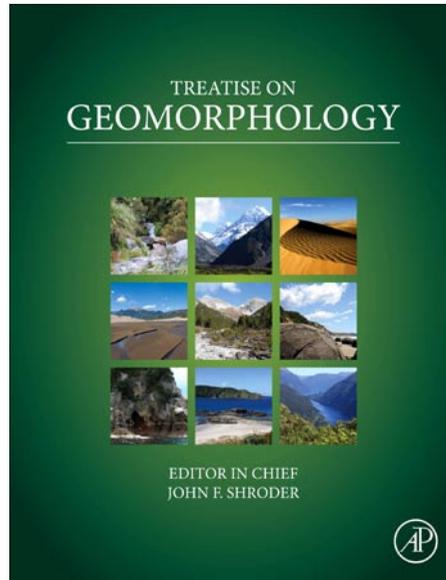


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7.29 Changing Hillslopes: Evolution and Inheritance; Inheritance and Evolution of Slopes

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Glossary

Glacial buzzsaw The theory of the strongest glacial erosion occurring just below the warm-based, erosive zone of ice with melt-water availability, and the upper, cold-based zone of ice frozen to the bedrock where little erosion occurs.

Graded slope A slope of just such inclination and character that sediment input, throughput, and output remain in equilibrium on the slope. No change in the slope over time can be detected unless the balance of forces changes.

Kafkalla slope A slope eroded in secondary limestone that was precipitated by soil-forming weathering processes in the eastern Mediterranean.

Richter slope A slope segment of uniform gradient at the base of a retreating cliff with the gradient of the bedrock slope above equivalent to the gradient of the talus slope below.

Sigmoidal slope A typical hillslope with an upper convex part and a lower concave portion.

Transport-limited slope A slope covered with sediment above bedrock where the operant erosion processes are insufficiently active to remove weathered debris that is accumulating.

Weathering-limited slope A slope of largely bare rock where almost all weathering debris is removed fairly quickly.

Abstract

The antiquity and inheritance of hillslopes have long fascinated geologists seeking to unravel the impact of climate on hillslope morphology. Given the onset of profound climate oscillations in the last several million years, Neogene

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landscapes may have differed significantly from the modern Earth surface. Early views on climate–morphology linkages have also differed greatly; some ascribed nearly every feature of modern slopes to past climate regimes, whereas others noted the ubiquity of slope forms worldwide and thus rejected a primary role for climate. Efforts to differentiate between these divergent views were hampered by a lack of model testing. The revival of topographic surveys in the 1950s encouraged quantitative analysis of slope forms and explicit treatment of hillslope processes. More recently, the coupling of process-based models for sediment transport, erosion rate estimates via cosmogenic radionuclides, and widespread topographic data has enabled the testing and calibration of process-based models for hillslope interpretation and prediction. In soil-mantled terrain, models for soil transport and production predict that hillslope adjustment timescales vary nonlinearly with hillslope length; the adjustment timescale for typical settings should vary from 10 000 to 500 000 years, similar to the timescale for glacial–interglacial and other climate fluctuations. Because process-based models for bedrock landscapes are poorly understood, we have limited ability to quantify, for example, post-glacial rockfall and scree slope formation. Although the paradigm of steady-state hillslopes has facilitated the testing of numerous process models in the last several decades, this assumption should be relaxed such that climate–hillslope linkages can be more clearly defined.

7.29.1 Introduction

Hillslope morphology is perhaps the most accessible and frequently documented characteristic of landscapes. Even the casual traveler instinctively notices the dramatic differences, for example, between steep, rocky talus slopes and rolling, gentle, soil-mantled terrain. More than 150 years ago, geologists began documenting the myriad of hillslope forms and pondering their origin and evolution (Powell, 1875; Gilbert, 1877; Darwin, 1881; Davis, 1899). Concurrent investigations into the profound climatic variations in Earth history (e.g., Agassiz, 1840) caused these investigators to consider the possibility that the hillslopes we encounter today may owe some fraction of their character to prior climatic regimes. The extent to which landscapes retain a paleoclimate signature became infused into the rapidly growing inquiry into landscape evolution. From a generalized and perhaps overly simplistic perspective, highly contrasting views were promulgated; at one extreme, workers such as Budel (1982) suggested that nearly all mid-latitude landforms were inherited from glacial time periods, whereas others such as King (1957) noted the ubiquity of certain hillslope forms and minimized the role of climate in driving slope differentiation. The interpretation of inheritance in hillslope morphology thus deserves a systematic examination, particularly in light of technological advances (including computing power, cosmogenic dating, and airborne laser mapping) that have moved the geosciences forward in the last 20 years.

Correlation between sedimentation events and climate has been offered as evidence for a profound climatic control on hillslope and landscape evolution in the late Cenozoic. Sedimentary basins in diverse settings experienced a 2- to 10-fold increase in sedimentation rate in the last 2–4 million years, commensurate with the global onset of cool conditions and a change in the amplitude of Milankovitch-driven, glacial–interglacial cycles (Zhang et al., 2001; Molnar, 2004). Although ambiguity exists as to the feedbacks between weathering, erosion, and CO₂ that drive this cooling (e.g., Raymo and Ruddiman, 1992; Willenbring and von Blanckenburg, 2010), the impact of these fluctuations on landscapes has been significant. Zhang et al. (2001) hypothesized that climate variability enhanced sediment production by subjecting landscapes to periods of intensified sediment production followed by periods of increased sediment transport. Oscillations between enhanced transport and production are thought to

function as a positive feedback and enhance the net export of sediment relative to the stable state (Zhang et al., 2001; Molnar, 2004; Norton et al., 2010). Although these patterns have been challenged in some settings due to concerns over extrapolating borehole data and sediment preservation (Metivier, 2002; Clift, 2006; Schumer and Jerolmack, 2009), the ubiquity of increased sediment delivery leading into the Quaternary (even in unglaciated regions) causes one to ponder the morphologic implications of a cooling climate and glacial–interglacial climate oscillations. At the landscape scale, changes in climate may prevent landscapes from obtaining an equilibrium state whereby morphology is adjusted to the current conditions and erosion rates are balanced by rock uplift via tectonic forcing or isostatic compensation (Zhang et al., 2001). This assertion compels us to ask: (1) How persistent are landforms associated with a given climate, or in other words, what are the timescales by which hillslopes adjust to changing climate? (2) More fundamentally, do landscapes retain the topographic signature of climate conditions prior to Plio–Pleistocene cooling? Although most sedimentary systems may be challenged to record high-frequency flux variability (Castelltort and Van den Driessche, 2003), investigations of colluvial deposits in favorable settings provide compelling evidence for strong climate modulation of sediment production on Milankovitch timescales (Pederson et al., 2000, 2001; Clarke et al., 2003; Dreibrodt et al., 2010). However, the associated morphologic adjustments remain poorly constrained. In this chapter, we strive to clearly differentiate between climate control of sedimentation and climate control of landscape morphology. Despite the accessibility of the Earth's surface, our interrogation of hillslope forms has achieved minor progress in comprehensively addressing the aforementioned queries.

The interpretation that changes in global sediment yield correlate with the amplitude of climate variability is compelling and intuitive because it matches many of the geological observations of erosional disequilibrium in catchments within bedrock (and soil-mantled) landscapes (e.g., Bull and Knuepfer, 1987; Bull, 1991; Pan et al., 2003; Anders et al., 2005). Unraveling these effects has proved to be challenging. Ideally, one could explore the influence of climate change on slope form by developing a suite of climo–hillslope relationships from well-chosen study sites (e.g., Peltier, 1950; Tricart and Muslin, 1951; Toy, 1977; Budel, 1980) and analyzing how transitions between these characteristic forms manifest, providing a roadmap for identifying inheritance in real landscapes. Unfortunately, efforts

to link slope morphology to specific climate regimes have been widely attempted, but heavily criticized because of uncertainty in the climate regime responsible for shaping specific landforms (e.g., Melton, 1957; Toy, 1977; Dunkerley, 1978). As it stands, few settings have escaped significant climate variation. Perhaps tropical landscapes provide an opportunity for associating morphologic trends with climate characteristics, given the relatively minor climate adjustment in these areas (Ruxton, 1967; Young, 1972; Pain, 1978). Of course, other factors such as rock type and base-level lowering further confound this endeavor and motivate a more systematic approach.

The morphologic signature most relevant to our purposes is that defined by ridges and valleys, features that define drainage basins, rather than orogen-scale features, such as forearc highs, or more localized features, such as patterned ground and solifluction lobes. The list of landscape metrics used to define ridge-and-valley terrain and drainage basins is surprisingly short, and includes local relief, slope, curvature, and drainage density. For decades, slope profiles served as the primary technique for portraying and communicating theories for slope development, although few geologists endeavored to objectively measure slope angles in the field until the 1950s (Strahler, 1950; Savigear, 1952; Koons, 1955; Schumm, 1956a, 1956b; Melton, 1957). The formalization of mathematical models for hillslope evolution further motivated the systematic analysis of field data for comparison with model predictions (Scheidegger, 1961; Ahnert, 1970; Kirkby, 1977). The advent of digital elevation models (DEMs) has inspired some new metrics, such as area-slope plots to define hillslope convexity and the hillslope valley transition and spectral analysis to quantify hillslope–valley spacing, but additional tools are needed. In this chapter, our goal of characterizing slope inheritance highlights the need for improved morphologic metrics with increased diagnostic capability.

Here, we review work that addresses the inheritance and evolution of hillslopes. For this effort, we focus on the form and dynamics of soil- and bedrock-mantled slopes not significantly influenced by denudation from aeolian or solution processes. Although base-level lowering ultimately driven by tectonic forces carves valleys and generates relief, here, we focus on climate as the primary driver of landscape change. Although climate is generally simplified in terms of annual temperature and precipitation, in fact, a multitude of climate-driven factors control hillslope evolution, such as vegetation and frost processes. Thus, the geomorphic implications of both direct and indirect consequences of climate change are of interest here. Although we do not address glacial erosion directly, we do describe the post-glacial modification of glacierized regions, particularly due to periglacial processes. We seek to balance a process-based perspective with the impressive heritage of work on landscape inheritance and evolution. The body of literature on the topic of slope evolution and inheritance is vast, and while highlighting major contributions, we focus on two primary climate-driven questions: (1) What did landscapes look like before the advent of Plio–Pleistocene climate change and (2) how are glacial–interglacial fluctuations recorded in landscape morphology? Tackling these questions is crucial for informing our predictions of geomorphic response to man-made climate changes.

7.29.2 Hillslope Evolution

7.29.2.1 Historical Context

The evolution of hillslope form has traditionally been depicted in one dimension in both theoretical and empirical studies because cross sections are relatively easy to illustrate schematically and measure in the field. Prediction of hillslope evolution began with Fisher's (1866) profile model for the evolution of bedrock quarry faces. After his initial work, a wide range of conceptual models attempted to explain realistic hillslopes using relatively simple geometries (Davis, 1930; Penck, 1953). Until the 1950s, however, these models were seldom tested or calibrated with actual field data (e.g., Savigear, 1952). The analysis of diverse geologic settings promoted the notion that characteristic slopes and slope forms occur within different climatic and tectonic settings. In the early twentieth century in which the Davisian geographic cycle dominated geomorphic disciplines, interest in slope profile evolution spawned several concepts that have emerged in the more recent realm of process geomorphology, notably the concept of threshold hillslopes (Carson and Petley, 1970). These theories have been discussed and reviewed in varying detail over the past century (e.g., Carson and Kirkby, 1972; Young, 1972; Clark and Small, 1982; Parsons, 1988; Tucker and Hancock, 2010). This chapter briefly reviews six major hillslope evolution theories that provide the basis for much of the theoretical treatment of slope profile evolution.

