# River Width Adjustment. I: Processes and Mechanisms

By the ASCE Task Committee on Hydraulics, Bank Mechanics, and Modeling of River Width Adjustment[1]

**Abstract:** In 1993 a Task Committee (TC) of the ASCE was established to study the hydraulics, bank mechanics, and modeling of width adjustment in alluvial channels. The work of the TC in reviewing width adjustment processes and mechanisms is reported in this paper. A companion paper presents the findings of the TC with regard to width adjustment modeling. This paper first establishes the geomorphic context within which width adjustments occur, and it demonstrates that width adjustment may take place over a wide range of scales in time and space. In the past engineering analyses of channel width have tended to concentrate on prediction of the equilibrium width for stable channels. Most commonly the regime, extremal hypothesis, and rational (mechanistic) approaches are used, and these are reviewed herein. More recently, attention has switched to channels that are adjusting their morphology either due to natural instability or in response to changes in watershed land use, river regulation, or channel engineering. Characterizing and explaining the time-dependent behavior of width in such channels requires an understanding of the fluvial hydraulics of unstable channels, especially in the near-bank regions. Existing knowledge is reviewed, useful engineering tools are presented, and gaps requiring further field and laboratory research are identified. Finally, this paper considers the mechanics of bank retreat due to flow erosion and mass failure under gravity, and bank advance due to sedimentation and berm building. It is demonstrated that, while rapid progress is being made, most existing analyses of bank mechanics are still at the stage of being research tools that are not yet suitable for design applications. This paper ends with a series of conclusions and recommendations that synthesize the findings of the TC.

## INTRODUCTION

### Background

A variety of methods is used to describe channel morphology and morphological adjustments for river engineering purposes. Available approaches range from equations that predict the regime or graded morphology of equilibrium channels to mathematical models that simulate channel changes in time and space. Most mathematical models, however, neglect time-dependent channel width adjustments and do not simulate processes of bank erosion or deposition. Although changes in channel depth caused by aggradation or degradation of the river bed can be simulated, changes in width cannot. When attempting to model natural systems this is a significant limitation because channel morphology usually changes with time, and adjustment of both width and depth (in addition to changes in planform, roughness, and other attributes) are the rule rather than the exception (Leopold et al. 1964; Simon and Thorne 1996). As a result, our ability to model and predict changes in river morphology and their engineering impacts is limited. This is unfortunate because width adjustments can seriously impact floodplain dwellers, riparian ecosystems, bridge crossings, bank protection works, and other riverside structures, through bank erosion, bank accretion, or bankline abandonment by the active river channel.

Considerable research effort has recently been directed towards improving this situation. In view of this effort, a Task Committee (TC) of the American Society of Civil Engineers was formed in 1993 to address the topic of river width adjustment. The objectives of the TC were as follows:

- Review the current understanding of the fluvial processes and bank mechanics involved in river width adjustment
- Evaluate methods (including regime analysis, extremal hypotheses and rational, mechanistic approaches) for predicting equilibrium river width
- Assess our present capability to quantify and model width adjustment
- Identify current needs to advance both state-of-the-art research and the solution of real world problems faced by practicing engineers

To achieve these objectives TC members first presented individually authored papers in a special session at the 1994 ASCE Hydraulics Conference in Buffalo, New York (Cotroneo and Rumer 1994, pp. 944–984). The TC then prepared two substantial review papers for the *Journal of Hydraulic Engineering.* This paper, which is the first, addresses processes and mechanisms of width adjustment. The second paper (ASCE 1998) addresses width adjustment modeling.

### Geomorphic Context of River Width Adjustment

River width adjustments may occur due to a wide range of morphological changes and channel responses (Figs. 1 and 2). Widening can occur by erosion of one or both banks without substantial incision [Fig. 1(a)] (Everitt 1968; Burkham 1972; Hereford 1984; Pizzuto 1994). Widening in sinuous channels may occur when outer bank retreat, due to toe scouring, exceeds the rate of advance of the opposite bank, due to alternate or point bar growth [Fig. 1(b)] (Nanson and Hickin 1983; Pizzuto 1994) while, in braided rivers, bank erosion by flows deflected around growing braid bars is a primary cause of widening [Fig. 1(c)] (Leopold and Wolman 1957; Best and Bristow 1993; Thorne et al. 1993). In degrading streams, widening often follows incision of the channel when the increased height and steepness of the banks causes them to become unstable [Fig. 1(d)]. Bank failures can cause very rapid widening under these circumstances (Thorne et al. 1981; Little et al. 1982; Harvey and Watson 1986; Simon 1989). Widening in coarse-grained, aggrading channels can occur when flow acceleration due to a decreasing cross-sectional area, coupled with current deflection around growing bars, generates bank erosion [Fig. 1(e)] (Simon and Thorne 1996).

Morphological adjustments involving channel narrowing are equally diverse (Fig. 2). Rivers may narrow through the for-

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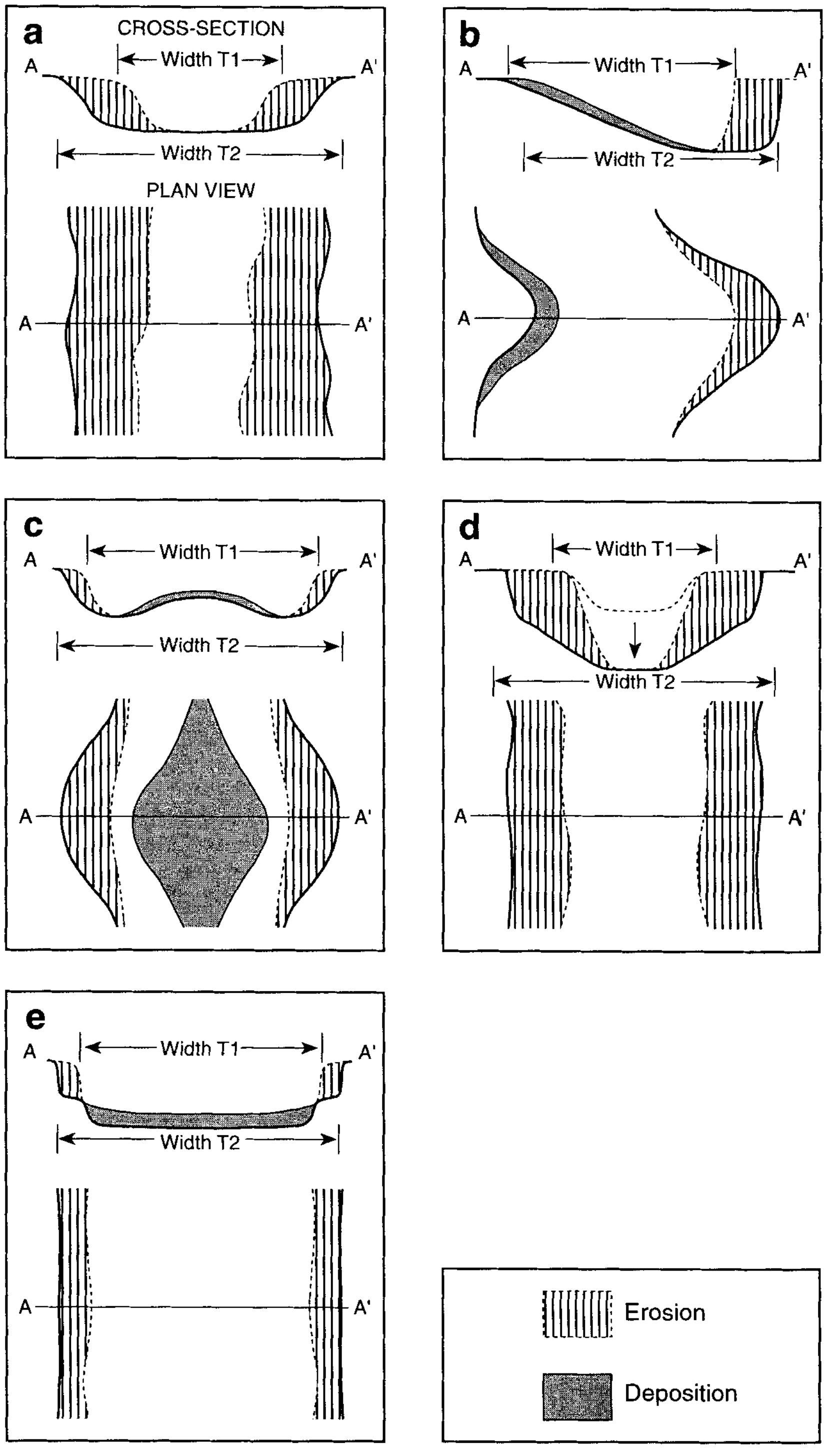


**FIG. 1. Geomorphology of Channel Widening: (a) Channel Enlargement by Bank Erosion without Incision; (b) Erosion of Outer Bank in Sinuous Channel at Faster Rate Than Accretion on Bar Opposite; (c) Deflection of Flows by Growing Braid Bar; (d) Bank Failure and Retreat Due to Mass Instability following Channel Incision; (e) Bank Erosion Due to Flow Acceleration and Deflection in Coarse-Bedded, Aggrading River**

mation of in-channel berms, or benches at the margins [Fig. 2(a)]. Berm/bench growth often occurs when bed levels stabilize following a period of degradation and can eventually lead to creation of a new, low-elevation floodplain and establishment of a narrower, quasi-equilibrium channel (Woodyer 1975; Harvey and Watson 1986; Simon 1989; Pizzuto 1994). Encroachment of riparian vegetation into the channel is often identified as contributing to the growth, stability, and, in some cases, the initiation of berm or bench features (Hadley 1961; Schumm and Lichty 1963; Harvey and Watson 1986; Simon 1989). Narrowing in sinuous channels occurs when the rate of alternate or point bar growth exceeds the rate of retreat of the cut bank opposite [Fig. 2(b)] (Nanson and Hickin 1983; Pizzuto 1994). In braided channels, narrowing may result when a marginal anabranch in the braided system is abandoned [Fig. 2(c)] (Schumm and Lichty 1963). Sediment is deposited in the abandoned channel until it merges into the floodplain. Also, braid bars or islands may become attached to the floodplain, especially following a reduction in the formative discharge [Fig. 2(d)]. Island tops are already at about floodplain elevation and attached bars are built up to floodplain elevation by sediment deposition on the surface of the bar, often in association with the establishment of riparian vegetation. Attached islands and bars may, in time, become part of the floodplain

