Geomorphic thresholds: the concept and its applications[1]

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ABSTRACT. Geomorphic thresholds were defined initially as the condition at which there is a significant landform change without a change of external controls such as base level, climate and land use. Landforms evolve to a condition of incipient instability following which change or failure occurs. Subsequently, through usage, the definition has been broadened to include abrupt landform change as a result of progressive change of external controls. Therefore, it is now appropriate to recognize both intrinsic and extrinsic geomorphic thresholds. The threshold concept has practical significance. If the threshold conditions can be recognized, not only will different explanations for some landforms emerge but also the ability to identify incipiently unstable landforms and to predict their change will be of value to land managers and engineers. For example, the development of gullies and fan-head trenches can be explained by the depositional steepening of valley floors and fan-heads to threshold slope. As a consequence, as yet ungullied but potentially unstable areas can be recognized. In addition, channel pattern variations and the conversion of meandering channels to braided ones, and of braided channels to single-thalweg sinuous ones can occur naturally at pattern thresholds. Such changes can also be accomplished artificially, when it is recognized that a channel is near a pattern threshold. Sediment yield variations will be related to these periods of instability. Recognition of this will aid in the explanation of some hydrologic, sedimentologic, and stratigraphic anomalies.

INTRODUCTION

IN the past geomorphologists concentrated their efforts on the development of an understanding of the erosional and depositional evolution of landforms through geologic time. More recently we have departed from this approach, when it was realized that it is the details of landscape evolution that require elaboration and explanation if traditional geomorphic problems are to be solved and if geomorphic research is to be of value to those who are managing and attempting to control various components of the landscape (rivers, slopes, flood-plains). Therefore, an understanding of the functioning of geomorphic systems over short spans of time is mandatory. When this is attempted it is soon apparent that the extrapolation of measured average rates of erosion and deposition to longer periods of time is misleading, in the sense that they do not reveal the natural complexity of landform development or the variability of existing landforms.

Owing to the complexity of Quaternary climatic and tectonic histories, topographic and stratigraphic variability can be conveniently explained as a result of climatic and tectonic events. In this way the compulsion to fit geomorphic and stratigraphic details into a Quaternary chronology is satisfied as well as the basic scientific need to identify cause and effect. However, as the details of Holocene stratigraphic and terrace chronologies are studied, a bewildering array of changes are required to explain the behaviour of a drainage system. In fact, it is now accepted that some major erosional adjustments can be induced by rather insignificant changes in the magnitude and frequency of storm events (Leopold, 1951).

The numerous deviations from an orderly progression of the erosion cycle have led many to discount the erosion cycle concept completely. Current practice is to view the evolutionary

development of the landscape within the conceptual framework of the erosion cycle, but to consider much of the modern landscape to be in dynamic equilibrium. There are obvious shortcomings in both concepts. For example, although the cycle involves continuous slow change, evidence shows that periods of relatively rapid system adjustment result from external causes. This, of course, is equally true of geomorphic systems in dynamic equilibrium. That is to say, for an abrupt change to occur in either the cycle or a system in dynamic equilibrium there must be an application of an external stimulus. Hence, landscape changes and changes in rates of depositional or erosional processes are explained by the influence of man, by climatic change or fluctuations, by tectonics, or by isostatic adjustments.

One cannot doubt that major landscape changes and shifting patterns of erosion and deposition have been due to climatic change and tectonic influences and that man's influence is substantial. Nevertheless, it is the details of the landscape, the last inset fill, the low Holocene terrace, modern periods of arroyo cutting and gulleying, alluvial fan-head trenching, channel aggradation and slope failure that for both scientific and practical reasons of land management require explanation and prediction. These geomorphic details are of real significance, but often they cannot be explained by traditional approaches.

Another aspect of the problem is that within a given region all landforms did not respond to the last external influence in the same way, and some seem hardly to have responded at all. This is a major geomorphic puzzle that is commonly ignored. If land systems are in dynamic equilibrium, components of the system should respond in a similar way to an external influence. Hence, the effects of hydrologic events of large magnitude should not be as variable as they appear to be.

In summary, the cyclic and dynamic equilibrium concepts are not of value in the location of incipiently unstable landforms because within these conceptual frameworks, system change is always due to external forces. There is now both experimental and field evidence to indicate that this need not be true.

The answer lies in the recognition of additional geomorphic concepts that are necessary for an understanding of landform development and evolution. The most important of these is the concept of geomorphic thresholds. The additional concepts of complex response and episodic erosion are a logical consequence of the threshold conception.

These concepts are certainly not new, and one need not search long to find a geomorphic paper or book mentioning thresholds or the complexities of geomorphic systems (for example, Chorley and Kennedy, 1971; Tricart, 1965; Pitty, 1971). However, they have not been directed to the solution of specific problems nor has their significance been fully appreciated. The assumption that all major landform changes or changes in the rates and mechanics of geomorphic processes can be explained by climatic or tectonic changes has prevented the geologist from considering that landform instability may be inherent.[2]

THRESHOLDS

Thresholds have been recognized in many fields and their importance in geography has been discussed in detail by Brunet (1968). Perhaps the best known to geologists are the threshold velocities that are required to set in motion sediment particles of a given size. With a continuous increase in velocity, threshold velocities are encountered at which movement begins, and with a progressive decrease in velocity, threshold velocities are encountered at which movement ceases. These are Brunet's (1968, pp. 14 and 15) 'thresholds of manifestation' and 'thresholds of extinction', and they are the most common types of thresholds encountered. However, when a third variable is involved, Brunet (1969, p. 19) identified 'thresholds of reversal.' An example of this type of threshold is Hjulström's (1935) curve showing the velocity required for movement

of sediment of a given size. The curve shows that velocity decreases with particle size until cohesive forces become significant, and then the critical velocity increases with decreasing grain size. Another example of this type of relationship is the Langbein-Schumm (1958) curve which shows sediment yield as directly related to annual precipitation and run-off until vegetation cover increases sufficiently to retard erosion. At this point there is a decrease in sediment yield with increased run-off and precipitation. Perhaps thresholds is not a good word to describe the critical zones within which these changes occur, but it is a simple and easily understood term.

The best known thresholds in hydraulics are described by the Froude and the Reynolds numbers, which define the conditions at which flow becomes supercritical and turbulent. Particularly dramatic are the changes in bed-form characteristics at threshold values of stream power.

In the examples cited, an external variable changes progressively, thereby triggering abrupt changes or failure within the affected system. Response of a system to an external influence occurs at what will be referred to as *extrinsic thresholds*. That is, the threshold exists within the system, but it will not be crossed and change will not occur without the influence of an external variable.

Thresholds of the other type are *intrinsic thresholds*, and changes occur without a change in an external variable. An example is long-term progressive weathering that reduces the strength of slope materials until eventually there is slope adjustment (Carson, 1971) and mass movement (Kirkby, 1973). Following failure, a long period of preparation ensues before failure can occur again (Tricart, 1965, p. 99).

Glacial surges, that are not the result of climatic fluctuations or tectonics (Meier and Post, 1969) probably reflect periodic storage and release of ice, as an intrinsic threshold of glacial stability is exceeded. In semi-arid regions sediment storage progressively increases the slope of the valley floor until failure occurs by gullying. This is a special type of intrinsic threshold, the geomorphic threshold (Schumm, 1973). It is a result of landform (slope) change through time to a condition of incipient instability and then failure. Another example of a geomorphic threshold is provided by Koons (1955) who described morphologic changes resulting in the collapse of sandstone-capped shale cliffs in the Mesa Verde region in south-western Colorado. Beneath a vertical cliff of Mesa Verde sandstone is a 32 to 38 degree slope of weak Mancos Shale. Through time, the basal shale slope is eroded and reduced in height, thereby producing a vertical shale cliff beneath the sandstone cap. At some critical height the cliff collapses and the cycle begins again. The episodic retreat of this escarpment is the result of the change in cliff morphology under constant climatic, base level and tectonic conditions.

Intrinsic thresholds are probably common in geologic systems, but only geomorphic examples will be considered here. A *geomorphic threshold* is one that is inherent in the manner of landform change; it is a threshold that is developed within the geomorphic system by changes in the morphology of the landform itself through time. It is the change in the landform itself that is most important, because until it has evolved to a critical situation, adjustment or failure will not occur.

The concept of geomorphic thresholds, which involves landform change without a change in external controls challenges the well-established basis of geomorphology, that landform change is the result of some climatic, base-level or land-use change. Therefore, the significance of the geomorphic threshold concept for the geomorphologist is that it makes him aware that abrupt erosional and depositional changes can be inherent in the normal development of a landscape and that a change in an external variable is not always required for a geomorphic threshold to be exceeded and for a significant geomorphic event to ensue.

It is important to stress again that the geomorphic threshold as defined above is an intrinsic threshold. If as a result of a climate change, discharge and flow velocities in a stream channel are increased, the resulting bank erosion and meander cut-offs may convert a meandering stream to a braided channel. In this case the cause of the pattern conversion is extrinsic, but if, as a result of increasing channel sinuosity under unchanging average discharge, cut-offs occur which convert a meandering reach to a straight reach the control is intrinsic. In the first case the change may be regional in nature and in the second it may be local. However, explanation of locally anomalous, erosional or depositional features is what is required for rational landform management.

During recent discussions with students and colleagues, it became apparent that the term geomorphic threshold is being used in a much broader sense than originally intended, and a revision of the definition is required.

The concept of intrinsic geomorphic controls was stressed because it initially provided a new approach to the understanding of the details of the landscape that could be used for predictive purposes. Nevertheless, there can also be extrinsic geomorphic thresholds. For example, in common usage 'thresholds' can be the result of either cause or effect. That is, we speak of hydraulic, velocity, shear, and stream power thresholds above which sediment moves or banks fail, but we can also speak of bank, channel and slope stability thresholds, when the forces causing the failures are not clearly identified and understood. Therefore, geomorphic thresholds can be of two types, and they can be redefined in the following way. A geomorphic threshold is a threshold of landform stability that is exceeded either by intrinsic change of the landform itself, or by a progressive change of an external variable. Although the original definition is broadened, the concept of abrupt landform change remains.

Evidence for geomorphic thresholds

Recent field and experimental work supports the concept of geomorphic thresholds. The best examples result from investigations of gully distribution and stream pattern variability.

Gullies   Field studies in valleys of Wyoming, Colorado, New Mexico, and Arizona revealed that discontinuous gullies, short but troublesome gullied reaches of valley floors, can be related to the local oversteepening of the valley-floor surface (Schumm and Hadley, 1959). For example, the beginning of gully erosion in these valleys tends to be localized on steeper convex reaches of the valley floor. Carrying this one step farther, with the concept of geomorphic thresholds in mind, for a given region of uniform geology, land use and climate, a critical valley slope will exist above which the steepest reach of the valley floor is unstable. In order to test this hypothesis, measurements of valley-floor gradient were made in the Piceance Creek Basin of western Colorado. The area is underlain by oil shale, and the potential environmental problems that will be associated with the development of this resource are considerable.

Within this area, valleys were selected in which discontinuous gullies were present. The drainage area above each gully was measured on maps, and critical valley slopes at the point of gully development were surveyed in the field. No hydrologic records exist, so drainage-basin area was selected as a substitute for run-off and flood discharge. When this critical valley slope is plotted against drainage area, the relationship is inverse (Fig. 1), with gentler slopes being characteristic of large drainage areas. As a basis for comparison, similar measurements were made in ungullied valleys, and these data are also plotted on Figure 1. The lower range of critical slopes of the unstable valleys coincides with the higher range of slopes of the stable valleys. In other words, for a given drainage area it is possible to define a critical valley slope above which the valley floor is unstable.

Note that the relationship does not pertain to drainage basins smaller than about four
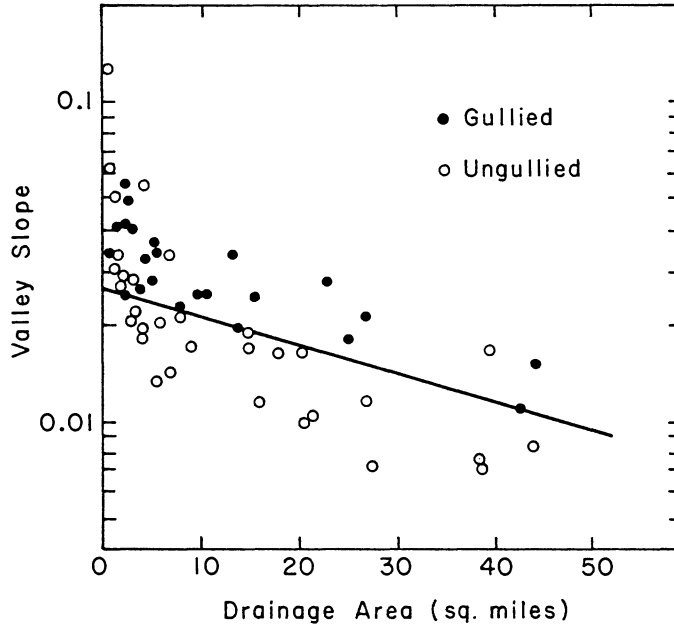
FIGURE 1. Relation between valley slope and drainage area, Piceance Creek Basin, Colorado. Line defines threshold slope that separates gullied from ungullied valley floor. (FROM PATTON AND SCHUMM, 1975)

square miles. In these small basins variations in vegetative cover, which are perhaps related to the aspect of the drainage basin or to variations in the properties of the alluvium, prevent recognition of a critical threshold slope. Above four square miles there are only two cases of stable valley floors that plot above the threshold line, and one may conclude that these valleys are incipiently unstable and that eventually a flood will cause erosion and trenching in these valleys.

