Review

Hillslope soils and vegetation

Ronald Amundson [a,*], Arjun Heimsath [b], Justine Owen [a], Kyungsoo Yoo [c], William E. Dietrich [d]

[a] Department of Environmental Science, Policy, and Management, 130 Mulford Hall, University of California, Berkeley, CA 94720, USA
[b] School of Earth and Space Exploration, Arizona State University, Tempe, AZ 85287, USA
[c] Department of Soil, Water, and Climate, University of Minnesota, 439 Borlaug Hall, 1991 Upper Buford Circle, St. Paul, MN 55108, USA
[d] Department of Earth and Planetary Science, 307 McCone Hall, University of California, Berkeley, CA 94720, USA

ARTICLE INFO

ABSTRACT

Assessing how vegetation controls hillslope soil processes is a challenging problem, as few abiotic landscapes exist as observational controls. Here we identify five avenues to examine how actively eroding hillslope soils and processes would differ without vegetation, and we explore some potential feedbacks that may result in landscape resilience on vegetated hillslopes. The various approaches suggest that a plant-free world would be characterized by largely soil-free hillslopes, that plants may control the maximum thickness of soils on slopes, that vegetated landforms erode at rates about one order of magnitude faster than plant-free outcrops in comparable settings, and that vegetated hillslope soils generally maintain long residence times such that both N and P sufficiency for ecosystems is the norm. We conclude that quantitatively parameterizing biota within process-based hillslope models needs to be a priority in order to project how human activity may further impact the soil mantle.

© 2015 Published by Elsevier B.V.

1. Introduction

'Over nearly the whole of the earth's surface there is a soil' (Gilbert, 1877). Yet, why does a soil mantle occur so pervasively on a tectonically active planet, with variable topography, an active hydrological cycle, and oscillating climate conditions? Here on Earth gravity, assisted by water, is a pervasive force for causing loosened material mantling solid rock to travel downhill (Culling, 1963). Gilbert suggested that 'the general effect of vegetation is to retard erosion; and since the direct effect of rainfall is the acceleration of erosion, it results that its direct and indirect tendencies are in the opposite directions.' Thus, it may be hypothesized that if land plants had not evolved, the common experience of soil-mantled uplands might be an exception on Earth rather than the rule.

While this hypothesis seems reasonable, testing it is not a trivial exercise. The face of the Earth is covered by life, and thus nature provides few lifeless landscape *controls* to which plant-mantled land surfaces can be directly compared. What controls the thickness of upland soils? What processes control their fertility? More fundamentally, do plant–soil interactions respond in ways that optimize conditions for plant-based ecosystems? About 80 years before Gilbert, James Hutton had deciphered an outline of the production and removal processes that maintain hillslope soils and suggested that they are balanced in a way 'so contrived that nothing is wanting … for the pleasure and propagation of created beings' (Hutton, 1795). Stated somewhat differently,
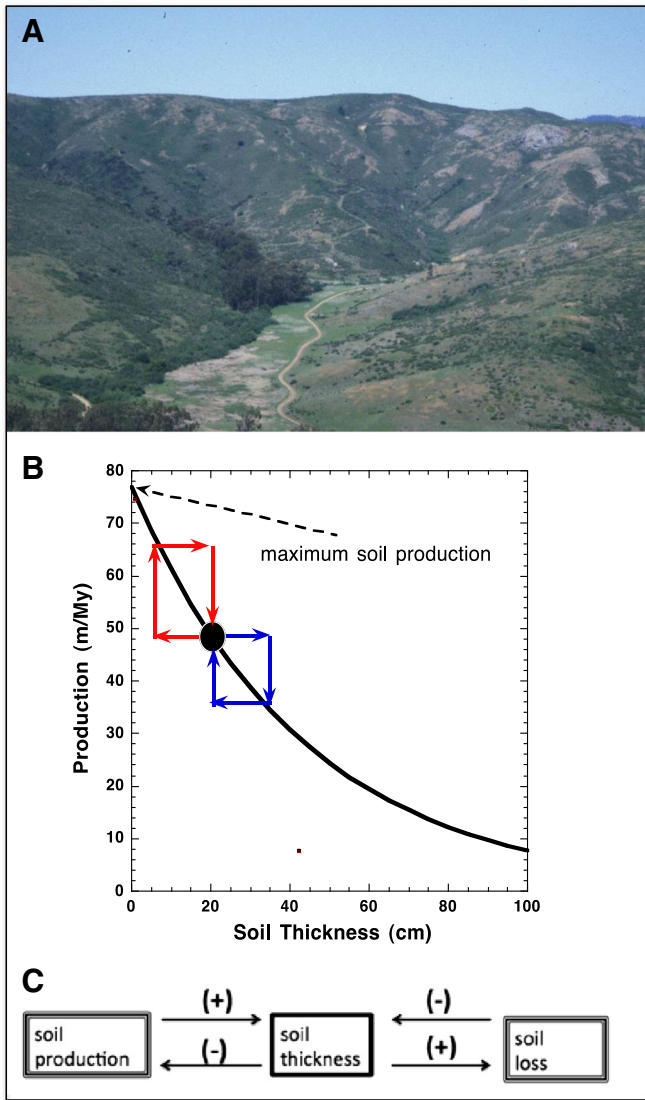
for Hutton, hillslope soils exist *for* life (Gould, 1987), rather than because of it. While Hutton had a unique set of philosophical and historical constraints for arriving at this hypothesis (Montgomery, 2012), the contrast between Gilbert and Hutton's ideas underscore that we know little about the feedbacks that must exist between abiotic and biotic processes on soil-mantled hillslopes. This theme has been the basis for a recent special issue of *Geomorphology* (Hession et al., 2010) and is a topic of ever-growing interest among earth scientists.

Plants are arguably the key component of the biota on landscapes, and how different would hillslopes be without their influence and feedbacks? In this paper, we review and interpret five different approaches that help us evaluate the effect of plants on hillslope soils, and from these analyses arrive at a few tentative ways that hillslope soils and plants interact. While we can only begin to perceive the outlines of these processes and feedbacks; an additional motivation here is to articulate reasons why we may wish to develop new methods to clarify the nature of biotic/abiotic couplings on an increasingly human-dominated and -managed planet.

2. The dynamics of hillslope soils

Soil mantled hillslopes are the setting for vast areas of the Earth's forests and grazing land. Of these landscapes, the gentle convex-up segments have received the most research attention and are considered here. The pace at which these landscapes evolve is ultimately dictated by local base-level changes, which are driven by tectonics on regional scales over geological time. A landscape typical of soil-mantled hillslopes is illustrated in Fig. 1A.

definition used here is particularly relevant to plants, for this is the component of soil that is in direct physical and chemical interaction with plants and their roots and with the associated organisms (insects to mammals) that exist because of the plants.

Mobile soil thickness on slopes is the balance between production and erosion. Erosion is the divergence of soil flux, which is facilitated by mechanisms that move particles diffusively down slope (Fernandes and Dietrich, 1997; Roering et al., 1999; Heimsath et al., 2005). The soil removed is replaced by soil production— the physical disruption of the underlying bedrock or saprolite and its emplacement in the soil column. If soil thickness is time invariant, soil production can alternately be viewed as landscape denudation. The time-dependent mathematical formulation of this situation is (Dietrich et al., 1995)
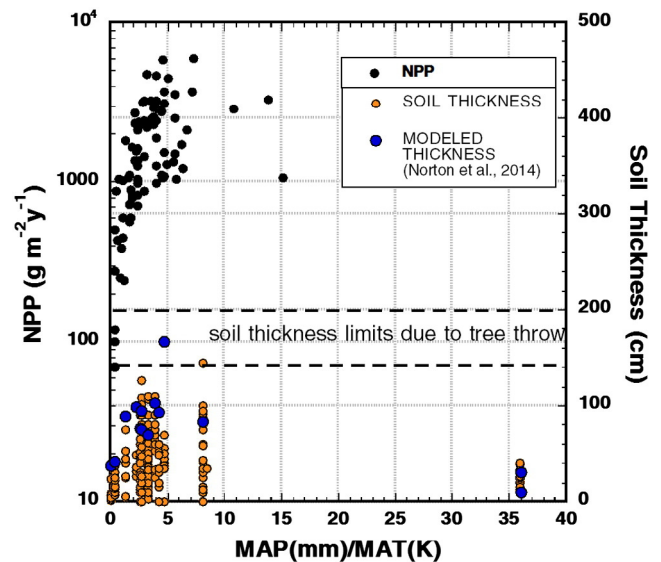
$$\underbrace{\frac{dH}{dt}}_{\substack{\text{change} \\ \text{in} \\ \text{soil} \\ \text{thickness}}} = \underbrace{P_s}_{\substack{\text{soil} \\ \text{production}}} - \underbrace{E}_{\substack{\text{soil} \\ \text{erosion}}} \tag{1}$$

where H = soil thickness (L), t = time, P_s = soil production rate (conversion of rock/sediment to mobile soil) (L / T), and E = erosion rate (L / T). Soil production also includes atmospheric inputs of dust and salt (Owen et al., 2010), which are significant mainly in arid conditions. *Soil production* is not the same as *soil formation*. Soil production refers to the conversion of rock or saprolite into mobile soil. Soil formation is a far more complex set of processes that includes weathering advances through the mobile and immobile materials in the profile, transfers of organic matter and clay downward, physical and biological mixing.