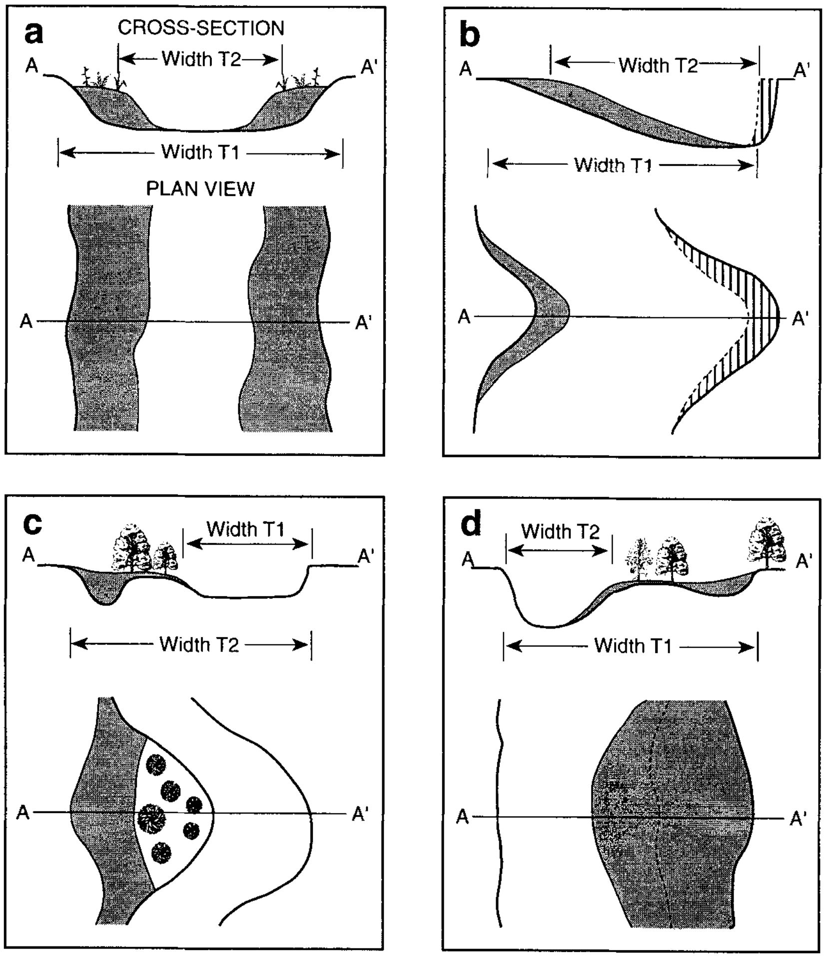


**FIG. 2. Geomorphology of Channel Narrowing: (a) Channel Reduction by Berm or Bench Formation; (b) Accretion on Advancing Bar at Faster Rate Than Erosion of Bank Opposite; (c) Abandonment of Marginal Anabranch in Braided Channel; (d) Closure of Marginal Channel When Braid Bars or Island Becomes Attached to Floodplain**

bordering a much narrower, sometimes single-threaded, channel (Williams 1978; Nadler and Schumm 1981).

If the flow regime and sediment supply are quasi-steady over periods of decades or centuries, the morphology of the river adjusts to create a metastable, equilibrium form (Schumm and Lichty 1965). Such rivers are described as being graded or in regime (Mackin 1948; Leopold and Maddock 1953; Wolman 1955; Leopold et al. 1964; Ackers 1992). Although the width of an equilibrium stream may change due to the impact of a large flood or some other extreme event, the stable width is eventually recovered following such perturbations (Costa 1974; Gupta and Fox 1974; Wolman and Gerson 1978). Unfortunately, predicting the time-averaged morphology of equilibrium channels remains, despite years of effort, a difficult problem (Ackers 1992; White et al. 1982; Ferguson 1986; Bettess and White 1987).

Many rivers, however, cannot be considered to have equilibrium channels even as an engineering approximation. These rivers display significant morphological changes, including width adjustments, when viewed over decades or centuries. For example, rivers in arid and semiarid regions of the American West change their morphologies drastically as the volume of annual precipitation, frequency of flood events, and other factors vary stochastically around steady long-term averages (Schumm and Lichty 1963; Everitt 1968; Burkham 1972; Osterkamp and Costa 1987). Because the width, and other morphological attributes, of these streams vary so dramatically they cannot be considered as graded or regime channels (Stevens et al. 1975) but are perpetually enlarging rapidly in response to a period of relatively high flows, or contracting during periods of less than average runoff (Schumm and Lichty 1963; Stevens et al. 1975; Pizzuto 1994).

Other nonequilibrium rivers may be actively adjusting to changes in flow regime and sediment supply (Andrews 1977; Madej 1977; Smith and Smith 1984), changing valley slope (Patton and Schumm 1975), succession of riparian vegetation (Hadley 1961; Graf 1978), climate change (Schumm 1968; Knox 1983; Hereford 1984), watershed land-use change (Hammer 1972), neotectonic valley floor tilting (Burnet and Schumm 1983; Schumm and Winkley 1994), or sea-level rise (Brammer et al. 1993). The resulting width adjustments can occur at various rates and in different temporal sequences. For example, Hammer (1972) suggested that rivers of southeastern Pennsylvania adjusted to the impacts of urbanization in <5 years, but Andrews (1986), Jacobson and Coleman (1986), and other researchers have documented flood event-induced disruptions to river morphology that will persist for more than a century.

Width adjustments not only encompass a variety of time scales, they are also accomplished by a wide range of fluvial processes and geotechnical mechanisms associated with different discharge, climatic, and environmental conditions. Bank erosion processes provide a useful example. Wolman (1959) noted that significant bank erosion on Watts Branch in the Maryland Piedmont occurred more than ten times per year during relatively small but frequent flow events. However, field scientists working in other alluvial systems have reported that significant bank erosion was caused mostly by large floods with recurrence intervals of decades or centuries (Williams and Guy 1973; Costa 1974; Gupta and Fox 1974; Gardner 1977; Osterkamp and Costa 1987). In other cases, bank retreat has been found to be almost entirely unrelated to flow stage and intensity but more closely correlated with precipitation events and ground-water levels that generate erosion through sapping or piping (Brunsden and Kesel 1973; Ullrich et al. 1986; Hagerty 1991a,b). Thus, identification of the dominant erosion processes and failure mechanisms and selection of the appropriate discharge or climate events to be included in either conceptual or mathematical width adjustment models remain very difficult tasks.

These few examples indicate that channel changes involving width adjustment occur in a wide variety of geomorphic contexts, that width adjustment will usually be accompanied by changes in other morphological parameters such as channel depth, roughness, bed material composition, riparian vegetation, energy slope, and channel planform, and that the processes responsible for width adjustment are diverse. Furthermore, adjustment processes display a variety of spatial and temporal patterns and operate over a wide variety of time scales. Because of this diversity, it is unlikely that a single method can be developed to predict the trends and rates of width adjustment for all rivers. Therefore, engineers must establish the morphological context of width adjustment and identify the major processes and mechanisms involved before selecting appropriate methods for analysis, modeling, and solution of width adjustment problems. This is best achieved through systematic stream reconnaissance of the bank and its context within the fluvial system (Thorne 1993; Simon and Downs 1995). Bank reconnaissance is considered more fully in the accompanying paper in the section titled "Rapid Assessment Techniques."

## EQUILIBRIUM APPROACHES

The hydraulic geometry of alluvial rivers and canals that have attained a steady-state morphology is considered in this section. Such information is especially relevant for the design of irrigation canals and channelization schemes and for predicting the response of a stream to flow regulation. Existing methods describing equilibrium river morphology can be grouped into empirical (regime theory and power laws), extremal hypotheses, and rational or mechanistic approaches.

### Regime Theory and Power Law Approach

The initial approach to prediction of equilibrium channel form was based on empirical methods, developed from field observations and regression equations and applied to the design of stable channels. The first regime relation was proposed by Kennedy (1885) over a century ago. Several regime relations followed, and these have been repeatedly refined and enhanced. The regime equations attributed to Lindley (1919), Lacey (1920), Simons and Albertson (1963), and Blench (1969) are probably the most widely known.

While regime equations are extensively used by engineers with successful outcomes, they suffer several shortcomings, including the facts that they are not dimensionally homogeneous and that their validity is limited to the basins and data from which they were derived. More sophisticated regime relations have been proposed by employing computers to obtain regression equations based on much larger data sets (Brownlie 1983). Most work, including that previously cited, pertains to sand-bed streams, but equivalent regime relations have also been proposed for gravel-bed streams; reviews are presented by Bray (1982) and Hey and Thorne (1986). More recently, semianalytical work by Julien and Wargadalam (1995) has attempted to refine the regime approach within a framework based on the governing principles of open channel flow.

Geomorphologists have used data from natural streams and laboratory flumes to develop power-law hydraulic geometry relations between channel top width, average depth, average velocity, and bank-full discharge (Leopold and Maddock 1953). The exponents in these relations exhibit surprising universality, particularly the one for channel width, which has been found to be ~0.5 for rivers with widely varying flow regimes and sediment characteristics located in different physiographic regions of the world. However, the regression coefficients are found to vary significantly from one locality to another, which renders power-law hydraulic geometry relations inappropriate as tools for general design purposes. It is worth mentioning that hydraulic geometry relations developed by geomorphologists have independently affirmed several trends shown in the corresponding regime relations.

The relevant empirical formulas developed in these approaches, as well as others that are not mentioned here due to lack of space, are described in detail in standard river mechanics books [e.g., Garde and Ranga-Raju (1977) and Simons and Senturk (1992)] and earlier review papers [e.g., Ferguson (1986)].

### Extremal Hypothesis Approach

The last two decades have seen the proliferation of approaches that employ an extremal hypothesis as part of their formulation for predicting channel morphology. Equations for sediment transport and alluvial friction are combined with a third relationship to determine channel width and to predict regime or equilibrium conditions. This third relationship has frequently been expressed in terms of the maximization or minimization of a parameter, such as stream power, energy dissipation rate, or sediment concentration. Extremal hypotheses have been introduced by Chang (1980, 1988a), Yang et al. (1981), Yang and Song (1986), Bettess and White (1987), Yang (1992), Chiu and Abidin (1995), and Millar and Quick (1997), among others. An extremal hypothesis, based on stream power, also forms the basis of the analytical approach to the river regime of White et al. (1982) and the Wallingford tables for the design of stable channels (White et al. 1981).

The theoretical justification for such hypotheses and the relationships between them are still not entirely clear. Also, when applying extremal hypotheses, a clear understanding is required of the physical constraints presented by geological or other boundary conditions on the evolution of a channel toward a form that minimizes its rate of energy expenditure. The predictions based on such methods, however, provide global, if not exacting, agreement with a wide range of observations.