Using Figure 1, one may define the threshold slope above which trenching or valley instability will take place in the Piceance Creek area. This has obvious implications for land management for, if the slope at which valleys are incipiently unstable can be determined, corrective measures can be taken to artificially stabilize such critical reaches, as they are identified.

Note also that both intrinsic and extrinsic threshold conditions can be recognized on Figure 1. The steepening of the valley slopes by deposition will result in the exceeding of an intrinsic geomorphic threshold, whereas an increase of run-off would have the effect of shifting the points to the right until some cross the threshold line. Obviously, drainage area cannot be increased, but, if a hydrologic variable were plotted on the abscissa of Figure 1, the points would shift to the right with an increase of discharge.

It seems possible that future work will demonstrate that similar relationships can be established for other alluvial deposits. For example, fan-head trenches are another erosional feature of variable distribution, and within a given region some fans are trenched and some are not trenched (Bull, 1968). The usual explanation for these features involves the extrinsic controls of climate, tectonics, and land use, but strong geomorphic controls may also be present. For example, during experimental studies of fan growth, (Schumm, 1977) precipitation was delivered to a sediment source area at a constant rate. The resulting run-off and sediment moved

out of the drainage basin to a piedmont area where a miniature fluvial fan was formed. The fan is termed fluvial rather than alluvial, because the streams that were carrying the water and sediment from the source area were perennial. During fan growth, the fan was trenched repeatedly, as the fan-head steepened as a result of aggradation and then adjusted to this oversteepened condition by trenching (Fig. 2). It should be emphasized that the trenching and the reworking of the sediment and the subsequent back-filling of the channel by aggradation were not due to any change in the experimental procedure or to changes in the intensity or duration of the precipitation applied to the source area. The fan-head trenching occurred as a result of the oversteepening of the fan-head, and the trenching was the result of exceeding an intrinsic geomorphic threshold.

It is important to realize that the threshold slope as indicated on Figure 2 will probably not be a single value; rather there should be a range of slopes in a threshold zone that will depend on vagaries of hydrology, land use, and lithology of the source area.

The concept of thresholds as applied to alluvial deposits in the western U.S.A. is illustrated by Figure 3, where the decreasing stability of an alluvial fill is represented by a line indicating increase of valley slope with time. Of course, a similar relation would pertain if, with constant slope, sediment loads decrease slowly with time. Superimposed on the ascending line of increasing slope are vertical lines showing the variations of valley floor stability caused by flood events of different magnitudes. The effect of even large events is minor until the stability of the deposit has been so reduced by steepening of the valley gradient that during one major storm, erosion begins at time A. It is important to note that the large event is only the most apparent cause of failure, as it would have occurred at time B in any case.

Studies of alluvial deposits in drylands suggest that large infrequent storms can be significant but only when a geomorphic threshold has been exceeded. It is for this reason that
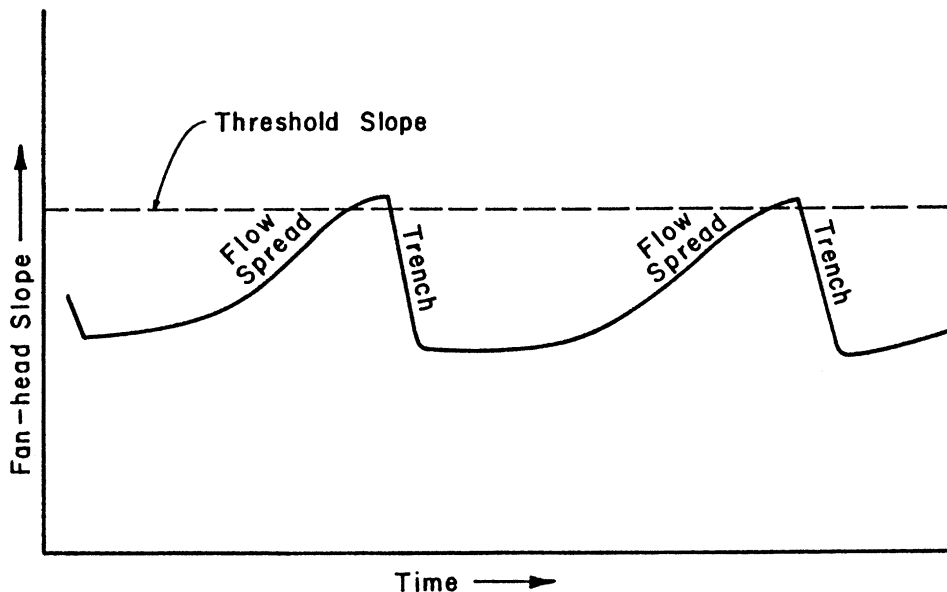

FIGURE 2. Diagram showing change in fan-head morphology through time, during an experimental study of fan growth. When flow is spread at the fan-head, deposition increases the slope of the fan-head to a critical slope, at which time the fan-head is trenched, and the flow is confined to the trench. Backfilling of the trench leads to a spread-flow condition at the fan-head. This sequence of events occurred repeatedly during the experimental study
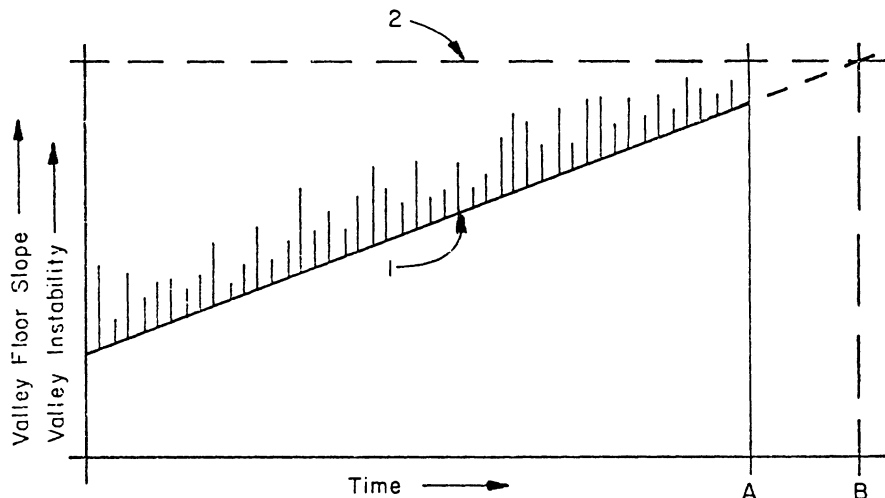
FIGURE 3. Hypothetical relation between valley-floor gradient and valley-floor instability with time. Superimposed on line 1, representing an increase of valley-floor slope, are vertical lines representing instability of the valley floor as related to flood events. When the ascending line of valley-floor slope intersects line 2 representing the maximum slope at which the valley is stable, failure or trenching of the valley alluvium will occur at time B. However, failure may occur earlier, at time A as the result of a major storm or flood event

high-magnitude low-frequency events may have only minor and local effects on a landscape in this morphoclimatic region.

River patterns Experimental studies of river meandering have been performed by both hydraulic engineers and geologists over many years. Such a study was designed to investigate the influence of slope and sediment loads on channel patterns (Schumm and Khan, 1972). It was found, during these constant-discharge experiments, that if a straight channel was cut in alluvial material at a very low slope, the channel would remain straight. However, at steeper slopes the channel meandered (Fig. 4). As the slope of the alluvial surface on which the model stream was flowing (valley slope) was steepened, the velocity of flow increased, and shear forces acting on the bed and bank of the miniature channel increased. Slope in this case is an index of sediment load and the hydraulic character of the flow as expressed by stream power. At some critical value of stream power, bank erosion and shifting of sediments on the channel floor produced a sinuous course. As slope increased beyond this threshold, meandering increased until at another higher threshold the sinuous channel became a straight braided channel. The experiments revealed that there is not a continuous change in stream patterns with increasing slope from straight through meandering to braided, but rather the changes occur essentially at two threshold values of slope, or stream power.

If the slope of a valley floor varies according to the sediment contribution from tributaries, to variations in deposition during the geologic past or to tectonics, then the river should reflect these changes of valley slope by changes in pattern. A comparison of the experimental results with Mississippi River patterns was made possible by data obtained from the Vicksburg District, Corps of Engineers. When these data are plotted, they show that variations in the channel pattern of Mississippi River are related to changes of the slope of the valley floor (Fig. 5).

Summary

Although the increase of valley slope by deposition to a condition of instability is obviously an
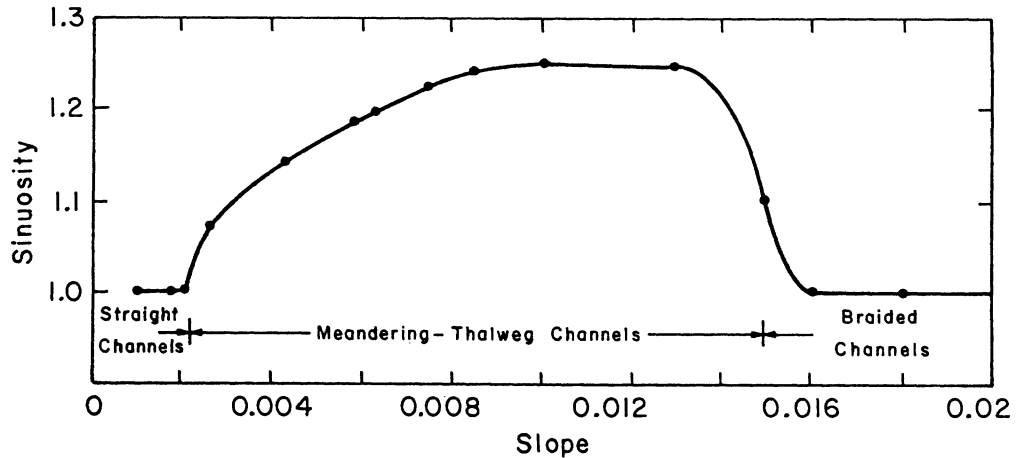
FIGURE 4. Relation between flume slope and sinuosity (ratio of channel length to length of flume or length of valley) during experiments. The change from a straight to a sinuous pattern, and from a sinuous to a braided pattern occurs at two threshold slopes. The absolute value of slope at which such changes occur will be influenced by discharge. Discharge was maintained at 0·15 cfs during the experiments. *Source:* Schumm and Khan, 1972
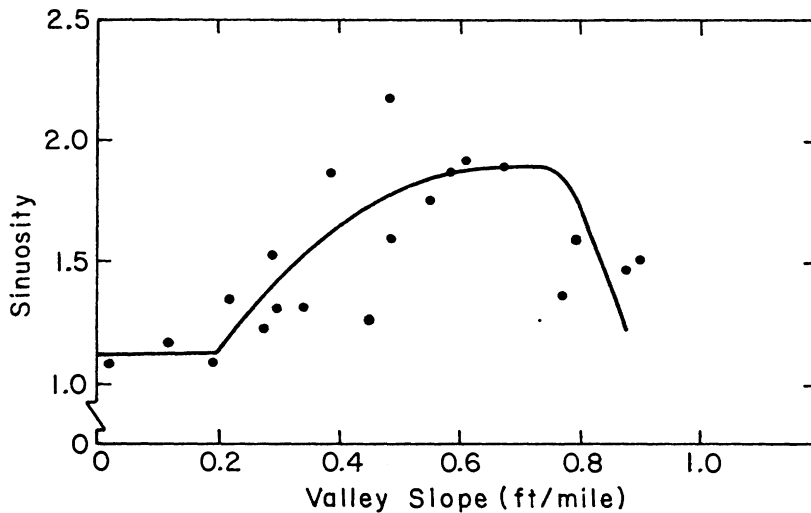

FIGURE 5. Relation between valley slope and sinuosity, Mississippi River between Cairo, Illinois and Head of Passes, Louisiana. Data from Potamology Section, U.S. Army Corps of Engineers, Vicksburg, Mississippi. Scatter about curve reflects natural sinuosity variation due to meander growth and cut-off. *Source:* Schumm *et al.*, 1972

intrinsic control, the variations of valley-floor gradient that produce river-pattern changes are an extrinsic control, as would be any change in sediment load or stream power that forces a channel across a pattern threshold (Fig. 4).

