Sediment transport mechanisms on soil-mantled hillslopes

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ABSTRACT

Landscape evolution is modeled widely using a simple creep law for complex processes of sediment transport. Here, field data show how a new transport model, combined with an exponential soil production law, better captures spatial variations of soil thickness on hillslopes. We combine parameterizations of simple and depth-dependent creep with overland flow to predict soil thickness and suggest how soil distribution evolves in response to climatic and tectonic forcing. We present an empirical expression for the response time of the system to external forcing that shows strong dependence on relief and is independent of soil production rate. We suggest that this parameterization may be used to quantify upland carbon storage and removal and predict impacts of deforestation or rapid climatic changes.

Keywords: soil dynamics, transport, dating, geomorphology, landscape evolution.

INTRODUCTION

A proper parameterization of the transport laws operating on soil-mantled landscapes has basic implications for predicting how landscapes evolve and for understanding better how human land management can affect the landscape. Many landscape evolution models have used a linear sediment transport law (Anhert, 1967; Willgoose et al., 1991; Kooi and Beaumont, 1994; Dietrich et al., 1995; Howard, 1997; Braun and Sambridge, 1997), which sets the transport flux equal to a linear function of topographic slope. This ''simple creep'' law has its origins in pioneering studies of convex hillslopes by Davis (1892) and Gilbert (1909). There is some field evidence for a linear transport law (McKean et al., 1993; Small et al., 1999), which leads to a simple analytical solution for mass balance (Dietrich et al., 1995), but observations suggest that landsliding, depth-dependent creep, and overland flow contribute to soil transport (Carson and Kirkby, 1972; Selby, 1993; Montgomery and Dietrich, 1995; Hovius et al., 1997; Prosser and Rustomji, 2000). Recently, nonlinear transport laws have been used to capture landsliding (Anderson and Humphrey, 1994; Anderson, 1994; Howard, 1994, 1997) or explain the planar topography on steep slopes (Roering et al., 1999).

Soil transport flux depends on soil production. Here, we use soil production rates determined from concentrations of ^{10}Be and ^{26}Al (Heimsath et al., 2000), as well as high-resolution measurements of soil depth and topographic form to calibrate a parameterization of soil-mantled landscape evolution that combines soil transport by simple creep, overland flow, and depth-dependent viscous creep. Measurements of soil production rates from the Bega Valley, southeastern Australia, from 53 m/m.y. with no soil cover to 7 m/m.y. under 100 cm of soil, show that production declines exponentially with increasing soil thickness (Heimsath et al., 2000):

$$P = P_0 \exp\left(-\frac{h}{h_0}\right), \tag{1}$$

where $P_0 \approx 50$ m/m.y. and $h_0 = 0.5$ m. Similar results were found in northern California (Heimsath et al., 1999).

The simple creep law can be tested by examining the dependence of soil thickness on surface curvature. Assuming, at steady state, that soil production is balanced by local transport dominated by simple soil creep, then

$$K_D \nabla^2 z + \kappa P = 0, \tag{2}$$

where $\nabla^2 z$ is surface curvature, K_D is a constant, κ is the ratio of soil density to bedrock density, and P is the soil production rate (Heimsath et al., 1997). It follows from equations 1 and 2 that curvature should decrease exponentially with soil thickness, and that this should hold only for convex areas of the landscape—i.e., that no concave (or convergent) areas can develop on a steady-state landscape. Field data from northern California and southeastern Australia clearly indicate that this is not the case (Heimsath et al., 1997, 2000); soil thickness appears to increase linearly from divergent to convergent topography. Two options would resolve this apparent paradox: either the studied landscapes have not reached steady state or other mechanisms beside simple creep contribute to soil transport. Recent data from the Australian site support the steady-state assumption (Heimsath et al., 2000). Here we propose a new parameterization for the transport of soil that combines the effects of simple soil creep, depth-dependent creep, and transport by overland flow. Our parameterization reproduces the observed relationship between soil thickness and curvature and, combined with measurements of soil production rate, it indicates the time required for soil distribution to reach a balance between production and transport.

BASIC EQUATIONS

On hilly landscapes, with no eolian inputs and without significant losses by dissolution, soil thickness is controlled by the balance between local soil production from the underlying bedrock and removal or accumulation by erosion or deposition. Under these conditions, the continuity equation can be expressed as an evolution equation for soil thickness, h:

$$\frac{\partial h}{\partial t} = -\nabla \cdot Q + \kappa P, \tag{3}$$

where Q is downhill soil flux, P is soil production (i.e., the rate at which the soil-rock [or soil-saprolite] interface migrates downward by weathering), κ is the ratio of soil density to rock density, t is time, and ∇ is the divergence operator (Carson and Kirkby, 1972; Dietrich et al., 1995). The rate of lowering of the ground surface, $\partial z/\partial t$, is related to the rate of change of soil thickness by

$$\frac{\partial z}{\partial t} = -\left(P - \frac{\partial h}{\partial t}\right)\sec \alpha, \tag{4}$$

where α is the local slope angle.