### Tractive Force Methods

Mechanistic, or tractive force, methods employ the basic laws of mechanics to obtain expressions that specify the geometry of stable channel cross sections. This approach was initiated in the late 1940s by the U.S. Bureau of Reclamation, and it resulted in the threshold channel theory (Glover and Florey 1951; Lane 1955). The theory is based on a fluid momentum balance to obtain the local boundary shear stress and a stability criterion for the sediment particles that make up the channel perimeter. It assumes that the channel is straight, that secondary flow is negligible, and that sediment is noncohesive

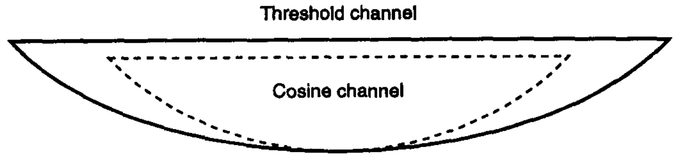


**FIG. 3. Comparison between Threshold Channel Profile Obtained from Momentum-Diffusion Model and Cosine Profile [Adapted from Vigilar and Diplas (1997)]**

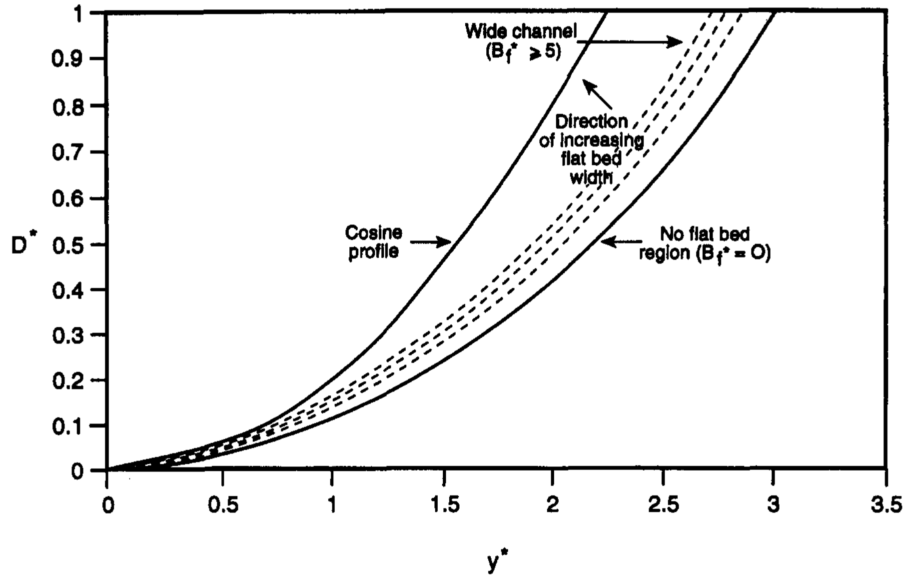


**FIG. 4. Bank Profiles Generated by Momentum-Diffusion Model for Different Values of Flatbed Width of Channel**

and does not vary within the channel. With these assumptions a cosine profile is predicted for the stable cross section (Fig. 3).

In theory, a threshold channel does not allow for bed-load transport (Diplas 1990), and Parker (1979) showed that the Glover and Florey method cannot be extended to generate channels capable of transporting sediment while they maintain threshold banks. This result is contrary to numerous observations from natural streams and flume experiments, which attest to the possible coexistence of a mobile bed and stable banks. Parker (1978) overcame this inconsistency by employing the momentum balance of Lundgren and Jonsson (1964), which accounts for lateral turbulent diffusion of downstream momentum. Due to the complexity of the corresponding differential equation, his solution was limited to the flatbed region, while the bank geometry was solved as a first-order solution, yielding a cosine profile. Thus, Parker was able to reconcile the existence of sediment movement within a stable channel. Ikeda et al. (1988) extended the results of Parker (1978) to include sediment heterogeneity, Ikeda and Izumi (1990) considered the effect of bank vegetation, while Parker (1978) and Ikeda and Izumi (1991) examined the influence of suspended load on channel dimensions. The results are in general agreement with data from natural streams and laboratory flumes.

The tractive force model, in the form proposed by Parker, was recently refined by Diplas and Vigilar (1992). The main differences from the previous work were that the governing equations were solved numerically and that the bank geometry was not assumed but became part of the solution. As a result, the threshold channel shape turned out to be different to a cosine curve, having a larger top width and center depth (Diplas and Vigilar 1992; Vigilar and Diplas 1994, 1997). For the example shown in Fig. 3, the longitudinal slope is 0.00081, the value of the critical Shields parameter is 0.056, and the sediment is semiangular, with $D_{50}$ = 45 mm and $D_{90}$ = 75 mm. The cross-sectional area of the threshold channel and the water discharge that it conveys are more than twice the values corresponding to a cosine channel under the same conditions. This is attributed to the role of momentum diffusion, which results in decreased stresses in the central region of the channel (thus allowing for a deeper flow) and increased stresses in the upper bank regions (forcing banks to assume gentler slopes to prevent erosion). Knowledge of the local topography, sediment size and shape, and the value of the critical Shields parameter uniquely determines the dimensions of a threshold channel and its discharge.

In the case of a channel with stable banks and a mobile bed the bank profile changes with the width of the flatbed section (Fig. 4) (Vigilar and Diplas 1997). However, beyond an aspect ratio of 12, which is typical of natural streams, the bank profile remains constant, and the channel is termed "wide." The stable channel dimensions and bed-load transporting capacity can be determined for known local bed slope, sediment size and shape, value of the critical Shields parameter, and water discharge. If the bed-load discharge is specified, the channel bed slope becomes part of the solution.

## FLUVIAL HYDRAULICS

### Fluvial Complexity

The hydraulic phenomena significantly responsible for width adjustment in nonequilibrium streams are briefly described in this section. The interactions between near-bank fluvial processes and bank materials are themselves considered more fully in the accompanying paper in the section titled "River Bank Mechanics."

In rivers, the bed topography, resistance to flow (roughness), sediment type, and the water and sediment motion are usually all highly variable in space and time. In this section specific sources of variability are reviewed.

Topographic variability includes not only complexity in the shape of individual cross sections but also streamwise variations of both cross-sectional shape and planform geometry. The effects of these channel features may change markedly when the stream flows out-of-bank at times of flood. The high discharges associated with over-bank flow may be especially influential in producing geomorphological change due to higher, and in some cases lower, than usual velocities and boundary shear stresses in the channel, which may lead to patterns of erosion and deposition that differ radically from those associated with in-bank flows.

Roughness variability occurs not only due to the variation of hydraulic resistance around the wetted perimeter of an individual cross section, which arises from the nonuniform lateral distribution of grain size and the influence of the cross-sectional geometry, but also due to alteration of local resistance coefficients as bed forms (ripples and dunes in sand-bed rivers, pebble clusters in gravel-bed rivers) adjust according to changes in depth and/or unit discharge.

Sediment variability results from the heterogeneous nature of streambed materials that can lead to armoring, through selective entrainment, or fining, through selective deposition.

Also, the properties of bank materials and bank-derived sediments may differ markedly from the bed material. Intact bank materials may be cohesive or noncohesive and may possess other important geotechnical attributes that vary widely over short distances.

Flow and sediment transport variabilities are complex because the flow in natural channels is usually unsteady and fully tubulent, with coherent flow structures that are not presently amenable to complete mathematical analysis. The application of turbulence models is therefore difficult (Rodi 1980; ASCE 1988; Nezu and Nakagawa 1993; Cokjlat and Younis 1994a,b; Thomas and Williams 1995). Consequently, the length and time scales in the formulation of the problem are unknown. Also, the sediment load is often supply-limited and is not predictable on the basis of limiting hydraulic transport capacity alone. It is difficult, therefore, to link flow magnitude to morphological change, because floods of similar size may transport widely differing amounts of sediment depending on sediment availability at the time of the event (Newson 1980; Beschta 1987).

Finally, in natural systems the bed topography, sediment characteristics, boundary roughness, discharge capacity, and channel form are mutually adjustable and interrelated. Detailed values for each are a unique product of the historical sequence of past events in forming and modifying the channel. Bearing in mind these complexities, it is little wonder that understanding of bank erosion, bank accretion, and width adjustment are in their infancy.

The flow of water and sediment in rivers is described by Newton's laws, which are straightforward until turbulence is to be represented. The simplified, one-dimensional (1D) St. Venant equation is commonly applied in river engineering because the flow can be considered to occur predominantly in the downstream direction. An appropriate rigid boundary resistance law is then usually adopted to relate the conveyance capacity with geometry [see for example, Keulegan (1938), ASCE (1963), Chow (1959), Cunge et al. (1980), and Yen (1993)]. For alluvial channels the resistance law must also take into account the additional energy losses arising from bedforms (Engelund 1966; Alam and Kennedy 1969; Garde and Ranga-Raju 1977; White et al. 1980; van Rijn 1984). For meandering channels additional resistance terms are required for form drag due to channel curvature (Nelson and Smith 1989).

A 1D representation is simple but is inadequate to define the processes and the mechanics of river width adjustment. A three-dimensional (3D) formulation is better able to deal with the complexity of channel width adjustment, but 3D models are so complex to solve that at present they are of little value in practice except in the most well funded of projects. A two-dimensional (2D) representation (depth integration of 3D equations) can be solved more readily, but it is still limited because it is not strictly applicable in the near-bank zone and does not give any motion in the vertical direction. Vertical motion is considered to be particularly important at river bends, where most bank erosion and bar deposition occurs. However, a 2D formulation is still frequently used because it yields useful information about the lateral variation of most of the important hydraulic parameters that impact bank erosion and accretion. A detailed review of those aspects of channel hydraulics that directly influence width adjustment is given here. A wider, general review of river and floodplain hydraulics is given elsewhere by Knight and Shiono (1995).

## Channel Boundary

### *Cross-Sectional Shape*

The shape of a river cross section influences the isovel, secondary flow, and boundary shear stress distributions in a number of ways. A typical example (Fig. 5) is a rectangular cross section in a straight river with vertical banks. The data are from a rectangular duct experiment with an equivalent open channel width/depth ratio of 20. Even in this relatively wide case the isovels and boundary shear stress distribution indicate the presence of secondary flow cells and 3D effects in the near-bank zone, which in this case is 15% of the channel width. Fig. 6 shows isovels and boundary shear stresses for flow in a narrow trapezoidal channel with an aspect ratio of 1.5. In this case the narrowness of the channel causes the flow in the entire cross section to be influenced by 3D flow structures, unlike the case shown in Fig. 5. In the wide channel case, flow in the central region is almost 2D provided that $y/2b < 0.85$, and for this condition standard boundary layer distributions may be assumed for velocity and Reynolds stress. However, it should not be forgotten that even in a wide ($B/H > 20$) channel, although it may be acceptable to ignore bank effects for many hydraulic and geomorphic analyses, the banks still influence the flow in the near bank zones sufficiently to require that the resulting 3D flow structures are accounted for in models of width adjustment.