A pattern change in response to intrinsic controls can occur when during meander development, sinuosity and meander amplitude become too great, and the stream gradient is reduced to the point where aggradation and meander cut-offs are induced.

When the valley slope is near a threshold, a major flood will significantly alter the stream pattern. This conclusion has bearing on the work of Wolman and Miller (1960), concerning the

geomorphic importance of events of high magnitude. They concluded that, although a major amount of work is done by events of moderate magnitude and relatively frequent occurrence, nevertheless, the large storm or flood may have a major role in landscape modification. However, evidence on the influence of rare and large events on the landscape is equivocal. Major floods have destroyed the flood-plain of the Cimarron River (Schumm and Lichty, 1963), but they did not significantly affect the Connecticut River (Wolman and Eiler, 1958).

These and other observations indicate that a major event may be of either major or minor importance in landscape modification, and an explanation of the conflicting evidence requires further consideration of the threshold concept. Some landscapes or components of a landscape have apparently evolved to a condition of geomorphic instability and these landforms will be significantly modified by a large infrequent event whereas others will be unaffected. Therefore, there will be, even within the same region, different responses to the same conditions of stress.

When within a landscape some components fail by erosion whereas others do not, it is clear that erosional thresholds have been exceeded locally. An important task for the geomorphologists is to locate incipiently unstable components of a landscape. The recognition of geomorphic thresholds within a given region will be a significant contribution to the understanding of the details of regional morphology as well as providing criteria for identification of incipiently unstable land forms.

COMPLEX RESPONSE AND EPISODIC EROSION

It is inherent in the threshold concept that a landscape is not always in a condition of grade, balance or equilibrium. The existence of the threshold suggests an inability of the landform to adjust readily to a new equilibrium condition. During experimentation, a small drainage system (Schumm and Parker, 1973; Schumm, 1977) was rejuvenated by a slight (10 cm) change of base level. As anticipated, base-level lowering caused incision of the main channel and development of a terrace (Fig. 6 A, B). Incision occurred first at the mouth of the system, and then progressively upstream, successively rejuvenating tributaries and scouring alluvium previously deposited in the valley (Fig. 6B). As erosion progressed upstream, the main channel became a conveyer of upstream sediment in increasing quantities, and the inevitable result was that aggradation occurred in the newly cut channel (Fig. 6C). However, as the tributaries eventually became adjusted to the new base level, sediment loads decreased, and a new phase of channel erosion occurred (Fig. 6D). Thus, initial channel incision and terrace formation was followed by deposition of an alluvial fill, channel braiding, and lateral erosion, and then, as the drainage system achieved stability, renewed incision formed a low alluvial terrace. This low surface formed as a result of the decreased sediment loads when the braided channel was converted into a better defined channel of lower width–depth ratio.

Somewhat similar results were obtained by Lewis (1949) in a pioneering experiment performed in a small wooden trough 4 m long and 50 cm wide. Lewis cut a simple drainage pattern in sediment (four parts sand, one part mud). He then introduced water at the head of the flume into both the main channel and into two tributaries. The main channel debouched onto a 'flood-plain' before entering the 'sea' (tail box of flume).

During the experiment, the break in slope or knickpoint at the upstream edge of the 'flood-plain' eroded back, rejuvenating the upstream drainage system. Initially, erosion in the head waters was rapid as the channels adjusted, and deposition occurred on the 'flood-plain', thereby increasing its slope. As the upstream gradients were decreased by erosion and the stream courses stabilized, sediment supply to the 'flood-plain' decreased. Because of the reduction of the sediment load the stream cut into the alluvial deposits in the upper part of the flood-plain to form a terrace. Lewis concludes that 'perhaps the most significant fact
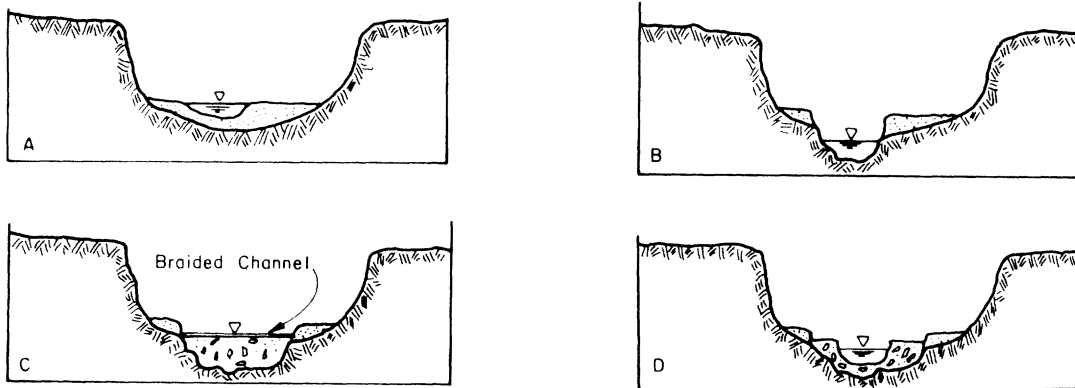
FIGURE 6. Diagrammatic cross sections of experimental channel 1·5 m from outlet of drainage system (base level) showing reponse of channel to one lowering of base level.

A. Valley and alluvium, which was deposited during previous run, before base level lowering. The low width-depth channel flows on alluvium.

B. After base level lowering of 10 cm, channel incises into alluvium and bedrock floor of valley to form a terrace.

C. Following incision, bank erosion widens channel and partially destroys terrace. An inset alluvial fill is deposited, as the sediment discharge from upstream increases. The high width-depth ration channel is braided and unstable.

D. A second terrace is formed as the channel incises slightly and assumes a low width-depth ratio in response to reduced sediment load. With time, in nature, channel migration will destroy part of the lower terrace, and a flood plain will form at a lower level

revealed by the ... experiment is that terraces were built in the lower reaches without any corresponding change of sea levels, tilt or discharge.'

The experimental results indicate that an event causing an erosional response within a drainage basin (tilting, changes of base level, climatic and/or land use) automatically creates a negative feedback (high sediment production) which results in deposition; this is eventually followed by incision of alluvial deposits as sediment loads decrease.

A similar sequence of events may be under way in Rio Puerco, a major arroyo in New Mexico, as well as in other south-western channels. For example, the dry channel of Rio Puerco, although previously trenched to depths of 13 m, is now less than four m deep near its mouth. This is due to deposition caused by very high sediment loads produced by the rejuvenated drainage system.

The complex response observed during experiments and in the field is apparently a quest for a new equilibrium by a complex system. In fact, the changes documented in Figure 6 are minor as compared to those accompanying major periods of aggradation or degradation which result from major climatic or base level changes. Under these circumstances the complex response becomes episodic erosion and deposition.

It is well established in the geomorphic literature that a major reduction of base level will cause progressive downcutting and readjustment of the stream gradients until a new graded or equilibrium situation has been developed. In fact, the fluvial system may not be capable of regrading in this way when sediment movement is great (Hey, 1979). For example, the fan-head trenching described earlier (Fig. 2) represented an episodic growth of the fan. Periods of deposition and growth are interrupted by brief periods of incision and reworking of the previously deposited sediment.

Where major incision has occurred in alluvium and relatively weak rocks, evidence of pauses in the erosional downcutting are found, but this is usually attributed to some external

influence, such as variations in climate, the rate of base-level change, or variations in the rates of uplift of the sediment source area. However, recent studies in the Douglas Creek drainage basin of western Colorado support the idea of discontinuous downcutting. The investigation of recent erosional history of this valley shows that modern incision of the valley fill began after 1882. Yet, there are four surfaces now present below the two pre-1882 surfaces (Fig. 7). These surfaces are unpaired, discontinuous terraces that elsewhere have usually been explained by the shifting of a channel laterally across the valley floor, during progressive downcutting (Davis, 1902). In the Douglas Creek Valley, however, downcutting was discontinuous. In fact, during pauses in downcutting there was deposition. The denudation scheme as sketched on Figure 7 portrays the sequence of events in this 1000 km² drainage basin in western Colorado. That is, during incision of the main channel, there is rejuvenation of tributaries and a progressive increase in sediment yield from upstream. Sediment loads become so great that downcutting ceases and deposition begins. Deposition continues until it is possible for the channel to incise again and to continue the down-cutting process.

The Douglas Creek situation appears to conform to the observations of Born and Ritter (1970) who mapped six discontinuous and unpaired terraces at the mouth of Truckee River, where it enters Pyramid Lake in Nevada. A reduction of the water level in Pyramid Lake reduced the base level of the lower Truckee River, but instead of simple downcutting commensurate with the lowering of the base level, the channel in fact paused as many as six times.
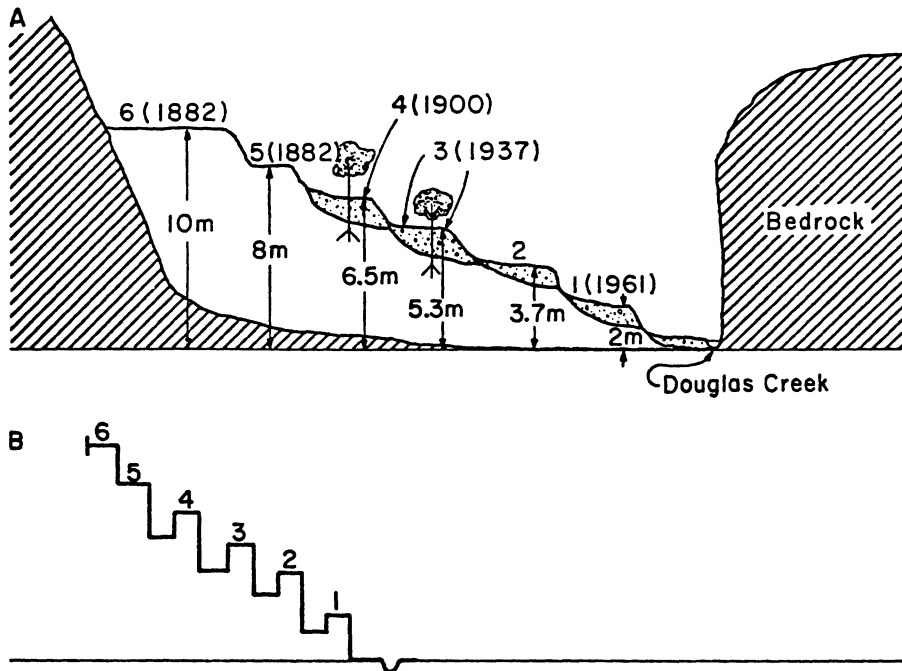

FIGURE 7.
  A. Sketch of Douglas Creek valley showing erosion surfaces formed since 1882. Age of surfaces is based on tree-ring dating and historical data. Note burial of trees by deposition. Surfaces 5 and 6 were present before modern erosion began after 1882.
  B. Summary of behaviour of Douglas Creek. Vertical segments indicate incision or deposition, horizontal segments periods of relative stability. *Source:* Womack and Schumm, 1977

Gage (1970) cites an example of rapid deposition, which caused aggradation from 3 to 25 m in the Waiho River of New Zealand. This glacier-fed river then proceeded to clear the deposited sediment over a period of a few weeks. The erosion produced a flight of 3 m terraces. Gage attributed this and similar events to 10-year weather patterns, and he cautioned that if some of these terraces were preserved they could easily be mistaken for surfaces of considerable antiquity.

Other examples can be cited of multiple unpaired terraces that may have formed as a result of rapid but episodic incision similar to that which occurred in the Douglas Creek Valley (Davis, 1902; Small, 1973), and the terraces of the Rangitata River in New Zealand (Fig. 22B) the Snake River and the Shoshone river in the U.S. are possible examples.

Another influence that can produce episodic erosion is armouring. As a channel degrades it may concentrate the coarse fraction of its sediment as a gravel or cobble lag which protects the channel. This armour can provide a protection that prevents further incision until a flood occurs that breaches the armour. In this case both intrinsic and extrinsic controls act to produce episodic erosion.

