CHAPTER 6

Fundamentals of Fluvial Geomorphology

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Geomorphology is the study of landforms and the processes responsible for making and modifying them. Fluvial geomorphology is the study of landforms whose genesis and evolution are affected by flowing water. A river or stream constitutes a geomorphic system and in working on a natural watercourse the complete system must be considered because, even though a project may directly involve only a small portion of the system, it has the potential to trigger morphological responses in any part of the system. It is impossible to predict the types and locations of morphological responses without a good understanding of the fluvial system, and this demands thorough knowledge of the water and sediment regimes of the river, because the water discharge and associated sediment load drive morphological processes in the system. Water and sediment regimes in rivers derive from the natural climatic, geologic, topographic, and biologic characteristics of the watershed, together with land use and water resource management effects in developed watersheds. It is these watershed attributes and activities that control runoff and sediment sources, the magnitude and distribution of flows, the caliber and type of sediment, and the manner in which water and sediment are supplied to the channel network. In turn, interaction of the flow and sediment load with the materials forming the bed and banks of the channel dictate the three-dimensional morphology of the alluvial channel and its propensity for stability or change (Schumm 1977).

The purpose of this chapter is to present an overview of some basic concepts of fluvial geomorphology and river mechanics, with an emphasis on their application to engineering design of channel rehabilitation projects. In this chapter, "channel rehabilitation" is used in a broad sense that encompasses all aspects of channel modification to achieve a desired channel improvement, whether for river restoration, flood control, navigation, water supply, channel stability, sediment control, or other beneficial use. Regardless of the goals of the rehabilitation project, sound understanding of geomorphic processes and forms in fluvial systems is essential to successful performance of channel rehabilitation projects.

6.1 BASIC CONCEPTS

The six fundamental concepts that should be considered in designing engineering works in rivers and watersheds are:

1. The channel in the project reach is only part of the broader fluvial system;
2. The fluvial system is dynamic;
3. The fluvial system behaves with complexity;
4. Adjustment and response in the system are nonlinear, and abrupt changes can be triggered by relatively small external perturbations or the crossing of geomorphic thresholds;
5. System evolution and response are time-scale-dependent and engineering geomorphic analyses must include a historical perspective; and
6. System evolution and response are space-scale-dependent and the physical size of the system or subsystem must be considered in engineering geomorphic analyses.

6.1.1 The Fluvial System

Schumm (1977) provides an idealized sketch of a fluvial system (Fig. 6-1). Zone 1 is the upper portion of the system, that is, the watershed or drainage basin; this portion of the system functions as the zone of sediment supply. Zone 2 is the middle portion of the system, that is, the river; this portion of the system functions as the sediment exchange and transfer zone. Zone 3 is the lower portion of the system, which may be an estuary, delta, lake, floodplain, wetland, or reservoir; this portion of the system functions as the zone of sediment deposition. These three zones are idealized, because in real

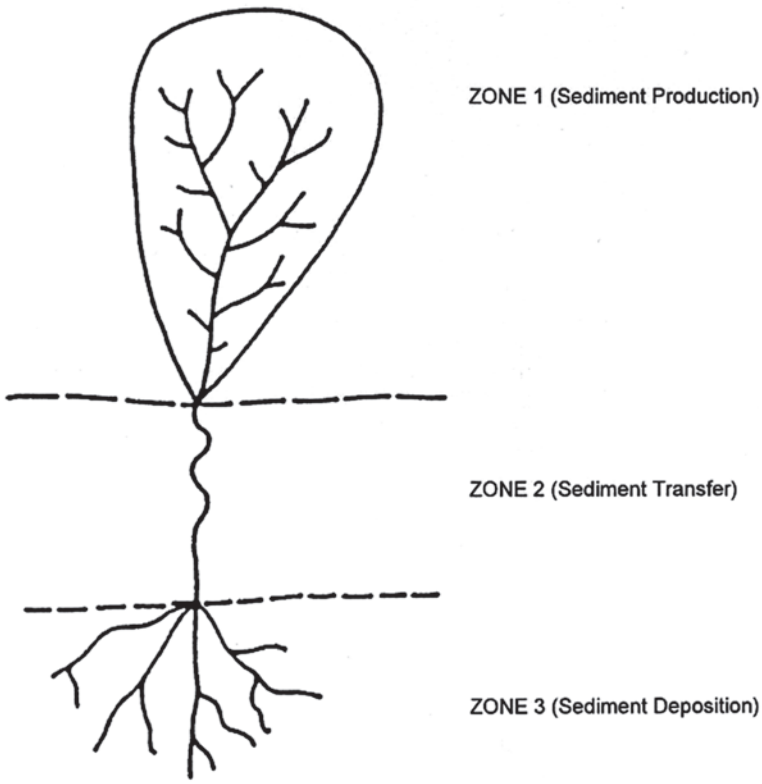


Fig. 6-1. The fluvial system (Schumm 1977, with permission from S. Schumm).

systems sediments can be eroded, transported, and stored in any of the zones. However, within each zone one process is usually dominant, and Schumm's idealized schematization illustrates graphically how sediment processes throughout the system are connected.

In planning any type of engineering alteration to a stream, the potential impacts of disrupting or breaking sediment connectivity in the fluvial system must be considered. For instance, if a channel rehabilitation project is planned for a specific reach of stream in Zone 2, the design engineer must ensure, from a system viewpoint, that the scheme does not interfere unduly with the transfer of sediment from the source zone upstream (Zone 1) to the storage zone downstream (Zone 3).

The fundamental concept that a stream is part of a larger, complex system was eloquently encapsulated by Hans Albert Einstein (1972):

> If we change a river we usually do some good somewhere and "good" in quotation marks. That means we achieve some kind of a result that we are aiming at but sometimes forget that the same change, which we are introducing, may have widespread influences somewhere else. I think if, out of today's emphasis of the environment, anything results for us it is that it emphasizes the fact that we must look at a river or a drainage basin or whatever we are talking about as a big unit with many facets. We should not concentrate only on a little piece of that river unless we have some good reason to decide that we can do that.

6.1.2 The System Is Dynamic

Fluvial processes in each of Schumm's idealized zones are dominated by activity. Zone 1 is the sediment source zone, implying that erosion dominates, driving channel change through net incision or valley widening. Zone 2 is the exchange and transfer zone, implying that as runoff and sediment yield from the watershed increase, the transport capacity of the stream is able to keep pace, exchanging sediment between transport and storage while maintaining dynamic equilibrium in channel form and reworking the floodplain. Zone 3 is the zone of sediment accumulation, implying that deposition dominates, with channel change and long-term storage increasing as sediment accrues in this zone. The functioning of each zone indicates that the system is dynamic and that change in the fluvial system is not only natural, but also essential to its operation.

From an engineering viewpoint the impacts of these dynamics and channel changes may be very significant. For example, loss of 100 ft of stream bank due to channel migration may endanger a home or destroy valuable agricultural land. From a geomorphic viewpoint, channel migration is to be expected and channel shifting represents a natural manifestation of the fluvial system. Indeed, it may not even signal a departure from conditions of natural, dynamic equilibrium. In planning channel rehabilitation measures, engineers must realize that when faced with having to work on a dynamic fluvial system we must try to understand the fluvial system and avoid disrupting it unduly while we are accomplishing our design task. Where disruption, for example, through perturbing the balance between sediment supply and transport capacity, is inevitable, we must predict the system response and take appropriate steps to prevent or at least mitigate adverse responses.

6.1.3 Complexity

Landscape changes are usually complex (Schumm and Parker 1973). The stream and its watershed are a landscape system; change to one portion of the system may result in complex changes, both locally and throughout the remainder of the system.

During complex response, the system responds through the activation of different processes at different locations and times in response to one triggering event or intervention. Consequently, when a fluvial system is subjected to an engineering intervention, changes should be expected to occur throughout the system and over a prolonged period. For example, channelization of a reach of the stream usually accelerates stream velocities, disrupting the sediment-transfer system by increasing sediment transport capacity and allowing the stream to carry away more sediment than is being supplied from upstream. This sediment imbalance results in bed erosion that can migrate upstream through the headcutting process and in increased sediment output that can migrate downstream as a wave of deposition. Through time, headcutting migrates upstream, increasing sediment supply to the channelized reach and eventually causing aggradation there. Thus, in response to a single external intervention, channelization, the affected reach can experience an initial degradational response followed by a secondary aggradational response. This type of complex response not only is theoretically possible, but also has been observed in nature. For example, several Yazoo Basin streams in north Mississippi that were channelized in the 1960s responded initially through degradation, but later exhibited aggradation (Harvey and Watson 1986; Watson et al. 1997). Over the 40 years since the initial perturbation, repeated waves of degradation, temporary stability, and aggradation have occurred, but dynamic equilibrium has still not yet been reestablished.

6.1.4 Thresholds

Rivers and watersheds are described theoretically as nonlinear, complex systems (Richards and Lane 1997) in that they display discontinuous responses to progressive and incremental change in control variables. In the context of fluvial geomorphology, threshold behavior is characterized by progressive change in one variable that eventually results in abrupt change in the system. In engineering terms, the crossing of a geomorphic threshold may be evidenced either by an abrupt change in the rate, direction, or type of change in a naturally evolving fluvial system, or by a disproportionately strong response to a perturbation by an engineering intervention. Bank collapse due to channel incision has been cited as an example of threshold behavior (Thorne and Osman 1988) and may be used to illustrate the phenomenon and related consequences. As an alluvial river accumulates sediment on its bed, morphological evolution occurs through progressive channel aggradation. As aggradation continues, the channel slope gradually increases until, eventually, a limiting condition for bed slope with respect to sediment transport is reached. At this moment the trend of morphological evolution switches from aggradation to degradation as a geomorphic threshold (critical channel slope) is crossed.

Schumm (1973) argued that drainage basins exhibit both *extrinsic* thresholds and *intrinsic* thresholds. In the preceding example, channel change was driven by gradual accumulation of sediment on the bed, which could occur as part of sediment storage in Zone 3 in the natural system. This would be characteristic of an intrinsic threshold. Extrinsic thresholds are crossed when the system is perturbed by an external factor that triggers a disproportionate morphological response. The design engineer must be aware of the existing geomorphic thresholds, the possibility that a natural system may be close to an intrinsic threshold, and the widespread adverse effect that an ill-planned channel stabilization project may have if it causes the system to cross a threshold.

Alluvial channels have a measure of resilience that enables perturbations and imposed changes to be absorbed by morphological adjustments without widespread disequilibrium in the system. Greater resilience implies that the system is a greater distance from a geomorphic threshold than a less resilient system. Systems of greater resilience are less sensitive to change, and those of low resilience are highly sensitive to perturbation.

Threshold theory is often expressed in terms of apparently simple examples, such as the transition between meandering and braiding. This is often quoted as representing a geomorphic threshold, even though Leopold and Wolman recognized as long ago as 1957 that there is actually a continuum of planforms (Leopold and Wolman 1957). Bledsoe (1999) demonstrated that whereas a meandering stream may respond to an increase in bed-material mobility by braiding, it may also respond by incising. In fact, for a given sediment size (D_x), increasing energy (expressed as a mobility index) can result in either a braided or an incised channel, depending on the relative erosion resistance of the bed and bank materials. Also, the threshold mobility index is not single-valued, but is better characterized by a stochastically determined range of values (Fig. 6-2). These findings illustrate that in practice, the geomorphic threshold behavior of alluvial streams may be complex.

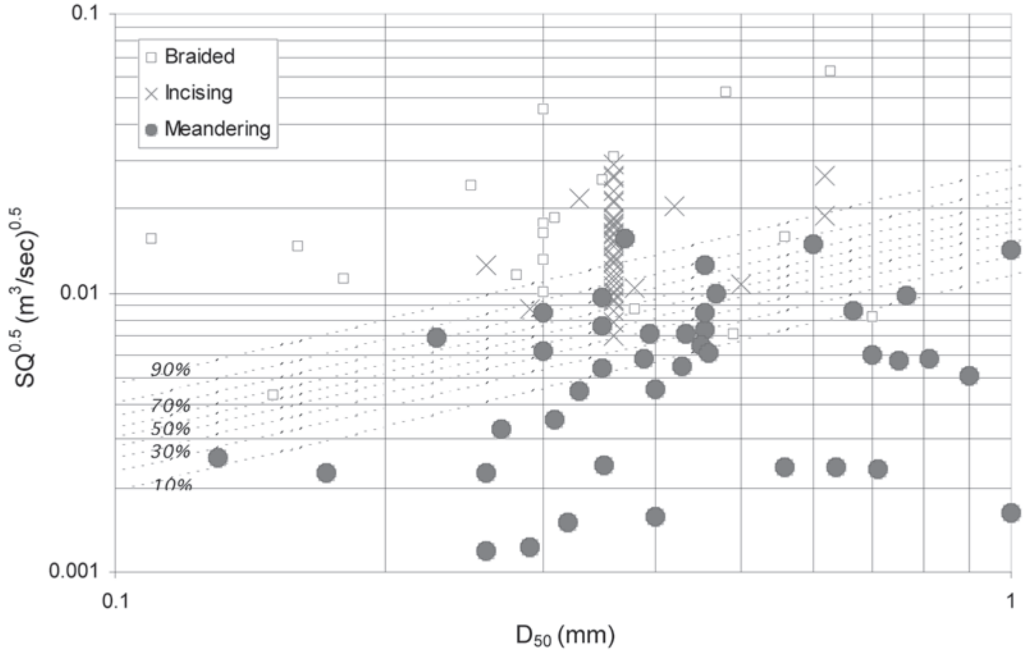


Fig. 6-2. Probability (%) of incising or braiding (dashed lines) is shown as a function of SQ (vertical axis) and D_{50} (horizontal axis) for sand beds. Discharge is represented by annual flood as first priority and then by bank-full flow (reprinted from Bledsoe and Watson 2001, with permission from Elsevier).

6.1.5 Time

Geomorphologists usually refer to three time scales when analyzing rivers:

1. Geologic time,
2. Modern time, and
3. Present time.

Geologic time is expressed in thousands or millions of years, and in this time frame the river is affected by major geologic and climatic changes such as formation of mountain ranges, changes in sea level, and climate change. Equilibrium is not possible over geologic time because, inevitably, the system evolves as material is washed from the mountains to the plains and responds to external changes. The modern time scale describes a period of tens of years to several hundred years, and has also been called the graded time scale (Schumm and Lichty 1965). During this period, a river may adjust to a balanced condition; that is, it may be fully adjusted to prevailing watershed water and sediment regimes and largely retain the same form as it operates in dynamic equilibrium. Present time is considered to be an even shorter period, perhaps 1 year to 10 years. Within this very short time frame, equilibrium may be static—that is, change in the system may be insignificant. Although the duration of these time scales is suggested, no fixed rules govern these definitions. The design of a major project may require less than 10 years, and numerous minor projects are designed and built within the limited scope of observations made during present time. However, project life often extends into graded time, when static equilibrium cannot be assumed to apply. From a geomorphological viewpoint, engineers build major projects in an instant of time, and base their design on an instantaneous snapshot of the river, but still expect these projects to operate successfully and last for a significant period in a dynamically changing system.

Recognition of the importance of time is especially important in considering the postconstruction performance of a project. Society demands a quick return on its investments and projects are expected to produce positive results almost instantaneously. Often, success or failure of a project is judged within one or two years, regardless of whether formative events have occurred to drive geomorphic recovery from construction impacts, or design events have occurred to test whether the project works as intended. With respect to the morphological impacts of a river engineering project, it must be remembered that short-term channel stability or adjustment is not necessarily indicative of long-term behavior. For this reason, the morphological performance of channel projects should be monitored and appraised over a period longer than a few years before a project is declared to have been successful.