Secondary flow cells may be generated by anisotropic turbulence (stress-induced secondary currents) or streamwise curvature (skew-induced secondary currents) and are always present in any turbulent flow along a channel with a noncircular cross section, such as a natural river channel (Einstein and Li 1958; Liggett et al. 1965; Tracy 1965; Perkins 1970; Melling and Whitelaw 1976; Chiu and Hsiung 1981; Naot and Rodi 1982; Nezu 1993; Meyer and Rehme 1994). In straight channels, stress-induced secondary velocities are usually small, typically being 1–2% of the primary velocity. Modeling these weak motions is especially difficult in complex cross sections of the type that characterizes natural channels. In meandering channels, skew-induced secondary velocities may be as great as 10–20% of the primary flow, and they are known to affect the distributions of primary velocity and bed-shear stress sig-

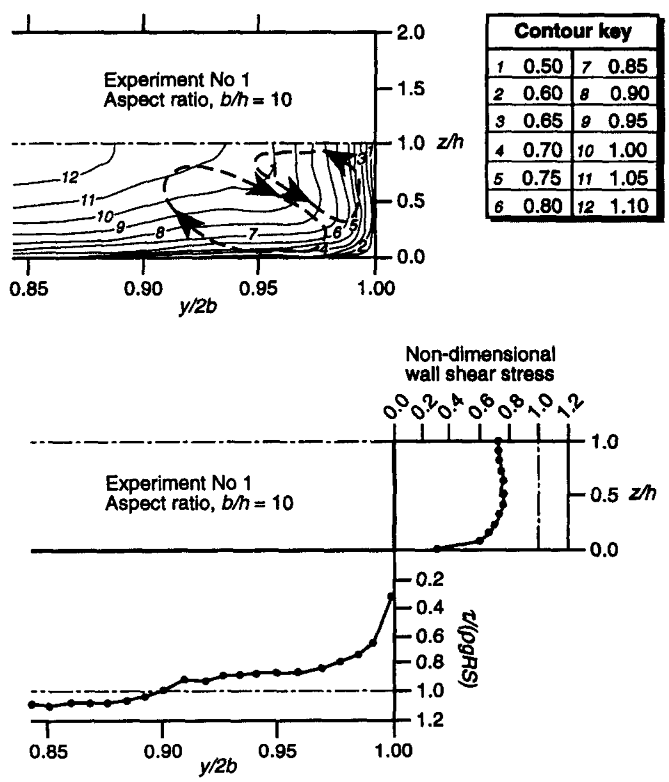


**FIG. 5. Typical Influence of Vertical River Bank on Velocity and Boundary Shear Stress in Wide Rectangular Channel (Rhodes and Knight 1994)**

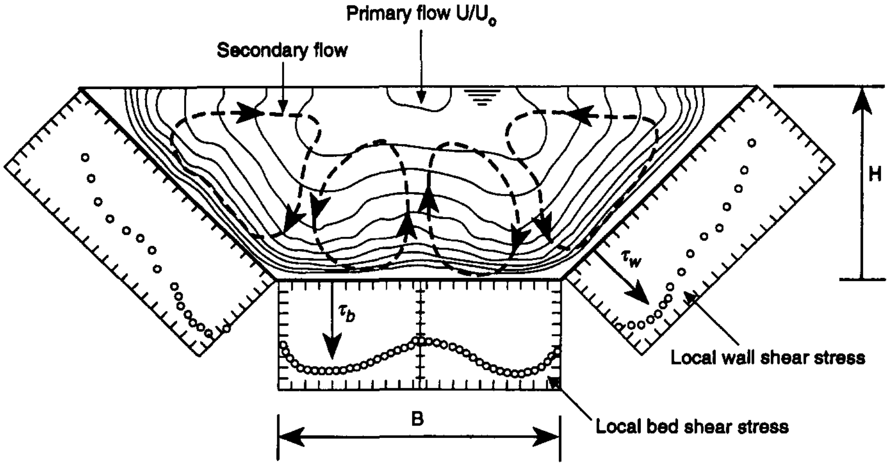


**FIG. 6. Typical Relationship between Boundary Shear Stress Distribution, Secondary Currents, and Primary Velocities in Trapezoidal Channel. Froude Number = 3.24, Width/Depth Ratio = 1.52 [Adapted from Knight et al. (1994)]**

nificantly (Bathurst et al. 1979; Ikeda and Parker 1989; Nelson and Smith 1989; Shiono and Muto 1993; Knight and Shiono 1995).

River engineers are often concerned with the parameters at the channel boundary. A depth-averaged form of the streamwise equation of motion for flow in a straight channel is given by Shiono and Knight (1988, 1991) in the form

$$\rho g H S_0 - \frac{1}{8}\rho f U_d^2\left(1 + \frac{1}{s^2}\right)^{1/2} + \frac{\partial}{\partial y}\left\{\rho\lambda H^2\left(\frac{f}{8}\right)^{1/2} U_d\frac{\partial U_d}{\partial y}\right\} = \frac{\partial}{\partial y}[H(\rho UV)_d] \quad (1)$$

where $\rho$ = water density; $g$ = acceleration due to gravity; $H$ = water depth; $U$ = long-stream velocity; $V$ = cross-stream velocity; $y$ = lateral distance across the channel; $S_0$ = streamwise channel slope; $s$ = local channel side slope of the banks; and $U_d$ = depth-averaged mean velocity defined by

$$U_d = \frac{1}{H}\int_0^H U\,dz \quad (2)$$

Three coefficients, $f$, $\lambda$, and $\Gamma$, are introduced to deal with the local friction factor, dimensionless eddy viscosity, and secondary flow parameter, defined respectively by

$$\tau_b = \left(\frac{f}{8}\right)\rho U_d^2; \quad \bar{\tau}_{yx} = \rho\bar{\varepsilon}_{yx}\frac{\partial U_d}{\partial y};$$

$$\bar{\varepsilon}_{yx} = \lambda U_* H; \quad \frac{\partial H(\rho UV)_d}{\partial y} = \Gamma \quad (3a\text{–}d)$$

where $\bar{\varepsilon}_{yx}$ = depth-averaged eddy viscosity; and $\tau_b$ = local boundary shear stress. Eq. (1) gives lateral distributions of $U_d$ or $\tau_b$ across the channel width, provided that appropriate values of $f$, $\lambda$, and $\Gamma$ are specified for each boundary element. Applications of this model to both in-bank and over-bank flow are described in Shiono and Knight (1990), Abril (1995), and Knight and Abril (unpublished paper, 1996).

The depth-mean apparent shear stress $\tau_a$ acting on a vertical plane in the streamwise direction of a river cross section may be determined by laterally integrating (1) to give

$$\tau_a = -\frac{1}{H}\int_0^y\left[\rho g H S_0 - \tau_b\left(1 + \frac{1}{s^2}\right)^{1/2}\right]dy \quad (4)$$

It should also be noted from a comparison between (1) and (4) that the depth-averaged apparent shear stress has two distinct components, one arising from depth-averaged secondary flow motion and the other from turbulent Reynolds stresses. It is this apparent shear stress that is often required in sediment transport models, as it links together the local boundary shear

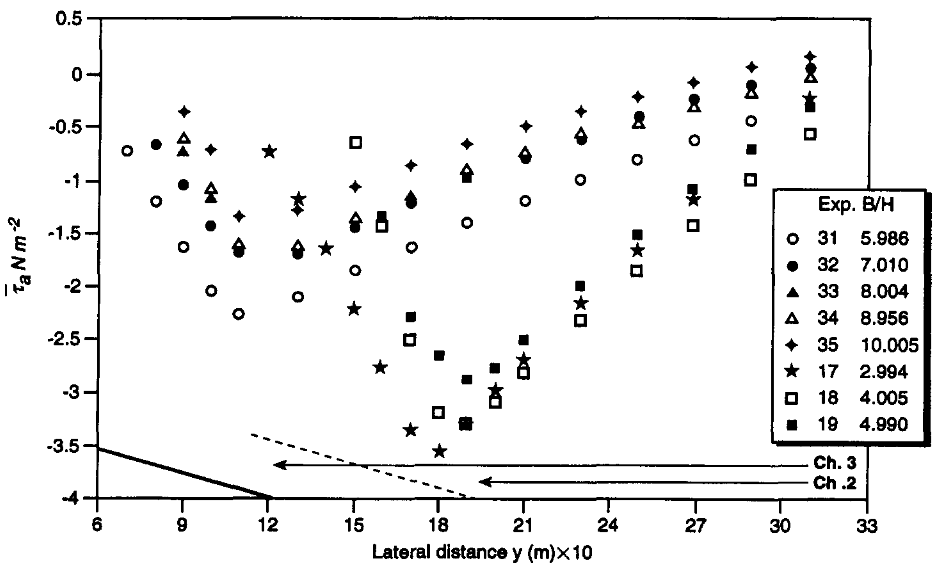


**FIG. 7. Lateral Variation of Depth-Averaged Apparent Shear Stress $\tau_a$ for Trapezoidal Channels with Roughened Walls and Smooth Bed**

stress at a point on the wetted perimeter with the streamwise resolved weight force and the resultant net stress. Some values of $\tau_a$ are shown in Fig. 7 for trapezoidal channels with roughened banks and width/depth ratios between 4 and 10.

Significantly, the right-hand side of (1) contains a depth-averaged secondary flow term. This term is often ignored in stream-tube models and in some eddy viscosity/width-adjustment models (Wark et al. 1990; Darby and Thorne 1992; James and Wark 1994; Kovacs and Parker 1994). However, this term is, in fact, important as is well illustrated by the examples of Knight and Abril (unpublished paper, 1996) using benchmarked experimental data from the U.K. Flood Channel Facility (Knight and Sellin 1987; HR Wallingford 1992).

From the point of view of sediment transport, although it is known that the entrainment and motion of grains may be correlated with turbulent bursts and sweeps (Jackson 1976; Raudkivi 1995), inclusion of burst and sweep phenomena in a practical model of width adjustment is at this stage premature owing to our lack of knowledge concerning all the details of coherent structures in the boundary layer close to the bed (Tehrani 1992; Ashworth et al. 1996). As a result, one of the flow parameters still most closely associated with sediment motion is the local time-averaged boundary shear stress. Many researchers have attempted either to predict or measure the lateral distribution of local time-averaged boundary shear stress around the wetted perimeter of channels of various shapes and the longitudinal distribution of local boundary shear stress over sand dunes or in channel reaches. These studies have usually been conducted at laboratory scale using, for convenience, rectangular, trapezoidal, or lenticular shaped cross sections (Engelund 1964; Lundgren and Jonsson 1964; Knight et al. 1994; Rhodes and Knight 1994), and only a few studies have been undertaken in the field at full scale (Bathurst et al. 1979; Dietrich and Whiting 1989; Nece and Smith 1970).

While the applicability of laboratory-based work to field situations is limited, at present no other way exists in which certain fundamental issues can be studied except under carefully controlled laboratory conditions, particularly those issues relating to flood phenomena. Available information, therefore, tends to be biased towards laboratory data. Figs. 8 and 9 illustrate how the average wall or bank stress $\tau_w$ and bed shear stress $\tau_b$ vary for uniformly roughened trapezoidal channels (side slope angles of 45, 68, and 90°) and how they compare with simple exponential equations, which are adequate for design purposes. These data all relate to in-bank flows. Similar plots and equations are available for the maximum stresses on the bed and banks, together with their locations, for both uniformly and nonuniformly roughened channels. High differential roughness may typically occur in engineered channels (with portions of the wetted perimeter especially rough due to riprap) or in natural channels (with banks that are significantly rougher than the bed due to dense riparian vegetation).