During an experimental study of gully erosion the sediment used contained a small fraction of granules and pebbles, that progressively armoured the incising channel (Begin, 1979). What was of interest, however, was the manner in which sediment production was influenced by the armour (Fig. 8). Sediment production declined exponentially with time, as in experiments
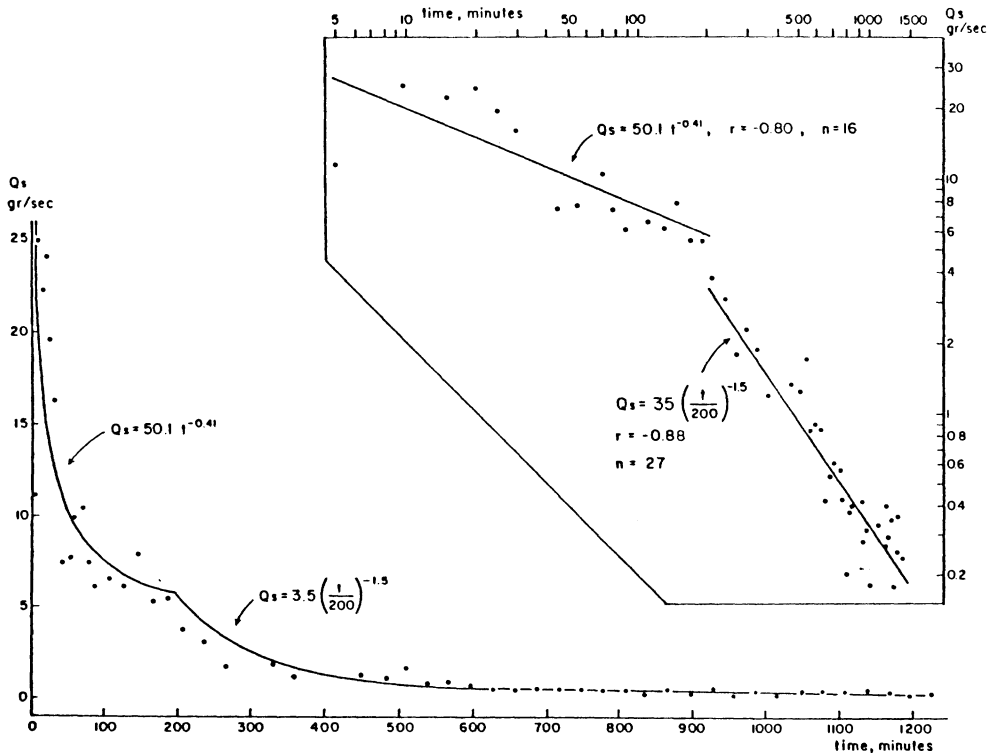

FIGURE 8. Effects of development of armours on sediment discharge, Qs, during experimental study of gully erosion. At about 200 minutes into the run the armour became effective and abruptly reduced sediment discharge. The change is displayed very clearly in the logarithmic plot

without the development of armour, but the effect of the armour eventually caused a very abrupt and significant decrease of sediment production.

Apparently, as the armour formed, it decreased the amount of sediment contributed by bed erosion, but the flow attacked the banks and widened the gully. Hence, the armour did not obviously influence the quantity of sediment produced. However, a point was reached at which channel widening and bank erosion ceased, and the channel then flowed on the armour and could not acquire a large load from the banks. The abruptness of the decrease of sediment production is startling, and reminiscent of the decrease in the Colorado River sediment in the 1940s (Schumm, 1977, p. 326).

Begin's experiments were run at constant discharge, but in a natural situation the armour would be breached by high discharges. Destruction of the armour and its renewal by further reworking of alluvium would produce pulses of sediment from the channel.

An analogy to episodic erosion that occurs under constant conditions is provided by the hydraulic behavior of Medano Creek in Great Sand Dunes National Monument, Colorado (Bean and Schumm, 1979). A steady flow from the Sangre de Cristo Mountains moves through a cobble-bed channel until it enters the aeolian-sand-bed channel near the mountain front. In the sand channel antidunes form and wash out, thereby causing large variations of channel roughness. Water is stored in the rough antidune reaches and it is released as a surge when the antidunes wash out. A pulsating flow is generated by this process that converts the steady flow to surges with a periodicity of about 30 seconds. The surges, although only about 2 cm high, can be followed downstream for at least 2 km. The pulsating flow is not a result of variations in external controls, rather it is generated by the repeated crossing of bed-form thresholds under steady flow conditions.

In summary, the concept of geomorphic thresholds leads away from the generally accepted ideas of progressive erosion and progressive response to altered conditions. Rather the fluvial system must hunt for a new equilibrium (complex response) and when the change is major it is overwhelmed by the quantity of sediment that requires movement and the results are episodes of erosion and deposition (episodic erosion). The processes, when understood, provide a basis for the development of a revised model of geomorphic evolution and for predictive geomorphology.

APPLICATIONS

The concepts of geomorphic thresholds, complex response and episodic erosion have significant applications to classic geomorphic problems as well as to modern land management and river engineering problems.

Landscape Evolution

It is stimulating to consider the grand changes of a landscape during the millions of years of its erosional evolution. Unfortunately, this overview provides little assistance to those concerned with the short-term behaviour of landforms, as the previous discussion suggests.

The reason that most models of geomorphic evolution are unsatisfactory for short-term interpretation is that they are oversimplified, and this is largely because they are based on very limited information. For example, the extrapolation of average denudation rates from a ten-year record to a thousand or million years of erosional evolution of a landscape produces a model of landscape evolution that is based on an assumption of progressive slow change, which probably is not correct (Gage, 1970).

The criticism of the Davis geomorphic cycle in the writings of Penck and John Hack reflect their concern with this simplistic model. Some years ago, Lichty and I (1965) attempted to
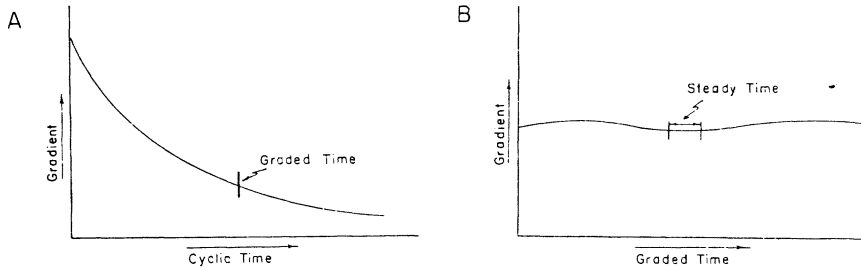
FIGURE 9. Diagram illustrating change of channel gradient during different time spans. *Source:* Schumm and Lichty, 1965
   A. Progressive reduction of channel gradient during cyclic time. During graded time, a small fraction of cyclic time, the gradient remains relatively constant.
   B. Fluctuations of gradient above and below a mean during graded time. Gradient is constant during the brief span of steady time.

resolve some of the controversy by considering the landscape during very different spans of time. The time required for the denudation of a landscape was subdivided into cyclic, graded and steady time periods (Fig. 9). Under the category of cyclic time are time spans of geologic duration, that is, the period of time required for the denudational evolution of a landscape. For example, during this period, one expects an essentially exponential decrease of stream gradient, which is a landscape component that reflects changes in the fluvial system. However, cyclic time can be subdivided into graded time and steady time periods. During graded time, average gradient will remain relatively constant, but there will be, through time, fluctuations about this mean. Graded time, therefore, conforms to the definition of graded stream, as expressed by Mackin (1949). During the very short period of steady time, there is no change. When considering a landscape or its components, it is helpful to think in terms of the time spans and how a landscape is altered during the time spans under consideration. In short, the period of time referred to as being cyclic or geologic in duration can be represented by the Davis curve showing the erosional evolution of the landscape (Fig. 10). Of course, it seems unlikely that denudation will continue for such an appreciable period of time without interruption by climate change or by isostatic adjustment. So the smooth curves presented by Davis (Fig. 10) and by Schumm and Lichty (Fig. 9A) can be expected to be complicated. For example, the progressive erosion of a region will probably cause short but dramatic periods of isostatic adjustment (Fig. 11). With an essentially constant rate of denudation a condition is reached when isostatic compensation is necessary, and this probably takes place during a short period of relatively rapid uplift (Schumm, 1963).

It is understood that a change of an external variable will interrupt the progress of the geomorphic cycle, and changes of stream profile and variations of gradient during graded time are readily understood as reflecting variations of discharge and sediment load (Fig. 9B).
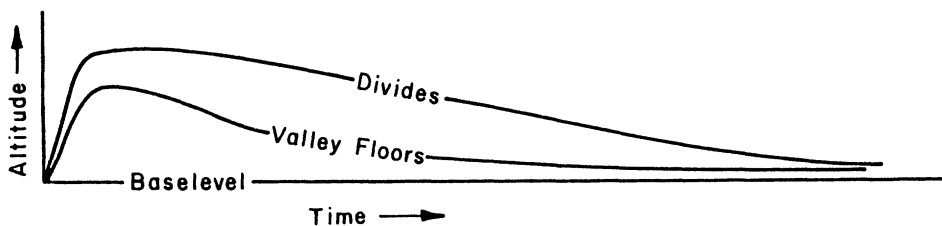

FIGURE 10. The Davis geomorphic cycle as usually presented

Nevertheless, Figure 9 poses a problem. It is difficult to imagine how the graded time curve (Fig. 9B) can be compatible with the cyclic time curve (Fig. 9A). The progressive reduction of gradient shown on Figures 9a and 10 seems reasonable, but this in turn prevents a graded condition from developing until 'old age' (Fig. 10). Conversely, if graded conditions exist, then progressive reduction of the valley floor and stream gradient is impossible.

This line of reasoning apparently requires the elimination of either the concept of progressive erosion or grade. However, there is an alternative solution. If valley floors (Fig. 10) and stream gradients (Fig. 9A) do not evolve progressively but change rapidly, during brief periods of instability that separate longer periods of grade, then a model incorporating both progressive change and grade can be proposed.

Another important aspect of landscape denudation is that, as a landscape changes, the components of the landscape, hill-slopes, tributaries, and main channels, will not necessarily be adjusted to one another or be graded. That is, a channel adjusting to uplift may not be able to cope with the effects of the rejuvenation in the watershed upstream. Hence, an actively eroding system will be continually searching for a stability that cannot be maintained (complex response and episodic erosion). Interestingly enough, Davis was to some extent aware of this problem. For example, when the diagram illustrating the geomorphic cycle that was prepared by Davis himself is inspected (Fig. 12), it is apparent that part of his scheme is missing in Fig. 10. For example, following a description of the erosion cycle (Davis, 1899, p. 254–255) Davis returns to
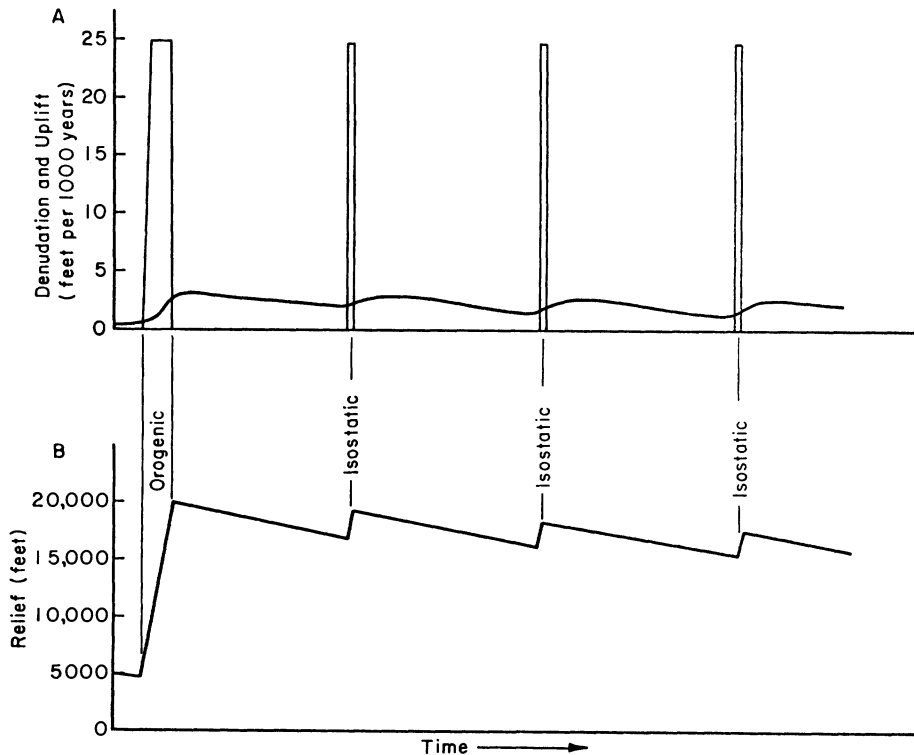

FIGURE 11. The effect of isostatic adjustment on the geomorphic cycle. *Source:* Schumm, 1963a
  A. Hypothetical relation of rates of uplift (25 feet per 1000 years) and denudation (3 feet per 1000 years) to time.
  B. Hypothetical relation of drainage basin relief to time as a function of uplift and denudation shown in A)
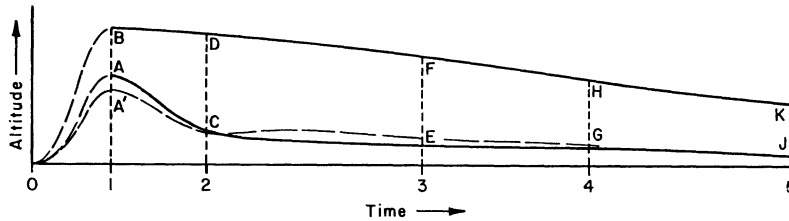
FIGURE 12. The geomorphic cycle according to Davis (1899)

the topic of valley deepening and the effect of sediment load on this process, and he clearly states that downwearing of the valley floor may not be a continuous process. That is, at some stage the main river adjusts by aggradation to the increased quantities of sediment being delivered from upstream, and a valley-fill deposit is formed within the valley. This is shown by the broken line C, E, G on Figure 12. Davis, thus, envisaged a period of aggradation during the progressive erosion of the drainage basin. This concept was also stated in an earlier paper (Davis, 1895, p. 130).

