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11 Landscape Dissection and Drainage Area–Slope Thresholds

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ABSTRACT

Development of the distinct ridges and valleys that define landscape dissection may be controlled by either a spatial competition between slope-dependent and area-dependent sediment transport processes or by exceedence of a spatially and temporally variable channel initiation threshold. Field observations indicate that channel heads in many humid, soil-mantled environments represent a change in sediment transport processes, rather than a change in process dominance. In environments where sediment transport by sheetwash and rainsplash occur, however, the development of defined incision may reflect the spatial dominance of overland flow over rainsplash that infills incipient channels. Simple models of channel initiation processes predict drainage area–slope thresholds for channel initiation by overland flow and landsliding that are similar to observed relations in regions where these processes control channel initiation. A drainage area–slope threshold for soil saturation also can be used to define the spatial extent of saturation overland flow. Plotted together on a graph of drainage area versus slope these thresholds predict both a region of the landscape dominated by diffusive sediment transport that corresponds to undissected hillslopes and a region of generally non-erosive overland flow corresponding to unchanneled valleys where occasional evacuation of stored sediment occurs, and regions of the landscape where sediment transport occurs by overland flow and landsliding. Over geologic time scales, the short-term variability of these thresholds may result in valley incision upslope of the mean channel head location. Maintenance of the upslope ends of the valley network thus may reflect either the spatial signature of temporal variability of channel head locations or a spatial transition in the dominant sediment transport processes. In either case, landscape dissection is sensitive to both long-term change in the controls on channel initiation and short-term variance in channel head locations.

INTRODUCTION

The degree to which a landscape is dissected traditionally is described by the drainage density, the total stream length divided by drainage area. Unfortunately, the extent of the

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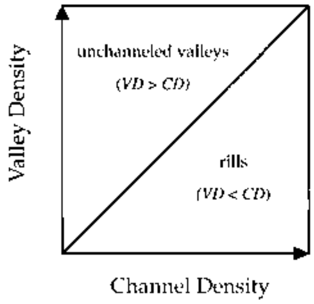


Figure 11.1 Schematic illustration of potential relations between channel and valley density. Landscapes in which the valley density exceeds the channel density will have unchanneled valleys. Landscapes in which the channel density exceeds the valley density have channels, or rills, developed on convex or planar hillslopes

stream network is variable (Gregory & Walling, 1968) and thus there is not a unique drainage density for a given landscape. Less ambiguous descriptions of landscape dissection are provided by the extent of valleys (areas of topographic convergence) and channels (avenues of sediment transport within definable banks). Valley and channel densities are morphologically and spatially distinct, as the extent of channel and valley networks only rarely coincide (Figure 11.1). Unchanneled valleys often separate channel heads from drainage divides in humid, soil-mantled landscapes, whereas in arid or disturbed landscapes small channels or rills often extend beyond the confines of the valley network and onto the intervening hillslopes. Channel and valley densities thus provide a less ambiguous description of landscape dissection than does the drainage density. Understanding the process of landscape dissection into distinct hillslopes and valleys depends, however, on establishing relations between the controls on valley and channel densities, as erosion by streamflow, dissolution, and debris flows incises valleys into the landscape.

In soil-mantled landscapes the distinction between valleys and ridges is pertinent, as there is a scale beyond which smaller valleys are not present (Dietrich *et al.*, 1987; Montgomery & Dietrich, 1992). However, there may not be such a scale in rocky landscapes in which sediment transport is limited by the rate at which weathering processes generate material available for transport. In such landscapes, diffusive processes may be essentially inactive and incisive transport processes may act over virtually all scales. Consequently, the present discussion is restricted to transport-limited landscapes.

Most landscape evolution models assume either what we term the Gilbert hypothesis or the Horton hypothesis about processes controlling the development of landscape dissection. In a series of papers, Gilbert (1877, 1909) argued that ridge erosion is controlled by slope-dependent processes that give rise to convex slopes, whereas valley development is a product of both runoff and slope (see review in Montgomery, 1991). In the Gilbert hypothesis then, ridge and valley topography arises from the varying spatial dominance of diffusive (slope-dependent) and incisive (slope and discharge-dependent) processes. Smith & Bretherton (1972) subsequently demonstrated mathematically, for the particular case of a landscape undergoing steady state erosion, that lateral perturbations would tend to grow (and presumably form valleys) where incisive processes dominate over diffusive processes. Instead of the dominance of one process over another, Horton (1945) proposed that erosion and subsequent valley development only occurs where a threshold of ground surface resistance to overland flow is exceeded. He formulated a quantitative theory that predicted this threshold distance, x_c, downslope from the drainage divide. Subsequent models which assumed that a threshold of erosion must be crossed for channels to develop (e.g., Schaefer, 1979; Dietrich *et al.*, 1986; Montgomery & Dietrich, 1988, 1989; Montgomery, 1991; Willgoose *et al.*, 1991) are similar in spirit to the Horton hypothesis. Montgomery and Dietrich (1992) further argued that the variance in the position of channel head locations may define the headward extent of valleys. This observation suggests a resolution of the Gilbert and Horton hypotheses in which the Gilbert hypothesis is most relevant to valley maintenance over geologic time scales (i.e., 10^4 to 10^6 yr) and the Horton hypothesis is most relevant to channel initiation and sediment transport over shorter geomorphic time scales (i.e., 10^2 to 10^3 yr).

The extent of both channels and valleys reflect the interaction of erosive processes, the resistance of the ground surface to those processes, and processes acting to infill incipient incision. Channel heads may be defined on the basis of the upslope extent of sediment transport concentrated within clearly demarkated banks. The extent of valleys, on the other hand, may be defined by the upslope extent of convergent topography. Incisive processes act to create topographic depressions, generate relief, and incise hillslopes, while diffusive processes tend to infill topographic depressions, degrade relief, and round hillslopes. Consequently, either the transition from convex to concave slopes or from convergent to divergent topography has been considered to correspond to the transition from diffusive to incisive transport, in accordance with the Gilbert hypothesis.

Several previous workers in effect equated this transition with both channel and valley heads (Smith & Bretherton, 1972; Kirkby, 1980, 1987; Tarboton *et al.*, 1992). This is consistent with the Gilbert hypothesis and involves the testable assumptions that above the channel head diffusional sediment transport prevents incision by infilling incipient channels, while below the channel head incisive sediment transport processes overwhelm diffusional transport and maintain a valley. In general, sediment transport at any point on the landscape may be expressed as the sum of diffusive and incisive components, which may be expressed as

$$q_s = \mathrm{f}(S) + \mathrm{f}(q,S) \tag{1}$$

where q_s is the sediment flux, S is the slope, and q is discharge. The first term on the right-hand side of equation (1) represents sediment transport by diffusive processes and the second term represents transport by incisional (or advective (Loewenherz, 1991)) processes. This model, first presented by Smith and Bretherton (1972), assumes that channels and valleys correspond to areas where $q_s/q < \delta q_s/\delta q$. In this case, discharge-dependent sediment transport occurs upslope of the convex/concave slope transition, but the tendency to incise will be overwhelmed by diffusive transport (Dunne, 1980).

