Landscape instability in an experimental drainage basin

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ABSTRACT

Do drainage basins develop static river networks when subject to steady forcing? While current landscape evolution models differ in formulation and implementation, they have the common characteristic that when run for long times at constant forcing, they evolve to a static steady-state configuration in which erosion everywhere balances uplift rate. This results in temporally stationary ridge and valley networks. We have constructed a physical model of a drainage basin in which we can impose constant rainfall and uplift conditions. The model landscapes never become static, and they are not sensitive to initial surface conditions. Ridges migrate laterally, change length, and undergo topographic inversion (streams occupy former ridge locations). Lateral stream migration can also produce strath terraces. This occurs without any change in external forcing, so the terraces must be considered autocyclic. The experimental drainage basin also exhibits autocyclic (internally generated) oscillations in erosion rate over a variety of time scales, despite constant forcing. The experimental landforms are clearly not perfect analogs of natural erosional networks, but the results raise the possibility that natural systems may be more dynamic than the current models would suggest, and that features like strath terraces that are generally interpreted in terms of external forcing may arise autocyclically as well.

Keywords: landscape evolution, strath, topographic inversion, autocyclic.

INTRODUCTION

Historically, the idea of a static erosional network with uplift everywhere balanced by erosion has deep roots. Playfair (1802) described drainage basins as trees, each stream delicately adjusted such that at each joining of streams, the slopes were delicately balanced. The systematic change of slope within landscapes suggested to Playfair that an equilibrium existed between erosion and sediment transport over the entire basin, and a stable geometry resulted from this balance.

Gilbert (1877) noted that erosional landforms have convergent stream networks and divergent ridge networks, and proposed that the typical concave-up profile of streams is due to the increased volume of water moving through downstream sections in the drainage network. He postulated that divides between adjacent streams must migrate toward the stream with a shallower gradient; stable channel networks are achieved once gradients in adjacent streams are similar. Instability of drainage lines could be explained in terms of differential resistance to erosion, differential uplift, time, and possibly the interaction between stream transport capacity and availability of sediment for transport. For Gilbert, the network of streams and hillslopes is a strongly interactive system, delicately adjusted at dynamic equilibrium to a stable form.

Strahler (1950) characterized erosional landscapes as open mass-transport systems that adjust their morphology to attain a time-independent form. He measured valley-side slope angles from several completely dissected natural drainages, and showed that a given area maintains a characteristic slope with a narrow range of values. The presence of a characteristic slope lends support to the hypothesis of a stable landform.

Hack (1960) hypothesized that every stream-hillslope pair is adjusted one to the other, and given constant forcing conditions, all elements of the landscape erode at the same rate, similar to Gilbert's dynamic equilibrium. Differences in form could, under those conditions, be related only to differences in resistance to flow, such as variable lithology and vegetation. Changes in the form could also result from changes in the forcing conditions, but responses to perturbations were fast enough to restore a dynamic steady state adjusted to the new boundary conditions. He explicitly viewed landscapes as spatial structures with time-independent forms.

NUMERICAL FORMULATIONS OF LANDSCAPE EROSION

In general, erosion is controlled by the resistance of the substrate to surface and body forces. The resistance is set by rock properties (crystal structure and chemistry, rock strength or cohesion, grain size), vegetative cover, and degree of saturation. Applied forces vary widely across natural landscapes: the scratching paws of burrowing animals; the pounding impact of rain drops; the torque of bending trees under the wind's impulse; soil expansion and contraction during saturating events and freeze-thaw; episodic failures of oversteepened slopes; the thrashing torrents of streams; and the grinding mass of a sliding glacier, to name a few. The rate of disintegration of crystalline bedrock to smaller particles also sets a limit on the availability of transportable material. For modeling purposes, simplification is required at some level to obtain solutions to mass transport across the landscape.

Several landscape erosion models have been developed in recent years (Willgoose et al., 1991; Kramer and Marder, 1992; Chase, 1992; Leheny and Nagel, 1993; Howard, 1994; Tucker and Bras, 1998, for overview of models), and they differ in the means and degree of simplification of erosional processes that they employ. However, all of the models assume that forces applied by surface runoff dominate erosional processes in landscapes, and thus are based on routing water down the steepest slope in a numerical grid. Runoff is treated as steady, uniform flow, and this assumption allows the use of upstream drainage through a point as a proxy for stream flow at that point. Calculation of erosion depends on assumptions concerning the availability and transportability of sediment, but in general erosion rate is a function of local slope, discharge, uplift rate, and substrate resistance. The models develop landscapes with branching stream networks similar to natural drainage patterns (Chase, 1992; Leheny and Nagel, 1993), and even capture transient evolutionary features of the network such as extension and abstraction of streams (Glock, 1931; Kramer and Marder, 1992).

