a qualitative scaling from 1 - 3 based on the thickness and continuity of these accumulations (e.g., Machette, 1985). Values of 1 represent very thin, mm-scale, carbonate coatings at the soil bedrock contact and on fracture surfaces, while values of 3 represent continuous accumulations > 2 cm thick. A value of 2 was assigned to conditions between these values (Fig. 3.4).

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bedrock contact and on fracture surfaces, whereas values of 3 represent continuous accumulations > 2 cm thick. A value of 2 was assigned to conditions between these values (Fig. 3.4).

Detailed field mapping efforts were supplemented by remote geomorphic mapping of the ~ 40 km² Wildhorse Valley at a scale of 1:3,000. Remote mapping relied on 1-m Airborne Laser Swath Mapping (ALSM) elevation data and 1-m-resolution aerial imagery from the National Agriculture Imagery Program (NAIP) (Fig. 3.3). Contacts were mapped by identifying concave and convex breaks in slope that separate mappable geomorphic domains. In some cases, slopes vary gradually between distinct morphologic regions, or changes in slope direction separate distinct domains - domains were separated at these locations accordingly. Where the contact between domains is uncertain or gradational, we indicated this in the extent of the mapped feature.

In remote geomorphic mapping of the Wildhorse Valley, we lumped the features categorizing the surface of arroyos documented in the detailed geomorphic maps into a single morphologic unit designated “arroyo” (Fig. 3.3). In addition to arroyos, we identified 4 other geomorphic domains. Low-elevation, low relief valley-floor surfaces whose width greatly exceeds that of the trunk-stream channels were designated terrace surfaces, and assigned a relative age based on inset relationships. Second, we identified fans as generally conical landforms forming positive relief on terrace surfaces whose up-slope portions often extend into constricted valleys. Third, we demarcated convergent topography that merges in flat, U-shaped, or in rare cases, slightly domed valley floors that lack evidence of self-constructing channel features as colluvial hollows (Hack and Goodlett, 1960; Dietrich and Dunne, 1978). Finally, regions where hillslopes meet valley floors in steep, wedge shapes were labeled as toe-slope aprons.

The Pancho Rico Fm. is a light-colored unit, and exposures of this unit in satellite imagery appear as white regions that contrast with the green, tan, and

brown hues of the surrounding grassland. The scree-mantled slopes found within arroyos maintain this white color and contrast with the surrounding hillslopes (Fig. 3.2 A & C). We used this observation to map the slope breaks bounding arroyos across much of the northern Gabilan Mesa using 1-m aerial imagery (Fig. 3.1).

3.5.3 Topographic analysis

We measure channel morphology within different geomorphic domains with morphometrics computed along longitudinal profiles. A single longitudinal profile is derived for each of the major tributary basins within the Wildhorse Valley. We identified major tributary basins by determining the extent of all non-overlapping tributary basins whose drainage area, A [m^2], is greater than or equal to 0.05 km^2 and less than 0.5 km^2 , where drainage area was calculated with a D_8 flow routing scheme. We excluded all tributary basins that did not include any mapped geomorphic domains and whose relief did not exceed 100 m. These ranges of drainage area and relief were selected to capture the range of tributary sizes common in the Wildhorse Valley and to avoid basins defined by the D_8 routing that do not span the majority of the relief between the Wildhorse Valley bottom and drainage divides. Longitudinal profiles were computed by following the path of greatest drainage area upstream from each tributary basin outlet. And we associate topographic attributes along this profile with the mapped geomorphic domain that contains them. We excluded depositional landscape elements by masking out points that fell within mapped fan and terrace units.

We identify variations in hillslope lengths from longitudinal profiles in plots of curvature as a function of the product of drainage area and slope, AS [m^2]. We compute the longitudinal slope along this path by measuring the average slope, S , within 3-m-vertical windows (Wobus et al., 2006) and curvature was determined from

the parameters of the parabolic surfaces that best fit elevation data within a 7-m-radius window (e.g., Perron et al., 2009). Past work recognizes that where hillslopes transition to valleys, a point we refer to as AS_{crit} , curvatures tend to increase with AS (Perron et al., 2009). With this in mind, we fit a piecewise linear function to these data that minimizes the sum of the squared residuals. We identified the AS value at which curvatures begin to steadily increase (AS_{crit}) as the intersection point between these flat and sloping linear segments.

Channels often obey a power law scaling between slope and drainage area (Flint, 1974), and the parameters of this power law have been used to distinguish areas of landscapes dominated by different processes (Stock and Dietrich, 2003). We fit two power law functions to slope area data from different mapped geomorphic units; each power law has the form:

$$S = kA^{-\theta} \quad (3.1)$$

where k [$\text{m}^{2\theta}$] is typically referred to as the ‘steepness’ and θ is the ‘concavity’. The two power law regressions are equivalent at the drainage area that separates them, A_t [m^2]. We identified the four best fitting parameters that minimize the sum of the squared errors between the regression and data.

3.5.4 Relating geomorphic process and landscape morphology to aspect

We measure aspect and use it to classify measurements to contrast landscape properties associated with different microclimates. We define 3 aspect classes and group data within these classes. North- and south-facing aspect classes contain data whose aspects fall within 45° of north or south, respectively. The remaining data are classified as east- or west-facing.

The aspect at individual soil pits was computed from the the north and east

components of slope determined from the parabola that best fit elevations within a prescribed window. Where 1-m data was available we use a 7-m radius window, with 10-m data we use a 3x3 pixel window. We assign an aspect to tributary basins and longitudinal profiles by computing the plan-view orientation of a straight line connecting the outlet of that drainage basin to the point along the longitudinal profile where the upstream area first exceeds 1,000 m² (this value was chosen to minimize the influence of divergent, low drainage area portions of the profile).

We calculated the fraction of tributary basin areas occupied by different mapped geomorphic process domains. We compare these distributions among north- and south-flowing tributaries to assess how micro-climates have influenced their formation. To assess differences in landscape morphology with aspect we contrast longitudinal profile morphometrics (see previous section) in basins that flow north and south.

To highlight how the cross-sectional asymmetry of drainage basins developed through incision into the Paso Robles surface we construct topographic profiles of the mean elevations within a prescribed radius of section lines from 10 m elevation data using a 15 m radius swath along paired ridge and valley profiles extending from the northern to the southern boundary of basins (Fig. 3.1). We select locations of basin profiles to highlight asymmetric regions, particularly where the preservation of the mesa surface along ridges provides constraints on channel incision during landscape development. We identify the limit of the Paso Robles surface in these profiles as the point separating a broad, low slope surface from steep valley walls. We reference the elevations and the distance along swath profiles to the location of the lowest point along the swath, and normalized distance and elevation by the total length and relief.

Differences between distributions of metrics within aspect classes are evaluated by recording the larger p -value (p_{max}) of two tests: the Wilcoxon Rank Sum test (for differences in median values) and the two-sample t-test (for differences in mean values). We restrict comparisons to data within north- and south-facing aspect classes.

We assign significance when $p_{max} < 0.001$.

3.6 Results

3.6.1 Characterizing the extent and physical properties of surface materials

The tensile strength of dry Pancho Rico Fm., measured in 12 of the 14 samples analyzed, varied from 0.30 - 0.75 MPa (median = 0.51 MPa), which is comparable to other weak sedimentary rocks (Sklar and Dietrich, 2001; Johnson and Finnegan, 2015). Two of the 14 samples fractured in a manner inconsistent with expectations for the Brazilian splitting test, either developing a fracture at an angle to the direction of loading or developing a fracture that didn't fully separate the sample; these data were excluded from analyses.

All but one of the cores inundated in water developed through-going fractures within 10 minutes of being submerged - these cores have a tensile strength of zero (Johnson and Finnegan, 2015) in the context of the Brazilian splitting test. The remaining core developed a fracture extending nearly half of the cores diameter in the first ten minutes, which grew to span $\frac{3}{4}$ of the diameter after 3 hours of submersion. Within the first ten minutes of wetting, cores developed fracture spacings of 4 - 42 mm (median: 7 mm). After 3 hours of submersion, the median fracture spacing was unchanged, but 4 of the cores had shattered beyond our ability to identify individual fractures.

Of the 15 cohesive soil samples 12 developed a single through-going fracture parallel to the direction of loading, consistent with the expectations of the Brazilian splitting test. The tensile strength varied from 0.02 - 0.11 MPa (median = 0.06 MPa) among soil cores, a significantly smaller distribution than the unslaked bedrock (p_{max}

$= 3.7 \times 10^{-5}$) and larger than the slaked bedrock ($p_{max} = 2.8 \times 10^{-7}$).

Particle size distributions are shown and summarized for soil pits on north- and south-facing slopes in Fig. 3.5. Within individual profiles particle size distributions are fairly uniform across a range of depths. While there are exceptions, there are no bulk trends observed in this small number of samples. Silt content is relatively similar between profiles from north- and south-facing slopes. However, soil pits on north-facing slopes contain smaller fractions of sand and larger fractions of clay (Fig. 3.5). On south-facing slopes, field tests with hydrochloric acid revealed carbonate either distributed throughout soil columns or found in high concentrations at the soil bedrock contact in the form of continuous horizons, similar to those mapped around arroyos (Fig. 3.4). On north-facing slopes, carbonate coated root casts were present near the base of some deep profiles and carbonate was distributed in some soil profiles at significant depths (e.g., $> \sim 1$ m), but not detected in shallower parts of profiles.

Distributions of soil thickness within different aspect classes are shown in Fig. 6. When correcting for slope with 10-m elevation data, soil thickness distributions on north- and south-facing hillslopes are significantly different ($p_{max} = 5 \times 10^{-15}$).

3.6.2 Relating geomorphic process distribution to landscape Form

Geomorphic mapping

Detailed geomorphic maps show that bedrock is primarily exposed near steep headwall knick-zones that separate downstream alluvial channels from a generally convex-up, low-relief landscape upslope (Fig. 3.4). In many cases, meter-scale beds of sandstone and bioclastic conglomerate form a prominent vertical scarp that typically exceeds the scale of these resistant beds. The tops of these scarps are often defined by the exposure of the more resistant beds, with less-resistant mudstone exposed in the

vertical face beneath. In some of these cases, the resistant beds overhang the less resistant stratigraphy. Areas adjacent to the channel are typically mantled with scree that extends up to or near the break in slope that defines arroyos. In some arroyos, Pancho Rico Fm. bedrock is exposed between the limit of these scree aprons and the bounding slope break (Fig. 3.4). Scree mantles range from thin, discontinuous veneers to accumulations at least 10s-of-centimeters thick that support vegetation. Thicker, more vegetated scree accumulations are more common near the outlets of arroyos, and occupy a greater fraction of large arroyos than of small ones. Small fan-shaped accumulations of fresh talus at the foot of slope beneath rills suggest that granular flows may transport sediment across the scree aprons to channels, and at least one of these fresh talus accumulations was deposited following a rainstorm during the course of this fieldwork. Where pedogenic carbonates are present at exposed soil-bedrock contacts, they can commonly be traced laterally around a large fraction of the arroyo (Fig. 3.4).

