Channel changes in badlands

ALAN D. HOWARD *Department of Environmental Sciences, University of Virginia, Charlottesville, Virginia 22903*
GORDON KERBY *Environmental Quality Division, Virginia Division of Highways and Transportation, Richmond, Virginia 23219*

ABSTRACT

Stripping of the vegetation and soil from a 13-hectare site in Virginia underlain by coastal plain sediments created a rapidly evolving badland topography. Two types of channels developed: (1) sand-bed alluvial channels were graded to transport the bed material load supplied from slope erosion with available runoff, but they also generally eroded their beds slowly, and (2) steeper, bedrock-floored channels incised rapidly. In bedrock channels the erosion rate was proportional to the 4/9ths power of drainage area and the 2/3rds power of gradient. These exponents are consistent with a model in which the erosion rate is proportional to the bed shear during high flows.

Due to rapid mass wasting and reduced runoff, the alluvial channels became as much as 50% steeper during the winter than the summer, with an attendant yearly cycle of winter aggradation and summer entrenchment. The gradients, their seasonal variability, and their downstream hydraulic geometry were consistent with the predictions of total load transport formulas for sand beds and high loads. The hydraulic geometry of alluvial channels in the Virginia badlands were similar to that on the Morrison Formation in the western United States.

INTRODUCTION

The long time scale over which processes must act to produce most landscapes means that their cumulative effects are not easily measurable over the human time scale. In particular, the downcutting of bedrock-floored channels and the readjustment of channel gradient in response to climatic change or human interference generally occur so slowly that models of these processes are crude and of uncertain validity.

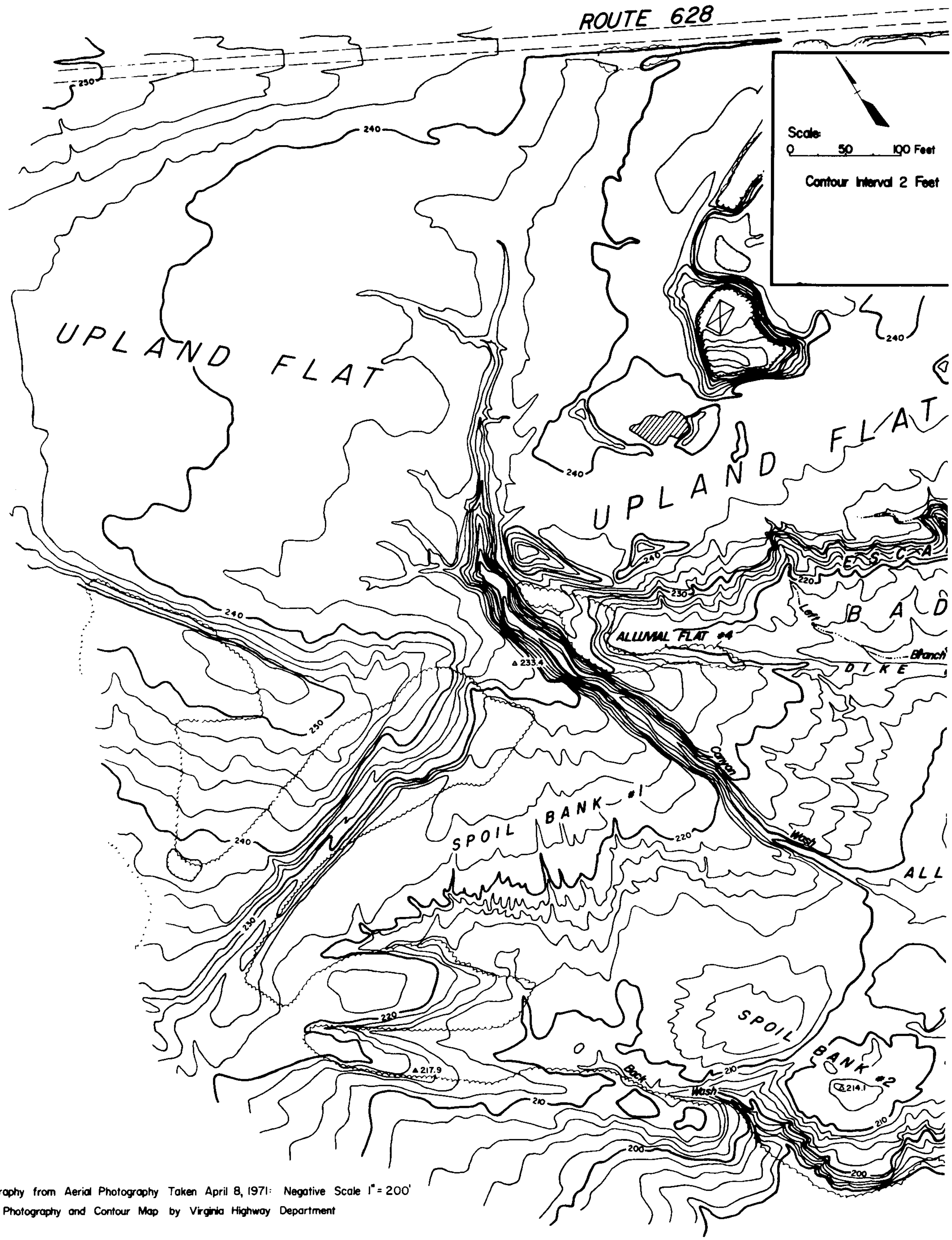
Occasionally, nature presents us with a microcosm, scaled-down temporally and/or spatially while maintaining many essential features of larger systems. Such appears to be the case in natural and man-induced badland landscapes, where the absence of vegetation and the soft rocks greatly accelerate erosion and landform development. Also, owing to the enhancing efficiency of runoff erosion as compared to mass wasting, the drainage density is high. Thus, badlands have been intensively studied both because of their morphometric similarity to large-scale landforms and because of the possibility of short-term measurements of process rates (Schumm, 1956b). Curiously, little attention has been directed to the mechanics and dynamics of badland fluvial features; most previous studies have concerned either slope processes (Schumm, 1956a, 1964, 1967; Schumm and Lusby, 1963; Engelen, 1973) or the evolution of the landscape and the drainage pattern as a whole (Schumm, 1956b), although Schumm (1962) measured rates of erosional regrading of miniature badland pediments. Yet, the processes acting in the channels within badlands are probably more directly comparable to those in larger streams than are badland slopes analogous to vegetated or alpine slopes.

Data from two sets of badland channels are presented here; most observations were made in a channel system that developed over a period of about 15 yr in poorly consolidated Tertiary sediments exposed in an abandoned borrow pit along Interstate 95 near Stafford, Virginia (Fig. 1). A few comparative observations were taken in natural badlands near the Henry Mountains, Utah; these badlands are discussed more fully in Howard (1970). The various small areas of man-induced badlands in the eastern United States (for example, the Perth Amboy badlands studied by Schumm, 1956b) afford the advantage of rates of landform evolution much greater than those in natural badlands, for the yearly rainfall may be more than 20 times greater. Badlands of the two regions differ also in their degree of response to seasonal climates. In most natural badlands, the slowness of processes means that slopes and channels show only minor changes due to seasonal contrasts in runoff and mass wasting, whereas marked seasonal alterations in slope morphology (Schumm, 1956b) and channel gradients (this paper) characterize the eastern badlands. Climatic changes acting over several decades or more would be required in western badlands to produce effects equivalent to those resulting from seasonal changes in eastern badlands.