7.29.2.2 Fisher's Cliff Retreat (1866)

The first slope evolution model was a geometrical exercise that predicted the evolution of a vertical bedrock quarry face in southern England. If the quarry face were vertical and bounded by horizontal surfaces, then any material that was weathered would accumulate as scree (**Figure 1(a)**). If the rock is equally exposed to weathering along its face, then at each timestep a layer of rock of uniform thickness is removed and deposited as talus below the face. The resultant bedrock profile beneath the scree slope is characteristically parabolic and Fisher proposed a mathematical expression for this bedrock slope based on the geometry of the bedrock slope and the scree slope. Although lacking in information about the processes driving slope evolution, this model is significant in two ways: (1) it provided an early example that geomorphic processes can be mathematically explained and (2) its simple geometrical argument produced realistic slope geometries that influenced qualitative studies of hillslope form (Lawson, 1915; Wood, 1942). Fisher's approach was further generalized by Lehmann (1933), who included a wider range of hillslope and scree geometries. The theory was later modified by Bakker and Le Heux (1952) such that the bedrock slope evolved at the angle of repose of the overlying talus creating a Richter slope (a uniform gradient slope segment at the base of a retreating cliff with bedrock gradient equal to the talus gradient).

7.29.2.3 Davis' Graded Slopes and Their Progressive Downwearing (1899)

Davis described how slopes evolved through time in the context of his "cycle of erosion" (Davis, 1899). In his model,

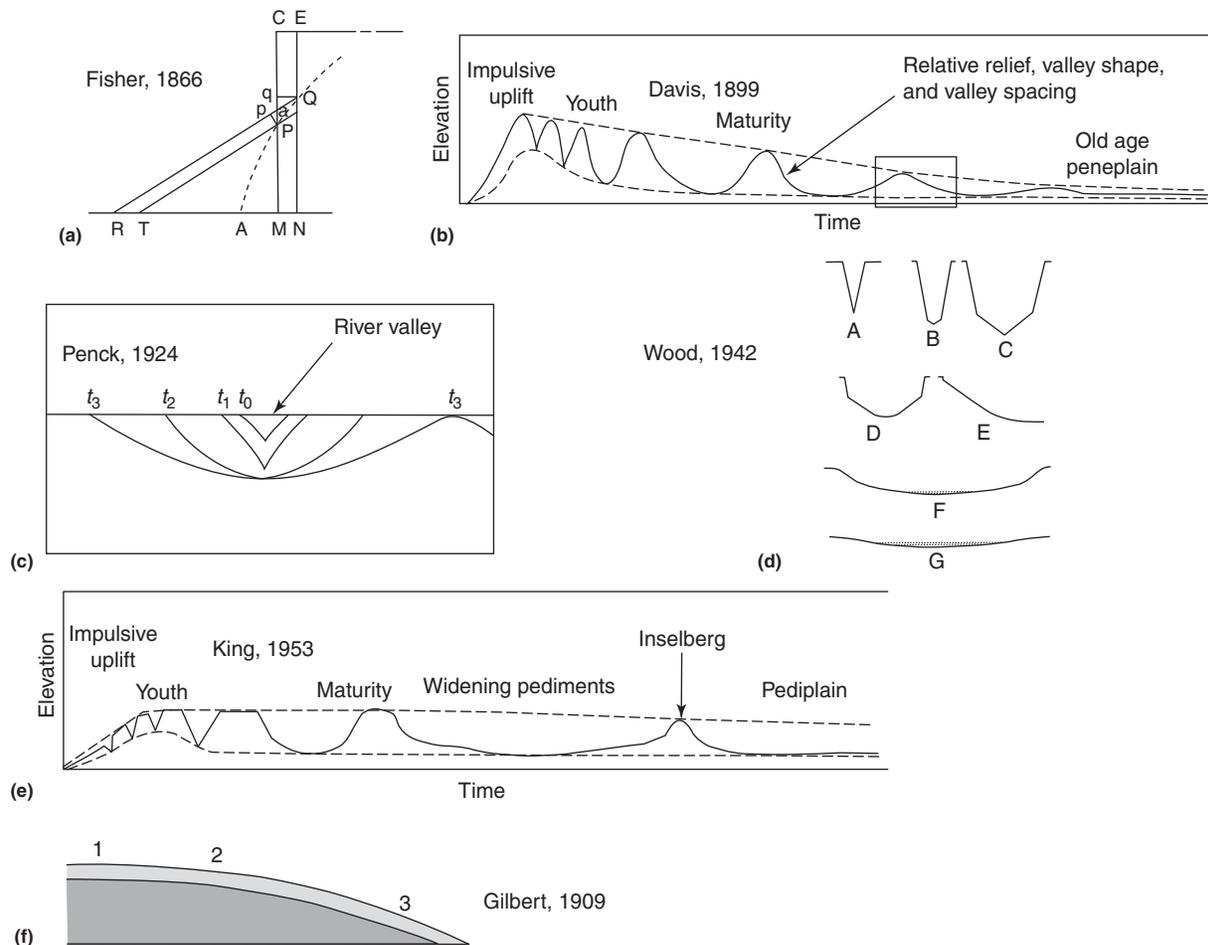


Figure 1 (a) Profile view of Fisher's (1866) cliff retreat model for chalk bedrock and scree deposition. CP is the face of the cliff, A is the bottom of the quarry, TP is the talus surface. Retreat of the headwall to QE raises the talus slope to RQ and a curved interface PQ results. Reproduced from Fisher, O., 1866. On the disintegration of a chalk cliff. *Geological Magazine* 3, 354–356. (b) The cycle of erosion proposed by Davis (1899) and modified by Summerfield (1991) showing declining relief, valley widening, and slope downwearing. Reproduced from Davis, W.M., 1899. The Geographical Cycle. *Geographical Journal* 14, 481–504. (c) Penck's slope replacement model (1924, 1953) showing successive retreat of an infinite number of inclined valley bottoms leading to a smooth concave profile. Reproduced from Penck, W., 1953. *Morphological Analysis of Land Forms: A Contribution to Physical Geology*. Macmillan, London, 417 pp. (d) Wood's slope cycle showing retreat and the maintenance of constant slope angles (a)–(d) before slope consumption and replacement by alluvial fill. Reproduced from Wood, A., 1942. The development of hillside slopes. *Proceedings of the Geologists Association* 53, 128–140. (e) King's parallel slope retreat model (1953) modified by Pazzaglia (2003) allowing for lateral translation of slope forms and the persistence of wide pediments and inselbergs of substantial relief. Reproduced from King, L.C., 1953. *Canons of landscape evolution*. *Geological Society of America Bulletin* 64(7), 721–752 (apply thru CCC). (f) Hack (1960) and Gilbert (1877) model of steady state (or dynamic equilibrium) suggests that slope forms are time independent. The soil-mantled profile shown here maintains a convex form by increasing flux rates with gradient (Gilbert, 1909). Reproduced from Gilbert, G.K., 1909. The convexity of hilltops. *Journal of Geology* 17(4), 344–350.

the deep incision of gorges into an initially uplifted surface created steep linear hillslopes (Figure 1(b)). Through time hillslopes develop a sigmoidal form, with a concave base and convex upper slope. Once the initial shape of this hillslope has been attained, it is maintained through time, but the gradient of the hillslope decreases by downwearing until a flat, penultimate plain forms (Davis, 1899). Davis discussed slopes in the same terms that he used for fluvial systems, referring to the characteristic form of hillslopes at each stage of the cycle as a graded profile. This condition was achieved when the slope achieved a consistent regolith cover without rock outcrops (a full definition of grade and other terminology from that era can be found in Young, 1970). In most discussions of the geographical cycle, the progressive downwearing required to

create graded slopes was discussed without reference to process (with erosional processes being described as agencies of removal). Nonetheless, Davis discussed his cycle with specific reference to the character of hillslope sediment at each stage of his cycle, suggesting that the linear hillslopes of youth contain coarse regolith of moderate thickness, while the sigmoidal, old age profiles, with their low slope and weak agencies of removal, contained fine-grained regolith of great thickness.

Much has been written about Davis' cycle of erosion (Carson and Kirkby, 1972; Young, 1972; Parsons, 1988; Pazzaglia, 2003), particularly by prominent dissenters of his theory of slope evolution (Bryan, 1940; King, 1953; Penck, 1953). Interestingly, both Penck (1953) and King (1953), who worked in bedrock landscapes in European Alps and

southern Africa, respectively, formed alternate hypotheses as a direct response to the lack of significant soil-mantled sigmoidal slopes. Their insights appear to have also had an impact on Davis' own theory; toward the end of his career he described the evolution of graded slopes as a 'normal' or humid cycle of erosion. He contrasted this with arid and semi-arid hillslopes, suggesting that "the various processes of arid and of humid erosion are thus seen to differ in degree and manner of development rather than in nature, and as their differences in degree and manner are wholly due to differences in their climates, so the forms produced by the two erosions may be shown to differ in the degree to which certain elements are developed rather than in the essential nature of the elements themselves" (Davis, 1930, p. 147). The explicit recognition of climate as driving differences in geomorphic process mechanisms and rates is a significant departure from the simple concept of a single, graded hillslope profile. Discussion of the debate about the differences between Davis and Penck's theories is extensive and we suggest that the reader explore the excellent reviews by Bryan (1940), King (1953), Carson and Kirkby (1972), and Parsons (1988) for more detailed discussion.

7.29.2.4 Penck's Slope Replacement (Originally Published in 1924 in German with the Title 'Die Morphologische Analyse', Translated and Published in English in 1953)

Penck's model for hillslope evolution is driven by replacement of progressively shallower slopes along the slope base and directly contrasts Davis' graded slopes (Figure 1(c)). Penck's initial condition is similar to Davis' youthful landscape, whereby a flat plateau is incised by a stream that is no longer eroding but has the capacity to remove material. The steep, linear hillslope is subjected to equal weathering across its length, increasing the mobility of the regolith. The mobile regolith is removed by denudation at a rate that is proportional to its slope, leaving a low-gradient lower slope segment. Through time, the steeper upper slope (or *steilwand*) retreats and the shallower slope remnant (or *haldenhang*) increases in length. During retreat of the *steilwand*, the *haldenhang* is also subjected to weathering and denudation of this slope occurs when the surface layer is fine-grained enough to be mobile. Through time, progressively shallower slopes form, the surface layer of each containing progressively finer material. Penck's contribution is significant in two ways: (1) he discussed his slope retreat model in the context of a series of process rules (e.g., the notion that weathering leads to an increase in the mobility of material) and (2) he described his model using geometrical and mathematical arguments, supporting the idea first proposed by Fisher (1866) that landscape evolution can be formalized mathematically.

7.29.2.5 Wood's Slope Cycle

Wood (1942) suggested a model for the development of slopes that included the effects of weathering and transport. Although still rooted in the general framework of the cycle of erosion, Wood's contribution was to explicitly describe how processes acting on slopes control their evolution, leading to the

suggestion that changes in slope profiles represent changes in the active transport processes. His paper began by discussing the evolution of a vertical bedrock slope beneath which a talus pile forms. He suggested that progressive weathering of the free face creates a talus slope of constant angle (Figure 1(d)). The length of the free face shrinks through time creating a convex buried bedrock slope (Fisher, 1866; Lawson, 1915; Lehmann, 1933). Taking this model he applied it to valley slopes formed during a cycle of erosion. After deep incision and formation of a free face, a decrease in incision rate and widening of the valley allow the constant (talus) slope to form. Below the talus slope a convex-upward, waning slope develops. The waning slope forms because of the size sorting that occurs during transport of material from the constant slope, where finer material is carried farther from the main slope. Finally, weathering at the junction of the free face and the incised plateau creates a convex-upward waxing slope, primarily by rainwash and hillslope creep. Through time, valley alluviation and lateral corrasion cause the free face and constant slope to reduce in size leaving a broad peneplain (Wood, 1942).