The Horton hypothesis requires that channel and valley maintenance results from exceedence of an erosional threshold. Although Horton originally maintained that no erosion occurs up slope of x_c, the Horton hypothesis can be extended to include the more general case in which sediment transport above the channel head occurs only by slope-dependent processes, whereas below the channel head sediment transport also is a function of discharge, or drainage area. Thus, according to the Hortonian model the channel head is a point where sediment transport processes change, rather than a transition in the dominant transport process, as held under the Gilbert hypothesis. Dietrich & Dunne (1993) expressed this condition as

$$q_s = \mathrm{f}(S) \qquad x < x_c \tag{2a}$$

above the channel head, whereas at and downslope of the channel head sediment transport may be described as

$$q_s = f(S) + f(q,S) \qquad x \geqslant x_c \tag{2b}$$

where x_c is the threshold for sediment transport by incisive processes. A threshold control on channel initiation implies that channel heads need not correspond to the transition from convex and concave slopes. Rather, at any given point in time channeled transport may start some distance downslope from drainage divides regardless of slope curvature. The channel head position may fluctuate up and downslope in response to climatic, land-use, or even seasonal variation, and the extent of valleys (and the consequent transition from convex to concave topography) will reflect a balance between periodic headward extention of the channel head and relatively continuous diffusive processes. These hypotheses for channel initiation and valley maintenance involve testable assumptions about sediment transport and channel initiation processes, which are discussed in the sections below.

CHANNEL INITIATION

While there are a number of mechanisms for initiating channels and maintaining channel heads, we restrict our discussion to landscapes in which channel initiation is dominated by overland flow (Horton, 1945; Kirkby & Chorley, 1967), seepage erosion (Dunne, 1980; Montgomery, 1991), and landsliding (Dietrich *et al.*, 1986; Montgomery & Dietrich, 1989). Channels initiated and maintained by these processes tend to have characteristic forms, which can be grouped into gradual and abrupt morphologies (Montgomery & Dietrich, 1988, 1989), although a more elaborate classification scheme has been offered by Dietrich & Dunne (1993). Within any given area, several channel initiation processes may be active, reflecting variations in slope, soil type and thickness, and vegetation type and density.

Processes controlling each of these channel initiation mechanisms can be described as threshold phenomena. Channel initiation by overland flow where diffusional dampening is effective (see Dunne (1980) and Loewenherz (1991)) may be assumed to occur wherever the flow chronically exerts a boundary shear stress in excess of critical for initial motion. This is the assumption explicitly proposed by Horton (1945) and subsequently adopted by others (i.e., Schaefer, 1979; Montgomery, 1991; Willgoose *et al.*, 1991; Dietrich *et al.*, 1992). Channel initiation by seepage erosion may be considered to be controlled by either seepage gradients capable of entraining surficial materials (Dunne, 1990) or, in cohesive material, by sufficient drainage area to support channel head spalling and overland flow to remove failed material (Montgomery, 1991). Channel initiation by landsliding is controlled by critical pore pressures in the colluvium and thus can be modelled as occurring when sufficient throughflow converges at a location to generate pore pressures necessary to trigger slope instability (Dietrich *et al.*, 1986; Montgomery & Dietrich, 1988, 1989). Consequently, the channel initiation threshold in equation (2) may be formulated as a function of either the overland flow discharge necessary to overcome the erosional resistance of the ground surface, the head gradient necessary for seepage to entrain surface materials, subsurface discharge sufficient to

Figure 11.2 Small channels, or rills, developed on a convex hillslope in a badlands in New Mexico. Note the film canister for scale. The view is from the drainage divide looking downslope

saturate the soil thickness and generate erosive overland flow, or the pore pressure necessary to initiate slope instability. Each of these relations allows models for channel initiation to be formulated in terms of thresholds related to the contributing drainage area. We propose that such a threshold-based view of channel initiation is most representative of field conditions in many landscapes.

Simple analytical models of channel initiation may be developed by assuming a steady state rainfall and associating channel initiation with the exceedence of erosional thresholds (see also Dietrich *et al.*, 1992). Drainage area may be considered to be a surrogate for surface or subsurface runoff in each of the cases discussed above. Models for channel initiation by Horton overland flow, saturation overland flow, and landsliding, as well as for channel head advance by sloughing, predict different relations between the contributing drainage area and the local slope at the channel head. The limitations of using the steady state assumption for modelling a transient process are obvious, but this assumption does allow useful conceptual relations to be drawn that are not possible for the transient case. These relations can be compared with published data for channel head locations to allow prediction of trends in landscape response to disturbance and environmental change and to examine linkages between the development of channel and valley networks. Perhaps equally important, none of the simple models that follow attempt to explain the size of a channel head. Hence these models are useful only in illustrating controls on where erosion due to runoff would tend to occur for various erosional processes.

Horton Overland Flow

Horton overland flow controls the formation of gradual channel heads in semi-arid to arid landscapes and often results in the development of rills (channels that extend onto hillslopes) in arid, poorly vegetated, or disturbed lands (Figure 11.2). Rill development has been argued to be associated with a critical shear stress, a critical shear velocity, or a critical stream power (see review in Slattery & Bryan, 1992). However, sediment transport by overland flow does not always incise channels and may not always act to maintain valleys. For example, sediment transport by sheetflow may occur upslope of the point where diffusional processes dominate incisional processes, and thus an overland flow transport threshold may be exceeded on convex hillslopes. Furthermore, Dunne (1991) showed that if rainfall duration is short relative to the time required to transport sediment off of a hillslope, then even though rills may be incised, the entrained material will be redeposited lower on the slope. In this case, rills function as channels that act to maintain convex slope profiles.

Channel initiation by overland flow may occur under either laminar or turbulent flow regimes. While unchanneled overland flow generally is laminar (see discussions in Dunne & Dietrich, 1980; Reid, 1989), rill flow and rill initiation may be associated with turbulent or transitional conditions and the development of supercritical flow (Slattery & Bryan, 1992). Although flow over bare surfaces may be turbulent (Emmett, 1970), even high-velocity flow over grassy, or well-vegetated surfaces generally is laminar (see Moody diagrams in Dunne & Dietrich (1980) and Reid (1989)). Consequently, we develop models for overland flow erosion exceeding surface resistance by both turbulent and laminar flow.

Assuming a steady state rainfall intensity R over a surface with uniform infiltration capacity, I, the discharge per unit contour length q can be expressed as

$$q = (R - I)a \tag{3}$$

where a is the drainage area per unit contour length. For turbulent open-channel flow, the hydraulic radius is approximately the flow depth. Using the Manning equation to calculate mean flow velocity, we can set the unit discharge equal to the product of the flow velocity and depth, and thus

$$q = (1/n)h^{5/3}S^{1/2} \tag{4}$$

where n is the Manning resistance coefficient, h is the flow depth, and S is the water surface slope. The critical shear stress for incipient motion is determined by soil properties. A depth–slope product sufficient to generate a basal shear stress greater than the critical shear stress is required for sediment transport:

$$\tau_{cr} = \rho_w g(hS)_{cr} \tag{5}$$

where ρ_w is the density of water and g is gravitational acceleration. Thus the critical discharge (q_{cr}) can be found by substituting (5), rearranged in terms of h, into (4):

$$q_{cr} = \frac{\tau_{cr}{}^{5/3}}{(\rho_w g)^{5/3} n S^{7/6}} \tag{6}$$

Equating (6) and (3) and solving for the critical drainage area per unit contour length a_{cr} necessary for erosion by overland flow yields

$$a_{cr} = \frac{\tau_{cr}{}^{5/3}}{(R - I)(\rho_w g)^{5/3} n S^{7/6}} \tag{7}$$

which describes an inverse loglinear source area–slope relation in which the contributing area per unit contour length for channel initiation by turbulent Horton overland flow is approximately proportional to the inverse of the local slope and is a function of surface roughness, critical shear stress, and the difference between rainfall intensity and infiltration capacity. According to equation (7), an increase in the critical shear stress or infiltration capacity of the soil will result in larger source areas, whereas an increase in rainfall intensity, surface roughness, or bulk density will result in smaller source areas (Figure 11.6A).