Since lenticular shapes more closely approximate the shape of natural alluvial channels they have often been the focus of river studies (Lundgren and Jonsson 1964; Ikeda 1981; Kovacs and Parker 1994). Five methods for determining the local boundary shear stress were reviewed by Lundgren and Jonsson (1964) with the area method being found to be most suitable for general use. Distributions of boundary shear stress based on a particular lenticular shape (Fig. 10) are shown for all five

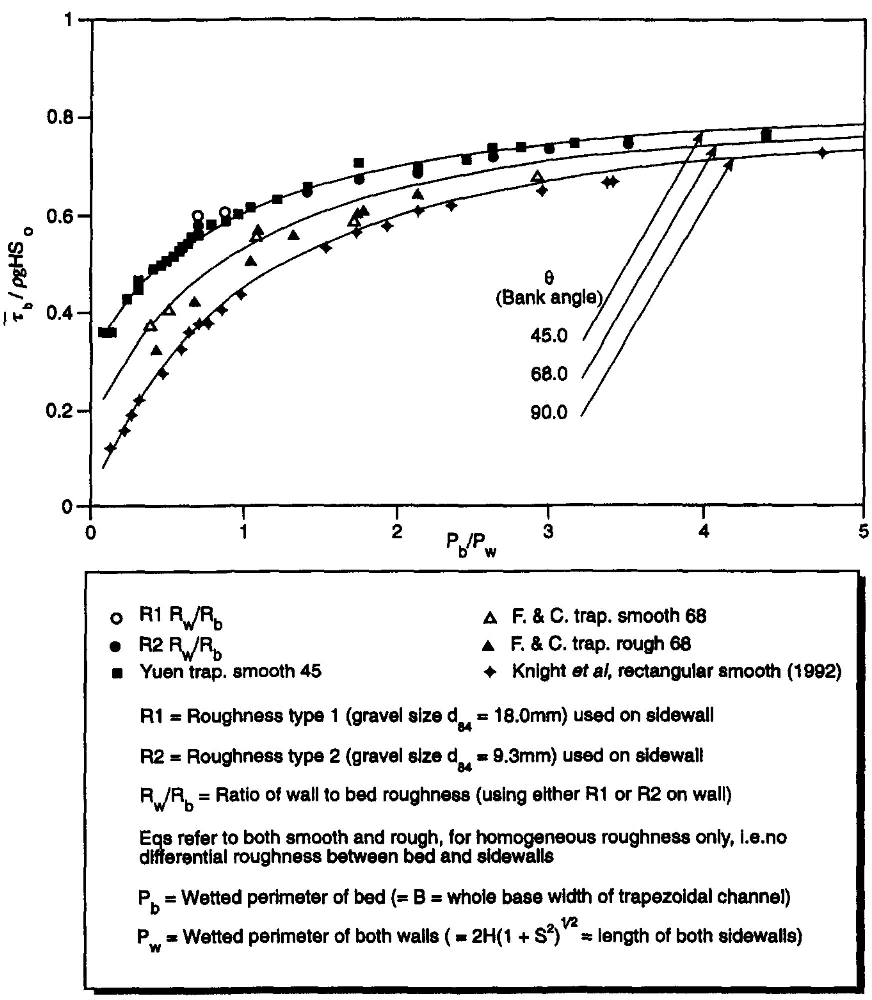


**FIG. 8. Average Wall Shear Stress $\bar{\tau}_w/\rho gHS_0$ for Smooth and Rough Trapezoidal Channels with Different Side Slope Angles**

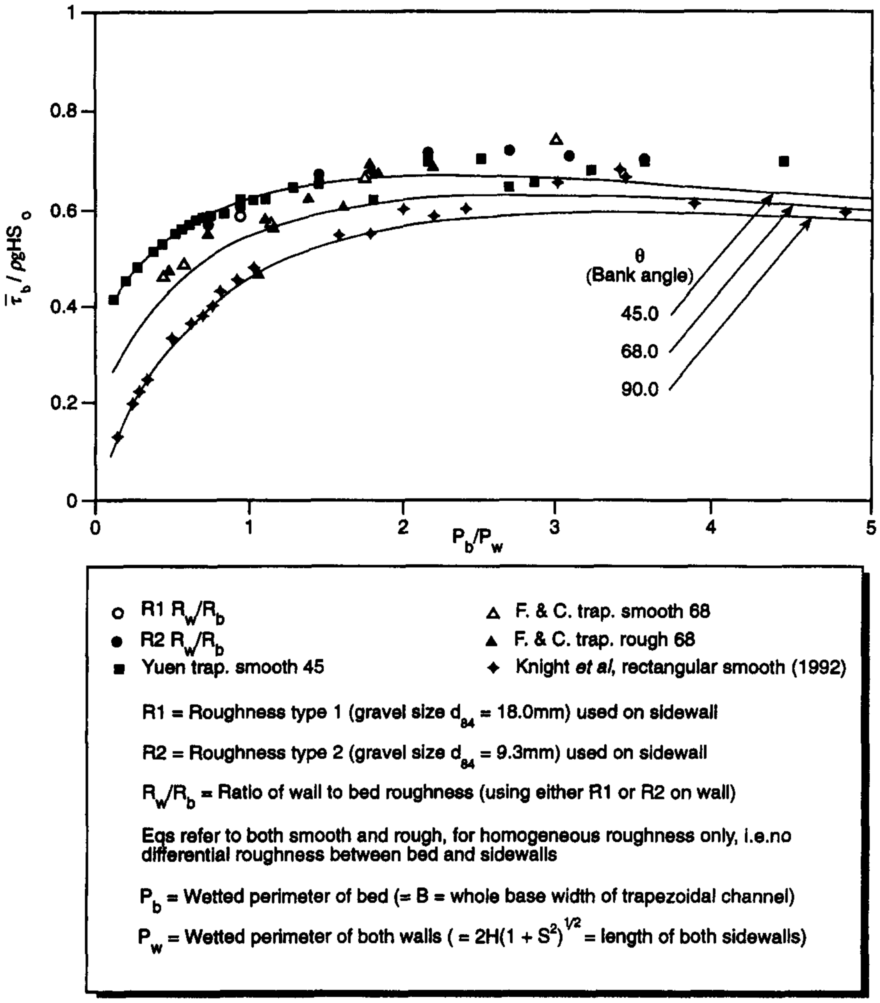


**FIG. 9. Average Bed Shear Stress $\bar{\tau}_b/\rho gHS_0$ for Smooth and Rough Trapezoidal Channels with Different Side Slope Angles**

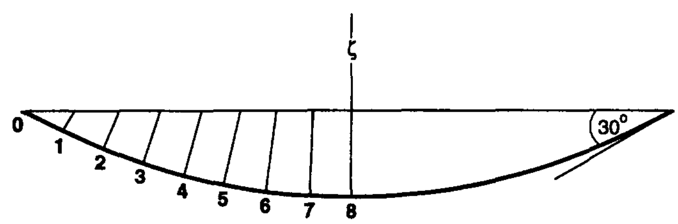


**FIG. 10. Cross Section Used in Numerical Example (Circular)**

methods in Fig. 11. The way in which a river bank might be eroded into an equilibrium shape under the action of these distributions of applied shear stress is illustrated in Figs. 12–14, taken from Kovacs and Parker (1994). In these sketches the lateral distribution of streamwise boundary shear stress is of special significance since those regions of the cross section in which the stress is above transport threshold will become active with regard to bed material transport and will form the active width. The resulting cross-sectional shapes for straight alluvial channels under both threshold and mobile bed conditions have also been the subject of much investigation, as reported in the previous section. However, it should be noted that, since the discharge in a natural river is highly variable, the actual channel cross-sectional shape is constantly responding to the changing stage, possibly with associated alteration to the bed forms and, hence, channel resistance. Hence, it is only possible to predict medium-term (5–10 year), time-averaged, cross-sectional shapes, unless detailed records of flow and stage changes at a section are available.

For over-bank flows in straight and meandering channels, considerably less experimental data is found with regard to both velocity and boundary shear stress (Knight and Demetriou 1983; Knight et al. 1989, 1990, 1992; Tominaga and Nezu 1991, 1993; Ackers 1992, 1993; Sellin et al. 1993). In general terms, for straight channels with floodplains the boundary shear stresses under over-bank flows vary in a more complex way than for in-bank flows, with stresses in the main river channel decreasing because of the influence of the slower floodplain flows. Conversely, the floodplain boundary shear stresses rise above their expected 2D values because of the effect of the faster flowing, main river flow. Interaction between channel and floodplain flows results in some localized and complex effects in the vicinity of the main channel river banks. These features are discussed by Knight and Cao (1994) in relation to large-scale experimental studies and by Knight and Shiono (1995) in relation to the relative importance of the three terms in (1).

The redistribution of boundary shear stress within the cross section during over-bank flow also has a profound effect upon sediment transport rate, as shown by Ackers (1992), Abril (1995), and Knight and Abril (unpublished paper, 1996). It will consequently also have an effect on bank adjustments that occur when the river flows in an over-bank condition.

## Longitudinal Changes

Unlike artificially constructed channels, river channel cross sections are not generally uniform trapezoids. Consequently, width, depth, slope, and planform change significantly in the streamwise direction. Some schematization of the natural river is therefore necessary before a numerical model is constructed (Samuels 1990) and, inevitably, some streamwise averaging of

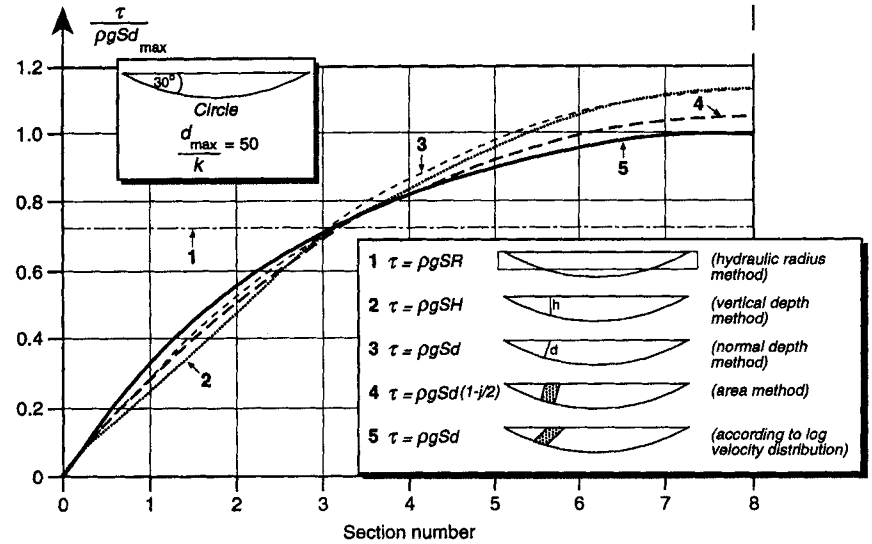


**FIG. 11. Comparison between Five Different Methods of Determining Shear Stress Distribution**

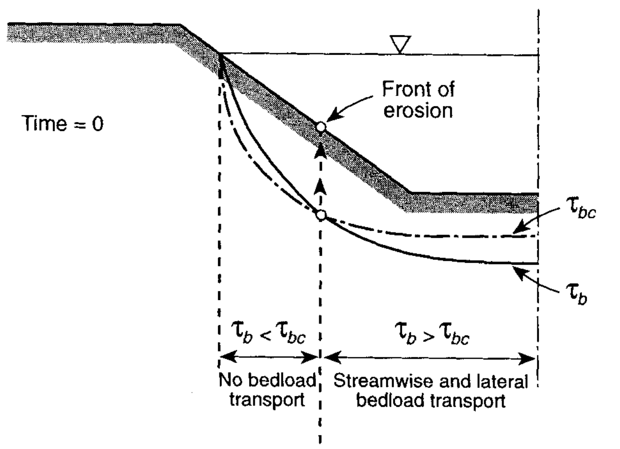


**FIG. 12. Definition of Front of Erosion for Initially Trapezoidal Cross Section at Time $\tau = 0$**

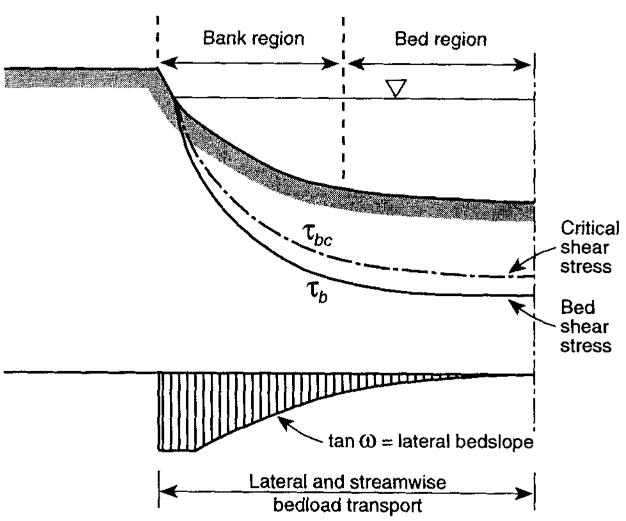


**FIG. 13. Sketch of Distribution of Shear Stress $\tau_b$ and Lateral Bed Slope tan ω along Perimeter of Straight Channel Cross Section during Development of Stable Profile**