7.29.2.6 King's Parallel Slope Retreat (1953)

King (1953) discussed the inconsistencies in the Davisian model of hillslope evolution based on empirical evidence derived from his home country of South Africa. Possibly the key contention put forth by King related to peneplanation, which culminates in a shallow, convex-upward slope (King, 1953). King's alternative model, which was strongly influenced by Penck (1953), Bryan (1940), and Wood (1942), suggested that a landscape was composed of a series of nested, retreating escarpments (Figure 1(e)). Each escarpment is composed of a free rock face contributing sediment to a talus slope below which sits a convex-upward pediment (the equivalent of Wood's waning slope). Escarpments maintain this general form as they retreat across the landscape creating a landscape of shallow, concave-upward pediments. These pediments persist until consumed by another retreating escarpment formed during a subsequent fall in base level. King was strongly influenced by his local geography, but suggested that parallel slope retreat was not restricted to arid landscapes, but could also be applied to soil-mantled, humid settings.

7.29.2.7 Gilbert (1877) Revisited: Hack's Dynamic Equilibrium (1960)

Gilbert's (1877) law of equal action posits that hillslopes (and channels) adjust their morphology such that rates of erosion tend toward uniformity (Figure 1(f)). On hillslopes, Gilbert (1909) used uniform erosion to suggest that sediment flux increases with slope angle. Hack (1960) revisited this concept using process-oriented observations from the Appalachian Mountains, USA. The suggestion that landscapes tend toward a balance between uplift and erosion implies time independence. This concept refocused studies toward process and, in doing so, established the notion that "landscapes are essentially modern and a reflection of modern processes" (Pazzaglia, 2003, p. 254). Interestingly, the rapid reemergence of the steady-state landscape paradigm moved the focus away

from identifying the signature of past climates in topography in favor of a stricter interpretation of process mechanisms, rates, and current forms.

7.29.2.8 Process-Based Modeling and the Continuity Equation

The form of hillslopes derives from sediment production and transport processes acting in conjunction with base-level forcing realized by the vertical and/or horizontal movement of channels. For our purposes, channels communicate the effect of vertical motions of the Earth's surface (such as that caused by rock uplift) to hillslopes and thus constitute boundary conditions that drive slope evolution. In a given setting, the suite of active hillslope processes depends on diverse variables that include rock type (e.g., rock mass strength), climate (e.g., storm frequency and intensity), biology (e.g., ecosystem composition), human activity (e.g., timber harvest), and tectonic forcing (e.g., earthquake magnitude and frequency). The formulation of quantitative relationships for sediment production and transport processes that account for these variables in a particular landscape implies that characteristic (although perhaps nonunique) hillslope forms will emerge given sufficient time. This construct indicates, for example, that hillslope erosion rates do not directly depend on climate variables, such as annual precipitation; instead, hillslopes adjust their form according to climate-related processes and erode at rates that match channel incision.

Although early mathematical models for slope evolution used geometric arguments to predict cliff retreat and debris slope evolution (e.g., Fisher, 1866), subsequent efforts incorporated the sediment continuity equation to systematically explore the implications of different sediment-transport processes and boundary conditions. Advancing the work of Culling (1960, 1963, 1965), who pioneered the use of realistic boundary conditions for modeling slope evolution and formalized particle motions on hillslopes, Kirkby (1971) proposed a suite of characteristic hillslope forms that derive from sediment-transport models with varying efficacy of hydraulic and gravitational forces. In essence, the vast majority of geomorphic models in use today is an extension or direct application of Kirkby's framework. Kirkby emphasized through his models that characteristic hillslope forms emerge and thus obliterate the initial or inherited slope configuration. That said, he certainly recognized the unlikelihood that most landforms are a reflection of the contemporary climate, a point emphasized by contemporaneous empirical studies (Arnett, 1971) as well as critiques of landscape evolution models (Selby, 1993).

The model established by Kirkby and colleagues has been instituted in a multitude of geomorphic papers (e.g., Willgoose et al., 1991; Anderson, 1994; Howard, 1994; Tucker and Slingerland, 1997). Here, we present a simple (neglecting bulk density differences between soil and bedrock) and commonly used pair of continuity expressions for exploring climate controls on hillslope form:

$$\frac{\partial z}{\partial t} = U - P + \frac{\partial h}{\partial t} \quad [1a]$$

$$\frac{\partial h}{\partial t} = P - \nabla \cdot \tilde{q}_s \quad [1b]$$

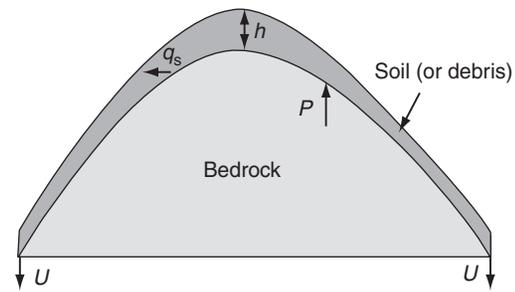


Figure 2 Schematic hillslope illustrating the components of continuity-based hillslope evolution models. Rock uplift acts at the margins of hillslopes through channel incision (U), which is depicted vertically but could also be accomplished horizontally. The rock-to-soil conversion rate is represented by P and sediment-transport rate by q_s . Soil depth is defined vertically as h .

where z is the elevation, t the time, U the rate of base-level lowering (positive values correspond to incision), P the production rate of mobile soil or debris, h the soil or debris depth, and q_s the sediment flux or transport (Figure 2). Equation [1a] describes the rate of change of the hillslope surface and eqn [1b] describes the rate of change of soil (or debris) thickness as the balance between production and the divergence of sediment transport. On slopes devoid of a soil or debris mantle, landscape lowering is limited by the production rate, P , although most applications of these equations do not explicitly account for the transport of material across bare bedrock surfaces. As elaborated by Dietrich et al. (2003), expressions for q_s and P constitute geomorphic transport laws for soil transport, soil production, overland flow erosion, and landslide processes that shape hillslopes as well as valley-forming processes such as fluvial incision. Transport laws should be testable independent of the model and retain the essence of a physically based formulation. Here, we focus on soil (or debris) production, P , and sediment transport, q_s , as the primary means by which climate modulates hillslope form. As illustrated in modeling studies (Smith and Bretherton, 1972; Izumi and Parker, 1995; Rinaldo et al., 1995; Howard, 1997; Moglen et al., 1998; Tucker and Bras, 1998; Perron, 2006), the relative importance of sediment transport on hillslopes and in channels dictates drainage density and thus hillslope relief and drainage basin structure. Because these processes are highly dependent on climate, the evolution of entire catchments ought to contain a climate signature. For example, changes in the efficacy of soil transport on hillslopes (such as biogenic soil disturbances) can significantly influence relief, average slope angle, and land surface curvature. The effect of rock type is subsumed within the expressions for production and transport and is not explicitly treated here. This conceptual framework enables us to broadly cast our review in terms of how climate-driven changes in P and q_s may be manifested in hillslope morphology.

7.29.3 The Inheritance of Landforms Predating Plio–Pleistocene Climate Change

"The day is long past when it was the fashion to ascribe detailed landscape forms... to pre-Pleistocene morphogenesis" Cotton (1958).

The discovery that late Cenozoic cooling accompanied by amplified seasonal and glacial–interglacial fluctuations coincided with a significant increase in sediment yield from the world's rivers causes one to ponder what landscapes looked like prior to this profound change in the global climate regime. Certainly, glaciation would have been much less extensive and limited primarily to polar regions in the Neogene, but what morphologic properties characterized hillslopes during the relatively protracted period of warm and moist conditions in the Neogene?

Acknowledging Cotton's reticence, this is a challenging endeavor and addressing this question in a given setting requires constraints on paleogeography, base-level conditions (i.e., tectonic forcing), and the regional climate regime; it has even been argued that this endeavor is hopelessly complicated (Chorley, 1964; Selby, 1974). The deeply weathered interior lowlands of Australia, for example, feature many ancient landforms (with some perhaps even dating to the Precambrian) that derive from epeirogenic or isostatic tectonics combined with their current position in a zone of high aridity (Twidale, 1976; Ollier and Pain, 1996; Bloom, 1997; Twidale, 1998). In the Mesozoic and early Cenozoic, Australia's mid-latitude position induced warm, humid, and forested conditions that fostered deeply oxidized regoliths (e.g., laterites) developed on broadly rolling landscapes (Vasconcelos et al., 2008). Given these observations in Australia, late Cenozoic cooling inherited a deeply weathered and highly differentiated landscape that formed because the rate of weathering front propagation exceeded the erosion rate (eqn [1b]) for an extended period of time (Vasconcelos et al., 2008). This low-relief Neogene template apparently discouraged significant topographic modification across broad areas. Deep weathering and relicts of broadly planated surfaces also characterize portions of southern Africa (King, 1953; Partridge and Maud, 1987; Beauvais et al., 2008). Interestingly, pre-Pliocene landforms in the British Isles, central Europe, and Oregon also appear to feature deeply weathered regoliths and are characterized by rolling plains and hilly terrain with low to moderate relief (Fisher, 1964; Battiau-Queney, 1987; Mignon, 1997; Brault et al., 2004; Benito-Calvo et al., 2008). In Southern France, Late Neogene volcanic activity preserved a vast landscape and outcrops reveal a low-relief, mildly incised surface that is considered representative of the region at that time (Bonnet et al., 2001). The change to a dissected landscape is interpreted to reflect Holocene increases in precipitation as well as vegetation change. Early Pleistocene till deposits in Scotland sit atop deeply weathered bedrock and low-relief terrain (Fisher, 1964). More general conjecture on the role of preglacial topography on landscape evolution during glacial periods also supports a low-relief interpretation (Klimaszewski, 1960). These studies provide consistent glimpses of landscapes that existed across diverse settings prior to Plio–Pleistocene climate change. The characterization of Neogene landscapes has been closely allied with studies of residual deposits, particularly laterites, silcretes, and bauxites (McFarlane, 1976; Retallack, 2010). The conditions necessary for the formation of these deposits have been well studied and include a period of slow incision of low-relief terrain that predates protracted subaerial exposure and climate stability. The definition and use of paleosurfaces for landscape reconstruction are

discussed by Widdowson (1997) and paleosol-based landscape characterizations are discussed in detail by Retallack (2001). In active tectonic settings, paleosurfaces are much more difficult to identify such that paleotopographic reconstructions are highly data-limited (Abbott et al., 1997).

This brief literature survey highlights the challenges of establishing a generalized approach for characterizing Neogene hillslope forms (a more extensive treatment can be found in Bloom 1997); many sites provide suggestions of low drainage density and low-relief terrain. Nonetheless, issues of preservation bias, uncertainty in base-level forcing, and inheritance persist. Quantitative documentation of hillslope morphology and landscape dissection that might reveal the relative importance of hillslope and fluvial processes and facilitate the calibration of process-based models, however, are lacking although some attempts have been made to map paleolandscapes using abundant outcrop data and GIS analysis (Benito-Calvo et al., 2008). Advances in quantifying paleo-relief using thermochronology are beginning to provide constraints that may illuminate Neogene landscape properties (Reiners, 2007), although it is unclear whether these constraints will enable quantification of process-scale hillslope forms.

7.29.4 The Inheritance of Landforms during Glacial–Interglacial Fluctuations

Climatic variability, particularly glacial–interglacial cycles, causes different suites of surface processes to act on a landscape through time. This was first described in detail by Gilbert (1900, p. 1010), who suggested that “a moist climate would tend to leach the calcareous matter from the rock, leaving an earthy soil behind, and in a succeeding drier climate the soil would be carried away.” The implication of this for landscape morphology is significant, suggesting that at any point in time a landscape will contain morphologic elements that reflect both modern processes and those inherited from past climatic regimes (Ahnert, 1994). Morphologic inheritance is particularly obvious in arid and glaciated landscapes where talus slopes, alluvial fans, and pluvial lake deposits attest to past climates. For arid environments, increased physical and chemical weathering during wet, vegetated periods leads to increased sediment transport during drier climates via rilling in unvegetated soils (Harvey et al., 1999). In high mountains, the expansion and contraction of glaciers change catchment morphology and sediment flux in a profound fashion. Glaciers are efficient at eroding bedrock, transporting sediment (Hallet et al., 1996), and creating large U-shaped valleys. Deglaciation is commonly rapid and hillslopes adjust via rockfall, which creates talus, bedrock landslides, debris flows, and large fans (Ballantyne, 2002). Goodfellow (2007) presented a systematic and thoughtful analysis of preserved glacial surfaces emphasizing the implications for postglacial variations in denudation.