Arroyos are common throughout the Gabilan Mesa, including in the Wildhorse Valley (Fig. 3.1 and 3.3). The slope-breaks that define the margins of arroyos' scree aprons and bedrock knick-zones truncate a lower-relief, generally convex-up landscape (Fig. 3.2 A and C). In some cases, concave, soil mantled topography and channels are upstream of the heads of arroyos. However, in many other cases the arroyo head knickpoint is directly downstream from convex soil mantled slopes. In a few cases the head of an arroyo appears to have penetrated beyond where watershed divides project from adjacent, truncated ridgelines (Fig. 3.3 A). On the walls of arroyos, and occasionally extending above their bounding slope break, we typically found closely spaced, parallel to sub-parallel channels.

We identified two low-relief valley floor surfaces. An older terrace surface (Q_{tso}) extends throughout the valley, but occupies the majority of the valley extent in the

upstream areas of Wildhorse Valley. This geomorphic surface is truncated by a necessarily younger terrace surface (Q_{tsy}), which is found exclusively in the lower reaches of the watershed. The position and inset relationship of Q_{tsy} suggests that headward incision of the downstream portion of Wildhorse Creek is creating the Q_{tsy} surface, which largely replaces the Q_{tso} surface (Fig. 3.3). Arroyos are observed both upstream and downstream of this change in relative base level, however arroyos proximal to the limit of this incision have escarpments that can be traced directly to incised terrace surface margins.

Fans form positive relief along a substantial areal fraction of the Q_{tso} and Q_{tsy} surfaces (Fig. 3.3). The downstream limits of these landforms are hosted on what field observations reveal to be largely aggradational terraces.

Topographic analysis

We identified 129 drainage basins containing mapped geomorphic units (Fig. 3.3). In drainage area-slope plots arroyo and colluvial hollow data show a steeper decline of slope with drainage area at high drainage areas than at low drainage areas (Fig. 3.7). Arroyos are typically steeper than colluvial hollows and show a more rapid decline in slope with drainage area (Fig. 3.7). Relevant components of power-law fits are summarized in Table 3.1. Plots of curvature versus AS show that the AS_{crit} value differs between mapped geomorphic process domains (Fig. 3.7), as summarized in Table 3.1. The smallest AS_{crit} is seen in arroyos (41 m^2) and these features have positive curvature values across a broad range of AS . The AS_{crit} value of colluvial hollows is the largest observed ($1.3 \times 10^4 \text{ m}^2$) and curvature within these features is slightly positive (mean: 0.02) and does not vary strongly as a function of AS .

3.6.3 Relating geomorphic process and landscape morphology to aspect

Incision into the Q_{tso} surface is associated with different geomorphic domains on north- and south-facing slopes. Near the midpoint of the Wildhorse Valley, just downstream of a major tributary junction, incision and lateral channel migration appear to have undercut both north- and south-facing hillslopes (Fig. 3.3 C). On the undercut south-facing hillslopes, bedrock is exposed on hillsides with scree-mantled slopes and in arroyos. On opposing north-facing slopes, a clear break in slope separates the characteristically planar north-facing slopes from a series of three concave depressions immediately upslope from where the mainstem channel is adjacent to the valley sidewall. Colluvial hollows fill these concave depressions.

Of the tributaries identified, 32 and 67 were identified as north- and south-flowing, respectively. The drainage area of north-flowing tributary basins varies from 0.05 - 0.27 km² (median: 0.09 km²) while south-flowing tributary basins range in drainage area between 0.05 and 0.5 km² (median: 0.18 km²), the maximum range allowed by our method. These populations of tributary size are significantly different ($p_{max} = 7.7 \times 10^{-5}$). While arroyos were found throughout the Wildhorse Valley, they are almost exclusively hosted within south-flowing drainage basins, where they occupy a 15% (median) of tributary basin areas (Fig. 3.8). With three exceptions, arroyos are absent from north-flowing tributaries, and in those cases where an arroyo is present in a north-flowing tributary, it is limited to the more south- or west-facing aspects within these watersheds (Fig. 3.3). Distributions of the relative extent of arroyos within north- and south-facing tributaries are significantly different ($p_{max} = 7 \times 10^{-15}$). Colluvial hollows occupy a larger fraction of north-flowing drainage basins, a median of 7-10%, than those drainage basins flowing south (median of <1%, $p_{max} = 4 \times 10^{-19}$, Fig. 3.8).

Longitudinal profile morphometrics highlight differences in the morphology of north- and south-flowing tributary basins, whether including or excluding data within mapped colluvial hollows and arroyos (Fig. 3.7 & Table 3.1). In all cases north- and south-flowing data show a transition to a more rapid decline in slope with increasing area at intermediate areas (Fig. 3.7). However this transition, A_t , occurs at a larger area in south-flowing tributary basins (Table 3.1). Including arroyos and colluvial hollows in measurements shifts the mean slope and curvatures of south facing basins up for equivalent values of A and AS , respectively (Fig. 3.7). Including these data does not produce as large a shift in data from north-flowing basins. Curvatures begin to increase more steeply with AS at large values of AS in both north- and south-flowing basins, however this transition is subdued in north-flowing data (Fig. 3.7). The AS_{crit} value of south-flowing basins is smaller than north-flowing basins, an effect that becomes more pronounced when including data from colluvial hollows and arroyos (Table 3.1).

We present paired ridge and valley profiles traversing basins throughout the Gabilan Mesa (Fig. 3.1 & 3.9). In these examples, south-flowing portions of basins occupy 57-88% (median: 74%) of the total distance across drainage basins (50% would define a symmetrical basin). This narrow range includes an example from one of the largest and most visually asymmetric basins, where the Paso Robles surface is no longer preserved (Fig. 3.1). Where the Paso Robles surface remains (i.e., excluding the example mentioned above), 21-62% (median: 37%) of the basin width is occupied by south-flowing channels south of the limit of this surface (Fig. 3 & 11). The northward penetration of channels into this surface accounts for 12-58% (median: 37%) of the ridge-to-ridge distance.

3.7 Discussion

3.7.1 Strength of soils and the Pancho Rico Formation

The tensile strength of dry Pancho Rico Fm. is ~ 0.5 MPa, similar to what is observed in other poorly consolidated and fine-grained sedimentary rocks, including mudstones (Sklar and Dietrich, 2001; Johnson and Finnegan, 2015). This is greater than the tensile strength measured in dry cores from cohesive topsoil blocks. While one of the tensile strengths measured in the soil cores was not distinguishable from 0 (below our detection limit), the majority of cores recorded some, albeit low, internal strength. However, these soils formed in an environment where they were repeatedly wetted and dried. In contrast, cores of Pancho Rico Fm. bedrock that we saturated rapidly developed closely spaced fractures, permanently eliminating the tensile strength of these cores.

While tensile strength has been shown to be a good predictor of the resistance of bedrock to erosion by saltating bedload (Sklar and Dietrich, 2001), there is no evidence suggesting that this is the mechanism responsible for channel erosion in upland soil-mantled channels and within arroyos. Further efforts to assess the properties of surface materials that govern the resistance to and mechanics of incision, and techniques to measure these properties (Sklar and Dietrich, 2001; Bursztyn et al., 2015), will greatly aid in our understanding of how erosion sculpts the heterogeneous surface of the earth.

The majority (12 of the 15 tested) of soil cores subjected to the Brazilian splitting test fractured as expected for these experiments, despite the fact that over time-scales longer than the duration of these experiments (on the order of minutes) creep may dissipate stresses in soils (Vutukuri and Lama, 1974; Selby, 1993). Rather than showing evidence of internal deformation, the remaining 3 cores developed auxiliary fractures, perhaps due to the weak and anisotropic nature of these samples. If creep

did occur in any of these tests, the estimated tensile strength should be regarded as a minimum. Assuming that tensile strength provides some indication of the susceptibility of materials to incision (Sklar and Dietrich, 2001; Bursztyn et al., 2015), our measurements confirm that oscillations in moisture content produce soils that are cohesive and stronger than the slaked Pancho Rico Fm. Similar behavior has been seen in other mudstones (e.g., Franklin and Chandra, 1972), and in the context of landscape development has been argued to facilitate the lateral erosion of bedrock rivers (Montgomery, 2004; Johnson and Finnegan, 2015; Schanz and Montgomery, 2016).

We simulated the slaking process by rapidly inundating cores in water. Given the arid climate in the Gabilan Mesa this may be a poor representation of conditions that produce slaking in this landscape. However we observed that Pancho Rico Fm. exhumed by ground squirrels developed networks of through-going fractures after rainstorms, which divide clasts into a number of smaller fragments. Some fractures cross cut squirrel tooth marks, establishing that the fractures post-date their exposure at the surface and affirm that the slaking process occurs in this environment.

Past work highlights at least three mechanisms for imparting soil strength, or more generally erosional resistance. First, the growth of root networks increases the resistance of soils to detachment (Vannoppen et al., 2015). Second, carbonate distributed within the soil column cements soil and creates an armored surface (Alonso-Zarza et al., 1998; Eppes et al., 2002). Alternatively, perhaps wetting and drying of clays produces mineralogic configurations that impart cohesiveness to the soil (e.g., Sposito et al., 1999). We did not perform the micro-scale characterization of soils to determine the mechanism by which cohesiveness develops in these landscape elements, and so one or more of these processes may be responsible for the soil cohesion we observed. Further characterization of the structure of these soils may help to ascribe a particular set of mechanisms to this mechanical transformation.

