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Predicting the topographic limits to a gully network using a digital terrain model and process thresholds

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Abstract. A digital terrain model is used with process thresholds to predict the extent of a stable gully network in a 5 km^2 catchment of the southeastern highlands of Australia. The model, developed by *Dietrich et al.* [1992, 1993], predicts the topographic controls on channel networks and interprets these in terms of a critical shear stress for channel incision (τ_c) applied by saturation overland flow. We adapt the model slightly to compare the shear stress applied by Hortonian overland flow to that applied by saturation overland flow. The limits to gully erosion in the catchment are controlled strongly by a topographic threshold that has an inverse relationship between upslope catchment area and local gradient. The topographic threshold for channel incision is reproduced using a simple model of Hortonian overland flow and a τ_c appropriate for incision into a degraded grass surface (τ_c = 245 dyn/cm^2). This is consistent with historical evidence for the timing of gully erosion. The study confirms a strong topographic control on the extent of the channel network in a catchment significantly different from the western North America catchments where the topographic threshold was first demonstrated. Despite its simplicity, the model for incision by overland flow appears capable of distinguishing the hydrological processes responsible for channel incision when these are reflected in the relationship between channel network and landscape morphology. The model requires relatively simple inputs, suggesting it may be useful for mapping gully erosion hazard in actively eroding catchments.

Introduction

Extensive gully erosion accompanied the introduction of modern agriculture to Australia, as it did in several other continents [*Eyles*, 1977; *Cooke and Reeves*, 1976; *Boast*, 1990]. Such erosion is responsible for massively increased sediment delivery, reduced water quality, and the destruction of buildings, fences, and roads. In southeastern Australia the main phase of gully erosion occurred last century and analysis of sequential aerial photography shows that many gully networks have not expanded in the last 40 to 50 years [*Eyles*, 1977; *Starr*, 1989; *Prosser*, 1991]. This suggests that these gullies have now reached an equilibrium extent. However, some gullies are still extending headward, and in other areas, recently intensified land use is resulting in gully initiation [*Prosser and Winchester*, 1996]. In both cases it would be useful to distinguish those sites that are most susceptible to future gully erosion and to predict the eventual limits of the gullies. Landscape dissection is also a crucial aspect of landform evolution, so relatively simple means of predicting the extent of channel networks are also useful in models of landform evolution [e.g., *Rinaldo et al.*, 1995].

The extent of the channel network is controlled by processes at the channel head. These processes include incision by overland flow, seepage and tunnel erosion by subsurface flow, and mass failure or landsliding (see review by *Dietrich and Dunne* [1993]). Channelized flow is then responsible for transporting eroded sediment downstream. Topography is a strong influence on these processes through gradient and by concentrating surface and subsurface flow. Therefore we could expect strong topographic controls on the extent of channel networks.

The potential importance of topography was formalized first by *Horton* [1945], who proposed a critical hillslope length required to generate a channel. The critical length was interpreted as that required to generate a boundary shear stress of Hortonian overland flow sufficient to overcome surface resistance to scour. Subsequent analyses have shown a range of relationships between channel network properties and topography, although the physical processes behind these remain less certain [*Abrahams*, 1984]. High-resolution digital terrain models (DTMs) enable rapid, detailed investigation of topography and suggest that characterizing topography may be sufficient to predict the extent of channel networks to reasonable accuracy. *Moore et al.* [1988a], for example, used a DTM to demonstrate that ephemeral gully erosion in an agricultural catchment was restricted to locations with relatively large upslope contributing area and steep gradient. This topographic control reflected the conditions required for soil saturation and was also related to the unit power of overland flow.

Factors other than topography influence erosion processes, such as the strength of materials, soil hydraulic properties, and

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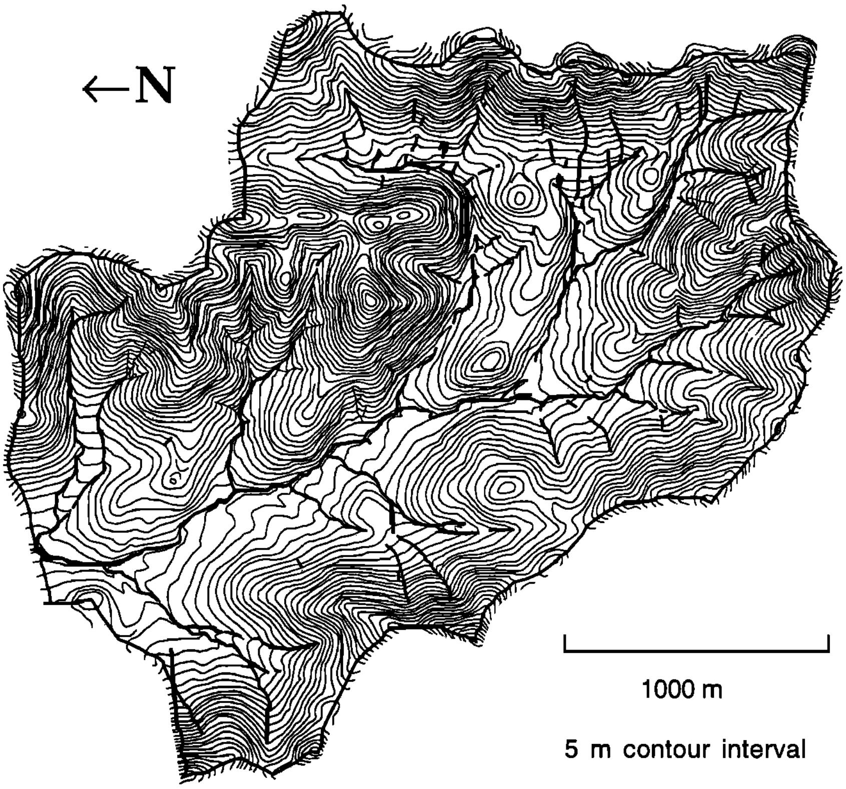

Figure 1. Topographic map of Gungoandra Creek catchment showing field-surveyed channel network. Gullies are indicated by thick lines and prehistoric channels are shown by thin lines.

vegetation cover. These attributes have been incorporated into DTM analysis of erosion [*Vertessy et al.*, 1990], but defining their spatial distribution takes considerable effort, and it is not clear that such effort is justified by improved prediction.