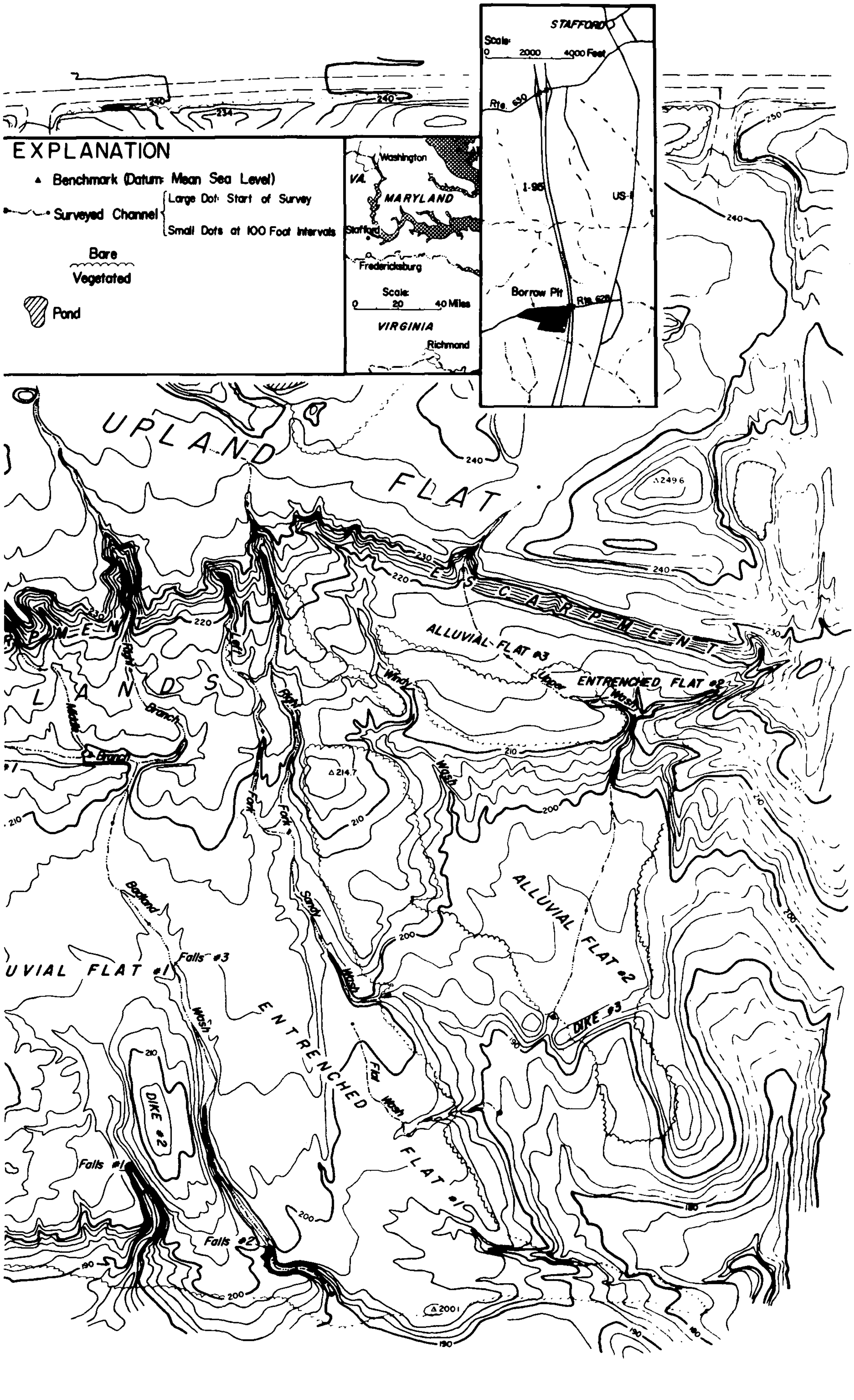
The rapidity of evolution of the borrow-pit channels has permitted quantification of two types of stream processes normally too slow-acting for direct observation: (1) erosion of bedrock-floored channels and (2) fluctuation of gradients of alluvium-floored channels in response to seasonal and long-term variation in runoff and sediment load. This division into *bedrock* and *alluvial* channels corresponds to a natural-process threshold separating two channel morphologies with distinct hydraulic behavior and negligible intergradation of form (Howard, 1980). The alluvial channels are carpeted with a blanket of sand that is transported during runoff events (Figs. 2B, 3). The steeper bedrock channels lack a sediment bed (Figs. 2A, 3), except for sand and pebbles filling isolated scour holes. Downstream transitions between these two types of channels are abrupt (Fig. 3). The presence or absence of an alluvial bed is determined primarily by the gradient; alluvial channels have, in general, lesser gradients than bedrock channels for the same drainage area (Fig. 4). In addition, the gradients of alluvial channels have a well-defined inverse dependence upon drainage area, whereas the two factors are largely inde-

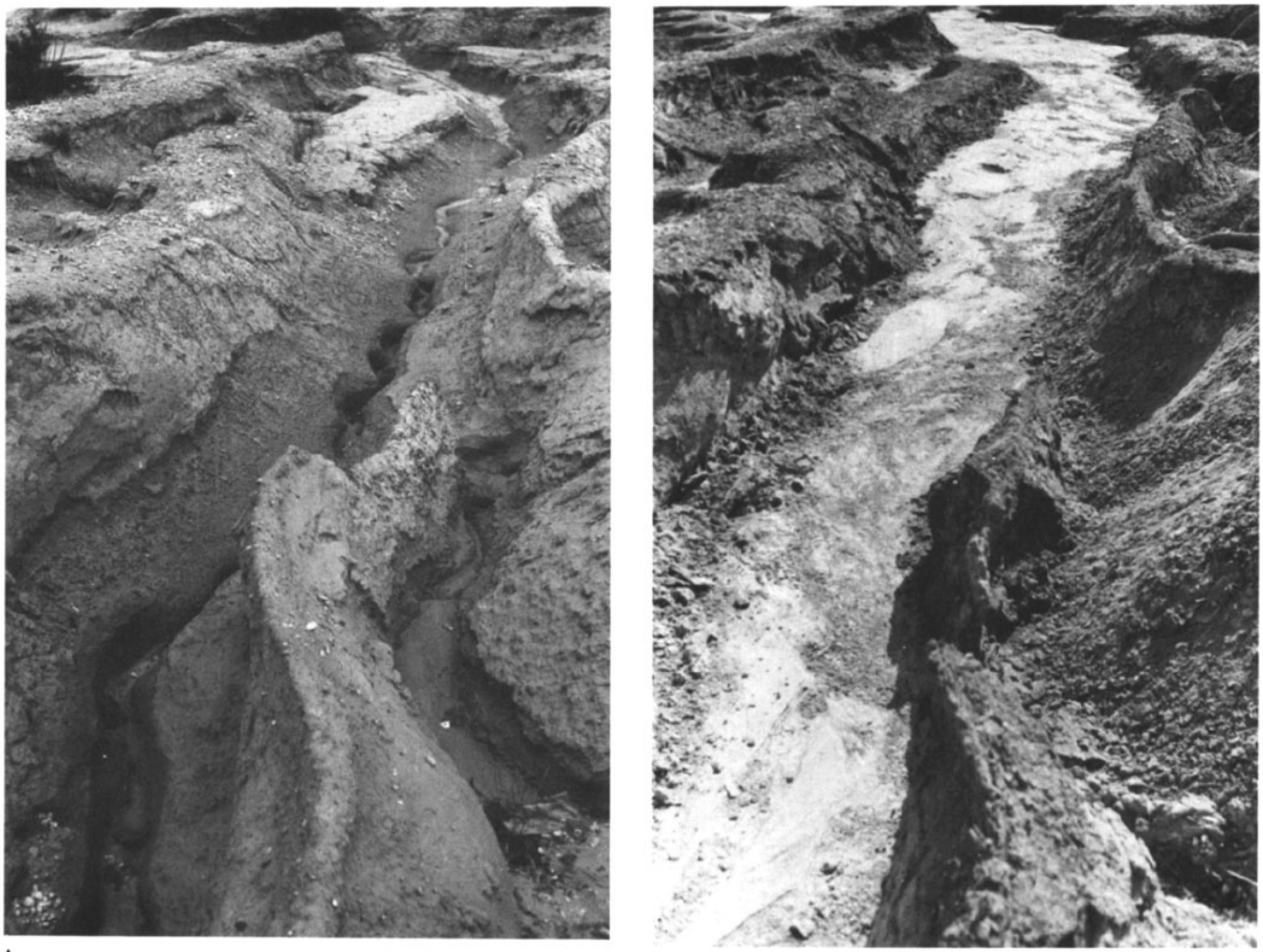
Geological Society of America Bulletin, v. 94, p. 739–752, 13 figs., 4 tables, June 1983.

Figure 1. Topographic map of the borrow pit study area following 8 yr of erosion. Place names assigned for convenience.



Topography from Aerial Photography Taken April 8, 1971: Negative Scale 1" = 200'
Aerial Photography and Contour Map by Virginia Highway Department


EXPLANATION
Benchmark (Datum: Mean Sea Level)
Surveyed Channel
Large Dot: Start of Survey
Small Dots at 100 Foot Intervals
Bare
Vegetated
Pond
Washington
VA.
MARYLAND
Stafford
Fredericksburg
Scale:
0 20 40 Miles
VIRGINIA
Richmond
STAFFORD
Scale:
0 2000 4000 Feet
Rte. 630
I-95
US-1
Borrow Pit
Rte. 628
UPLAND FLAT
ESCARPMENT
ALLUVIAL FLAT #3
ENTRENCHED FLAT #2
Upper Wash
Windy Wash
Right Fork
Left Fork
Middle Branch
Right Branch
Sandy Fork
Sandy Wash
Badland Wash
Falls #3
UVIAL FLAT #1
ENTRENCHED FLAT #1
ALLUVIAL FLAT #2
DIKE #3
DIKE #2
Flat Wash
Falls #1
Falls #2
Δ249.6
Δ214.7
Δ200.1



A B

Figure 2. Seasonal changes in the Left Branch of Sandy Wash (looking downstream). (A) October 1970, showing sinuous bedrock channel and slopes smoothed by rain splash and runoff. Channel changes abruptly to alluvial near top of picture. A large remnant of a winter terrace occurs on the left bank near the top of the picture. (B) March 1971, showing aggraded alluvial channel, slopes loosened by needle-ice erosion, and debris accumulating in channel. The surface of the alluvial channel has been slightly roughened by needle ice.

pendent for bedrock-floored channels (Fig. 4).

Field Sites and Measurement Program

The correlation of rock units within the borrow pit is based upon field work by Wayne Newell (1981, personal commun.). The oldest units were exposed along the channel walls below Falls 3 along Badland Wash and below Falls 1 on Canyon Wash (Fig. 1); they were predominantly cohesive, cross-bedded, clayey sands with occasional thin, discontinuous hardpans and cemented stringers of pebbles, and they were part of the Potomac Group. The Entrenched Flats 1 and 2 were underlain by similar clayey sands. The Badlands area was underlain by a mottled, greenish-gray, glauconitic, silty sand with weak cohesion and a pronounced vertical jointing (Aquia Formation). A few discontinuous, well-fractured hardpan layers also occurred within this unit, which formed the lower slopes of the Escarpment. Capping the Escarpment was a layer of fissured clay about 1 to 2 m thick (Calvert Formation). The small unexcavated remnants of the original surface above the Upland Flat were underlain by reddish, sandy, gravels that are correlated simply as "Upland Gravels." The Upland Flat was floored by residual "desert pavement" gravels from this formation. The Spoil Banks and Dikes were material displaced during borrow operations, mainly clay from the Calvert Formation. The Alluvial Flats were underlain by a thin cover of aggradational sand.