The functions that describe the production and erosion of soils are becoming more widely understood after the advent of cosmogenic nuclide-based methods of determining soil production functions (Heimsath et al., 1997). Here, we illustrate two common forms of the functions: soil thickness-dependent production (Heimsath et al., 1997):

$$P_s = P_o e^{(-aH)} \tag{2}$$

Fig. 1. (A) A view of an example of a gentle, soil-covered landscape with significant convex-up components characterized by hillslope ridges and noses. View to the NW in Tennessee Valley, Marin County, CA. Note that the focus of this paper is on the nature of the convex-up regions of the landscape, while the hollows and floodplains in the photograph are driven by somewhat differing processes. (B) A graphical illustration of the P term in Eq. 1, using the soil production function from Tennessee Valley, CA by Heimsath et al. (2005). A curvature of 0.01 m^{-1} is used. The black circle represents the steady-state soil thickness. A change in soil thickness with time that thickens (blue line) or thins (red line) a soil from steady state (resulting from a hypothetical perturbation) results in a respective decrease or increase in rates of soil production. The graph suggests an increasing rate of soil production the further soil thickness trends from steady-state values (inspired by Carson and Kirkby, 1972). (C) A simple feedback loop model of Tennessee Valley, CA, using soil thickness-dependent production and loss laws. The figure illustrates that soil thickness, in some landscapes, is the balance between two processes with negative feedbacks. Soil production vs. soil thickness is an overall negative feedback, as is (for soil thickness-dependent transport) soil thickness vs. diffusive soil loss.

There are two scientific definitions of *soil* (see Yoo and Mudd, 2008). Here, we use the geomorphic definition, where soil is viewed as the mobile portion of the weathering profile, as a material that no longer retains the fabric of the parent rock or sediment. In many locations, this commonly restricts *soil* to the A horizon or biologically mixed portion of a soil profile. Pedologists and geochemists view soil as the vertical weathering profile—one that includes the mobile geomorphic *soil*, but also extends into highly chemically weathered material that may still contain remnants of rock or sediment structure (and which is certainly not mobile) (Jenny, 1941; Yoo and Mudd, 2008). The geomorphic

Fig. 2. The changes in global NPP (Lieth, 1973; Sanderman et al., 2003) with climate (the MAP (mm)/MAT (K) ratio). The measured soil thicknesses for sites in Table 1 are plotted, but show no significant trend with NPP or climate but do seem to have a maximum thickness, one consistent with the maximum depths of tree throw discussed by Roering et al. (2010). The predicted soil thickness, a balance between SPR and denudation, calculated from Norton et al. (2014), are also illustrated and show a very good relationship to observed values.

where P_o = maximum soil production at a site (L / T), α = a constant (1 / L)

$$E = \vec{\nabla} \cdot \vec{q_s} \qquad (3)$$

where $\vec{\nabla} \cdot$ = vector differential operator, $\vec{q_s}$ = sediment flux (L³ / LT), The values of the various parameters in the functions are site dependent and reflect the integrated effects of rock composition, climate, and vegetation. The nature of the laws that control production and transport has profound impacts on the stability of the hillslope soil system. For example, Gilbert (1877) speculated that soil production (Eq. 2) may, instead, be a *humped* function with soil thickness (i.e. maximum soil production rate occurs at some shallow soil thickness) (e.g., Cox, 1980):

$$P_s = P_o e^{(-bH)}(1 + cH) \qquad (4)$$

where b is a scaling factor for the decrease in soil production with depth. If $c = 0$, the model is equivalent to Eq. (2). If $c / b > 1$, the relationship is humped and P_s reaches a maximum at thickness $(c - b) / bc$. Analyses of this situation (Carson and Kirkby, 1972; Dietrich et al., 1995; D'Odorico, 2000; Norton et al., 2014) show that such a law leads to an inherently unstable system at shallow soil thicknesses, with a bimodal landscape of soil and bare rock. If the soil thickness of the peak production value is small (i.e., less than say 10 cm) the two functions may seem quite similar overall, but for the important difference that exposed bedrock in the case of the humped production function is expected to shed particles much slower than when buried. Whether this proposed system instability exists extensively in nature is still widely debated (Heimsath et al., 2009), but observational evidence remains ambiguous as for which production functions may dominate at a particular location. An exponential relationship between soil thickness and production rate (Eq. 2), illustrated in Fig. 1B using model parameters is derived from
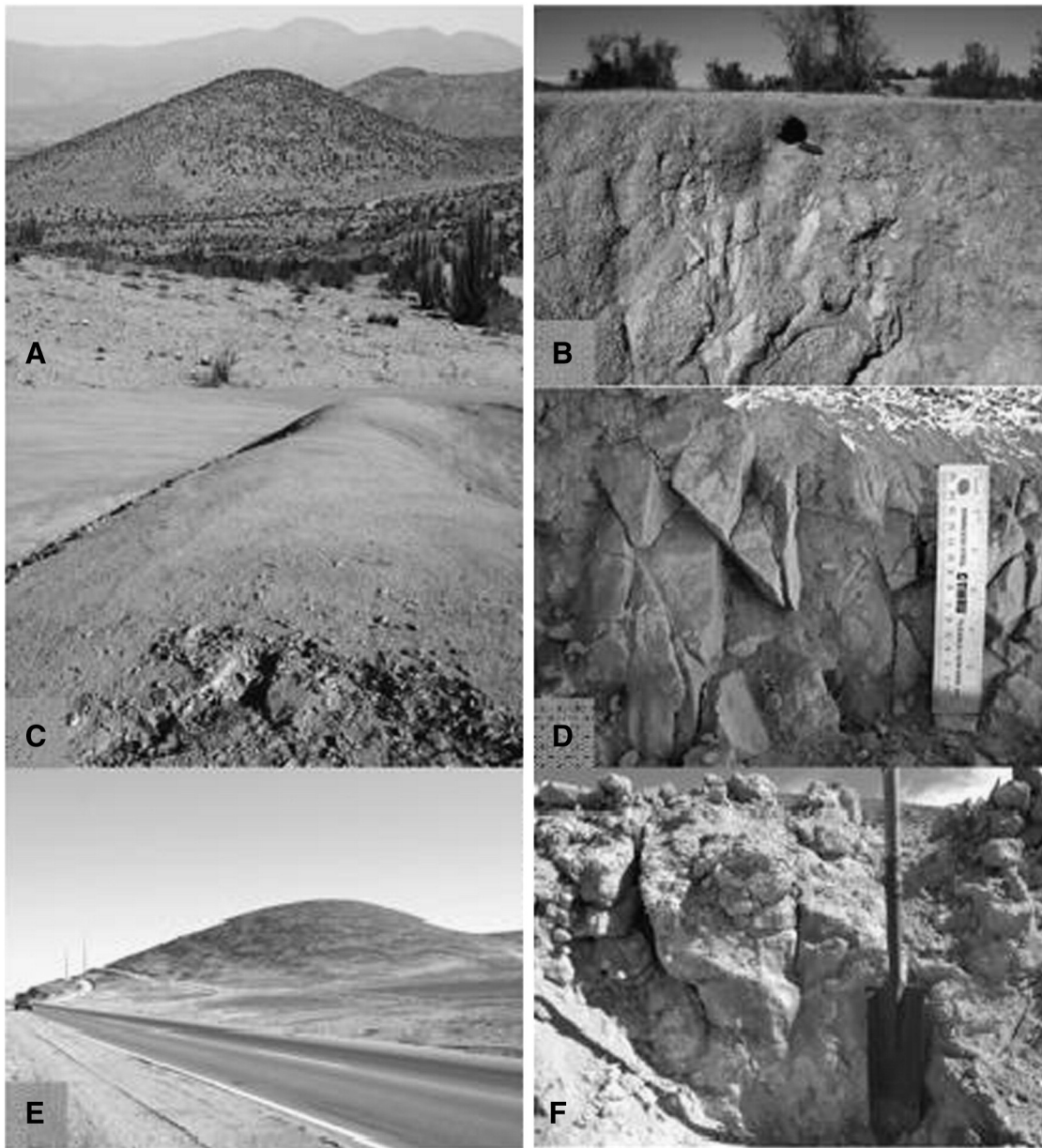
Fig. 3. Hillslope shapes and soil cover along a S–N rainfall gradient. (A) and (B) hillslope shape and soil cover at 100 mm MAP, (C) and (D) at 10 mm MAP, and (E) and (F) at 1 mm MAP. In (B), the soil is a thin (~30 cm—note the animal burrow in the middle of the photo) sandy soil over somewhat weathered granitic bedrock/saprolite. In (D), where vegetation is largely absent, an ~1 cm-thick layer of soil material overlies fresh bedrock. The apparent mechanism by which this rock is converted to soil is through chemical alteration of the uppermost rock fragments by rare rainfall combined with salt. (E) Shows a soil that reflects a mix of atmospherically derived sulfate and other salts and dust and bedrock fragments heaved into the profile by salt shrink/swell.

studies in Tennessee Valley, CA (Heimsath et al., 1997, 2005), which suggests system stability, and negative feedbacks to production such that soil thinning and thickening lead to acceleration (or deacceleration) of production to drive the site back to steady state values. Notably the increases in soil production rates (P_s) as soil thickness is perturbed from steady state are modest, and the rates of these processes are slow. The response shown in Fig. 1A to perturbation represents a built-in *resilience* of the hillslope soil system to perturbation: negative feedbacks that appear to drive the system back to local steady state (Fig. 1C).