Dietrich et al. [1992, 1993] and *Montgomery and Dietrich* [1988, 1989, 1992, 1994a, b] explored topographic controls on channel networks in more detail using both field measurements and DTMs. They observed strong inverse relationships between contributing area and slope at channel heads from field measurements over several catchments, suggesting that there is a threshold contributing area required to support a channel [*Montgomery and Dietrich*, 1988]. In steep terrain, this topographic threshold was interpreted in terms of a process threshold for the onset of landsliding using field and DTM data for three catchments [*Montgomery and Dietrich*, 1989, 1994a; *Dietrich et al.*, 1992, 1993].

In relatively low gradient hollows the threshold contributing area required to support a channel was found to be inversely proportional to the square of slope. This relationship is consistent with control of the channel network by a critical shear stress for incision by saturation or Hortonian overland flow [*Montgomery and Dietrich* 1988, 1989, 1992, 1994b]. The channel network of a small catchment in Marin County, California, was successfully simulated by combining a DTM with the process threshold for incision by saturation overland flow [*Dietrich et al.*, 1992, 1993]. Field experiments conducted in the catchment showed that the modeled critical shear stress was realistic if convergence of flow along the axis of hollows was taken into account [*Prosser et al.*, 1995; *Prosser and Dietrich*, 1995]. The model assumed spatially uniform soil properties and vegetation, demonstrating the strong control of topography on the channel network.

The model developed by *Dietrich et al.* [1992, 1993] may provide a practical tool for predicting the extent of gully erosion. However, until now the applicability of the process threshold for incision by overland flow has not been tested beyond the Marin County catchment. Herein, we test if there is a topographic threshold to the extent of channels in a typical gullied catchment of southeastern Australia. This provides an independent test of the model in an environment quite different from that where it was first applied. We then use the model to interpret the extent of the channel network in terms of thresholds for incision by Hortonian and saturation overland flow and compare these predictions to field observations.

Site Description

We selected the upper 5 km^2 of the Gungoandra Creek catchment as a site for digital terrain modeling. Gungoandra Creek is 80 km south of Canberra on the Southern Tablelands of New South Wales. The site was chosen because it is a typical steep catchment of southeastern Australia with a well-developed and stable network of gullies (Figure 1). Comparison of aerial photographs showed that the position of 95% of gully heads had not moved beyond the measurement resolution of 20 m between 1944 and 1992. No soil conservation works have been conducted in the catchment, and as land use and climate have not changed since the 1950s, we assume that the gully network is in equilibrium with the present conditions.

Historical evidence suggests that the gullies formed in the mid-19th century after the introduction of European-style agriculture [*Eyles*, 1977]. The higher-elevation gullies are often discontinuous, both with each other and with larger trunk gullies, demonstrating that gullies were initiated at many locations and did not form by retreat of a single headcut. In the lower parts of the catchment, gullies have vertical heads of 1 to 2 m height, and on the upper slopes they have gradual heads on slopes which show evidence of widespread surface wash erosion. Gullies were defined as channels incised to bedrock through at least 30 cm of soil, colluvium or alluvium, with an easily discerned edge to the top of the channel bank. A few valleys have shallow channels with rocky sides, where regolith has not accumulated or where channels have formed on colluvial deposits. Such channels were defined as prehistoric channels because they have a smooth vegetated transition from channel floor to hillslope, indicating considerable age. Definition of the channel network is crucial to the interpretations of this paper, and the definitions used here identify long lasting, morphologically distinct features that are likely to have a significant role in landform evolution and have practical significance to soil conservationists.

The catchment has steep, straight, upper hillslopes and long concave footslopes that merge with a narrow alluvial flat. In planform the hillslopes are broadly rounded with a low density of narrower valleys (Figure 1). The bedrock is Silurian acid volcanics and fine-grained sandstones which have been folded, sheared, and metamorphosed into vertically dipping, north-south striking beds [*Jenkins*, 1993]. Median annual rainfall is 500 mm, relatively evenly distributed throughout the year but unreliable from one year to the next. Summer rains are characterized by intense convective storms of short duration, while cold fronts provide winter rains of longer duration and lower intensity [*McAlpine and Yapp*, 1969].

The natural eucalypt woodland of the catchment was cleared in the mid-19th century and was replaced by pastures of native and introduced species under the influence of sheep grazing. Valley floors are naturally treeless with a cover of tussock grass and sedge [*Costin*, 1954]. Vegetation cover is often sparse (20 to 50%) because of poor soil fertility and a short growing season. Stock numbers were much higher in the early part of the century, which was also a time of severe droughts and rabbit plagues [*Durham*, 1961]. Presumably vegetation cover was even lower in those times.

The upper slopes of the catchment have thin soils and a cobble-strewn surface, with frequent minor outcrops of bedrock. Hollows and footslopes are mantled by weathered Pleistocene colluvial deposits of 1 to 5 m thickness. The soils are typical of the region with shallow lithosols on the steeper slopes and yellow and red podsolics developed on the colluvial materials. The A horizon is up to 40 cm thick, weakly pedal, and hard setting with a loam to clay loam texture. On the lower slopes there is a sharp break to the B horizon, which has a clay loam to sandy clay texture, and the clays disperse easily.