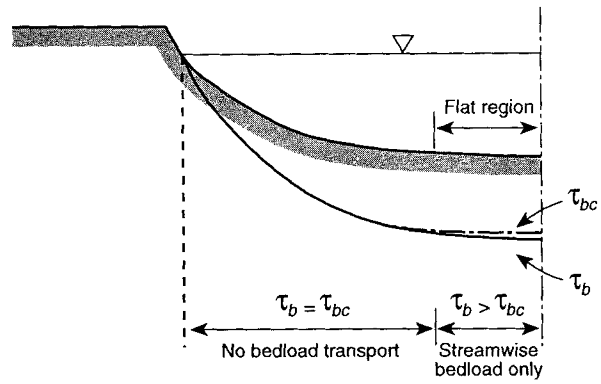


**FIG. 14. Shear Stress Distribution in State of Dynamic Equilibrium**

cross-sectional area, hydraulic radius, energy gradient, mean boundary shear stress, etc. must be performed. This is potentially a source of considerable error in any representation of the channel hydraulics and must be treated with appropriate care (McBean and Perkins 1975; Cunge et al. 1980; Laurenson 1986).

Samuels (1989, 1990) used perturbation analysis of the steady flow equation to show that weighting the friction slope towards the upstream section gives considerably improved accuracy. For example, the longitudinal spacing between channel cross sections may be doubled if weighting coefficients are used rather than the arithmetic mean, with the same accuracy in water levels being achieved. However, the correct representation of the energy gradient or water surface slope is difficult to achieve in natural channels where width, pool-riffle sequences (especially in gravel-bed rivers), and pool-crossing geometry (in sinuous rivers) introduce marked channel variability. In these cases channel schematization, even using a weighting technique, may be inadequate. Further work is needed on the derivation of representative, reach-averaged parameters and their significance in 1D models.

The longitudinal energy gradient is an important parameter describing the work done by the river per unit channel length, and it appears in the stream power per unit channel length parameter, $\rho gQS_f$. In the extremal hypothesis approach described in the previous section, this parameter is minimized to allow closure of the equations for equilibrium channels. This is a specific case of the more general theory of minimum energy dissipation rate.

The values selected to represent the streamwise friction slope $S_f$ and schematized hydraulic radius $R$ are also important because they contribute to the streamwise reach-averaged boundary shear stress $\tau_0$ ($=\rho gRS_f$). Hence, the streamwise averaging of either $S_f$ or $R$ becomes relevant, as does the neglect or inclusion of various streamwise acceleration terms in the governing equations (Nelson and Smith 1989). The expression for the longitudinal variation of stream power involves at least two elements, one for width and the other for depth (Cao and Chang 1988; Cao and Knight, unpublished paper, 1996). It therefore follows that width adjustments should not be considered in isolation at a particular cross section, but some longitudinal balance should be considered. This is particularly important in sinuous channels but may also be important in straight channels.

In meandering alluvial rivers boundary shear stress distribution, bed topography, and channel cross-sectional shape all vary considerably over the meander wavelength due to flow curvature effects (Dietrich and Whiting 1989; Knight et al. 1992). These variations are considerably more complex during over-bank flow. Laboratory studies (Tominaga and Nezu 1991, 1993; Sellin et al. 1993; Wark et al. 1994; Knight and Shiono 1995) show that new flow structures are introduced during over-bank flow, and field studies (Fukuoka 1993, 1994; Lawler 1993a) indicate that in the natural environment over-bank flows are strongly influenced by local morphological features. This makes shrewd engineering judgment essential in schematization of the channel at reach and grid scales.

## Near-Bank Zone

Knowledge of the velocity, boundary shear stress, secondary flows, and turbulence structure close to a river bank is required before fluvial processes can be linked to channel width adjustment. The preceding sections have highlighted current difficulties in predicting the details of near-bank hydraulics, even for relatively simple prismatic channels, whether flowing in-bank or over-bank. The 3D nature of near-bank flow leads to nonlinear distributions of Reynolds stresses normal to the boundary and velocity profiles that are not logarithmic (Knight and Shiono 1990, 1995; Shiono and Knight 1991; Meyer and Rehme 1994). The effect of these flow complexities on boundary shear stress distributions in the vicinity of river banks has been illustrated by Knight and Cao (1994).

At present no simple formulas exist to characterize the lateral distribution of local time-averaged boundary shear stress around the wetted perimeter of a natural channel, although Figs. 8 and 9 give some guidance, and experimental data, such as that shown in Figs. 5 and 6, indicate how local values may vary about the cross-sectional mean. Consequently, boundary shear stresses in the near-bank zone have either been estimated from trends established experimentally, like those shown in Figs. 5–11, or been obtained from turbulence models, of which the nonlinear $k$-$\varepsilon$ and large eddy simulation models are probably the most appropriate (Rodi 1980; Nezu and Nakagawa 1993; Thomas and Williams 1995; Younis 1996). Assuming that these local values are known, Figs. 12–14 show how the boundary shear stress and bed-load region are often conceptualized in a typical width adjustment model (Ikeda and Izumi 1991; Kovacs and Parker 1994; Knight and Yu 1995). It should be remembered, however, that although the boundary shear stress is arguably one of the more important connecting links between the flow field and the distributions of erosion (or scour), deposition, and channel change (Breusers and Raudkivi 1991), it is not necessarily the dominant parameter responsible for bank erosion or width adjustment. Process dominance also depends on the geomorphic context within which width adjustment is taking place, the channel type (Fukuoka et al. 1993), and location within the watershed (Lawler 1992).

## Adjustment of Channel Boundaries in Near-Bank Zone

The adjustment of alluvial channel boundaries is usually related to spatially averaged hydraulic parameters, such as boundary shear stress, streamwise stream power, and energy gradient, together with data defining the net sediment supply to the system and the bank material properties (Molinas and Yang 1986; Hasegawa and Mochizuki 1987; Chang 1988b; Hasegawa 1989; Wiele and Paola 1989; Pizzuto 1990; Lawler 1993a; Parker 1995). The variability of channel morphology and complexity of the turbulent flows described earlier might suggest that a probabilistic approach is more suitable than the deterministic treatments described previously. Whichever approach is selected, the same key hydraulic parameters are still likely to be included in process equations or functions and should be represented as faithfully as possible, despite some implicit longitudinal smoothing of localized flow structures or morphological features. The appropriate inclusion of 3D phenomena into either a 1D system model or a depth-averaged model is still awaited and is likely to be derived from detailed 3D numerical simulations (ASCE 1988; Li and Wang 1994; Thomas and Williams 1995; Younis 1996). However, even allowing for future advances in representing the 3D flow, a comprehensive theoretical framework still needs to be developed for determining the equilibrium form of stable alluvial channels before attempting to simulate changes from the equilibrium profile.

Identification of the junction point between active (that is eroding) and inactive (noneroding) elements of the bank, together with characterization of the erosion front that moves this point, appears to be crucial to quantifying width adjustment. Hydraulic conditions at the active-inactive junction are especially difficult to determine with precision, even for in-bank flows. Near-bank hydraulic conditions for over-banks are strongly influenced by secondary flow structures close to the banks and must be represented carefully through the correct use of local friction factors, eddy viscosities, and depth-averaged secondary flow values. In this respect, the depth-averaged approach described earlier is worthy of further study. Although major drawbacks with this approach remain, a need still exists for calibration of specific channel shapes and no details of vertical motion are available from the process equations.

It is also important to know the rates at which width, depth, slope, and local morphological adjustments are made, so that errors can be assessed when an incremental series of quasi-steady-state discharges are used to simulate a hydrograph. The dominant or effective discharge responsible for forming channel morphology, although easy to define in theory, is still poorly understood and, except for work by Ackers (1992), very little attention has been paid to the influence of over-bank flows on dominant discharge. The hypothesis that in an equilibrium channel the bank-full stage corresponds to the dominant or effective discharge has some theoretical basis, but it may be a special case within a variety of associations between important features of channel morphology and a range of effective flows (Hey 1975; Thorne et al. 1993; Biedenharn and Thorne 1994). Further experimental work on equilibrium and nonequilibrium alluvial channels is required before linkages between dominant discharge, the range of effective flows, and channel morphology can be substantiated.

# ANALYSIS

The limitations of our knowledge of fluvial hydraulics and some suggestions for a research agenda to address present constraints on understanding flow processes that drive width adjustment are discussed next.

### *Local Boundary Shear Stress*

If the local time-averaged boundary shear stress continues to be used as a parameter defining sediment entrainment, transport, and deposition in morphological models, then the knowledge of its spatial and temporal variation in different types of channels must be improved. At present cross-sectional boundary stress distributions are predicted from empirical equations for channels with simple cross-sectional shapes and attributes. These simple equations deal only with heterogeneous roughness, in which bed and bank roughness heights are similar. New equations for local boundary shear stress values relative to spatial and temporal mean values should be developed for irregular cross-sectional shapes and roughness distributions. The experimental work necessary has to be laboratory-based but should include field studies in engineered prismatic channels and alluvial rivers with natural cross sections under a wide range of flow conditions to test the validity of results from more idealized situations.

### *Turbulence and Bank Shear Stresses for In-Bank and Over-Bank Flows*

Coherent turbulent structures and secondary patterns must be studied in more detail close to the bed and in the near-bank zones of river banks. Not only are temporal variations in the near boundary Reynolds stresses due to turbulent burst and sweep important for sediment entrainment and motion, but so are the large-scale motions induced either by nonuniform boundary roughness or skewing of the flow at bends (secondary currents of Prandtl's first and second kinds).

### *Accurate Longitudinal Averaging of Streamwise Terms in Governing Equations*

River channel topography demands that some longitudinal averaging is performed during modeling. Studies should be made into the effects of using 1D resistance laws to simulate reach-scale and local morphological features. Until energy gradients and resistance coefficients are refined to represent local variability, uncertainties are likely to persist.

### *Laboratory and Field Studies on Rates of Adjustment*

There is a dearth of high-quality data on rates of change of parameters at both field and laboratory level. Well-focused laboratory studies (Ikeda 1981), small field studies (Fujita et al. 1990; Hasegawa 1993; Fukuoka 1994; Hagerty et al. 1995), and large-scale field studies (Ergenzinger et al. 1994) have shown that it is possible to investigate the many facets and processes at work using a variety of experimental approaches.