Several sources of data recorded the evolution of the badland topography. Near the cessation of excavation in 1963, a series of profiles spaced at intervals of 25 ft (8.2 m) and running perpendicular to Dike 1 (Fig. 1) were surveyed by the Virginia Department of Highways to measure the volume of the excavated sediment. Although the three dikes and the spoil banks were emplaced after the survey, most of the borrow pit was not subsequently regraded. In 1971, the Virginia Department of Highways took low-level aerial photographs

Figure 3. Transition from bedrock (foreground) to alluvial channel above a short falls on Sandy Wash below the junction of the Right and Left Branches. The March 1970 photograph (looking upstream) shows angle-of-repose side slopes shaped by needle-ice wasting and infilling of the channel bottom. Runoff from one strong rain could re-expose the bedrock floor. The Escarpment appears in the background.

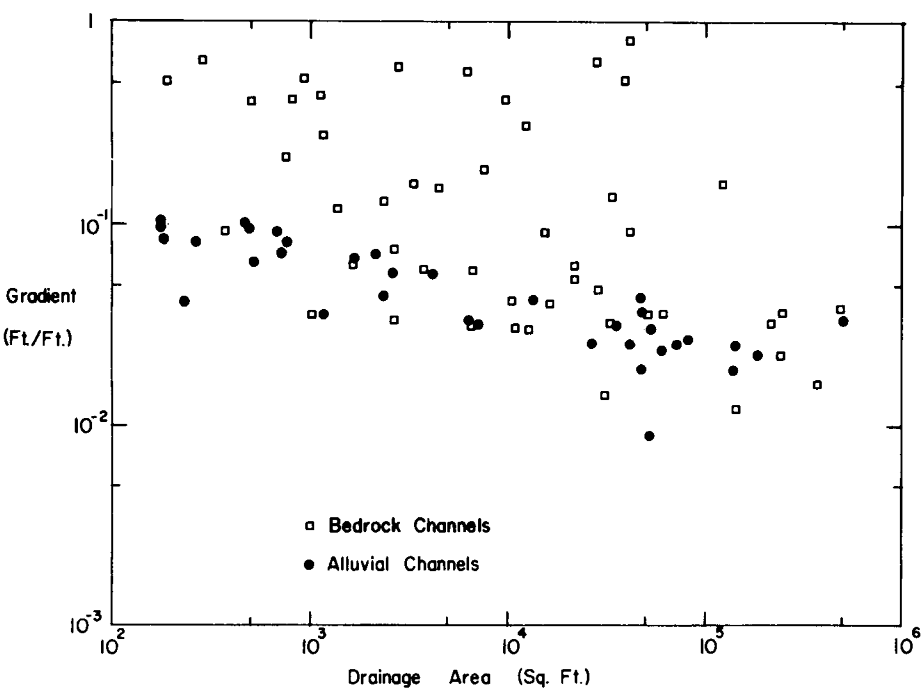


Figure 4. Plot of gradient versus drainage area for alluvial and bedrock channels in the borrow pit.

that were used to prepare a topographic contour map with a 2-ft (0.6-m) contour interval (Fig. 1). From the fall of 1970 until the fall of 1972, theodolite surveys of the major channels were made early each spring and late each fall. Occasional resurveys continued through 1976. From the spring of 1970 through the fall of 1972, photographs were taken of the badlands and channels to give a pictorial record of changes. During the summer of 1972, the Canyon Wash (Fig. 1) was filled in by bulldozer, and a combination of reseeding, fertilizing, and drainage diversions forestalled re-entrenchment. The slow natural revegetation by *Pinus Virginiana* over 15 yr from 1963 had not diminished erosion rates appreciably, when, sometime during 1978, the borrow pit was regraded and reclaimed.

The western United States measurements were made in badlands developed on the Morrison Formation, which is composed of variegated, sandy montmorillonitic shales and interbedded channel sandstones.

On account of the original map scale, measurements were made in feet, and so appear on the figures. However, metric units have been used in the text.

Erosion Rates

The average rate of erosion and the sediment yield within the borrow pit were calculated by measurement of the volume of sediment removed from 1963 until the 1971 map compilation. The 1963 topography was reconstructed from the 1971 map by continuation of divide contours across eroded valleys, correcting for divide erosion by comparison with the 1963 surveyed profiles. The hypsometric integrals of the 1963 and 1971 topography were measured by planimeter and the volumetric change calculated by subtraction. The average rate of downcutting over the 13 hectares was 0.02 m/yr, occurring for the most part as erosion of channels and development of associated side slopes. This is equivalent to a yearly sediment yield of approximately 315 metric tons per hectare, which is close to the values measured from suburban construction sites in northern Virginia and Maryland underlain by soils of similar erodibility (Guy, 1965, 1972; Vice and others, 1968; Wolman, 1967).

CHANNEL DYNAMICS

The main drainage courses were established very shortly after cessation of borrow operations; aerial photography taken 3 yr later (1966) revealed a drainage network differing only in detail from the pattern in 1971. Alluvial and bedrock sections alternated along most of the channels. The bedrock sections occurred either where the postexcavation surface was steep, where resistant beds forced waterfalls or rapids, or where confining sidewalls locally increased the bed-material transport capacity. Conversely, alluvial channels occurred where the original gradient was small and erosional resistance was low.

Several segments of alluvial channel occupied former depressions, such as behind Dikes 1 to 3 and their associated settling ponds, which were emplaced in a futile attempt to impede downstream sediment delivery. By 1966, alluviation and storm runoff had caused Dike 1 to be breached and incision had begun again. Dike 2 had similarly been circumvented before 1966 by Canyon and Badland Washes. However, resistant beds at Falls 1 to 3 limited downcutting. As a result, deposition had filled in the settling pond, creating Alluvial Flat 1, so that alluvial streams carried sediment across the former depression. Subsequently, minor incision occurred in these streams as the falls retreated upstream, but the rates of downcutting were not rapid enough to steepen the channels to the point where the alluvial channel cover would be stripped, creating bedrock channels. In filling behind Dike 3 had created Alluvial Flat 2 by 1969. Alluvial Flats 3 and 4 resulted from fanlike aggradation across nearly level original surfaces.

Resistant beds inhibited stream downcutting, and because of the nearly stable base level, alluvial streams formed upstream, graded to a "control point" at the outcrop of a resistant bed. The beds forming waterfalls or rapids were commonly hardpans or cemented conglomeratic lenses only a few centimetres thick. Examples of alluvial channels above control points included Canyon Wash above Falls 1, Badland Wash above Falls 3, the midsection of Sandy Wash, and Upper Wash at Alluvial Flat 3, where dense vegetation restricted downcutting. Vertical erosion in these alluvial sections was limited by the rate of lowering of the control point. The alluvial channel of Badland Wash upstream from Falls 3 eroded at a rate of about 0.2 m/yr from 1970 to 1972 as a result of retreat of the waterfall (Fig. 5). By contrast, the alluvial section of Sandy Wash had eroded only half as fast during the same period (Fig. 6), and Alluvial Flat 3 on Upper Wash aggraded due to a stable control point and an increased sediment supply (Fig. 7).

At several locations (Entrenched Flats 1–3 and the Upland Flat), the borrow operations produced a nearly level surface perched above a steep bordering scarp. Drainage