Models based solely on stream erosion, however, develop slopes approaching infinity near drainage divides. While some natural landscapes have near vertical slopes, most hillslopes flatten near ridge crests. If a diffusive short-length scale process is added to the models, the resultant land-

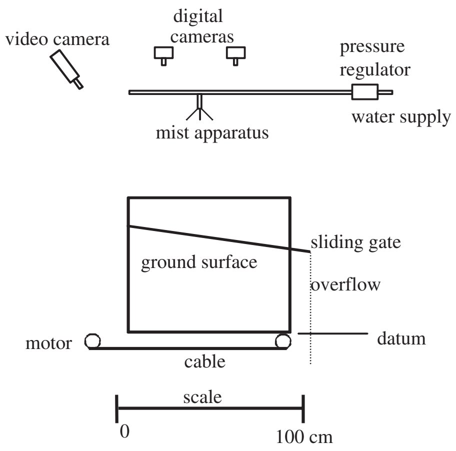

Figure 1. Schematic cross section of erosional facility.

Figure 2. Vertical photograph of drainage basin nearing complete dissection. Note that upper right corner retains remnants of initial flat surface. Basin is 87 cm wide, 98 cm long. Original digital photograph is 1280 × 960 pixels.

forms develop rounded ridge crests, and the mean interfluve distance increases (Chase, 1992; Howard, 1994; Tucker and Bras, 1998). Local diffusion is meant to capture effects such as raindrop impact, bioturbation on hillslopes, and hillslope failures. We note that while the diffusive terms change the form of the numerical landscape, the models nonetheless achieve static landscapes at constant forcing.

Numerical models suggest that the only requirements for a dendritic network to develop are detachment and removal of material from the system as water flows downhill. Furthermore, the strong negative feedback between stream erosion and slope dampens perturbations and drives landscapes to stable forms over longer time scales. Constant boundary conditions inevitably result in stable networks. Thus, numerical models appear to validate conjectures by geomorphologists that landscapes can attain static perfectly adjusted forms.

Tests of landform stability under constant forcing are difficult to conduct on natural landscapes because of the long time scale for significant landform change, and the uncertainties in climatic and tectonic history. Physical models of eroding drainage basins (Parker, 1977; Phillips and Schumm, 1987; Schumm et al., 1987; Hancock, 1997) represent a possible means of testing landscape stability, because steady climatic and tectonic boundary conditions can be enforced, and lithologic variability can be minimized. Numerical formulations of landscape erosion based on discharge and slope do not contain an inherent length scale, so in principle they should apply to both small (i.e., sandbox) and large (i.e., continental) scales.

Although physical experiments have been used for some time to study drainage-basin evolution (Parker, 1977; Phillips and Schumm, 1987; Schumm et al., 1987; Hancock, 1997), to date no physical model has been developed to monitor landscape change under constant forcing conditions for extensive erosional periods. Numerical results predict that a landform should attain a static geometry after eroding through a vertical distance approximately three times the total instantaneous elevation range (Howard, 1994). We have built a physical apparatus capable of monitoring landscape evolution over this length scale. We note that while eroding through this length may seem extraordinary, fission-track data from South Island, New Zealand, imply an average denudation of ~10 km in the past 10 m.y. (Tippett and Kamp, 1993). Local relief is ~3 km, so, assuming temporally steady relief, about three relief units have been eroded.

EXPERIMENTAL SETUP AND MONITORING

We constructed our experimental system to monitor drainage-basin behavior at steady rainfall and uplift rates. The main purpose of our experiment was to provide a physical test for numerical landscape evolution models. We have attempted to match the simplifying assumptions that go into the models: erosion is controlled by surface runoff, uplift rate, and substrate resistance to surface and body forces. In natural landscapes all of these quantities vary spatially and temporally. We tried to remove complications associated with random variability, and thus to isolate erosional behavior that results from the elementary dynamics of the eroding system. A simple way to do this is to maintain a uniformly erodible substrate, and force the system at constant rates. We provided the model system with one outlet in the expectation that this would lead to a simply structured dendritic network.