The deduction of Davis that deposition will naturally follow initiation and presumably rejuvenation of a drainage system was a very astute one. This idea seems to have been ignored; nevertheless, the possibility that deposition takes place naturally within the drainage system is of great importance.

In summary, it appears that an uplifted drainage system will have difficulty in disposing of all the sediment delivered to its major channels from minor tributaries and interfluve areas, and sediment will be stored within the system. The sediment may be stored until an intrinsic threshold is exceeded and erosion is renewed. Depending on the volume of sediment that is to be eroded and the distance to base level, the river may be unable to adjust and a period of erosional and depositional complexity results.

When the concepts of geomorphic thresholds, complex response, and episodic erosion are applied to landscape evolution the model becomes as summarized by Figure 13, which is essentially a modification of the left side of the Davis scheme shown in Figures 10 and 12 (youth and early maturity), but the progress of denudation is interrupted by periods of isostatic adjustment. As in Figures 10 and 12, the upper line of Figure 13A represents divide elevations and the lower line valley-floor elevations. The divide elevations probably change as shown by Figure 13A. That is, only major external influences affect the divides, and they are subjected to a relatively uniform downwearing. However, if the valley floor is considered in greater detail over a shorter span of time (Fig. 13B), a stepped pattern of valley floor reduction emerges, as a result of storage and flushing of sediment from the valleys. This model ignores variations due to external influences, and it shows a system that is in dynamic metastable equilibrium (Chorley and Kennedy, 1977).

A steady state equilibrium involves fluctuations about an average, but a metastable equilibrium occurs when an external influence carries the system over some threshold into a new equilibrium regime. The effects of external variables on equilibrium systems are expected, but in the case of landscape denudation the dynamic metastable equilibrium may reflect the response of the system to inherent geomorphic thresholds (Schumm, 1973), that is, the accumulation of sediment to an unstable condition. When a geomorphic threshold is crossed, the drainage system will be rejuvenated, and the complex response will come into play (Fig. 13C). Figure 13C shows periods of instability separated by longer periods of dynamic equilibrium or grade. Because periods of erosion are followed by periods of deposition, the bedrock
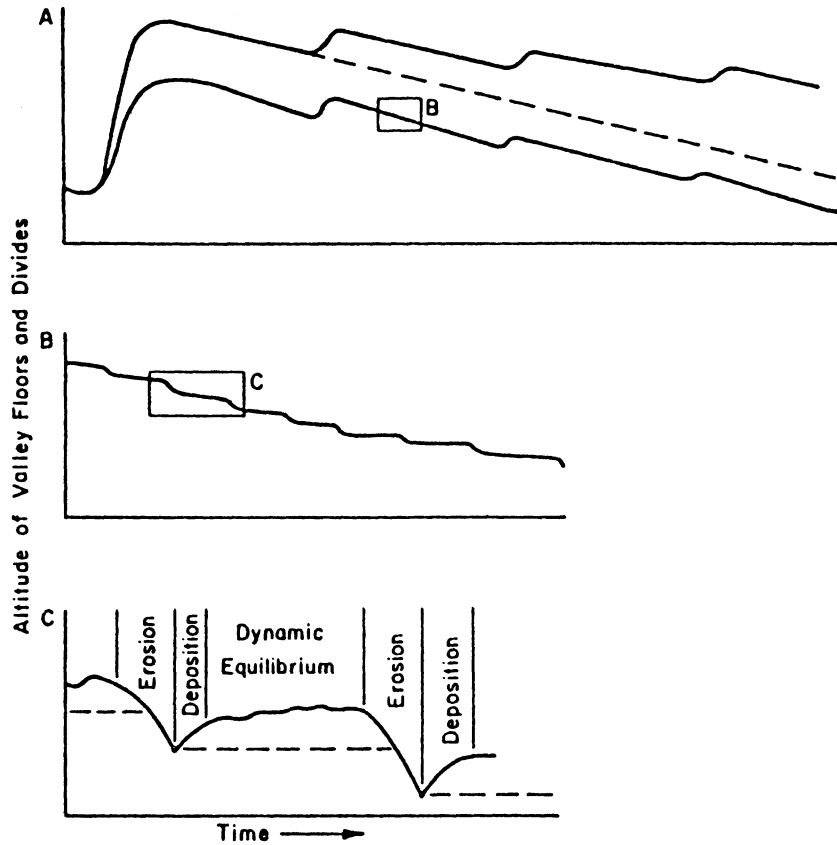
FIGURE 13. Modified concept of geomorphic cycle.
   A. Erosion cycle, as envisioned by Davis (broken line), following uplift, and as affected by isostatic adjustment to denudation.
   B. Portion of valley floor in A above, showing episodic nature of decrease of valley-floor altitude.
   C. Portion of valley floor in B above, showing periods of instability separated by longer periods of dynamic equilibrium. *Source:* Schumm, 1977

floor of the valley is reduced in a steplike manner through time, as shown by the broken lines. Hence, separating the periods of erosion there will be periods of deposition and storage of alluvium. In addition, during these periods of relative stability, channel pattern may change from straight to increasingly sinuous, as the nature of the sediment that is moved through the channel changes (Schumm, 1963b). Hence, sinuosity may also increase to a condition of incipient instability, when a large flood can cause an abrupt shortening of the river course, of the type documented by Fisk (1944) for Mississippi River.

   If episodic erosion takes place, many of the details of the landscape, small terraces, and recent alluvial fills do not need to be explained by the influence of external variables because they develop as an integral part of system evolution. The Davis curve of Figure 10 averages out these variations during cyclic time (Fig. 9A).

   It is unlikely that episodic erosion and the dynamic metastable equilibrium model of landform evolution will apply in situations of low relief and low sediment yield. Therefore, it is suggested that it will dominate during the early stages of landform evolution (Fig. 14) and today
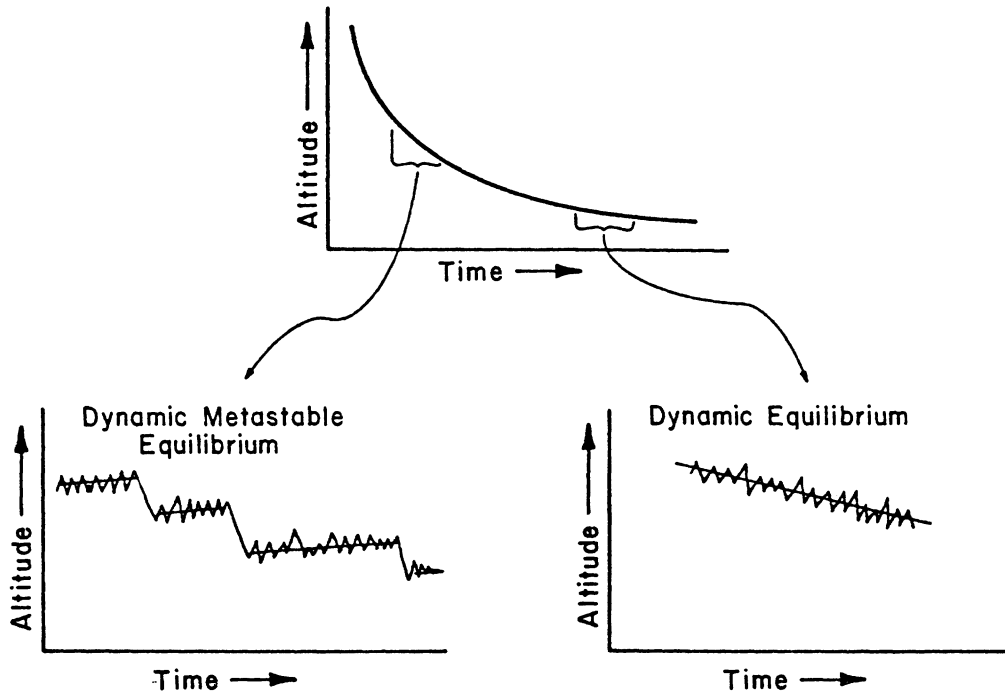
FIGURE 14. Diagrams illustrating erosional evolution of valley floors in youth by episodic erosion (dynamic metastable equilibrium) and later by progressive erosion (dynamic equilibrium). *Source:* Schumm, 1977

in areas of high relief and high sediment production. The dynamic equilibrium model will apply during later stages of the geomorphic cycle.

The model of landscape evolution advanced in Figures 13 and 14 may seem to pose problems for the interpretation of some landscape details, but of course the usual appeal to climate fluctuation or tectonics is no solution. In fact, one need not attempt to explain all of the details of a landscape as a result of climate change or diastrophism because some of the complications are inherent in landscape evolution. To one concerned with river stability and erosion control this model may be of considerable aid, because it suggests that it is possible to recognize unstable components of the landscape through the identification of geomorphic threshold slopes, and to take measures to prevent further development of the instability (Patton and Schumm, 1975).

In geologically and geomorphically homogeneous regions the high variability of sediment yield and run-off characteristics of drainage basins may also be explained by this model, that is, even within the same geomorphic region, similar drainage basins need not be producing comparable quantities of sediment, as sediment is stored and flushed. In effect, the revised model explains why the attempt to relate geomorphic variables to hydrology has been successful but less satisfying than one would like.

Terraces and pediments   If, as described above, channels incise as the result of the exceeding of intrinsic thresholds, and early cycle erosion is episodic, then it may not be possible to correlate some erosional and depositional surfaces except locally. For example, it is well known that in a piedmont region stream capture causes dissection and abandonment of terraces (Ritter, 1972) and pediments (John Rich, 1935, Charles Hunt *et al.,* 1953), yielding a history of episodic erosion unrelated to external variables.

Figure 15 shows pediment development at the base of an escarpment. This is the situation that exists along the base of the Book Cliffs between Grand Junction, Colorado and Price, Utah. Streams draining from the Book Cliffs transport coarse sediments, and they require a steeper gradient than the piedmont streams that are eroding into shale (Fig. 15a). The aggressor shale streams (A), therefore, have a lower gradient, and the higher streams are vulnerable to the capture by shale streams (Figs 15b and c).

There are four erosion surfaces that have been preserved on the Mancos shale in this area. The high-level surfaces are beautifully developed pediments with very clear contacts between the shale bedrock and the overlying gravels. In a longitudinal section the pediments sweep away from the cliffs in a smooth curve; however, where the contact can be seen in transverse sections the bedrock surfaces of the pediment are very irregular (Carter, 1979), but at most locations the irregularity of the redrock surface is masked by the alluvial or gravel deposits above it.

The shale streams at the base of the scarp (Fig. 15) erode a broad piedmont area at a level
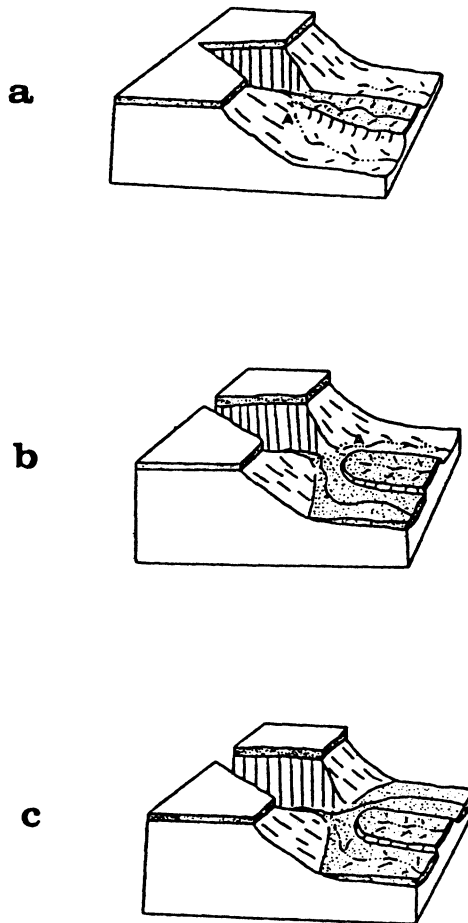

FIGURE 15. Development of multiple pediments by stream capture along an escarpment. Capture of the high-level stream by the southern stream (A) will form a second pediment level (a, b). Continued erosion of the weaker piedmont rocks by the northern stream A will eventually cause capture and shift of the channel to the northern lowland to form a third pediment (c.) *Source:* Carter, 1977

below the gravel-bearing streams, and their tributaries produce an irregular shale surface at this low level. When capture of the high level stream takes place, the gravels from the scarp inundate this lower surface and convert it from an irregular bedrock surface to a smooth gravel-covered surface that is a new lower-level pediment surface.