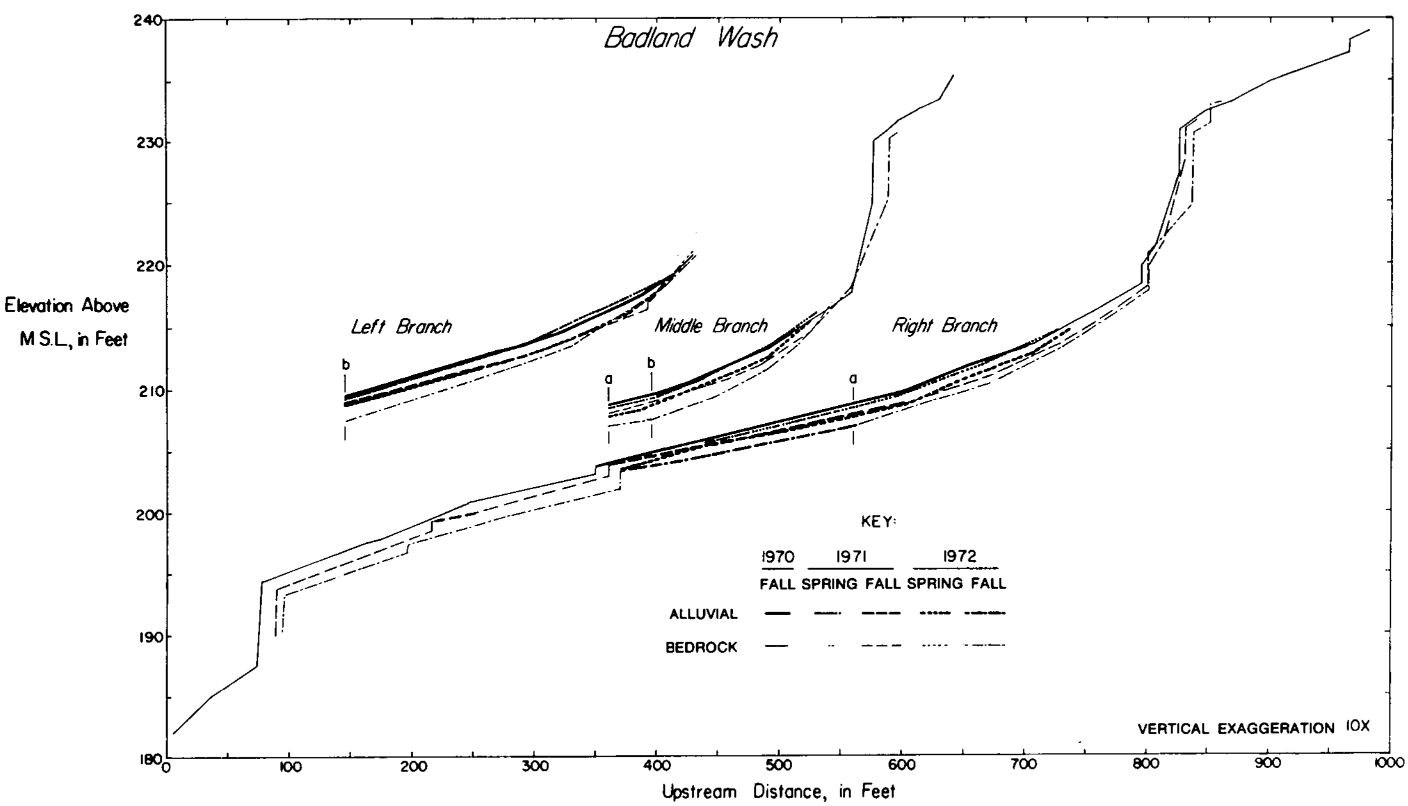


Figure 5. Profiles of Badland Wash and its tributaries from October 1970 through November 1972, showing seasonal and long-term changes.

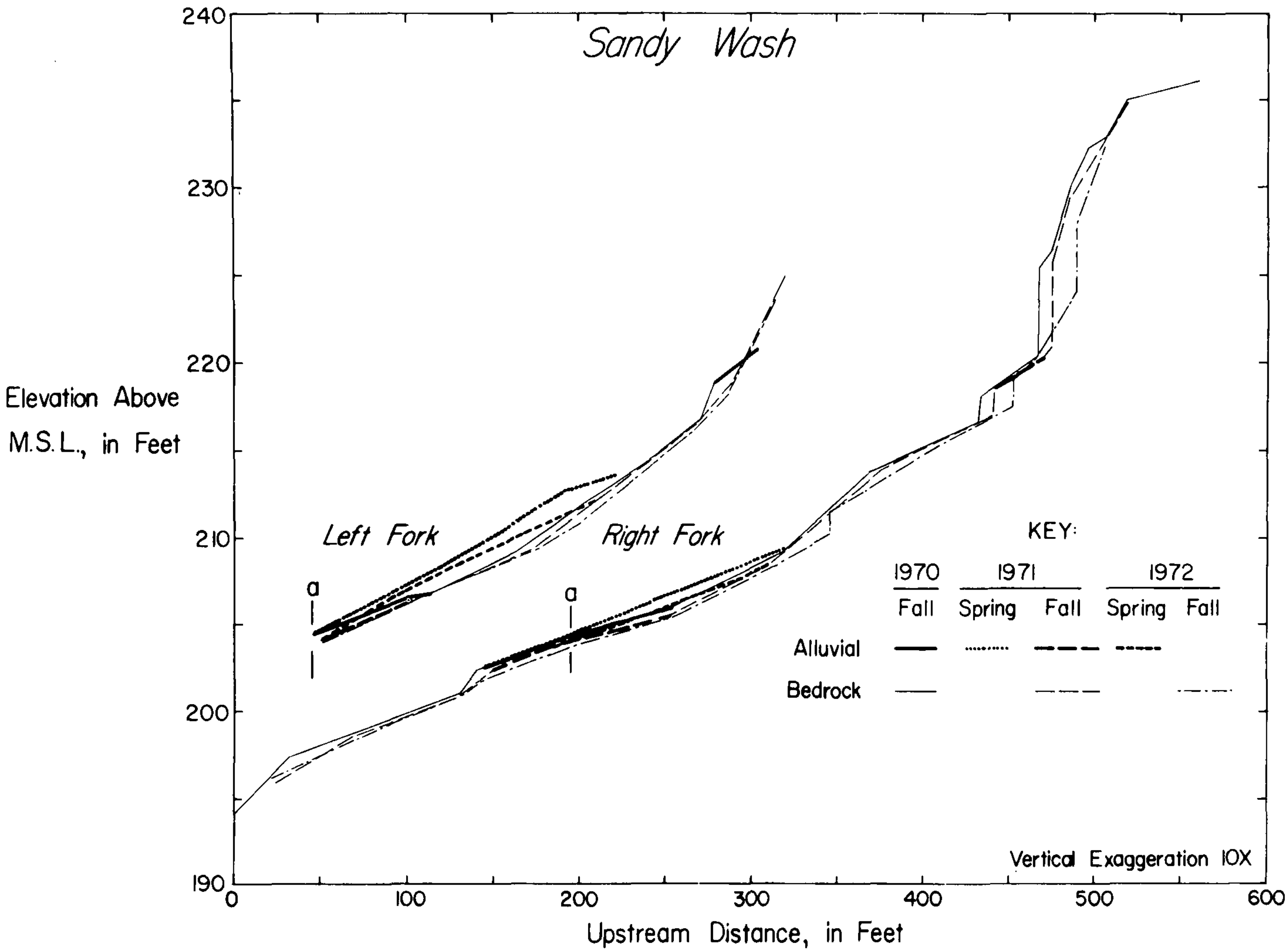


Figure 6. Profiles of Sandy Wash and its tributaries.

from the upland passing over the steep scarp caused the headward migration of a knickpoint and an associated "wave of dissection" (Figs. 1 and 5 through 7).

Two time scales could be distinguished in the evolution of the channel network: seasonal changes were superimposed upon long-term evolution. The seasonal changes were due to summer-winter contrasts in runoff characteristics and sediment load resulting from differences in slope processes, as discussed by Schumm (1956b, 1964) and Schumm and Lusby (1963). During the winter (November-March), mass wasting, primarily by needle ice (Outcalt, 1969, 1971), predominated on most slopes. The extruded prismatic ice needles incorporated thin layers of the soil-like bedrock, which moved downslope on melting of the ice. The process was especially rapid on slopes steeper than the angle of repose, for the dislodged soil was delivered to the base of the slope as a fan-like deposit (Figs. 8A, 2B, 3). Channels adjacent to steep, high slopes became buried, sometimes to a depth of 1 ft or more (Figs. 2B, 3). Winter and spring storms were observed to rapidly entrain much of the loose soil in the channels and fan deposits.

Rain splash and runoff erosion were most effective during the summer season. Figure 9 shows seasonal hourly rainfall data. Transport rate in sand-bed channels increases with about the square of discharge (Howard, 1980). Assuming nearly complete runoff on the shaly badlands, the high-intensity summer rains had about ten times the cumulative transporting potential as winter rains for equivalent channel gradients. Rain splash and runoff also eroded the slopes, producing ephemeral summer rills. However, the intense summer rains smoothed, compacted, and pitted the soil surface (Fig. 8B), with the crust so formed presumably limiting the rate of late summer slope erosion, as has been commonly observed in rain splash erosion studies (Farres, 1978; Thornes, 1980).

Relative rates of sediment delivery to the channels during the winter and summer were uncertain, but the intense winter mass wasting suggests relatively high winter rates. Summer runoff had much greater transporting potential than winter. Alluvial channels responded to this changing balance of load and discharge by aggrading to increase gradients during the late winter and early spring and degrading with gradient reduction during the late summer and fall (Figs. 5–7). Winter-season gradients averaged 6% to 50% higher than summer (Table 1). The depth of alluviation increased upstream because the alluvial sections are

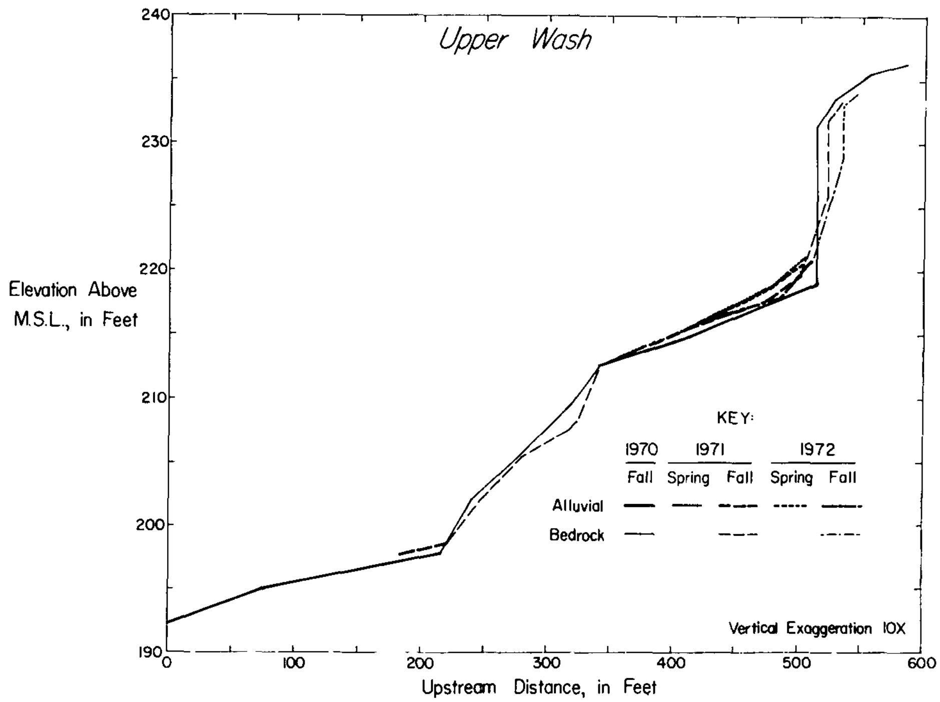


Figure 7. Profile of Upper Wash.

graded to a downstream control point. The downstream control points of most streams continued to erode during the winter, so that the gradient increase was accomplished with less alluviation than occurred along other channels where the base level was stable.