3.7.2 Variations in soils with facing direction

Soil thicknesses vary between north- and south-facing slopes in a variety of landscapes (e.g., West et al., 2014; Olyphant et al., 2016) including the Gabilan Mesa (Fig. 3.6). This suggests that variations in incoming solar radiation impact the processes that govern soil production and control soil thickness in these settings. Our analysis of soil properties was limited to a small number of samples; however, across all samples and nearly all depths, soils on south-facing slopes contained higher sand and lower clay fractions than those on north-facing slopes. If the variations in grain size within these profiles reflects weathering of similar source materials, the reduction in grain size on north-facing slopes suggests that weathering on these aspects is more intense (Birke-land, 1984). Variations in the distribution of carbonate are consistent with deeper cycling of water on north- versus south-facing slopes. Greater water penetration and retention on north-facing hillslopes may favor enhanced weathering and deeper soils, and soils in the Gabilan Mesa have been observed to retain greater moisture on north-facing slopes (Kane, 1978). Alternatively, variations in the water balance also drive changes in the vegetation residing on north- and south-facing slopes, with larger woody vegetation restricted to north-facing slopes (Fig. 3.2 C). Plant roots may facilitate the breakdown of fresh rock and the transport of soil (Roering et al., 2010), and so it is possible that the oak trees that typically reside on north-facing slopes expedite the breakdown of rocks beyond what can be accomplished solely by the grasses currently found on south-facing slopes.

3.7.3 Synthesis and interpretations of mapping and topographic analysis

The floor of the Wildhorse Valley consists of at least two terrace surfaces. Field observations and analogies with valleys to the south suggest that these surfaces reflect

some combination of lateral erosion and infilling (García and Mahan, 2009). The younger of these surfaces (Q_{tsy}) is inset and incising into the older surface (Q_{tso}), which is correlative with large terraces of the Salinas River preserved within the Salinas Valley. Other terraces within the Gabilan Mesa that appear to grade to a similar elevation above the Salinas River were dated with optically stimulated luminescence to be $\sim 12\text{-}15$ ka (García and Mahan, 2009). This is coincident with the end of the last glacial maximum, when a stormier climate appears to have triggered the headward expansion of channel heads and enhanced evacuation of sediment from hillslopes (Reneau et al., 1986). This change in climate and sediment flux appears to have shifted the balance between the transport capacity and sediment supply of channels in a way that favors the creation of terraces (Gilbert, 1877; Fuller et al., 2009).

Arroyos occur throughout the Wildhorse Valley. This includes locations where incision into terrace surfaces, sometimes accompanied by lateral erosion, has lowered the channel or hillslope base level (Fig. 3.3). However, arroyos are also present in the upper parts of the catchment where the preservation of the Q_{tso} surface suggests that the base level of these hillslopes and channels has remained stable or steadily aggraded since the establishment of this surface. If arroyos represent the response of this landscape to a change in the rate of base level lowering, then the episode that triggered the formation of these features, and the arroyos themselves, must predate the Q_{tso} surface. Some arroyos are likely triggered or refreshed by variations in base-level lowering that excite erosion rates based on the direct connection between incised terraces and arroyo bounding breaks in slope (Fig. 3.3 B) and the ability of lateral and vertical incision to expose the bedrock that hosts these features (Fig. 3.3 C). Given that this is recorded by incision into the Q_{tso} surface, some arroyo incision must post-date the likely establishment of this surface $\sim 12\text{-}15$ ka (García and Mahan, 2009). Alternatively, perhaps arroyos arose due to changes in sediment transport accompanying a

change in climate that enhanced erosion on hilltops and in low-order channels. In this latter case arroyos may have formed synchronously with the establishment of the Q_{tso} surface. Indeed, in the nearby Pancho Rico Creek, García and Mahan (2009) recognized the rapid construction of terrace deposits at the Pleistocene-Holocene transition from debris flow deposits sourced from tributaries containing arroyos.

The convex forms of hillslopes can require 10^4 - 10^5 years to develop and reach equilibrium (Fernandes and Dietrich, 1997). In conglomerates, quasi-continuous, surface-parallel pedogenic carbonates like those observed in the Gabilan Mesa along south-facing slopes can also take 10^5 years to develop (Machette, 1985), although this process may be more rapid in the Gabilan Mesa where the fine-grained Pancho Rico Fm. presents a hydrologic barrier. The convex nature of hillslopes and the continuity of carbonate horizons (Fig. 3.2, 3.3 & 3.4) suggests that these slopes were relatively long-lived, stable landforms prior to being cross-cut by arroyos, suggesting that arroyos have retreated headward into soil mantled topography.

The cross-cutting nature of slope breaks separating arroyos from convex soil-mantled hillslopes and drainage divides (Fig. 3.3 A) and the presence of strongly concave topography at small AS values within these features (Fig. 3.7 & Table 3.1) are consistent with the headward expansion of bedrock-incising arroyos into formerly unchannelized, soil-mantled topography. Changes in process and hydrology accompany the transition from low relief, soil-mantled topography to steep topography floored by bedrock. In particular, the debris fans found beneath rills at the foot of talus slopes (Fig. 3.4) suggest that granular flows within arroyos may contribute to drainage development. The low concavity of upstream portions of arroyos (Fig. 3.7 & Table 3.1) are in line with expectations for regions thought to be dominated by these processes (Stock and Dietrich, 2003). Exposure of bedrock may also influence the frequency and magnitude of overland flow in a way that favors channel growth, but we do not have the hydrologic data necessary to evaluate the importance of this

effect. However, over landscape-forming timescales, drainage density (Horton, 1945) has been argued to increase as incision increases relative to hillslope transport (Perron et al., 2009). In light of this, the observation that arroyos expand drainage networks beyond what is observed in soil-mantled portions of the landscape suggests that channel incision is more efficient within arroyos. This is consistent with the observations of bedrock and soil physical properties, which indicate that the bedrock may be readily incised once exposed from beneath a cohesive soil mantle.

3.7.4 Conceptual model of arroyo development

We integrate our observations from laboratory measurements, field mapping, and landscape morphometrics to infer the sequence of processes responsible for the initiation, development, and refreshing of arroyo-related incision (Fig. 3.10). Laboratory experiments indicate that the bedrock of the Pancho Rico Fm. is likely easier to detach than the soils that typically overlie it, once exposed and fractured via slaking. If an acceleration in incision exposes the Pancho Rico Fm. slaking produces bedrock fragments that are readily detached (Fig. 3.10, t_1). The headward migration of the channel through highly detachable bedrock undermines the relatively strong soils and allows expansion of arroyos into higher areas of the landscape (Fig. 3.10, t_2). The rapid production of heavily fractured bedrock and the development of relief as the arroyo propagates may promote the emergence of granular flow processes that further shape arroyo evolution. Eventually (Fig. 3.10, t_3), infilling of arroyos may protect the vulnerable bedrock beneath and stabilize the downstream and cross-channel slopes of these features. Where the heads of arroyos are coincident with relatively resistant units (e.g. a bioclastic conglomerate bed as in Fig. 3.4 B) vertical knickpoints often exceed the scale of these beds and these strata often overhang underlying units. This suggests that some strata within the Pancho Rico Fm. may offer resistance to arroyo expansion. Encountering these variations in the stratigraphy could allow erosion

within the arroyo to be slowed, limiting further expansion and facilitating the infilling and burial of weak bedrock necessary to stabilize these features.

In this view, the event that initiates and refreshes these incisional features is the exposure of the bedrock to unconfined wetting and drying. Stripping soils to expose bedrock requires either high instantaneous erosion rates or sustained erosion rates in excess of surface soil-production rates. Our field measurements show that soils are significantly thicker on north-facing slopes than on those that face south (Fig. 3.6). If the thicker north-facing soils reflect more efficient soil production on north-facing slopes, the greater efficiency of soil production and the greater thickness of soil on these north-facing slopes may more effectively buffer bedrock from being exposed on these aspects. Within Wildhorse Valley, base level lowering represented by the Q_{tsy} surface has locally removed the Q_{tso} surface and truncated both north- and south-facing hillslopes that abut the channel (Fig. 3.3 C). A continuous soil mantle is maintained along north-facing slopes, while bedrock is exposed in the corresponding south-facing slopes. This supports the idea that north-facing slopes host soils that more effectively shield bedrock from exposure than south-facing slopes, assuming that incision affecting these slopes is equivalent in rate and magnitude. In some cases, the bedrock exposed along south-facing slopes appears to have initiated small arroyos along truncated noses. In other cases, where slope breaks initiating at the margins of Q_{tso} surfaces are continuous with arroyo margins, the incision represented by the Q_{tsy} surface may have rejuvenated larger, pre-existing arroyos. Arroyos are absent in corresponding north-facing watersheds, further substantiating the link between pedogenesis, the exposure of bedrock, and the initiation and rejuvenation of arroyos.

3.7.5 Implications for landscape Asymmetry

In the Gabilan Mesa, south flowing slopes are longer and more dissected than their north flowing counterparts. The production of asymmetric topography as a result of

aspect-controlled microclimates might be viewed as the result of one of two general mechanisms. In one model, referred to here as locally driven asymmetry, differences in the relative intensity of different erosion and transport processes operating at the hillslope scale produce the majority of the asymmetry seen in landscape morphology (Bass, 1929; Istanbuluoglu et al., 2008; McGuire et al., 2014; West et al., 2014; Yetemen et al., 2015). Alternatively, these initial contrasts in erosion may yield aspect-dependent discrepancies in the flux of sediment to trunk channels, favoring lateral migration of trunk channels in a particular direction and greatly exacerbating asymmetry (Melton, 1960). This latter mechanism was favored by previous work in the Gabilan Mesa (Dohrenwend, 1978); we refer to it here as non-locally driven asymmetry.

Tilting can also drive asymmetric drainage development (Cox, 1994), and the tilt of the Gabilan Mesa may have contributed to asymmetry in some locations in the Gabilan Mesa (Richardson, 2015). However, tilting does not explain the observed asymmetry in geomorphic process (Fig. 3.8), and much of the asymmetry observed in the Gabilan Mesa is commonly orthogonal or at high angles to the dip direction of the inclined mesa surface and underlying stratigraphic units (Dohrenwend, 1978). The cross-basin profiles assessed here are at a high angle to this dip (Fig. 3.9 B), so we assume that tilting has had a negligible effect on these profiles.

Arroyos appear to breach drainage divides (Fig. 3.3 A) and significantly extend valley topography (Fig. 3.7 & Table 3.1), which suggests that elongation of south-flowing watersheds may result from the headward expansion of the arroyo-hosting basins. However, our geomorphic mapping also highlights evidence of lateral migration, including the preservation of terrace surfaces on the north sides of valleys, the preferential development of south flowing fans, and the tendency of valley thalwegs to be located to the south side of valleys (see supporting material section 3.10.2 for details). Thus, the central question remains as to which mechanism dominates the

production of the watershed-scale topographic asymmetry observed in the Gabilan Mesa.