Morphologic inheritance and its role in controlling sediment flux from catchments have been discussed in detail by prominent workers in arid and glaciated environments (e.g., Bull, 1991; Ahnert, 1994). In the absence of climate changes and under conditions of constant tectonic forcing, these

authors argue that catchments approach a time-independent form through the negative feedback between erosion and base-level forcing (as suggested by Gilbert, 1877; Hack, 1960). When one of the external (called *eksystematic* by Ahnert, 1994) drivers of erosion (essentially tectonics, climate, and humans) perturbs this system, process rates change to nudge the landscape toward a new equilibrium (Bull, 1991; Ahnert, 1994). The suite of surface processes associated with the new climatic regime will act on the inherited morphology of the previous climatic regime until those forms are erased.

Perhaps the most widely discussed example of this approach is the process-linkage model (Bull, 1991), which suggested that a catchment will respond to an external perturbation in two stages, called the reaction time and the relaxation time. The reaction time is the length of time between a change in the external driver and a geomorphic response. Bull (1991) argued that because many geomorphic processes (e.g., fluvial sediment entrainment and landsliding) are nonlinear or threshold processes, the reaction time is dependent on the relative magnitude of the external perturbation and its persistence. Relaxation time represents the amount of time required to attain steady state after a process threshold has been crossed.

This conceptual model is best described by way of example. The Charwell catchment, New Zealand, is a nonglaciated catchment with a series of alluvial terraces preserved at its outlet due to progressive movement of the strike-slip Hope Fault. Bull and Knuepfer (1987) suggested that terrace aggradation occurred primarily during the transition from cool, dry periglacial conditions during the Last Glacial Maximum (LGM) to warmer, wetter conditions in the early Holocene. Intense periglacial weathering in the catchment during the LGM produced abundant coarse sediment exceeding the transport capacity of the Charwell River. An increase in precipitation in the early Holocene increased stream power, allowing the periglacially derived sediment to be transported to the outlet of the catchment creating alluvial terraces. As the supply of periglacially derived sediment decreased, stream power remained constant causing the stream to incise and approach a new equilibrium between sediment supply and transport. The Charwell River is an example of a catchment that moves between a bedrock periglacial-dominated landscape during glacial periods and a soil-mantled landscape during interglacial periods. The process-linkage model is not limited to these particular climatic conditions, as examples of rapid geomorphic change due to consequence of climate change have been discussed in arid (e.g., Bull and Schick, 1979; Anders et al., 2005), glaciated (e.g., Church and Ryder, 1972; Ballantyne, 2002; Anderson et al., 2006), and soil-mantled settings (e.g. Bull, 1991).

The concept of a time-dependent process response or relaxation time controlling the process response has been discussed in glacial settings via the theory of paraglaciation (Church and Ryder, 1972; Ballantyne, 2002). The term paraglacial, which is defined as “glacially conditioned sediment availability” (Ballantyne, 2002), conditions a general response that includes process changes associated with a morphologic legacy. Paraglacial theory combines the processes that occur after glaciation, emphasizing the morphologic response of all processes, rather than attempting to separate a single process and track its morphologic implications.

The notion that climate transitions generate increases in sediment production is largely historic. Differences in climatically and tectonically driven process rates were seen as the defining difference between Davis' normal, concavo-convex soil-mantled hillslope profiles and those of more arid landscapes (Davis, 1930; King, 1953), with Davis suggesting that all erosional processes act on a landscape, but that form is driven by those processes that are most efficient in a particular climatic or tectonic setting. For soil-mantled landscapes, weathering and erosion are driven by mechanical and chemical weathering processes that are facilitated by the presence of a soil mantle. In the absence of soil, a weathering-limited landscape may be subject to a more diverse suite of weathering processes (e.g., freeze-thaw, wetting and drying, and salt weathering) that strongly depends on rock properties (Yatsu, 1988; Graham et al., 2010). As a result, conceptual models for soil and bedrock dominated hillslope evolution tend to feature very different descriptors and we will address them separately below.

7.29.5 Bedrock Landscapes

7.29.5.1 What is a Bedrock Landscape?

Bedrock-dominated landscapes provide some of the most dramatic scenery in the globe with impressive cliffs of exposed bedrock, deep canyons, extensive talus deposits, and alluvial fans (Figure 3). Bedrock landscapes impress because they contrast the soil-mantled topography that most of the world's population lives on, where ubiquitous soil and vegetation commonly obscure bedrock exposure. Bedrock-dominated landscapes occur in areas where the net sediment flux from a hillslope exceeds the rate of soil production (Heimsath et al., 1997). This condition typically occurs in areas of steep topography and/or climates that limit soil production (usually arid or cold climates) as in desert, glacial, and periglacial landscapes (Figure 4). Bedrock landscapes can be broadly divided into cold, glacial, or periglacial landscapes, where soil formation is limited by cool temperatures and steep, generally tectonically active topography and warm, arid desert landscapes, where soil formation is limited by aridity and tectonics can vary. Although less extensive than soil-mantled landscapes, bedrock landscapes commonly occur along active tectonic margins that account for a significant proportion of the world's sediment supply (Hay et al., 1988). Sediment production from bedrock landscapes has been implicated in the cooling of Quaternary climate (Raymo and Ruddiman, 1992) and climate-driven changes in erosion appear to control Cenozoic sedimentation (Zhang et al., 2001; Molnar, 2004; Clift, 2010).

Sediment yield from bedrock landscapes is strongly controlled by the distribution of sediment and bedrock inherited from previous climate conditions. In particular, decoupling of periods of intense weathering with little erosion and vice versa appears to strongly control the sediment flux from bedrock catchments (Anderson, 2002; Anderson et al., 2006; Gunnell et al., 2009). The magnitude and timing of the decoupling between weathering and erosion can vary, but the transition between periods of global cooling and warming (glacial-interglacial cycles) is particularly important (Bull and Schick, 1979; Bull and Knuepfer, 1987; Bull, 1991). The

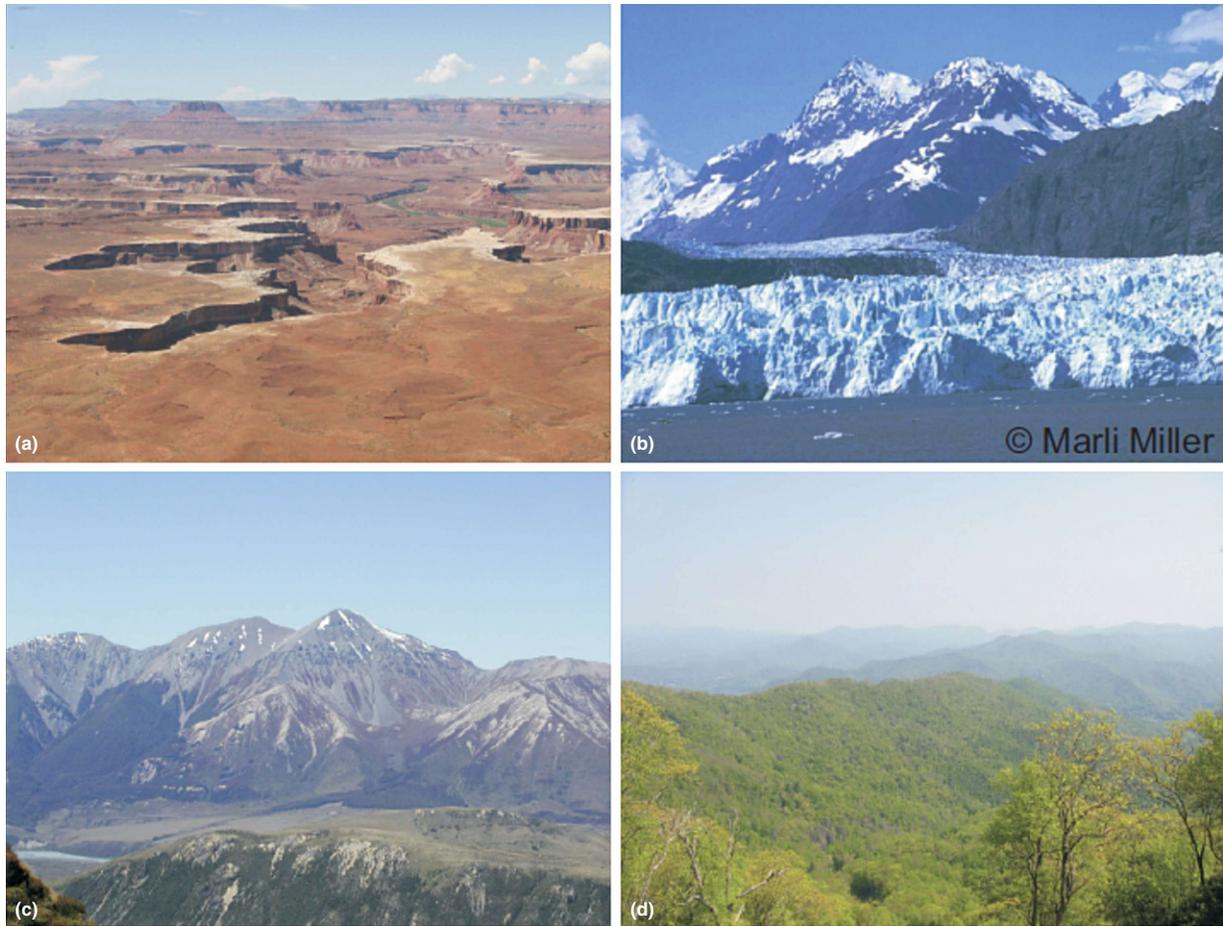


Figure 3 Imagery of bedrock and soil mantled landscapes. (a) Canyonlands, UT, is an example of a bedrock landscape formed in an arid landscape. (b) A glaciated bedrock landscape, Margerie Glacier, AK. (c) A periglacially dominated landscape of the eastern Southern Alps, New Zealand. By contrast, (d) shows the soil-mantled southern Appalachian Mountains, NC.

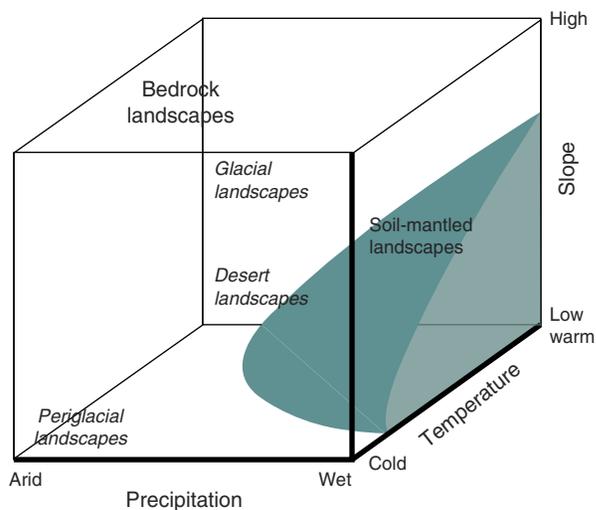


Figure 4 Schematic relationship showing the combination of temperature, precipitation, and topography (slope) that may dictate the occurrence of bedrock- and soil-mantled terrain. Because erosion rates strongly correlate with relief (and thus slope), this axis also acts as a surrogate for erosion rate.

concept of inheritance, where geomorphic processes and their rates are dependent on the history of a catchment, is essential to any system where weathering, erosion, and sediment transport are decoupled in space and time. A recent review reports the abundant literature conducted in arid settings using hillslope forms as a recorder of stability or activity (Schmidt, 2009). Many studies summarized therein seek to attach unique climate histories to specific hillslope landforms, such as flatirons. Here, we summarize how bedrock hillslopes evolve due to global climate fluctuations at a range of spatial scales from the landscape or orogen scale (10^2 – 10^4 km²) through to the scale of an individual hillslope (10^{-2} – 10^0 km²).