The only seasonal change occurring along most sections of bedrock channels was reduced erosion rates during the winter. However, some sections of channel alternated between alluvial during the winter and bedrock during the summer. The alluvial channel also extended headward (Figs. 2, 5, 6), because the depth of winter alluviation increased upstream. Winter alluvial and summer bedrock channels also alternated where the summer gradient was intermediate between the summer and winter gradients of an alluvial channel of equivalent discharge, bed material supply, grain size, and channel width.

Local changes in alluvial channel gradient occurred in response to fluctuations in the rate of mass wasting. During 1970, Upper Wash, where it crossed the Escarpment above Alluvial Flat 3 (Figs. 1, 7), flowed over an overhanging cliff. However, the cliff collapsed between October 1970 and March 1971, resulting in rapid backwasting and the exposure of the collapse debris to runoff, rain splash, and needle ice erosion. As a result, the gradient across Alluvial Flat 3 increased 50% and the maximum depth of alluviation exceeded 0.6 m. A relatively rapid backwasting continued through 1976, maintaining the steeper gradients.

The entrenched summer channels were commonly narrower than their aggraded counterparts, resulting in a portion of the winter alluvium remaining as terraces (Fig. 2A). Other terraces formed by brief periods of alluviation due to temporary heavy sediment loads following upstream slope failure.

BEDROCK CHANNELS

The lack of alluvial bed on bedrock channels, even during periods of no flow, is due to their steepness; they always carry below-capacity loads of bed sediment, and during waning flows, the alluvium is preferentially deposited on the lower-gradient alluvial beds. Another factor leading to preferential deposition on alluvial sections is the enhancement of grain rebound from the hard bed, so that bedrock channels presumably can transport more bed material than can alluvial channels for the same gradient. This difference in rebound also partially accounts for the very sharp downstream transitions between adjacent sections of alluvial and bedrock beds (Howard, 1980).

The rate of downcutting of bedrock channels increased with bed shear stress, so that the erosion rate was an increasing function of gradient and drainage area. This dependence can be quantified because of the historical record of landform evolution. Most channels became deeply incised, whereas the intervening divides were little eroded. Stream profiles at the close of operations were reconstructed from these interstream divides, which, when compared with the channel survey made in fall 1970, or the topographic map based on aerial photography taken in spring 1971, gave an estimate of channel erosion over a 7- to 7½-yr period (Fig. 10). A least-squares logarithmic regression of the rate of erosion (m/yr) versus the drainage area (m^2) and the gradient (m/m) yields the following estimating equation ($r^2 = 0.85$, $n = 50$):

$$\delta z/\delta t = -.11\ A^{.44}\ \delta z/\delta x^{.68}, \qquad (1)$$

where $\delta z/\delta x$ is the downstream gradient. The 95% confidence interval for the area exponent is 0.38 to 0.51, for the gradient exponent is 0.58 to 0.78, and for the constant is -.06 to -0.21.

The behavior of the bedrock channels is consistent with a simple model that assumes that the erosion is proportional to a power of the dominant bed shear, although hydration of clay minerals, scour by sediment in transport, and minor chemical weathering may also be involved. Neglecting these, the erosion rate can be expressed as a function of the shear stress, τ, exerted by the dominant discharge on the channel perimeter:

$$\delta z/\delta t = -K_1\ \tau^b, \qquad (2)$$

where b is the *shear stress exponent.* K_1 relates erosion rates to shear stress and therefore measures bed erodibility. It is assumed that stream gradients are not more than about 15°. A more realistic model would replace τ by $(\tau - \tau_c)$, where τ_c is the shear stress required to initiate erosion. However, most bed erosion probably occurs at high discharges, where τ may be much greater than τ_c, so that for downstream comparisons the latter may be neglected. The Manning equation is used for flow resistance, and it is assumed that bed roughness does not vary downstream, which seems true for the bedrock channels. The downstream relationship between the cross-sectional area, A_x, and the hydraulic radius, R, is approximated as follows:

$$A_x = K_2\ R^m, \qquad (3)$$

A

B

Figure 8. Seasonal changes of slopes and channels in the Badlands. (A) March 1970, showing angle-of-repose slopes, loose surface texture, and polygonal, frost-widened fractures. (B) October 1970, showing rain-smoothed texture, narrow divides, and tendency to form vertical, joint-controlled upper slopes. The alluvial channels have been entrenched slightly since (A), forming small terraces.

where m is termed the *hydraulic radius exponent*. Finally, the dominant discharge is assumed to be proportional to a power of the drainage area:

$$Q = K_3 A^e. \quad (4)$$

These assumptions imply the following relationship between the rate of bed erosion, the drainage area, and the gradient (Howard, 1970, 1971):

$$\frac{\delta z}{\delta t} = -K_4 A^{[3eb/(3m+2)]} \frac{\delta z}{\delta x}^{[b(6m+1)/2(3m+2)]} \quad (5)$$

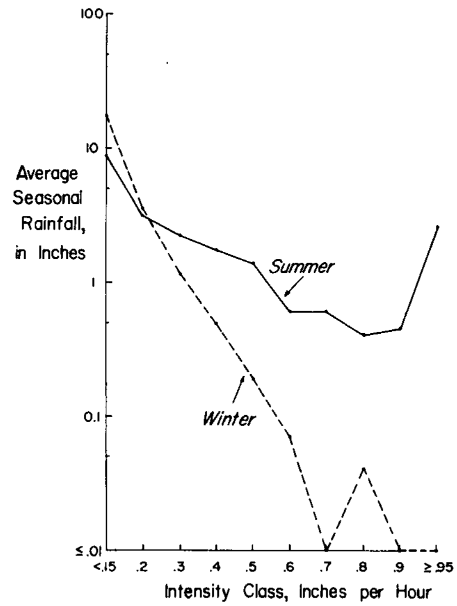


Figure 9. Average winter (October 15–April 14) and summer rainfall, by intensity class, based upon 10 yr of hourly observations at Fredericksburg, Elkton, and Quantico, Virginia.

where K_4 is a dimensional constant incorporating K_1 through K_3, the unit weight of water, and the constant of proportionality in the Manning equation. For a given system of units, a fixed value for b, and equivalent runoff characteristics, K_4, like K_1, expresses bed erodibility, which may vary from place to place.

The borrow-pit observations can be used to infer values for the exponents b, m, and K_4, assuming that the model is correct and that erodibility does not vary systematically with drainage area or channel gradient (Table 2). The inferred value of the shear-stress exponent lies near unity for reasonable values of the exponent e, which is consistent with laboratory measurements of the rate of erosion of cohesive soils (Arulanandan and others, 1973; Akky and Shen, 1973). A direct proportionality was also assumed in a model of overland flow erosion proposed by Horton (1945, p. 324). The predicted values for the downstream hydraulic radius exponent are in the range between the value for constant-width rectangular channels (m = 1) and a semicircular or parabolic channel of constant shape (but not size) m = 2), but they are well below those observed in perennial alluvial channels with self-formed banks ($m \approx 2.5$). The width of the bedrock channels was difficult to estimate because the channels merged

TABLE 1. SEASONAL CHANGES IN GRADIENT OF ALLUVIAL CHANNELS

Channel	Average gradient		Seasonal gradient
	Fall	Spring	ratio
Lower Badland Wash	0.0212	0.0229	1.08
Badland Wash, Left Branch	0.0271	0.0229	1.08
Badland Wash, Middle Branch	0.0361	0.0423	1.17
Badland Wash, Right Branch	0.0308	0.0326	1.06
Sandy Wash, Left Fork	0.0414	0.0604	1.46
Sandy Wash, Right Fork	0.0271	0.0400	1.48
Upper Wash, below Escarpment	0.0337	0.0513	1.52

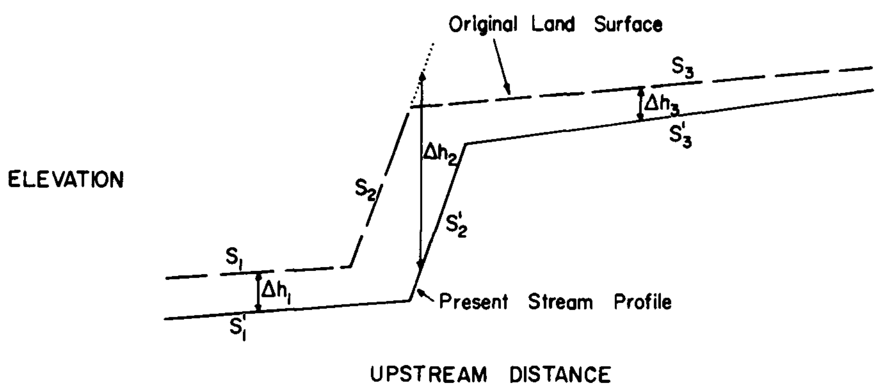


Figure 10. Measurement of erosion rates in bedrock channels. For a specific section of channel, the amount of dissection is measured in the center, and the measured gradient is the average of the original and later values.

indistinctly with the valley walls, so that direct measurement of m was not possible. However, both large and small bedrock channels seemed similar in shape, so that a value of m near 2 is reasonable. Substituting b = 1 and m = 2 into equation 5 and assuming e = 1 lead to the prediction that the drainage area exponent would be 0.38 and the gradient exponent would be 0.81, compared with the observed values of 0.44 and 0.68, respectively (equation 1). Although the predicted values are near the limits of the 95% confidence intervals for the empirical data, they seem close enough to conclude that the model is appropriate, with the discrepancy probably due to a combination of possible systematic downstream changes in bed erodibility, inaccurate choice of parameter values, and correlation between the logarithms of gradient and drainage area in the observed channels (r = −0.47).