3. Examining hillslope soil processes without plants

Our understanding of how geomorphic processes function on plant-covered hillslopes is considerable, but few comparative studies of vegetated vs. unvegetated hillslopes exist (see below for discussion). Our goal is to derive ways to examine the way the processes may operate in the absence of vegetation to underscore the additive effect of biota on the land surface evolution. Five different avenues are considered: (i) field studies of landscapes that climatically lie outside the range of plants, but that still have liquid water; (ii) the examination of large combined data sets from a broad range of climates and plant densities; (iii) inferences about pre land–plant landscapes and geochemical processes; (iv) modeling; and (v) experiments. While likely not exhaustive, these five categories are discussed sequentially below, and are then used to arrive at preliminary perceptions of the cumulative effect of plants on hillslope soil systems.

3.1. Hyperarid landscapes

Plant productivity, and the ability of plants to exist, depend on liquid water. Areas where net primary productivity (NPP) drops below 100 g m^{-2} y^{-1} exist only where rainfall is negligible (Fig. 2). We focus on northern Chile, where rainfall and plant cover decline continuously with decreasing latitude while many other geological and geographic factors remain constant.

Owen et al. (2010) observed the changes in the rate and mechanisms of hillslope processes that occurred along a climate gradient from semiarid to a plant-free hyperarid condition in northern Chile, and the changes observed appear to reveal some important clues to biotic vs. abiotic landscapes. Hillslopes with ~100 mm mean annual precipitation (MAP) have typical biotic soil production mechanisms—root penetration, animal and insect burrowing—that convert saprolite to mobile soil material and have biotically mediated soil transport leading to erosion (Fig. 3A, B). The resulting soil mantle reflects the balance between these processes. With a decline in precipitation to ~10 mm MAP (Fig. 3C, D), nearly all plants disappear (only rare succulents, surviving on fog moisture and with no appreciable root systems, occur). Soil

production appears to be due largely to the chemical breakdown of the exposed granitic bedrock surface, creating coarse loose, sand grains. Erosion appears to be largely the result of advective overland flow during infrequent rainfall events, reflected in the presence of sorted sand and gravel bands along slope contours. Soil is nearly absent, and erosion approximately matches maximum soil production rates. Finally, when both plants and rainfall essentially disappear (~1 mm MAP) (Fig. 3E, F), processes change further. Soil production is now the combined accumulation of atmospheric dust and sulfate salt, which additionally penetrates the bedrock and pries rock fragments into the soil mantle by shrink–swell mechanisms. Erosion is a combination of rare overland flow (indicated by the presence of *Zebra stripes*, a surface sorting of stones; Owen et al., 2013) and sulfate shrink–swell. Rates of erosion are slower than maximum production rates and a thin soil mantle persists. In summary, this climate transect clearly shows the ways in which production and erosion processes change as landscapes become plant-free and shows that, when plants disappear but occasional rainfall remains, hillslopes are nearly soil-free.

3.2. Data compilations

Since the advent of the use of cosmogenic nuclides to determine rates of soil production (Heimsath et al., 1997), a limited, but growing, number of studies have examined soil thickness and production rates in various settings (Table S1), with the goal of establishing local soil production functions (Table 1). A recent detailed compilation of these data was made by Stockmann et al. (2014). These studies now span an enormous range in rainfall and plant density. Here, we use these data to try to explore the interrelated role of climate—and plants—on soil thickness, soil production, and soil residence times.

We plot the data from these studies as a function of the ratio of the mean annual precipitation (mm) to the mean annual temperature (K), a ratio called the *aridity index*, a metric developed or modified by numerous people over the past century (see Quan et al., 2013, for a discussion). In the aridity index for a given MAP, the index declines (becomes more arid) with increasing MAT. While other meteorological calculations, like a detailed soil water balance, might be more physically informative, we lack monthly rainfall and temperature data for many of our sites, as well as information on other meteorological parameters. However, it is important to include precipitation and temperature in the evaluation of landscape processes, as they interactively control water availability and rates of physical processes.

3.2.1. Soil thickness

Hillslope soil thickness (Eq. 1) is the balance between soil production and erosion and is locally correlated with landscape curvature (Heimsath et al., 1997). Thickness does not appear to correlate to

Table 1
Compilation of sites with soil production functions.

Site	Symbol	MAT (C)	MAP (mm)	Equation (m/My) (H = cm)	R	Reference
Yungay, Chile	Ch1	16	2	$0.96 \times e^{(-0.00024H)}$	0.04	Owen et al. (2010)
Chanaral, Chile	Ch2	15	10	$4.40 \times e^{(0.40H)}$	0.42	Owen et al. (2010)
La Serena, Chile	Ch3	13.6	100	$33.00 \times e^{(-0.030H)}$	0.53	Owen et al. (2010)
Blasingame, CA	SN1	16.6	370	$40.39 \times e^{(-0.0077H)}$	0.43	Dixon et al. (2009)
Frog Hollow, AU	Au 1	12.8	650	$35.78 \times e^{(-0.021H)}$	0.97	Yoo et al. (2007)
Nunnock River, AU	Au 2	11.4	720	$65.92 \times e^{(-0.020H)}$	0.88	Heimsath et al. (2001a)
Point Reyes, CA	PR	13	760	$84.58 \times e^{(-0.017H)}$	0.94	Heimsath et al (2005)
Bretz Mill, CA	SN2	12	770	$43.71 \times e^{(0.00047H)}$	0.03	Dixon et al. (2009)
Providence Creek, CA	SN3	8.9	920	$77.45 \times e^{(-0.0086H)}$	0.64	Dixon et al. (2009)
San Gabriels, CA	SG	13	950	$160.77 \times e^{(-0.033H)}$	0.76	Heimsath et al. (2012)
Sierra Summit, CA	SN4	4	1060	$17.86 \times e^{(0.0014H)}$	0.05	Dixon et al. (2009)
Tennessee Valley, CA	TV	14	1200	$86.14 \times e^{(-0.024H)}$	0.94	Heimsath et al. (1997)
Tin Camp, AU	AU 3	27	1400	$45.53 \times e^{(-0.020H)}$	0.84	Heimsath et al. (2009)
Coos Bay, OR	OR	11	2300	$157.35 \times e^{(-0.010H)}$	0.56	Heimsath et al. (2001b)
New Zealand	NZ1	5	10,000	$1815 \times e^{(-0.058H)}$	0.90	Larsen et al. (2014)
New Zealand	NZ2	5	10,000	$3199 \times e^{(-0.055H)}$	0.83	Larsen et al. (2014)

increases in NPP on regional scales (Fig. 2). However, no measured thickness exceeds 150 cm, and most are <100 cm. This depth was suggested by Dietrich et al. (1995); Roering et al. (1999) to be the maximum soil thickness in forested landscapes affected by tree throw, and the effect of tree throw on soil horizon depth and spatial variability is being explored through various modeling approaches (Finke et al., 2013). Thus, while no data exist to suggest that average soil thickness responds to changes in NPP, the maximum thickness that soils attain on hillslopes may be plant regulated.

3.2.2. Soil production

For each soil production study in Table 1, we derived an exponential soil production function, calculated the maximum rate of production at zero soil thickness: P_0 in Eq. 2, and plotted this in Fig. 4A. While for simplicity we use an exponential model, we are agnostic about the nature of the soil production function at shallow thicknesses, and we discuss this further below. The data indicate a strong relationship of maximum soil production rate with aridity index, and soil production appears to be unrelated to regional rates of tectonic uplift (Table S1). For comparison, Norton et al. (2014) recently developed a model of soil production by assuming that the soil production rate is controlled by an Arrhenius formula:

$$SPR_{max} = a_o Pe^{\frac{-E_a}{R}\left(\frac{1}{T}-\frac{1}{T_0}\right)} \tag{5}$$

where SPR_{max} = maximum soil production (L / t), a_o = a factor to scale precipitation (P) rate (L / t) to soil production function, E_a = activation energy for silicate weathering (kJ mol^{-1}), R is the gas constant, T = MAT (K) and T_0 a reference temperature (278 K). Eq. 5 derives from studies of chemical weathering and suggests that the maximum rate of production is largely controlled by chemical alteration that liberates particles. In Fig. 4A, we plot the calculated SPR_{max} value for the sites with CRN determined production rates. The trend reveals that SPR_{max} values are generally similar to measured P_0 as a function of the aridity index. The SPR_{max} is dependent on T and precipitation, so the correlation with the aridity index is expected. The fact that P_0 shows strong climate relations is a relatively new finding, one that differs from the earlier analyses that suggested very weak climate effects on watershed erosion rates (von Blanckenburg, 2006). But here we focused on the potential maximum production rate, not on the actual soil production rate, which may be adjusted through soil depth to match the incision rate at channels bordering the hillslope. The lack of correlation with uplift in our data appears to be due in part to sites having not adjusted to uplift that drives stream incision and thus the boundary condition of hillslope and soil co-evolution. In Chile for example, the climate is so dry that stream incision simply cannot keep up with regional uplift (Amundson et al., 2012).