Considering that simple creep is only one of several processes that contribute to soil transport (Carson and Kirkby, 1972; Donohue, 1986), we propose a parameterization for downhill soil flux involving three possible processes: simple creep, depth-dependent viscous creep, and transport by overland flow.

Simple soil creep flux is commonly regarded as operating in a shallow surficial layer and as dependent only on slope (Dietrich et al., 1995):

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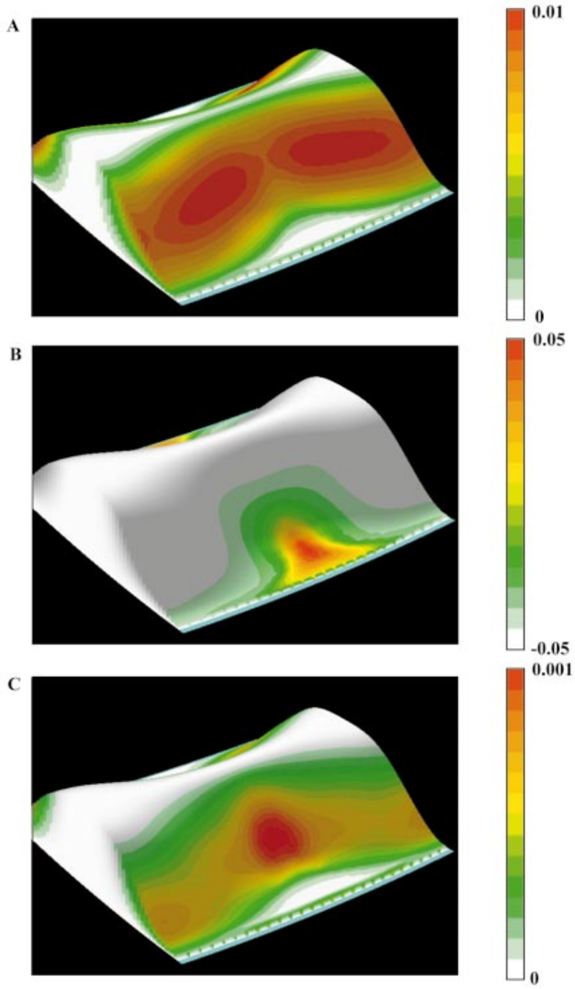

Figure 1. Results of model after 100 k.y., in form of (A) soil thickness (in centimeters), (B) surface curvature (per meter) calculated at resolution of numerical grid (10 m), and (C) soil production rate (in meters per year). Transport parameter values (see text) are: $K_D = 3 \times 10^{-3}$ $m^2 \cdot yr^{-1}$, $K_V = 3 \times 10^{-3}$ $m^{2-m} \cdot yr^{-1}$, $K_W = 3 \times 10^{-9}$ $m^{2-3k} \cdot yr^{k-1}$, $m = 1.67$, $n = 0.5$, $k = 1.67$, $p = 1.3$, $P_0 = 53 \times 10^{-6}$ $m \cdot yr^{-1}$, $h_0 = 0.5$ m. The precipitation rate is assumed to be uniform at 0.5 $m \cdot yr^{-1} \cdot \kappa$; the ratio of soil to bedrock density is 0.5. The time step used in the explicit time integration is 10 yr.