6.1.6 Scale

The size or scale of the fluvial system has a bearing on the way in which it evolves toward a natural equilibrium, adjusts to catchment and climate change, and responds to engineering interventions. The time taken for the system to evolve, adjust, or respond increases with the scale of the system. As a general rule, a small stream will react more rapidly to engineering works than a large stream. For instance, channel adjustments in the Mississippi River are still occurring in response to artificial meander cutoffs constructed in the 1930s, and it may require over 100 years before morphological changes triggered by the cutoffs are completed (Biedenharn 1995; Biedenharn and Watson 1997). Conversely, some small bluff line streams in north Mississippi that were channelized in the 1960s have adjusted through initial degradation, secondary aggradation, and dynamic stability within a period of less than 25 years (Watson et al. 2002).

The physical size of a stream also conditions and may limit the type of engineering works that are appropriate and feasible. Although the materials involved in alluvial stream mechanics (basically water, sediment, and vegetation) are scale-independent, the ways that they interact are not. For example, the morphological impact and significance of a large tree on the bank of a small stream is quite different from that of a similar tree on the bank of a large river. From an engineering perspective, it is particularly important to recognize that analyses, techniques, and solutions designed for one scale of stream may not be directly transferable to another. Deciding whether an analytical tool, stabilization technique, or channel enhancement solution developed for streams of a particular size is transferable to streams at other scales demands a thorough understanding of the underpinning science and engineering principles involved. It is not enough to have demonstrated repeatedly that a given approach *works* when applied to streams of a particular scale. Before tools, techniques, or solutions developed for one scale of system are promulgated for wider application, it must be established *how* and *why* they work. Principles, such as stabilizing a retreating bankline by increasing bank erosion resistance and mass stability or by retarding near bank velocities, are transferable across different scales of river; however, the hydraulic models, bank stability analyses, and structural measures appropriate to control bank retreat successfully may not be.

6.2 CHANNEL MORPHOLOGY

Alluvial rivers and streams are dynamic and continuously change position, shape, and other morphological characteristics in response to variations in discharge, sediment load, and boundary conditions. It is therefore important to study not only the existing morphology of the river but also possible variations during the lifetime of the project. The river morphology is determined by the water discharge, quantity and character of the sediment load, characteristics of the bed and bank materials (including vegetation), geologic controls, and valley topography. Morphological changes and adjustments take place in response to variations in any of these parameters through time or human activities. To predict the behavior of a river in a natural state or as affected by human activities, we must understand how fluvial and geotechnical processes operate on the boundary materials to form and adjust the morphological features of the channel through time.

A schematic diagram defining the morphological features associated with straight and meandering channels is shown in Fig. 6-3. The *thalweg* is the trace of the deepest point of the channel. The thalweg and associated line of maximum velocity cross from side to side within the channel, and this pattern of flow affects the overall cross-sectional geometry of the stream. At a bend, there is a concentration of flow in the outer half of the channel due to secondary flow. This causes the scour depth to increase at the outside of the bend, to produce a *pool*. As the thalweg crosses the channel downstream of a bend, the velocity distribution and cross-sectional shape become more symmetrical, and scour depths decrease

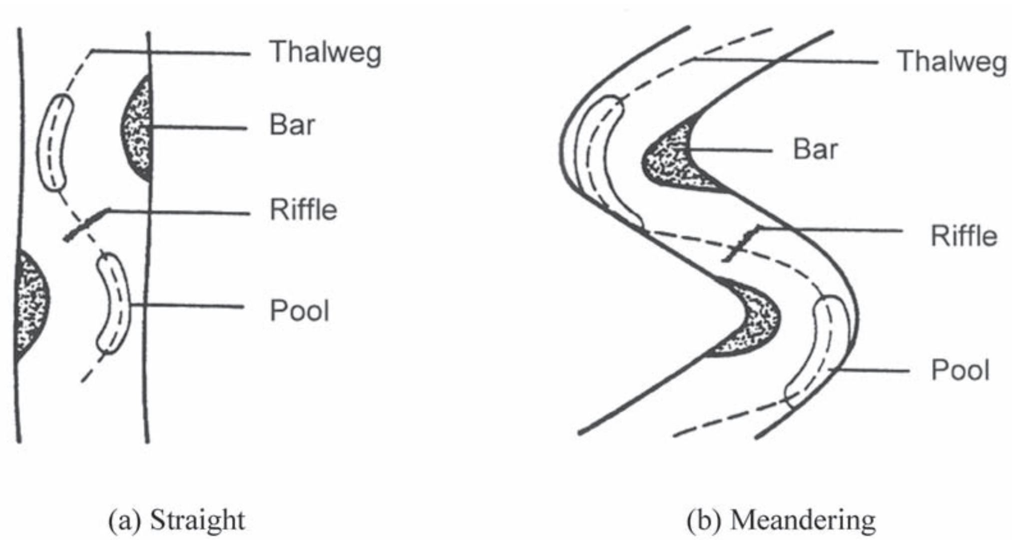


Fig. 6-3. Features associated with (a) straight and (b) meandering rivers.

because of deposition of sediment eroded from the pool upstream. This area is known as the *riffle* or *crossing*.

Pool-riffle sequences are characteristic of cobble, gravel, and mixed load rivers of moderate gradient ($S < 5\%$) (Sear 1996). Riffles are topographic high points in an undulating bed profile and pools are low points. Typically, sediment grain size is coarser on riffles than in pools. A sorting mechanism was proposed by Keller (1971) to explain this variation. According to Keller, fine sediment is removed from riffles during low flows and deposited in pools because velocities and bed shear stresses are higher at riffles (Keller 1971). As discharge rises, velocity and shear stress in the pool increase quickly, with little, if any, increase over the riffle. Consequently, the formative flow velocities and shear stresses in pools are higher than at riffles, resulting in scour of large sediment from the pools and deposition on the next riffle downstream. However, field evidence for this conceptual explanation is equivocal. Ashworth (1987), Petit (1987), and Clifford (1990) have measured the shear stress reversal hypothesized by Keller, but other studies have suggested that pool and riffle velocities equalize at bank-full flow, but do not reverse (Lisle 1979; Carling 1991).

Yalin (1971) suggests that pools and riffles may be explained by macro-turbulent eddies generated at the boundaries of a straight, uniform channel, which produce alternate acceleration and deceleration of flow. Yalin showed theoretically that the longitudinal spacing of faster and slower zones would average πw (w = channel width) for macro-turbulent eddies with diameters similar to the channel width. This is about half the riffle spacing of five to seven times

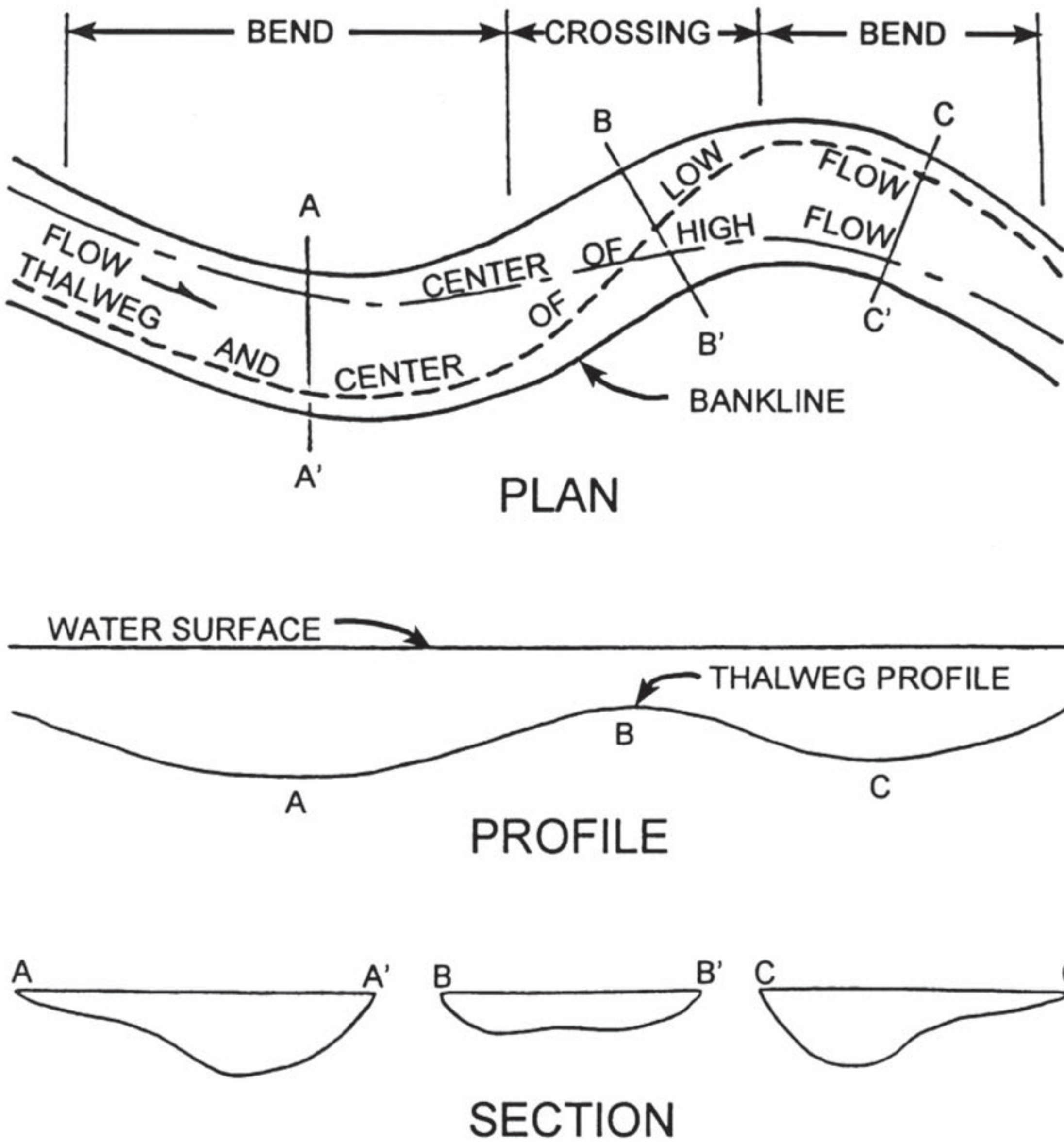


Fig. 6-4. Typical plan, profile, and cross-sectional views of pools and crossings.

the channel width observed in nature (Keller and Melhorn 1973). Hey (1976) proposed a resolution to this difference between theory and observation by proposing that the largest eddies in a stream do not scale on the width, but the half-width, with the centerline of the channel acting as a line of symmetry. According to Hey's hypothesis, riffles would be spaced at $2\pi w$, which better accords with observations.

The cross-sectional shape of a stream varies systematically with distance along the channel in relation to the plan geometry, the type of channel, and the characteristics of the sediment that is formed and transported within the channel. The cross section at a bend is typically deeper at the *concave* (outer bank) side, with a nearly vertical bank, and has a sloping bank formed by the point bar at the *convex* side. The cross section is more trapezoidal or rectangular at a crossing (Fig. 6-4). Cross-sectional dimensions and shape are described by a number of variables. Some of these, such as the *area (A), width (w)*, and *maximum depth (dm)*, are self-explanatory. Other commonly used parameters warrant explanation. *Wetted perimeter (P)* refers to the length of the wetted cross section measured normal to the direction of flow. *Average depth (d)* is calculated by dividing the cross-sectional area by the channel width. *Width-depth ratio (w/d)* is the channel width divided by the average depth. *Hydraulic radius (R)*, which is important in hydraulic computations, is defined as the cross-sectional area divided by the wetted perimeter. In wide channels, with w/d greater than about 20, the hydraulic radius and the mean depth are approximately equal. The *conveyance*, or capacity, of a channel is related to the area and hydraulic radius and is defined as $AR^{2/3}$.

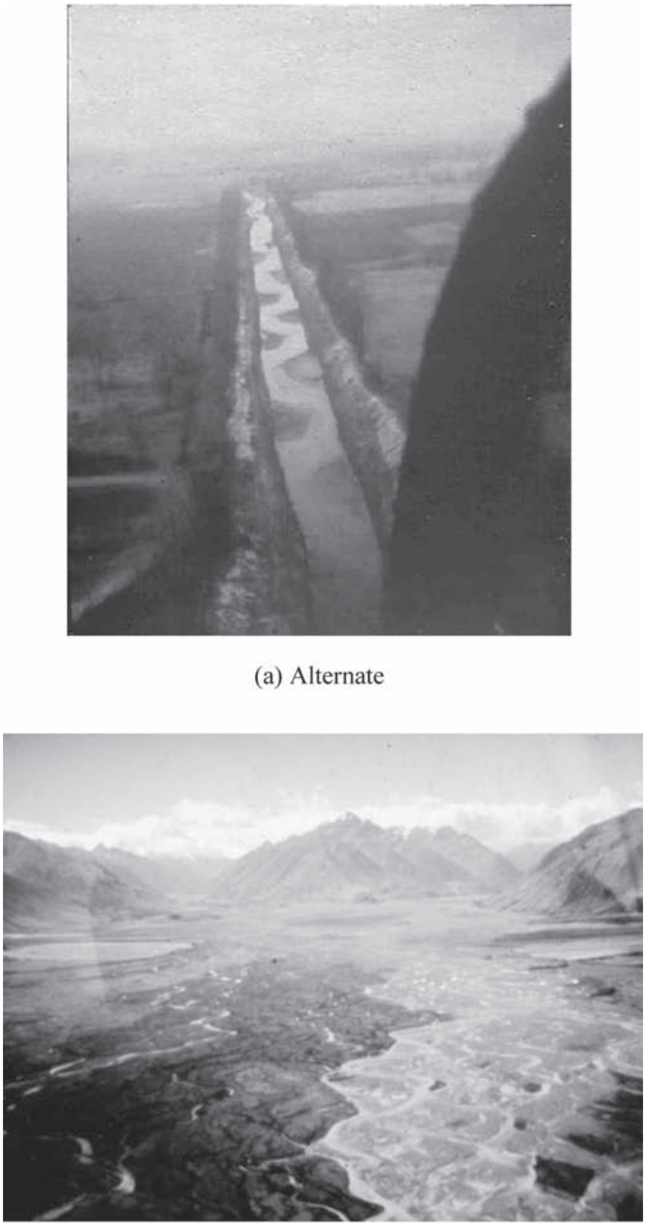
Bars are depositional features that occur within a channel. The types, sizes, frequency of occurrence, and locations of bars are related to the quantity and caliber of the sediment load, local sediment transport capacity, and the morphology of the reach. The most common types of bars are point bars, middle bars, and alternate bars.