Hydrological investigations of other catchments in the region show that B horizons are essentially impermeable with saturated hydraulic conductivities of 0 to 5 mm/h [*Hook*, 1990; *Mackenzie et al.*, 1991; *Lane et al.*, 1994]. Intense rainfall can generate Hortonian overland flow on all parts of the landscape [*Lane et al.*, 1994], predominantly in summer. As winter progresses, perched water tables saturate the B horizon from its interface with the underlying bedrock. The transmissivity and water holding capacity of the B horizon are so low that by the end of winter the B horizon is saturated across the lower slopes [*Hook*, 1990]. Water tables then spread quickly into the A horizon during winter storms causing rapid throughflow, downslope exfiltration, and saturation overland flow [*Hook*, 1990; *Lane et al.*, 1994]. Runoff can also be generated by rainfall intensity exceeding the transmissivity of the A horizon [*Hook*, 1990].

The Model

There are four components to the model developed by *Dietrich et al.* [1992, 1993]: generation of the DTM, prediction of steady state saturation overland flow, prediction of channel incision by overland flow, and prediction of channel incision by shallow landsliding. The first three components are of concern here and are outlined below.

Dietrich et al. [1992, 1993] used the Topog DTM, which divides a catchment into a network of irregularly shaped elements defined by upper and lower contour sections and flow lines drawn normal to the contours [*O'Loughlin*, 1986; *Moore et al.*, 1988b]. It is assumed that both shallow subsurface flow and overland flow follow the flow lines with little lateral diffusion [*O'Loughlin*, 1986]. The Topog package provides calculation of terrain attributes for each element, such as upslope catchment area (a), lower contour width (b), specific catchment area (a/b), and gradient ($\tan\theta$). The shape of an element is described by the convergence index, defined as $(b_2 - b_1)/(b_2 + b_1)$, where b_1 and b_2 are the contour lengths of the upslope and downslope boundaries of the element, respectively.

The discharge of saturation overland flow (Q) across the lower boundary of an element is

$$Q = qa - TMb \tag{1}$$

where q is total surface and subsurface runoff per unit area, T is soil transmissivity (m^2/d), and M is the local slope ($\sin\theta$) [*O'Loughlin*, 1986; *Dietrich et al.*, 1992, 1993]. Total runoff per unit area equals precipitation less evaporation and deep drainage. Soil transmissivity is the integral of saturated hydraulic conductivity over the depth of the soil profile. The second term on the right-hand side of (1) is the ability of soil to transmit water parallel to the flow lines, so that soil saturation and overland flow occurs wherever

$$\frac{a}{b} \geq \frac{T}{q} M. \tag{2}$$

The prediction of saturation overland flow assumes that all rainfall on unsaturated areas infiltrates the soil and that saturated areas are fed by throughflow moving from upslope under the influence of gravity.

For Hortonian overland flow the discharge of overland flow is simply

$$Q = qa \tag{3}$$

where q is now precipitation minus infiltration. It is assumed that Hortonian overland flow occurs on all parts of the landscape and that infiltration is spatially uniform, although spatially variable infiltration can be accommodated by the Topog models. Infiltrated water is assumed not to contribute to Hortonian overland flow.

The model assumes that overland flow incises a channel wherever boundary shear stress (τ_b) exceeds a critical value (τ_c) representing surface resistance to scour. For nonaccelerating flows,

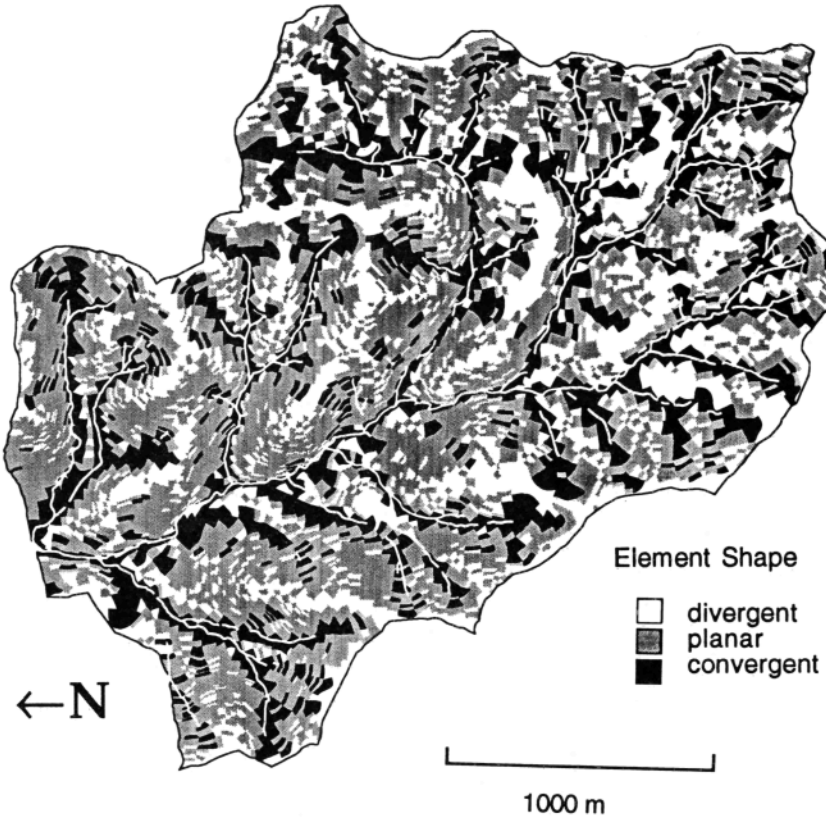

Figure 2. Spatial pattern of divergent, planar, and convergent Topog elements in the Gungoandra Creek catchment.

$$\tau_b = \rho g d M \tag{4}$$

where ρ is fluid density, g is acceleration due to gravity, and d is mean depth of overland flow. To determine the τ_b applied by Q requires knowledge of flow resistance, characterized by the Darcy-Weisbach friction factor (f):

$$f = 8gdM/u^2 \tag{5}$$

where u is mean velocity. For flow over grassed surfaces, f can be expressed as a power function of Reynolds number ($f = kRe^c$). Noting that $Q = udb$ and $Re = ud/\nu$, where ν is the kinematic viscosity of fluid, (1) (4), and (5) can be combined to yield the τ_b applied by saturation overland flow:

$$\tau_b = \left(\frac{g^2 k \rho^3}{8\nu^c}\right)^{1/3} M^{2/3}\left(q\frac{a}{b} - TM\right)^{(2+c)/3}. \tag{6}$$