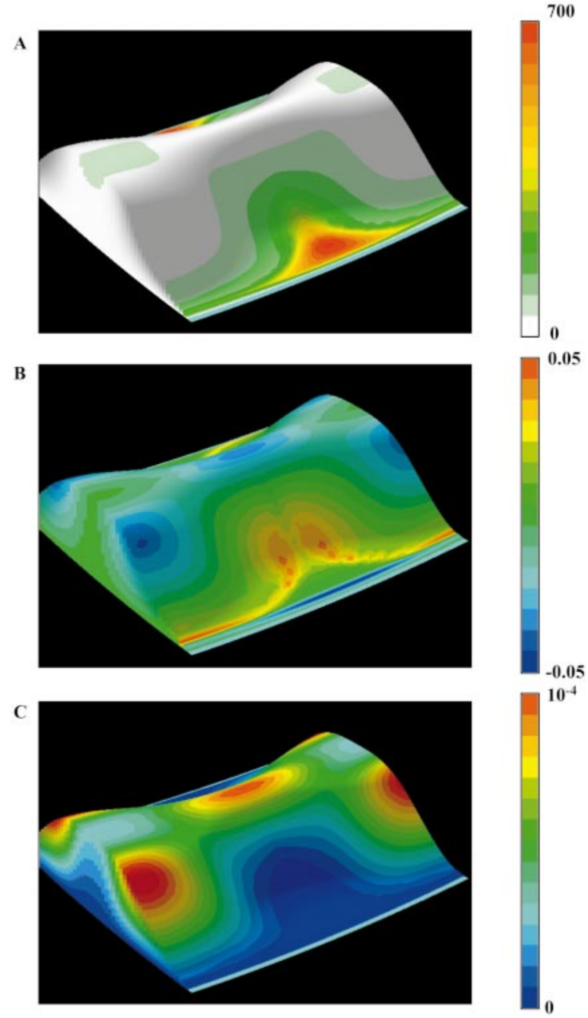

Figure 2. Contributions to total local soil flux (in $m^2 \cdot yr^{-1}$) from (A) linear creep, (B) depth-dependent creep, and (C) overland flow as computed by model after 100 k.y. Parameters are same as for Figure 1.

$$Q_D = -K_D S. \tag{5}$$

Depth-dependent creep formulations, on the other hand, are usually based on Bingham rheology (Selby, 1993), although experiments suggest that the yield strength for soils is very low (Van Asch et al., 1989), implying that strain rate is simply proportional to stress. Thus, we adopt a general viscous flow model for depth-dependent creep:

$$Q_V = -K_V h^m S^n. \tag{6}$$

Without experimental data for m and n, we initially adopt values from Manning's equation for liquid flow ($m = 1.67$ and $n = 0.5$), expecting that these are likely to be altered in the future. Assuming that $m = 1.67$ in equation 6 is consistent with a velocity profile of the form: $v(z) \propto Z^{2/3}$, which is compatible with recent measurements (Clarke et al., 1999).

A similarly general expression for soil transport by overland flow is given by Moore and Burch (1986):

$$Q_W = -K_W A^k S^p, \tag{7}$$

where A is upstream catchment area and the exponents k and p have values between 1 and 2 (Prosser and Rustomji, 2000). The total soil flux is the sum of all three fluxes.

SOLUTION METHOD

Combining the individual expressions for Q with the exponential soil production function (equation 1) leads to a nonlinear partial differential equation (PDE) that relates soil thickness to local slope, curvature, and drainage geometry. This equation is solved by dividing the landscape into small hexagonal cells of locally uniform soil thickness, h_i. Integrating both sides of the PDE over the surface of each cell, Ω, and, making use of Gauss's theorem, we can write:

$$-\int \nabla \cdot Q \, d\Omega = -\oint n \cdot Q \, d\Sigma \approx -\sum Q_i^1 l_i, \tag{8}$$

where Q_i^1 is the contribution to the flux Q through cell side i of length l_i. This formulation transforms the two-dimensional mass conservation problem into an evolution equation for soil thickness expressed in terms of a weighted sum of one-dimensional fluxes:

$$\frac{\partial h_i}{\partial t} = -\Omega \sum_{i=1}^{6} Q_i^1 l_i + \kappa P_0 \exp\left(-\frac{h_i}{h_0}\right). \tag{9}$$