Norton et al. (2014) also derived a steady state soil thickness model that reflects the balance between soil production and denudation (which for steady state soils is equivalent to P_s (Eq. 2)). Using the P_0 values for sites in Fig. 4A, and Norton et al.'s (2014) Eq. 11, we calculated the predicted soil thicknesses for the sites (Fig. 2). There is a very consistent relation between the predicted values and the range of values found in any site, showing the interplay between physical and chemical processes on controlling soil thicknesses.

The correlation of P_0 to climate mirrors the response of NPP to climate (at least for the less humid end members) (Fig. 4B) and raises a question of what role plants play in the observed soil production patterns. For example, biotic effects must be inherently embedded into chemical weathering functions such as Eq. (5), as studies have shown that the rates of weathering in unvegetated watersheds are 3 to 5 times slower than watersheds with plants (Moulton et al., 2000; Berner et al., 2003). How can one develop comparative vegetation mantled and vegetation-free sites for soil production rate comparisons? Within many soil-mantled landscapes are rock outcrops of various
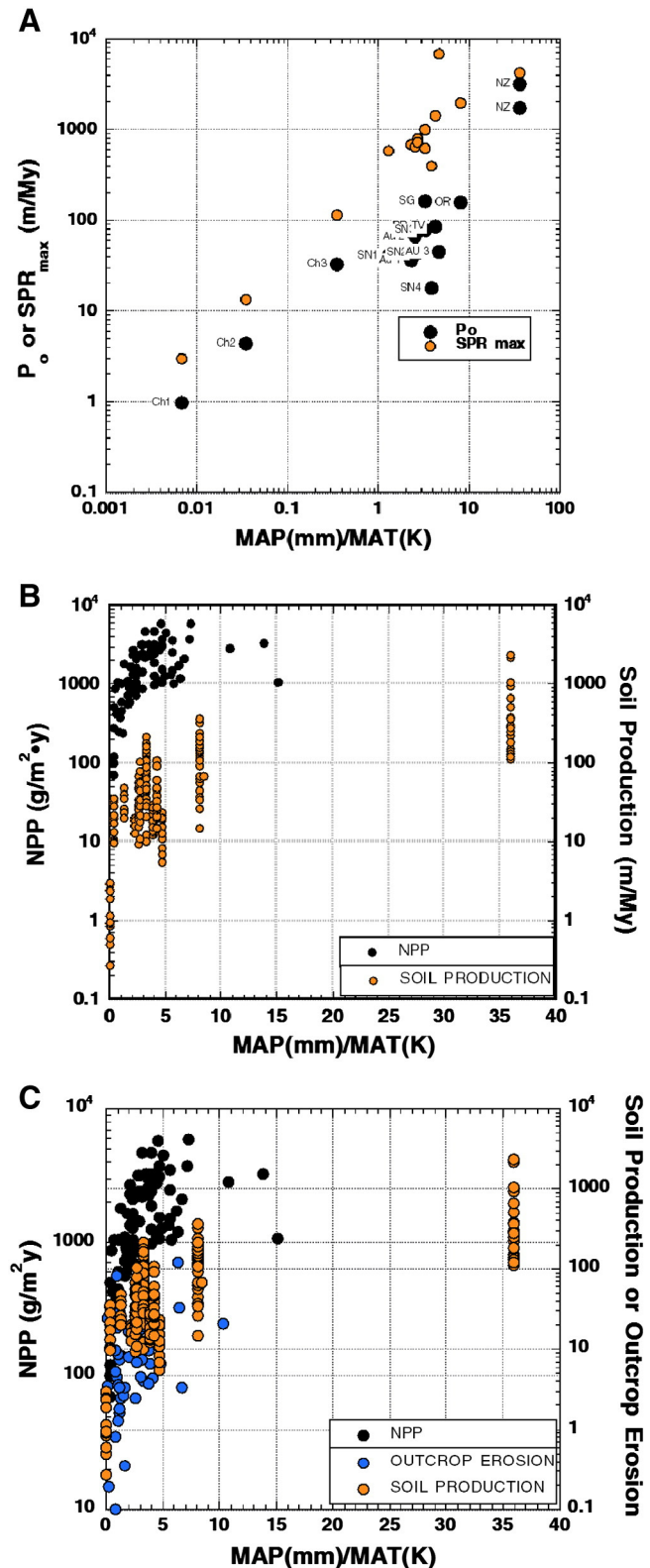

Fig. 4. (A) The change in maximum soil production rates (Table 1) vs. climate. The symbols adjacent to a point are listed in the second column of Table 1. Also shown is the calculated SPR_{max} of Norton et al. (2014), which emphasizes the combined production of soil and saprolite by chemical weathering. (B) The measured soil production rates (Table 1) and NPP vs climate. (C) As in (B) but with the addition of outcrop erosion rates from Portenga and Bierman (2011). See text for more discussion.

types; and a growing number of studies, compiled by Portenga and Bierman (2011), have examined outcrop erosion rates across a broad climate spectrum. Portenga and Bierman (2011) found that rates of

vegetated basin erosion based on stream sediment samples were greater than the outcrops: 218 m My^{-1} versus 12 m My^{-1}. We consider these outcrop rates to approximate plant-free controls (though local variations in rock composition may also contribute to their emergence), which we can then use to compare to soil production rates on the plant-mantled soilscapes (Fig. 4C). Although the data are quite variable, outcrop erosion rates are roughly a 0.5 to 1 order of magnitude slower than plant-covered landscape denudation (Portenga and Bierman, 2011; Hahm et al., 2014). The ability of plant-covered land surfaces to experience higher rates of denudation is not unanticipated and reflects combined physical (root penetration, tree throw) and chemical (increased soil CO$_2$, organic acids and chelates) weathering enhancements by the vegetation (Hahm et al., 2014). Erosion rates with no soil mantle are consistently lower than mantled landscapes which not only implicate the importance of plants, but also suggest that a humped soil production function, with a critical soil thickness, may be more descriptive of many landscapes. The peak may be very close to zero thickness. Identifying the critical thickness in bi-modal landscapes and determining why other landscapes lack a clear bi-modal exposure of soil vs. no soil are clearly areas for more research.

3.2.3. Soil residence time and fertility

One of the paradigms of ecosystem ecology that has emerged in the past 40 years is that the nutritional status, and potential productivity, of terrestrial ecosystems varies with soil age (Walker and Syers, 1976). Sediment and rock have little N, and N accumulates in soils from atmospheric deposition and secondarily from biological N fixation, reaching steady state values in a given climate on the order of 10^2 to 10^3 y (Amundson et al., 2003). In locations with organic-rich sedimentary rock, lithologic sources of N can also be an important source of soil N (Morford et al., 2011). Conversely, essential rock-derived nutrients such as Ca and K, and especially P (which along with N controls most ecosystem productivity,) are at maximum total levels at time 0 and decline with time caused by chemical weathering of the primary minerals and by the leaching of released ions to groundwater and streams. However, as argued by Porder et al. (2007), biologically available P may increase initially as rock-derived P is released, but this also eventually begins a decline in availability. Thus, a window of time for maximum soil fertility appears to exist, a period where neither N nor P deficiencies limit production (Vitousek et al., 1997). Studies that have demonstrated this relationship have been conducted on level and stable landscapes with minimal erosion, where soil age is equivalent to the elapsed time since the landform stabilized.

The *age* of hillslope soils can be equated with their residence or, more accurately, turnover times (τ), the time required to replace the soil thickness by soil production:

$$\tau = H/P \qquad (6)$$

assuming for simplicity that rock and soil bulk densities are equal. In reality, soil is less dense than rock and residence times will be lower (about two-thirds of the calculated values), but this simple comparison reveals some interesting relationships. Two intriguing questions that the data set allows us to address are (i) what is the range of residence times for hillslope soils, and (ii) how do these times compare to the rates of processes that drive ecosystem fertility? Fig. 5A illustrates the

calculated hillslope soil residence times vs. effective precipitation. The residence times range between 1 and 100 Ky, and average ~10 Ky. We examined how these residence times compare to N and P availability as follows. Nitrogen cycling rates closely match those of C as both are bound in organic matter (Brenner et al., 2001). Several compilations show how soil C decomposition constants (which control the time to steady state) vary with temperature and precipitation. Using data from Amundson (2001) (see Trumbore et al., 1996 for discussion of different pool behaviors and temperature sensitivity), we calculated an approximate time to N steady state (Fig. 5A) as a function of MAP/MAT. To test whether the calculated relationship of hillslope N to residence time is valid, we used data collected by Yoo et al., 2005a, 2005b, and plotted soil C storage (which should mirror N storage) vs. soil residence time. The relation (Fig. 5B) shows that the total C storage responds to residence time in the manner predicted by time-dependent soil C models (Amundson, 2001). As a caveat, however, we also note that the C is also somewhat correlated with soil thickness (Fig. 5C), so that rates of soil removal may not be as critical as the total volume of soil available to accumulate C and N.