TABLE 2. ESTIMATED PARAMETERS FOR BEDROCK CHANNEL EROSION

e	b	m
0.8	0.96	1.07
0.9	0.92	1.22
1.0	0.90	1.38

Column 1. Assumed value for exponent in equation 4.
Column 2. Estimated value of exponent in equation 2.
Column 3. Estimated value of exponent in equation 3.

ALLUVIAL CHANNELS

Most of the alluvial channels gradually downcut during the period of observation (Figs. 5 to 6), indicating that the alluvial bed accumulated by deposition during the waning stages of runoff events, rather than by long-term aggradation. (However, winter aggradation was superimposed upon the epicycles of scour and fill). The flow in the alluvial channels probably carried a capacity load of sand except during peak flows, when the alluvial bed may have been scoured to bedrock. These conclusions, together with the consistent relationship between drainage area and gradient (Fig. 4), suggest that the alluvial channels were adjusted, or graded, to transport the sediment supplied from upstream with the prevailing pattern of runoff (after Mackin, 1948). However, due to the ephemeral flow and seasonal changes in sediment supply and runoff, the gradients continuously readjusted, so that there may never have been exact equilibrium. In the following sections, alluvial channel data are examined to determine their patterns of behavior within the borrow pit, including downstream hydraulic geometry and seasonal gradient variations.

Downstream Hydraulic Geometry

The relationships between sediment transport capacity, stream gradient, grain size, and flow properties at equilibrium have been investigated in many flume studies and summarized in numerous formulas. Were such data available for the badland channels, the closeness to equilibrium could be directly determined. Because of the irregularity of flows and the difficulty of field measurement, however, the sediment concentration and flow properties were not measured. Rather, indirect methods were used to compare the downstream hydraulic geometry of the badland alluvial channels with predictions of sediment transport relationships. Several quantities were measured on a sample of large-to-small alluvial channels:

1. *Local gradient* was measured either from the surveyed profiles (Figs. 5–7) or , in the case of small, steep channels, by Brunton compass.

2. *Equivalent length.* The drainage area and channel width were measured at each site. Their ratio is termed the "equivalent length" (Howard, 1980). Width and drainage area were combined into a single measure for two reasons. First, channel width is not an explicit variable in sediment transport formulas; discharges of sediment and water are expressed per unit width. Equivalent length is used as a surrogate measure of these specific discharges, as is discussed below. Second, runoff from some slopes travels as unconcentrated sheet flow across alluvium-floored pediments or small alluvial fans. A unique channel width or drainage area does not exist on such alluvial surfaces. However, the equivalent length can be measured by finding the area contributing flow across an arbitrary line drawn on the alluvial surface perpendicular to the gradient (Fig. 11). Equivalent lengths

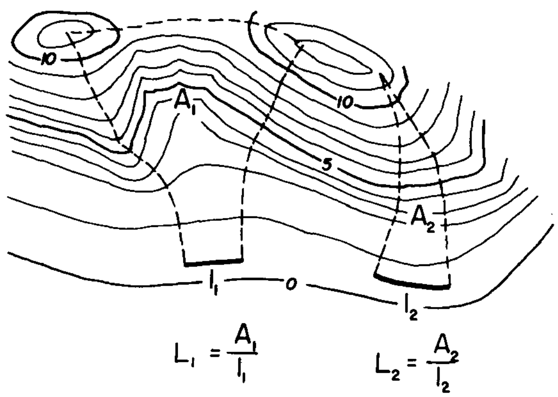


Figure 11. Measurement of equivalent length, L, where runoff occurs by unconcentrated sheet flow over an alluvial surface at the base of a badland slope (idealized contours). The heavy lines are arbitrary widths, and the dashed lines show the contributing drainage area. Case 1 indicates converging flow; case 2, diverging.

for both alluvial channels and alluvial surfaces are included in the measured data.

3. *Grain size* of the alluvial bed channels varied little within the borrow pit, with a median near 0.3 mm. Occasional pebbles derived from the Upland Gravels were scattered through the alluvium. The alluvium was not systematically sampled because of the small grain size range.

4. *Interfluvial slope steepness.* Sediment yield should be greater where slopes are steeper. Slopes were classified according to steepness and position within the basin: (1) alluvial surface and alluvial channel, (2) upland flats sloping less than 5°, (3) moderate slopes (5° to 15°), and (4) steep slopes. Alluvial surfaces and channels generally eroded very slowly, or were stable, so that they contributed very little sediment directly. Thus, sediment yields should have increased in the order listed. Runoff per unit area is probably less sensitive to slope steepness, particularly during the intense or long-duration storms that presumably transported the majority of the sediment.

For a basin containing only one slope class, the gradient is assumed to be related to the equivalent length:

$$S_i = C_i L^z \qquad (6)$$

Where S_i is the channel gradient associated with slope class i with a characteristic constant C_i. For a basin containing portions of more than one slope class, the gradient is assumed to be a porportional logarithmic sum of the gradients characteristic of each of the fore slope classes present:

$$\log S = \sum_{i=1}^{4} P_i \log S_i, \qquad (7)$$

or

$$\log S = \sum_{i=1}^{4} P_i \log C_i + z \log L, \qquad (8)$$

where P_i is the proportion of the drainage area in slope class i (including cases with individual $P_i = 0$). This model was chosen because the coefficients to be estimated entered linearly.

Measurements of L, P_i and S were taken at 36 locations in the borrow pit during October 1971. Values of L and P_i were estimated using low-level aerial photography. Least-squares estimates of the coefficients z and C_i were determined by regression, with L measured in metres (Fig. 12 and Table 3, part A). The coefficients explain 89% of the variance of the logarithm of the gradient. The coefficients increase in order, as expected, although there is considerable overlap in their confidence intervals. The unexplained variance probably resides in differences in grain size and in erodibility among the bedrock slopes.

A similar relationship was found for alluvial surfaces and alluvial channels on natural badlands developed on the Morrison Formation in Utah. Only two slope classes, (1) alluvial surface and channel and (2) badland slopes, were defined for the western data, but the grain size was additionally measured at each site, so that the model becomes:

$$\log S = P_a \log C_1 + (1-P_a) \log C_2 + u \log d + z \log L, \qquad (9)$$

where P_a is the proportion of the drainage area underlain by alluvial surface and d is the median grain size in millimetres. Regression coefficients for a sample of 31 locations explained 91% of the variance (Fig. 12 and Table 4).

Alluvial-channel gradients in the borrow pit obey the following relationship to drainage area (m^2):

$$S = .15\ A^{-0.19}. \qquad (10)$$

The width of alluvial channels was related to about the 1/4th power of drainage area in both the borrow pit and the Morrison badlands, but the western channels are about twice as wide for an equivalent drainage area (Fig. 13).

Howard (1980) has shown that the observed gradient relationships are consistent with predictions of downstream changes in sand-bed alluvial-channel gradients based upon total-load sediment-transport formulas under conditions of high sediment loads.

Erosion Rates in Alluvial Channels

The survey data indicate that alluvial channels may erode their beds through time if their downstream control point lowers. However, there is an upper limit to their capacity for erosion, beyond which their gradient steepens, waning flows no longer redeposit alluvium, and they become bedrock channels. Combination of equations 1 and 10 suggests that, for channel width, runoff, and bed erodibility characteristic of the borrow pit, this maximum erosional rate increases with drainage area:

$$\delta y/\delta t_{max} = -0.0089\ A^{0.31}. \qquad (11)$$

Alluvial channels that lie near the upper limit of this criterion require steeper gradients than do their stable counterparts. The requirement for steeper gradients in more rapidly eroding channels follows directly from equation 1. In rapidly eroding channels, the alluvial cover is stripped during high discharges, whereas the flow carries a below-capacity bed material load, so that the eroding alluvial channel transports the same sediment load as a noneroding alluvial channel of lesser gradient.