The multiple pediment levels in this area can be attributed to stream capture without any external influence, although it is probable that, during the erosional history of the area, changes in the level of the Colorado River and climate changes have influenced pediment development. Nevertheless, the pediment surfaces can be explained primarily as a result of the normal and intrinsic process of stream capture in this area.

The concept of geomorphic thresholds and episodic erosion is particularly important in this respect, because the geomorphologist cannot automatically assume that each erosional or depositional discontinuity in the landscape is a result of external influence.

To summarize, when the influence of external variables such as isostatic uplift is combined with the effects of complex response and geomorphic thresholds, it is clear that denudation, at least during the early stage of the geomorphic cycle, will not be progressive. Rather, it should be comprised of episodes of erosion separated by periods of relative stability, a complicated sequence of events. Much of this complexity is the result of a delayed transmission of information through the system. That is, channel changes that take place near the mouth of a drainage basin following incision are in response to the conditions at that time and location. Therefore, the channel at that point is not prepared for the subsequent changes that its incision induces within the system upstream; hence, downcutting is followed by deposition when the upstream response occurs (Figs 6 and 7). Application of the new concepts therefore, yields a revision of the standard geomorphic cycle from that of Figure 10 to that of Figures 13 and 14.

Practical applications

Gully erosion   The concept of geomorphic thresholds is useful in identifying those conditions at which a landform is incipiently unstable. Following this identification some action can be taken either to prevent failure from occurring or to minimize the effect of the change when it does occur. For example, regime equations are an attempt to design a stable channel based on the forces acting on the materials forming the perimeter of a canal. A critical velocity is identified above which erosion will take place, and the canal is designed so that this velocity will not be exceeded.

Erosion begins at some extrinsic threshold condition of tractive force or shear. Recent work by Graf (1979) in small valleys near Evergreen, Colorado has demonstrated that one can locate the threshold value of tractive force at which a valley is going to be subjected to gullying. Figure 16 shows this relationship between the density of vegetation on the valley floor, (biomass), and tractive force. It is clear that for a given biomass there is a critical tractive force at which incision of the valley floor will take place, and for a given tractive force there is a minimum value of biomass below which the valley floor becomes unstable and gullying begins.

Changes of both biomass and tractive force can be the result of land-use changes and climatic fluctuations, which weaken valley-floor vegetation and which increase peak discharge and tractive force. However, with external influences unchanged, the tractive force can be increased if the valley-floor slope is steepened by aggradation, an intrinsic control.

Elsewhere in western Colorado the relationship between drainage area and valley-floor inclination (Fig. 1) permits identification of a threshold valley-floor slope above which incision of the valley floor is likely to take place. However, the task is difficult because in small drainage basins there are other factors that act to obscure this relationship. For example, unless climate,
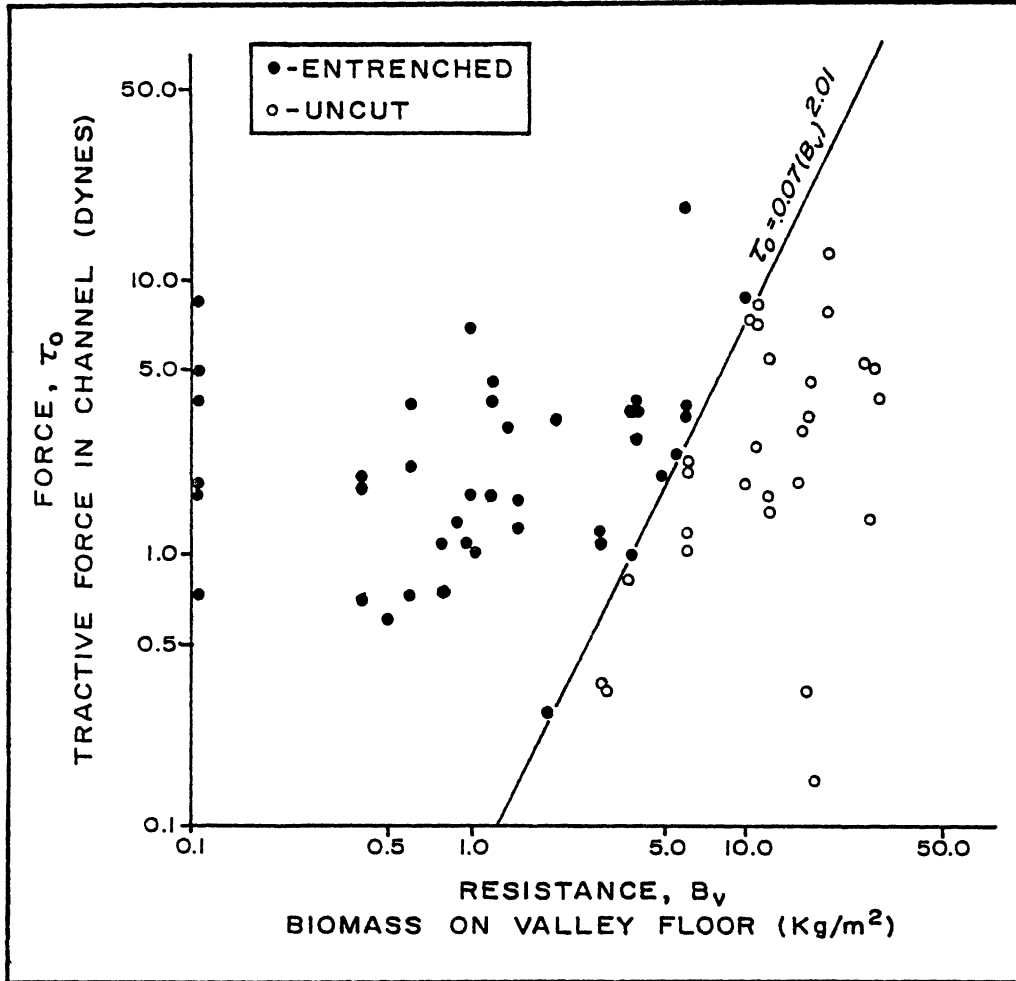
FIGURE 16. Tractive force (calculated for flood of 10-year recurrence interval) and biomass of valley floors with discriminating function separating stable and unstable valley floors and channels. *Source:* Graf, 1978

land use, and vegetative cover are essentially uniform throughout an area, the valley-floor slope criterion may be masked.

More recent studies along the Chalk Bluffs of north-eastern Colorado (Bradley, 1979) showed that valley width is also an important variable. In these small drainage basins narrowing of the valley increases the depth of flow and tractive force or stream power, which in turn causes erosion on gentler slopes in the narrower valleys. A ratio of critical valley slope to valley width incorporates these two variables and when plotted against drainage area, a plot similar to that of Figure 1 results. However, instead of a threshold line there is a zone in which both gullied and ungullied valleys plot (Fig. 17). This obviously is a more realistic relation considering the other variables that have not been included.

Further analysis of the data of Figure 1, which involved the estimation of 10-year flood discharge and calculations of the tractive force exerted on the valley floor by that flood, showed that when plotted on logarithmic paper, a zone rather than a line of discrimination exists
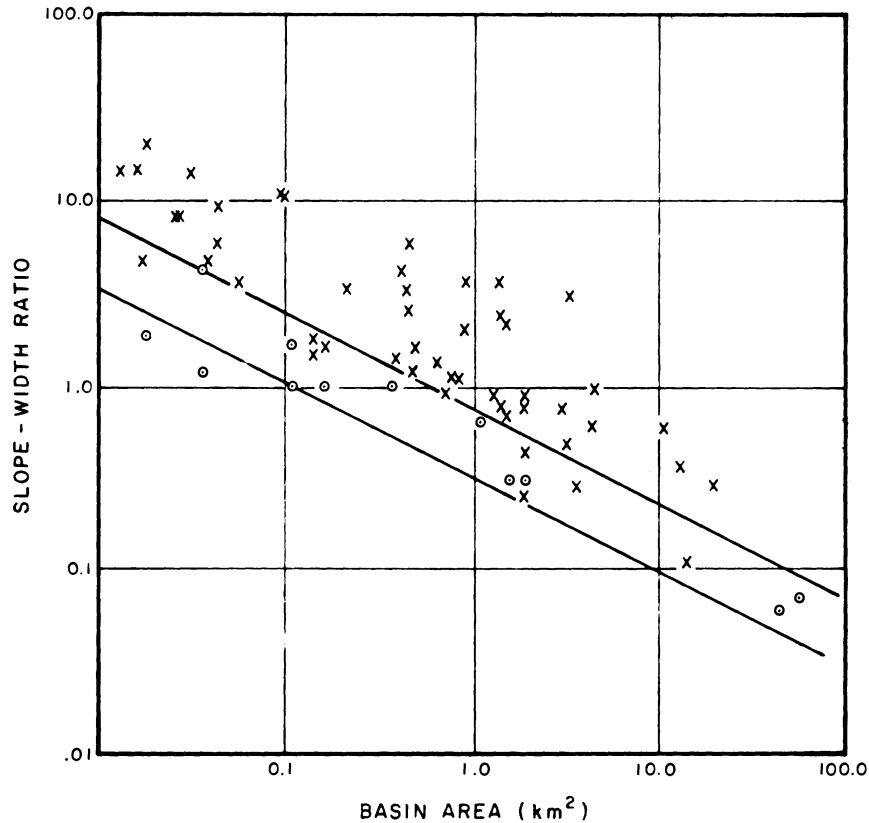
FIGURE 17. Relation between ratio of critical valley slope (S, tan.) and valley floor width (W in m.) to drainage basin area showing threshold zone between gullied and ungullied valley floors, Chalk Bluffs area, northeastern Colorado. *Source:* Bradley, 1979

between gullied and ungullied valleys (Begin and Schumm, 1979). The higher the ungullied valley plots, the greater is the tractive force exerted on the valley floor and the greater is the probability of incision of that valley (Fig. 18).

In addition to the desirability of identifying incipiently unstable valley floors, it is also desirable to determine after failure or gullying if deposition and healing will occur. The concept of complex response assures that this will take place, but in order to accelerate the process geomorphic studies may be able to suggest where in the valley aggradation is most likely. At least for five trenched channels that location is in a reach of little tributary contribution (Schumm, 1961). If further work supports this tentative conclusion steps can be taken to accelerate the healing process at these locations.