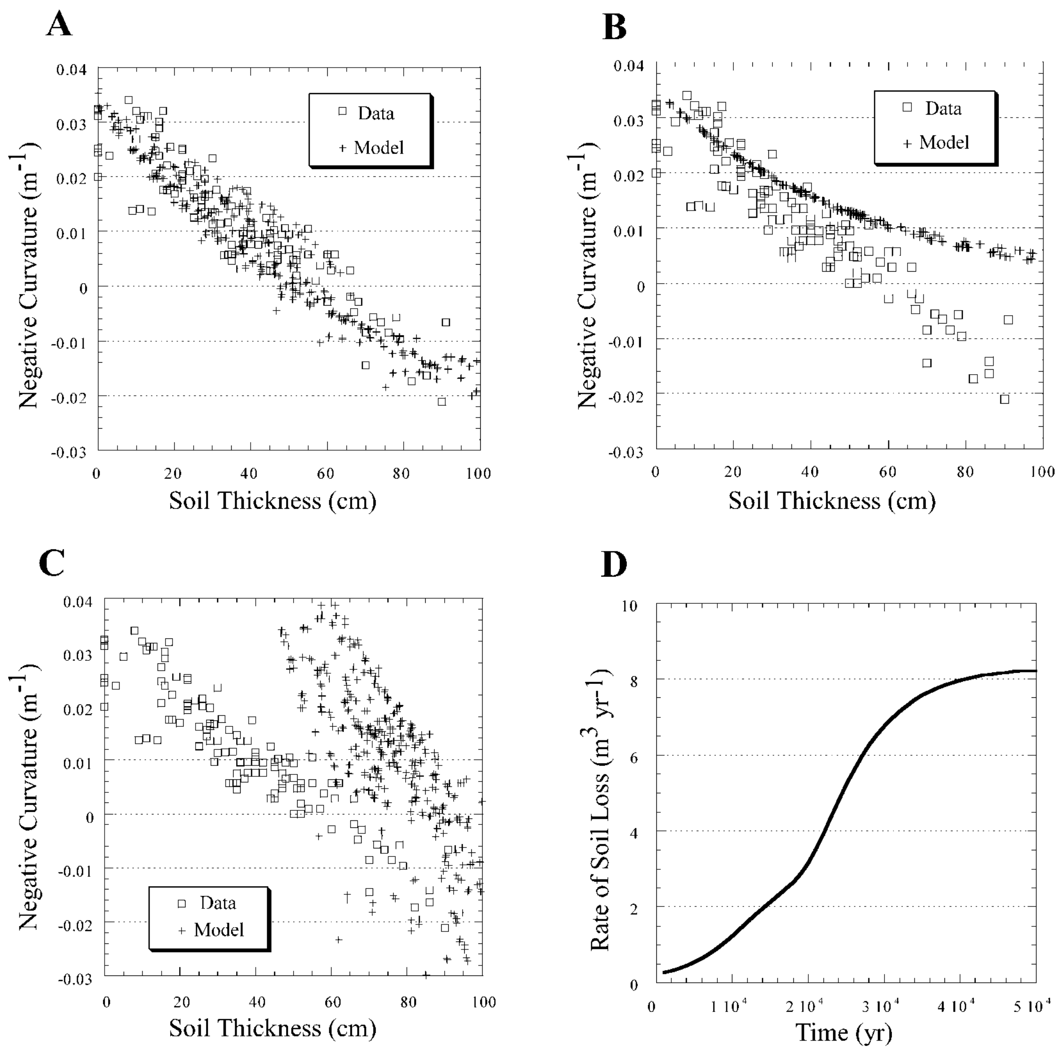

Figure 3. Computed curvature-thickness curves compared to observations at Bega Valley site (Heimsath et al., 2000) by using (A) model parameters given in Figure 1 caption (see text), (B) with K_V and K_W both set to zero, and (C) K_D and K_W set to zero. Rate of soil loss through upper and lower open boundaries is given in D for set of parameters given in Figure 1 caption.

Time evolution is treated explicitly—i.e., $h_i(t + \Delta t) = h_i(t) + \Delta t \partial h_i / \partial t$—and the time step Δt is taken small enough (10 yr) to ensure stability.

SYSTEM BEHAVIOR

We tested our new parameterization by solving a simple generic problem. We constructed a smooth bedrock surface on which soil thickness is initially assumed to be nil and allowed to evolve in accordance with equation 9. The initial geometry of the synthetic hill is a large-amplitude, symmetrical cosine bell on which a smaller amplitude trough has been added to create regions of varying curvature in both the *x* and *y* directions. The hill is 200 m high and extends 300 m in the *x* direction and 600 m in the *y* direction. This geometry was chosen to generate surface slope and curvature values similar to those observed in the Bega Valley site surveyed by Heimsath et al. (2000).

A steady-state soil distribution is reached after approximately 50 000 yr (Fig. 1A). Soil thickness is maximum at the base of the hill and thinnest at the top of the hill and along the crest of the steep ridges flanking the hill on either side. The regions of minimum soil cover are characterized by maximum negative curvature (Fig. 1B) and soil production rate (Fig. 1C). The regions of maximum soil production also have the most rapid surface lowering. The triangular region of maximum soil accumulation at the base of the hill (Fig. 1A) has a relatively flat surface, sloping gently toward the model boundary. Topography in the transition zone between the basal regions and the hill slopes is characterized by a marked change in slope and a positive curvature (Fig. 1B).