At the onset of incision into the Paso Robles surface, valleys could not have existed north of where the Paso Robles surface is still preserved today (Fig. 3.1 & 3.9). This allows us to constrain the maximum amount of southward migration of trunk channels and the minimum amount of northward growth of tributary channels, the relative magnitudes of which are similar (Fig. 3.9 A & C, median of 37% of the ridge to ridge distance for both measurements). However, it is important to emphasize that one is a minimum and the other a maximum, and our manual definition of the limit of the Paso Robles surface is subjective. What is striking about the profiles in Fig. 3.9 A is the symmetry in ridgeline slopes near the mainstem channel. Given that lateral migration was responsible for the lengthening of south flowing channels, one would expect to see south-facing ridgeline elevations that gradually descend from the divide to the basin axis, tracking the proposed vertical incision and horizontal migration of axial channels. Examples of these low-slope, south facing ridgelines are seen in the Gabilan Mesa, as is the case in the bold profile in Fig. 3.9 A, but they are not the majority in the profiles surveyed here. Instead, among the other profiles north and south facing ridgelines have similar slopes. Excluding the bold profile in Fig. 3.9 A, on north and south facing ridgelines the median normalized relief at $\frac{1}{4}$ of the basin width away from the mainstem channel is the same, 0.75. This is inconsistent with a gradual southward migration of the trunk channel throughout incision, and when combined with the observation that south-flowing tributaries often penetrate drainage divides and the mesa surface (Fig. 3.3 A, D & 3.9) suggests that northward expansion of south flowing tributaries, driven in part by the expansion of arroyos in south flowing drainages, is essential to establishing the observed asymmetry. Hence, in this part of the Gablian Mesa, it appears that local pedogenic processes may set in motion a series of events that produce a pervasive landscape-scale asymmetry.

3.7.6 The impact of surface armoring on landscape stability

The apparent instability of landscapes like those studied here, where small incisional perturbations that expose bedrock could further accelerate erosion, is an important concern for land managers evaluating changes in land use and for scientists attempting to understand the evolution of landscapes. While the specific slaking mechanism proposed here for removing the strength of the Pancho Rico Fm. is certainly not valid for all rocks, other weathering processes may act to disaggregate substrates and/or produce strong surfaces in a way that yields similar inversions of strength.

In southern California, Eppes et al. (2002) described the pedogenic armoring of surfaces. The ultimate incision into these surfaces produced amphitheater-headed channels, with morphologies similar to the arroyos observed here. However this process only occurred where local substrate favored pedogenic carbonate formation (Eppes et al., 2002). In Swaziland, rapid gully erosion on grazing land yields a concerning amount of soil loss (Mushala, 1997). These gullies incise deeply into otherwise unchannelized portions of hillslopes, accessing deep (up to 10 m thick) saprolite derived from gneissic and granitic bedrock. In this location, the saprolite has a 2 - 3 times lower shear strength than the soils that overlie it (Morgan and Mngomezulu, 2003). Similarly in Madagascar, features called lavakas incise through “hardened” laterite soils and into saprolite (Wells et al., 1991). The process of lavaka growth is thought to be promoted by the contrasts in strength between these materials, which favors incision into saprolite over laterite (Wells et al., 1991). These lavakas are also seen to cross-cut soil-mantled hillslope topography and in many cases breach drainage divides.

3.8 Conclusions

The Paso Robles Fm. overlies the Pliocene Pancho Rico Fm. and defines the shallowly dipping low-relief surface that caps much of the Gabilan Mesa (Galehouse, 1967; Durham, 1974; Dohrenwend, 1974). As a result of the tilt of this surface, the large rivers draining the Gabilan Mesa flow predominantly west-southwest, with smaller tributaries draining into them from the north and south. The south-facing aspects within Gabilan Mesa drainages contain long catchments with well-developed ridges and valleys, while north-facing aspects are a fraction of the length and are often planar and undissected (Fig. 3.1 & 3.9) (Dohrenwend, 1978). Within the Wildhorse Valley, we find that different geomorphic processes dominate north- and south-flowing tributaries. North-flowing valley networks are typically filled in by extensive colluvial hollows, while south-flowing networks are often occupied by steep sided, bedrock-incising arroyos that cross-cut lower relief, soil mantled topography and often extend upslope of typical hillslope-channel transitions (Fig. 3.3 & 3.7). Arroyos are even seen to penetrate north through drainage divides (Fig. Fig. 3.3). The Pancho Rico Fm. exposed within these arroyos pervasively fractures upon wetting and drying when unconfined, producing low-tensile-strength materials. The emergence of arroyos is commensurate with the headward expansion of drainage networks into hillslopes and transitions in geomorphic process. Arroyos are common throughout the Gabilan Mesa, and are restricted to south-flowing tributaries.

We propose that the expansion of arroyos into hillslopes results when incision accesses bedrock that is easier to erode than the soils that formerly mantled it. Soil thickness, like the presence of arroyos, is aspect dependent, with thicker soils residing on north-facing slopes. We suggest that variations in the efficiency of weathering and the thickness of soils may allow arroyos to dominantly form on south-facing slopes where soils are thin and weak bedrock more readily accessed. South flowing channels

are longer than their north flowing counterparts, a phenomena driven in part by the headward expansion of south flowing channels - a process we document occurring by the erosion of arroyos. If true, pedogenic processes may, under some circumstances, produce conditions that shape the morphology of entire drainage basins.

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3.10 Supporting material

3.10.1 Impact of arroyo expansion on drainage density

Geomorphic mapping within the Wildhorse Valley reveals what appears to be a high concentration of drainages within arroyos. Here we formally assess this observation and explore the implications.

Methods

We characterize the extent and locations of the drainage network manually (see supplemental geomorphic map). In larger channels, banks defining the bounds of channels were directly identified from shaded topography. This could be accomplished for some deeply incised headwater channels, but channel banks were not always identifiable in 1-m elevation data. We extended channel networks along regions where curvatures are strongly concave (e.g., $\sim 0.1 \text{ m}^{-1}$ or greater) along elongate bands no more than a few meters wide. Within each of the tributary basins (Fig. 3.3) defined by our automated procedure we measure drainage density (cumulative channel length divided by watershed area) and relate this to the fraction of those drainage basins occupied by arroyos (Fig. 3.11).

We evaluate the significance of correlations between measurements using the larger of two p -values determined from the non-parametric Spearmans rank test and with linear regressions, where we assess the significance of the slope of the correlation. We assigned significance when $p < 0.001$.

Results and Discussions

The drainage density measured in the Wildhorse Valley is 4.2 km^{-1} , and is distributed disproportionately among north- and south-flowing tributary basins. Within north-flowing tributaries drainage density varies from 0 to 7.3 km^{-1} , with a median value of

2.2 km^{-1} . This distribution is significantly different ($p_{max} = 1 \times 10^{-10}$) than that of south-flowing tributary basins, whose drainage density ranges from 0 to 13.4 km^{-1} , with a median of 5.5 km^{-1} (Fig. 3.11 A). Across all drainage basins, drainage density increases with the fraction of these basins occupied by arroyos (Fig. 3.11 B). The best-fitting linear regression to drainage density data as a function of the fraction of a basin occupied by arroyos gives an intercept of 2.5 ± 0.5 and a slope of 19.5 ± 3.2 ($p = 8 \times 10^{-21}$ and 3×10^{-22} , respectively).

The total drainage density of a drainage basin containing arroyos and soil mantled regions can be cast as a sum of the contributors to drainage density:

$$D_t = \frac{L_s + L_b + L_a}{A_c} \quad (3.2)$$

Here $D \text{ [L}^{-1}\text{]}$ is drainage density and the subscript t denotes the total drainage density of a basin. This is given by the sum of the lengths of soil mantled channels, $L_s \text{ [L]}$; the lengths of bifurcations within arroyos, $L_b \text{ [L]}$; and the length of the longest channel within the arroyo, $L_a \text{ [L]}$, divided by the catchment area, $A_c \text{ [L}^2\text{]}$. By approximating the area of an arroyo as a rectangle of length L_a and width $w \text{ [L]}$, and converting the ratio of component lengths to the drainage density within those components by accounting for the fraction of the basin occupied by arroyos, $f \text{ []}$, the respective drainage densities and arroyo geometry are related as:

$$D_t = D_s + (D_b + \frac{1}{w} - D_s)f. \quad (3.3)$$

Where the subscripts s and b denote drainage densities in soil-mantled terrain and due to bifurcations within arroyos, respectively. The drainage density due to arroyo bifurcations is the difference between the total drainage density within arroyos and the drainage density of arroyos from the longest channel, $1/w$. We measure D_s as the drainage density in basins without arroyos where channels were mapped (i.e., $f =$

0), which have a mean drainage density of 2.8 km^{-1} (standard deviation, 1.5 km^{-1}). Within arroyos, which are typically about 75 m wide, we measure a mean drainage density of 28.9 km^{-1} (standard deviation 14.7 km^{-1}). In the context of Equation 3.3 this gives a D_b of $\sim 15 \text{ km}^{-1}$. The similarity between D_b and $1/w$ suggests that the addition of channels within the bedrock and scree walls of arroyos effectively doubles the drainage density within these features. In the framework of Equation 3.3 the best fitting linear regression to D_t data as a function of f (Fig. 3.3) yields a D_s of $2.5 \pm 0.5 \text{ km}^{-1}$ and a value of $22.0 \pm 3.2 \text{ km}^{-1}$ for the sum of D_b and $1/w$, within the range of observations for these parameters obtained above. Therefore, increases in drainage density above that observed in soil mantled catchments (Fig. 3.3) should reflect a combination of the lengthening of mainstem channels by headward expansion of arroyos and the addition of channel segments within arroyos.

3.10.2 Evidence of lateral channel migration in the Wild-horse Valley

Past work in the Gabilan Mesa suggested that relatively high sediment loads were delivered to mainstem valleys from south flowing tributaries, and preferentially deflected streams to the south (Dohrenwend, 1978). An effect that over time was believed to give rise to the observed asymmetry in the size of north- and south-flowing drainages. Using our geomorphic mapping and topographic analysis we highlight evidence in support of the basic tenants of this lateral migration hypothesis, although these findings do not predict how much of the observed asymmetry has resulted from lateral migration.