The turbulent flow assumption, however, may not be generally correct for channel initiation by overland flow. Moody diagrams for overland flow over grass surfaces (Dunne & Dietrich, 1980; Wilson, 1988; Reid, 1989) suggest laminar-type flow to Reynolds numbers in excess of 30 000. Consequently, even though rapid overland flow along hollow axes may appear turbulent, a laminar flow model may be more appropriate, especially where a well-vegetated grass surface protects the ground from surface runoff.

A model for channel initiation by laminar overland flow also may be constructed based on a critical shear stress argument. Assuming that a channel is initiated when a critical boundary shear stress is exceeded, we may again express the channel initiation criterion in terms of equation (5). Flow velocity u may be expressed as

$$u = (2\,ghS/f)^{0.5} \tag{8}$$

where f is a dimensionless friction factor, which itself may be expressed as

$$f = k\nu/q \tag{9}$$

where k is a dimensionless surface roughness coefficient, ν is the kinematic viscosity, and q is the unit discharge. Note that the unit discharge is equal to the product of the average flow velocity and depth, or that

$$q = uh \tag{10}$$

Consequently, substituting (10) into (9) and then (5), rearranged in terms of h, and (9) into (8) allows an expression for the critical discharge:

$$q_{\mathrm{cr}} = \frac{2\tau_{\mathrm{cr}}^{3}}{k\nu\rho_{\mathrm{w}}^{3}g^{2}S^{2}} \tag{11}$$

During steady state runoff, the discharge is given by equation (3). Thus combining (11) and (3) results in an expression for the critical contributing area as a function of slope where

$$a_{\mathrm{cr}} = \frac{2\tau_{\mathrm{cr}}^{3}}{(R-I)k\nu\rho_{\mathrm{w}}^{3}g^{2}S^{2}} \tag{12}$$

Equation (12) predicts an inverse loglinear relation where the contributing area is inversely proportional to the square of slope (Figure 11.6B).

Differences in the models for channel initiation for turbulent and laminar overland flow indicate that a change in hillslope flow hydraulics could have a pronounced impact on the extent of the channel network. The critical drainage area for turbulent overland flow is approximately proportional to the square of the critical shear stress, whereas the critical drainage area for laminar overland flow is a function of the cube of the critical shear stress. Moreover, channel initiation by turbulent overland flow is inversely proportional to the slope, whereas channel initiation by laminar overland flow is inversely dependent on the square of slope. The difference in the critical area between these two scenarios is expressed by the ratio of equations (12) and (7):

$$\frac{a_{\mathrm{crt}}}{a_{\mathrm{crl}}} \propto \frac{S^{5/6}}{\tau_{\mathrm{cr}}{}^{4/3}} \tag{13}$$

where a_{crt} is the critical area for turbulent flow and a_{crl} is the critical area for laminar flow. This ratio indicates that the difference between the source area size under turbulent and laminar flow is approximately proportional to slope and is inversely proportional to the critical shear stress. Thus a change from laminar to turbulent flow, such as might accompany extreme changes in vegetation cover, would be expected to lead to significant expansion of the channel network.

Seepage Erosion and Soil Saturation

Dunne (1990) developed an expression for the critical head gradient for seepage erosion and concluded that for cohesive material seepage gradients would have to be unrealistically high to cause liquefaction and thus initiate a channel. Consequently, seepage erosion is only a viable channel initiation mechanism in cohesionless materials or in materials in which weathering causes pronounced cohesion loss. In vegetated landscapes the root strength of plants provides an effective cohesion to the soil, suggesting that channel initiation by seepage forces is restricted to poorly vegetated landscapes. Abrupt channel heads in cohesive soils on low-gradient slopes, however, may be maintained by spalling of material from the seepage face at the channel head (Figure 11.3) (Bradford & Piest, 1985; Montgomery, 1991). Channel head advance at an abrupt channel head in the Tennessee Valley area of Marin County, California, for example, was correlated with peak piezometric levels at the base of the channel head, but stability calculations suggest that channel head spalling reflects loss of restraining pressure for the pedons exposed on the channel head (Montgomery, 1991). Overland flow over the channel head sufficient to remove the failed material is required to prevent buttressing and allow continued channel head advance, but does not directly contribute to erosion at the channel head. The burrowing activity of fossorial mammals appears to be the dominant sediment transport process upslope of the channel heads in this area (Black & Montgomery, 1991). Overland flow upslope of the channel head was not observed to transport sediment over the channel head and grain-by-grain seepage entrainment of material from the face of the channel head was not observed. Hence controls on the location of such channel heads involve a change in process. Maintenance of these abrupt channel heads may be modeled as dependent upon sufficient drainage area to saturate the soil profile in order to both fail the channel head and provide overland flow

Figure 11.3 Abrupt headcut in the Tennessee Valley area of Marin County, California. Note alcove formed by spalling of the seepage face at left side of headcut. Piezometers above headcut extend 0.5 m above ground surface

runoff to remove the failed material. This mechanism, however, is not a channel initiation mechanism. Disturbance or locally accelerated incision is necessary to develop abrupt channel head morphology (Leopold & Miller, 1956; Reid, 1989; Montgomery, 1991). Once initiated, however, such an abrupt channel head probably will advance until the drainage area becomes insufficient to generate either soil saturation to cause spalling or overland flow to remove the failed material (Montgomery, 1991). This indicates that stable locations for such abrupt channel heads are governed by a contributing drainage area sufficient to saturate the soil profile and generate overland flow. Consequently, for the present discussion we assume that the critical area for 'seepage' erosion occurs when the soil is saturated.

For the case where the infiltration capacity of the soil exceeds the rainfall intensity, but flow does not recharge to a bedrock groundwater table, the subsurface discharge due to flow through the soil is given by

$$q = Ra \tag{14}$$

Further assuming that the hydrologic gradient is equal to the ground slope, Darcy's law can be expressed as

$$q = KzhS = ThS \tag{15}$$

where K is the saturated hydraulic conductivity, z is the soil thickness, h is the proportion of the soil column that is saturated, and T is the transmissivity of the soil profile. Combining (14) and (15) the relative saturation of the soil profile may be expressed as

$$h = Ra/TS \tag{16}$$

For saturated areas $h = 1$ and thus the contributing area per unit contour length necessary to saturate the soil column a_{cr} is given by

$$a_{cr} = TS/R \tag{17}$$

This approach is equivalent to those used by Beven & Kirkby (1979) and O'Loughlin (1986) to estimate the distribution of relative soil moisture in a catchment. Equation (17) shows that with greater rainfall a smaller contributing area per unit contour length is necessary to produce saturation at a given location. Equation (17) predicts a positive linear relation between the drainage area and slope (Figure 11.6C). This saturation threshold essentially defines a limit to potential channel head advance by seepage erosion and gullying.