The experimental device consists of an oval tank ~1 m in diameter and 1 m deep with a single outlet dammed by a motor-controlled gate (Fig. 1). The motor operates continuously and drops the outlet at a slow, constant rate, in effect uniformly lifting the basin relative to base level. The height to width ratio of the tank permits relative uplift of three to six times the instantaneous drainage relief, defined as the maximum elevation above the outlet. A mister sprinkles rain (droplet size <200 μm) over the basin to generate runoff. The experimental basin forms third- to fifth-order drainage networks. Stream incision and transport dominate erosional processes in the experiment, but we also observe hillslope failures.

For each experiment, well-sorted silica silt (median grain diameter [d_{50}] = 45 μm) is mixed with kaolinite (100:1 by weight) and water in a cement mixer, poured into the basin, and allowed to settle overnight. The settling process produces a flat surface pocked with small sediment volcanoes (<4 cm diameter) generated by groundwater overpressuring during loading. The initial runoff pattern on this surface is essentially random.

Prior to the inception of uplift, rainfall is spatially calibrated by collecting rainfall in pans distributed over the basin. Rainfall is generated by a commercial radial mister. Oscillating fans above the mister break up persistent rainfall pat-

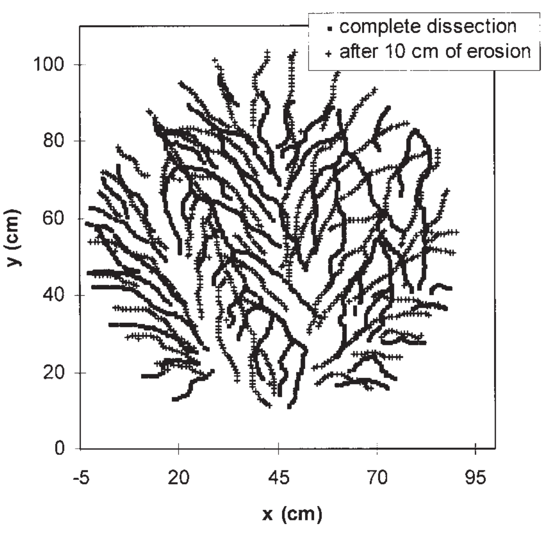

Figure 3. Ridge networks for two sequential photographs. We sketched ridge segments on digital photographs (1280 × 960 pixels), extracted photographic coordinates of ridges, and projected coordinates to ground reference frame using three-dimensional projective coordinate equations. Eroded distance between photographs is 10 cm.

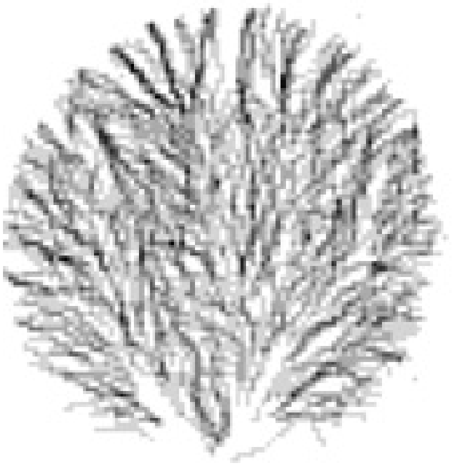

Figure 4. Density plot of cell occupation by ridge. We counted number of times that 1 cm^2 cell was occupied by ridge in eight photographs after complete dissection, and plotted as gray scale. Dark regions have been more frequently occupied by ridge. Few ridges have maintained position, and these tend to occur at basin margins. Note that maximum value that cell can have is 8. No cells in plot have values exceeding 6. Basin width is 87 cm.

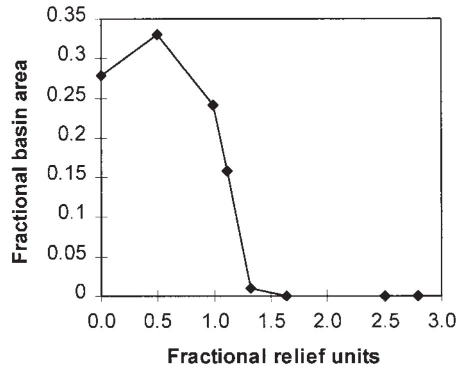

Figure 5. Fraction of basin area occupied continuously by ridge. Eroded distance since complete dissection was normalized to H_r (H_r = eroded distance from initial flat surface to complete dissection). Note that by 2 H_r, no cell has been continuously occupied.

terns, and produce a reasonably uniform rainfall distribution with a spatial variation ~12% (standard deviation/mean) for a measurement interval of 12 min.

A run is initiated by starting the motor-controlled outlet and turning on rainfall. The run is terminated when the outlet reaches the bottom of the tank. In the meantime, we monitor landscape development with still photographs and time-lapse video, and measure sediment flux rates at the outlet of the basin by capturing basin effluent in a cylinder of known volume, and recording the weight and time required to fill the cylinder.