As a measure of the gradient increase with erosion, the erosion rate was added into the regression of equation 8 in a semiquantitative manner. Nominal values were assigned to measured erosion rates as follows: 4.0 for rates greater than .15 m/yr; 3.0 for rates between .09 and .15; 2.0 for rates between .03 and .09; 1.0 for rates between .005 and .03; and 0.0 for rates less than .005 (including aggradation). A nominal classification was used because only about one-third of the data could be directly measured (*for the most part, the larger and more* rapidly eroding channels); values were assigned to the other channels by estimation based upon measurements in nearby streams

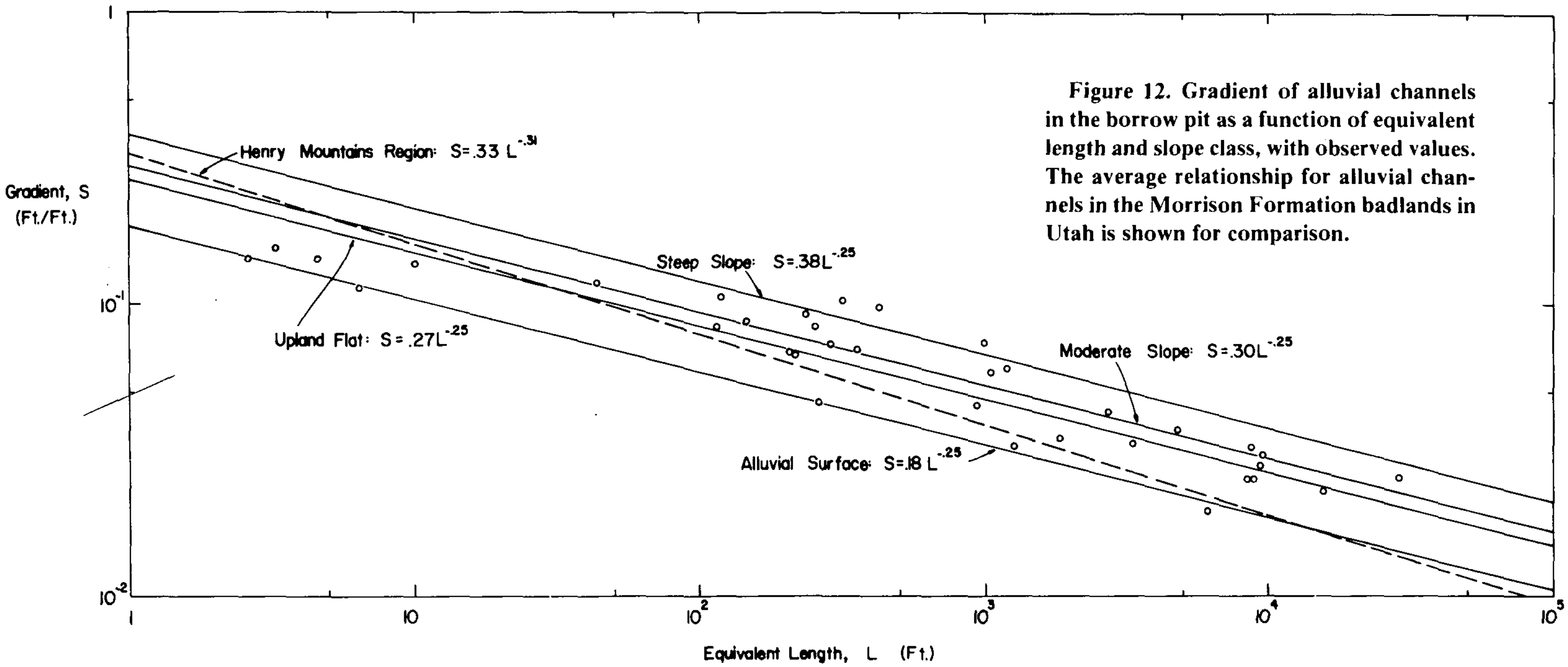


Figure 12. Gradient of alluvial channels in the borrow pit as a function of equivalent length and slope class, with observed values. The average relationship for alluvial channels in the Morrison Formation badlands in Utah is shown for comparison.

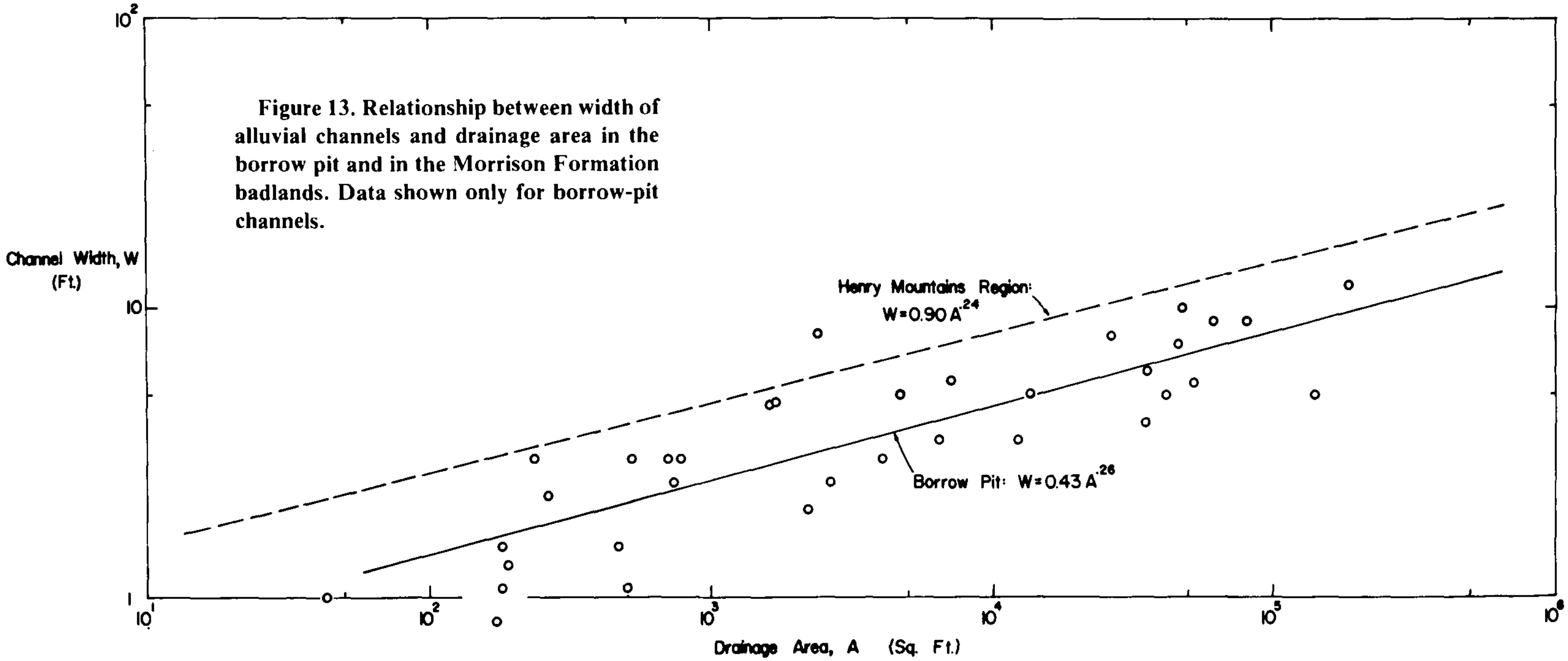


Figure 13. Relationship between width of alluvial channels and drainage area in the borrow pit and in the Morrison Formation badlands. Data shown only for borrow-pit channels.

TABLE 3. REGRESSION COEFFICIENTS FOR ALLUVIAL CHANNELS IN THE BORROW PIT

(A)

Estimated value	95% confidence interval	Category
C_1 = 0.14	0.11–0.18	Alluvial surface
C_2 = 0.19	0.10–0.36	Upland flat
C_3 = 0.21	0.14–0.31	Moderate slope
C_4 = 0.26	0.15–0.45	Steep slope
z = −0.24	-0.19–−0.30	Equivalent length exponent

(B)

Estimated value	Category
C_1 = 0.14	Alluvial surface
C_2 = 0.19	Upland flat
C_3 = 0.23	Moderate slope
C_4 = 0.23	Steep slope
z = −0.25	Equivalent length exponent
j = 0.04	Erosion rate exponent

TABLE 4. REGRESSION COEFFICIENTS FOR ALLUVIAL CHANNELS IN THE MORRISON SHALE BADLANDS

Estimated value	95% confidence interval	Category
C_1 = 0.15	0.10–0.23	Alluvial surface
C_2 = 0.46	0.30–0.69	Slopes
u = 0.17	0.03–0.36	Grain-size exponent
z = −0.31	−0.26–−0.35	Equivalent-length exponent

and the total erosion that had been experienced within the drainage basin based upon comparison of the 1971 map with previous data. The nominal data were then normalized to a zero mean and unit standard deviation over the sample. Equation 8 was modified to include an additional term, $j \times E$, where j is the estimated regression coefficient and E is the normalized erosion rate. Table 3, part B lists the estimated coefficients. As indicated, the gradient increases with erosion rate, and there is a significant increase in variance explained. The relative magnitudes of the coefficents of the slope classes are somewhat altered by the inclusion of erosion rates, but because of the positive correlation between erosion rates and both relief and size of the drainage basin, and because of the nominal classification, the coefficients in part B of Table 3 should not be used in quantitative predictions.

Time Scales, Equilibrium, and Grade in the Alluvial Channels

Gradients within a channel network can be considered to be in equilibrium, that is, to be *graded*, if the gradients are continually adjusted to the prevailing hydraulic regime (supply of sediment and runoff from slopes). The presence or absence of equilibrium is determined jointly by the temporal history of the hydraulic regime and by the time scale of response of the channel gradients. The time scale for response of a channel network to regime changes is approximately (Howard, 1982):

$$T = D^2 S/(4 q_s), \qquad (12)$$

where D is the mainstream length, and q_s and S are measured at the downstream end of the network. Regime fluctuations that occur with very high frequencies compared to this response time have negligible effect upon the gradients and are ignored in the determination of equilibrium. On the other hand, the channel can completely adjust to variations that occur with very low frequencies, as well as to very slow trends in the regime. Disequilibrium occurs if there are large, recent step changes or rapid trends in regime or if there are large cyclical components to the regime with frequencies of the same order of magnitude as the time scale of response (Howard, 1982).