Saturation Overland Flow

Saturation overland flow controls the formation of gradual channel heads on many low-gradient slopes in humid, soil-mantled landscapes. Gradual channel heads in grasslands of Marin County, California, provide a well-studied example. A three-year monitoring program in this area did not reveal obvious sediment entrainment by overland flow upslope of channel heads even during high discharge events in which significant sediment transport is observed downstream of the channel head (Montgomery, 1991). Instead, sediment transport upslope of these channel heads is dominated by burrowing activity (Black and Montgomery, 1991). This change in sediment transport processes suggests a threshold condition that may be modeled using a critical shear stress. Sediment transport by overland flow is possible only after development of sufficient tractive force to overcome the resistance of the vegetation covering the soil surface, especially where interwoven grass shields the ground surface (Figure 11.4). In such areas, runoff occurs over this grass surface and sediment transport occurs only where runoff breaks through this protective mat. During extreme discharge events the channel head may extend upslope of its typical position and temporarily transport material until the channel is progressively infilled in the interval between extreme discharge events. Fire and grazing also can reduce the resistance of this material and allow upslope extensions of the channel head. Given sufficient time, a threshold-controlled channel head thus may intermittently extend significantly upslope of the typical channel head location.

Kirkby (1987) developed a model for channel initiation by saturation overland flow that predicts a positive source area–slope relation. This model considers erosion to be directly proportional to overland flow. Montgomery & Dietrich (1988), however, reported field data that indicated an inverse source area–slope relation from areas in which saturation overland flow controls channel heads on low-gradient slopes. Consequently, they (Montgomery & Dietrich, 1988) commented that theories for channel initiation by saturation overland flow required revision. Formulation of such a model in terms of an erosional threshold leads to a model more consistent with the available field data.

Figure 11.4 A mat of vegetation protects the ground surface from erosion by overland flow in the unchanneled valleys upslope of channel heads in many grassland catchments, such as the Tennessee Valley area of Marin County, California

For the case of steady state rainfall, the overland flow discharge at a point in a catchment is that portion of the rainfall (Ra) that cannot be accommodated by subsurface flow (TS). This is given by

$$q = Ra - TS \tag{18}$$

Assuming the surface flow to be laminar, equation (18) can be set equal to (11) to solve for the critical drainage area per unit contour length:

$$a_{cr} = \frac{2\tau_{cr}^{3}}{Rk\nu\rho_{w}^{3}g^{2}S^{2}} + \frac{TS}{R} \tag{19}$$

This is equivalent to equation (3) of Dietrich *et al.* (1992). For low-gradient slopes the left-hand term will dominate. With increasing slope the right-hand term will increase in importance and will dominate the relation for steep slopes. Thus equation (19) predicts a non-linear drainage area–slope relation for channel initiation by saturation overland flow that for low-gradient areas is similar to the Horton overland flow threshold, but that with increasing slope becomes asymptotic to the threshold for soil saturation (Figure 11.6D). On steep slopes, however, landsliding processes may dominate channel initiation.

Landsliding

In many steep landscapes channel heads coincide with small-scale debris flow scars in topographic hollows (Figure 11.5) (Montgomery & Dietrich, 1988). In the Oregon Coast

Figure 11.5 Channel head associated with a debris flow at the base of a steep topographic hollow on Mettman Ridge near Coos Bay in the Oregon Coast Range

Range, for example, overland flow does not occur on steep slopes due to the high hydraulic conductivity of the low-density colluvial soils (e.g., Yee and Harr, 1977; Montgomery, 1991) and thus landsliding is a primary channel initiation mechanism. Although recurrent debris flows are a primary means of transporting sediment from hillslopes into downslope channels (Dietrich & Dunne, 1978), they typically excavate only a portion of the colluvium upslope of a channel head and only rarely, if ever, extend to drainage divides. Under a constant climatic regime, the frequency of debris flow excavation, and thus the long-term rate of sediment transport, probably tends to decrease from a relatively short recurrence interval low on the slope to essentially never at the drainage divide (Dunne, 1991). Other types of landsliding also may be associated with channel heads. For example, in some areas channels begin within and on the margins of earthflows. We will not consider such cases further, but will instead discuss debris-flow-controlled channel heads in well-defined topographic hollows, in part for simplicity of modeling and in part because of the long-term influence of debris flows on hollow form, which contrasts with the more ephemeral geomorphic expression of channels associated with other landslide types. The model outlined below (Dietrich *et al.*, 1986; Montgomery & Dietrich, 1989) is based on coupling a subsurface flow model (Iida, 1984) with a simple slope stability model.

As in the model for soil saturation, the steady state subsurface discharge per unit contour length may be expressed by equation (14). Assuming Darcy flow parallel to the ground surface and uniform saturated conductivity, this subsurface discharge also is given by

$$q = hzK\sin\theta\cos\theta \tag{20}$$

where K is the saturated hydraulic conductivity, θ is the ground surface slope and the proportion of the soil depth that is saturated, h, is measured vertically, rather than normal to the ground surface. Equations (14) and (20) allow an expression for the proportion of the soil thickness that is saturated as a function of the upslope contributing area:

$$h = RA/zK\sin\theta\cos\theta \tag{21}$$

The simplest, and most appropriate stability model is the infinite slope model, which can be expressed as

$$h = \frac{C'}{\rho_w g z \cos^2\theta \tan\phi} + \frac{\rho_s}{\rho_w}\left[1 - \frac{\tan\theta}{\tan\phi}\right] \tag{22}$$

where C' is the effective cohesion of the soil, ρ_s is the bulk density of the saturated soil, and ϕ is the angle of internal friction of the soil. Combining (21) and (22) yields an expression for the critical drainage area per unit contour length where

$$a_{cr} = \frac{zK\sin\theta\cos\theta}{R}\left[\frac{C'}{\rho_w g z \cos^2\theta \tan\phi} + \frac{\rho_s}{\rho_w}\left(1 - \frac{\tan\theta}{\tan\phi}\right)\right] \tag{23}$$

This equation predicts a highly non-linear relation between source area and slope (Figure 11.6E) and is only valid for

$$\tan\theta \geqslant [(\rho_s - \rho_w)/\rho_s]\tan\phi \tag{24}$$

which for many soils may be approximated by $\tan\theta \geqslant \tan\phi/2$. For cohesionless soils, the steepest slope stable under this relation is given by the friction angle of the soil (i.e., $\tan\theta = \tan\phi$). For cohesive soils steeper slopes may be stable.

DRAINAGE AREA–SLOPE PROCESS REGIMES

In a relatively uniform landscape, a single channel initiation process may be the dominant control on channel initiation. In landscapes with a variety of slopes and soil properties, however, all of these processes may influence channel initiation. In this case, we would