This is the more general form of (10) and (11) used by *Dietrich et al.* [1993] which are for the special cases $f = k/Re$ and $f = k$, respectively. Substituting τ_c for τ_b in (6), saturation overland flow will erode channels wherever the following topographic condition is satisfied:

$$\frac{a}{b} \geq \left(\frac{\tau_c^3 8\nu^c}{g^2 k \rho^3}\right)^{1/(2+c)} \frac{1}{qM^{2/(2+c)}} + \frac{T}{q}M. \tag{7}$$

Following the same reasoning, but using (2) instead of (1), the τ_b generated by Hortonian overland flow is

$$\tau_b = \left(\frac{g^2 k \rho^3}{8\nu^c}\right)^{1/3} M^{2/3}\left(q\frac{a}{b}\right)^{(2+c)/3} \tag{8}$$

and channels will erode wherever,

$$\frac{a}{b} \geq \left(\frac{\tau_c^3 8\nu^c}{g^2 k \rho^3}\right)^{1/(2+c)} \frac{1}{qM^{2/(2+c)}} \tag{9}$$

which is the more general form of (12) derived by *Montgomery and Dietrich* [1994b] for $f = k/Re$.

The model does not consider explicitly the role of seepage processes, which may be important for eroding a channel head initiated by overland flow. Although seepage processes were observed in the catchment modeled by *Dietrich et al.* [1992, 1993], the extent of the channel network could be predicted without their inclusion. Nevertheless, seepage processes can be inferred from the model. Seepage loosens sediment from the gully head and makes it available for transport by surface flow. In the model this is equivalent to a negligible τ_c for erosion by overland flow, so that (7) reduces to (2): the threshold for soil saturation [*Montgomery and Dietrich*, 1994b]. In the model the position of a channel head controlled by seepage is therefore limited by the upslope extent of soil saturation. This approximation is supported by field studies which show that seepage erosion is associated with local ground saturation [e.g., *Montgomery and Dietrich*, 1995].

We will apply the process thresholds for Hortonian and saturation overland flow to the Gungoandra Creek catchment, but first we will describe how the DTM was derived and use it to investigate the presence of topographic thresholds.

Topographic Analysis

A base map of the Gungoandra catchment (Figure 1) was commissioned from 1:25,000 color aerial photography. The map was prepared to a scale of 1:7,500 with 5-m contour interval using analogue photogrammetry. Channels, remnant forest, and other visible features were also plotted on the map. The channel network was represented well on the map, because of low vegetation cover, but was checked and modified by field mapping.

The base map was digitized and registered to the Australian Map Grid using ground control points. Elevation and horizontal errors of the input data were estimated to be ±0.5 m and ±10 m, respectively. The digital contours were then used to construct the Topog DTM, producing 9130 discrete elements with an average lower width of 40 m (Figure 2). A variable contour interval was specified to create a network of reasonably uniform element size. The contour interval is 2 m between 772 m and 840 m, 3 m between 840 m and 900 m, and 5 m between 900 m and 1015 m.

The convergence index of each element, calculated by Topog, is a convenient method of classifying the shape of each element in a way that also represents the broader upslope topography. Convergent, planar, and divergent elements of the catchment were defined by convergence index ranges of <-0.07, -0.07 to 0.07, and >0.07, respectively (Figure 2). These boundaries are arbitrary and were chosen on visual inspection of the DTM. Convergent elements map the valleys, divergent elements define the ridge system, and planar elements represent the intervening slopes (Figure 2). Specific catchment area decreases from convergent to planar to divergent elements. Channels are restricted to convergent and planar elements, so only these are considered in the following analyses.

A scattergram of a/b plotted against $\tan\theta$ shows clear separation of channeled and unchanneled elements (Figures 3a and 3c), and can be used to investigate topographic thresholds empirically. Elements containing channel heads plot at the boundary of the channeled and unchanneled fields (Figure 3b). This is to be expected as channel heads represent the transition between slope and channel processes. Figure 3 also shows a strong negative relationship between a/b and $\tan\theta$ at channel heads. Linear regression through the log-transformed channel head data yields the topographic threshold for channel erosion:

$$a/b = 30 \tan \theta^{-1\,6} \qquad R^2 = 0.52 \tag{10}$$

This threshold defines the channel network well (Plate 1a), correctly discriminating channels in 85% of planar and convergent elements. A similar relationship can be demonstrated for a plotted against $\tan\theta$, as b is reasonably constant at channel heads ($a = 1421 \tan\theta^{-1.59}$). Catchment area alone, however, does not define the channel network well, as it overpredicts channels on the low-gradient footslopes (data not shown). Interpreting the topographic threshold (10) in terms of (9), (2), and (7) moves the investigation from the purely empirical to potential process interpretations of the channel network.