Phosphorus is bound in primary minerals. As a proxy for P-bearing mineral weathering, we here calculate the albite weathering front advance rate through soils, examining the loss of Na. While Na is not a plant–essential element, feldspars as a group also contain K and Ca, which are important nutrients, and thus reflect the chemical release and loss of rock-derived nutrients. Additionally, reported field-based weathering rates of the two minerals are similar: apatite = 6.8×10^{-14} mol m^{-2} s^{-1} (Buss et al., 2010) vs. albite = 10^{-12} to 10^{-16} mol m^{-2} s^{-1} (White and Brantley, 2003). This approach also enables us to examine recently published data compilations of Na weathering and removal in soils (Rasmussen et al., 2011), allowing us to test some of our assumptions. The weathering front advance rate of albite through soils was calculated using the expression (White et al., 2008; Maher, 2010):

$$\omega = \text{weathering advance rate } (L/t) = q_h[m_{sol}/M_{total}] \qquad (7)$$

where q_h = fluid flux (m/y), M_{total} = total moles of mineral (2300 mol/m^3) (initial mass of albite content in protolith), and m_{sol} = the mass of plagioclase dissolved in a thermodynamically saturated volume of pore water (mol/m^3), using an lnK (albite) = ΔG^o_R / $-RT$ (White et al., 2008). We assumed (see White et al. (2008)) that one fifth of total precipitation becomes fluid flux. In Fig. 5A, the time for the front to pass through a 50-cm-thick soil layer is illustrated.

To test our albite weathering front calculations, we turn to a compilation of studies of weathering of granitic terrains (Rasmussen et al., 2011). In this compilation, the fractional loss of Na (in albite) from the soil surface (upper 10 cm) and cosmogenically derived denudation rates (D) were reported for numerous watersheds. If our calculations of weathering advance rates (ω) are correct, there should be a relationship between the fractional loss of Na and the ratio of weathering front advance/denudation rates. Ideally, when the ratio is <1, the pace of denudation exceeds chemical weathering, and soils should retain Na (while the reverse should be true for ratios >1). This hypothesis assumes no biotic soil mixing, which will counteract the effect of weathering, although even in highly bioturbated soils weathering fronts are observable (White et al., 2008). Fig. 5D shows Na losses vs. ω/D, and

Fig. 5. (A) Soil residence times (from sites in Table 1) and calculated time to reach N steady state (using data from Amundson, 2001) and time to weather the upper 50 cm using the weathering front advance model of White et al. (2008) (see text for more discussion). Also illustrated are the measured fractional losses of mobile elements from the upper 50 cm or whole soil if thickness is l < 50 cm (data from Rasmussen et al., 2011; Larsen et al., 2014). (B) Soil C storage as a function of residence time for Tennessee Valley, CA. The pattern is consistent with the concept that soils are N limited in initial stages of development and possibly P limited from inadequate time to develop bioavailable P sources (see Porder et al. (2007)). (C) NPP vs. soil thickness at Tennessee Valley, CA. While soil thickness at Tennessee Valley is related to residence time (Yoo et al., 2006), NPP seems most strongly related to residence time as illustrated in (B). (D) The fractional loss of soil surface Na (0 = no loss, −1 = 100% loss) (data from Rasmussen et al., 2011) vs. the weathering advance rate (ω) divided by denudation rate (D). The data indicate that the retention of nutrients is dependent on the balance between downward migration of a chemical weathering front and the rate of soil production, and only sites with high rates of weathering and modest rates of denudation become nutrient limited. The ratio ω/D is analogous to the definition of CDF by Riebe et al. (2004), with the distinction that ω here is based on a chemical weathering model. (E) Fractional loss of total elements (in upper 50 cm/whole soil if less than 50) vs. residence time for data from Rasmussen et al. (2011); Larsen et al. (2014). The trends suggest that for soil profiles, the degree of chemical weathering loss is dependent on denudation rates, and that locally, changes in soil thickness and resulting production rates (see Eq. 2) can modulate the soil nutrient status and result in changes in plant productivity.
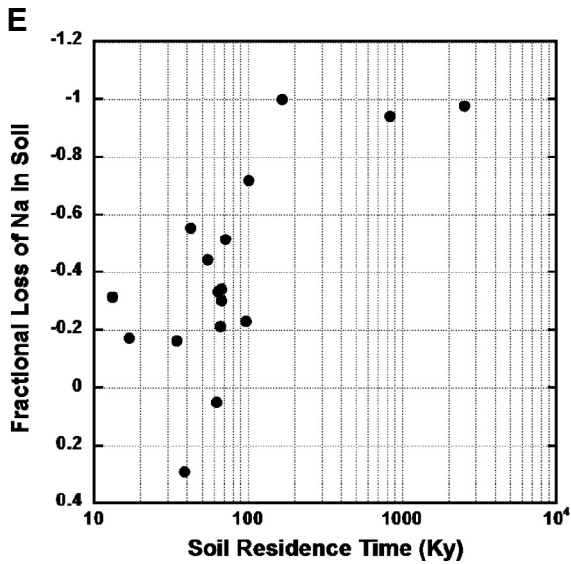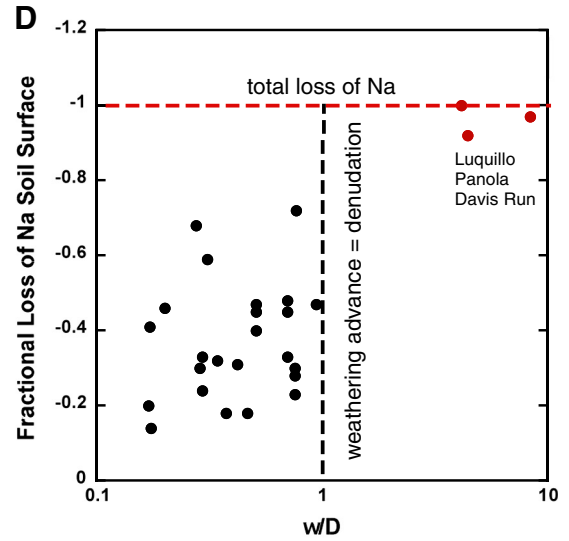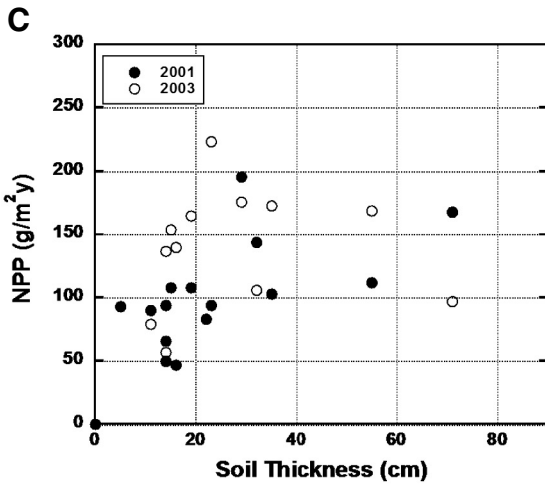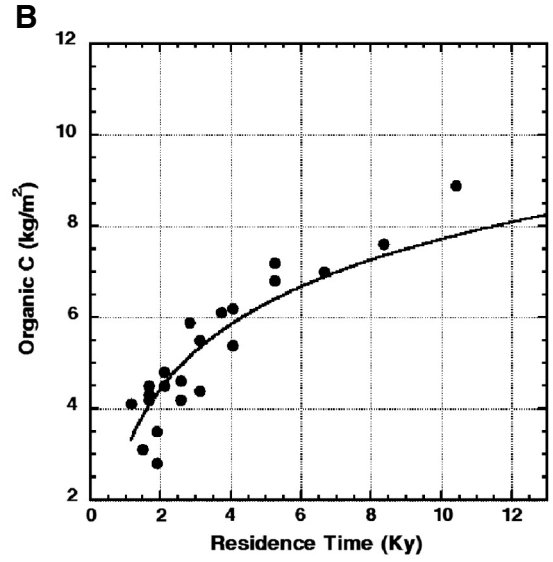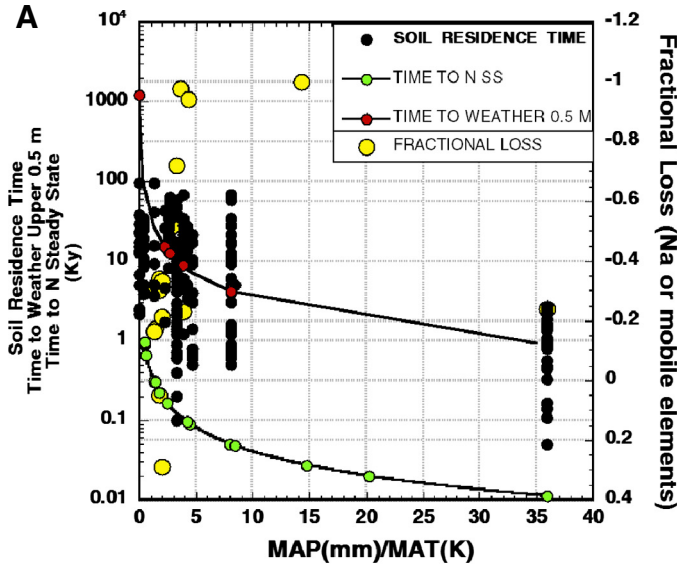
the results generally conform to our hypotheses. Only sites in warm and moist environments (with rapid chemical weathering rates and high ω/D values) have lost all or nearly all Na at the soil surface. This supports our calculations shown in Fig. 5A and illustrates how upland landscapes may be, as Hutton proposed, indefinitely fertile because of the continuous soil replenishment of relatively fresh bedrock by erosion.