Determination of the time scale for the borrow-pit channels can be done only indirectly, because no measurements of q_s exist. However, the average erosion rate in the borrow pit (0.02 m/yr) can be used as an estimator. As portions of the borrow pit were essentially non-eroding, the erosion rate in the badland sections was at least twice the average rate. On the other hand, only one-half or less of the eroded sediment was in the sand size range, so that the 0.02 figure will be used as the areal supply rate. The largest badland channels had a mainstream length of about 100 m and a drainage area of about 6,000 m^2. Using equation 10 and Figure 12, the corresponding gradient and channel width are .03 and 2.3 m, respectively, giving an estimated time scale (equation 12) of about 1.4 yr. This time scale includes both periods with runoff and the long intervals with none. The time scale measured in terms of runoff duration was much shorter, about 200 hr, because appreciable runoff occurred only about 150 hr/yr.

The hydraulic regime in the borrow pit had two major quasicyclical time scales of change: storm-period fluctuations and seasonal variations. Simulations by Howard (1982) suggested that a channel network can continuously readjust to cyclical regime changes, obeying the following inequality:

$$2\pi \, T \, W < .5, \qquad (13)$$

where W is the frequency in cycles/yr. Howard (1982) also showed that high-frequency components obeying the following inequality are "filtered out" in channel response; that is, they are unimportant:

$$2\pi \, T \, W > 20. \qquad (14)$$

The response of channel networks to frequency components lying between these two inequalities has finite damping and delay, that is, disequilibrium behavior.

The storm-period fluctuations, with frequencies of several tens per yr, produced little change in channel gradients. However, the seasonal cycle fell within the disequilibrium range, so that the seasonal gradient changes were an incomplete response to the seasonal variations in the hydraulic regime. In addition, the gradients in rapidly eroding alluvial channels were appreciably biased by the erosion, as discussed above. Therefore, the alluvial channels in the borrow pit were not graded in a strict sense. However, the yearly average of channel gradients should have been adjusted to the average hydraulic regime (excepting the erosional bias), which can be considered to be a restricted type of grade.

The western badland channels have a much longer time scale due to the sparse rainfall (about 0.1 m/yr compared to the 1.3 m/yr in Virginia). In view of the fact that the sediment yield should be approximately proportional to the rainfall, the western badland channels show negligible gradient response to seasonal regime changes, and they would be graded if the climate remained statistically constant over a period of tens of years.

CONCLUSIONS

The miniature channel system studied in Virginia provided a limited analogue to much slower responses that occur in large channel systems. The comparison is most direct for natural badlands in arid environments, where slope and channel processes are nearly identical. The major difference is the large seasonal cycle, which probably has no analogue for the equivalent time scale (decades or centuries) in the natural badlands.

The pattern of alluvial channel regrading in response to both variations in hydraulic regime and change in base level should be comparable to those occurring in larger channels, with the exception that the shallow, braided badland channels did not change appreciably in sinuosity during aggradation or erosion. By comparison, most large alluvial channels meander, and changes in sinuosity constitute an additional element of channel response that affects the amount of aggradation or erosion following a change in regime (Schumm, 1968, 1971).

The transient response of natural alluvial channels to changes in hydraulic regime is poorly characterized (Howard, 1982). Owing to their marked and rapid seasonal regrading, alluvial channels in eastern badlands would be ideal sites for more detailed studies.

The relationship between erosion rate, gradient, and drainage area in bedrock channels (equation 1) should be applicable to predicting channel erosion rates in soft sediments and soils with appropriate scaling for erodibility and runoff characteristics. The empirical observations of erosion rates clearly validate the model of drainage development proposed by Horton (1945), which suggests that a dendritic network develops because of the competitive advantage of large channels.

The borrow-pit observations also have implications for methods of reducing erosion and sediment yields from erosion-susceptible areas such as construction sites. Equation 1 emphasizes the necessity of dispersing runoff and of minimization of slope gradients. Also, the pattern of evolution of borrow-pit topography indicates that most of the sediment yield derived from channel erosion and development of adjacent steep side slopes as opposed to general sheetwash, underscoring the necessity of protecting erodible soils from development of rills and gullies.

ACKNOWLEDGMENTS

The Virginia Highway Research Council provided funds for aerial photography and the compilation of the topographic map; special thanks to Mike Ozol. John Fisher, Stanley Schumm, M. G. Wolman, Tom Dunne, and Mike Church offered helpful comments on earlier drafts. Wayne Newell provided stratigraphic correlations for the borrow pit.

REFERENCES CITED

Akky, M. R., and Shen, C. K., 1973, Erodibility of a cement-stabilized sandy soil, *in* Soil erosion: Causes and mechanisms: Highway Research Board Special Report 135, p. 30–41.

Arulanandan, K., Sargunam, A., Loganathan, P., and Krone, R. B., 1973, Application of chemical and electrical parameters to prediction of erodibility, *in* Soil erosion: Causes and mechanisms: Highway Research Board Special Report 135, p. 42–51.

Engelen, G. B., 1973, Runoff processes and slope development in Badlands National Monument, South Dakota: Journal of Hydrology, v. 18, p. 55–79.

Farres, P., 1978, The role of time and aggregate size in the crusting process: Earth Surface Processes, v. 3, p. 243–254.

Guy, H. P., 1965, Residential construction and sedimentation at Kensington, Maryland, *in* Federal inter-agency sedimentation conference proceedings, 1963: U.S. Department of Agriculture Miscellaneous Publication 970, p. 30–37.

—— 1972, Urban sedimentation—In perspective: Proceedings of the American Society of Civil Engineers, Journal of the Hydraulics Division, v. 98, HY12, p. 2009–2016.

Horton, R. B., 1945, Erosional development of streams and their drainage basins: Geological Society of American Bulletin, v. 56, p. 275–370.

Howard, A. D., 1970, Study of process and history in desert landforms near the Henry Mountains, Utah [Ph.D. thesis]: Baltimore, Maryland, Johns Hopkins University, 198 p.

—— 1971, Problems in interpretation of simulation models of geologic processes, *in* Morisawa, M., ed., Quantitative geomorphology: Some aspects and applications: Binghamton, New York, Publications in Geomorphology, p. 61–82.

—— 1980, Thresholds in river regimes, *in* Coates, D. R., and Vitek, J. D., eds., Thresholds in geomorphology: London, George Allen and Unwin, p. 227–258.

—— 1982, Equilibrium and time scales in geomorphology: Application to sand-bed alluvial streams: Earth Surface Processes and Landforms, v. 7, p. 303–325.

Mackin, J. H., 1948, Concept of the graded river: Geological Society of America Bulletin, v. 59, p. 463–512.

Outcalt, S. I., 1969, Weather and diurnal frozen soil structure at Charlottesville, Virginia: Water Resources Research, v. 5, p. 1377–1382.

—— 1971, An algorithm for needle-ice growth: Water Resources Research, v. 7, p. 394–400.

Schumm, S. A., 1956a, The role of creep and rainwash on the retreat of badland slopes: America Journal of Science, v. 254, p. 693–706.

—— 1956b, Evolution of drainage systems and slopes in badlands at Perth Amboy, New Jersey: Geological Society of America Bulletin, v. 67, p. 597–646.

—— 1962, Erosion of miniature pediments in Badlands National Monument, South Dakota: Geological Society of America Bulletin, v. 73, p. 719–724.

—— 1964, Seasonal variations of erosion rates and processes on hillslopes in western Colorado: Annals of Geomorphology, Supplement no. 5, p. 215–238.

—— 1967, Rates of surficial rock creep on hillslopes in western Colorado: Science, v. 155, p. 560–561.

—— 1968, River adjustments to altered hydraulic regime—Murrumbidgee River and paleochannels, Australia: U.S. Geological Survey Professional Paper 598, 65 p.

—— 1971, Fluvial geomorphology: Channel adjustment and river metamorphosis, *in* Shen, H. W., ed., River mechanics, Volume 1: Fort Collins, Colorado, Water Resources Publications, chap. 5, 22 p.

Schumm, S. A., and Lusby, 1963, Seasonal variation of infiltration capacity and runoff on hillslopes in western Colorado: Journal of Geophysical Research, v. 68, p. 3655–3666.

Thornes, J. B., 1980, Erosional processes of running water and their spatial and temporal controls: A theoretical viewpoint, *in* Kirkby, M. J., and Morgan, R.P.C., eds., Soil erosion: New York, John Wiley & Sons, chap. 5, p. 129–182.

Vice, R. B., Guy, H. P., and Ferguson, G. E., 1968, Sediment movement in an area of suburban highway construction—Scott Run Basin, Virginia—1961–64: U.S. Geological Survey Water-Supply Paper 1591-E, 41 p.

Wolman, M. G., 1967, A cycle of sedimentation and erosion in urban river channels: Geografiska Annaler, v. 49A, p. 385–395.

Manuscript Received by the Society May 15, 1981
Revised Manuscript Received May 24, 1982
Manuscript Accepted June 8, 1982