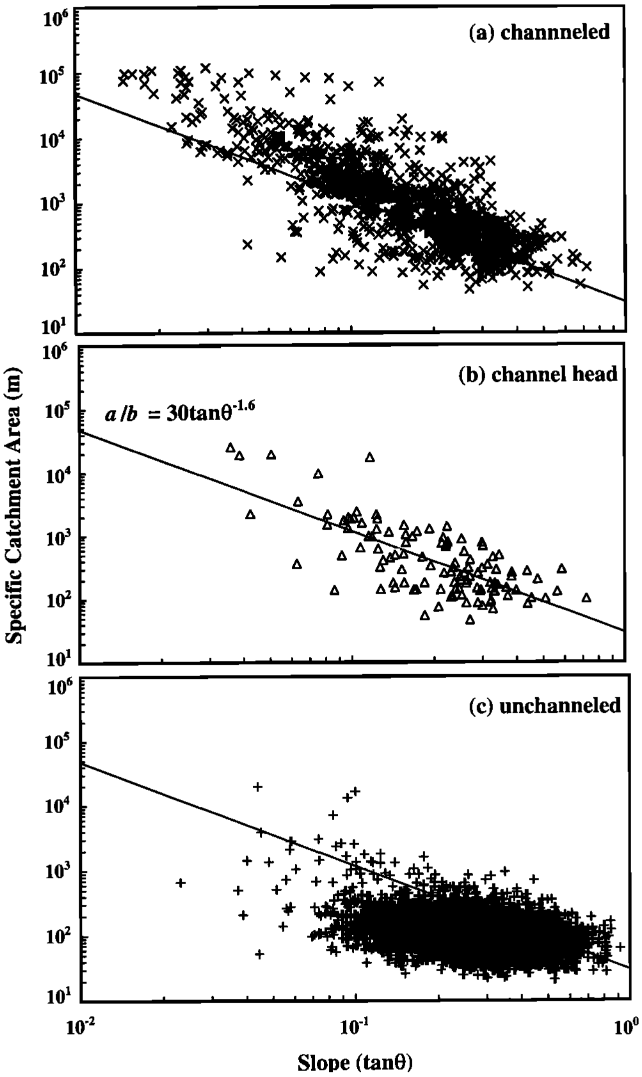

Figure 3. Specific catchment area and slope for (a) channeled, (b) channel head, and (c) unchanneled convergent and planar elements at Gungoandra Creek, showing the threshold line that separates channeled from unchanneled elements.

Threshold of Hortonian Overland Flow

To predict τ_b applied by Hortonian overland flow using (8) requires us to specify the runoff per unit area (q) and the variation of flow resistance with Reynolds number ($f = kRe^c$). The channel network has been stable for 50 years. Consequently, we chose the 1 in 50 year, 1-hour duration rainfall intensity of 42 mm/h [*Pilgrim*, 1987] to represent a typical channel-forming event. Little difference in τ_b was observed using a 1 in 100 year rainfall event, and larger-magnitude events are probably too infrequent to play an important role in channel formation. The steady state infiltration rate was estimated to be 20 mm/h based on soil texture [*Craze and Hamilton*, 1991] and rainfall simulator experiments on similar soils [*Fisher*, 1993]. This gave q a value of 22 mm/h.

The relationship $f = 1500Re^{-0.7}$ was used to characterize flow resistance. This is a typical relationship for concentrated overland flow on degraded grass [*Prosser et al.*, 1995]. The flow resistance relationship needs to differ by at least an order of magnitude from that above before it affects the prediction of τ_b. Using relationships from other degraded plots [e.g., *Prosser and Slade*, 1994] therefore has minimal impact on the results.

Using (8) and the inputs described above, the model predicts that τ_b at channel heads ranges from 100 to 590 dyn/cm^2. This reflects the scatter of a/b plotted against $\tan\theta$ for channel heads, but 81% of channel heads fall within the range of 150 to 400 dyn/cm^2 (Figure 4a). To define a single τ_c for channel

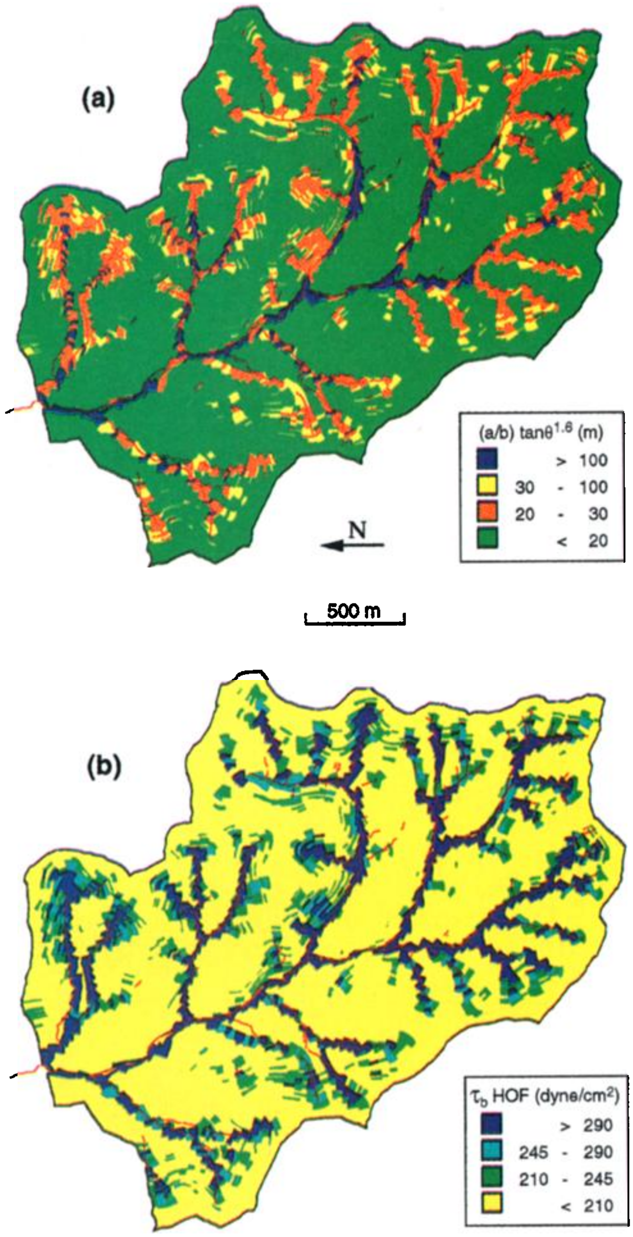

Plate 1. (a) Map comparing the topographic threshold, (a/b) tan $\theta^{1.6}$, with the channel network at Gungoandra Creek. (b) Map comparing the channel network at Gungoandra Creek with τ_b applied by Hortonian overland flow under degraded grass cover as predicted by (8).