The ratio ω/D is somewhat related to the definition of the *chemical depletion fraction* (CDF) (Riebe et al., 2004):

$$CDF = 1 - C_{i,p}/C_{i,x} = W/D \qquad (8)$$

where $C_{i,p}$ and $C_{i,s}$ = immobile element concentration (e.g. Zr, Ti) in parent material and soil, respectively; W = chemical weathering flux and D = total denudation flux; CDF is the negative of *tau* (Brimhall et al., 1992; Ewing et al., 2006), the fractional loss of a mobile element (or all mobile elements) from a soil. We used calculated chemical weathering advance rates and measured D to estimate fractional losses of Na (e.g. CDF of Na). For a given ω, the trend in Fig. 5D suggests that the fractional loss of an element at the soil surface horizon should decline with increasing D. We also calculated the fractional loss of Na in the upper 50 cm (or whole soil if <50 cm) and the soil residence time (for the entire reported soil thickness divided by cosmogenically derived denudation rates) using the data from Rasmussen et al. (2011); Larsen et al. (2014) (Fig. 5E). The plot shows that fractional loss appears dependent on residence time (and hence D), with total loss of Na at residence times exceeding 10^5 y. Sites that exceed this residence time and that are stripped of nutrients are the warm, wet, and relatively slowly uplifting regions in Puerto Rico or the SE USA (Davis Run, Panola, Luquillo). In Fig. 5A, we include the Na fractional loss values of the sites in Fig. 5E vs. the aridity index, again showing how the high ω/D sites fall outside the zone of fertility. Porder et al. (2005a, 2005b, 2007) found that foliar N and P are lowest on the stable edges of volcanic flow escarpments in Hawaii (with very long residence times), while actively eroding escarpments have much higher nutrient contents. Additionally, Porder et al. (2007) examined the relationship between soil residence time and the kinetics of P weathering and immobilization, providing estimates of conditions (soil thickness vs. erosion rate) that may lead to P limitations, particularly under low erosion rates.

The apparent relations in Fig. 5D, E are important in that it shows that locally a perturbation that increases erosion (D), decreases soil thickness and will, in turn, decrease the nutrient leaching in the soil relative to the parent material. This implies that potential feedbacks may occur between plants and physical processes in terms of maintaining adequate soil fertility. Do thickening soils on eroding hillslopes with longer residence times become nutrient limited such that plant productivity declines, leading to thinning soils and associated increases in soil production? NPP data for soils with relatively short residence times (e.g. thin soils in Fig. 5C) show that NPP responds to soil thickness. Possibly a minimum soil thickness exists where soils become too thin to hold root mass, thus limiting NPP. At the other extreme, as soil residence times continue to increase, they cross a transition to soils on stable landforms, where numerous studies reveal nutrient limitations. But are there eroding hillslopes where residence times are long enough that P limitations become a potential limit to plant growth? These uncertainties point toward opportunities for combined geomorphic, biogeochemical, and ecological studies of hillslope soils (e.g. soil production rates/soil nutrient contents/NPP) that may illuminate biotic–abiotic feedbacks that may greatly enhance present models (see Buendia et al., 2014).

3.3. Pre-plant Earth

Primitive land plants (e.g. lichens, mosses) are suspected of having evolved and spread in the late Precambrian (Kennedy et al., 2006), while vascular land plants evolved in the Silurian (Corenblit and Steiger, 2009). Corenblit and Steiger (2009) reviewed the physical and chemical impacts that land plants have on geomorphic processes and

also considered the evolutionary feedbacks between abiotic conditions and resulting plant biology (and their engineering capabilities). Recently, two separate approaches to evaluating the chemical, mineralogical, and biotic changes have been published that shed light on Precambrian soils and possible changes in the hillslope soil mantle following the evolution of soil stabilizing organisms. Kennedy et al. (2006) compiled a multiproxy chemical and mineralogical record for the past 2 By. The authors suggested enhanced terrestrial cover of moss, lichens, etc. developed by ~700 Ma. At the time that these organisms are believed to have emerged, a corresponding increase in ^{87}Sr in marine carbonates is interpreted as reflecting increased rates of chemical weathering on land. An apparent increase in expandable clays (e.g. smectites, vermiculite) in the sedimentary record is interpreted in the record to suggest that colonization of land surfaces stabilized the soil cover, increased chemical weathering (drawing down CO_2), created expandable clay (a common soil mineral) that bound with organic matter and was buried—ultimately driving a rise in atmospheric O_2. In a subsequent paper, Knauth and Kennedy (2009) examined the C isotope record of various carbonate sedimentary rock assemblages, argued that a strong biological (plant) signal in late Precambrian carbonates is associated with waters that passed through terrestrial landscapes, and concluded that soil processes had commenced and were forming clays and altering the C and O cycles. Both these papers, which include pre- and post-land stabilization evidence, indicated profound changes in properties associated with soils, and suggested that while soils were likely present in some form throughout the Precambrian, they underwent a fundamental change coincident with the evolution of plants and/or related organisms.

3.4. Process modeling

Marston (2010), in a recent review, pointed out that modeling of combined biotic (especially plants) and geomorphic processes, and their feedbacks, is a poorly developed field—but one of emerging interest. In recent years, some mechanistic animal–soil feedback models have been developed, which serve as guides for future vegetation-centric models. For example, Yoo et al (2005a, b), built on Roering et al.'s (1999) model of nonlinear sediment transport resulting from the effects of tree throw in generating sediment. Yoo et al. (2005a, 2005b) devised a feedback model between gopher density, soil thickness, and sediment transport. The authors evaluated the impact and response time of gopher dominated hillslope to climate change and subsequent changes in gopher populations and sediment movement. More generally, they found that gopher-mediated landscapes have more homogeneous erosion rates, suggesting that biotic (gophers) landscapes maintain topographic relief over time. More recent process-based research has focused on the role of gophers in creating "Mima mounds" and vernal pools in California grassland (Reed and Amundson, 2012; Gabet et al., 2014).

The modeling of bare vs. vegetated landscapes is less advanced than the impact of burrowing animals on hillslope soils. Collins et al. (2004) developed a Channel-Hillslope Integrated Landscape Development (CHILD) model to create a feedback between erosion and plant cover (soil cover was not evaluated). For the bare vs vegetated experiments, the authors found that the vegetated landscapes had greater relief—as suggested by Yoo et al. (2005a, 2005b) for gopher-dominated landscapes. Roering (2008) proposed that tree root growth and throw introduced a depth-dependency in soil flux processes that linked soil thickness, flux and hillslope curvature. Furbish et al. (2009) showed theoretically that the normal to the surface lofting of particles by biological activity leads to diffusive-like soil transport, and to a depth dependency in the flux rate. Gabet and Mudd (2010) developed a numerical model of soil production, erosion, and thickness to explore the effect of tree growth and throw on hillslope soil processes for the Pacific Northwest. As part of their exercise, they conducted scenarios where trees were removed from the experiments, and the results led them to suggest that 'prior to trees, bedrock erosion rates, largely driven by chemical

weathering processes and small-scale physical disturbances, may have been unable to keep pace with transport rates, leaving slopes glazed by only a thin cover of weathered material.' Martin et al. (2013) derived effective diffusion coefficients for slope dependent transport driven by periodic tree throw, as influenced by stand replacing forest fires. Hoffman and Anderson (2014) employed discrete element modeling to show that the cycle of tree root growth and decay can be an effective transport agent of soil. Pawlik (2013) offers an extensive review of the influence of trees on hillslope geomorphic processes.

Numerous opportunities exist to more mechanistically include the role of plants (and climate) in hillslope soil models. First, the soil production function appears to be both climate and plant driven. Second, the nutrient state of soil is impacted by the soil residence time, which should in turn impact plant productivity. Does site specific variation in soil thickness (and thus residence time) cause variations in plant cover as that proposed in the case of animals (e.g. Yoo et al., 2005a, 2005b)? Can plant cover be quantitatively linked to the effective friction coefficient or similar parameter that expresses its ability to resist particle movement? The growing sparse list of modeling studies listed here illustrates the numerous opportunities that exist in this area.

3.5. Experiments

One obvious way to create plant-free controls is to remove vegetation, and a considerable history of experimental removal of plants to examine changes in erosion rates exists. However, the largest plant removal *experiment* that has been replicated millions of times is cultivation, and the effects of cultivation practices on accelerated erosion is a very well studied issue, with several compilations (Wilkinson and McElroy, 2007; Montgomery, 2007). When cultivation leaves soil bare or partially bare for significant periods of time, it shifts its occurrence in soil erosion mechanisms from slow, biotically driven particle movement to rapid advective losses by running water. Using the compilation of soil erosion rates by Montgomery (2007), we compared soil production rates to "plant-free" erosion rates as a function of effective precipitation (Fig. 6). This comparison clearly shows that hillslopes devoid of plants would rapidly lose their soil mantle—a mantle derived under biotic conditions—within centuries to millennia (Montgomery, 2007). Many of the agricultural erosion rates are from landscapes underlain by loess, till, or other relatively erodible materials—areas where we lack natural soil production rate data.
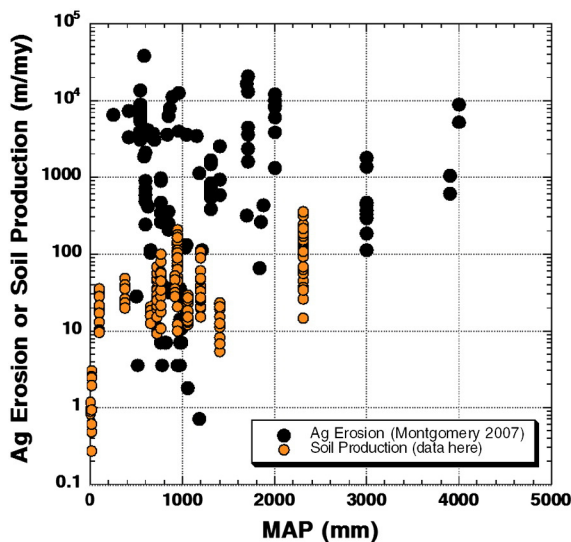

Fig. 6. Variations in soil production rates (from the sites in Table 1) and agricultural erosion rates (from Montgomery (2007)) vs. MAP. The agricultural data show little variation with MAP but do show that in many locations the rates of agricultural erosion greatly exceed the natural rates of soil production.