SEGREGATION OF TRANSPORT MECHANISMS

After 50 000 yr, soil distribution has evolved to favor spatial segregation of transport mechanisms. This segregation results from the multiprocess parameterization: simple creep dominates in regions of maximum slope (equation 5), depth-dependent creep dominates in regions of maximum soil thickness (equation 6), and transport by overland flow dominates in regions of drainage convergence (equation 7) (Fig. 2).

Our results confirm those of previous investigations (such as the one-dimensional model described in Kirkby, 1992) that even on the scale of a single hill (i.e., a few hundred meters), three common transport mechanisms can all be activated simultaneously, but each mechanism is dominant in a separate region of the landscape, depending on the local combination of slope, soil thickness, and drainage area. It is important that the partitioning of transport mechanisms is not the result of predetermined threshold parameters (as assumed in Howard, 1994; Montgomery and Dietrich, 1995).

CURVATURE-THICKNESS RELATIONSHIP

We compared curvature-thickness relationships from a range of model runs to observations from the Bega Valley, southeastern Australia (Heimsath et al., 2000). By combining our three transport parameterizations, an excellent fit to the data is obtained over the complete range of surface curvatures (Fig. 3A). However, the results are sensitive to the relative weightings of the transport parameters (Fig. 3, B and C). If simple creep is the only transport mechanism, then the regions of finite soil thickness and positive curvature cannot be matched (Fig.

3B), but if viscous creep is dominant, then the areas of relatively thin soil cover are missed (Fig. 3C). Our results therefore demonstrate that, if our process parameterizations are accepted and if the production function is well delineated by cosmogenic data, then field measurements of curvature and soil thickness can be used to limit the value of the various transport parameters.

TRANSIENT BEHAVIOR

Using the transport parameters that yield the best fit with the observed curvature-thickness data (Fig. 3A), we investigated the transient behavior of the system. After 50 000 yr of evolution, soil thickness becomes constant in all parts of the landscape. This situation does not correspond, however, to a ''true geomorphic steady state'' as soil production and transport lead to ongoing lowering of the hill, as expressed in equation 4. During the first 20 000 yr of model evolution, accumulation is maximum in regions of topographic convergence and/or maximum slope, approximately one-third of the way down the hillside, after which the region of maximum soil accumulation shifts toward the bottom of the hill. This progression is seen in the computed rate of soil loss from the landscape (Fig. 3D), which shows an inflection point at about 20 000 yr, corresponding to the transition to a situation where depth-dependent creep acting across a relatively small area at the foot of the hill is able to remove from the landscape all the sediment transported to that region by linear creep and overland flow from the other areas of the hill.

The response time of the system, τ (i.e., the time taken for soil thickness to reach steady-state following a change in boundary conditions or transport coefficients such as triggered by climate variation) is in fact a function of the geometry (height, H, and extent, L) of the underlying bedrock topography as well as the parameters K_D and K_V (K_W plays a much less important role). We used the results of a large number of model runs in which all parameters were varied to derive an empirical expression for τ:

$$\tau \propto L^{5/3} H^{-2/3} K_D^{-1/3} K_V^{-2/3}, \tag{10}$$

which implies that both mechanisms (simple creep and depth-dependent creep) contribute to the removal of soil from the landscape.[1] However, their roles are clearly different: simple creep transports soil from the source areas along the slopes of the hill, whereas depth-dependent creep transports soil toward base level (typically a river channel) across the concave or low-gradient areas at the base of the hill. It is important to note that the response time τ appears independent of the rate of soil production P_0 or the depth scale h_0, which should be the case for hilly soil-mantled landscapes that are transport limited. Equation 10 therefore highlights the importance of relief and length scale in determining how quickly soil distribution reaches steady state on a given landscape. The values of K_D and K_V are determined by bedrock type, vegetation, and other environmental factors. Consequently, the relative response time of the various components of a landform could be inferred from estimates of the distribution of relief as a function of length scale (Chase, 1992).

Finally, we propose that our estimate of the value of the time scale τ provides a unique measure of the response time of soil-mantled landscapes to environmental changes by either local factors, such as tree clearing, or global factors, such as rapid climate changes. We also suggest that this study should lead to a reassessment of how soil production and transport can be incorporated in large-scale surface process models, which are used, for example, to study the interplay between tectonics and erosion. We also suggest that by quantifying the rate of transport and storage of soil on upland landscapes, as we have done here, a more accurate estimate of terrestrial carbon sequestration may be achieved. Namely, detailed measurements of soil carbon (Stallard, 1998) combined with soil production and transport rates for hilly landscapes can extend our understanding of carbon storage across a large part of Earth's surface.

[1]The exact form of τ depends on the value of the exponents m and n in equation 6.

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