River patterns The variability of sinuosity and pattern changes from meandering to braided proved excellent examples of the effect of both intrinsic and extrinsic threshold conditions. Fisk (1944) and Winkley (1970) show that the sinuosity and length of the Mississippi River varies dramatically through time. Sinuosity decreases to a minimum when an avulsion or a series of cut-offs straighten the channel. Such changes can be related to major changes of sediment load or an increase of peak discharge, but they can also be due to a progressive increase of sinuosity with an accompanying reduction of channel gradient to the point where aggradation
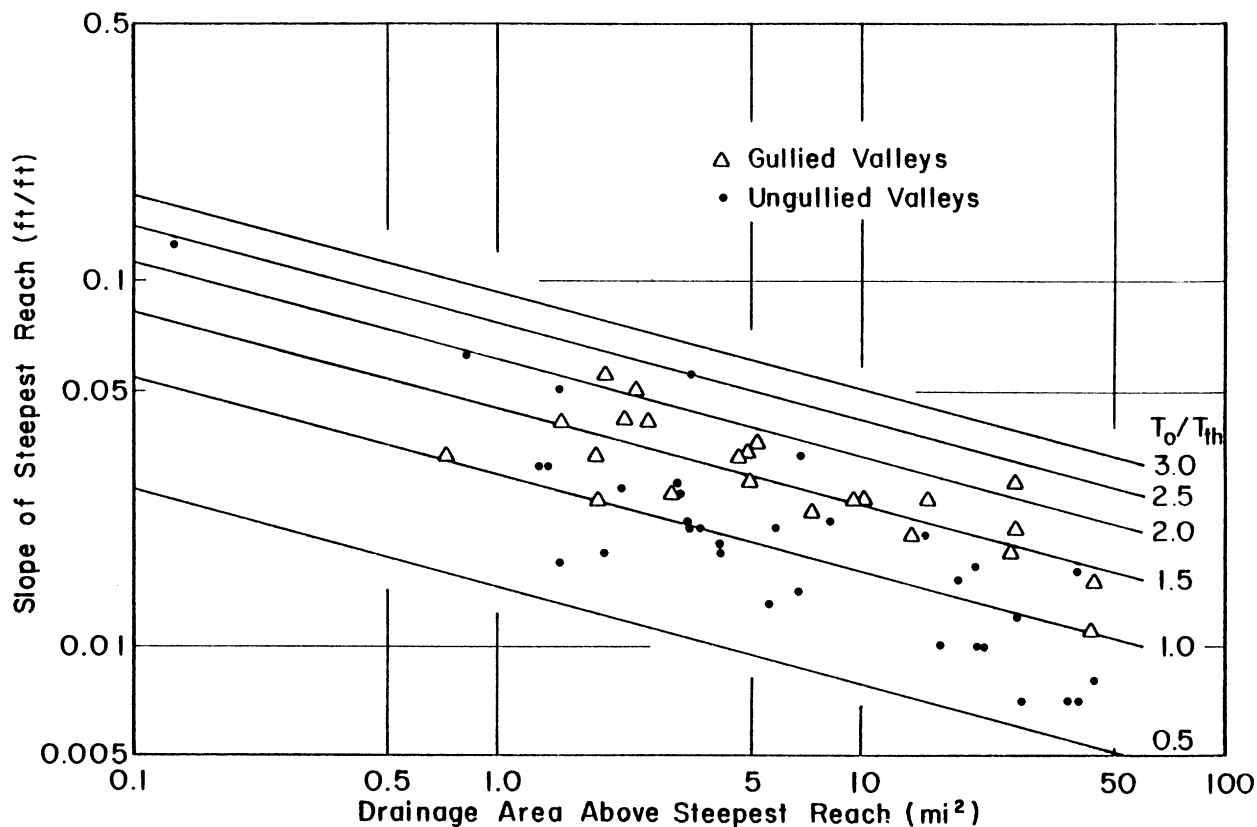
FIGURE 18. Data of Fig. 1 plotted on logarithmic paper. *Th* is threshold shear stress calculated for 10-year flood: by definition no gullies exist below this value. *To* is shear stress calculated for each point. Ratio of *To/Th* reveals that some ungullied valleys in small watersheds have *To* values three times that of its threshold shear stress *(Th)*. Either other conditions are controlling the situation or the probability of gullying is high in these valleys. Note that in the smaller drainage basin a ratio 2·5 can be sustained but for drainage basins greater than 10 sq miles, a ratio of 1·5 is maximum.
Source: Begin and Schumm, 1979

and cut-offs or avulsion result. Such a situation appears to exist along the sinuous parts of the Rio Puerco arroyo, New Mexico, where meander amplitude has increased to the point that in some reaches sediment is being deposited in the upstream limb of each meander, and the bends are being cut off (Schumm, 1977, Fig. 9.3). These changes reflect an intrinsic control by the channel pattern itself.

The work of Lane (1957) and Leopold and Wolman (1957) indicates that there is a gradient or discharge threshold above which rivers tend to be braided (Figs 19 and 20). The experimental work reported by Schumm and Khan (1972) shows that for a given discharge, as valley-floor slope is progressively increased, a straight river becomes sinuous and then eventually braided at high values of stream power and sediment transport (Fig. 4). Rivers that are situated close to the meandering-braided threshold should have a history which is characterized by transitions in morphology from braided to meandering and vice versa.

The suggestion made here is that if one can identify the natural range of patterns along a river, then within that range the most appropriate channel pattern and sinuosity probably can be identified. If so, the engineer can work with the river to produce its most efficient or most stable
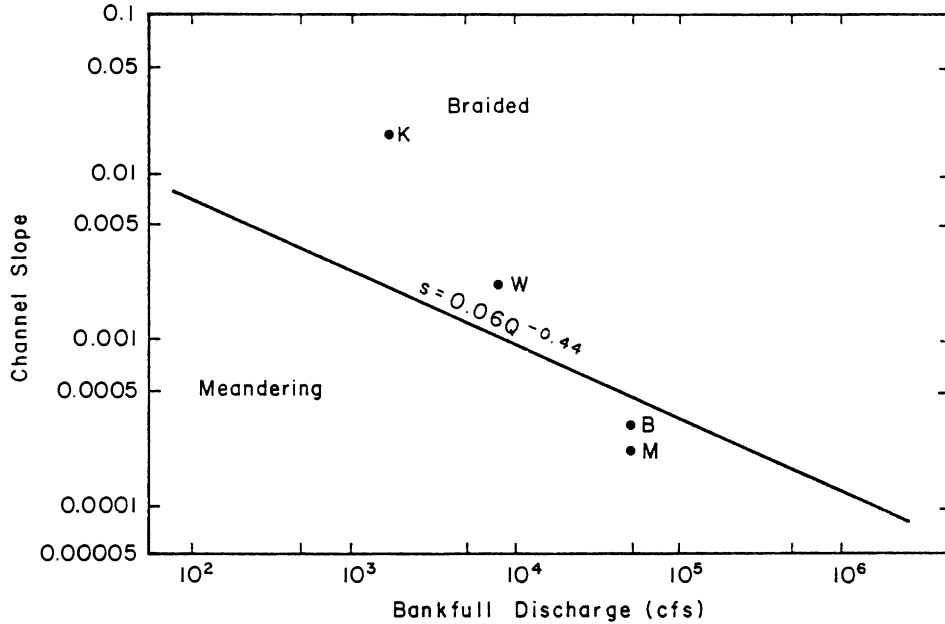
FIGURE 19. Leopold and Wolman's (1957) relation between channel patterns, channel gradient, and bankfull discharge. Letters B and M identify braided and meandering reaches of Chippewa River. Letters K and W refer to Kowhai and Wairau Rivers. *Source:* Schumm and Beathard, 1976
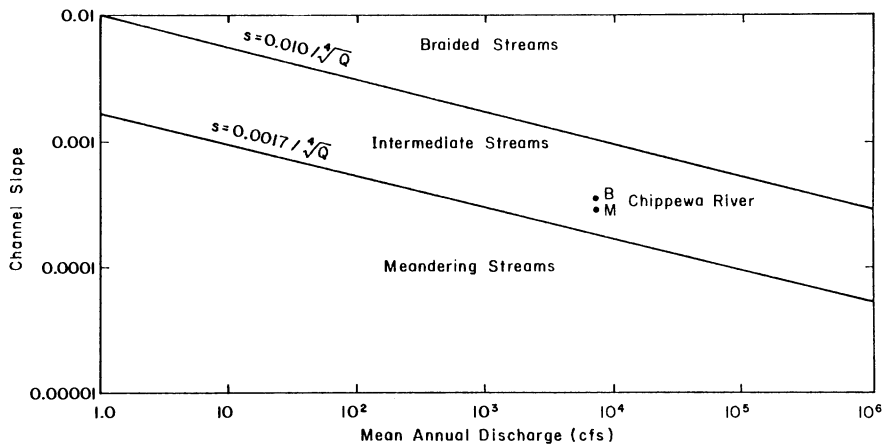

FIGURE 20. Lane's (1957) relation between channel patterns, channel gradient, and mean discharge. The regression lines were fitted to data from streams that Lane classified as highly meandering to braided. Letters B and M identify position of the braided and meandering reaches of the Chippewa River. *Source:* Schumm and Beathard, 1976

channel. Obviously, a river can be forced into a straight configuration or it can be made more sinuous, but there is a limit to the changes that can be induced beyond which the channel cannot function without a radical morphologic adjustment as suggested by Figure 4. Identification of rivers that are near the pattern threshold would be useful, because a braided river near the threshold might be converted to a more stable single-thalweg stream. On the other hand, a

meandering stream near the threshold should be identified in order that steps could be taken to prevent braiding due perhaps to changes of land use.

In most cases it will be difficult to determine if a river is susceptible to the type of treatment discussed in the preceding sections. Perhaps the best qualitative guide to river stability is a comparison of the morphology of numerous reaches, and the determination of whether or not there has been a change in the position and morphology of the channel during the last few centuries. Another approach might be to determine the position of the river on the Leopold-Wolman (1957) or Lane (1957) gradient-discharge graphs (Figs 19 and 20). If a braided river plots among the meandering channels or vice versa, it is a likely candidate for change because it is incipiently unstable.

An example of the way that this could be done is provided by the Chippewa River of Wisconsin (Schumm and Beathard, 1976), a major tributary to the Mississippi River. The Chippewa River rises in northern Wisconsin and flows 320 km to the Mississippi River, entering it 120 km below St Paul. It is the second largest river in Wisconsin, with a drainage basin area of 24 600 km².

From its confluence with the Mississippi to the town of Durand 26·5 km up the valley, the Chippewa is braided (Table I). The main channel is characteristically broad and shallow, and it contains shifting sand bars. The bankfull width as measured from U.S. Geological Survey topographic maps is 333 m. The sinuosity of this reach is very low, being only 1·06. However, in the 68 km reach from Durand to Eau Claire, the Chippewa River has a meandering configuration with a bankfull width of 194 m and a sinuosity of 1·49. The valley slope and channel gradient are different for each reach of the river. The braided section has a gentler valley slope than the meandering reach upstream, 0·00035 as opposed to 0·00040, contrary to what is expected from Figures 4 and 5, but the situation is reversed for channel slope. The braided reach has a channel gradient of 0·00033, whereas the meandering reach has a gradient of 0·00028.

There is no evidence to suggest that the Chippewa river is either progressively eroding or aggrading its channel at present. In fact, the river below Durand has remained braided during historic time. It has maintained its channel position and its pattern, but a significant narrowing as the result of the attachment of islands and the filling of chute channels has occurred downstream of Durand, which resulted in a recent decrease in channel width of over 40 per cent. The only other change in the lower river is a noticeable growth of its delta into the Mississippi valley. The delta deposits show increased vegetational cover as well as progradation into the Mississippi River valley.

TABLE I

Chippewa River morphology

Location	Channel pattern	Channel width (m)	Sinuosity	Valley slope tan. (ft./mile)	Channel slope tan. (ft./mile)
Below Durand	Braided	333	1·06	0·00035 (1·83)	0·00033 (1·76)
Above Durand	Meandering	194	1·49	0·0004 (2·14)	0·00028 (1·46)
Buffalo Slough	Meandering	212	1·28	0·00035 (1·83)	0·00027 (1·43)

The relations described by Leopold and Wolman (1957) and Lane (1957) provide a means of evaluating the relative stability of the modern channel patterns of the Chippewa River. The bankfull discharge was plotted against channel slope on Figure 19 for both the braided and the meandering reaches of the Chippewa. The value used for the bankful discharge is 53 082 cfs, which is the flood discharge having a return period of 2·33 years. The braided reach plots higher than the meandering reach, but both are well within the meandering zone, as defined by Leopold and Wolman. This suggests that the braided reach is anomalous; that is, according to this relation the lower Chippewa would be expected to display a meandering pattern rather than a braided one. Even when the 25-year flood of 98 416 cfs is used, the braided reach still plots within the meandering region of Figure 19.

When the Chippewa data are plotted on Lane's graph (Fig. 20) the same relation exists. The Chippewa falls in the intermediate region, but within the range of scatter about the regression line for meandering streams. Again the braided reach is seen to be anomalous because it should plot much closer to or above the braided stream regression line. The position of the braided reach as plotted on both figures indicates that this reach should be meandering.

This conclusion requires an explanation that can be based on the geomorphic history of Chippewa River. For example, there is an as yet unmentioned significant morphologic feature on the Chippewa River flood-plain. This is Buffalo Slough, which occupies the south-eastern edge of the flood-plain (Fig. 21). It is a sinuous remnant of the Chippewa River that was abandoned, and it is evidence of a major channel change in the Chippewa River valley (Table I).