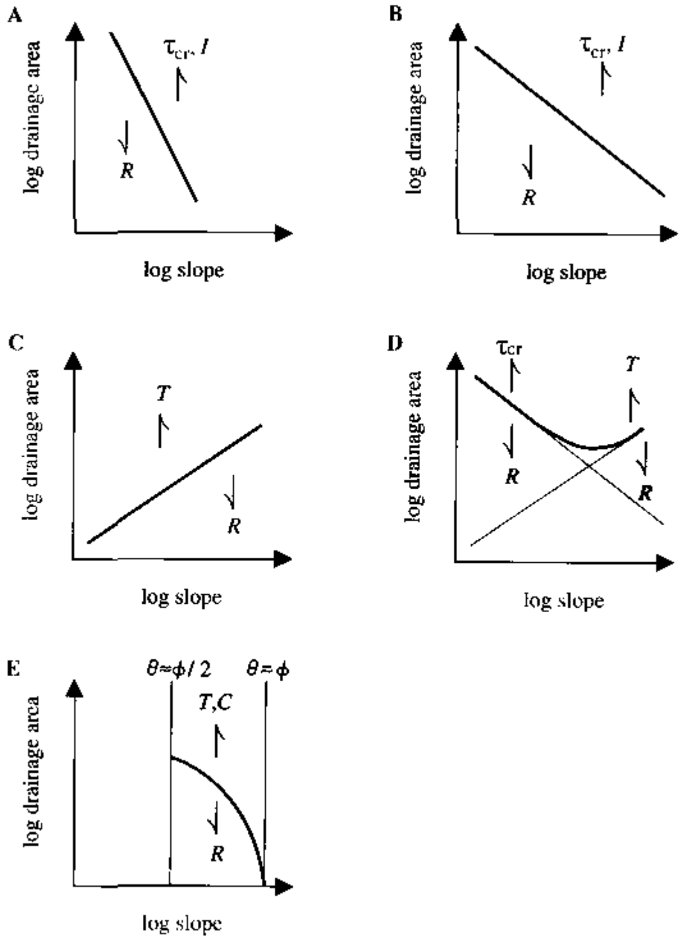


Figure 11.6 Schematic illustration of the relation between drainage area and slope predicted from simple threshold theories of channel initiation by (A) turbulent Horton overland flow, (B) laminar Horton overland flow, (C) soil saturation, (D) saturation overland flow, and (E) shallow, small-scale landsliding. Arrows indicate effect of increases in rainfall intensity (R), critical shear stress (τ_{cr}), infiltration capacity (I), transmissivity (T), cohesion (C), and angle of internal friction (ϕ)

expect different portions of the landscape to be dominated by different processes. Channel initiation in steep areas would be dominated by landsliding, whereas in undisturbed, low-gradient areas overland flow would dominate. Seepage erosion and gullying would be expected to occur in response to local disturbance in low-gradient areas. In many real landscapes, therefore, a complex interplay between channel initiation processes will control channel head locations. Models of channel initiation processes can be used to interpret this interplay and allow prediction of those portions of a landscape in which different channel initiation processes provide the primary control on channel head locations. Consequently, we may divide the landscape into process regimes by plotting the erosional thresholds for saturation overland flow, soil saturation,

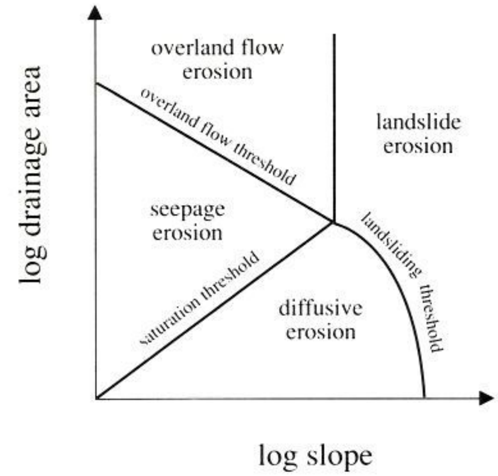


Figure 11.7 Combining the thresholds depicted in Figure 11.6 allows division of the landscape into process regimes where different sediment transport and channel initiation mechanisms operate. Below the thresholds for transport by soil saturation, overland flow, and landsliding only diffusive sediment transport is effective and hence this region should correspond to hillslopes stable to channelization tendencies

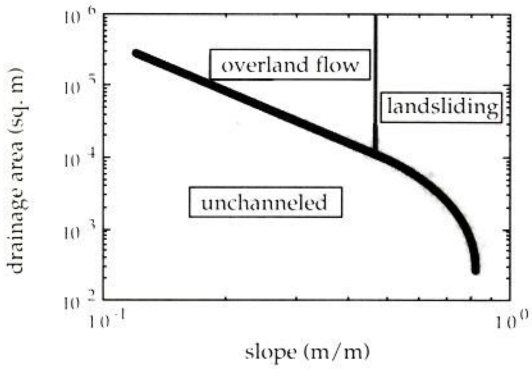
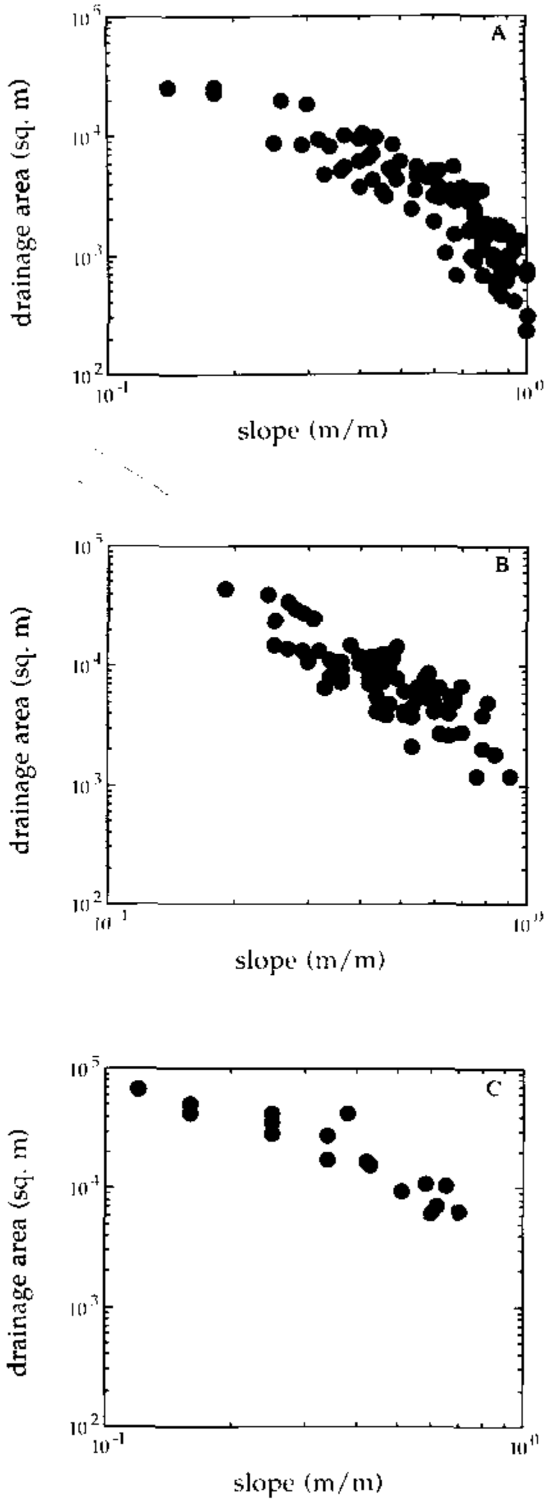
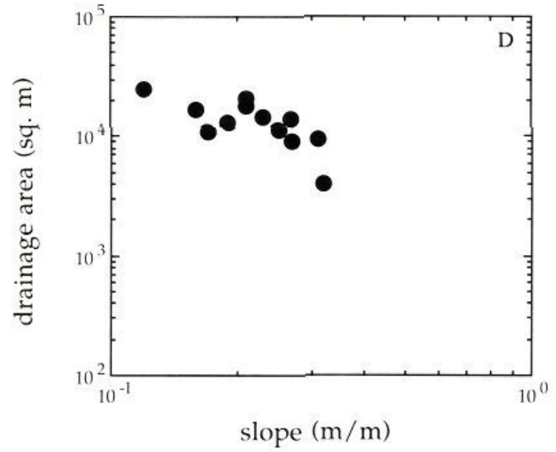