### *Unsteady Flow and Dominant Discharge Approaches*

Studies of river width adjustment in most fluvial systems require that fieldwork is sustained over relatively long time periods. As in natural streams the hydrological regime is uncontrollable, continuous measurement is essential to guarantee gathering of useful data, and the long-term reliability of monitoring equipment to collect the hydraulic data is crucial.

### *Extremal Hypotheses*

At present no consensus exists regarding the applicability of extremal hypotheses to the analysis of width adjustments. Before adjustment models incorporating extremal hypotheses are further developed there is a need to explore the basis, limitations, and prerequisites for application of these hypotheses in disequilibrium channels with complex geometries and, consequently, there is still a need for both theoretical and laboratory studies.

### *Bank Roughness*

The effects of bank and riparian vegetation on the river channel roughness and morphology are still poorly understood (Ikeda and Izumi 1990; Masterman and Thorne 1992; Darby and Thorne 1996b), and further research in both laboratory flumes and natural channels is essential to elucidate the effect of submerged and emergent stems on near-bank velocities, turbulence, bank shear stresses, and channel conveyance capacity. Research should incorporate studies of both flexible and stiff-stemmed plants.

## BANK MECHANICS

The fundamental processes responsible for channel width adjustment are fluvial erosion, fluvial deposition, and mass bank failure. The following seven topics concerned with the mechanics of bankline movement are germane to width adjustment and are addressed in this section: (1) Bank erosion; (2) weakening of resistance to erosion; (3) bank stability with respect to mass failure; (4) basal endpoint control; (5) effects of vegetation; (6) seepage effects; and (7) bank advance.

### Bank Erosion

Water flowing in an alluvial channel exerts forces of drag and lift on the boundaries that tend to detach and entrain surface particles. To remain in place the boundary sediment must be able to supply an internally derived force capable of resisting the erosive forces applied by the flow. The origin of these resisting forces varies according to the grain size, size distribution, and the nature of electrochemical bonding that may exist between particles. Alluvial bank materials are formed primarily by fluvial deposition and are often stratified, with a general fining-upward sequence. Therefore, the engineering characteristics and erodibility of the bank may vary with elevation. Also, floodplain deposits typically include alluvial sands and gravels, clay plugs, and strongly cohesive backswamp deposits, so that bank material properties vary spatially over relatively short distances. While the distribution of sustained bank retreat along the course of a river depends primarily on the distribution of boundary shear stress in the near-bank zones, outcrops of particularly resistant material may act to slow the local bank retreat rate and to distort the fluvially driven pattern of channel planform evolution.

In the case of noncohesive sands and gravels, the forces resisting erosion are generated mainly by the immersed weight of the particles, although close packing of grains in imbricated patterns can also wedge particles in place, greatly increasing the critical boundary shear stress necessary for entrainment. Generally, the mobility of noncohesive bank materials can be predicted using a Shields-type entrainment function, but this must be modified to take into account the destabilizing effect of the channel side slope. Also, the critical value of the dimensionless shear stress must be adjusted to allow for excessive tightness or looseness in packing of bank material particles (Thorne 1982).

Fine-grained bank materials, containing significant amounts of silt and clay, are to some degree cohesive and resist entrainment primarily through interparticle, electrochemical bonding rather than through the immersed weight of the particles. When cohesive bank materials are entrained by the flow, it is aggregates of grains (such as soil crumbs or peds that have been produced by soil-forming processes) that are detached. Fluvial entrainment, therefore, requires that the local boundary shear stresses exceed the critical value to initiate motion of crumbs or peds rather than that related to the primary soil particles. Ped size, stability, and interped bonding strength are not conservative soil properties as they depend to some degree on the local history of soil development, in general, and recent antecedent conditions of wetting and drying, in particular. It follows that the conditions of incipient motion for cohesive bank materials are complex, time-dependent, and difficult to define.

A TC (ASCE 1968) summarized early studies into the mechanics of cohesive bank erosion. The TC recorded the results of noteworthy contributions by, among others, Smerdon and Beasley (1961) and Flaxman (1963) on channel stability in cohesive materials and Grissinger and Asmussen (1963) and Grissinger (1966) on the erodibility of cohesive soils.

For example, Grissinger and Asmussen (1963) found that the erosion resistance of clayey soils increased with the time that the materials were wetted. They postulated that when clay is initially wetted the free water releases the bonds between particles, but that, as free water is absorbed, the clay minerals hydrate and interparticle bonds are strengthened. This illustrates how the chemical bonding of clay particles may vary with time and the history of soil moisture changes.

A wealth of subsequent work has further addressed fundamental aspects of cohesive soil behavior, leading to important papers by Partheniades and Passwell (1970), Kandiah and Arulanandan (1974), Arulanandan (1975), Arulanandan et al. (1975), Ariathurai and Krone (1976), Ariathurai and Arulanandan (1978), Abt (1980), Grissinger (1982)—who presents an excellent review of progress achieved up to the early 1980s, Kamphuis and Hall (1983), Parchure and Mehta (1985), Springer et al. (1985), Shaikh et al. (1988a,b), and, most recently, Annadale and Parkhill (1995) and Kranenburg and Winterwerp (1997). Space limitations preclude an in-depth review of the findings of these papers here. However, in summary it can be concluded that while critical boundary shear stresses for cohesive bank soils are extremely difficult to predict accurately (Grissinger 1982), they tend to be higher than for noncohesive bank materials. As a result, erosion rates for cohesive banks are generally lower than for noncohesive materials (Vanoni, 1975; Thorne and Tovey 1981).

Once entrained, crumbs and peds disintegrate rapidly due to corrasion at the channel boundaries and turbulent buffeting in the flow, so that most fine sediment derived from bank erosion is transported in suspension and is conventionally classified as wash load. However, Hey (1997) has recently challenged this long-held view. He points out that, as bank material load is derived from erosion processes operating within the channel, it should be differentiated from watershed-derived wash load, the supply of which is not controlled by any in-stream process. The concentration of wash load tends to be supply-limited and is independent of the transport capacity of the flow. Hey (1997) suggests that bank material load should be classified, with the bed material load, as part of the transport-limited sediment load. As the dynamics and deposition of transport-limited and wash load components of the total load differ markedly within the fluvial system (Vanoni 1975), the reclassification of bank-derived sediment from wash load to channel-derived sediment load would have important morphological implications. These issues should be further explored.

## Weakening of Resistance to Erosion

The erodibility of bank soils can be increased markedly by processes of weakening and weathering. The processes responsible for loosening and detaching grains and aggregates are closely associated with soil moisture conditions at and beneath the bank surface. In poorly drained soils positive pore-water pressures act to reduce bank stability, which can lead to bank failure, particularly during rapid drawdown of the channel stage following a high flow. Conversely, rapid immersion of a dry bank can lead to slaking, which is the detachment of aggregates by positive pore pressures due to compression of trapped air.

Swelling and shrinkage of soils during repeated cycles of wetting and drying can contribute to cracking that significantly increases erodibility and reduces soil shear strength. Shrinkage is especially damaging to the strength of the bank when intense drying of the soil leads to desiccation cracking. Heaving, due to the 9% increase in water volume on freezing (Ritter 1978), and the growth of needle ice crystals at the bank surface, followed by collapse of ice wedges and needles during thawing of soil moisture, are highly effective in increasing the susceptibility of cohesive bank materials to flow erosion (Lawler 1993b).

Temporal variability in the erodibility of bank soils due to the operation of weakening processes means that the effectiveness of a given flow event in eroding the bank depends not only on the magnitude and duration of a particular event but also on antecedent conditions (Wolman 1959).

## Bank Stability with Respect to Mass Failure

Fluvial erosion drives bank retreat directly by the removal of material from the bank face, but it often also causes bank retreat by triggering mass instability. The stability of the bank with regard to mass failure depends on the balance between gravitational forces that tend to move soil downslope and the forces of friction and cohesion that resist movement. Failure of the bank occurs when flow scour of the bed next to the bank toe increases the bank height, or undercutting increases the bank angle, to the point that motivating forces exceed restoring forces on the most critical potential failure surface, and the bank collapses in a gravity-induced, mass failure.

The analysis of slope stability with respect to mass failure has been the topic of considerable research effort, primarily by geotechnical engineers, but also by geomorphologists and geophysicists. Engineering research has concentrated on development of engineering designs for artificial slopes and embankments, but rather little of this work is applicable to the very steep slopes, undisturbed soils, complex stratigraphies, and unspecified drainage conditions found in eroding, natural river banks. Also, application of most geotechnical analyses requires detailed site investigation to provide the necessary data on profile geometry, soil properties, bank stratigraphy, and ground-water flow net. While it is possible to collect such detailed information for a specific construction site or key location, the data obtained cannot easily be generalized to represent bank conditions along a reach of river, due to inherent variability in the properties of natural alluvium and uncertainty concerning the local bank environment.

In fact, relatively little research specifically concerned with streambank stability has been undertaken, and the following brief review of slope stability literature concentrates on more recent work that is directly relevant to mass failure of river banks. However, it is acknowledged that treatment of river bank stability shares a common origin with that of engineered slopes and embankments in the fundamental work of researchers, such as Bishop (1955), Peck and Deere (1958), Bishop and Morgenstern (1960), Morgenstern and Price (1965), Spencer (1967), Terzaghi and Peck (1967), Vaughan and Walbancke (1973), Fredlund and Krahn (1977), and Poulos et al. (1985).

There is a clear contrast in failure mechanics between noncohesive and cohesive materials because of significant differences in their soil mechanics. In a noncohesive bank, shear strength increases more rapidly with depth than does shear stress, so that critical conditions are more likely to occur at shallow depths. In a cohesive bank, shear stress increases more quickly than shear strength with increasing depth so that critical surfaces tend to be located deep within the bank (Terzaghi and Peck 1967).

Noncohesive materials usually fail by dislodgement and avalanching of individual particles or by shear failure along shallow, very slightly curved slip surfaces. Deep-seated failures occur in cohesive materials with a block of disturbed, but more or less intact, bank material sliding into the channel along a curved failure surface. In banks with shallow slope angles ($\theta$

< 60°), the failure surface is curved and the block tends to rotate back toward the bank as it slides, in a rotational slip (Fig. 15). Steep banks characteristically fail along almost planar surfaces, with the detached block of soil sliding downward and outward into the channel in either a planar slip or a toppling failure (Fig. 16) (Thorne 1982).

Rotational slips may be defined as a base, toe, or slope failure depending on where the failure arc intercepts the ground surface [Fig. 15(a)] and are analyzed using conventional geotechnical procedures (Bishop 1955; Fredlund 1987). The risk of failure is usually expressed by a factor of safety, defined as the ratio of restoring-to-disturbing moments about the center of the failure circle. In the method of slices [Fig. 15(b)], the soil body within the failure arc is divided into vertical slices with forces acting as shown. To obtain a determinate solution for the factor of safety it is necessary to make an assumption regarding interslice forces, for example, to assume that these forces act horizontally. The critical slip failure circle cannot be simply located, and usually a computer program is used to explore the large number of possible solutions to determine the position of the most critical arc.

In nature, many eroding river banks are very steep, and near-vertical banks, termed river cliffs, often occur at the outer margin of meander bends and along severely incised channels.