Flow through Buffalo Slough has decreased during historic time, and indeed flow was completely eliminated in 1876, when the upstream end of Buffalo Slough was permanently blocked. The abandonment of the former Buffalo Slough channel by the Chippewa River is the result of an avulsion, but one that took many years to complete. The channel shifted from Buffalo Slough to a straighter, steeper course along the north-western edge of the flood-plain. This more efficient route gradually captured more and more of the total discharge. The new
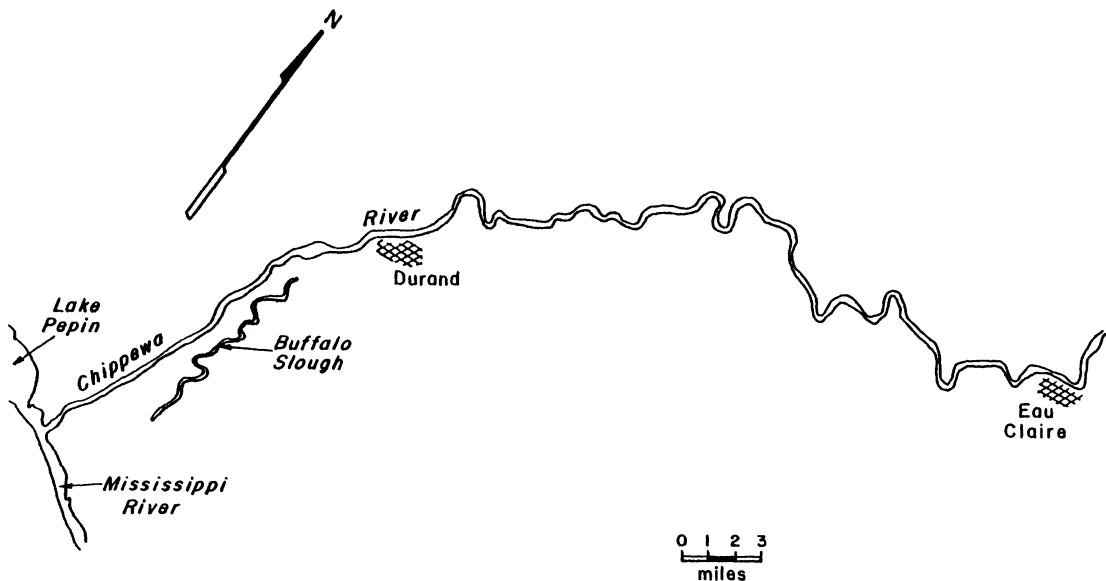

FIGURE 21. Map of lower Chippewa River. The river is braided below Durand, Wisconsin. Buffalo Slough is an old course of Chippewa River. *Source:* Schumm and Beathard, 1976

Chippewa channel produced a divided flow or braided configuration due to a higher flow velocity and the resulting bank erosion. As the new channel grew, the old course deteriorated and eventually its discharge was so reduced that the Mississippi was able to effectively dam the old channel mouth with natural levees and plug its outlet.

The sinuosity of Buffalo Slough is approximately 1·28. Although sinuosity is usually defined as the ratio of stream length to valley length, it is also the ratio of valley slope to channel slope. A sinuosity of 1·28 is, therefore, the ratio of the present valley slope, 0·00035, to a channel slope of about 0·00027. This channel slope value is very similar to the channel slope of the meandering reach upstream of Durand, 0·00028, therefore, the meandering pattern of the Buffalo Slough channel was appropriate. Although delta construction was responsible for the lower valley slope of the lower Chippewa River valley, the Buffalo Slough channel had a gradient that was not appreciably different from that in the upstream reach. Channel sinuosity decreased from 1·49 to 1·28 in a downstream direction, thereby maintaining a channel gradient of about 0·00027.

This sinuous channel could not have transported the large amounts of sediment that the present braided channel carries to the Mississippi River, or it too would have followed a straight braided course. Therefore, the present sediment load carried by the Chippewa River is greater than that conveyed by the Buffalo Slough channel, but this is due almost entirely to the formation of the new straight, steep, braided channel, which is 121 m wider than the old sinuous Buffalo Slough channel. It appears that the lower Chippewa has not been able to adjust as yet to its new position and steeper gradient, and the resulting bed and bank erosion has supplied large amounts of sediment to the Mississippi. The normal configuration of the lower Chippewa is sinuous, and if it could be induced to assume such a pattern, the high sediment delivery from the Chippewa could be controlled. An appropriate means of channel stabilization and sediment load reduction in this case is the development of a sinuous channel.

Since the above suggestions were made (Schumm and Beathard, 1976), more detailed studies of the Chippewa River basin indicate that upstream sediment production must be controlled, especially where the upper Chippewa River is cutting into the Pleistocene outwash terraces. If the contribution of sediment from these sources were reduced, the lower Chippewa could resume its sinuous course.

An indication that the pattern conversion of the Chippewa could be successful if the upstream sediment sources were controlled is provided by the Rangitata River of New Zealand. The Rangitata River is the southernmost of the major rivers which traverse the Canterbury Plains of South Island. It leaves the mountains through a bedrock gorge. Above the gorge, the valley of the Rangitata is braided (Fig. 22A), and it appears that the Rangitata should be a braided stream below the gorge, as are all the other rivers which cross the Canterbury Plain. However, below the gorge, the Rangitata is meandering. A few miles farther downstream, the river cuts into high Pleistocene outwash terraces, and it abruptly converts from a meandering to a braided stream (Fig. 22C). The braided pattern persists to the sea. If the Rangitata could be isolated from the gravel terraces, it probably could be converted to a single-thalweg sinuous channel, because the Rangitata is obviously a river near the pattern threshold.

Other New Zealand rivers are near the pattern threshold and therefore, they are susceptible to pattern change. In fact, New Zealand engineers are attempting to accomplish this pattern change in order to produce 'single-thread' channels which will cause less flood damage and be less likely to acquire large sediment loads from their banks and terraces. For example the engineers of the Marlborough Catchment Board in Blenheim, New Zealand have had success in converting the Wairau River, a major braided stream, from its uncontrolled braided mode to that of a slightly sinuous, single-thalweg, relatively more stable channel (Pascoe, 1976). The
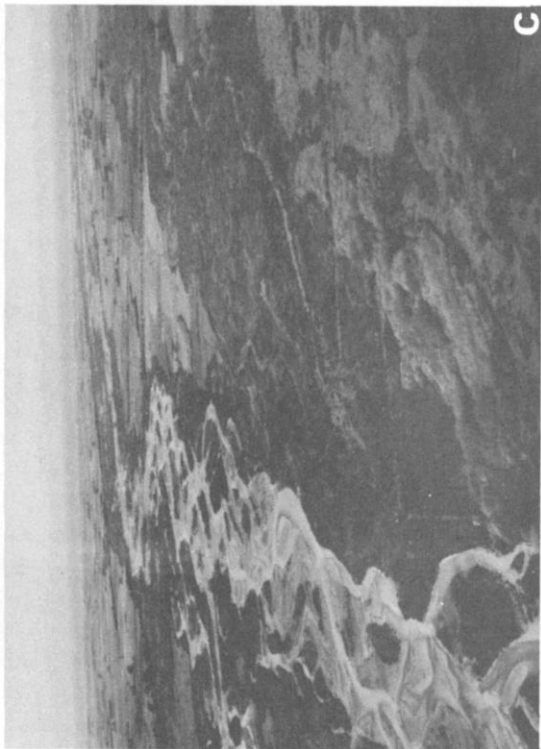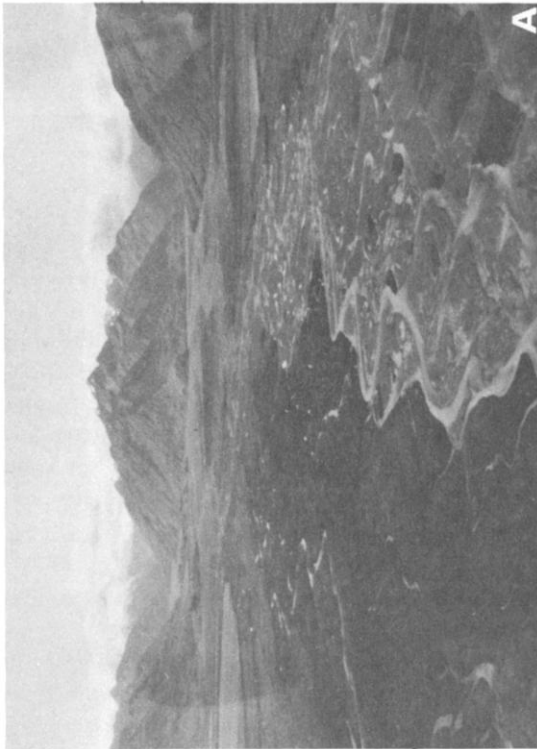
FIGURE 22. Rangitata River, New Zealand.
  A. Above the gorge.
  B. Meandering reach below gorge.
  C. Braided reach below reach of terrace erosion

increase in sinuosity is only from 1·0 to 1·05, and this was accomplished by the construction of curved training banks. On Figure 19 the Wairau River plots close to the threshold line, and with the reduction of sediment load produced by bank stabilization it appears that the pattern threshold can be crossed successfully.

Farther to the south near Kaikoura, the Kowhai River is being modified in the same manner as was the Wairau (Thomson and MacArthur, 1968). Whereas much of the sediment load in the Wairau River is derived from bank and terrace erosion which can be controlled, high sediment loads are delivered to the Kowhai River directly from steep and unstable mountain slopes. On Figure 19 the Kowhai River plots well above the threshold line, and without a major reduction in upstream sediment, it may be difficult to maintain a single-thalweg channel at this location.

The variability of the Rangitata River pattern indicates that braided to single-thalweg conversions should be possible for the Chippewa and Wairau Rivers. However, not all braided rivers can be so readily modified, as this depends on their position with regard to the line defining the pattern threshold on Figure 19.

SUMMARY AND CONCLUSIONS

The concepts advanced here provide, it is hoped, a basis for truly predictive and applied geomorphology. The fact that, at least locally, geomorphic thresholds of instability can be defined quantitatively suggests that they can be identified elsewhere and then used as a basis for recognition of potentially unstable landforms in the field. This approach provides a basis for preventive erosion control. Using geomorphic principles the land manager can spend his limited funds in order to prevent erosion rather than spending it in a piecemeal fashion to attempt to restore seriously eroding areas to their natural conditions.

In addition to the applications described previously, the complexity of erosional behaviour and the variability of sediment production will be of considerable interest to sedimentologists (Schumm, 1977). If the papers appearing in a recent volume on fluvial sedimentology (Miall, 1978) are any indication, the concept of geomorphic thresholds is a welcome addition to the sedimentologist's bag of tricks (Heward, 1978; Minter, 1978, p. 805; Nami and Leeder, 1978, p. 439).

Newtonian physical principles are utilized by engineers to control the landscape and by geomorphologists to attempt an explanation for the inception, evolution, and character of geomorphic situations, but their predictive power is reduced by the complexity of the field situation. For example, an increase in gravitational forces would probably not everywhere cause an equal acceleration in erosional rates. That is, increasing stress may not produce commensurate strain, but local failures will occur. Thus, the application of stress over time will not everywhere have the same result especially as the system to which stress is applied is itself changing through time. The logical consequence of the above situation, as outlined in this paper, is that high magnitude events will not everywhere produce dramatic erosional events; rather the result depends on the character of the geomorphic system.

Geomorphic thresholds can be of two types, intrinsic or extrinsic, depending on whether the crossing of the threshold is due to influences that are internal or external to the landscape. Fan-head trenching, arroyo cutting, gully development, stream capture, river pattern change, meander cut-offs can be due to both, and it is the geomorphologist's responsibility to identify each type and to advise the engineer and land manager how to work with, rather than against, the natural forces that modify landforms. For example, if a river is considered as a dynamic entity with a history, then its relative stability and propensity for change can be more fully comprehended. In many cases, channel instability may, in fact, be a natural and predictable

adjustment to geomorphic controls. This perspective can be expected to have potential for the better understanding of rivers and their management.

Erosion or deposition is not always progressive and the crossing of thresholds in high relief situations may lead to episodic erosion or deposition, which can significantly affect sediment yields and the correlation of pediments, terraces, and alluvial deposits.

As demonstrated, threshold and episodic erosion concepts have practical implications. They can be used to explain anomalous erosion features and to predict future erosional and deposition changes that will occur, either as a result of man's activities or as a normal part of the erosional and depositional evolution of landscape. They also lead to a modified version of the geomorphic cycle of erosion (Fig. 13) which stresses dynamic metastable equilibrium at least in the early stages of the cycle (Fig. 14).

NOTES

1. At the January, 1979 annual meeting of the Institute of British Geographers I presented the guest lecture on the topic of geomorphic thresholds. The presentation was essentially a review of past research by me and by my students, the results of which have been published in the proceedings of three geomorphology conferences held at the State University of New York, Binghamton (Schumm, 1973, 1975, 1979) and of an American Society of Civil Engineers Symposium on inland waterways held at Colorado State University (Schumm and Beathard, 1976). In addition, the results and their implications have been summarized in a book (Schumm, 1977). Nevertheless, both Professor Eric Brown and Dr David Stoddart invited me to prepare my remarks for publication in the *Transactions* on the grounds that the ideas had not received wide dissemination among European geomorphologists. I have borrowed liberally from the previous publications in order to do this. Those that are familiar with my work will find this paper redundant, others, I hope, will find it of interest and of use.

2. The ideas presented here were developed during several research projects supported by the U.S. Army Research Office, National Science Foundation and Colorado Agriculture Experiment Station. I thank Michael Harvey, New Zealand Ministry of Works, for his helpful review.

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