Figure 11.8 Illustration of the expected source area-slope relation in a landscape with a range of slopes in which both overland flow and landsliding control channel initiation. The central tendency in the source area-slope relation (solid line) is governed by long-term changes in climate and the other factors illustrated in Figure 11.6. Specific channel head locations also define a distribution about the central tendency (shaded area) that reflects both spatial and short-term temporal variability in the processes influencing channel initiation

and landsliding on a graph of drainage area and slope (Figure 11.7) (Dietrich *et al.*, 1992). Portions of the landscape that plot above the overland flow threshold may be expected to support channels. Non-erosive overland flow occurs between the thresholds for soil saturation and overland flow erosion and this field essentially defines low- to moderate-gradient unchanneled valleys. Areas within this zone may be subject to channelization by seepage erosion in response to disturbance. Portions of the landscape plotting above the landslide threshold also will be subject to channelization. Together these thresholds surround a portion of the landscape within which none of these processes


A
drainage area (sq. m)
slope (m/m)
B
drainage area (sq. m)
slope (m/m)
C
drainage area (sq. m)
slope (m/m)



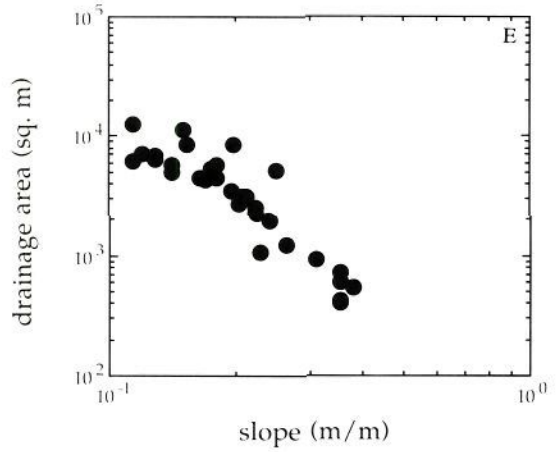


Figure 11.9 Plot of source area versus local slope at the channel head for studies areas in: (A) Coos Bay, Oregon, (B) Tennessee Valley, California, (C) Southern Sierra Nevada, California, (D) Stanford Hills, California, (E) northern Humboldt Range, Nevada. Channel initiation in the Coos Bay, Tennessee Valley, and Southern Sierra Nevada catchments is controlled primarily by overland flow on low-gradient slopes and landsliding on steeper slopes. The source area-slope relations for each of these areas reflect this transition, but it is best expressed in the Coos Bay data where landsliding is the dominant channel initiation process. Channel initiation in the Stanford Hills and Humboldt Range study areas is controlled by overland flow. The source area-slope relation in these catchments is consistent with the form predicted by an overland flow model. See text for sources of data

are effective at transporting sediment and in which diffusive processes dominate sediment transport [see also Dietrich *et al.* (1992)].

OBSERVED SOURCE AREA-SLOPE RELATIONSHIPS

Few field data exist against which to test even the general form of predicted source area–slope relations. Abrahams (1980) measured source areas from Schumm's (1956) mapping of badlands and found that source area is inversely correlated with the average slope of first-order basins. He (Abrahams, 1980) also found that source-area size and average first-order basin slope were unrelated for another area where field inspection suggested that channel initiation was dominated by subsurface flow. Unfortunately, the use of the average first-order basin slope severely compromises comparisons of these data with predicted source area–local valley slope relations. Dietrich *et al.* (1986) reported an inverse relation between source area and average source-area slope for a limited number of field sites from several different regions. Dietrich *et al.* (1987) found that source-area size is slope-independent in catchments underlain by basalt where relatively impermeable dikes control groundwater flow paths. Montgomery & Dietrich (1988, 1989) subsequently found inverse source area–local valley slope relations in three study areas where channel initiation occurs by landsliding, overland flow, and to a lesser extent seepage erosion and channel head sloughing. Repka (unpublished data) also found an inverse source area–local slope relation for an area in which channel initiation occurs by Horton overland flow. Although we cannot at present constrain or fully characterize all of the parameters in the models presented above, we can test the general form of the predicted drainage area–slope relations for some of these models by comparison with available field data from areas in which these processes are the dominant channel initiation mechanisms.

Most of the available source area–slope data are from areas in which overland flow and/or landsliding are the dominant channel initiation mechanisms. In an area with both gentle and steep slopes, and in which both overland flow and landsliding occur, the threshold requiring the smallest drainage area should be reflected in the source area–slope relation. The models presented above predict that $a_{cr} \propto S^{-2}$ for low-gradient slopes dominated by laminar overland flow, and a more rapid decrease in source-area size with increasing slope for landslide-controlled channel intiation. Landsliding is only a relevant process when equation (24) is satisfied, a condition that commonly will be approximated by $\tan\theta > 0.5$. Conversely, overland flow is rare on steep hillsides in soil-mantled environments. Therefore, it is reasonable to expect different source area–slope relations for low and high-gradient slopes (Figure 11.8).

Montgomery & Dietrich (1988) reported inverse source area–slope relations across a wide range of slopes ($0.1 < \tan\theta < 1.0$) for three catchments in the western United States. They documented that landsliding dominated channel initiation on steeper slopes, whereas overland flow dominated channel initiation on low-gradient slopes and showed that a model for channel initiation by landsliding provided a reasonable fit to the observed data for $\tan\theta > 0.5$ for one of these areas (Montgomery & Dietrich, 1988, 1989). Willgoose *et al.* (1991) subsequently argued that the general inverse trend across the full range of these data indicates a single underlying channel initiation mechanism [an assertion directly contradicted by field observations (Montgomery & Dietrich, 1988, 1989)]. Examination of the available data confirms the expected generality of different source area–slope relations for low- and steep-gradient slopes. Furthermore, the observed data are consistent with the predicted form of the source area–slope relation for overland flow on low-gradient slopes and for landsliding on steeper slopes.

Nine source area–slope data sets are available at present. Five data sets report source area and local slope at the channel head (Figure 11.9), while four report average source-area gradient (Figure 11.10). On steep slopes gradient does not vary greatly and hence average source-area slopes are probably similar to local slopes at channel heads.