Point bars form at the inside (convex) bank of bends in a meandering stream (Fig. 6-5). The size and profile of the point bar are influenced by the characteristics of the flow, the degree of sinuosity, and the quantity and caliber of the sediment deposited at the bend. The development of a point bar is driven by reduction in the sediment transport capacity at the inner bank and sediment sorting due to the action of transverse flows and secondary currents (Dietrich et al. 1984), often coupled with flow separation at the inside of the bend downstream of the apex (Leeder and Bridges 1975). *Middle bar* is the term given to areas of deposition lying within, but not connected to, the banks. Middle bars in meandering rivers may form at riffles, especially where the crossing reaches between consecutive bends are long, and in bends, due to the development of a chute channel that separates part of the point bar from the inner bankline. Figure 6-6 shows a typical middle bar on the Mississippi River formed by this process. *Alternate bars* are regularly spaced depositional features positioned on opposite sides of a straight or slightly sinuous channel (Fig. 6-7a) and may be precursors to meander initiation or braiding. *Braid bars* are sediment features found between the subchannels of multithread, braided rivers (Fig. 7b). Braid bars are highly mobile, and deflection of flow due to bar movement is responsible for the shifting pattern of anabranches and the frequent bank attack that characterize braided river morphology.

Sinuosity (P) is a commonly used parameter to describe the degree of meandering in a stream. Sinuosity is defined as the ratio of distance measured along the channel (channel length) to distance measured along the valley axis (valley length). A perfectly straight channel has a sinuosity of unity, whereas a channel with a sinuosity of 3 or more would have tortuous meanders. *Meander wavelength (L)* is the straight-line repeating distance for the meander waveform, as depicted in Fig. 6-8, and is twice the inflection point spacing. The *meander path length* is the channel length between inflection points. *Meander amplitude (A)* is the width of the meander bends measured perpendicular to the valley or straight-line axis (Fig. 6-8). The ratio of amplitude to meander wavelength is generally within the range from 0.5 to 1.5.

Fig. 6-5. Typical meandering channel with point bars.

Fig. 6-6. Typical middle bar.



(a) Alternate

(b) Braided

Fig. 6-7. Typical bar patterns: (a) alternate, (b) braided.

Meander wavelength and amplitude are primarily dependent on water and sediment discharge, but are usually locally modified by spatial variation in the erodibility of the material in which the channel is formed. The effects of different bank materials are responsible for the irregularities found in the alignment of natural channels. In rare cases where the material forming the banks is practically homogeneous, meanders take a form that may be approximated by a sine-generated curve with a uniform meander wavelength. The *meander belt* is formed by and includes all the locations historically held by a stream due to meander development and migration. It should be noted that the width of the meander belt is usually greater than the meander amplitude and, in many cases, may include all of the active floodplain.

The *radius of curvature* (r_c) is the radius of the circle defining the centerline curvature of an individual bend, measured between the bend entrance and the bend exit (Fig. 6-8). The *arc angle* (θ) is the angle swept out by the radius of curvature. The ratio of radius of curvature to width (r_c/w) is a very useful parameter in the description and comparison of meander behavior and, in particular, bank erosion rates. The radius of curvature is dependent on the same factors as the meander wavelength and width. Meander bends generally develop a radius-of-curvature-to-width ratio (r_c/w) of 1.5 to 4.5, with the majority of bends falling in the range from 2 to 3. Nanson and Hickin (1986) examined the influence of r_c/w on bend migration rate and reported that maximum bank erosion rates occurred when the channel acquired an r_c/w between 2 and 3. This finding has been supported by many empirical studies, for example, Thorne (1991). Plots of erosion rate versus r_c/w do, however, display wide scatter and Biedenharn et al. (1989) showed that part of this scatter could be explained by variations in the erodibility of the outer bank material (Fig. 6-9).

River slope is one of the best indicators of the ability of a river to do morphological work. In general, rivers with steep slopes are much more active with respect to channel changes achieved through sediment movement, bed scour, bar building, and bank erosion. Slope can be defined in a number of

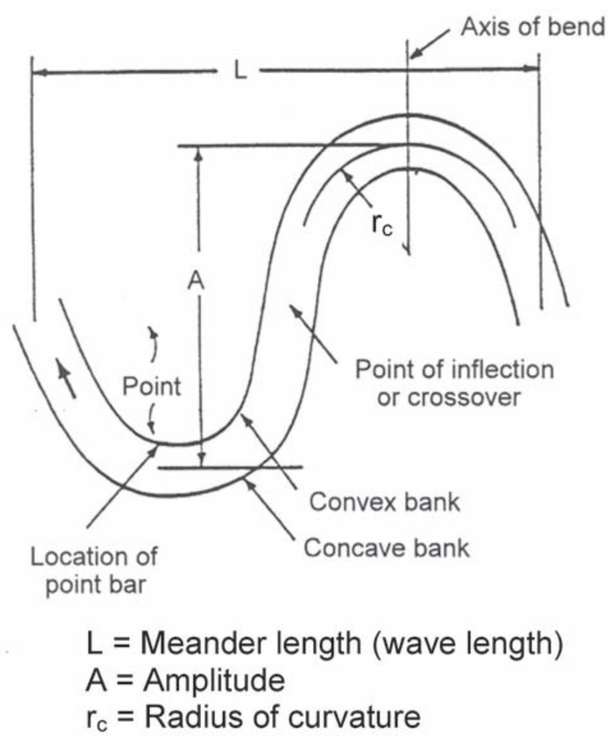


Fig. 6-8. Definition sketch for channel geometry (FISRWG 1998, with permission from the USDA).

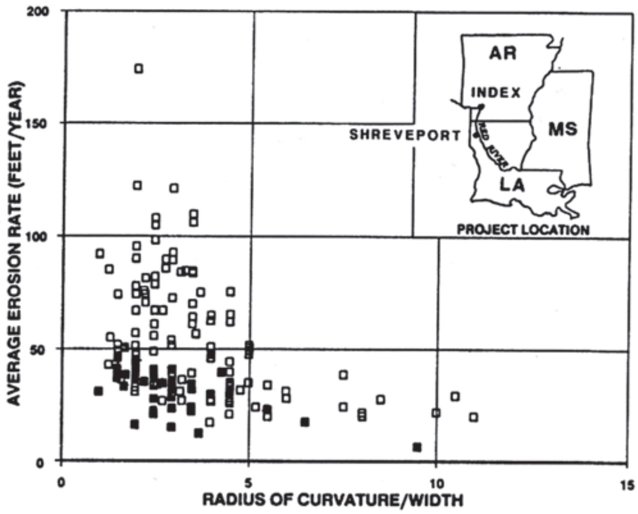


Fig. 6-9. Average annual erosion rate versus *r*/*w* for meander bends of the Red River. Open symbols represent free, alluvial bends and closed symbols, constrained bends (Biedenharn et al. 1989, with permission of ASCE Publications).

ways, however, leading to inconsistency in the way slope is used to represent the ability of a river to do morphological work. Ideally, energy slope should be used to calculate stream power, but the data required are seldom available. In gauged streams, water surface slope may be calculated using stage readings at consecutive gauging stations along the channel. However, many small streams are ungauged. In ungauged streams, thalweg slope is often used to calculate stream power. The thalweg profile not only provides a reasonable basis for calculation of stream power, but also may aid in locating bed controls due to geologic outcrops, other nonerodible materials, or inputs of relatively immobile sediments from steep tributaries. Repeat thalweg profiles are particularly useful in identifying bed-level adjustments through aggradation, degradation, local scour, and fill. When different slopes are used to calculate stream power, it must be kept in mind that the thalweg, water surface, and energy slopes are not necessarily equal.

6.3 SEDIMENT TRANSPORT

One aspect of river engineering that causes considerable confusion and misunderstanding is the terminology associated with sediment transport. In discussing the sediment transport, it is important to be familiar with the terminology adopted and the nature of the load being discussed. Over an extended period, a common terminology has emerged, and although it is not universally agreed upon or applied, it provides the basis for at least reducing inconsistency.

Total sediment load is the mass of granular sediment transported by a stream. It can be broken down by source, transport mechanism, or measurement status (Table 6-1). *Bed load* is a component of total sediment load made up of particles moving in continuous or frequent contact with the bed. Transport occurs at or near the bed, with the submerged weight of particles supported by solid-solid contact with the bed. Bed load movement takes place by processes of rolling, sliding, and saltation. *Suspended load* is a component of the total sediment load made up of sediment particles moving in continuous or semicontinuous suspension within the water column. Transport occurs above the bed, with the submerged weight of particles supported by anisotropic turbulence within the body of the flowing water. *Bed-material load* is the portion of total sediment load composed of grain sizes that are found in appreciable quantities in the streambed. The bed-material load is the bed load plus the coarser portion of the suspended load, that is, particles of a size that are found in significant quantity in the bed. *Wash load* is the portion of the total sediment load composed of grain sizes finer than those found in appreciable quantities in the streambed. *Measured load* is the portion of total sediment load that is sampled by conventional suspended load samplers. The sediment sampled in

Table 6-1 Classification of the Sediment Load

Measurement method	Transport mechanism	Sediment source
Unmeasured load	*Bed load*	
		Material load
Measured load	*Suspended load*	
		Wash load

deriving the measured load includes a large proportion of the suspended load, but excludes that portion of the suspended load moving very near the bed (that is, below the sample nozzle) and all of the bed load. *Unmeasured load* is that portion of the total sediment load that passes beneath the nozzle of a conventional suspended load sampler, moving in near-bed suspension and as bed load.

6.4 CHANNEL-FORMING DISCHARGE

Morphological studies have revealed that channel form depends on a delicate balance between the flows of water and sediment that shape the channel, the processes by which channel form is changed, and the ability of the boundary materials to resist change. Variability of water and sediment discharges is a characteristic of the watershed and, over a sufficiently long period, the morphology of the channel will adjust to accommodate the range of flow events responsible for regulating the balance between the erosive and resistive forces that mold the channel. Consequently, the shape and dimensions of an alluvial river channel are adjusted to and reflect the wide range of flows that entrain, transport, and deposit boundary sediments (Lane 1955). The concept that there is a single discharge that, if it prevailed all the time, would produce the same width, depth, slope, hydraulic roughness, and planform as those produced by the actual range of discharges is attractive, but viewed in this context it is clearly a gross simplification. The single discharge best able to represent the actual spectrum of sediment-transport events to yield the same bank-full morphology as that shaped by the natural sequence of flows is referred to as the channel-forming flow or the dominant discharge. Dunne and Leopold (1978) define channel maintenance flow as the most effective discharge for moving sediment, forming or removing bars, forming or changing bends and meanders, and generally doing work that results in the average morphological characteristics of channels. Their definition of channel maintenance flow is very similar to the concept of channel-forming discharge.

In a regulated canal system, the dimensions of the channel can appropriately be based on a single design discharge. Empirical analysis of the relationship between that discharge and the dimensions for a stable, unlined canal formed in alluvial materials produced the regime theory. Early work on regime theory stems from design of straight canals in the Indian subcontinent (Inglis 1941; 1947; 1949), and North America (Blench 1952; 1957). Later, flume experiments extended the regime approach to channels with meandering planforms (Ackers and Charlton 1970a; 1970b). However, for widely varying flows emanating from a natural watershed, the problem of identifying the single channel-forming discharge is both challenging and critical.

Soar (2000) recently reviewed the huge literature pertaining to the concept of channel-forming flow. This concept is closely related to the theory of dynamic equilibrium, which is characterized by fluctuations of channel form around an average condition that persists through time. In perennial rivers, recovery of equilibrium following a major event occurs relatively quickly, partly because rapid vegetation growth encourages sedimentation (Hack and Goodlett 1960; Gupta and Fox 1974). Hence, the long-term time-averaged condition is a valid representation of the channel form. Recovery in the ephemeral channels of semiarid regions tends to take longer, reflecting the influence of relatively wet and dry periods on vegetation growth (Schumm and Lichty 1965; Burkham 1972). In arid areas, infrequent floods impart long-lasting imprints on channels because more frequent flows do not have the power to restore a regime condition (Schick 1974). It has been concluded that the channel-forming flow concept may be inapplicable to ephemeral rivers that exhibit highly variable flow regimes, because there may not be a single discharge that can explain channel form (Stevens et al. 1975; Baker 1977). This is because channel morphology is likely to be perpetually in disequilibrium with the prevailing flows rather than fluctuating around an average state.

Channel-forming flow or dominant discharge is actually a geomorphological concept and not strictly a measurable parameter. However, a number of discharges that may be taken to represent the channel-forming flow can be defined and calculated using prescribed methodologies. The first approach is to identify a candidate flow based on channel morphology, such as the bank-full discharge. A second approach is to select a discharge based on a specified recurrence interval discharge, typically between the 1- and 3-year events in the annual maximum series. The third approach is analytical and involves calculating the effective discharge.

6.4.1 Bank-Full Discharge

Based on both theoretical and empirical arguments, bank-full discharge is generally recognized as being the moderate flow that best fits Wolman and Miller's (1960) dominant discharge concept for rivers in dynamic equilibrium. Leopold et al. (1964) proposed that the bank-full discharge was responsible for channel maintenance and form, and therefore that it was equivalent to the channel-forming discharge. Dury (1961) also suggested that the channel-forming discharge is approximately equal to the bank-full discharge and Dunne and Leopold (1978) concluded that their maintenance discharge corresponded to the bank-full stage. Field identification of bank-full discharge is, however, problematic (Williams 1978). It is usually based on identification of the minimum width-to-depth ratio (Wolman 1955; Pickup and Warner 1976), together with the recognition of some discontinuity in the nature of the channel, such as a change in sedimentary or vegetative characteristics. Nixon (1959) defined the bank-full state as the highest flood of a river that can be contained within its channel without spilling water on the river floodplain. Wolman and Leopold (1957) defined the bank-full stage as the elevation of the active floodplain. Woodyer (1968) suggested that bank-full discharge corresponds to the elevation of the middle bench of rivers having several overflow surfaces. Schumm (1960) defined bank-full stage as the height of the lower limit of perennial vegetation, primarily trees. Similarly, Leopold (1994) states that bank-full stage is indicated by a change in vegetation, such as herbs, grasses, and shrubs. Finally, the bank-full stage is also defined as the average elevation of the highest surface of the channel bars (Wolman and Leopold 1957). Harrelson et al. (1994) provide explanations of field methods for determining bank-full discharge using vegetation, gradation of bank materials, and elevation of sedimentary features. Although several criteria have been identified to assist in field identification of bank-full stage, ranging from vegetation boundaries to morphological breaks in bank profiles, considerable experience is required to apply these in practice, especially on rivers that have in the past undergone aggradation or degradation.

6.4.2 Specified Recurrence Interval Discharge

Problems and subjectivity in the field identification of bank-full elevation and discharge make it attractive to use an objectively defined discharge such as a specific recurrence interval flow. This recurrence interval flow can, in turn, be related to the bank-full elevation (Table 6-2). Wolman and Leopold (1957) suggested that the bank-full frequency has a recurrence interval of 1 to 2 years. The most often quoted recurrence interval is 1.5 years. Dury (1973) concluded that the bank-full discharge is approximately 97% of the 1.58-year discharge, or the most probable annual flood. Hey (1975) showed that for three British gravel-bed rivers, the 1.5-year flow in an annual maximum series passed through the scatter of bank-full discharges measured along the course of the rivers. Richards (1982) suggests that, in a partial duration series, bank-full discharge equals the most probable annual flood, which has a 1-year return period. Leopold (1994) concludes that most investigations have found that the recurrence interval for bank-full discharge ranges from 1.0 to 2.5 years. However, there are many instances where the bank-full discharge does not fall within this range. For example, Williams (1978) showed that for 35 floodplains in the United States the recurrence interval of bank-full discharge varied between 1.01 and 32 years, and found that only about one-third of those streams had a bank-full discharge with a recurrence interval between 1 and 5 years. In a similar study, Pickup and Warner (1976) determined that bank-full recurrence intervals ranged from 4 to 10 years on the annual series.