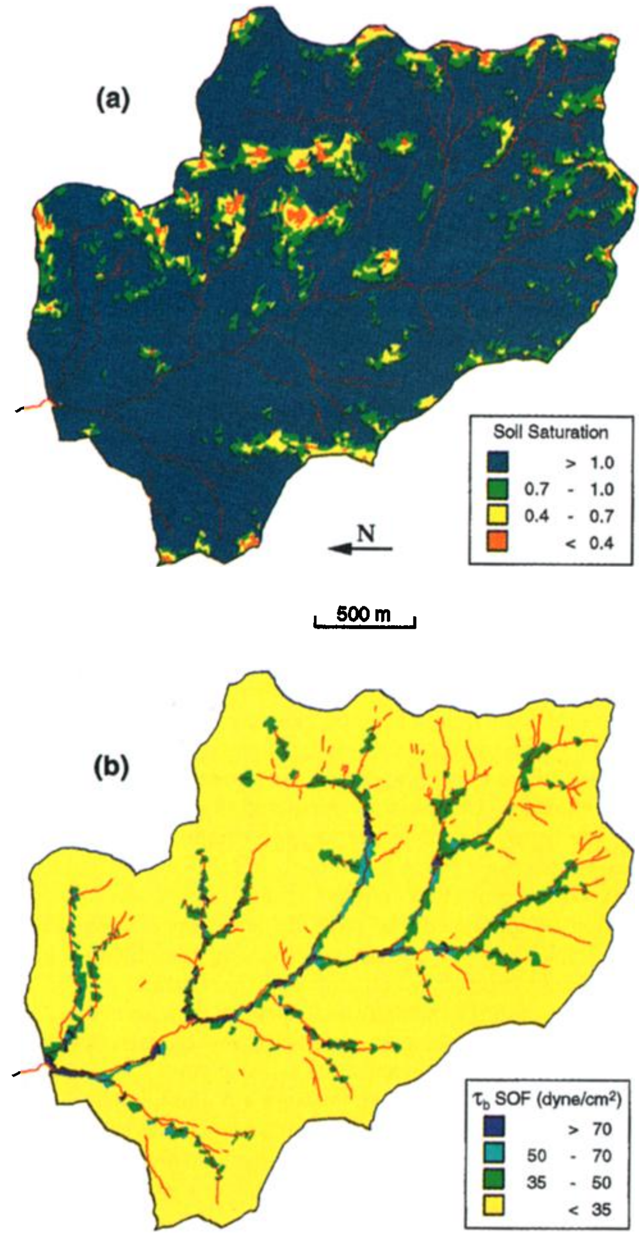

Plate 2. (a) Map of soil saturation predicted by (2) using T/q = 169 m. Values are subsurface drainage as a proportion of the capacity of the soil to transmit water. Values ≥1 indicate soil saturation. (b) Map of τ_b applied by saturation overland flow predicted using (6) and T/q = 1187 m for degraded grass cover.

incision is partly arbitrary because of the range of τ_b, but the median τ_b at channel heads of 245 dyn/cm^2 reproduces the topographic threshold of (10). Plate 1b shows that a τ_c between 210 and 290 dyn/cm^2 maps the channel network well, and τ_b increases below most channel heads. Some unchanneled valleys and steep slopes are predicted to contain channels. These generally have a south or east aspect, where greater soil moisture results in better vegetation cover and presumably greater resistance to incision. It was not possible to distinguish gullies from prehistoric channels on the basis of τ_b, so factors other than topography determine the type of channel observed.

Channel networks were poorly developed in most catchments of the Southern Tablelands before the introduction of modern agriculture [*Eyles*, 1977]. Swampy alluvial flats, with no continuous channels, dominated the trunk valleys, and only the steeper confined valleys supported channels [*Prosser et al.*, 1994]. Flume experiments conducted on a remnant swampy alluvial flat suggest that for the dense cover of tussock grass and sedge, f = 1 350 000 $Re^{-1.2}$ [*Prosser and Slade,* 1994]. Using this f/Re relationship and q = 22 mm/h, (9) predicts that the Gungoandra Creek catchment would be largely unchanneled if $\tau_c \geq$ 1000 dyn/cm^2 (Figure 4b).

Threshold of Saturation Overland Flow

Saturation overland flow may be an important process for channel incision at Gungoandra Creek during long winter storms. If the present channel network was eroded by saturation overland flow or by seepage of subsurface flow, then the ground above channel heads would saturate during storms. Therefore (2) must be satisfied, which is dependent upon the hydrological ratio T/q. Plotting (2) on a scattergram of a/b against tan θ shows that $T/q \leq 169$ m is required to saturate all channel head elements (Figure 4b). Saturated hydraulic conductivity data from similar soils in the region suggest that T lies in the range 0.1 to 3.4 m^2/d [*Hook*, 1990; *Mackenzie et al.*, 1991; *Lane et al.*, 1994]. Thus q lies between 0.6 and 20 mm/d to saturate all channel heads. The lower limit of q is similar to the mean daily rainfall for winter months, and the upper limit is similar to a 2-year recurrence interval, 72-hour duration, storm.

Mapping the extent of soil saturation for $T/q = 169$ m suggests that almost all of the catchment is saturated, including long planar slopes with gradients up to 0.9 (Plate 2a). These slopes are unlikely to saturate by the expansion of soil saturation from lower in the catchment, applying the definition of saturation overland flow of *Freeze* [1974]. Such extensive soil saturation is certainly not the average winter condition, as is implied by the lower end of the range of T and q. Rather than reflecting expansion of saturated areas from below, the almost complete extent of soil saturation probably reflects locally generated saturation as a result of low water storage capacity and transmissivity in the shallow clayey soils. Under relatively low T, runoff is more likely to be the result of short periods of intense rainfall, much of which will not infiltrate. Thus it is more appropriate to predict channel head position using (9) and a relatively high value for q: that is by assuming incision by Hortonian overland flow.