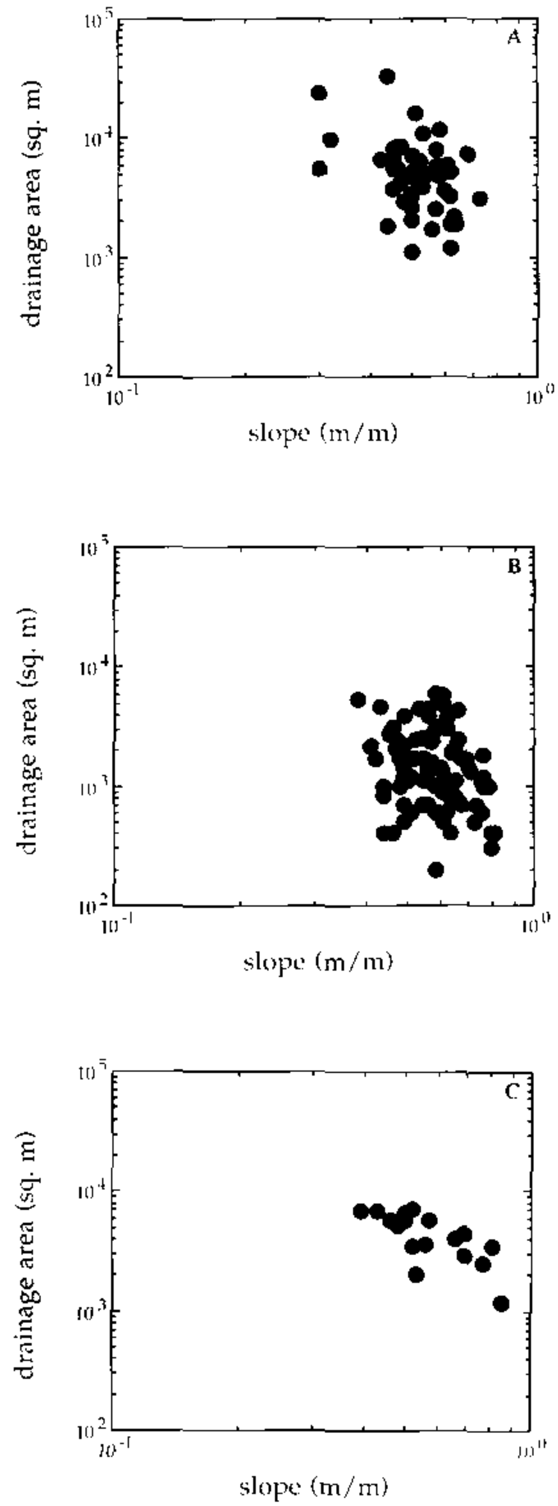
Together these data represent a broad range in geology, vegetation, and climate. Furthermore, observations pertaining to channel initiation mechanisms are available for each of these areas. The most intensively studied areas are a series of catchments in Coos Bay, Oregon, and the Tennessee Valley area of Marin County, California, in which all channel heads were mapped and local slopes at the channel head were measured in the field. Montgomery & Dietrich (1988, 1989, 1992) discuss the study areas, data collection methods, and field observations from these areas.

The Coos Bay area is underlain by nearly flat-lying Eocene sandstone, is covered by a managed coniferous forest, and receives an annual rainfall of approximately 1500 mm. Data from the catchments mapped in this area span a wide range of slopes (Figure 11.9A). On low-gradient slopes ($\tan\theta < 0.5$) the data followed an inverse trend consistent with $a_{cr} \propto S^{-2}$, while data for steeper slopes exhibit a more rapid decline in source area with increasing slope. Field observations from these catchments indicate that landsliding is a major channel initiation process on steep slopes (Montgomery & Dietrich, 1988; Montgomery, 1991), whereas overland flow occurs only on gentler slopes in this area due to the high conductivities of the colluvial soils.

A similar pattern of channel initiation by overland flow on low-gradient slopes and by landsliding on steeper slopes was observed in the Tennessee Valley catchments (Montgomery & Dietrich, 1988, 1989; Montgomery, 1991). A number of the channel heads in these catchments appear to be controlled by seepage erosion associated with the development of abrupt channel heads in response to historic overgrazing. These channel heads are discussed in greater detail elsewhere (Montgomery & Dietrich, 1989; Montgomery, 1991). The area is underlain by chert, greenstone, and greywacke, is covered by grassland and coastal prairie vegetation, and receives approximately 760 mm annual rainfall. Source area-slope data for this area are generally consistent with the relation predicted by overland flow for low-gradient slopes and exhibit an apparent steepening for slopes greater than 0.5 (Figure 11.9B). Montgomery & Dietrich (1988, 1989) reported that the data for the steeper slopes are reasonably modeled by a threshold-based equation for channel initiation by landsliding. Dietrich *et al.* (1992) show that the general inverse relationship is consistent with the saturation overland flow model [equation (19)].

Data from three other gentle-gradient study areas also reflect channel initiation by overland flow. The Southern Sierra data (Figure 11.9C) (Montgomery & Dietrich, 1988, 1992) are from an area underlain by deeply weathered, unglaciated granitic rocks covered by open oak forest and grasslands, which receives approximately 260 mm of rainfall annually. The Stanford Hills data (Figure 11.9D) are from an arena on the Stanford University campus south of San Francisco, California, underlain by basalts and sandstones and covered by open oak forest and grasslands. The data from Nevada (Figure 11.9E) are from a dissected Pleistocene alluvial fan in an arid region (Repka, unpublished data). Field observations suggest that channel head locations in each of these areas are dominated by laminar overland flow and the data from each of these areas exhibit an inverse relation consistent with the form of the predicted relation; $a_{cr} \propto S^{-2}$.

Three additional data sets using average source–area slope are available for steep environments. Data reported by Dietrich *et al.* (1987) from San Pedro Ridge, California (Figure 11.10A) suggest a steepening of the source area–slope relation at about $\tan\theta = 0.5$. The area is underlain by sandstone and is covered by a hardwood forest. Small-scale


A
B
C
drainage area (sq. m)
slope (m/m)
10^{2}
10^{3}
10^{4}
10^{5}
10^{-1}
10^{0}

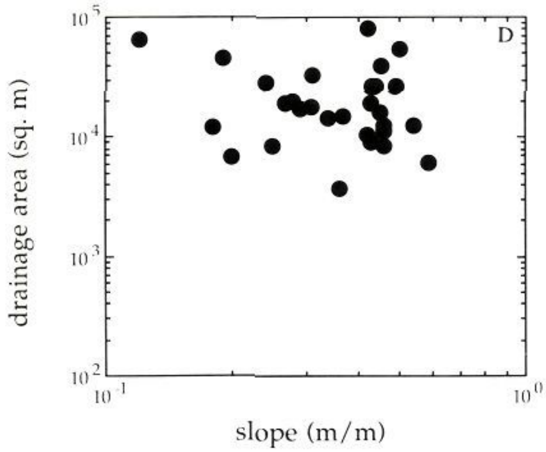


Figure 11.10 Plot of source area versus average source-area slope for study areas in (A) San Pedro Ridge, California, (B) San Dimas, California, (C) Japan, and (D) Rock Creek, Oregon. Channel initiation in the San Pedro Ridge, San Dimas, and Japan study areas is dominated by landsliding. The steep source area–slope relation in each of these areas is consistent with the form predicted by the landslide model. Channel head locations in the Rock Creek study area, however, are controlled by the location of impermeable dikes. In this case, the source areas are essentially slope independent on both gentle and steep slopes. See text for sources of data

landsliding is common at steep channel heads in this area. Data for the San Dimas area (Figure 11.10B) were derived from Maxwell's (1960) map of field surveyed channel networks in the Fern basin of the San Dimas National Forest, California. Landsliding is a common process on the steep slopes in this area and the data suggest a very steep source area–slope relation. Dietrich *et al.* (1987) reported data from Japan (Figure 11.10C) derived from figures in Tsukamoto *et al.* (1978). These data are from areas in which landsliding is an important sediment transport process. The form of the data from each of these steep landscapes conforms to the steeper relation expected from channel initiation by landsliding.

Dietrich *et al.* (1987) also reported data from Rock Creek, Oregon (Figure 11.10D), which do not exhibit a systematic source area–slope relation over low-gradient slopes ranging from $0.1 < \tan\theta < 0.6$. This area is underlain by gently dipping basalt flows and breccias cross-cut by fine-grained dikes. They (Dietrich *et al.*, 1987) found that fine-grained basalt was commonly exposed at the channel head and illustrated how such low-permeability zones in the underlying bedrock could strongly control channel head locations. This local bedrock hydrologic control on channel initiation may dominate source area–slope relations in areas with high bedrock conductivities and local low-permeability zones.