If a specified recurrence interval flow is used to estimate the channel-forming discharge, a range of 1 to 3 years should be used. However, because of the uncertainties discussed above, it is recommended that discharges in this range be compared to the bank-full stage in the field to verify that they do have morphological significance.

Table 6-2 Recommended Frequencies for Bank-Full Discharge (after Soar 2000)

Discharge frequency	Recommended by
1 to 5 years	Wolman and Leopold (1957)
1.5 years	Leopold et al. (1964); Hey (1975); Leopold (1994)
1.58 years	Dury (1973, 1976); Riley (1976)
1.02 to 2.69 years	Woodyer (1968)
1.01 to 32 years	Williams (1978)
1.18 to 3.26 years	Andrews (1980)
1 to 10 years, 2 years	USACE (1994)
2 years	Bray (1973, 1982)

6.4.3 Effective Discharge

The effective discharge is defined as the increment of discharge that transports the largest fraction of the annual sediment load over a period of years (Andrews 1980). The effective discharge incorporates the principle prescribed by Wolman and Miller (1960) that the channel-forming or dominant discharge is a function of both the magnitude of sediment-transporting events and their frequency of occurrence. An advantage of using the effective discharge is that it is a calculated value that integrates the discharge and sediment-transport regimes of the stream.

Equivalence between bank-full and effective discharges for natural alluvial channels that are in regime has been demonstrated for a range of river types (sand, gravel, cobble, and boulder-bed rivers) and in different hydrological environments, if the flow regime is adequately defined and the appropriate component of the sediment load is correctly identified (Andrews 1980; Carling 1988; Hey 1997). However, Benson and Thomas (1966), Pickup and Warner (1976), Webb and Walling (1982), Nolan et al. (1987), and Lyons et al. (1992) report that the effective and bank-full discharges are not always equivalent. This suggests that the effective discharge may not always be a direct surrogate for the channel-forming flow or the bank-full discharge.

Although the effective discharge is straightforward conceptually, and has been used for many years, many engineers have expressed concerns that the effective discharge calculations do not yield reasonable results in some instances. These problems may be attributable to data limitations, insufficient understanding of the morphology of the stream, or improper calculation procedure. To minimize these uncertainties a standardized procedure for the determination of the effective discharge has been developed and is outlined in the following paragraphs. This procedure is intended to help investigators avoid many of the potential problems that the authors have experienced in the calculation of effective discharge. Interested readers are referred to Biedenharn et al. (2000a) for a more detailed discussion of effective discharge calculation.

The method most commonly adopted for determining the effective discharge is to calculate the total load (tons) transported by the range of flows over a period of time by multiplying the frequency of occurrence of selected discharge classes (number of days) by the median magnitude of the sediment load (tons/day) transported by that class of flows. Although this approach has the merit of simplicity, the accuracy of the estimate of the effective discharge is clearly dependent on the calculation procedure adopted. The basic inputs required for calculation of effective discharge are (1) flow-duration data and (2) sediment transport as a function of stream discharge.

The first step in an effective discharge calculation is to group the discharge data into classes and determine the number of events occurring in each class during the period of record. This is usually accomplished from a flow-duration curve, which is a cumulative distribution function of measured discharges. A flow-duration curve shows the percentage of time a specific discharge is equaled or exceeded during the period of record, for which the curve was developed. From the flow-duration curve, the number of days that discharges within the specified class interval occurred can be calculated. The three critical components that must be considered in developing a flow-duration curve are the time base, the number of class intervals, and the period of record.

Conventionally, values of mean daily discharge are used to compute the flow-duration curve. Although this is convenient and uses readily available mean daily flow data that are published by the U.S. Geological Survey (USGS), it can, in some cases, introduce bias into the calculations. Mean daily values underestimate the influence of the high flows that occur within the averaging period and overestimate the significance of the low flows. On large streams such as the Mississippi River, the use of mean daily values is acceptable because differences between mean daily and daily peak discharges are negligible. However, on flashy streams, the time from the flood peak to base flow may be only a few hours, so mean daily flow cannot adequately describe the hydrograph. Missing flood peaks and associated high sediment loads can result in the effective discharge being underestimated. Rivers with a high flashiness index, defined as the ratio of the instantaneous peak flow to the associated daily mean flow, are most affected.

To avoid this problem it may be necessary to increase the temporal density from 24 h (mean daily) to 1 h, or even 15 min, especially on flashy streams. This will ensure that the hydrograph is adequately described, enabling a more representative effective discharge to be determined.

Class intervals should be arithmetic and must be of equal width. It has been demonstrated that the use of logarithmic or non-equal-width arithmetic classes introduces systematic bias into the calculation of effective discharge (Soar 2000; Soar and Thorne 2001). However, interested readers should review Holmquist-Johnson (2002) for guidance in calculating effective discharge for conditions under which equal-width class intervals are not usable. The selection of class interval may influence the calculated effective discharge. There are no definitive rules for selecting the most appropriate interval and number of classes. Yevjevich (1972) stated that the class interval should not be larger than $s/4$, where s is an estimate of the standard deviation of the sample. For hydrological applications he suggested that the number of classes should be between 10 and 25, depending on the sample size. Hey (1997) found that 25 classes with equal, arithmetic intervals produced a relatively continuous flow-frequency distribution and a smooth sediment-load histogram with a well-defined peak, indicating an effective discharge that corresponded exactly with bank-full flow. However, in the authors' experience, 25 classes may not always produce satisfactory results. It is recommended that in difficult cases the number of intervals be increased, but not to the extent that individual classes have zero events or only one event.

The period of record must be sufficiently long to include a wide range of morphologically significant flows, but not so long that changes in the climate, land use, or runoff characteristics of the watershed produce significant changes with time in the data. If the period of record is too short, there is a significant risk that the effective discharge will be inaccurate because of the occurrence of unrepresentative flows. A reasonable minimum period of record for an effective discharge calculation is about 10 years, with 20 years of record providing more certainty that the range of morphologically significant flows is fully represented in the data. Records longer than 30 years should be examined carefully for evidence of temporal changes in flow and/or sediment regimes.

The next step in the determination of the effective discharge is to develop a sediment-rating curve that relates the sediment transport and discharge. The sediment-rating curve can be developed from observed, measured sediment loads or using a computational procedure. Effective discharge is very sensitive to the slope of the sediment-discharge relationship.

The sediment load that is responsible for shaping the channel should be used in the calculation of the effective discharge. The suspended sediment load reported by USGS publications usually includes a portion of the bed-material load and most of the wash load. If measured suspended-sediment data are used for the effective-discharge calculation, then the fine sediment load, consisting of particles not found in appreciable quantities in the bed, should be omitted. If the bed load in the stream is only a small percentage of the total bed-material load, it may be acceptable to use only the measured suspended bed-material load in the effective discharge calculations. However, if the bed load is a significant portion of the load, it should be calculated using an appropriate sediment-transport function and then added to the suspended bed-material load to provide an estimate of the total bed-material load. If bed-load measurements are available, which seldom is the case, observed data may be used.

Once the fines have been removed from the data set, a sediment-rating curve is developed from the concentration data by plotting sediment load (concentration times discharge) against discharge, and then calculating a best-fit regression curve through the data, or, as required in some cases, multiple segments of best-fit regression.

The discharges used to generate the bed-material load histogram are the arithmetic mean discharges in each class of the flow-frequency distribution. The bed-material transport rate for each discharge class is found from the rating curve equation. This load is multiplied by the frequency of occurrence of that discharge class to find the total amount of bed material transported by that discharge class during the period of record. Care should be taken to ensure that the time units in the bed-material load rating equation are consistent with the frequency units for the distribution of flows. The results are plotted as a histogram. The bed-material load histogram should display a continuous distribution with a single mode (peak). If this is the case, the effective discharge corresponds to the mean discharge for the modal class (that is, the peak of the histogram). If the modal class cannot be identified readily, the peak of a smooth curve drawn through the tops of the histogram bars can be used to estimate the effective discharge by interpolation.

6.4.4 Overview

All three approaches to estimating the channel-forming flow or dominant discharge (bank-full estimate, discharge of a selected return period, and effective discharge) present challenges. The selection of the appropriate method will be based on data availability, the physical characteristics of the study stream, the level of study, and time and funding constraints. It is recommended that all three methods be used and the results cross-checked to reduce the uncertainty in the final estimate of the channel-forming flow. If the effective discharge method is used, then it is recommended that the standardized procedure presented here be followed.

6.5 RELATIONSHIPS IN RIVERS

Given the evident complexity of fluvial processes and their interactions with channel morphology, it is perhaps surprising that the characteristic forms adopted by alluvial rivers are limited in number and frequent in occurrence. For example, the planforms of meandering rivers display clear similarity in their proportions. Brice (1984) suggested that the similarity of meanders accounts for the fact that, if scale is ignored, all meandering rivers tend to look alike in plan view. It is the familiar and almost ubiquitous nature of the forms and features displayed by alluvial streams of different sizes, in widely varying landscapes, that makes these complex systems amenable to description by relatively simple empirical relationships. For example, relationships developed by Williams (1986) illustrate how Brice's recognition of the similarity of meanders may be expressed quantitatively through empirical relationships relating the geometric properties of channel meander to one another (Table 6-3).

Similarly, in regime theory the concept that the width, depth, slope, and planform of a river are adjusted to a channel-forming discharge is expressed numerically in simple power-law equations. The *Stream Corridor Restoration Manual* (FISRWG 1998) provides the selected summary of regime equations reproduced in Table 6-4.

Independent of regime theory, Leopold and Maddock (1953) compiled important statistical equations linking various channel dimensions to discharge using USGS gauging records. These equations, termed *hydraulic geometry relationships*, describe how width, depth, velocity, and other hydraulic characteristics vary both with stage at a station and with changing bank-full discharge downstream for some streams in the United States. The hydraulic geometry relationships are of the same general form as the regime equations of Kennedy (1895):

$$W = \mathrm{a}\ Q^{\mathrm{b}}$$
$$D = \mathrm{c}\ Q^{\mathrm{f}}$$
$$V = \mathrm{k}\ Q^{\mathrm{m}}$$

where W = channel width, Q = discharge, D = depth, and V = velocity. Later versions of these hydraulic geometry relationships (listed in Table 6-5) add the median bed sediment size (D_{50}) to improve the predictive power of the equations, and appear in the following format:

$$W = k_1\ Q^{k2}\ D_{50}^{\ k3}$$
$$D = k_4\ Q^{k5}\ D_{50}^{\ k6}$$
$$S = k_7\ Q^{k8}\ D_{50}^{\ k9}$$

The relationships presented here are only a small sample of those available in the literature. Regime relationships are empirical, which means that the relationships are derived from observed physical correlations and are strictly only applicable to the data sets from which they were derived. In this regard, Rinaldi and Johnson (1997) are correct to point out the inappropriateness of using simple regression equations in the design of meander restorations when fluvial processes and channel morphology in the project stream differ manifestly from conditions in the rivers used to develop the equations. In practice, hydraulic geometry and other empirical relationships may be widely and usefully applied, provided that conditions in the study watershed are similar to those in the watersheds for which the equations were developed. However, even under ideal conditions these equations remain incomplete representations of the factors that actually

Table 6-3 Derived Empirical Equations for River Meander and Channel Size (FISRWG 1998, with permission from USDA)

Equation number	Equation	Applicable range (meters)
	Interrelations between meander features	
2	$L_m = 1.25\ L_b$	$5.49 < L_b < 13{,}293$
3	$L_m = 1.63\ B$	$3.69 < B < 13{,}689$
4	$L_m = 4.53\ R_c$	$2.59 < R_c < 3{,}598$
5	$L_b = 0.8\ L_m$	$7.93 < L_m < 16{,}494$
6	$L_b = 1.29\ B$	$3.69 < B < 10{,}000$
7	$L_b = 3.77\ R_c$	$2.59 < R_c < 3{,}598$
8	$B = 0.61\ L_m$	$7.93 < L_m < 23{,}201$
9	$B = 0.78\ L_b$	$5.49 < L_b < 13{,}293$
10	$B = 2.88\ R_c$	$2.59 < R_c < 3{,}598$
11	$R_c = 0.22\ L_m$	$10.06 < L_m < 16{,}494$
12	$R_c = 0.26\ L_b$	$6.80 < L_b < 13{,}293$
13	$R_c = 0.35\ B$	$4.88 < B < 10{,}000$
	Relations of channel size to meander features	
14	$A = 0.0054\ L_m^{1.53}$	$10.06 < L_m < 23{,}201$
15	$A = 0.0085\ L_d^{1.53}$	$6.10 < L_d < 13{,}293$
16	$A = 0.0103\ B^{1.53}$	$4.88 < B < 11{,}616$
17	$A = 0.0669\ R_c^{1.53}$	$2.13 < R_c < 3{,}598$
18	$W = 0.0167\ L_m^{0.89}$	$7.93 < L_m < 23{,}201$
19	$W = 0.0228\ L_b^{0.89}$	$4.88 < L_b < 13{,}293$
20	$W = 0.0279\ B^{0.89}$	$3.05 < B < 13{,}689$
21	$W = 0.7108\ R_c^{0.89}$	$2.59 < R_c < 3{,}598$
22	$D = 0.0267 L_m^{0.66}$	$10.06 < L_m < 23{,}201$
23	$D = 0.0361 L_b^{0.66}$	$7.01 < L_b < 13{,}293$
24	$D = 0.0367 B^{0.66}$	$4.88 < B < 11{,}616$
25	$D = 0.0848\ R_c^{0.66}$	$2.59 < R_c < 3{,}598$
	Relations of meander features to channel size	
26	$L_m = 29.99\ A^{0.65}$	$0.04 < A < 20{,}914$
27	$L_b = 21.42\ A^{0.65}$	$0.04 < A < 20{,}914$
28	$B = 18.57\ A^{0.65}$	$0.04 < A < 20{,}914$
29	$R_c = 5.86\ A^{0.65}$	$0.04 < A < 20{,}914$
30	$L_m = 7.50\ W^{1.12}$	$1.49 < W < 3{,}963$
31	$L_b = 5.07\ W^{1.12}$	$1.49 < W < 2{,}134$
32	$B = 4.27\ W^{1.12}$	$1.49 < W < 3{,}963$
33	$R_c = 1.50\ W^{1.12}$	$1.49 < W < 2{,}134$
34	$L_m = 239.25\ D^{1.52}$	$0.03 < D < 18$
35	$L_b = 159.50\ D^{1.52}$	$0.03 < D < 18$
36	$B = 148.37\ D^{1.52}$	$0.03 < D < 18$
37	$R_c = 42.66\ D^{1.52}$	$0.03 < D < 18$
	Relations between channel width, channel depth, and channel sinuosity	
38	$W = 21.33\ D^{1.45}$	$0.03 < D < 18$
39	$D = 0.1492\ W^{0.89}$	$1.50 < W < 3{,}963$

(Continued)

Table 6-3 Derived Empirical Equations for River Meander and Channel Size (FISRWG 1998, with permission from USDA) *(Continued)*

Equation number	Equation	Applicable range (meters)
40	$W = 95.93\, D^{1.23}\, K^{-2.35}$	$0.03 < D < 17.99$ And $1.2 < K < 2.6$
41	$D = 0.08\, W^{0.05}\, K^{1.48}$	$1.49 < W < 3963$ And $1.2 < K < 2.6$

Note: A = bank-full cross-sectional area; B = meander belt width; D = bank-full mean depth; K = channel sinuosity; L_b = along-channel bend length; L_m = meander wavelength; R_c = loop radius of curvature; W = bank-full width. 1 ft = 0.3048 m.