7.29.5.2 Changes from V-Shaped to U-Shaped Valleys and Back Again

The fluctuation between periods of glacial expansion and contraction has a very measurable topographic signature and a distinctive suite of landforms. The most recognizable of these morphometries is the U-shaped cross-sectional profile of glacial valleys that is discussed in all physical geology texts. The formation of U-shaped valley profiles from the V-shaped fluvial profiles causes significant sediment removal and has

been implicated as a control on the height of mountain peaks (Molnar and England, 1990) and the mean elevation of mountain ranges (Brozovic et al., 1997). After glacial retreat, these U-shaped valleys degrade at a rate that is dependent on the magnitude of the postglacial response of a suite of weathering and erosional processes (the paraglacial response; Ballantyne, 2002). Formerly glaciated valleys generally become major stores of hillslope sediment cut off from fluvial action by wide floodplains and periodically evacuated by the advance of glaciers during cooling (Hales and Roering, 2009).

The change from a dominantly fluvial to glacial system is hypothesized to cause changes in the geodynamics of mountain ranges. The formation of U-shaped glacial valleys from V-shaped valleys removes mass from mountains and transports it to the alluvial plain. The mass that is removed thins the crust and results in an isostatic response, lowering the mean height of the mountain by $(\rho_c/\rho_m)T$, where T is the average thickness of material removed, and ρ_c and ρ_m are the densities of the crust and mantle, respectively. The isostatically driven rebound caused by this process increases the height of mountain peaks, despite a lowering of the mean elevation of an orogen (Molnar and England, 1990). In mountains with deep glacial valleys, this mechanism may increase peak height by hundreds of meters. Glacial erosion is thought to control the mean elevation of mountain ranges through efficient erosion concentrated at the equilibrium line altitude (ELA) (Brozovic et al., 1997). Brozovic et al. (1997) studied the distribution of slope and elevation for regions of the Himalayan range with different exhumation rates and showed that regardless of exhumation rate, mean and modal elevations and slope distributions correlate with the extent of glaciation. Their glacial buzzsaw hypothesis suggested that efficient glacial erosion adjusted its rate in response to tectonics, cutting through the rock fast enough to account for very large uplift rates. The correlation between ELA and the elevation statistics of mountains (particularly maximum elevation and hypsometry) has been demonstrated to correlate with the global distribution of mountain topography and glacial advances (Egholm et al., 2009).

At the catchment scale, glaciers control the topography of an orogen through progressive widening and deepening of valleys (Harbor et al., 1988; Kirkbride and Matthews, 1997; Montgomery, 2002). This process occurs because glaciers occupy a large proportion of valleys such that erosion is distributed across a broad area. Valley formation and erosion occur by abrasion and plucking occurring beneath the ice surface (Harbor et al., 1988) and headward retreat of cirques (Oskin and Burbank, 2005). Estimates of the timescale required to erode a V-shaped profile into a U-shaped profile have been quantified using a space for time substitution in the Two Thumb Range, Southern Alps, New Zealand (Brook et al., 2006). Situated atop the overriding Pacific plate, these mountains undergo progressively higher rates of rock uplift due to convergence toward the plate boundary defined by the Alpine Fault. Valleys widen over a number of glacial cycles, taking more than 400 000 years (four glacial–interglacial cycles) to develop a recognizable U-shaped form (Brook et al., 2006). This result contrasts with estimates of 10^5 years for U-shaped valley development estimated by modeling valley incision using an ice-flux based erosion law (Harbor et al.,

1988). Despite these differences, both studies suggest that a significant increase in sediment flux would be associated with the initial development of the glacial valley. By contrast, at the longer timescale both sediment flux and valley development are affected by erosion and inheritance from interglacial cycles.

The obvious topographic and sedimentologic signature of the postglacial (paraglacial) response has led to a considerable literature discussing the evolution of glaciated valleys once a glacier has retreated. In this case, the inherited valley is U-shaped, with steep bedrock sides and a flat base that will commonly be filled with sediment derived from the retreating glacier. A braided river commonly occupies the center of the valley and sediment produced on hillslopes tends to be disconnected from the fluvial systems. Postglacial valley evolution is generally described as happening rapidly (on the order of a few hundred to thousands of years). Steep bedrock slopes develop distinctive scree slopes at their base, large bedrock landslides occur sporadically, and large alluvial and debris fans form (Ballantyne, 2002). The rate of sediment production from hillslopes into the alluvial system (or from any sediment source within a glacial valley to its sink) is thought to decline exponentially with time, although few empirical results exist to support this intuition. A diverse range of processes have been proposed to contribute to this sedimentary response (e.g., glacial debuitressing and frost action) with many of the processes having the potential to create a similar exponential decline in sediment flux, also referred to as an exhaustion model (Ballantyne, 2002). The paraglacial theory is a convenient method for describing the sedimentary response of a system that has been preconditioned by glaciations but debate exists about the geomorphic processes that control this response.

Studies of processes that control the paraglacial response suggest that hillslopes in glacial valleys primarily respond to the effects of periglacial weathering and glacial debuitressing. The idealized U-shape of glacial valleys means that glacial valleys often become steeper than can be maintained by the strength of the rock mass (Selby, 1982). When glaciers occupy these valleys, their steepness is maintained by the buttressing effect of glacial ice and the constant undercutting of the bedrock through glacial action. The removal of glacial ice causes the debuitressing of these slopes and they readjust to a stable angle via bedrock landsliding (Augustinus, 1992). The removal of the topographic load associated with the glacier has also been suggested to create new fractures that can weaken the rock and promote instability (Augustinus, 1995). Efforts to model the distribution of stresses caused by valley creation and glacial action have been equivocal about the magnitude of this effect, as valley formation appears to generate tensile stresses required to fracture rock in the valley floor rather than on the sides of the valley (Miller and Dunne, 1996). The curvature of the bedrock surface also controls topographically induced stresses (Martel, 2006). Numerous glacial valleys have slopes that are consistent with their rock mass strength (Augustinus, 1992), suggesting that any glacial oversteepening is likely to cause this response.

Periglacial weathering (rock weathering by nonglacial ice) is widespread in paraglacial landscapes, likely acting as an important trigger for both relatively small rockfall events (Hales and Roering, 2005) and large landslides (Korup, 2005,

2007). Periglacial weathering is caused by ice lens growth and decay beneath the rock surface either seasonally (Hales and Roering, 2005) or as permafrost (Gruber et al., 2004). Recently, our understanding of the physics governing the growth and development of ice lenses has changed, from a focus on the volumetric expansion of water driving weathering to the cryosuction-driven segregation ice growth mechanism (Walder and Hallet, 1985; Hallet et al., 1991). A quantitative understanding of this process has led to the development of numerical models that predict the relative intensity of periglacial activity that promotes sediment transport (Hales and Roering, 2007). Sediment budgets created within periglacial landscapes highlight the high rates of periglacial erosion, which represent up to 80% of the glacial sediment budget (Harbor and Warburton, 1993; O'Farrell et al., 2009). Recent use of shallow subsurface geophysics has provided a more detailed picture of the volumes of sediment created within glacial valleys (Sass and Wollny, 2001), again highlighting the rapid response of glacial valleys to deglaciation.

The largely empirical studies discussed above highlight the complex nature of the process response of valleys to both the development of a U-shape and its subsequent deglacial evolution. This may explain the dearth of models that attempt to include glacial and periglacial processes. In part, we lack the suite of geomorphic transport laws for modeling the extent of postglacial geomorphic response. The first attempt to model the formation of glacial valleys from fluvial ones assumed that glacial erosion was driven by Hallet's (1979) abrasion law that related erosion to ice velocity. The Harbor et al. (1988) model simulated the cross-sectional development of valleys, from V- to U-shaped, providing a numerical test of potential controls on the evolution of slope form and one of the first estimates of the timescale for valley development. Since Harbor et al. (1988), numerical modeling of glacial systems has increased in complexity, primarily through the coupling of glacial erosion models to geodynamic models with ice growth models of increased complexity (Tomkin and Braun, 2002). Abrasion is still the primary erosion mechanism in these models. Apart from a study by Dadson and Church (2005), no numerical studies have simulated postglacial development of valleys. Without sufficient geomorphic transport laws to describe paraglacial processes, Dadson and Church (2005) drove postglacial valley evolution using geomorphic transport laws developed in soil-mantled landscapes (stream power, non-linear hillslope diffusion) and a stochastic landslide model.

One of the key challenges in the development of models for bedrock hillslope evolution is the range of processes and substrates occurring on a single slope. Given Fisher–Lehmann's model as a simple example, erosion of the upper bedrock face is controlled by the weathering rate of the cliff (called weathering-limited by Carson and Kirkby, 1972), whereas erosion of the scree slope is controlled by a suite of processes that transport coarse sediment, including grain flow (a transport-limited slope; Carson and Kirkby, 1972). Modeling bedrock hillslope evolution was pioneered by Kirkby and Statham in the 1970s when both created transport laws for the movement of material on scree slopes. The absence of a suitable weathering law required the authors to treat retreat rate as a linear function of gradient above a critical threshold (Kirkby, 1984). More sophisticated transport laws and the inclusion of stochastic

transport processes (usually to simulate bedrock landsliding) are characteristic of more recent bedrock hillslope models (e.g., Dadson and Church, 2005). Typically, the transport laws used in these models do not differ considerably from those developed in soil-mantled landscapes, either due to the similarity of process (i.e., frost-driven creep vs. biologically driven creep) or from the lack of alternative erosion laws.

7.29.6 Soil-Mantled Landscapes

Soil-mantled hillslopes exist when erosion rates do not exceed the maximum rate of soil production for extended periods of time. Thus, the evolution and inheritance of soil-mantled terrain depend on the relative magnitude of climate-driven variations in soil transport and production. In contrast to bedrock-dominated hillslopes governed by stochastic processes (such as landslides) that commonly deliver debris directly to the channel network, soil-mantled hillslopes offer the opportunity to readily observe and study the active transport medium (i.e., soil column). In addition, the presence of a soil mantle enables hillslopes to absorb variations in transport rate without imparting a persistent morphologic signature. More generally, soil-mantled hillslopes worldwide exhibit characteristic convexo-planar forms despite the remarkably diverse range of processes responsible for generating and mobilizing the soil mantle.

7.29.6.1 Soil Production Mechanisms and Rates

Early work on soil production was highly dispersed as summarized by a review of soil production functions by Humphreys and Wilkinson et al. (2007). In documenting the churning activity of earthworms in his garden, Darwin (1881) recognized the important balance between production and removal in maintaining the soil mantle and emphasized the importance of biota in regulating soil chemistry. Working in the Henry Mountains of the Southwestern US, Gilbert (1877) famously reasoned that processes of rock decay (i.e., soil or debris production) are retarded by excessive accumulation of disintegrated rock and that a finite soil thickness may be required to capture rain and maximize the rate of rock decay. Later termed the 'humped' soil production function, this notion was largely ignored for decades until it was analyzed along with other soil production functions in the 1960s and 1970s. These studies posited that soil production rates either decline exponentially with depth or attain a maximum value for a finite soil depth (Ahnert, 1976). Carson and Kirkby (1972) formalized the notion that soils with thickness less than that associated with the maximum production rate may be unstable and thus poorly represented in landscapes. Subsequent studies used cosmogenic radionuclides to systematically calibrate soil production functions in diverse settings around the world (McKean et al., 1993; Heimsath et al., 1997, 1999, 2000, 2009; Small et al., 1999; Wilkinson and Humphreys, 2005). Evidence exists for both exponential and humped production models and these studies have not been synthesized or analyzed in a generalized fashion. In other words, given information on climate, vegetation, and rock type, we do not have the ability to predict rates of soil

production or the form of the soil production function. Notably, a recent modeling study (Gabet and Mudd, 2010) that implicates tree throw (or turnover) as a primary soil production mechanism uses forest ecosystem information to demonstrate that a humped production model similar to one empirically determined by Heimsath et al. (2001) may be applicable to forested landscapes.