EXPERIMENTAL RESULTS

After a run is started, streams incise from the outlet and extend to the edges of the basin. Trunk streams near the outlet develop knickpoints that migrate upstream as waves, frequently triggering hillslope failures. A statistical balance between uplift and erosion is reached soon after complete dissection of the initial surface, as indicated by the sediment output measurements. Although we have done several runs that show comparable behavior, here we focus on one run, conditions for which are given in the Appendix.

We sketched ridge crests on photographs for eight sequential times after complete dissection to monitor landform stability (Fig. 2, a vertical photograph of the drainage basin nearing complete dissection of the initial flat surface). We chose ridge crests because they should be the most stable feature in a landscape, and because they are easier to identify in photographs than streams. Figure 3 is a plot of ridge crest location for two times, the first just after complete dissection, and the second after an additional 10 cm of erosion. Clearly, ridges have migrated, and several ridges (right side of Fig. 3) cut across previous ridge locations at angles approaching perpendicular. This plot is for times just after complete dissection, and one might consider such ridge migration as an early adjustment in an approach to a stable form. For a more comprehensive view, we collected all ridge location coordinates from eight photographs into a single grid (1 cm spacing), and ranked each cell in the grid according to the number of times a ridge occupied the cell (Fig. 4). Hence, the highest rank a cell could have is 8, and the lowest is 0. Darker regions delineate persistent ridge locations. Note that no cell is occupied continuously by a ridge.

To place ridge migration into a scaled context, we define a length scale, the relief unit (H_r), as the range in elevation of the landform at complete dissection, in this case, 21 cm. The photographic sequence of Figure 4 covers ~3 H_r of erosion. Figure 5 shows that the fractional basin area continuously occupied by a ridge rapidly decreases beyond 1 H_r, and by 2 H_r is essentially zero. Note that the increase in fractional area containing a ridge at H_r = 0.5 indicates that total ridge length (a proxy for drainage density) varies temporally.

Stable ridge locations require a uniform erosion rate and similar slope conditions on both sides of the ridge. These conditions are rarely met in the experimental basins, for several reasons. The migration rate of knickpoints is not uniform through the network. Their passage frequently destabilizes a hillslope, inducing hillslope failure and slumping. The ridge crest moves away from the slumping region. Increased sediment discharge from slumping overloads streams locally, and incision rates decrease temporarily. This results in sediment flux oscillations and migration of divides. Lateral ridge migration can result in topographic inversion: stream locations become ridges, and vice versa. Geomorphic signatures of ridge migration within our physical experiment include strath terraces (stream eroded flattened sections above the valley floor), islands in trunk streams, and stream piracy. Apparently, these features do not require changes in climate or tectonic forcing, because these are kept constant in our experiment. This observation suggests that a possible autocyclic origin of these features must be borne in mind when using them to infer climatic or tectonic changes in natural settings.

An additional measure of temporal variation in erosion rates is sediment yield at the outlet. Sediment fluxes (Fig. 6) oscillate 15%–25% around the average erosion rate. One of the causes of variability in sediment output is knickpoint migration. Using time-lapse video, we counted at least