Table 6-4 Limits of Data Sets used to Derive Regime Formulas (FISRWG 1998, with permission from the USDA)

Reference	Data source	Median bed-material size (mm)	Banks	Discharge (m^3/s)	Sediment concentration (ppm)	Slope	Bedforms
Lacey (1958)	Indian canals	0.1 to 0.4	Cohesive to slightly cohesive	2.37 to 237.3	< 500		
Blench (1969)	Indian canals	0.1 to 0.6	Cohesive	0.02 to 2,372.8	< 30[a]	Not specified	Ripples to dunes
Simons and Albertson (1963)	U.S. and Indian canals	0.318 to 0.465	Sand	2.37 to 9.5	< 500	0.000135 to 0.000388	Ripples to dunes
		0.06 to 0.46	Cohesive	0.12 to 2,095.2	< 500	0.000059 to 0.00034	Ripples to dunes
		Cohesive, 0.029 to 0.36	Cohesive	3.25 to 12.1	< 500	0.000063 to 0.000114	Plane
Nixon (1959)	U.K. rivers	Gravel		16.61 to 428.3	Not measured		
Kellerhals (1967)	U.S., Canadian, and Swiss rivers of low sinuosity, and lab	7 to 265	Noncohesive	0.03 to 1,675.2	Negligible	0.00017 to 0.0131	Plane
Bray (1982)	Sinuous Canadian rivers	1.9 to 145		4.60 to 3,284.0	"Mobile" bed	0.00022 to 0.015	
Parker (1982)	Single-channel Canadian rivers		Little cohesion	8.38 to 5,028.0			
Hey and Thorne (1986)	Meandering U.K. rivers	14 to 176		3.27 to 355.2	Qs computed to range up to 114	0.0011 to 0.021	

[a] Blench (1969) provides adjustment factors for sediment concentrations between 30 and 100 ppm. 1 ft^3/s = 0.0283 m^3/s.

influence channel form. For example, many popular hydraulic geometry equations express the stable width solely as a function of bank-full discharge. Intuitively, it would be expected that the width of a channel with sandy banks would be greater than that of an equivalent stream with clay banks. Indeed, Schumm's relationship between width-to-depth ratio (F) and the silt-clay weighted percentage in the channel perimeter (M) confirms this expectation empirically. If Schumm's relationship is valid, a width equation based only on discharge cannot fully account for observed width variability. Clearly, the generation of reliable results through application of simple and imperfect morphological relations

Table 6-5 Coefficients for Selected Hydraulic Geometry Formulas (FISRWG 1998, with permission from the USDA)

References	Data	Domain	k_1	k_2	k_3	k_4	k_5	k_6	k_7	k_8	k_9
Nixon (1969)	U.K. rivers	Gravel-bed rivers		0.5		0.545	0.33		$1.258n^{2b}$	−0.11	
Leopold et al. (1964)	Midwestern U.S.		1.65	0.5			0.4			−0.49	
	Ephemeral streams in semiarid U.S.			0.5			0.3			−0.95	
Kellerhals (1967)	Field (U.S., Canada, and Switzerland) and laboratory	Gravel-bed rivers with paved beds and small bed material concentration	1.8	0.5		0.33	0.4	−0.12[a]	0.00062	−0.4	0.92[a]
Schumm (1977)	U.S. (Great Plains) and Australia (Riverine Plains of New South Wales)	Sand-bed rivers	$37k_1^*$	0.38		$0.6k_4^*$	0.29	−0.12[a]	$0.01136k_7^*$	−0.32	
Bray (1982)	Canadian rivers	Gravel-bed rivers	3.1	0.53	−0.07	0.304	0.33	−0.03	0.00033	−0.33	0.59
Parker (1982)	Single-channel Alberta rivers	Gravel-bed rivers, banks with little cohesion	6.06	0.444	−0.11	0.161	0.401	−0.0025	0.00127	−0.394	0.985
Hey and Thorne (1986)	U.K. rivers	Gravel-bed rivers with									
		Grassy banks with no trees or shrubs	2.39	0.5		0.41	0.37	−0.11	$0.00296k_7^{**}$	−0.43	−0.09
		1-5% tree/shrub cover	1.84	0.5		0.41	0.37	−0.11	$0.00296k_7^{**}$	−0.43	−0.09
		Greater than 5-50% tree/shrub cover	1.51	0.5		0.41	0.37	−0.11	$0.00296k_7^{**}$	−0.43	−0.09
		Greater than 50% shrub cover or incised floodplain	1.29	0.5		0.41	0.37	−0.11	$0.00296k_7^{**}$	−0.43	−0.09

Notes: b_n = Manning n.
$k_1^* = M^{-0.39}$, where M is the percent of bank materials finer than 0.074 mm. The discharge used in this equation is mean annual rather than bank-full.
$k_4^* = M^{0.432}$, where M is the percent of bank materials finer than 0.074 mm. The discharge used in this equation is mean annual rather than bank-full.
$k_7^* = M^{-0.36}$, where M is the percent of bank materials finer than 0.074 mm. The discharge used in this equation is mean annual rather than bank-full.
$k_7^{**} = D_{54}^{0.84} Q_x^{0.10}$, where Q_x = bed material transport rate in kg s^{-1} at water discharge Q, and D_{54} refers to bed material and is in mm.
[a] Bed material size in Kellerhals' equation is D_{90}.

relies heavily on good insight and sound judgment on the part of the individual responsible for their application.

A misapplication of empirical relationships was lampooned by Mark Twain (1944) in *Life on the Mississippi.* Describing the Mississippi River cutoffs of which he had knowledge, he conceived a simple empirical relationship between river shortening and time, and then used it to predict the historical and future lengths of the Mississippi River, concluding that:

> Geology never had such a chance, nor such exact data to argue from! In the space of 176 years, the Lower Mississippi has shortened itself 242 miles. That is an average of a trifle over one mile and a third per year. Therefore, any calm person, who is not blind or idiotic, can see that in the Old Oölitic Silurian Period, just a million years ago next November, the Lower Mississippi River was upwards of 1,300,000 miles long, and stuck out over the Gulf of Mexico like a fishing rod. And by the same token, any person can see that 742 years from now the Lower Mississippi will be only a mile and three-quarters long, and Cairo and New Orleans will have joined their streets together, and be plodding comfortably along under a single mayor and a mutual board of aldermen. There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment of fact.

The primary points of this passage are that, no matter what their basis in fact and observation, empirical relationships cannot be extrapolated either backward or forward in time, and engineers must avoid falling into the trap of designing a project based solely on "…wholesale returns of conjecture out of a trifling investment of fact."

6.6 CHANNEL STABILITY AND INSTABILITY

In designing river enhancement and channel rehabilitation projects the design engineer must recognize that rivers are dynamic systems, and must consider both the existing and possible future channel morphologies in the design. The problem is compounded when engineering interventions are planned, because the future morphology of the channel depends not only on the natural, or autonomous, evolution of the system, but also on channel response to construction, operation, and maintenance of the project. For this reason, it is important for the design engineer to acquire a broad understanding of the current stability status of the project reach and the extended channel network and to use this understanding to predict the type and extent of adjustments to the fluvial system likely to be triggered by the project. The capability to predict system response to the proposed works is vital to ensure that the selected enhancement or rehabilitation measures will work in harmony with both existing and future river conditions. The concept of channel stability status (which incorporates instability) builds on the basic geomorphic principles introduced previously and may be applied to the river at system and local scales.

6.6.1 System Stability

The geomorphic concept underpinning stability assessment in rivers is that over time the cross-sectional dimensions and longitudinal slope of the channel of an alluvial stream adjust so that the channel is able to convey the discharges of water and sediment supplied from upstream with no net change in hydraulic geometry or planform. On this basis, a stream may be classified as either *stable* or *unstable*, depending on whether the channel has adjusted or is still adjusting to the flow and sediment regimes. Mackin (1948) expressed the stability concept in his definition of the *graded stream*:

> A graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and with prevailing channel characteristics, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium.

By definition, a graded stream does not have to have a channel that is static or fixed, and it may exhibit temporary morphological changes in response to the impacts of extreme events. Alluvial channel morphology is certain to be affected by major floods or protracted periods of low water, but provided that the time for moderate events to restore the graded morphology (termed the *recovery time*) is shorter than the return period for the extreme event (*recurrence interval*), the channel may be considered to be dynamically stable. The key attribute of a graded stream is that fluvial processes operating under formative flows tend to restore channel morphology to the graded condition following disturbance, rather than perpetuating or amplifying the changes imposed by the extreme event. A term commonly used for this type of stability is *dynamic equilibrium*.

The concept of dynamic equilibrium is inherent in a widely applied (and often misapplied), qualitative relationship for adjustment in alluvial streams proposed by Lane (1955):

$$QS \sim Q_s D_{50}$$

where Q = water discharge, S = slope, Q_s = bed-material load, and D_{50} = median size of the bed material. This relationship is commonly visualized as *Lane's balance* (Fig. 6-10). Mackin's explanation of how a graded stream responds to changes in the controlling variables is easily illustrated by Lane's balance, which shows how a change in any of the four driving variables will tend to produce a response in the others such that equilibrium is restored. When a channel is in dynamic equilibrium, it has adjusted these four variables so that the sediment transported into the reach is also transported out, without aggradation or degradation.

It should be noted that the map coordinates of a graded stream may change through time as the river reworks the

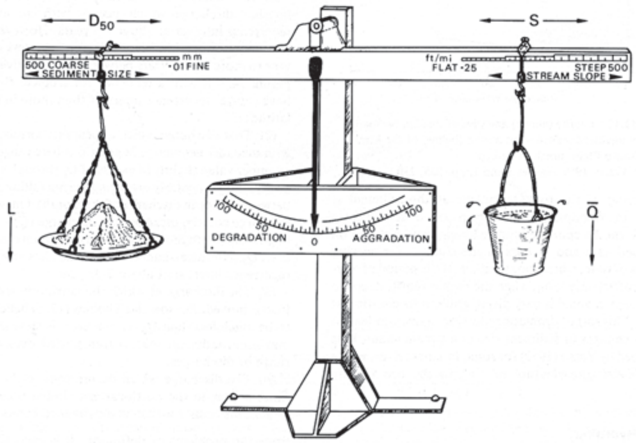


Fig. 6-10. Lane's balance (Rosgen 1996, with permission from Wildland Hydrology).

floodplain through meandering or braiding, provided that the reach-averaged values of width, depth, slope, and planform geometry are time-invariant. Indeed, meandering provides an important mechanism for an alluvial stream to adjust the slope relatively quickly and without transferring the large amounts of (relatively coarse) bed sediment necessary to alter slope materially through aggradation and degradation. Viewed in this context, changes in channel length achieved through meander extension and cutoff represent a natural adjustment mechanism, and planform changes do not necessary indicate disequilibrium. When natural cutoffs occur, the river may be obtaining additional length elsewhere through meander growth, with the net result being that the overall reach length, and therefore slope, remains unchanged.

In nature, few rivers actually attain a graded condition because the driving variables change through time. The concept still has value, however, because it provides an indication of the likely trend of channel evolution over engineering time scales, which are generally less than about 50 years. Although it is a mistake to assume that a river will be stable or unchanging over this period, the concept of dynamic equilibrium gives useful clues regarding the rates and types of adjustment that may be expected as the channel evolves toward a graded condition. Also, the proximity of the system to a graded condition gives an indication of how the river will respond to engineering interventions and, particularly, how sensitive it is to being destabilized. Finally, the geomorphic concept of the stable channel is valuable in that it establishes a reference point for the definition and treatment of morphological instability on a variety of scales.

6.6.2 Channel Instability

Channel instability is defined as temporal change in the hydraulic geometry, long profile, or planform pattern of a channel because of inequality between the supply and removal of sediment. Instability is, in a broad sense, inherent in the natural action of rivers in changing the landscape by eroding, transporting, and depositing sediment. In fact, the situation where sediment input exactly matches sediment output (dynamic stability) is actually a special case that can strictly occur only in subreaches of a fluvial system and that cannot persist for long periods.

Instability may result when the flow of water and transfer of sediment through a drainage network is disrupted or significantly perturbed. The fluvial system initially responds to disequilibrium by adjusting channel morphology in ways that tend to restore the previous equilibrium or graded condition. If stability is restored through a process-response that returns channel morphology to the predisturbance configuration (or something essentially similar), then the adjustments involved are restricted to the immediate vicinity of the disruption and, by definition, constitute *local instability*. However, if the magnitude of the change in driving variables is large, or the river is sensitive to destabilization because of channel characteristics (high stream power, easy availability of sediment, high erodibility of bed and bank materials, or absence of geologic or artificial controls) or proximity to a geomorphic threshold, then morphological adjustments can take the channel toward a new equilibrium configuration different from the predisturbance morphology. Under these circumstances, *system instability* propagates throughout the channel network and may spread into the watershed or even into neighboring systems.

6.6.3 Local Instability

Local instability refers to channel changes that result from adjustments to a fluvial system inherent in the maintenance of a dynamically stable configuration. There are three common causes of local instability. The first is channel response to temporary variations in discharge or sediment flux. Typically, discharge variations occur seasonally, or result from longer periods of above-average or below-average precipitation, whereas sediment input varies because of pulsing of sediment between storage and transport reaches or shifts in upstream channel alignment. The second cause is the series of adjustments that occur when channel morphology is altered by, and subsequently recovers from, the impact of a rare event such as a flood, drought, wildfire, or earthquake. The third cause is disruption of fluvial forms or processes associated with human activity or construction of infrastructure in or around a channel that triggers the channel changes necessary to accommodate the impacts of that disturbance within the existing, dynamically stable condition. Local instability is not symptomatic of significant disequilibrium in the system, but this does not mean that the processes of bed scour, bar deposition, and bank erosion associated with local instability are limited to a single location or that their consequences are negligible.

A good example of local instability is bankline movement due to planform evolution in a meandering river. Whereas the reach-averaged dynamically stable parameters of hydraulic geometry and slope remain steady, individual bends in a meandering river grow, migrate, and are abandoned. On average, channel lengthening through bank erosion along the concave bank in growing meander bends is offset by cutoffs at other bends as part of the natural meandering process. Under these circumstances, problems associated with bank erosion at a bend are amenable to local bank protection works, provided that the hydraulic geometry and slope of the reach are not significantly altered. However, it should be kept in mind that the channel may respond to stabilization of one bend through accelerated morphological activity in adjacent free bends. Hence, care must be taken to ensure that management of local instability at one location does not transfer or concentrate this natural process elsewhere in a way that is detrimental to the dynamic stability of the system.

The causes of local instability are not limited to the channel. This type of instability can also be triggered by activities in surrounding riparian and floodplain areas. For example, a reach of stream may display local channel widening due to trampling and overgrazing by cattle, while upstream and downstream reaches are not directly affected and are able to remain dynamically stable. In this situation, a local management solution, based on restriction of access by fencing, construction of suitably reinforced access ramps at water points, and reinstatement of the regime width, is all that is needed to alleviate a site-specific problem. Site-specific instability problems may respond satisfactorily to design alternatives developed using reference reach techniques.