Nonetheless, the effects of farming offer our clearest suggestion that largely soil-free hillslopes would dominate a plant-free planet.

4. Conclusions

Determining, on our green planet, what hillslopes might be like in a plant-free world is akin to *seeing through a glass, darkly*. The extreme range of climate, the ancient geological record, and large human disturbances provide sometimes fleeting glimpses of how profoundly plants impact the physical evolution of the planet. The goal of a growing number of geomorphologists and soil scientists is to begin to qualitatively and quantitatively understand the additive effect of vegetation on the physical processes that shape the earth's surface. Some provisional conclusions we arrive at are as follows:

- If water is available, a world without plants would likely have little or no soil on hillslopes.
- Plants may control maximum soil thickness.
- Soil production rates may exceed outcrop erosion rates by ~1 order of magnitude.
- Soil residence times are remarkably constrained within a broad window of nutrient sufficiency/optimization, yet environments with high weathering rates and low denudation rates may suffer from a deficiency of rock-derived elements.
- For sites from a wide range of climates, the degree of elemental loss apparently declines with decreasing soil residence time (and increasing denudation rate). This suggests that local feedbacks are possible between plants–nutrients–soil thickness.
- Modeling and paleochemical studies suggest that evolution of plants changed the earth's soil mantle, in turn changing the atmospheric chemistry and animal evolution. Another possibility, on long time scales, is that geomorphic conditions have impacted plant evolution.

Many existing hillslope models—which implicitly contain the effect of plants or plant processes—contain negative feedbacks between soil thickness and rates of production or erosion. This implies a degree of resilience of hillslope soil systems to natural perturbations—one greatly exceeded by direct human intervention. Bedrock landscapes, free of a continuous soil mantle, exist, and although this directly implies erosion prevents soil buildup, whether the emergence of such landscapes reflects a threshold condition in the soil production function or simply records where erosion chronically exceeds soil production rate remains unclear. The challenge now is how to explicitly and quantitatively account for the role of biota in the production of soil from bedrock and its transport downslope. This is needed to explore the co-evolution of the soil mantle and life and to explore landscape evolution and the influence of climate. As with the effects of warming on soil carbon, anthropogenic impacts may affect processes that operate on geological time scales, and measurable responses or impacts may be felt largely by future generations. Being able to parameterize the geomorphic role of plants on physical processes and to predict when and how the resulting impacts will be felt, is not only a scientific challenge, but arguably an ethical obligation to future generations.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.geomorph.2014.12.031.

References

Amundson, R., 2001. The carbon budget in soils. Ann. Rev. Earth Planet. Sci. 29, 535–562.

Amundson, R., Austin, A.T., Schurr, E.A.G., Yoo, K., Matzek, V., Kendall, C., Uebersax, A., Brenner, D., Baisden, W.T., 2003. Global patterns of the isotopic composition of soil and plant nitrogen. Glob. Biogeochem. Cycles 17, 1031. http://dx.doi.org/10.1029/2002GB001903.

Amundson, R., Dietrich, W., Bellugi, D., Ewing, S., Nishiizumi, K., Owen, J., Finkel, R., Heimsath, A., Stewart, B., Caffee, M., 2012. Geomorphic evidence for the late Pliocene onset of hyperaridity in the Atacama Desert. Geol. Soc. Am. Bull. 124, 1048–1070.

Berner, E.K., Berner, R.A., Moulton, K.L., 2003. Plants and mineral weathering: present and past. In: Holland, H.D., Turekian, K.K. (Eds.), Surface and Ground Water, Weathering, and Soils. Treatise on Geochemistry vol. 5. Elsevier.

Brenner, D.L., Amundson, R., Baisden, W.T., Kendall, C., Harden, J., 2001. Soil N and ^{15}N variation with time in a California annual grassland ecosystem. Geochem. Cosmochim. Acta 65, 4171–4186.

Brimhall, G.H., Chadwick, O.A., Lewis, C.J., Compston, W., Williams, I.S., Danti, K.J., Dietrich, W.E., Power, M.E., Hendricks, D., Bratt, J., 1992. Deformational mass-transport and invasive processes in soil evolution. Science 255, 695–702.

Buendía, C., Arens, S., Hildebrandt, A., 2014. On the potential vegetation feedbacks that enhance phosphorus availability — insights from a process-based model linking geological and ecological timescales. Biogeosciences 11, 3661–3683.

Buss, H.L., Mathur, R., White, A.F., Brantley, S.L., 2010. Phosphorus and iron cycling in deep saprolite, Luquillo Mountains, Puerto Rico. Chem. Geol. 269, 52–61.

Carson, M.A., Kirkby, M.J., 1972. Hillslope Form and Process. Cambridge University Press (475 p.).

Collins, D.B.G., Bras, R.L., Tucker, G.E., 2004. Modeling the effects of vegetation–erosion coupling on landscape evolution. J. Geophys. Res. 109, F03004 (10.1029/2003JF000028).

Corenblit, D., Steiger, J., 2009. Vegetation as a major conductor of geomorphic changes on the Earth surface: toward evolutionary geomorphology. Earth Surf. Process. Landf. 34, 891–896.

Cox, N.J., 1980. On the relationship between bedrock lowering and regolith thickness. Earth Surf. Process. 5, 271–274.

Culling, W.E.H., 1963. Soil creep and the development of hillslopes. J. Geol. 71, 127–161.

D'Odorico, P., 2000. A possible bistable evolution of soil thickness. J. Geophys. Res. 105 (B), 927–935.

Dietrich, W.E., Reiss, R., Hsu, M.-L., Montgomery, D.R., 1995. A process-based model for colluvial soil depth and shallow landsliding using digital elevation data. Hydrol. Process. 9, 383–400.

Dixon, J.L., Heimsath, A.M., Amundson, R., 2009. The critical role of climate and saprolite weathering in landscape evolution. Earth Surf. Process. Landf. 34, 1507–1521.

Ewing, S.A., Sutter, B., Owen, J., Nishiizumi, K., Sharp, W., Cliff, S.S., Perry, K., Dietrich, W.E., McKay, C.P., Amundson, R., 2006. A threshold of soil formation at Earth's arid-hyperarid transition. Geochim. Cosmochim. Acta 70, 5293–5322.

Fernandes, N.F., Dietrich, W.E., 1997. Hillslope evolution by diffusive proesses: the time-scale for equilibrium adjustments. Water Resour. Res. 33, 1307–1318.

Finke, P.A., Vanwalleghem, T., Opolot, E., Poesen, J., Deckers, J., 2013. Estimating the effect of tree uprooting on variation of soil horizon depth by confronting pedogenic simulations to measurements in a Belgian loess area. J. Geophys. Res. Earth Surf. 118, 2124–2139.

Furbish, D.J., Haff, P.K., Dietrich, W.E., Heimsath, A.M., 2009. Statistical description of slope-dependent soil transport and the diffusion-like coefficient. J. Geophys. Res. Earth Surf. 114. http://dx.doi.org/10.1029/2009JF001267.

Gabet, E.J., Mudd, S.M., 2010. Bedrock erosion by root fracture and tree throw: a coupled biogeomorphic model to explore the humped soil production runciton and the persistence of hillslope soils. J. Geophys. Res. 115, F04005 (10.10/2009JF001526).

Gabet, E.J., Perron, J.T., Johnson, D.L., 2014. Biotic origin for Mima mounds supported by numerical modeling. Geomorphology 206, 58–66.

Gilbert, G.K., 1877. Report on the Geology of the Henry Mountains. Dept of the Interior, US Govt Printing Office, Washington, DC.

Gould, S.J., 1987. Times Arrow, Time's Cycle: Myth and Metaphor in the Discovery of Geological Time. Harvard University Press, Cambridge, MA.

Hahn, W.J., Riebe, C.S., Lukens, C.E., Araki, S., 2014. Bedrock composition regulates mountain ecosystems and landscape evolution. Proc. Natl. Acad. Sci. http://dx.doi.org/10.1073/pnas.1315667111.

Heimsath, A.M., Dietrich, W.E., Nishiizumi, K., Finkel, R.C., 1997. The soil production function and landscape equilibrium. Nature 388, 358–361.

Heimsath, A.M., Chappell, J., Dietrich, W.E., Nishiizumi, K., Finkel, R.C., 2001a. Late Quaternary erosion in southeastern Australia: a field example using cosmogenic nuclides. Quat. Int. 83-83, 169–185.

Heimsath, A.M., Dietrich, W.E., Nishiizumi, K., Finkel, R.C., 2001b. Stochastic processes of soil production and transport: erosion rates, topographic variation and cosmogenic nuclides in the Oregon Coast Range. Earth Surf. Process. Landf. 26, 531–552.