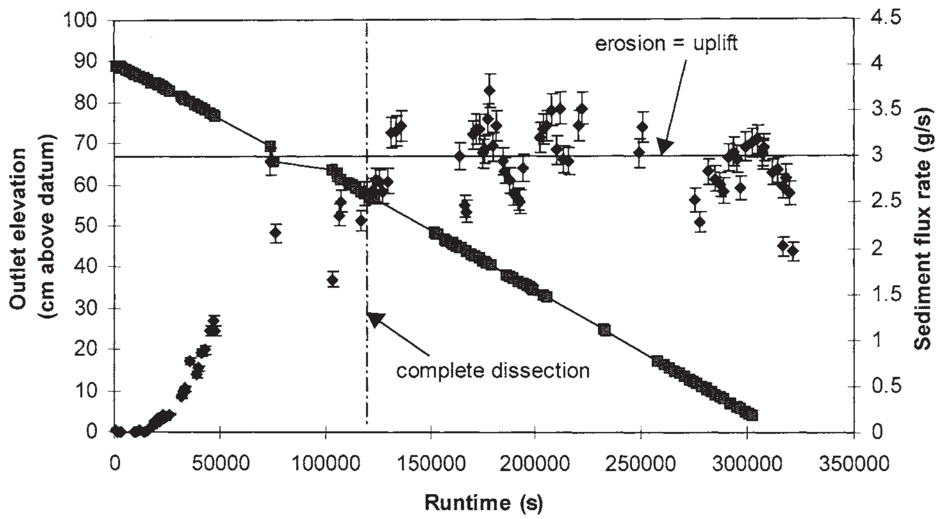

Figure 6. Base-level curve and sediment flux at outlet. Measurements of bulk density were deconvolved into water and flux rates using mixing relation, density of water and sediment grains (quartz), and fill time. Measurement errors for volume are 1%; for mass, 0.5%; and for time, 5%. Error bars are plotted as 5% of calculated mass rate. Elevations of outlet were taken from fixed tape measure by outlet. Squares—outlet elevation; diamonds—sediment flux rate.

30 knickpoints generated over a vertical erosion distance of 60 cm, or roughly 3 H_r. This suggests that about 10 knickpoints formed per H_r. The initial height of the wave is ~5 mm (approximately the flow depth at the outlet), or 2%–3% of H_r, and 1%–10% of the local valley-ridge relief. Assuming that the eroding wave maintains a constant height, erosion from kinematic waves may account for as much as 30% of total erosion rate. Time-lapse video also revealed that while the knickpoints tend to propagate uniformly upstream, hillslope failures interfered with knickpoint propagation, at times drowning the knickpoint. We stress that knickpoint generation cannot be attributed to abrupt base-level drops, because the outlet drops continuously (Fig. 6). The knickpoints appear to form spontaneously, and may be a result of instabilities in flow fields close to critical conditions (Parker and Izumi, 2000).

DISCUSSION

Our experimental results are quite different from the static behavior exhibited by numerical models. A possible source of the discrepancy could be residual random variability in rainfall and substrate resistance in our model. Temporal variability of rainfall in our basin was caused by water-pressure fluctuations in the laboratory. However, the time scale of ridge migration is much longer than that of rainfall fluctuations. There may also be minor variability in substrate resistance in the experiments, but it is difficult to see how this could directly cause systematic ridge migration.

Do natural eroding drainage basins behave like our model? Our experiment does not incorporate vegetation, chemical weathering, or orographic effects. Furthermore, there are scale distortions between our model and natural systems. The ratio of basin size to grain size in the experiments, although large, is nonetheless much smaller than in most natural systems. Sediment concentration in experimental streams (set by uplift/rainfall rate and substrate density) approaches 25%. Such concentrations are not unprecedented, but are unusual in natural rivers. Feedback between local deposition and erosion may be enhanced at this concentration, and could be a source of instability within model streams. The experimental streams are laminar (maximum Reynolds number ~750), whereas natural mountain streams are always turbulent. We have observed local small standing waves in model flows, which suggest that flows can be supercritical (Froude number >1). Natural mountain streams are frequently supercritical. Individual mass flows (hillslope failures) in the experiments have length scales in the range of 10 cm^2, so the ratio of event size to basin size is 0.0016. This ratio is plausible for small natural drainage basins but probably not for large ones.

We do not think that any of these scale effects would cause qualitatively different behavior in the experimental system than in natural ones. Furthermore, because current landscape models are formally scale independent, we consider our experiments to be a valid test of them. An obvious next step is to develop a model capable of predicting the existence of dynamic steady-state landscapes. Such a model would presumably be capable of predicting changes in the dynamics as a function of scale and forcing rate, and so would provide a means of resolving the scaling question. Experimentally, a clear next step is to do experiments like the ones described here at larger scales to test scale dependence of dynamic behavior. In the meantime, our experimental results suggest an alternative and much more dynamic view of the behavior of steady-state landscapes than that presented by the current suite of numerical models. The numerical models are clearly capable of capturing an average topographic form, but these models may not adequately represent the dynamic behavior of eroding landscapes.

APPENDIX

Model Run Parameters

Eroded distance:	80 cm
Rain rate (r):	6.5 μm/s
Uplift rate (u):	2.8 μm/s
Substrate density (ρ_b):	1.8 g/cm^3
Rainfall density (ρ_w):	0.998 g/cm^3
Grain size (d_{50}):	45 μm
Basin area:	6215 cm^2
Silt/clay (by weight):	100
Relief unit (H_r) (maximum relief):	21 cm
Maximum local valley-ridge relief:	8 cm
Maximum knickpoint height:	~0.5 cm
Average ridge density (total ridge length/basin area) (1/cm):	~0.16
Uplift/rainfall forcing ratio:	0.78 $(u\times\rho_b)/(r\times\rho_w)$ (dimensionless)

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