In practice, however, it is not always easy to establish whether a local instability problem results from and is amenable to a local solution or is symptomatic of more serious, system-scale impacts and adjustments. Even if the engineer suspects that local instability results from adjustments of the fluvial system to channel instability, human activities, or catchment land-use changes, they may lack the authority or resources to address off-site and nonpoint causes. Under these circumstances, the engineer may have to modify the adopted solution by constructing a local structure with the capability to continue functioning successfully even when system-driven channel adjustments have significantly altered local conditions. For example, local bank stabilization may be required at the outside of a migrating bend on a river that is predicted to degrade in the future because of system instability downstream. Ideally, the system-scale problem (degradation) should be addressed directly using one or more grade control structures, but this may be institutionally or financially unfeasible. Recognition that the problem is not entirely local is nonetheless still valuable, as it allows the engineer to determine the degree of additional toe scour protection necessary to ensure that the bank protection measures can withstand the additional bed lowering associated with degradation during the design life of the project.

6.6.4 System Instability

Adjustments involved in system instability typically involve *aggradation* (increasing bed elevation), *degradation* (decreasing bed elevation), or *planform metamorphosis* (abrupt alteration from one planform pattern to another). The response of an alluvial stream to an episode of system instability is, in detail, unique to that stream and the circumstances and timing of the events responsible for destabilization. Although channel evolution models (discussed later in the section on stream classification) have been developed to characterize commonly observed styles and sequences of adjustment in unstable systems, there is no generally applicable model for process-response to system instability.

Serious engineering and river-management problems often result from channel instability and may include endangerment of bridges, buildings, roads, and other infrastructure, undermining of pipeline and utility crossings, accelerated bed and bank erosion, loss of valuable environmental habitat, and increased sediment loads that adversely impact flood control and navigation channels, water quality, reservoir areas, and wetlands. Figure 6-11 illustrates some common consequences of system instability.

The causes of system instability can be grouped into three categories: downstream factors, upstream factors, and basin-wide factors.

6.6.4.1 Downstream Factors The stability of a fluvial system can be affected significantly by changes to downstream base level. *Base level* refers to the downstream limit of the channel network, the elevation of which defines the datum for measurement of potential energy in the system upstream. In subcritical flow, the water surface elevation at the downstream limit of the channel controls the longitudinal water surface profile for a stream. Similarly, the bed elevation at the downstream limit of the system represents the origin of the thalweg profile. It follows from these facts that changes in base level have strong potential to trigger system instability.

Base-level lowering, due to engineering interventions such as meander cutoffs or channelization (Fig. 6-12), triggers process-response by locally steepening the slope and increasing bed-material transport capacity. As capacity exceeds supply, the bed scours to make up the supply deficit as the channel adjusts through degradation. This adjustment may generate only local instability if armoring stabilizes the bed or a geological control prevents significant bed lowering. However, if unchecked by a local channel response or control, a wave of degradation migrates upstream through the system as a headcut or knickpoint. If degradation triggers bank instability, then a wave of channel widening may follow the headcut, generating further morphological adjustments and additional sediment input

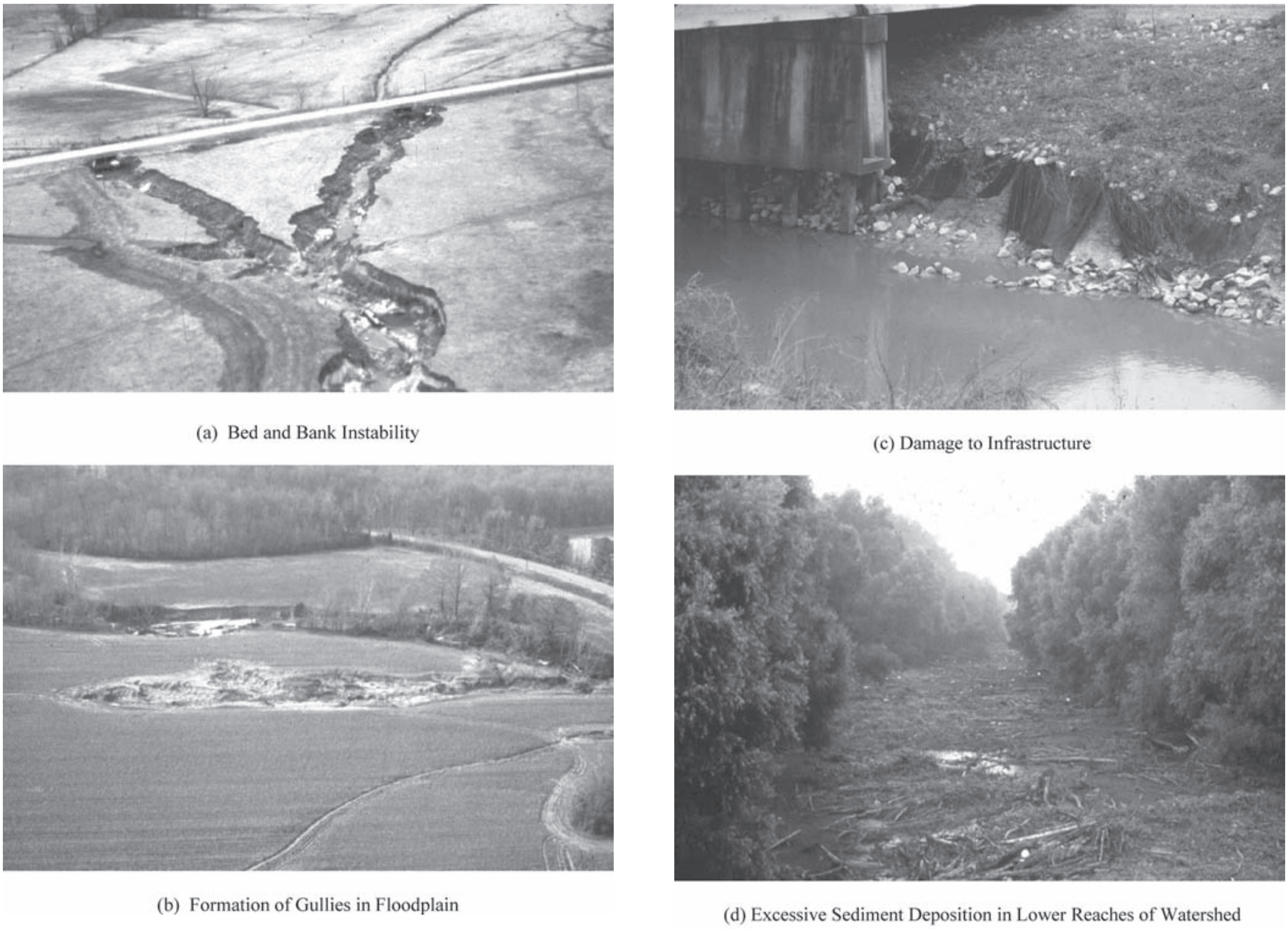


Fig. 6-11. Consequences of system instability: (a) bed and bank instability, (b) formation of gullies in floodplain, (c) damage to infrastructure, and (d) excessive sediment deposition in lower reaches of watershed.

Fig. 6-12. Channelized stream and abandoned old channel.

to the channel. As the degradational wave moves upstream, the zone of increased slope and additional sediment production moves with it. Sediment supply to the downstream reaches, coupled with local slope reduction due to upstream bed lowering, then triggers aggradation, which also migrates upstream through the system. Subsequently, sediment output and bed elevation at the downstream limit of the system display damped oscillation until, following a number of cycles of degradation/aggradation, the long profile is adjusted to the new base level and stability is restored.

6.6.4.2 Upstream Factors The stability of a fluvial system can also be significantly affected by changes to upstream reaches that alter the downstream discharge or sediment supply. The flow regime and sediment load together constitute the two main driving variables responsible for forming and maintaining the channel, and it is no surprise that the stability of an alluvial river may well be disturbed by changes in one or both of these factors. Upstream factors are often affected by engineering and river-management projects. River regulation by a dam or diversion structure is a common cause of downstream channel adjustment that serves to illustrate the types and complexity of system response that may result from such changes.

Channel response downstream of a dam or diversion structure depends on the way the works are constructed and operated. When the structure is built, sediment supply downstream may be elevated by disruption of the channel and floodplain during construction. This may increase supply over transport capacity, inducing an initial adjustment through aggradation. However, this response will be absent if appropriate sediment-control measures are applied on site. Once the works are complete, process-response downstream will depend on the balance of changes in the water and sediment regimes. Following closure of the dam or diversion, sediment is trapped in the pool upstream from the structure. Sediment-free water released from the structure then scours the bed downstream, generating degradation in the first few kilometers below the dam. Initially, the flush of sediment produced drives aggradation further downstream, but as the channel slope immediately downstream of the dam flattens, sediment output decreases, and the leading edge of the zone of degradation migrates downstream to re-erode recently deposited sediment and sends it further downstream as an aggradational wave. This river response to closure of a dam has been observed in many rivers, and yet this pattern of adjustment is by no means universal. To explain why, it is necessary to consider the other morphological responses that may dominate adjustment of the fluvial system. For example, if the bed downstream of the dam includes a widely graded, coarse-grained fraction, bed armoring may limit degradation and stabilize the bed at a slope *steeper* than that prior to dam construction. The same effect may result from the presence of a geological control, whereas widening with limited reduction in bed level may be triggered if the channel banks downstream of the dam are close to the critical height for mass instability (Thorne and Osman 1988). If regulation by the dam significantly reduces the magnitude or frequency of sediment-transporting flows, degradation may be limited or negligible, and if a reduction in the competence of the main stream is coupled with the input of a substantial sediment load from unregulated tributaries, aggradation may occur downstream of the dam, where degradation was expected (Biedenharn 1984). The point of citing these examples is to demonstrate that morphological response to change in one or more upstream factors is complex and difficult to predict. The specific attributes of system instability and channel response depend not only on the magnitude of changes imposed on the flow regime and sediment loads, but also on the sensitivity and boundary conditions of the downstream channel network.

6.6.4.3 Basin-Wide Factors In morphological studies, flow regime and sediment load are often cited as the independent variables controlling channel form and process. In reality, these variables are not truly independent, but depend in turn on the characteristics of the watershed, including factors such as climate, rainfall-runoff relationship, natural vegetation, land use, and resource management. Even if upstream and downstream factors remain constant, changes in the watershed may trigger instability in the fluvial system that leads to widespread morphological adjustments. For example, urbanization can increase peak flows and reduce sediment delivery to the channel network. These changes would reinforce one another (see Lane's balance in Fig. 6-10) to drive marked degradation in the channel draining the urbanized area, with morphological impacts migrating upstream and downstream through the system by slope adjustment and sediment transmission, respectively. Afforestation of the headwaters of a stream could produce very different morphological adjustments, depending on whether fluvial processes respond more strongly to decreases in runoff (due to increased consumptive use in the watershed) or elevated sediment delivery (due to erosion along forestry roads and ditches). In practice it is even more difficult to predict the morphological response of a fluvial system to basin changes than to upstream changes. In attempting to develop the capability to predict channel response to basinwide changes, engineers and river managers should make every effort to familiarize themselves with the geography of the basin, processes operating in the fluvial system, sedimentary features, and channel morphology. This knowledge, together with application of conventional hydrologic, hydraulic, and sediment-transport analyses, represents the best current option for regional sediment management and channel stabilization in a changing watershed.

6.7 CHANNEL CLASSIFICATION

The existence of a few distinctive channel forms provides the rationale for morphological classification of channels. The relationships linking channel form to fluvial process suggest that the morphological classification of a channel may allow the morphologist to infer process from classified channel form. The first step in classification is to identify whether the channel is either alluvial or nonalluvial. An *alluvial* channel is "self-formed" in that the bed and banks are composed of material transported by the river under present flow conditions. The channel is therefore free to adjust dimensions and location in response to changes in flow and sediment load. Conversely, a *nonalluvial river* is neither self-formed nor free to adjust. Examples of nonalluvial rivers include bedrock-controlled channels and streams flowing over very coarse glacial deposits.

Many classification schemes rest on channel planform pattern and stem from Leopold and Wolman's (1957) classification of channel planforms as straight, meandering, or braided. In this respect, the diagram produced by Brice (1975) is notable because it builds on earlier schemes to cover a wide range of commonly observed planforms and has proved useful in engineering geomorphic studies (Fig. 6-13). Schumm (1981; 1985) recognized an even broader range of channel patterns, although the basic straight, meandering, and braided patterns are still recognized within his classification of 14 basic patterns (Fig. 6-14).

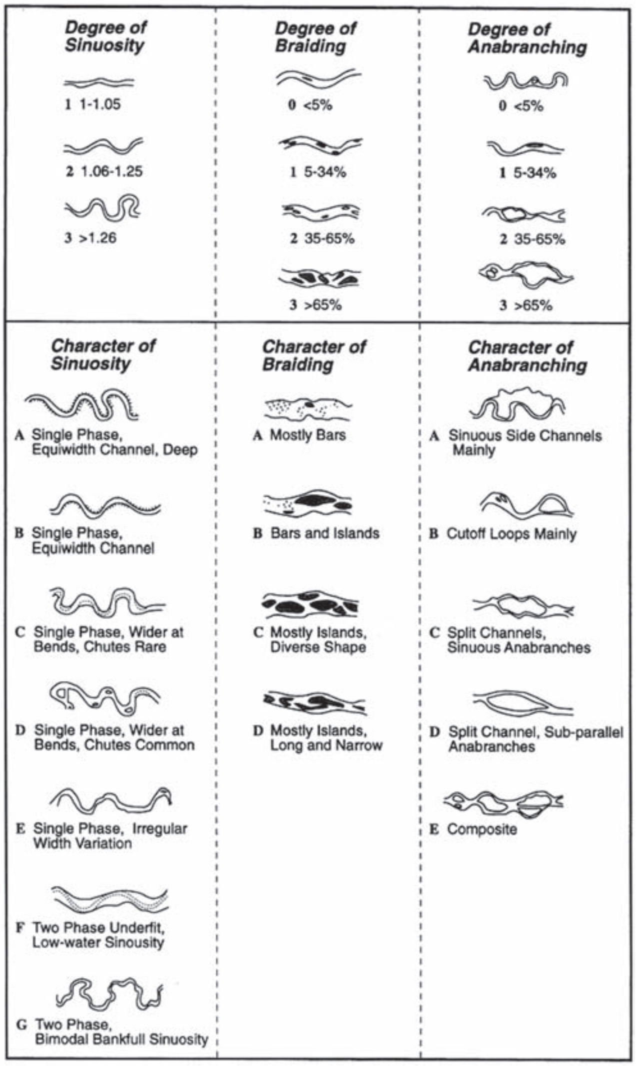


Fig. 6-13. Channel Pattern classification devised by Brice (after Brice 1975).

Knighton (1998) related Schumm's 14 patterns to investigations by Carson (1984a; 1984b); Knighton and Nanson (1993); and Nanson and Knighton (1996). Knighton (1998) noted that the patterns in Fig. 6-14 are related to the classification by the type of sediment load (Schumm 1976): bed load, mixed load, and suspended load. Types 1 through 5 are bed load streams, Types 6 through 10 are mixed load streams, and Types 11 through 14 are suspended load streams. Carson (1984a; 1984b) specified two types of wandering, gravel-bed rivers. The first is characterized by very rapid bend migration and frequent chute cutoffs of point bars, similarly to Type 3. A second wandering type is similar to Type 14, with vegetated islands separating most of the channels. Knighton and Nanson (1993) point out that coarse-grain, anastomosing channels do exist. Therefore, despite the variety of channel patterns that have been investigated and discussed, a continuum of channel patterns does exist and these patterns are controlled by the interaction of a series of continuous variables. Figure 6-14 suggests some of the variables that should be considered, such as sediment size and transport mechanism, whereas Bledsoe's (1999) logistic threshold approach indicates that specific stream power and sediment size are also important (Fig. 6-2).