Methods

We measured the density of fans within mainstem valleys and determined the mean downslope direction of these features to estimate relative differences in the flux of sediment delivered to mainstem valleys from north and south flowing tributaries. Fan density is calculated as the ratio of the fan areas within a search window to the area of all terrace surfaces and fans within that search window. We calculated independent fan densities for fans whose orientations place them in different aspect classes. Individual fans were assigned an aspect based on the orientation of the slope vector determined from the coefficients of the best-fitting plane computed from the 1-m elevation data contained within fan boundaries. We defined 26 search windows by first constructing lines that traverse mainstem valleys where a single valley floor deposit is present and extensive (see supplemental material). We segmented these lines into portions that are slightly longer than one valley width (300 m), and buffered them to create rectangular search windows 600 m wide.

We collected two sets of swath profiles across valley surfaces in ALSM data using a swath radius of 2 m to highlight how variations in sediment flux have influenced valley morphology. In the first set, cross valley profiles, locations of swath profiles across valley surfaces were selected to be roughly perpendicular to the trend of valleys, to have orientations within 45° of N-S, and to traverse valleys at locations away from tributary mouths (see supplemental material). Additionally, we trimmed this first set of profiles to show 20 m of relief. In the second set of swath profiles, fan-traversing profiles, we targeted locations of confluences between tributaries and mainstem valleys. These profiles were oriented approximately perpendicular to the mainstem valleys and were not restricted to any particular orientation. Profiles originated or were terminated either at the limit of mapped valley-fill surfaces or at the approximate location of the projection of this limit to the midpoint of tributary mouths (see supplemental material).

Using fan-traversing profiles we compared the asymmetry in fan length on either side of valleys to the valley floor asymmetry. Fan length asymmetry is measured as the difference between the length of fans derived from the left of the profile (looking upstream), $L_{f(l)}$, and the right of the profile, $L_{f(r)}$, divided by the total length of the profile, L_t . We measured valley asymmetry using the ratio of the distance to the lowest elevation point from the left end of the valley profile, L_{thal} , to L_t . We subtracted $\frac{1}{2}$ from the valley asymmetry and divided the fan length asymmetry by 2 so that both quantities may vary from $-\frac{1}{2}$ to $\frac{1}{2}$, such that the axial location of a channel within a symmetric valley would be located at 0.

We evaluate the significance of correlations between measurements using the larger of two p-values determined from the non-parametric Spearmans rank test and with linear regressions, where we assess the significance of the slope of the correlation. We assigned significance when $p < 0.001$.

Results and Discussions

The density of fans in any aspect class varies between 0.09 and 1, with a median value of 0.6. North-flowing fans make up 0-60% of valley floor deposits (median: 2%), while south-flowing fans occupy between 9% and 100% of valley floor deposits (median: 47%). The distributions of these fan densities are distinct from one another ($p_{max} = 9 \times 10^{-8}$, Fig. 3.12 B). Qualitatively, cross-valley profiles highlight that main-stem channels are typically offset toward the south side of valleys (Fig. 3.12 A). Mainstem channels often abut north-facing hillslopes, but are separated from the toes of south-facing hillslopes by a steady, south-directed slope. Asymmetry in the position of the thalweg within mainstem valleys is positively correlated with asymmetry in the length of fans sourced from either side of these valleys ($p_{max} = 3.8 \times 10^{-6}$). The predominance of south flowing fans and the bias of channels toward the south sides of mainstem valleys are recorded by the clustering of data within the

upper right quadrant of Fig. 3.12 C.

Near the upstream limit of incision by the Q_{tsy} surface, the Q_{tso} surface is a fill terrace, and so the fans sourced from tributary catchments probably facilitated a component of aggradation and valley filling (e.g., García and Mahan, 2009). However, without differential subsidence across the valley or preferential fan erosion, the areal extent of these fans provides constraints on the relative sediment delivery from north- versus south-flowing tributaries. South-flowing fans occupy a larger fraction of mainstem valleys than north-flowing fans (Fig. 3.12). Differences in the extent of fans suggest that more sediment is being delivered from south-facing aspects, consistent with the fact that south-flowing drainages are generally larger and may be expanding headward. While nearly all south-flowing tributaries have fans at the intersection of mainstem valleys, the same is not true for north flowing tributaries (Fig. 3.3). Either north-flowing fans are preferentially removed from these valleys, the events that construct fans are less frequent in north-flowing tributaries, or a combination of these mechanisms is at play. Along these lines, the much larger colluvial hollows present in north-flowing drainage basins indicate that these catchments may not flush the sediment trapped within hollows with the same frequency as south flowing catchments (Fig. 3.8). At timescales shorter than that of hollow excavation and filling, which may be related to cyclicity in climate (Reneau et al., 1986), this may also lead to a reduced sediment flux from north flowing drainages.

The Q_{tso} surface shows a southward cross-valley slope. Asymmetry in the position of the channel within this surface correlates with asymmetry in the extent of fans sourced from either side of mainstem valleys. Thus, it appears that the aspect-controlled asymmetry in sediment supply likely impacts the slope of the major valleys as well (Fig. 3.12). This may lead to the cross-valley migration of the major channels during accumulation of the Q_{tso} surface, which might have the effect of truncating north-facing hillslopes (Dohrenwend, 1978). Near the outlet of the Wildhorse Valley

the extensive and exclusive preservation of the Q_{tso} surface along the north side of the modern channel is consistent with the southward migration of channels in these large alluviated valleys.

3.11 Figures and tables

	AS_{crit}	A_t	$-\theta_l$	$-\theta_r$	k_r [m ^{2θ}]
Arroyos	41	6,817	0.15	0.39	14.3
Colluvial hollows	12,974	8,745	0.09	0.22	2.1
North flowing basins*	971	3,951	-0.08	0.27	3.7
South flowing basins*	254	14,412	-0.03	0.39	11.8
North flowing basins	603	3,995	-0.06	0.26	3.2
South flowing basins	112	6,781	-0.08	0.36	10.1

Table 3.1: Summary of morphometrics calculated from longitudinal profiles extracted from tributary basins in the Wildhorse Valley. Entries for north- and south-flowing basins marked with an asterisk were calculated without data from mapped colluvial hollows and arroyos. AS_{crit} is the AS value separating flat and sloping line segments fit to curvature, AS data. A_t is the area that separates two power law fits describing slope as a function of area, θ is the concavity of these power law fits and k the steepness. The subscripts l and r denote the regions where area is less than and greater than A_t , respectively

Figure 3.1: Hillshade of the northern Gabilan Mesa highlighting the study region. Inset map of California shows the location of the region depicted. We focus our analysis on Wildhorse Valley, and supplement with observations from neighboring regions. Although the majority of the Wildhorse Valley is hosted within the Pancho Rico Fm., the Paso Robles Fm. is preserved along many of the northern and southern drainage divides of the basin. Red dots show locations of arroyos mapped using aerial photographs, in the top right corner of the map thick vegetation prevented the mapping of arroyos in these regions. Orientation of mesa surface dip direction determined from the coefficients of the plane that best fits the elevation data within the mapped extent of that surface. Stars highlight generalized locations of soil thickness measurements, each star represents multiple measurements. Thin dashed lines crossing drainage basins were used for Fig. 3.9. Mapping of the Paso Robles Fm provided by (Ludington et al., 2005), 10 m resolution topographic data provided by the USGS National Elevation Database.

Figure 3.2: Photographs highlighting arroyos (A), colluvial hollows (B) and the asymmetric distribution of these features within the landscape (C). Photos A and B are looking approximately north and south, respectively. Photograph C is looking to the west, and is provided courtesy of Juan C. Fernandez Diaz.

Figure 3.3: Simplified geomorphic map of the Wildhorse Valley. See supplemental for the full geomorphic map. Extent is shown in Fig. 3.1. Colored polygons denote different geomorphic process domains. Small rectangles indicate the location of detailed maps of arroyos shown in Fig. 3.4. Basins bound by gray dashed lines were used to calculate basin specific calculations (e.g. Fig. 3.8). Small arrows highlight features of interest discussed in text, solid boxes and insets A and B highlight relationships at 2x scale. Arrows labelled A point to regions where arroyos appear to have crossed former drainage divides, B shows a location where the incised margins of a terrace correlate with the boundaries of an arroyo, C points to locations where incision patterns into terrace surface suggests some lateral erosion of slopes with opposing aspects, and D highlights regions outside the mapping area where south flowing channels extend a significant distance into the preserved mesa surface. Stars are adjacent to those soil pits sampled for grain size analysis.

Figure 3.4: Detailed geomorphic maps of arroyos. Location of each map is shown in Fig. 3.3. Linear ‘carbonate’ features denote exposures of pedogenic carbonate at the soil bedrock interface exposed in arroyo walls, description of these layers is given in main text as is the description of different scree layers. Unmapped regions within the bounding slope break of arroyos were covered by deep accumulations of tumbleweeds.

Figure 3.5: Variations in particle size distributions for north-facing (top panel) and south-facing (bottom panel) soil pits. Shaded regions represent the contributions from different grain size fractions. Bold black vertical lines represent the sampling intervals for the prominent horizons identified. In some situations transitions between horizons were gradual and the bounds of sampling intervals did not overlap, these are indicated with a diagonal boundary to shaded regions that has no bold outline.

Figure 3.6: Box plots of soil thickness on north-facing, south-facing, and other slopes. Numbers inside boxplots indicate the number of points that make up that distribution. Black boxes correspond to data corrected for slope and assigned direction on ALSM data, light grey boxes use 10 m data. Inset rose diagram correlates slope-facing direction and soil thickness(given by the distance from the center) of soil thicknesses based on the 10m data. Gray crosses denote outliers.

Figure 3.7: Morphometrics as a function of aspect and different geomorphic domains. A & B show the along channel slope as a function of drainage area, dashed lines in these plots are piecewise power-law regressions (see text). C & D show curvature as a function of the product of drainage area and slope, AS . Vertical dashed lines in C & D indicate the location of AS_{crit} , the point where curvatures transition from being roughly constant to steadily increasing. Means are plotted in logarithmic bins that contained at least 10 data points. A & C include means for data including and excluding regions mapped as either colluvial hollows or arroyos, regressions and AS_{crit} calculations exclude these regions. Parameters summarizing the relationships observed in these profiles are given in Table 3.1

Figure 3.8: Boxplots highlighting the percentage of tributary area occupied by arroyos (A) and colluvial hollows (B) within north-flowing, south flowing, and other catchments. Extent of catchments is shown in Fig. 3.3. Inset rose diagrams shows the full distribution of data. In the colluvial hollows plot, gray box plots include regions of hollows where classification was uncertain (Fig. 3.3).