The efficacy of soil production mechanisms (including bioturbation, frost cracking, hydration fracturing, crystal growth, etc.) strongly depends on rock properties (Dixon et al., 2009; Graham et al., 2010). Savigear (1960) indirectly identified the influence of rock properties on soil production rates in characterizing differing styles of slope evolution in West Africa. He observed that bedrock with closely spaced defects tended to exhibit a continuous debris mantle, facilitating evolution via slope decline. By contrast, similar areas with massive, unjointed bedrock lacked a significant debris mantle and tended to retreat via slope replacement. Although workers use rock mass strength data to explain hillslope morphology (Selby, 1980; Moon, 1984), we are not aware of studies that have systematically incorporated rock property data into the soil production function framework. Most soil production studies have been conducted in granitic or sedimentary bedrock settings and the rates generated via cosmogenic radionuclides generally do not exhibit significant variation with maximum production rates typically approaching 0.1–0.5 mm yr⁻¹ (McKean et al., 1993; Small et al., 1999; Wilkinson and Humphreys, 2005). Given the diversity of climate and vegetation at these study areas, this range of variation is surprisingly narrow. As a result, the morphologic legacy of climate-driven changes in soil production may not be readily apparent.

7.29.6.2 Soil Transport Rates

Rates of soil transport have been widely studied with varying success. In the 1960s and 1970s, numerous studies were performed to track human-timescale transport rates (Young, 1960; Kirkby, 1967; Schumm, 1967; Fleming and Johnson, 1975; Saunders and Young, 1983). Although some of these studies revealed slope-dependent soil movement for a given setting, others featured ambiguous (or uneven upslope) movement. These results reflect the challenges of characterizing highly stochastic processes over short timescales. Nonetheless, a prominent compilation of these studies by Saunders and Young (1983) purported to identify a climate control on rates of soil creep and surface wash. Their data suggested rapid creep rates for Mediterranean and tropical savannah settings and relatively low rates for temperate and semi-arid settings although the magnitude of these variations is less than a factor of 5. For wash processes, by contrast, their data suggested that lowering rates depend strongly on vegetation; in semi-arid settings, surface lowering is three orders of magnitude larger than in temperate regions. Subsumed within the data used to derive these patterns is the influence of vegetation, soil texture, and topography, confounding our ability to systematically access how these factors influence slope evolution. Despite more than 25 years of subsequent progress and refinement in our ability to estimate soil transport rates using cosmogenic radionuclides, ¹⁴C, optically stimulated luminescence, and

short-lived radionuclides, we have limited insights on how climate and vegetation regulate transport.

On many hillslopes, transport occurs in the absence of water and can thus be addressed using a slope-dependent (or diffusive) transport model whereby sediment flux varies as $q_s = KS$, where K is the transport coefficient (L² T⁻¹) and S is the slope gradient. This simple framework was originally inspired by Gilbert (1909) in order to explain the prevalence of convex hillslopes, and it also provides us with an opportunity to determine if K -values depend on climate or biologic variables. A compilation determined from diffusion studies on fault scarps indicates that K -values in semi-arid Basin and Range sites range from 0.1 to 2.0 m² ka⁻¹ and in forested sites along Lake Michigan and in the Oregon and California coastal K ranges from 3 to 12 m² ka⁻¹ (Hanks, 2000). Strictly interpreted, these results suggest that deeper rooted and more robust vegetation, as well as a reduced role for rainsplash transport may actually increase soil transport rates for a given hillslope. By characterizing the infilling rate of an unchanneled valley on the South Island of New Zealand, Hughes et al. (2009) demonstrated that the change from a grassland to forested ecosystem during the last glacial–interglacial transition instigated a near doubling of soil transport rates (and K -values) on gentle slopes not prone slope instability. This effect may be owed to the vigorous bioturbation associated with forested settings when compared to grassland regimes. Importantly, this interpretation may appear contrary to the traditional model posited for steep, slide-prone regions whereby dense vegetation adds cohesion to soils and subverts widespread transport and erosion. Instead, these results highlight the important role of topography in determining whether vegetation changes will have a more significant effect on impelling soil movement or stabilizing hillslope soils.

7.29.6.2.1 The modification of soil-mantled terrain during glacial–interglacial fluctuations

The question of whether hillslope form conveys more information about contemporary or past processes has been a contentious topic in geomorphology because the former interpretation de-emphasizes the value of modern process characterization (Chorley, 1964). Studies that quantify modern process rates and current landscape morphology characteristically note systematic relationships between process and form, implying that “slope form appears to determine contemporary geomorphic process” (Arnett, 1971). As such, the process–form linkage consists of substantial feedbacks rather than being a one-way, cause-and-effect relationship. However, if past climate states significantly modify topographic forms through a different suite of process rates and mechanisms, the interpretation of current process–form relationships becomes more challenging and complex (Parsons, 1988). Certainly, hillslopes formed through the operation of multiple processes further complicated this endeavor and, as a result, many research efforts have limited the scope and extent of their analyses in a quest for simplification. In one rather extreme example, Selby (1971) gathered climate and soil evidence to support his contention that more than 3 million years of relatively uniform climate conditions were responsible for the salt-driven (rather than frost processes) evolution of his

Antarctic hillslope study area. A similar situation may occur in the Atacama Desert of South America where it has recently been suggested that salt-driven hillslope modification over million-year timescales has generated broadly convex slope forms (Owen et al., 2011). The increasing availability of climate records may aid efforts to relate hillslope morphology to specific climate regimes. However, these examples are exceptional in that most of the Earth's surface experienced profound environmental changes during glacial–interglacial cycles in the last several million years, commonly with dramatic morphologic implications.

In the Southwestern US, Ahnert (1960) noted the prevalence of cliff-top convexities and suggested that these features may arise during pluvial epochs during which cold, wet

conditions prevailed and lakes dominated the region. He also noted that abundant landslide deposits that may have clustered exposure ages suggest a climate-related initiation mechanism (Ahnert, 1960). Everard (1963) worked in Cyprus and associated contemporary flatiron slopes with the dissection of former widely spaced convexo-concave hillslopes formed by creep during wet pluvial episodes (Figure 5). Everard recognized the combined influence of base level, climate, and vegetation on slope morphology and attempted to reconstruct paleo-topography using reworked fanglomerate deposits as evidence for the extent of the former active, creeping layer. Cotton (1958) also interpreted “smoothed, coarse-textured, whalebacked, relatively featureless relief” as a relict feature of efficient periglacial processes active during glacial advances and

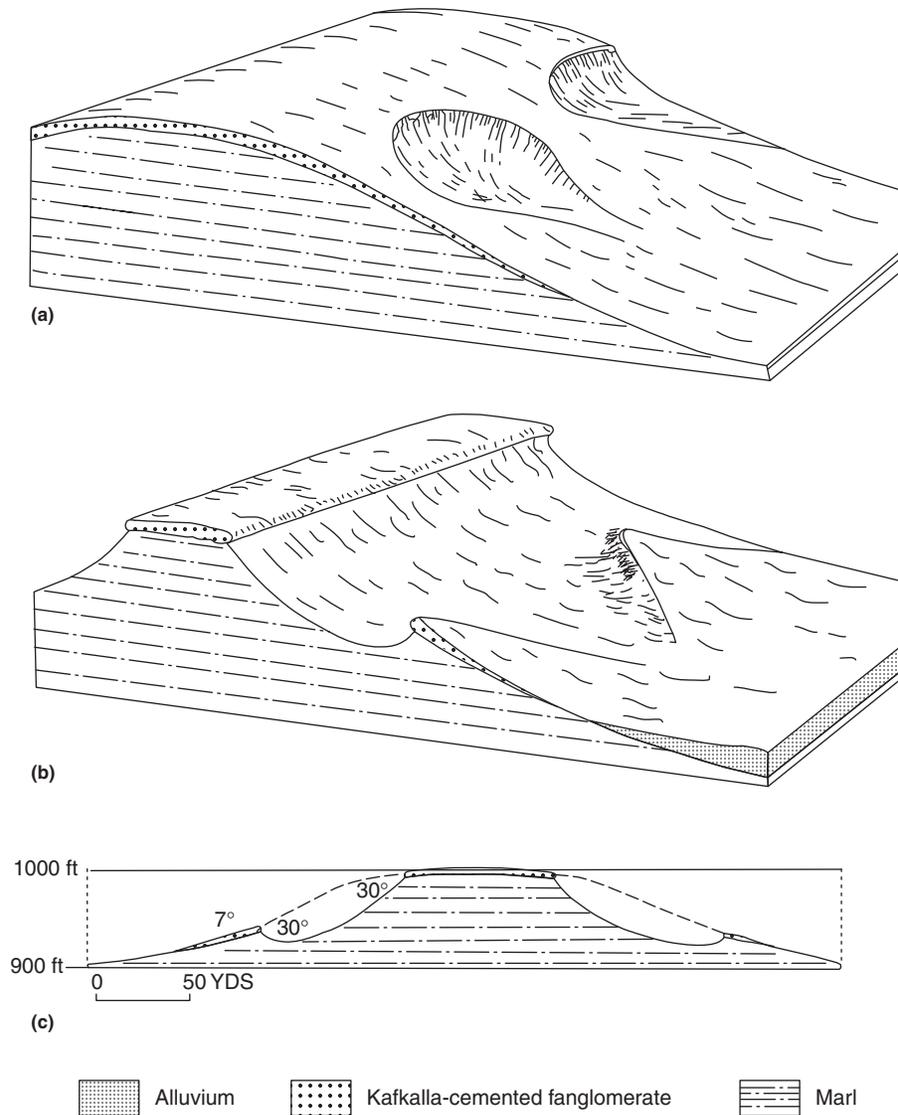


Figure 5 Schematic showing the formation of flatirons on the island of Cyprus due to glacial–interglacial transitions (Reproduced from Everard, C.E., 1963. Contrasts in the form and evolution of hill-side slopes in Central Cyprus. *Institute of British Geographers, Transactions* 32, 31–47). The broad, convex slopes of the glacial episodes are generated through pervasive soil creep of a mobile regolith. Interglacial conditions promote dissection and an increase in drainage density. (a) Gullies cutting into the flanks of concavo-convex ridges, which have been mantled with kafkalla-cemented fanglomerate detritus. (b) The gully heads have joined leaving isolated, undissected remnants of the former slopes (the flatirons) rising above the new hillside. This rapid erosion has aggraded the valley floor. (c) Sketch cross-profile of mesa 975625.