In parallel with the development of his morphological classification, Schumm (1977) considered of the type of sediment load being transported by the stream, the percentage of silt and clay in the channel bed and banks, and the stability of the channel to describe the morphology associated with stable conditions and the morphological changes expected in response to instability through aggradation or degradation (Table 6-6). For purposes of this classification system, a stable channel complies with Mackin's definition of a graded stream in that slope is adjusted to supply just the sediment transport capacity necessary to convey the sediment load supplied from upstream. An unstable stream may be either *degrading* (eroding) or *aggrading* (depositing). It is very important to remember that the work on which this classification was based was conducted in the Midwestern United States during the second half of the 20th century. Extrapolation or transfer of the classification or related implications to other times and places should, therefore, be done cautiously.

Other, more ambitious stream classifications have been developed by Neill and Galay (1967); Rundquist (1975); and Rosgen (1994). These classifications go well beyond a description of channel form to include description of land use and vegetation in the basin, geology of the watershed, hydrology, channel bed and bank materials, sediment concentration, channel pattern, and channel stability.

Rosgen (1994) presented a stream classification system similar to the earlier Rundquist (1975) system. Rosgen (1996) included classification of valley type and introduced an entrenchment ratio, defined as the ratio of the width of the flood-prone area to the surface width of the bank-full channel. Table 6-7 is a summary of delineative criteria for broad-level classification from Rosgen (1994). Each of the stream types can be associated with dominant bed material types as follows: bedrock—1, boulder—2, cobble—3, gravel—4, sand—5, and silt/clay—6.

Through modification of Fig. 6-14, Fig. 6-15 attempts to combine some of the concepts of Schumm and Rosgen. Schumm's classification system depends heavily on his Midwestern experience, whereas Rosgen's experience began in steep mountain streams. In addition, Schumm's (1977) classification does not specifically include incised channels, which are included in Rosgen's (1994) F and G classes. Figure 6-15 includes Rosgen's C, D, DA, and E classes, and could be expanded to include all of Rosgen's (1994) classes. The point of Fig. 6-15 is to demonstrate that moving from class to class is a somewhat predictable morphological response that manages energy, materials, and channel planform to reestablish the balance between the local capacity of

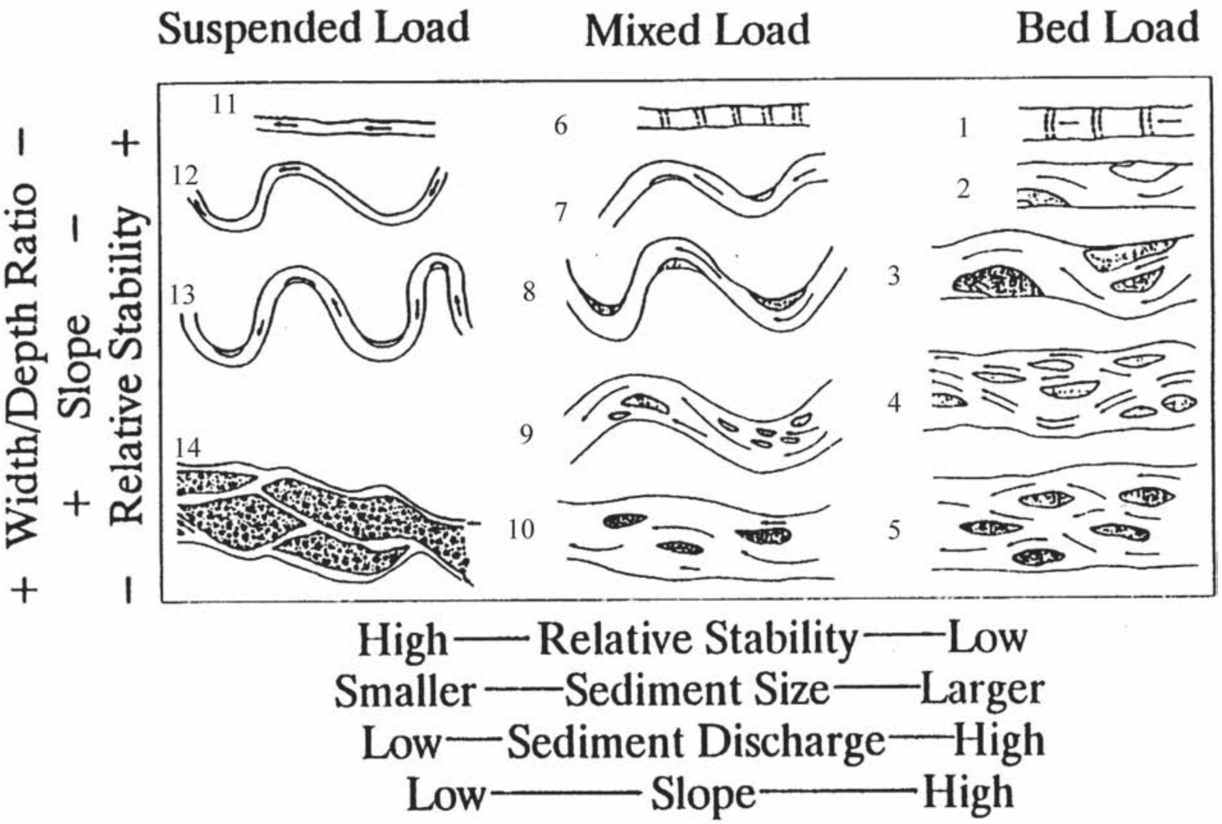


Fig. 6-14. Channel classification based on pattern and type of sediment load (Schumm 1981, with permission from SEPM [Society for Sedimentary Geology]).

Table 6-6 Classification of Alluvial Channels (Schumm 1977, with permission from S. Schumm)

Mode of sediment transport and type of channel	Channel sediment (M) (%)	Bedload (percentage of total load)	Channel stability: Stable (graded stream)	Aggrading (excess sediment discharge)	Degrading (deficiency of sediment discharge)
Suspended load	>20	<3	Stable suspended-load channel. Width/depth ratio <10; sinuosity usually >2.0; gradient, relatively gentle	Depositing suspended load channel. Major deposition on banks cause narrowing of channel; initial streambed deposition minor	Eroding suspended-load channel. Streambed erosion predominant; initial channel widening minor
Mixed load	5-20	3-11	Stable mixed-load channel. Width/depth ratio >10, <40; sinuosity usually <2.0, >1.3; gradient moderate	Depositing mixed-load channel. Initial major deposition on banks followed by streambed deposition	Eroding mixed-load channel. Initial streambed erosion followed by channel widening
Bed load	<5	>11	Stable bed-load channel. Width/depth ratio >40; sinuosity usually <1.3; gradient, relatively steep	Depositing bed-load channel. Streambed deposition and island formation	Eroding bed-load channel. Little streambed erosion; channel widening predominant

Table 6-7 Summary of Delineative Criteria for Broad-Level Classification (Rosgen 1994, with permission from Wildland Hydrology)

Stream type	Entrench. ratio	*w/d* ratio	Sinuosity	Slope	Meander belt/ bank-full width	Dominant bed material[a]
Aa+	<1.4	<12	1.0–1.1	> 0.10	1.0–3.0	1,2,3,4,5,6
A	<1.4	<12	1.0–1.2	0.04–0.10	1.0–3.0	1,2,3,4,5,6
B	1.4–2.2	>12	>1.2	0.02–0.039	2.0–8.0	1,2,3,4,5,6
C	>2.2	>12	>1.2	< 0.02	4.0–20	1,2,3,4,5,6
D	na	>40	na	< 0.04	1.0–2.0	3,4,5,6
DA	>2.2	variable	variable	< 0.005	na	4,5,6
E	>2.2	<12	>1.5	< 0.02	20–40	3,4,5,6
F	<1.4	>12	>1.2	< 0.02	2.0–10	1,2,3,4,5,6
G	<1.4	<12	>1.2	< 0.039	2.0–8.0	1,2,3,4,5,6

[a] Dominant bed material key: 1, bedrock; 2, boulders; 3, cobble; 4, gravel; 5, sand; 6, silt/clay.

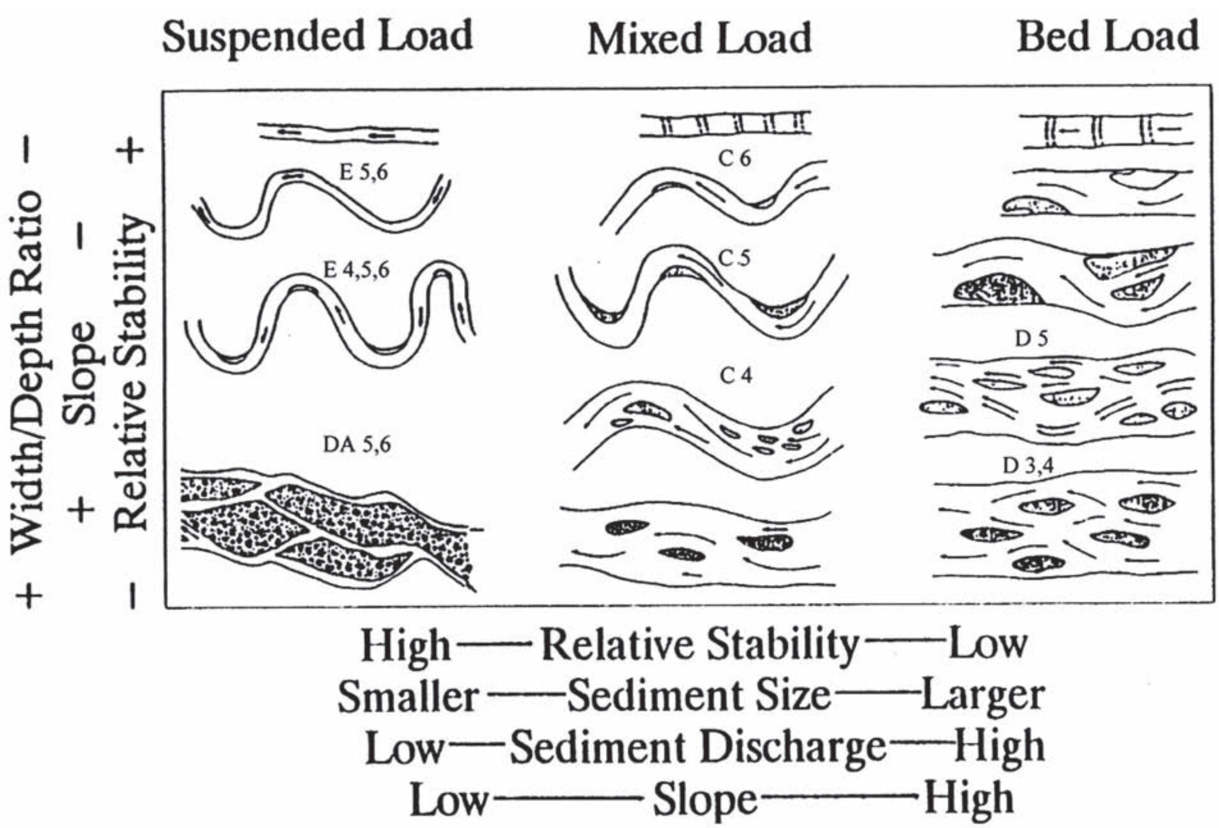


Fig. 6-15. Channel classification combining aspects of Schumm (1981) and Rosgen (1994) (Schumm 1981 and Rosgen 1994, with permission from SEPM [Society for Sedimentary Geology]).

the channel to convey water and sediment and the discharge and supply of sediment from upstream.

Thorne et al. (1997) point out that many classification systems fail to account for dynamic adjustment or evolution of the fluvial system. Downs (1995) developed a comprehensive system that incorporates the classifications of Brice (1975) and Brookes (1981) and builds on their earlier work by linking observed trends and patterns of adjustment to the fluvial and sediment processes responsible for driving channel change (Fig. 6-16). Adjustment-based classifications such

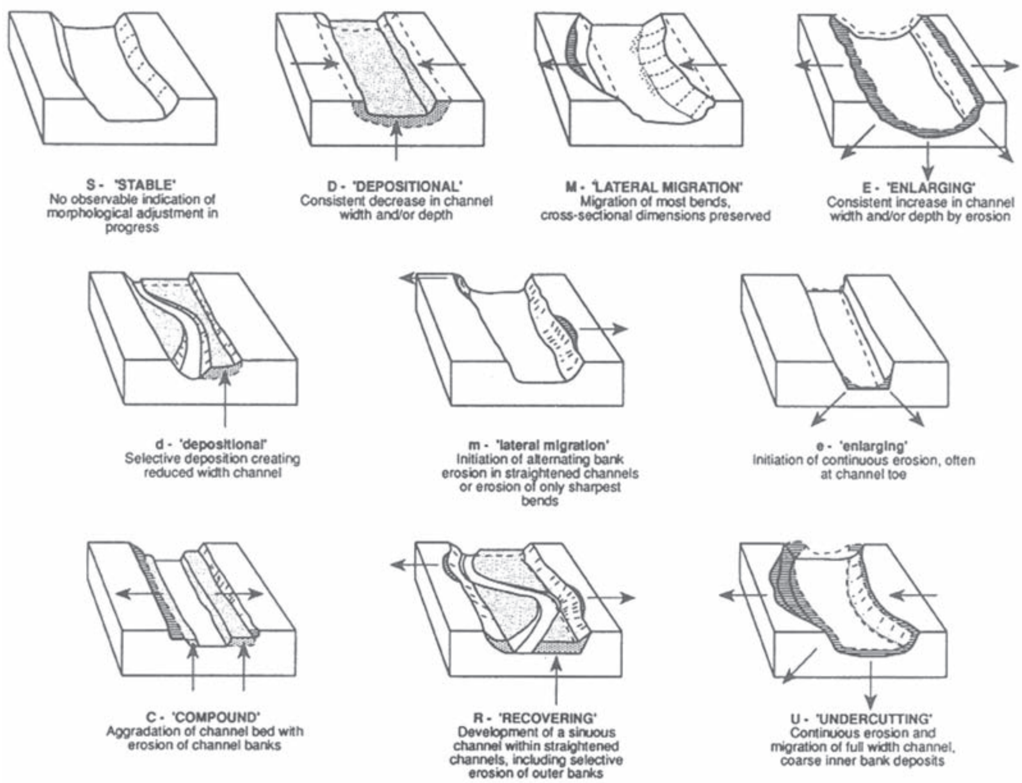


Fig. 6-16. Downs's channel classification, based on trends and types of morphological change (modified from Downs 1995).

as that of Downs differ fundamentally from morphology-based schemes in that each system requires the observer to determine the current stability status of the channel and the nature of channel adjustment processes. Because data may not be available to document change, these schemes require sound judgment on the part of the engineer, who must infer processes and trend of adjustment from channel form.

Although any conceivable morphological channel classification will oversimplify the variability of channel patterns in nature, the underlying concept of a continuum of channel patterns that is related to a limited number of controlling variables remains valid. The opportunity and challenge for the river engineer is to develop and refine associations between channel pattern characteristics and controlling variables and to use these relationships with care and caution to predict the manner in which pattern will change in response to alteration of controlling variables. Schumm (1976) points out that major alterations in pattern change, which he terms channel metamorphosis, may be triggered by a relatively minor change in a controlling variable, if the existing pattern is near a geomorphic threshold.