Figure 3.9: Ridge and Valley profiles from the locations shown in Fig. 3.1 highlighting the penetration of channels into the low-relief Paso Robles surface are plotted in A. The bold pair of ridge and valley profiles are from one of the largest valleys within the Gabilan Mesa, while asymmetry appears extreme here it is of a similar relative magnitude to the smaller valleys also shown here. Small circles and squares denote the heads of channels and the limit of the low-relief mesa surface, the difference between the location of these points for each profile provides a minimum estimate of headward channel expansion. (B) Rose diagram of asymmetry as a function of profile orientation, with asymmetry calculated as the channel head location in A minus 0.5. Black arrow shows the dip direction of the mesa surface. (C) Histograms of the minimum amount of headward channel expansion and the maximum amount of lateral migration estimated from the profiles in A.

Figure 3.10: Conceptual model of progressive arroyo development. Initially, t_0 , soil mantles underlying bedrock. Slopes are dictated by the transport of soil and pedogenesis has resulted in the accumulation of carbonate horizons (dashed lines) at the soil-bedrock contact. Erosion that causes the exposure of Pancho Rico bedrock at some time, t_1 , causes the rapid fracturing of that rock to readily detachable material. Heavily fractured material is easily detached and removed from that arroyo at time t_2 . Rapid erosion during this period causes the development of relief and the emergence of new transport and erosional processes. This erosion in turn exposes new bedrock that rapidly fractures and is detached. Eventually, material progressively infills the arroyo at time t_3 , protecting bedrock from further rapid disintegration.

Figure 3.11: Drainage density within the Wildhorse Valley. (A) Boxplots of drainage density within north-flowing, south-flowing, and other tributaries. Extent of catchments is shown in Fig. 3.3. Dashed line shows the drainage density measured on the whole of the Wildhorse Valley. (B) Relationship between the fraction of a basin occupied by arroyos, f (Fig. 3.8), and the drainage density of that basin. Dashed lines show the uncertainty associated with the regression coefficients.

Figure 3.12: (A) Cross valley profiles are smoothed swath profiles, the locations of profiles are shown in Fig. 3.3. (B) Box plots of fan density in different aspect classes in the upper portions of the Wildhorse Valley (see text for details). Dashed line shows the median density of fans in any aspect class. (C) Plots of the asymmetry in the thalweg position (e.g. the 0 position in A) against asymmetry in the length of fans draining from the left and right side ($L_{f(l)}$ and $L_{f(r)}$, respectively) of valleys (facing upstream).

Chapter 4

North-south topographic asymmetry reflects elongation of low-order tributaries in semi-arid landscapes underlain by weak rocks

4.1 Abstract

Differences in the average morphologies of pole- and equator-facing slopes are observed in a variety of landscapes. These morphologic differences are generally attributed to relative variations in the intensity of different mass transport processes on slopes receiving different magnitudes of solar radiation. We demonstrate that this asymmetry primarily arises from the preferential formation of low-order valley networks with a particular mean flow direction. This asymmetry is more severe in rocks expected to offer little resistance to erosion than in more resistant rocks when controlling for climate. This suggests that aspect-driven differences in surface processes that drive

differences in landscape evolution are also sensitive to local rock type.

4.2 Introduction

Climatic differences that arise due to aspect-dependent insolation can influence geomorphic process and, in turn, landscape forms (Bass, 1929; Emery, 1947; Melton, 1960; Dohrenwend, 1978; Burnett et al., 2008; Istanbuluoglu et al., 2008; West et al., 2014; McGuire et al., 2014). This ‘micro-climatic’ asymmetry is most commonly manifest in the northern hemisphere as north-facing slopes that are steeper than their south facing counterparts (Poulos et al., 2012). Opposing aspects also commonly display differences in the degree of drainage dissection (Bass, 1929; Dohrenwend, 1978; Istanbuluoglu et al., 2008; McGuire et al., 2014), which likely measures the voracity of hillslope transport relative to channel incision (Perron et al., 2009). This may directly reflect aspect-dependent variations in surface hydrology or indirectly result from the differences in vegetation that arise from this hydrologic configuration (Bass, 1929; Emery, 1947; Melton, 1960; Dohrenwend, 1978; Istanbuluoglu et al., 2008; McGuire et al., 2014). Models that drive landscape evolution with processes whose intensity varies with aspect reproduce both local (McGuire et al., 2014) and continental measurements of topographic asymmetry (Yetemen et al., 2015).

Early work in the central California Coast Ranges suggested that asymmetry in the cross-sectional shape of valleys was most pronounced in landscapes underlain by rock types expected to yield little resistance to erosion (Dohrenwend, 1978). We test the idea that asymmetry is more pronounced in weak rocks by comparing the asymmetry in geographically proximal watersheds that expose rock types of different erodibility.

Within a catchment, satellite-based estimates of vegetation productivity approximate the response of biota to microclimates. We use these satellite-based metrics

to define the range of climates where vegetation is less productive on south-facing slopes and therefore where microclimates may be influencing surface processes (Istanbulluoglu et al., 2008; Gutiérrez-Jurado et al., 2013). We identify four sites in this range of climates and compare topographic asymmetry in neighboring basins developed in contrasting rock types. Topographic profiles reveal that valley asymmetry is produced by both steeper N-facing hillslopes and the presence of low-order channels preferentially located along the northern margin of E-W trending valleys. We compare the lengths and orientations of observed flow paths to determine whether or not they might plausibly represent a random sampling of these values. We demonstrate that asymmetric valleys host flow paths that are preferentially oriented toward the south at length scales most likely associated with low-order valleys. Measurable topographic asymmetry is present to some degree in all four sites, but a strong preference for south-directed tributaries is only seen in sites hosting weak rocks. These results support the idea that development of strongly asymmetric topography is sensitive to local rock type (Dohrenwend, 1978).

4.3 Methods

4.3.1 Measuring microclimates

Variation in the distribution and density of flora on slopes of opposing aspects is common in arid and sub-humid landscapes (e.g., Istanbulluoglu et al., 2008). We measure asymmetry in vegetation in 12 regions that span the range of climates observed in the contiguous United States (Figure 4.1, basin locations provided in supplemental material). Here climate is defined by the aridity index, AI (ratio of precipitation to potential evapotranspiration; Zomer et al., 2008). In some regions, the Normalized Difference Vegetation Index, $NDVI$, varies as a function of aspect and past work

recognized that these variations track the response of vegetation to microclimates (McGuire et al., 2014). *NDVI* attempts to capture the photosynthetic productivity of vegetation by measuring the normalized difference in the intensity of the near infrared (*NIR*) and visible red (*VIS*) wavelengths:

$$NDVI = \frac{NIR - VIS}{NIR + VIS}. \quad (4.1)$$

We calculate the mean *NDVI* on north- and south- facing slopes within drainage basins, $\overline{NDVI_n}$ and $\overline{NDVI_s}$, and quantify asymmetry in this quantity with a normalized difference:

$$NDVI_{asym} = \frac{\overline{NDVI_n} - \overline{NDVI_s}}{\overline{NDVI_n} + \overline{NDVI_s}}, \quad (4.2)$$

such that positive values indicate greater *NDVI* on north-facing slopes (Fig. 4.2). We use catalogs of the 32-day average *NDVI* (provided by Google Earth Engine) to calculate $NDVI_{asym}$ every month from January 1st, 2000, to January 1st, 2016 (Fig. 4.2 A). By looking at both the sum of $NDVI_{asym}$ over the course of an average year, $\sum NDVI_{asym}$ (Fig. 4.2 B), and the day of the year and value where its magnitude is greatest, $\max(NDVI_{asym})$ (Fig. 4.2 C), we summarize the severity of differences in *NDVI* with aspect.

4.3.2 Topographic analysis

We measure topographic asymmetry at four sites where modern *AI* is in the range observed to favor more productive vegetation on north-facing slopes. Sites were selected where local geologic conditions produce a north-south topographic grain that favors the development of east- or west-trending valleys because this provides the easiest condition for detecting differences in the morphology of north- and south-facing topography. We focus on sites where nearby drainage basins are developed in rock

types expected to yield different degrees of resistance to erosion. Specifically, we limit our efforts to sedimentary rocks no older than the Pliocene, which we refer to as ‘weak’, as well as igneous and metamorphic rocks, which we term ‘strong’ (Table 4.1). Past efforts that have sought to quantify the physical properties of materials that govern their resistance to erosion support this general classification (Sklar and Dietrich, 2001; Bursztyn et al., 2015).

We use 10-m resolution digital elevation models (DEMs) to measure topographic attributes at our four sites. We fill internal sinks in DEMs to ensure continuous downstream flow paths and use this flow-corrected topography to identify 4-10 drainage basins within each rock type. We calculate the mean AI and proxies for microclimate within each polygon defined by the intersection between the outline of drainage basins and the mapped extent of selected geologic units. We refer to the union of these shapes as geology-basin (GB) polygons (locations of GB polygons provided in supplemental material).

The north and east components of slope were calculated using a second-order finite difference kernel, and the slope magnitude, S , and direction of slope, θ_s , were determined from these components. Where θ_s is within 45° of north or south we classified pixels as north- or south-facing, respectively. We further exclude points where local relief is less than 4 m in a 25-m radius search window to remove largely depositional areas of the landscape from consideration.

We use two similar metrics calculated within GB polygons to quantify topographic asymmetry. First, we collected up to 10 topographic profiles from each GB polygon, allowing their lengths to vary, while ensuring that they span the distance between opposing ridge lines and are oriented within 45° of north-south (locations of profiles provided in supplemental material). We use these profiles to compute the ratio of the average along-profile slope on north- and south-facing slopes (profile slope ratio; (Emery, 1947; Melton, 1960; Dohrenwend, 1978)). Second, we calculate

the ratio of the average slope of gridded data measured on north- and south-facing aspects (Poulos et al., 2012), which we refer to as the average slope ratio.

Identifying asymmetry in drainage network morphology

While hillslope asymmetry is widely observed (Poulos et al., 2012), past work has also noted systematic differences in the development of north- and south-flowing tributaries (Dohrenwend, 1978). We quantify differences in this drainage structure by tracking the maximum along-flow distance (L_{max} ; [L]) from each pixel to the corresponding drainage divide, as well as the mean direction of these flow paths (θ_f). Next, we determine the L_{max} value at which slopes transition from convex-up to concave-up geometries by fitting a two piece power-law function to the local slope in the down flow direction as a function of L_{max} . We refer to the value of L_{max} where these two power-law functions join as L_{hs} . This approach is similar to past efforts that have sought to quantify hillslope length, with the exception of using a direct measure of length rather than a transformation based on drainage area (e.g., Grieve et al., 2016). We normalize along-flow length to our estimate of hillslope lengths, $L^* = L_{max}/L_{hs}$ (Fig. 4.3 A), to compare sites with varying drainage densities.