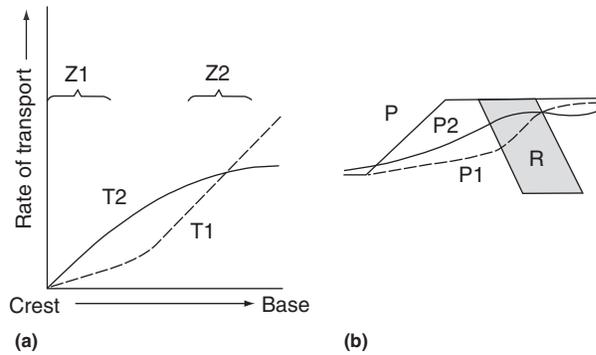


Figure 6 Conceptual model for glacial to interglacial changes in process (sediment transport) intensity and their morphologic implications proposed by Starkel (1964) and modified from Young, A., 1972. Slopes. *Geomorphology Texts*, 3. Oliver and Boyd, Edinburgh, 288 pp. T1 and T2 are transport curves for interglacial and glacial conditions, respectively. P1 and P2 are the associated morphologic profiles wherein R represents an outcrop of resistant bedrock.

noted the paucity of these paleo-features in New Zealand relative to Western Europe. In Wales, a similar interpretation has been offered in view of extensive fractured regoliths reflecting frost-shattering processes and rapid transport during the LGM (Young, 1958) that generated thick colluvial concave footslope deposits and broad hillcrests (Starkel, 1964). Starkel (1964) also attempted to associate different sediment-transport regimes with glacial and interglacial periods and analyze the morphologic implications (Figure 6). More recent efforts in the European Alps have used digital elevation data and erosion rate estimates to explore the timescale required to transform glaciated surfaces into ridge-valley-dominated drainage networks with convex soil-mantled hillslopes (Schlunegger et al., 2002; Schlunegger and Hinderer, 2003).

Chorley (1964) questioned the feasibility of associating slope forms to different climate and process regimes by pointing out the nonuniqueness of process-form constructs. Suggesting that slope studies are “subject to preconception,” Chorley concluded that “it is patently apparent that no distinctive slope form is uniquely linked to a given climatic or tectonic erosional environment.” This notion has since been eloquently demonstrated by Dunne (1991). In 1964, Chorley did however acknowledge the unrealized utility of mathematical models for interpreting slope-forming processes and distinguishing between the historical hangover of inherited forms and the dynamic equilibrium of contemporary process-form feedbacks.

7.29.6.3 Process-Based Models and Inheritance Timescales

Quantitative, process-based geomorphic models and erosion rate data sets that emerged in the 1960s and beyond provided an opportunity to objectively determine the timescale of hillslope response relative to climate fluctuations. In the simplest case, Parsons (1988) used typical erosion rates compiled by Saunders and Young (1983) to calculate that 50–500 ky are required to reduce a 100-m hillslope by half. Kirkby’s (1971) early process-response models indicated that similar timescales are required to attain characteristic hillslope

forms. Thus, it seems likely that most hillslopes owe some fraction of their form to past climate regimes, although differentiating changes from base-level lowering and climate-driven hillslope processes is not trivial (Summerfield, 1975; Armstrong, 1980; Trofimov and Moskovkin, 1984). The conceptual and quantitative framework for defining equilibrium and assessing adjustment timescales has been extensively addressed with some notably well-written analyses contributed by Howard (1988), Mayer (1992), Ahnert (1994), and Whipple (2001).

The extent of the inheritance likely depends on the magnitude and style of the associated change in process. To systematically address this question, several numerical modeling efforts have attempted to further refine hillslope adjustment timescales by applying and testing emerging hillslope transport models (i.e., equations for q_s in eqn [1b]). Ahnert (1976) studied the timescale for slope adjustment, moving beyond the interpretation of steady-state characteristic forms. Working on dissected marine terraces in Papua New Guinea, Chappell (1978) examined “the extent to which current landforms can be explained in the light of process studies” and, in doing so, questioned his own previous findings due to issues of process uniqueness. Koons (1989) used macro-diffusion to model mountain range evolution in the Southern Alps, New Zealand, accounting for climate-driven variation in hillslope transport rates and noting that “the erosional response time of any system is much greater than the time scale of climate variation.” Seeking to quantify the spatial extent of terrain subject to gelifluction under different climate regimes, Kirkby (1995) used a process model to demonstrate that the survival time of a feature of wavelength, L , is proportional to L^2/K , where K is the soil transport coefficient discussed above. Rinaldo et al. (1995) modeled complex changes in valley density through glacial–interglacial cycles and suggested that geomorphic signatures of past climates are less likely to be apparent in areas experiencing active uplift. Given observed K estimates, this suggests that small features ($L = \sim 1$ m) are erased over decadal timescales, whereas larger features such as entire hillslopes may require 10^4 – 10^6 years for morphologic adjustment. Fernandes and Dietrich (1997) used a similar framework to model hillslope adjustment to varying rates of base-level lowering and K -values; their slope relaxation times vary from 10^5 to 10^6 years. As a result, they argued that landscapes (and specifically hilltop convexities) likely reflect a legacy of past climate-driven processes in their morphology. This interpretation implies that climate fluctuations have a significant effect on rates of sediment transport.

By adapting a nonlinear transport model (whereby values of K increase rapidly as slopes approach a critical value), Roering et al. (2001) calculated significantly reduced adjustment timescales of order 4×10^4 years for parameter values characteristic of the Oregon Coast Range. Mudd and Furbish (2007) further refined the analytical framework for quantifying adjustment timescales and estimated hillslope adjustment timescales using a depth-dependent transport model. By contrast, their results suggest that longer timescales are required for slopes governed by depth-dependent transport than if rates vary with slope alone. Mudd and Furbish (2007) also proposed a framework for identifying the topographic signature of transient hillslope adjustment due to changes in

base-level lowering. A similar analysis should be possible, given changes in climate-driven erosional processes. Working at the sub-hillslope scale, Strudley et al. (2006) used a process model to conclude that tors may be the legacy of ancient landscapes. Most generally, these studies show that nonlinearities in the soil transport model can elicit order-of-magnitude effects on adjustment timescales, highlighting the need to better test and constrain existing transport formulations (Dietrich et al., 2003). Most importantly, these soil transport models highlight that adjustment timescales vary with the square of hillslope length such that gentle, broadly spaced slopes are most likely to include significant inheritance, whereas short, steep slopes are more likely to exhibit erosion rates and morphology tuned to the contemporary climate regime. Badlands are an extreme example of the latter.

Models that include transport due to landsliding are less common, primarily because process laws for slope instability are challenging to implement in landscape evolution models. Using thresholds of rock slope stability, Densmore et al. (1998) performed simulations of mountain front evolution and predicted relatively rapid morphologic change due to high rates of base-level lowering and threshold-driven transport processes. Tucker and Bras (2000) modeled the effect of two different rainfall regimes on slope failure and the evolution of upland drainage networks. Despite having the same annual rainfall, their results demonstrated that storm intensity and duration strongly modulated local relief and valley density. Explicit models of ecosystem feedbacks with hydrologic and geomorphic processes are beginning to quantify slope morphology associated with specific climate regimes; continued testing and calibration of these models should enhance our ability to interpret the topographic signature of past environments (Kirkby, 1989; Coulthard et al., 2000; Istanbuloglu and Bras, 2005; Collins and Bras, 2010).

7.29.7 Discussion and Conclusions

The role of climate in modulating hillslope form appears set to experience a revival in interest. Although there was considerable conviction in the inheritance of past climates in modern landforms, early efforts at placing climate in the context of slope studies were hampered by the lack of erosion rate information and process-based arguments. The advent of cosmogenic radionuclides for erosion rate estimation has invigorated efforts to revisit and test hillslope evolution theory (e.g., McKean et al., 1993; Heimsath et al., 1997; Bierman et al., 2001) and should continue to improve our ability to improve and formulate process models. The reemergence of dynamic equilibrium (or steady state) for interpreting landscapes firmly entrenched the view that contemporary processes and forms were strongly coupled, which tended to minimize the role of inheritance in slope interpretations. Perhaps more accurately, study areas were commonly selected in order to minimize the role of climate inheritance. Continued refinement of paleo-environmental records, even in areas once thought to be relatively free from major changes during glacial-interglacial transitions, for example, will force the geomorphic community to confront how changing process mechanisms and rates manifest in landscape form. Fortunately,

the process-based heritage of the past several decades provides a framework for tackling this problem in a systematic fashion.

Perhaps one of the profound difficulties we face is how to translate climate proxies into geomorphically meaningful information. Although paleoecology studies provide remarkably detailed constraints on vegetation change in various settings, how can we decode this information in quantitative constraints on storm intensity, bioturbation, or other factors driving hillslope response? As an example, recent numerical modeling of hillslope evolution in the Oregon Coast Range used a nonlinear depth-dependent and slope-dependent soil transport model to compare modeled hillslopes with current topography from airborne LiDAR data (LiDAR, light detection and ranging; Roering, 2008). The model was calibrated and tested using measured erosion rates as well as root density data of Douglas fir (which is the dominant species) to inform the transport model. Although the coarse hillslope properties are reproduced with reasonable fidelity, the calibrated model cannot account for the sharpness of modern hilltops. Does this imply a failure of the model or does it reflect the legacy of previous climate (i.e., vegetation) regimes? Prior to Douglas fir colonization nearly 13 000 years ago, the region was characterized by dry, open, parkland forests (Worona and Whitlock, 1995). Perhaps rates of bioturbation and soil transport were low during this pre-Holocene regime such that hilltops required greater convexity in order to match imposed base-level lowering and these sharp convexities persist today. To this effect, Chorley (1964) stated "...all slopes possess, in highly variable proportions, both the aspects of 'historical hangovers' and 'dynamic equilibrium', and one of the most pressing needs of current slope studies is that the co-existence of these attributes should be recognized and their relative importance evaluated." Ahnert (1994) expounded on the problem of slope inheritance and examined slope relaxation times across varying scales.

Given continued interest in modeling orogen-scale evolution to explore feedbacks between erosion and tectonic forcing, actual hillslopes in many simulations have been statistically embedded within large (> 500 m) grid cells. As such, the details of hillslope form are secondary to their fundamental traits, particularly relief and average gradient. Given this simplification, these efforts seek a straightforward linkage between slope and erosion rate. For a particular process formulation, one might expect a set of characteristic slopes to manifest (Strahler, 1950). Working in steep, tectonically active hillslopes, Strahler (1950) showed that although no mean slope angle dominated regionally, hillslopes within certain locales tend to exhibit slope angles with a symmetrical distribution and low dispersion (Figure 7). These statistical properties seemed to vary as a function of base-level lowering, as well as lithology, climate, and vegetation cover (Strahler, 1950). Hillslope angles may also reflect the engineering properties of the soil mantle; the consistency between soil friction angles and slope gradient thus lends support to the contention that slopes are adjusted to a limiting or threshold slope, an idea conceptualized in the nineteenth century (de la Noe and de Margerie, 1888) (Figure 8) and rooted in the observations of Bryan (1922) working in southern Arizona and later elaborated by Young (1961) and Carson and Petley (1970). This leads to a nonlinear relationship between sediment flux and slope close to this threshold angle (Anderson and Humphrey, 1989; Howard, 1994; Gabet, 2000;

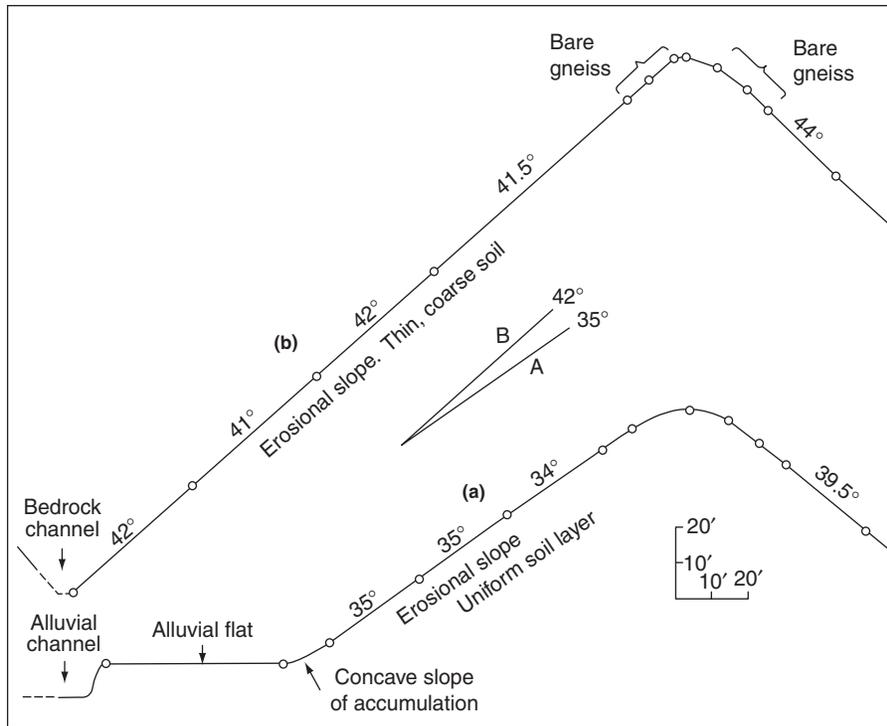


Figure 7 Hillslope profiles surveyed in the Verdugo Hills by Strahler, A.N., 1950. Equilibrium theory of erosional slopes approached by frequency distribution analysis. *American Journal of Science* 248, 673–696, showing the impact of base-level lowering on hillslope angle. Persistent channel–hillslope interaction (and incision) may promote the maintenance of steep slopes. (a) Steep hillslope profile subject to direct base-level lowering. (b) Gentle hillslope profile buffered from base-level lowering by broad alluvial plain.