6.8 CHANNEL EVOLUTION MODELS

Numerous geomorphological studies have used data developed from different locations to infer landform development through time, commonly employing a technique termed location-for-time substitution. This technique assumes that by observing channel form as one moves downstream along a channel, the effect of physical processes at one location through time can be predicted; that is, changing location is substituted for changing time. This technique was used to develop a channel evolution model (CEM) for Oaklimeter Creek, an incised stream in northern Mississippi (Schumm et al. 1984). Simon and Hupp (1987) later developed a similar model of channel evolution based on their observations of incised streams in western Tennessee.

The CEM (Fig. 6-17) consists of five channel-reach types, which describe the evolutionary phases typically encountered in an incised channel. These evolutionary phases range from strong disequilibrium to a new state of quasi-equilibrium. Quasi-equilibrium implies that the system is not static and changes through time, but over a period of years the average condition is one of stability. The model is based on the assumption that moving downstream through the system is equivalent to remaining in place and monitoring changes due to the passage of time. The response at any given location in the channel can then be predicted from the morphology of downstream channel locations.

The channel reach types in the CEM are labeled I through V and are assumed to occur consecutively in the downstream direction. The CEM assumes that each channel type will occur in turn at a given location as the channel evolves. The CEM channel types are shown in Fig. 6-17. Type I reaches are located upstream of the actively degrading reach and have not yet experienced significant bed or bank instabilities. These reaches are generally characterized by U-shaped cross sections with little or no recently deposited sediment stored in the channel bed.

Type II reaches are encountered immediately downstream of Type I reaches. Bed degradation is the dominant process in the Type II reach. Type II channels are over steepened reaches where the sediment transport capacity exceeds the sediment supply. Although the channel is actively degrading in a Type II reach, the bank heights (h) do not exceed the critical bank height (h_c), and therefore, reach-scale geotechnical bank instability is not encountered.

As bed degradation continues, the bank heights and angles continue to increase. When the bank heights exceed the critical bank height for stability in the Type III reaches, mass failures (geotechnical instability) begin. The dominant process in the Type III reach is channel widening. In places, the Type III reach may continue to be slightly degradational.

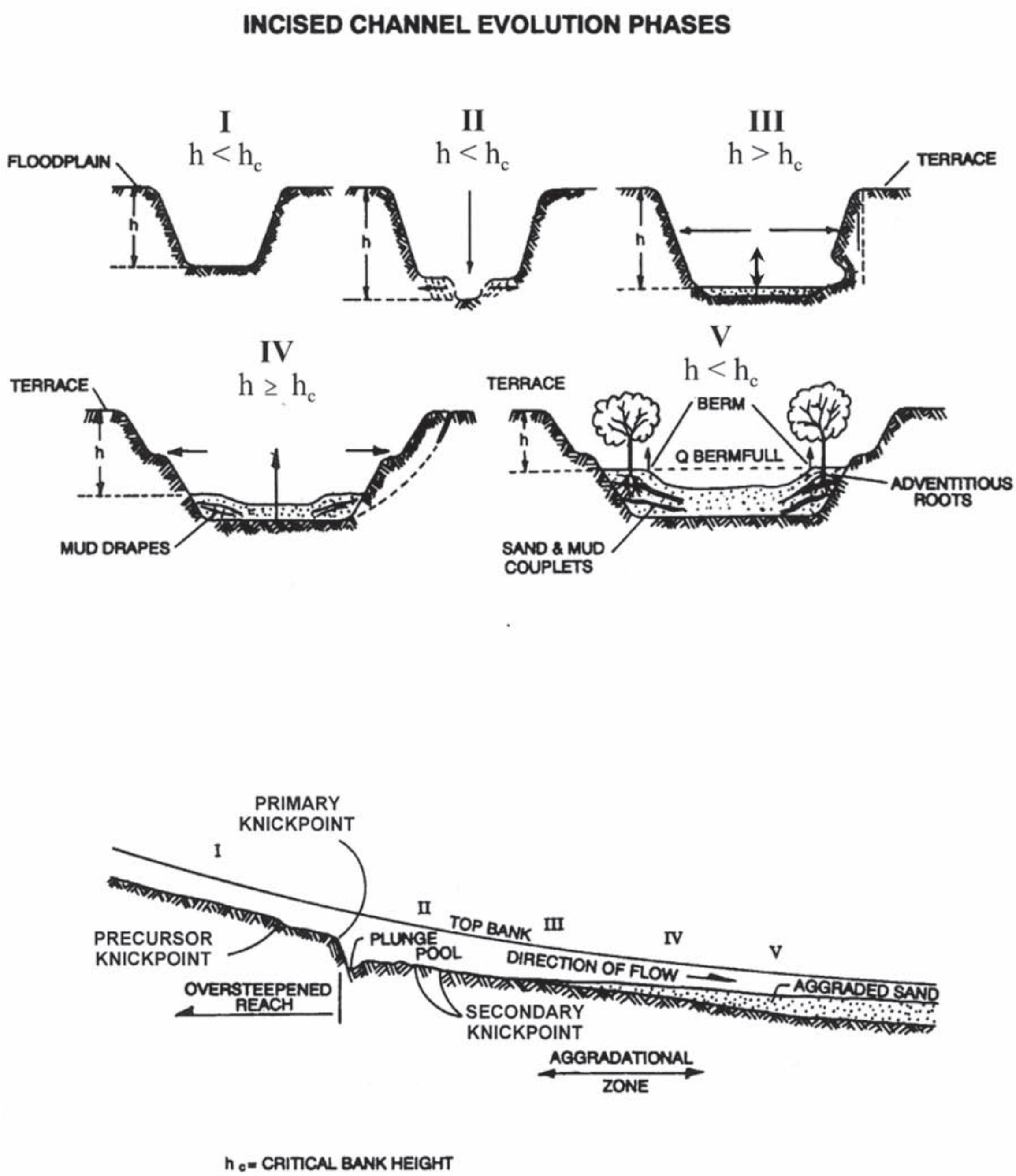


Fig. 6-17. Incised channel evolution sequence (after Schumm et al. 1984).

However, the reduced sediment transport capacity resulting from longitudinal channel slope decrease combined with increased sediment supply from upstream due to instability and from bank failures within the reach often results in the initiation of sediment deposition on the channel bed.

Type IV reaches are downstream of the Type III reaches and represent the first manifestation of the incising channel returning to a new state of dynamic equilibrium. In the Type IV reach, geotechnical bank instabilities and channel widening may continue, but at a much reduced rate. The sediment supply from upstream (Type III) exceeds the sediment transport capacity, resulting in aggradation of the Type IV channel bed. The Type IV reach is also characterized by the development of *berms*, which are depositional features along margins of the overwidened channel. These berms represent the beginning of a new inner channel with dimensions adjusted to the flow and sediment regime.

Type V reaches represent a state of dynamic equilibrium, with a balance between sediment transport capacity and sediment supply. Bank heights in the Type V channel are generally less than the critical bank height, and therefore, reach-scale geotechnical bank instability ceases. However, local bank failures can still exist as part of the meander process, or as the results of constrictions, obstructions, or other local factors. The berms that were initiated in the Type IV reach have now become colonized by riparian vegetation, forming a compound channel within the larger incised channel. The equilibrium channel of Type V is of a compound shape, with a smaller inner channel bounded by a narrow floodplain. The original floodplain of the Type I channel is now a terrace.

The channel evolution model addresses the channel stability status within a system context. Dynamic equilibrium in a Type V reach simply implies that system stability has been attained. A Type V reach may exhibit considerable erosion that is part of the natural meander process or some other local process, yet still be classified as being in dynamic equilibrium.

The primary value of the CEM sequence is to underpin identification of the evolutionary state of the channel from field reconnaissance. The morphometric characteristics of the channel reach types can also be correlated with hydraulic, geotechnical, and sediment-transport parameters (Harvey and Watson 1986; Watson et al. 1988). The evolution sequence provides an understanding that although reaches of a stream may differ markedly in appearance, the channel form in one reach is associated with those in adjacent and remote reaches by an evolutionary process. Form, process, and time relate dissimilar reaches of the stream linked to complex response and connectivity in the water and sediment-transfer systems.

6.9 GEOMORPHIC ASSESSMENT

Given the significance of fluvial geomorphology to engineering and management of rivers, the problem remains of gathering the data and qualitative information necessary to characterize and define channel form, process, and stability status in the project river. A thorough geomorphic assessment of the river and watershed is required. Unfortunately, many engineers charged with the design of river projects either fail to fully appreciate the importance of geomorphic assessments, or lack the education or training background to perform them adequately.

Geomorphic assessment is an essential part of the design process for schemes ranging from local bank protection through reach-scale habitat enhancement to master planning for water resource management in an entire watershed. The aims of geomorphic assessment are to provide the baseline information necessary to characterize process-form interactions in the river, identify control points and problem reaches, and support division of the system into geomorphically distinct subreaches that may be individually classified with respect to morphology. Once the system has been characterized and classified, the engineer may assess the stability status on a reach-by-reach basis and predict the medium- and long-term autonomous evolution under a do-nothing scenario. This provides a baseline against which to assess the morphological responses of the project reach and wider system to the proposed engineering, rehabilitation, or water resources project.

Perhaps the most important step in any geomorphic assessment is ensuring that the scope and content match the project goals, authority, channel and watershed characteristics, and available resources. There is no standardized or "cookbook" approach, but over the past two decades a number of assessment schemes have been developed, and these provide valuable guidance based on direct experience (Simons et al. 1982; Schumm et al. 1984; Richardson and Huber 1991; Schall and Lagasse 1991; Shirole and Holt 1991; Robinson and Thompson 1993; Biedenharn et al. 2000b). Typically, existing geomorphic assessment techniques may be subdivided into procedural steps dealing with

1. Assembly of existing and archived data/information in a desk study;
2. Establishment of current channel forms and sediment features through stream reconnaissance and field surveys;
3. Geomorphic analysis and interpretation of historical and contemporary information;
4. Stream classification and assessment of stability status at reach scale;
5. Prediction of past and future morphological evolution and response to proposed project; and
6. Integration of results into engineering design to optimize performance.

For more detailed reviews of practical and procedural issues in geomorphic studies and assessment, the reader is referred to articles by Thorne (1998; 2002).

The results of geomorphic assessment are rarely clear-cut. More often the individual elements of the assessment produce outcomes that are equivocal or even contradictory.

For example, the specific gauge record for a hydrological station may indicate that stages for a given discharge have decreased significantly, whereas the few available repeat cross sections from the same period show variations in bed topography but no evidence for discernible change, and stream reconnaissance indicates that the channel is hydraulically connected to its floodplain. The specific gauge record may suggest that the channel has degraded, but there may be a lack of supporting evidence from resurveyed cross sections and stream reconnaissance that the channel bed and floodplain levels are mutually-adjusted. In these situations, a level of confidence must be assigned to morphological conclusions based on different components of the assessment, based on the quantity, quality, and reliability of the data and the assessor's experience in applying the techniques involved. It may then be possible to reconcile apparently contradictory results by weighing the levels of confidence associated with each one. It is emphasized that sound judgment, based on insight and experience, is essential for accurate geomorphic assessment.

Obviously, geomorphic assessment alone can never provide a proper basis for engineering analysis or design. It is, however, of value when combined with computational and analytical methods for stable channel design. The wider contribution provided by geomorphic assessment is to establish the system context and framework within which the designer may

1. Select hydrodynamic and sediment transport equations appropriate to the stream and conditions;
2. Design stable channel dimensions that mimic natural channel forms and diversity while meeting project goals;
3. Use computer models matched to the alluvial setting and incorporating existing geologic and artificial controls to predict morphological response of the channel system to proposed rehabilitation measures;
4. Integrate environmental features effectively into morphological and engineering aspects of the project;
5. Anticipate maintenance requirements and optimize the design to ensure that the benefits are sustainable; and
6. Consider and propose the scope of post-project appraisal (PPA) and monitoring regime necessary to establish the strengths and weaknesses of project performance.

Geomorphic assessment alone is not sufficient to guarantee that a project will perform adequately with regard to morphological and environmental goals, but it is a valuable and necessary component of the integrated channel design process that is essential to ensure long-term sustainability in river engineering and management projects.

6.10 CLOSURE

Fluvial geomorphology, analytical river mechanics, and sound engineering judgment together provide the foundations for sound river engineering, rehabilitation, and management. Insights and understanding provided by geomorphic principles and identification of causal links between channel form, fluvial processes, and connectivity of the river system can be invaluable in the design of river projects and management strategies. This is the case not only because the engineering geomorphic approach is consistent with environmental goals such as minimizing negative impacts and maximizing biodiversity, but also because solutions that recognize and deal with the causes rather than the symptoms of channel problems represent better engineering. Many engineering stabilization and rehabilitation projects have failed not as the result of deficient hydraulic or structural design, but rather because the significance of geomorphology to the project and the project to geomorphology has not been identified and accounted for in the design. Experience is accumulating that engineering designs guided by knowledge of the fluvial system are able to avoid having to attempt to "tame the river," instead working with the river to produce schemes that have lower long-term maintenance requirements. Engineering-geomorphology thereby opens the door to cost-effective, sustainable solutions that do not commit future generations to heavy and expensive maintenance.

NOTATION

The following symbols are used in this paper:

A	=	area;
A	=	meander amplitude (Leopold et al. 1964);
A	=	bank-full cross-sectional area (in Table 6-3, from FISRWG 1998);
$AR^{2/3}$	=	conveyance;
a	=	coefficient;
B	=	meander belt width (in Table 6-3, from FISRWG 1998);
b	=	exponent;
b_n	=	Manning n (in Table 6-5, from FISRWG 1998);
c	=	coefficient;
D	=	depth (from Kennedy 1895);
D	=	bank-full mean depth (in Table 6-3, from FISRWG 1998);
D_x	=	given sediment size;
D_{50}	=	median size of the bed material (from Lane 1955);
D_{54}	=	bed material (in Table 6-5, from FISRWG 1998);
d	=	average depth of the channel;
dm	=	maximum depth;
F	=	width-to-depth ratio (from Schumm 1977);
f	=	exponent;
h	=	bank height;
h_c	=	critical bank height;

K = channel sinuosity (in Table 6-3, from FISRWG 1998);
k = coefficient;
L = meander wavelength (Leopold et al. 1964);
L_b = along-channel bend length (in Table 6-3, from FISRWG 1998);
L_m = meander wavelength (in Table 6-3, from FISRWG 1998);
M = channel sediment;
M = weighted percentage silt-clay in the channel perimeter (from Schumm 1977);
M = percentage of bank materials finer than 0.074 mm (Table 6-5, from FISRWG 1998);
m = exponent;
n = number;
P = wetted perimeter;
P = sinuosity;
Q = discharge (from Kennedy 1895);
Q = water discharge (from Lane 1955);
Q_s = bed-material load (from Lane 1955);
Q_x = bed-material transport rate in kg s^{1} at water discharge Q (Table 6-5, from FISRWG 1998);
R = hydraulic radius;
R_c = loop radius of curvature (in Table 6-3, from FISRWG 1998);
r_c = radius of curvature;
r_c/w = radius of curvature to width ratio;
S = slope (from Lane 1955);
S = estimate of the standard deviation of the sample (from Yevjevich 1972);
V = velocity (from Kennedy 1895);
W = width (from Schumm 1977);
W = channel width (from Kennedy 1895); and
W = bank-full width (in Table 6-3, from FISRWG 1998):
w = width;
w/d = width-depth ratio;
θ = arc angle;
π = pi.

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