Heimsath, A.M., Furbish, D.J., Dietrich, W.E., 2005. The illusion of diffusion: field evidence for depth-dependent sediment transport. Geology 33, 949–952.

Heimsath, A.M., Hancock, G.R., Fink, D., 2009. The 'humped' soil production function: eroding Arnhem Land, Australia. Earth Surf. Process. Landf. 34, 1674–1684.

Heimsath, A.M., DiBase, R.A., Whipple, K.X., 2012. Soil production limits and the transition to bedrock-dominated landscapes. Nat. Geosci. 5, 210–214.

Hession, W.C., Wynn, T., Resler, L., Curran, J., 2010. Geomorphology and vegetation: interactions. Depend. Feedbacks 116, 203–372.

Hoffman, B.S.S., Anderson, R.S., 2014. Tree root mounds and their role in transporting soil on forested landscapes. Earth Surf. Process. Landf. 39, 711–722.

Hutton, J., 1795. Theory of the Earth, Vol. II. Edinburgh. http://www.sacred-texts.com/earth/toe/toe22.htm.

Jenny, H., 1941. Factors of Soil Formation: A System of Quantitative Pedology. McGraw-Hill Books, NY.

Kennedy, M., Droser, M., Mayer, L.M., Pevear, D., Mrofka, D., 2006. Late Precambrian oxygenation: inception of the clay mineral factory. Science 311, 1446–1449.

Knauth, P.L., Kennedy, M.J., 2009. The late Precambrian greening of the Earth. Nature 460, 728–732.

Larsen, I.J., Almond, P.C., Egar, A., Stone, J.O., Montgomery, D.R., Malcolm, B., 2014. Rapid soil production and weathering in the Southern Alps, New Zealand. Science 343, 637–640.

Lieth, H., 1973. Primary production: terrestrial ecosystems. Hum. Ecol. 1, 303–332.

Maher, K., 2010. The dependence of chemical weathering rates on fluid residence time. Earth Planet. Sci. Lett. 294, 101–110.

Marston, R.A., 2010. Geomorphology and vegetation on hillslopes: interactions, dependencies, and feedback loops. Geomorphology 116, 206–217.

Martin, Y.E., Johnson, E.A., Chaikina, O., 2013. Interplay between field observation and numerical modeling to understand temporal pulsing of tree root throw processes, Canadian Rockies, Canada. Geomorphology 200, 89–105.

Montgomery, D.R., 2007. Soil erosion and agricultural sustainability. Proc. Natl. Acad. Sci. 104, 13,268–13,272.

Montgomery, D.R., 2012. The Rocks Don't Lie: A Geologist Investigates Noah's Flood. W.W. Norton and Co., NY.

Morford, S.L., Houlton, B.Z., Dahlgren, R.A., 2011. Increased forest ecosystem carbon and nitrogen storage from nitrogen rich bedrock. Nature 477, 78–81.

Moulton, K.L., West, J., Berner, R.A., 2000. Solute flux and mineral mass balance approaches to the quantification of plant effects on silicate weathering. Am. J. Sci. 300, 539–570.

Norton, K.P., Molnar, P., Schlunegger, F., 2014. The role of climate-driven chemical weathering on soil production. Geomorphology 204, 510–517.

Owen, J.J., Amundson, R., Dietrich, W.E., Nishiizumi, K., Sutter, B., Chong, G., 2010. The sensitivity of hillslope soil production to precipitation. Earth Surf. Process. Landf. http://dx.doi.org/10.1002/esp.2083.

Owen, J.J., Dietrich, W.E., Nishiizumi, K., Chong, G., Amundson, R., 2013. Zebra stripes in the Atacama Desert: fossil evidence for overland flow. Geomorphology 182, 157–172.

Pawlik, L., 2013. The role of trees in the geomorphic system of forested hillslopes — a review. Earth Sci. Rev. 126, 250–265.

Porder, S., Payton, A., Vitousek, P.M., 2005a. Erosion and landscape development affect plant nutrient status in the Hawaiian Islands. Oecologia 142, 440–449.

Porder, S., Asner, G.P., Vitousek, P.M., 2005b. Ground-based and remotely sensed nutrient availability across a tropical landscape. Proc. Natl. Acad. Sci. 102, 10909–10912.

Porder, S., Vitousek, P.M., Chadwick, O.A., Chamberlain, C.P., Hilley, G.E., 2007. Uplift, erosion, and phosphorus limitation in terrestrial ecosystems. Ecosystems 10, 158–170.

Portenga, E.W., Bierman, P.R., 2011. Understanding Earth's eroding surface with ^{10}Be. GSA Today 21, 4–10.

Quan, C., Han, S., Torsten, U., Zhang, C., Yu-Sheng, C.L., 2013. Validation of temperature-precipitation based aridity index: paleocliamte implications. Palaeogeogr. Palaeoclimatol. Palaeoecol. 386, 86–95.

Rasmussen, C., Brantley, S., Richter, D., de, B., Blum, A., Dixon, J., White, A.F., 2011. Strong climate and tectonic control on plagioclase weathering in granitic terrain. Earth Planet. Sci. Lett. 301, 521–530.

Reed, S., Amundson, R., 2012. Using LIDAR to model Mima mound evolution and regional energy balances in the Great Central Valley, California. In: Horwath Burnham, J.L., Johnson, D.L. (Eds.), Mima Mounds: The Case for Polygenesis and Bioturbation. Geological Society of America Special Paper 490 (http://dx.doi.org/10.1130/2012.2490(01)).

Riebe, C.S., Kirchner, J.W., Finkel, R.C., 2004. Erosional and climatic effects on long-term chemical weathering rates in granitic landscapes spanning diverse climate regimes. Earth Planet. Sci. Lett. 224, 547–562.

Roering, J.J., 2008. How well can hillslope evolution models "explain" topography? Simulating soil transport and production with high-resolution topographic data. Geol. Soc. Am. Bull. 120, 1248–1262.

Roering, J.J., Kirchner, J.W., Dietrich, W.E., 1999. Evidence for nonlinear, diffusive sediment transport on hillslopes and implications for landscape morphology. Water Resour. Res. 35, 853–870.

Roering, J.J., Marshall, J., Booth, A.M., Mort, M., Jin, Q., 2010. Evidence for biotic controls on topography and soil production. Earth Planet. Sci. Lett. 298, 183–190.

Sanderman, J., Amundson, R.G., Baldocchi, D.D., 2003. Application of eddy covariance measurements to the temperature dependence of soil organic matter mean residence time. Glob. Biogeochem. Cycles 17. http://dx.doi.org/10.1029/2001GB001833.

Stockmann, U., Minasny, B., McBratney, A.B., 2014. How fast does soil grow? Geoderma 216, 48–61.

Trumbore, S.E., Chadwick, O.A., Amundson, R., 1996. Rapid exchange of soil carbon and atmospheric CO_2 driven by temperature change. Science 272, 393–396.

Vitousek, P.M., Chadwick, O.A., Crews, T.E., Fownes, J.H., Hendricks, D.M., Herbert, D., 1997. Soil and ecosystem development across the Hawaiian Islands. GSA Today 7, 1–8.

von Blanckenburg, F., 2006. The control mechanisms of erosion and weathering at basin scale from cosmogenic nuclides in river sediment. Earth Planet. Sci. Lett. 242, 224–239.

Walker, T.W., Syers, J.K., 1976. The fate of phosphorus during pedogenesis. Geoderma 15, 1–19.

White, A.F., Brantley, S.L., 2003. The effect of time on the weathering of silicate minerals: why do weathering rates differ in the laboratory and field? Chem. Geol. 202, 479–506.

White, A.F., Schulz, M.S., Vivit, D.V., Blum, A.E., Stonestrom, D.A., Anderson, S.P., 2008. Chemical weathering of a marine terrace chronosequence, Santa Cruz, California I:

interpreting rates and controls based on soil concentration–depth profiles. Geochim. Cosmochim. Acta 72, 36–68.

Wilkinson, B.H., McElroy, B.J., 2007. The impact of humans on continental erosion and sedimentation. Geol. Soc. Am. Bull. 119, 140–156.

Yoo, K., Mudd, S.M., 2008. Towards process-based modeling of geochemical soil formation across diverse landforms: a new mathematical framework. Geoderma 146, 248–260.

Yoo, K., Amundson, R., Heimsath, A.M., Dietrich, W.E., 2005a. Erosion of upland hillslope soil organic carbon: coupling field measurements with a sediment transport model. Glob. Biogeochem. Cycles 19, GB3003. http://dx.doi.org/10.1029/2004GB002271.

Yoo, K., Amundson, R., Heimsath, A.M., Dietrich, W.E., 2005b. Process-based model linking pocket gopher (*Thomomys bottae*) activity to sediment transport and soil thickness. Geology 33, 917–920.

Yoo, K., Amundson, R., Heimsath, A.M., Dietrich, W.E., 2006. Spatial patterns of soil organic carbon on hillslopes: integrating geomorphic processes and the biological C cycle. Geoderma 130, 47–65.

Yoo, K., Amundson, R., Heimsath, A.M., Dietrich, W.E., Brimhall, G.H., 2007. Integration of geochemical mass balance with sediment transport to calculate rates of soil chemical weathering and transport on hillslopes. J. Geophys. Res. 112. http://dx.doi.org/10.1029/2005JF000402.