A more reasonable extent of soil saturation is predicted for the catchment if $T/q = 1187$ m, a value that saturates 50% of channel heads (Figure 4b). Using this value in (6), $q = 2.9$ mm/d, and the same f/Re relationship used for Hortonian overland flow ($f = 1500Re^{-0.7}$) gives a τ_b of saturation overland flow at channel heads of 0 to 97 dyn/cm^2 (Plate 2b). This is considerably lower than the range predicted for Hortonian overland flow using (8) and well below the $\tau_b > 200$ dyn/cm^2 required for sediment transport on degraded grass [*Reid*, 1989; *Prosser and Slade*, 1994; *Prosser et al.*, 1995]. Therefore channels are unlikely to be incised by saturation overland flow, but soil saturation could contribute to erosion of the lower-elevation channel heads through seepage processes.

Discussion

Analysis of the DTM shows that topography is a strong control on the limits of the channel network at Gungoandra Creek. Good definition of the position of channel heads was achieved using a threshold of a or a/b against tan θ (Plate 1a), and the form of that relationship is broadly similar to the inverse square relationship found for catchments of western North America [*Montgomery and Dietrich*, 1988, 1989, 1992, 1994b]. The topographic threshold of a and tan θ could also be derived from field measurements at channel heads, but the DTM allows rapid analysis of the whole catchment. The DTM may also provide a more accurate method as it is difficult to define the upslope catchment area in weakly convergent terrain such as this.

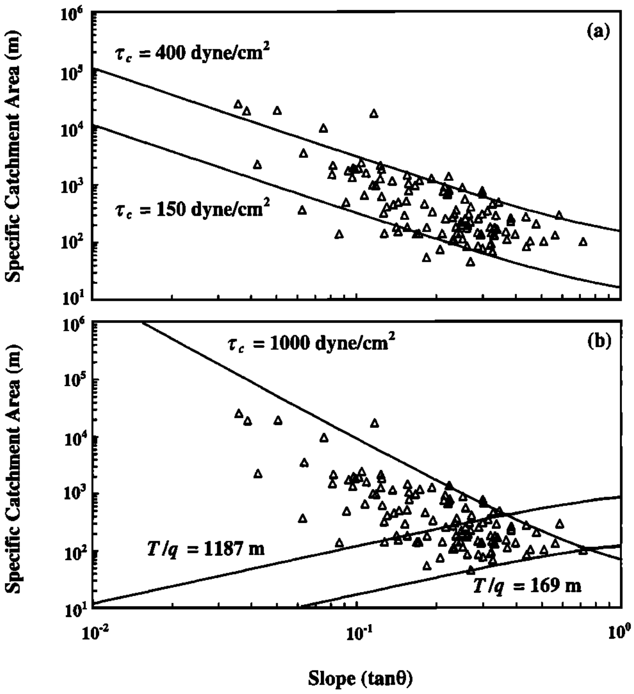

Figure 4. Specific catchment area and slope for channel head elements at Gungoandra Creek showing (a) thresholds for incision by Hortonian overland flow under degraded grass cover and (b) thresholds for incision by Hortonian overland flow under natural grass cover ($\tau_{cl} = 1000$ dyn/cm^2) and thresholds for soil saturation for $T/q = 169$ m and $T/q = 1187$ m.

Equations (2), (7), and (9) attempt to interpret the empirical topographic thresholds in terms of potential processes that formed the channel network. The representation of processes is highly simplified, but as will be argued below, it provides a useful means to discern catchment behavior. The assumptions of the model do, however, limit the level of interpretation and need to be addressed first.

The process thresholds assume uniform vegetation and soil properties across the catchment through the use of spatially uniform f, τ_c, and T. This does not mean that spatial variation in soil and vegetation does not contribute to the pattern of the channel network. The results merely imply that it is possible to constrain the channel network to within reasonable domains by considering topography alone. In many instances, variation in soils and vegetation are influenced by topography and are therefore implicitly modeled. On the basis of field observation, we varied T across the catchment in earlier simulations, but these refinements had little impact on model performance so the increased complexity was not warranted.

The model represents the topographically induced pattern of overland flow under typical channel-forming conditions but does not attempt to predict erosion from any individual event. During an intense 1-hour storm that would produce Hortonian overland flow, only the upper parts of the catchment would achieve anything near steady state conditions. The model will overestimate τ_b in the lower catchment, so it cannot be used to assess relative sediment transport capacity through the channel network. However, the channel heads are all relatively close to the divide and within the probable range of near steady state conditions. A dynamic model for overland flow being developed within the Topog framework (C. J. Wilson, personal communication, 1995) produced the same τ_b at gully heads as the steady state model (using a 1-hour storm and rainfall intensity of 42 mm/h), providing support for the steady state assumption.

The model calculates the mean τ_b across the lower boundary of an element. It was observed in the field that flow is usually restricted to the hollow axis; hence flow width is often much narrower than the width of a Topog element. This fine scale topography is not easily detected by a DTM, but it is important for channel incision because channels follow the axis of hollows closely. Thus the model underestimates the τ_b responsible for channel incision. At the Marin County catchment the τ_c for channel incision inferred from field experiments is equivalent to model predictions if flow width is constrained to within 7 to 11 m of a V-shaped hollow axis [*Prosser and Dietrich*, 1995]. Constraining flow width to 10 m at Gungoandra Creek produces a τ_b in the axis of hollows that is 3 times the model prediction for planar flow across 40-m-wide elements. These higher values will be compared to field estimates of τ_c below.

The model does not fully simulate processes operating at a channel headcut and therefore cannot be used to model headcut retreat. The modeled τ_b represents that of overland flow on an unchanneled surface, not within a channel, and thus predicts the relative tendency for overland flow to incise that surface. The presence or absence of seepage can be inferred from the prediction of soil saturation, but seepage processes themselves are not modeled. Similarly, plunge pool scour and mass failure of undercut blocks are not simulated. The model is most appropriate for gradual channels heads and those catchments where channels were initiated at many locations, such as at Gungoandra Creek.