We treat each pixel as a vector of length L^* and orientation θ_f and use these vectors to measure distributions of flow directions within drainage basins. For each basin in a study area, we first isolated L^* and θ_f values for each lithologic class. θ_f and L^* were then divided into $\pi/90$ rad increments and 25 length bins (from $3 < L^* < 50$ with logarithmic spacing), respectively (e.g., Fig. 4.4). For each L^* , θ_f bin, we compute the number of pixels of that L^* range within $\pi/8$ rads of that orientation and divide this by the total number of points within the L^* bin (^{rel}N ; the size of the θ_f smoothing window was chosen to reduce the impact of D_8 routing at short length scales). Finally, we aggregated ^{rel}N for each site using its mean value across all studied basins for each lithologic class. We also aggregated across basins

by weighting for their area (not shown), and found that such an approach yielded broadly equivalent results.

Quantification of network structure asymmetry requires estimation of the distribution of ^{rel}N expected in symmetric landscapes with identical watershed boundaries. We estimate these distributions using randomly seeded, synthetic topography (Fig. 4.3 B). First, we identified the boundaries and outlet of each basin. A space-filling network was simulated within these boundaries using a modified version of the priority flood algorithm (Barnes et al., 2014) in which filling priority was randomized, starting at the current outlet location and elevation. This procedure produces network geometries that are inconsistent with the fact that elevations must be equivalent on each side of a topographic divide and within the junction of a channel to prevent drainage divide migration (Shelef and Hilley, 2014). Thus, we migrate divides by reassigning their flow directions with an iterative procedure that ensures that the watershed areas and channel slopes globally obey the area-slope scaling required by the stream-power incision rule (e.g., Howard, 1990). The steepness used to produce the elevations of random basins is determined based on the relief and integrated drainage area (e.g., Perron and Royden, 2013) of the longest flow path in the observed basin and we use a concavity of 0.5 (Hack, 1957). Experiments with real data suggest that down-sampling from 10 m to 30 m resolution doesn't change the structure of ^{rel}N distributions, so we create our synthetic examples at this coarser resolution. Simulations of synthetic basins reveal that 10 realizations of each synthetic basin captures the variability in the distribution of simulated networks produced based on this procedure.

We used the value of ^{rel}N for the resulting simulations to infer the properties of drainage networks that lack the asymmetry observed within each studied watershed. Accordingly, we normalize observed distributions, $^{rel}N_{obs}$, by distributions computed from synthetic, simulated topography, $^{rel}N_{synth}$ as $N^* = \overline{^{rel}N_{obs}} / \overline{^{rel}N_{synth}}$. $\overline{^{rel}N}$ refers

to the mean ^{rel}N value computed from all GB polygons within each region and rock type or the mean of a series of simulated basins.

4.3.3 Assessing significance

We assess the significance of the N^* values observed at each L^*, θ_f bin by comparing the distributions of $^{rel}N_{obs}$ and $^{rel}N_{synth}$ within each bin. At each site, distributions of $^{rel}N_{obs}$ were created by treating each basin (whether real or a synthetic realization) as a sample (number of basins and their properties reported in Table 4.2). Examples of a real basin and a single random realization, as well as the resulting N^{rel} distribution are shown in Figures 4.3 & 4.4. We test the null hypothesis that observed and simulated ^{rel}N samples are drawn from the same distribution using the Wilcoxon rank-sum test (we reject the null hypothesis when $p < 0.05$; rejection of the null hypothesis requires the median of $^{rel}N_{obs} > ^{rel}N_{synth}$). To assess the significance of correlations between two variables, we record the larger of the two p -values determined from a linear regression and the non-parametric Spearman's rank test. We assign significance when $p < 0.05$.

4.4 Results

4.4.1 Modern microclimates

$NDVI_{asym}$ shows a significant correlation ($p = 4 \times 10^{-11}$) with AI , whether measuring this as $\sum NDVI_{asym}$ (Fig. 4.2 B) or $\max(NDVI_{asym})$ (not depicted). $\sum NDVI_{asym}$ transitions from positive to negative values around $AI = 0.5$. The greatest difference in $NDVI$ with aspect occurs in the middle of the year when $AI < 0.5$. This peak typically occurs either late or early in the year in regions where $AI > 0.5$. The mean slope of the GB polygon is also a significant predictor of $\max(NDVI_{asym})$. This

correlation shows opposing trends for data with opposite senses of $NDVI_{asym}$ (Fig. 4.2 C).

4.4.2 Topographic asymmetry

We identified four sites where $AI < 0.5$ (Table 4.1, Fig. 4.1) in both weak and strong rocks, excepting one site (Corning) where the average $AI = 0.76$ in strong rocks. The average ratio of slopes measured along topographic profiles varied from ~ 1 to ~ 2.6 (Fig. 4.5), while averaged north- and south-facing mean slopes varied between ~ 1 and ~ 1.6 .

Distributions of N^* as a function of θ_f where $N^* > 1$ (Figure 4.6) typically show large values of N^* in a narrow range of θ_f at $L^* > 25$. Large values of ^{rel}N oriented toward the outlet are present in both real and randomized grids because of the preferential elongation of high-order network segments in the direction of the axis watersheds (Fig. 4.3, Fig. 4.4). However, the variation in simulated values and the relatively small number of points occupying $L^* > 25$ yield N^* values larger than one with relatively small p -values (Fig. 4.6).

In some sites (e.g., Hualapai and Lower Kern), high N^* with low- p values at a given θ_f can persist across length scales (e.g., $5 < L^* < 25$; Fig. 4.6). At intermediate length scales ($L^* \approx 10$) in weak rocks, high N^* with low- p values are systematically oriented to the south, and to a lesser degree, the north (Fig. 4.6). Channels producing these N^* values are oriented orthogonal to the basin axis (selected to be E-W). In contrast, such N-S-oriented, large, significant values of N^* are absent at identical sites underlain by strong rocks (Fig. 4.6).

4.5 Discussion

4.5.1 Vegetation asymmetry and climate

North-facing slopes tend to have greater $NDVI$ where $AI < 0.5$. While the magnitude of $NDVI_{asym}$ scales with AI , the time of year at which these differences are most pronounced varies among climates (Fig. 4.2). North-facing slopes tend to have higher mean $NDVI$ values in dry climates, with the greatest normalized difference between opposing slopes occurring in the hot, summer months that receive the greatest incoming solar radiation (Yetemen et al., 2015). We interpret this as a water limitation, such that wetter north-facing slopes can support healthier vegetation when $AI < 0.5$. In contrast, when water is not limiting, the greatest difference between aspects may be the intensity of light, and the north-facing slopes in our northern hemisphere sites may be inhibited by this when incoming solar radiation is near its yearly minimum (Yetemen et al., 2015). The magnitude of differences in $NDVI_{asym}$ scales with mean slope (Fig. 4.2), consistent with the fact that increases in mean slope should drive greater variations in the incoming solar radiation received by north- and south-facing slopes as the incident angle between equator-facing slopes and the sun becomes less oblique.

4.5.2 Asymmetry in topography

Some degree of topographic asymmetry is observed globally (Poulos et al., 2012); for example, the mean slope of pole-facing slopes is typically greater than that of equator-facing slopes (Poulos et al., 2012), akin to what we observe (Fig. 4.5). However, this simple measure of asymmetry may underestimate the magnitude of asymmetry present in landscapes. Our data show that the average slope ratio within a particular

GB polygon underestimates the asymmetry in valley morphology seen in the topographic profiles taken from the most asymmetric sites (Fig. 4.5). The discrepancy between these two metrics is partly a reflection of the different scales over which these measurements are made. Pixel-to-pixel slopes in gridded data measure gradients between adjacent pixels, the majority of which are from hillslopes. Hence the average slope on north- and south-facing aspects will largely reflect these landscape elements (e.g., Poulos et al., 2012). In contrast, our cross-valley profiles are $10^2 - 10^3$ m long, at least an order of magnitude longer than typical hillslopes (Grieve et al., 2016). At the length-scales of our cross-valley profiles ($L^* > 5$) constructed within landscapes underlain by weak rocks, we document anomalously high N^* along S-, and to a lesser degree, N-directed orientations. These anomalous areas of the landscape correspond to the low-order portions of the channel network rather than along hillslopes, which may explain the discrepancy between slope-based measures of asymmetry and those measured in cross-valley profiles (Fig. 4.5). In particular, the greatest N-S asymmetry in N^* is seen in the Gabilans and Lower Kern sites in weak rocks (Fig. 4.6), and these sites show the greatest discrepancy between the profile slope and average slope ratios (Fig. 4.5).

Interestingly, previous studies have documented aspect-related changes in drainage density in asymmetric landscapes (Bass, 1929; Dohrenwend, 1978; Istanbuluoglu et al., 2008; McGuire et al., 2014), as well as the preferential elongation of S-oriented, low-order channels (Johnstone et al., *in review*). Simulations of bedrock-channel incision into soil-mantled hillslopes suggest that second-order valleys emerge when flow-lengths are approximately 3 – 30 times the hillslope length (Perron et al., 2009, 2012) (this ratio is equivalent to our L^* measurement). We observe that S-oriented channel segments in which $L^* \leq 25$ occur at substantially greater frequencies in weak lithologies than would be expected based on our synthetic fluvial landscapes. This indicates that preferential expansion of the drainage network and elongation of these

flow orientations occurs within low-order portions of drainage networks. This directional preference does not occur in landscapes underlain by strong lithologies, nor does it manifest when $L^* > 25$. At these long length scales flow paths are more commonly guided toward the outlet of the drainage basin, which is consistent with predictions of the realized synthetic networks.

The preferential elongation of low-order, south-flowing channels can be accomplished by the southward lateral migration of trunk channels or the northward expansion of south-flowing channel heads. Lateral migration is likely an important contributor to asymmetry in some settings (e.g., Melton, 1960), and may be favored in rocks with little resistance to erosion (Dohrenwend, 1978). Lateral erosion is thought to reflect asymmetry in sediment delivery from tributary valleys, which favors south-directed avulsions. Preferential preservation of valley fill deposits and strath terraces on the north sides of some valleys in the Gabilan Mesa (Garcia, 2006) indicate that lateral migration is certainly present in this area and contributes to topographic asymmetry (Johnstone et al., *in review*), despite the fact that an enhancement in erosion rates due to undercutting is rarely resolved (Richardson, 2015).