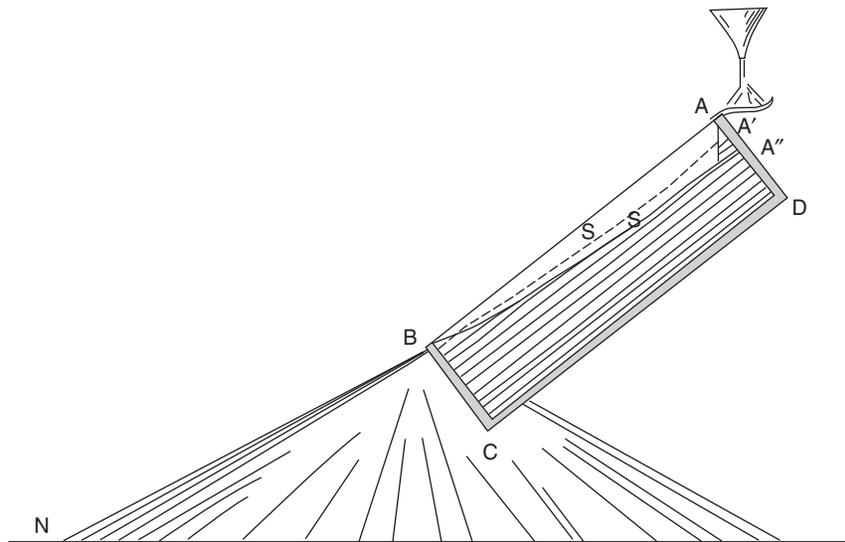


Figure 8 Depiction of threshold slope maintenance (by way of the sandpile metaphor). Reproduced from de la Noe, G., de Margerie, E., 1888. *Les Formes du Terrain*. Service Geographique de l'Armee. Imprimerie Nationale, Paris, 205 pp.

Roering et al., 2007), although the parameters defining the nonlinear curve should be anything but universal and instead vary with material and climate properties. By representing key hillslope traits with process-based relationships, hillslope relief can be successfully integrated with larger-scale orogen models. These efforts, however, commonly require that the typical hillslope length

(horizontally measured) be specified, which may change with climate parameters. This motivates the need to more clearly define how drainage density and slope angle vary with climate and erosion rate.

The comparison between soil-mantled and bedrock landscapes reveals our lack of physically based geomorphic transport laws for the suite of processes that sculpt bedrock

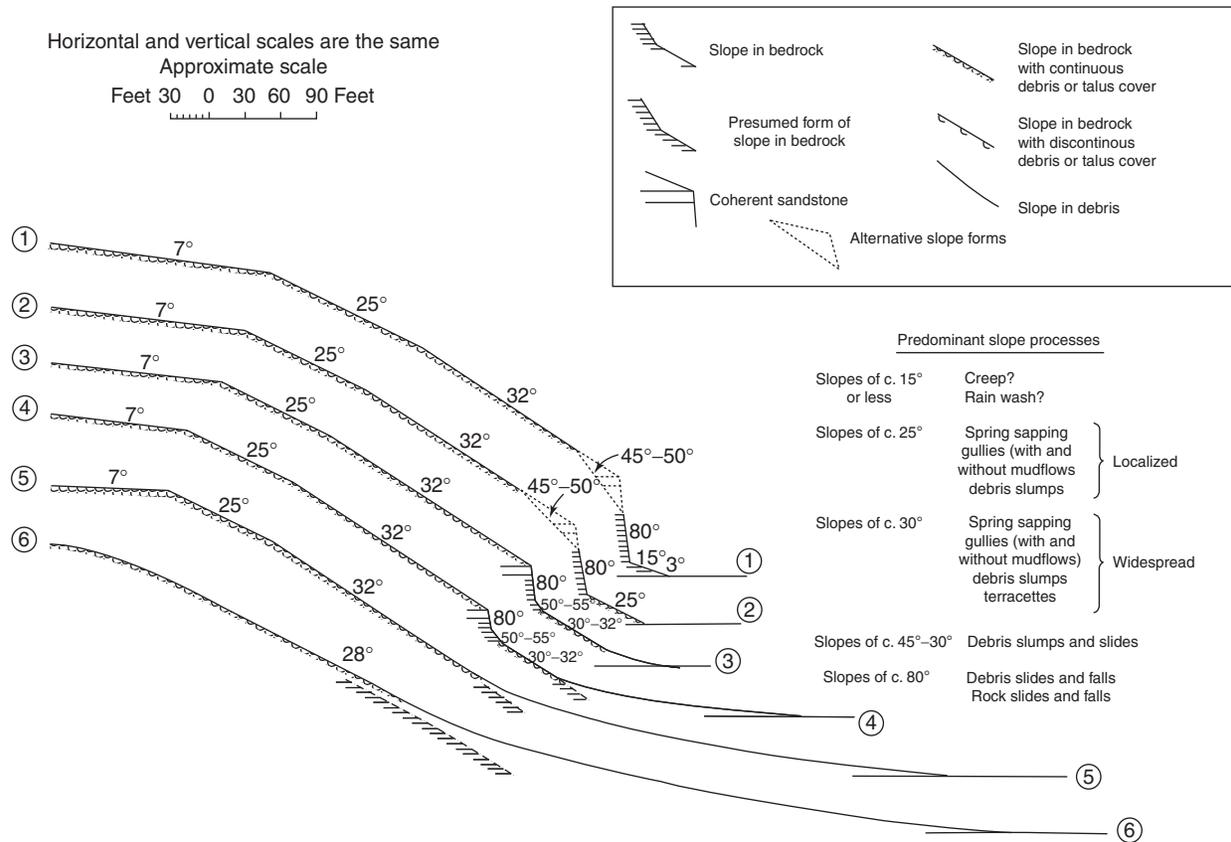


Figure 9 Summary diagram of hillslope profiles surveyed and analyzed by Savigear, R.A.G., 1952. Some observations on slope development in South Wales, Transactions and Papers (Institute of British Geographers) 18, 31-51, showing how progressive hillslope isolation from coastal processes promotes the relaxation and rounding of slope profiles.

landscapes. We still lack a quantitative framework for translating our understanding of mechanical weathering phenomena occurring on bedrock slopes into models that predict fluxes. In particular, rates of scree production are suggested to depend strongly on frost processes (Anderson, 1998; Hales and Roering, 2005, 2007; Delunel et al., 2010); yet, we have limited ability to predict production rates given constraints on climate variables and rock properties. In bedrock landscapes process laws have been developed for glacial erosion (Hallet, 1979; Harbor et al., 1988), but periglacial and arid weathering and transport processes have been neglected in this respect despite empirical evidence for process-form linkages similar to soil-mantled landscapes (the exception is the formation of rounded hillslopes via freeze-thaw-driven creep).

Slope development has been quantified in favorable settings where initial conditions are known and spatial trends serve as a surrogate for stages in temporal evolution (e.g., Savigear, 1952; Kirkbride and Matthews, 1997; Hales and Roering, 2005, 2007; Hilley and Arrowsmith, 2008). These studies provide an opportunity to document progressive degradation of the initial landform (e.g., marine terrace) assuming constant erosional forcing. Perhaps the first such study was performed in West Wales using slope profiles for cliffs that had been progressively cut off from active coastal erosion by the advance of a sand spit (Savigear, 1952). The useful geometry showed the progressive formation of a concavo-convex slope system from one that was initially convex

upward (Figure 9). Savigear's study of retreating cliffs provided the basis for the testing of an early process-form model (Kirkby, 1984). Savigear argued that where slopes were subjected to active erosion at their base they evolved through parallel retreat that created steep, linear slopes. When these slopes were cut off from basal erosion, they developed a convex basal slope below the declining rectilinear central slope. All of his slope profiles had a convex upper slope that represented erosion of the upper horizontal surface (Kirkby, 1984). Since Savigear's study, relatively few space-for-time studies have been performed, perhaps because each area has a specific (and thus nongeneralizable) lithologic, tectonic, or climatic setting. In addition, the role of climate-driven variability in erosional processes is seldom incorporated in these analyses. Nonetheless, the utility of well-constrained field experiments for developing and testing landscape evolution is invaluable, particularly if climate fluctuations can be thoughtfully incorporated.

Opportunities to improve our understanding of hillslope evolution and inheritance abound because of recent technological advances. Quantification of hillslope form has advanced little despite the abundance of digital data now available, including vast areas with airborne LiDAR coverage. Many LiDAR studies continue to use average slope angles for data analysis, although the novel resolution often incorporates small-scale features that are not relevant to long-term geomorphic development. As a result, systematic

methods (e.g., smoothing) for mapping topographic patterns relevant long-term trends are required, yet are commonly performed in an arbitrary fashion. Additionally, metrics to analyze the two- and three-dimensional structures of hillslopes have not advanced despite the ubiquity of LiDAR data. Only recently have spectral and flow algorithm approaches been proposed to quantify the typical length scale of hillslopes (Tucker et al., 2001; Roering et al., 2007; Perron et al., 2008). Exciting opportunities have emerged for measuring rates of slope change. Although cosmogenic radionuclides enable millennial-scale erosion rate data in most settings, spatially and temporally dense measurements of slope change have been accomplished in landslide-prone areas where deformation rates are perceptible on short timescales (Hilley et al., 2004; Roering et al., 2009). Future advances in remote-sensing capabilities may enable researchers to map the stochastic nature of slope-forming processes, such as tree turnover, dry ravel, mammal burrowing, etc., and directly test slope evolution models. Furthermore, the exploitation of cosmogenic radionuclides in well-chosen depositional settings should allow for temporal (and climate-driven) variations in erosion processes to be measured for direct testing of slope evolution and inheritance.

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Biographical Sketch



Dr. Tristram (T.C.) Hales received his PhD in geology from the University of Oregon in 2007. He worked for a short period of time as a postdoctoral fellow at the University of North Carolina before accepting his current position as a lecturer in geomorphology at Cardiff University. Dr. Hales is interested in how landscapes evolve in the presence of a wide range of geodynamic and Earth surface processes, particularly debris flows, soil creep, and periglacial processes.



Dr. Josh Roering received his BS and MS in geological and environmental sciences at Stanford University in 1995 and his PhD in geology from the University of California, Berkeley in 2000. He was a postdoctoral fellow at the University of Canterbury in New Zealand before accepting a faculty position at the University of Oregon in 2001. Dr. Roering studies hillslope processes and the evolution landscapes using laboratory, numerical, and field studies. He is particularly intrigued by the role of biota in modifying the Earth's surface.