Despite their deliberately simple representation, the process thresholds do provide valuable insights into channel incision at Gungoandra Creek which cannot be gained from the topographic thresholds alone. Comparing modeled τ_c values with results of field experiments implies that the channels formed under heavy degradation of the grass cover. Field experiments show that τ_b of 1050 to > 2400 dyn/cm^2 is required for sediment transport on undegraded grass plots and that considerably higher τ_b is required for channel incision [*Prosser and Slade*, 1994; *Prosser et al.*, 1995]. The model predicts that channels would be excluded from the catchment if $\tau_c \geq 1000$ dyn/cm^2, equivalent to a maximum τ_b of 3000 dyn/cm^2 in the axis of hollows. These predictions are consistent with historical evidence of unchanneled swampy valleys at the time of European settlement [*Eyles*, 1977; *Prosser*, 1991]. Taking into account the higher τ_b in hollow axes, the model predicts a τ_c for channel incision by Hortonian overland flow of 450 to 1200 dyn/cm^2, well below that for undegraded grass.

More precise validation of the predicted τ_c against field experiments on degraded grassland is difficult because experiments have defined conditions for significant sediment transport but have not produced channels. Thus the experiments provide only the minimum conditions required for channel incision. Experiments on grass clipped to the surface show that significant sediment transport begins at τ_b of 200 to 380 dyn/cm^2 [*Prosser et al.*, 1995]. It was inferred from these experiments that significant degradation of the root mat would be required for channel initiation, in addition to reduction of cover. A separate set of experiments conducted on ripped tussock grass showed that erosion began at τ_b = 700 dyn/cm^2 but discharge could not be maintained long enough to form a channel [*Prosser and Slade*, 1994]. Despite the limitations of the experiments the model predictions are consistent with channel initiation into a surface with degraded root mat and strongly reduced surface cover, conditions implied by the historical evidence.

Further experiments will improve our understanding of channel incision, but it will always be difficult to provide independent values of τ_c for use in the model. Model predictions are very sensitive to τ_c which will vary spatially and temporally, making it difficult to constrain. In addition, there is the problem of differences in measurement scale between model and experiment. Therefore it is probably more appropriate to calibrate the model on a stable channel network, as we have done here, and to recognize that the chosen τ_c incorporates all the model assumptions and simplifications. A similar approach is often taken when specifying flow resistance in simulations of catchment hydrographs [*Moore and Foster*, 1990].

The model appears capable of distinguishing the likely processes that control the channel network. As discussed above, simple representation of Hortonian overland flow produces a range of τ_b consistent with incision into degraded pastures. In contrast, simulation of saturation overland flow is only reasonable above 50% of channel heads, and τ_b is so low at these channel heads that it is unlikely to be responsible for channel incision. Therefore soil saturation is only likely to contribute to erosion of these channels through seepage processes. These predictions match field observations, where the higher-elevation channels have gradual heads with evidence of widespread surface wash erosion upslope. Lower-elevation channel heads are sharper and seepage has been observed after winter storms.

Some of the channels at Gungoandra Creek extend beyond the valleys and onto steep planar slopes, whereas the channel network in Marin County is confined within the catchment's valley network [*Montgomery and Dietrich*, 1992]. This suggests that the present channel network at Gungoandra Creek has extended beyond the average network extent responsible for incision of valleys. Furthermore, (10) shows that for any given slope, channel head catchment areas are somewhat lower at Gungoandra Creek than at the Marin County catchment (where $a = 1980 \tan \theta^{-1.65}$ [*Montgomery and Dietrich*, 1989]). Lower catchment areas at channel heads do not necessarily imply more severe land degradation at Gungoandra Creek since the predicted τ_c for incision is similar in both catchments. The lower topographic threshold does, however, reflect a difference in runoff process, for at Gungoandra Creek all parts of the source catchments generate overland flow. In essence, it is the continuation of channels onto more planar parts of the landscape that suggests Hortonian overland flow as the process of channel incision in the catchment.

Conclusions

One aim of the study was to investigate whether the topographic model for channel incision by overland flow proposed by *Dietrich et al.* [1992, 1993] could be applied to a catchment significantly different from the Marin County catchment where it was first developed. We found that the channel network at Gungoandra Creek was well defined by a topographic threshold of catchment area and slope. Furthermore, this topographic threshold could be interpreted in terms of a simple process model of threshold-controlled incision by overland flow. In common with the Marin County catchment, we found that model predictions of incision by overland flow were consistent with the historical evidence for channel incision under heavy grazing and degraded grass cover. In contrast to the Marin County catchment, the model indicates that Hortonian overland flow, rather than saturation overland flow, was responsible for channel incision. This is consistent with observed and perceived runoff processes in both catchments, and it suggests that the model can distinguish processes of incision where these are reflected in the relationship between channel network and landscape morphology.

Field experiments conducted elsewhere on the Southern Tablelands have shown that widespread gully erosion can be accounted for solely by degradation of valley floor vegetation, without need to invoke extreme climatic events, or increased hillslope discharge [*Prosser and Slade*, 1994]. The digital terrain analysis of Gungoandra Creek confirms this interpretation, with the advantage of providing a means to investigate the spatial extent of gully erosion which could not be achieved from the field experiments alone.

A second aim of applying the model to Gungoandra Creek was to evaluate its potential to predict the relative risk of gully erosion in catchments of the region. We believe that gully erosion at Gungoandra Creek has reached its limits, and the model predicts these limits well. The topographic threshold is a purely empirical approach to predicting gully erosion, but the threshold for incision by Hortonian overland flow is based on simple physical principles, requiring inputs of runoff per unit area, infiltration capacity, surface flow resistance, and τ_c. The model is insensitive to all but the last of these inputs, and they are relatively easily defined. However, the process threshold remains partly empirical because of the difficulty of independently specifying τ_c. Application of the model to Gungoandra Creek indicates that it has considerable potential, but it remains to be seen whether the thresholds derived here can be used to predict the location of future erosion in catchments where gully erosion is active.

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