A plot of the available data indicate a steepening in the source area–slope relation at gradients greater than $\tan\theta = 0.5$ associated with a change in process dominance from overland flow on gentler slopes to landsliding on steeper slopes (Figure 11.11). Data from the northern Humboldt Range are excluded because they are from an arid environment and plot significantly below the range of the data from more humid environments. Data from Rock Creek, Oregon, also are excluded because of the strong

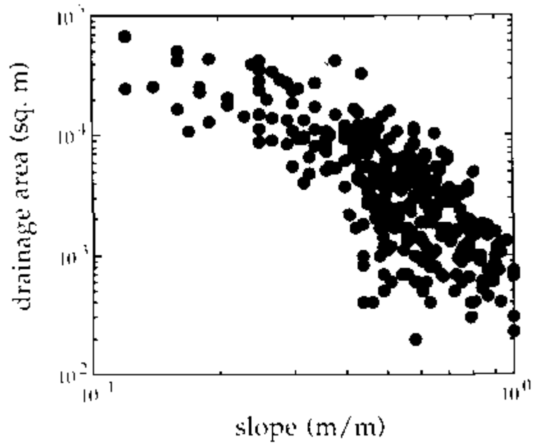


Figure 11.11 Plot of source area versus slope for all study areas except Rock Creek, Oregon, and northern Humboldt Range, Nevada

influence of bedrock dikes on channel head locations. The rapid decrease in source-area size on steeper slopes apparent in Figure 11.11 is consistent with the hypothesis that different channel initiation processes dominate different portions of a landscape (Figure 11.7), in this case reflecting the dominance of overland flow on low-gradient slopes and landsliding on steeper slopes. The approximately order-of-magnitude scatter about the central tendency of the composite data set reflects the variance of channel head locations both within and between study areas. The general agreement between the form of the observed source area–slope data and that predicted by models of relevant processes supports the general formulation of these models.

Source area–slope relations provide a conceptual framework within which to investigate landscape response to changes in the conditions affecting channel initiation. Slope may be thought of as the independent variable in the source area–slope relations due to the long response time and bedrock erosion required to significantly alter valley-long profiles. This leads to the question of what sets the valley slope and suggests that source-area size responds to changes in climate and process regimes through channel head migration along the valley axis. For all of the processes discussed above source area should decrease with either increasing rainfall or a decrease in infiltration capacity, hydraulic conductivity, or critical shear stress. The rainfall dependency suggests that channel network expansion should then accompany change to a wetter climate, whereas increased aridity would result in downslope retreat of channel heads and infilling of the upper reaches of first-order streams (Montgomery & Dietrich, 1988). However, the infiltration capacity, hydraulic conductivity, and critical shear stress may co-vary with changes in rainfall and the combined impact on channel head locations may be non-linear. For example, vegetation reduction in response to dramatically increased aridity may decrease the infiltration capacity, the critical shear stress, or possibly the flow resistance. Although flow resistance will tend to reduce flow depth and, consequently, the flow boundary shear stress, a relatively small reduction in critical boundary shear stress will greatly decrease the source area (Dietrich *et al.*, 1992). Similarly, land-use modifications that increase the effective rainfall intensity at the ground surface, or which decrease either the critical shear stress or the infiltration capacity of the soil would result in channel head advance, accelerated erosion, and channel network expansion. Furthermore, environmental perturbations resulting in a change in the controls on channel head locations would potentially destabilize the network of unchanneled valleys upslope of channel heads. As a whole then, these relations provide a quantitative framework within which to consider controls on channel initiation and thus channel network extent and landscape dissection.

CHANNEL INITIATION THRESHOLDS AND LANDSCAPE DISSECTION

Channel heads are not static and this dynamism is apparent in the variance in the source area–slope data, which integrate both spatial variations in the physical properties of the soil (e.g., τ_{cr}, K, z) and temporal variations in the position of individual channel heads (Figure 11.8). Over landscape-forming time scales, it is this range of channel head locations that defines the zone over which sediment transport occurs by channel processes. The variance about the general trend in the source area–slope data suggests that the threshold of channel initiation is best described as a zone of transition from channeled to unchanneled regions of the landscape. Such a transitional zone does not necessarily contradict models for valley stability based on the dominance of incisive and diffusive process. Rather this suggests the effect of both spatial and temporal dominance in maintaining the balance. Over long time scales, the extent of valley dissection represents an interplay between the frequency of channel head extention and the rate of diffusional infilling of the valley (e.g., Calver, 1978; Montgomery & Dietrich, 1992). In landscapes where the erosional threshold is significantly increased and the channel head is displaced far enough downslope, the range of channel head locations also will be displaced and the unchanneled valley network will expand downslope. In contrast, a decrease in the channel initiation threshold will result in expansion of the channel network and potentially the development of new valleys through incision into previously undissected hillslopes. This suggests a dynamic view of landscape development in which the relation between channel initiation and valley maintenance reflects both the spatial and temporal variance of channelization thresholds.

At the landscape scale, geomorphic response to environmental change may involve a shifting of the range over which channel heads oscillate. For example, a change from conditions favoring small source areas to conditions favoring larger source areas will result in a retraction of the zone defined by the short-term range in channel head locations. This will result in a decrease in the frequency that a channel will occupy any point upslope of the mean channel head location which, in turn, will allow diffusive transport to overwhelm channelization tendencies farther downslope from the drainage divide. Thus, a portion of the originally channeled valley will infill with colluvium, possibly to the point at which the original valley becomes obscured as a topographic feature. In contrast, a change from drier to wetter conditions (assuming that other factors, especially vegetation, do not co-vary with precipitation) will cause channel heads to advance upslope, excavate previously accumulated valley fills, and incise undissected hillslopes. The maximum distance upslope that channel extention may carve, and maintain, a valley depends upon the relation between the sediment transport function for extreme events and the rate of diffusive sediment transport. Therefore, the average channel head location will be significantly downslope of its long-term influence on valley maintenance. Consequently, the extent of landscape dissection may reflect the long-term interplay between a temporally variable channelization threshold and relatively continuous diffusive infilling.

CONCLUSIONS

The extent of channel and valley networks provide identifiable and morphologically distinct descriptions of landscape dissection. Channel initiation is controlled by a variety of processes which may be modeled as threshold phenomena, whereas valley maintenance may be modeled as reflecting either a spatial transition in process dominance, or a temporal variance in the exceedence of a channel initiation threshold. Simple analytical models for channel initiation processes predict source area–slope relations that agree with the available field data. These relations define areas of the landscape in which incisive (slope and area-dependent) and diffusive (slope-dependent) processes operate. In essence, we suggest that in vegetated, soil-mantled landscapes the extent of the valley network reflects the spatial signature of a dynamic channel network controlled by channelization thresholds. Consequently, the Gilbert hypothesis of landscape dissection, in which valley development reflects a stable spatial transition in process dominance, is best viewed as relevant to geologic time scales (i.e., 10^4–10^6 yr), whereas the Horton hypothesis of channel initiation and landscape dissection in response to exceedence of erosional thresholds is best viewed as relevant to geomorphic time scales (i.e., 10^2–10^3 yr).

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