Despite the appeal of this mechanism, field observations from some asymmetric landscapes suggest that it cannot be exclusively responsible for the observed topographic asymmetry. In weak rocks of the Gabilan Mesa, south-flowing channel heads appear to penetrate northward and cross-cut former drainage divides (Johnstone et al., *in review*). These features are also present in basins underlain by weak rocks of the Lower Kern site (Fig. 4.7). This suggests that northward expansion of south-flowing, low-order channels may be an important component of generating asymmetric landscapes.

Drainage density is set by the length of the hillslope - channel transition (Horton, 1945; Montgomery and Dietrich, 1988), which is in turn controlled by the strength of hillslope transport relative to channel incision (Perron et al., 2009). Many of the

local mechanisms for developing landscape asymmetry could cause hillslope transport rates to vary with the intensity of incoming solar radiation. Aspect-driven variations in frost heave impacts hillslope soil transport in cool climates (West et al., 2014). Heavily vegetated slopes may have higher hillslope transport rates than less vegetated slopes (Roering et al., 2010), which is consistent with observations from asymmetric landscapes (McGuire et al., 2014). On the other hand, variations in vegetation density in semi-arid climates drive contrasts in runoff with aspect (Gutiérrez-Jurado et al., 2013), which may increase the efficiency of channel incision and produce hillslope asymmetry (Yetemen et al., 2015). Given the large number of mechanisms that might produce aspect-related enhancement of drainage density or elongation of low-order channels, more focused studies that better constrain these factors may be required to understand the mechanisms in action at specific sites (e.g., Istanbuluoglu et al., 2008; Gutiérrez-Jurado et al., 2013; West et al., 2014; McGuire et al., 2014; Richardson, 2015).

Many mechanisms appear capable of producing topographic asymmetry, but only those that operate in semiarid landscapes underlain by weak rocks are consistent with our analysis (Fig. 4.6). In the Gablian Mesa, south-flowing catchments host channels that incise into the local bedrock (Fig. 4.7), which rapidly decomposes into fine sediment upon wetting and drying through slaking (Garcia, 2006). In contrast, soils formed from these clay-rich mudstones are cohesive despite experiencing similar changes in moisture content. This may produce a situation in which the soils offer more resistance to incision than the exposed bedrock, which allows bedrock-floored channels to expand headward at the expense of hillslopes. Finally, soil thickness appears related to local hydrology such that the dryer south-facing slopes have significantly thinner soils than north-facing aspects (Johnstone et al., *in review*). Thus, aspect-dependent variation in pedogenic processes may play a central role in the ability of channels to access bedrock and initiate headward-expanding, low-order channels

on south-facing slopes (Johnstone et al., *in review*). Aerial imagery from the Lower Kern site highlights features in south-flowing tributaries similar to those observed in the Gabilan Mesa (Fig. 4.7). Channel heads appear to abruptly truncate hillslopes and partially expose the underlying rock (Fig. 4.7), similar to what is observed in the Gablian Mesa. This set of processes provides a plausible mechanism for observed correlations between the preferential elongation of low-order channels and enhancement of drainage density along south-facing aspects in weak rocks. Intriguingly, a number of past studies that show profound asymmetry in the orientation of low-order tributaries are also focused on landscapes developed in young, likely weak, sedimentary rocks in climates where $AI < 0.5$ (Bass, 1929; Emery, 1947; Dohrenwend, 1978; Istanbuluoglu et al., 2008). Thus, it is possible that in each of these environments, pedogenic processes may play an important role in producing landscape-scale asymmetry.

Not all our sites developed in weak rocks are highly asymmetric, particularly those in wetter climates and qualitatively more resistant lithologies. In particular, the AI at the Corning site was the largest of all study areas, (Fig. 4.2 & Table 4.1) and past climates may have produced larger AI values than those observed today (e.g., Oster et al., 2015). The weak rocks of the Hualapai site are the oldest surveyed (Table 4.1), and may be more lithified than others falling in this designation. Such conditions may discourage headward expansion of equator-facing channels. While not all studied landscapes underlain by weak rocks show strong asymmetry, strongly asymmetric landscapes appear to be exclusively underlain by weak lithologies (Fig. 4.6). This may indicate that formation of asymmetric landscapes produced by preferential headward expansion of low-order channels requires a confluence of aspect-dependent critical zone processes that produce soils that are stronger than the underlying bedrock.

4.6 Conclusions

Aspect-controlled hillslope asymmetry occurs across a diverse range of climates and rock types (Poulos et al., 2012). However, cross-valley profiles demonstrate that this asymmetry extends into channel networks in some landscapes. We compare drainage network structure in asymmetric landscapes to synthetic symmetric channel networks to demonstrate that landscape-scale asymmetry is largely created by the elongation of low-order channels and increasing drainage density along equator-facing channels. These situations are exclusively found in landscapes underlain by rock types expected to yield little resistance to incision. This result is consistent with field observations that document rapid headward growth of low-order channels, instigated by the exposure of readily disaggregated bedrock beneath a more cohesive soil mantle. This suggests that critical zone processes that are shaped by microclimates and lithology can influence the structure of entire drainage basins.

4.7 Figures and tables

Region	Lithology	classification	Mean AI	Mean L_{hs} [m]
Corning	Tertiary andesite & rhyolite	strong	0.76	204
Corning	Miocene to Pleistocene SS & CGL	weak	0.44	39
Gabilans	Early to Late Cretaceous granodiorite and quartz monzonite	strong	0.39	116
Gabilans	Miocene to Pleistocene SS & MS	weak	0.22	107
Lower Kern	Permian to Tertiary, granodiorite and quartz monzonite	strong	0.35	180
Lower Kern	Miocene to Pleistocene SS & CGL	weak	0.16	60
Hualapai	Proterozoic - Early Tertiary granite & granodiorite	strong	0.28	147
Hualapai	Miocene to Pliocene CGL & SS	weak	0.17	105

Table 4.1: Mean AI , rock types, and mean L_{hs} for the sites where topographic asymmetry was measured. Locations are depicted in Figure 4.1. Abbreviations for sedimentary rocks are as follows; sandstone (SS), mudstone (MS), conglomerate (CGL). Ages and descriptions of rock units from (Ludington et al., 2005).

Region	Classification	Number of basins	Mean area (std) [km ²]	Area range [km ²]
Corning	strong	10	94 (76)	17 - 234
Corning	weak	9	59 (40)	16 - 140
Gabilans	strong	5	38 (22)	11 - 62
Gabilans	weak	8	34 (27)	5 - 81
Lower Kern	strong	8	77 (46)	2 - 130
Lower Kern	weak	6	22 (5)	16 - 30
Hualapai	strong	4	75 (57)	25 - 140
Hualapai	weak	5	32 (18)	16 - 63

Table 4.2: Rock types and summary statistics for the basins used to measure asymmetry. The number of basins is given. Statistics are derived from the area used to measure asymmetry within each basin (e.g., excluding low relief areas or areas outside GB polygons). Locations are depicted in Figure 4.1.

Figure 4.1: Map of the the ratio of precipitation to potential evapotranspiration, AI , from the CGIAR-CSI Global-Aridity database (Zomer et al., 2008) highlighting the regions used in this study. Lower values of AI correspond to more arid climates. In general, four or more drainage basins were selected from multiple geologic units within each site.

Figure 4.2: Methodology and observations of asymmetry in $NDVI$ as a function of aspect. **A** Example data from the Gabilans site highlighting the approach used for calculating $NDVI_{asym}$ in a single GB polygon. Small dots represent the average $NDVI$ value on north-facing (blue) and south-facing (red) slopes measured in $NDVI$ catalogs collected at 32 day intervals over a 15 year period. Large circles represent the monthly averages over the 15 year period. **B** Correlation in $NDVI_{asym}$ with AI . Different colored dots represent the day of the year when $NDVI_{asym}$ has the greatest magnitude. **C** Variations in $NDVI_{asym}$ as a function of AI . Red line is a linear fit to ‘dry’ sites where north-facing slopes have greater $NDVI_{asym}$ than south-facing slopes. Blue line is a fit to ‘wet’ sites with the opposite sense of $NDVI_{asym}$.

Figure 4.3: Maps of an observed drainage network and a synthetic realization of a random network with the same bounding geometry. Solid lines within basin bounds show flow paths where $L^* \geq 3$, dashed line shows the intersection of a geologic unit and the drainage basin boundaries (e.g., a GB polygon). Distributions of ^{rel}N for these basins are shown in 4.4.

Figure 4.4: Distributions of $relN$ as a function of L^* (distance from center of circle) and θ_f (radial coordinate with north pointed up). Dashed, black circles denote L^* values of 5, 25, and 50. Data are plotted in regions where $relN \geq 0.125$, indicating that flow paths in this length and orientation are relatively common. Panel **A** & **B** highlight the observed and randomly simulated drainage networks shown in 4.4, respectively.

Figure 4.5: **A** Relationship between the ratio of mean north- and south-facing slopes within a basin to the ratio of profile slopes measured within that region. Symbols denote rock types, triangles represent weak rocks while circles represent strong rocks. Symbols are color-coded for average AI (Table 4.1). Small gray box in lower left highlights region where north-facing slopes would be less steep than south-facing slopes; dashed line highlights a 1:1 relationship. **B** Normalized topographic profiles from the Lower Kern site (gray) and the form those profiles would take if asymmetry was perfectly represented by the basin averaged ratio of opposing slopes (red). Profiles are normalized by the total length, L_t , and relief, $Z_{max} - Z_{min}$, of the profile to scale to 1 in the x- and y-directions, and are shifted based on the elevation and along-channel position of the minimum elevation, Z_{thal} and L_{thal} .

Figure 4.6: Distributions of N^* as a function of L^* (distance from center of circle) and θ_f (radial coordinate with north pointed up). Dashed, black circles denote L^* values of 5, 25, and 50. Data are plotted in regions where $N^* \geq 1$. Varying degrees of transparency are used to indicate different p values for the comparison between observed and simulated ^{rel}N values.

Figure 4.7: Aerial imagery of the Lower Kern site (A) and Gabilan site (B) in unconsolidated sediments. White patches show likely areas of exposed bedrock in an otherwise grassy, soil-mantled landscape. Small black arrows highlight these white patches and areas where channel heads truncate convex drainage divides. Latitude and longitude coordinates of upper left corner of A & B are $36^{\circ}7.908'$, $-120^{\circ}56.055'$ & $35^{\circ}33.687'$, $-118^{\circ}58.782'$

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