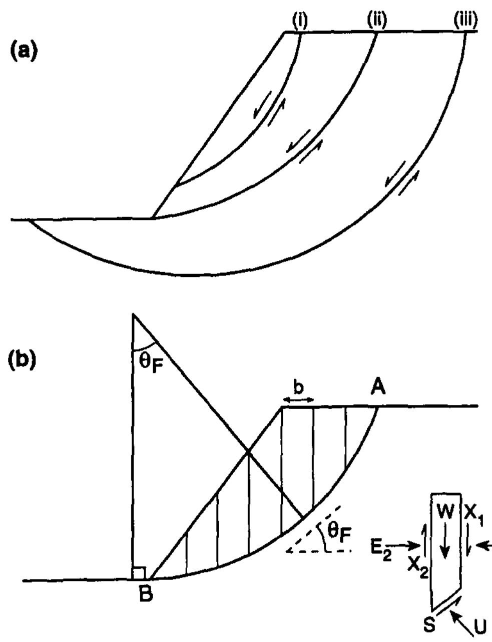


**FIG. 15. (a) Rotational Slip Failures in Cohesive Bank: (i) Slope Failure, (ii) Toe Failure, (iii) Base Failure; (b) Stability Analysis of Slip Circle by Method of Slices [after Thorne (1982)]**

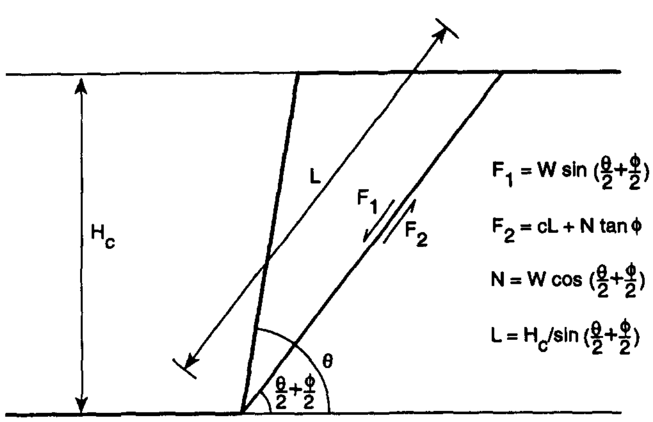


**FIG. 16. Culmann Analysis for Plane Slip Failure [after Thorne (1982)]**

Such steep slopes formed in friable soils are rarely encountered in hillslope and embankment studies and, consequently, stability analyses for planar slip have received relatively little attention in the geotechnical literature. Approaches that have been developed stem from the Culmann method, in which forces acting on the potential failure block are resolved normal to and along the failure plane, leading to an equation defining the critical height for mass failure that has the general form

$$H_c = 4c(\sin\theta\cos\phi)/\gamma[1 - \cos(\theta - \phi)] \quad (5)$$

where $H_c$ = critical bank height; $c$ = cohesion; $i$ = bank angle; $y$ = bank material unit weight; and $\phi$ = friction angle.

Tension cracks often develop downward from the ground surface, parallel to the bankline, behind steep banks due to horizontal, tensile stress in the soil at this location (Terzaghi and Peck 1967). Tension cracks truncate the effective length of the potential failure surface, tending to destabilize the bank and reducing its stability relative to that predicted from a Culmann analysis. Cracks may occupy as much as half of the bank height, isolating a column or slab of soil, which then slides and topples forward into the channel in a toppling failure (Fig. 17).

Stability analyses applicable to the very steep (almost vertical), deeply cracked river cliffs associated with eroding, unstable streambanks have been undertaken by researchers in hydraulic engineering and fluvial geomorphology, but much more testing and validation is required before these models could be adopted for routine application as engineering design tools (Osman and Thorne 1988; Darby and Thorne 1996a). As pointed out by Millar and Quick (1997), however, these analyses should only be applied to very steep banks.

Cantilevered or overhanging banks are generated when erosion of an erodible layer in a stratified or composite bank leads to undermining of overlying, erosion-resistant layers. Thorne and Tovey (1981) pointed out that cantilevered banks may fail by shear, beam, or tensile collapse (Fig. 18). Shear failure [Fig. 18(a)] occurs when the weight of the cantilever block exceeds soil shear strength, causing the overhanging block to slip downwards along a vertical plane. In a beam failure, a block rotates forward about a horizontal axis within the block [Fig. 18(b)] when disturbing moments about the neutral axis exceed restoring moments. Tensile failure [Fig. 18(c)] occurs when the tensile stress exceeds soil tensile strength and the lower part of the overhanging block falls away. Frequently, the strength of cantilever blocks is significantly increased by root reinforcement due to riparian and floodplain vegetation. Flow erosion and tensile failures occur below the root mat, leaving

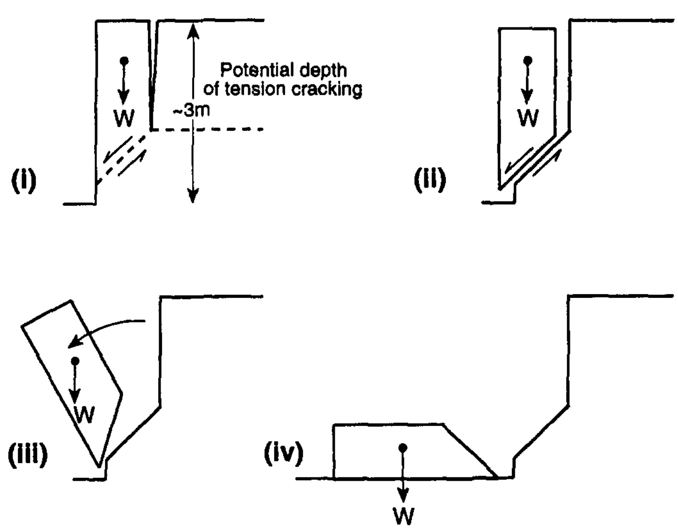


**FIG. 17. Sequence (i–iv) of Toppling Failure on Low Steep Stream Bank [after Thorne and Tovey (1981)]**

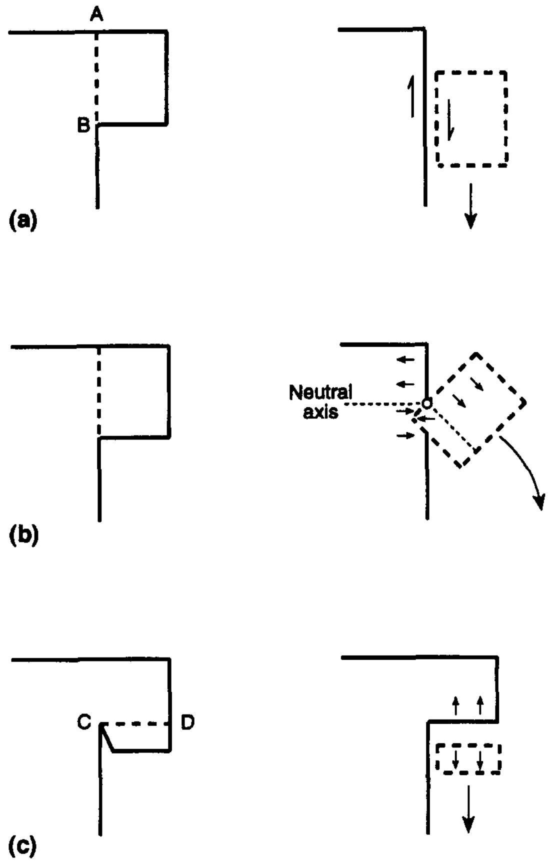


**FIG. 18. Mechanisms of Cantilever Failure: (a) Shear Failure along AB; (b) Beam Failure about Neutral Axis; (c) Tensile Failure across (CD) [after Thorne and Tovey (1981)]**

root-bound cantilevers that fail subsequently by either the beam or shear mechanism.

Whether bank failure occurs by rotational slip, toppling, or cantilever collapse, the primary force tending to move the failure block is the tangential component of the weight of the block. Fluvial erosion can increase the motivating force by increasing the bank height (through bed scour next to the bank toe) or by increasing the bank slope angle (through lateral erosion of the toe and lower bank). The weight of bank material also increases with the moisture content of the soil and failure often follows the change from submerged to saturated conditions that occurs when drawdown occurs in the channel.

Ample field evidence exists that bank failure may be triggered by any of these changes in motivating force. For example, Abam (1993) noted that bank failures in the Niger Delta, Nigeria, could often be attributed to increases in bank height and bank angle due to fluvial erosion and bed scour at the river banks. Abam also documented decreased bank stability due to rapidly falling water levels that led to (1) the loss of the confining pressure provided by the channel water level; (2) positive pore pressures due to poor drainage resulting from the low permeability of the soil; and (3) increases in the effective unit weight of the soil due to saturation.

## Basal Endpoint Control

While fluvial erosion processes and geotechnical failures are controlled by different aspects of bank geomorphology, they are actually linked. The key to characterizing this link lies in recognizing that mass wasting delivers the failed material to the toe of the slope, or basal area, but does not entirely remove it from the bank profile. The removal of failed material from the basal area depends primarily on its entrainment by current and wave action, followed by fluvial transport downstream. The concept of basal endpoint control explains how the medium- to long-term retreat rate of the bank is controlled by the rate of sediment entrainment and removal from the toe.

The concept of basal endpoint control was first developed by Carson and Kirkby (1972) to explain variations in hillslope profiles. Thorne (1982) applied the concept to river banks, proposing that bank retreat can only be sustained when the near-bank flow is able to remove failure debris and to continue to scour the basal area. In contrast, where the flow is unable to remove all the debris basal accumulation occurs and a berm or bench of failed material develops. This tends to protect the intact bank from fluvial erosion and, by acting as a buttress against gravity failures, increases bank stability. On this basis, the balance of basal supply and removal of sediment can be defined by one of the following three states of basal endpoint control:

1. *Impeded removal*—Bank failures supply debris to the base at a higher rate than it is removed. Basal accumulation results, decreasing the bank angle and height and, therefore, increasing stability with respect to mass failure. The rate of debris supply decreases, favoring the second state.
2. *Unimpeded removal*—Processes delivering debris to the base and removing it are in balance. No changes in basal elevation or slope angle occur. The bankline recedes by parallel retreat at a rate determined by the degree of fluvial activity at the base.
3. *Excess basal capacity*—Basal scour has excess capacity over the debris supply from bank failures. Basal lowering occurs, increasing bank angle and height and, therefore, decreasing stability with respect to mass failure. The rate of debris supply increases, favoring the second rate.

The state of basal endpoint control is useful in explaining the medium- to long-term rates of river bank retreat or advance. It also highlights the importance of considering the response of near-bank morphology to bank stabilization. The concept indicates that a reduction in debris supply that is due to bank stabilization may induce a state of excess basal capacity that generates very deep toe scour (Thorne et al. 1995). As pointed out by Maynord (1996), this additional scour must be properly accounted for in the design of the stabilization works if failure due to undermining is to be avoided.

Hagerty (1991a,b) proposed that not all sustained bank retreat depends on the state of basal endpoint control. This proposal was based on the fact that piping is a widespread cause of sustained bank retreat along the Ohio River that is, apparently, independent of the state of basal endpoint control. Even though the bank toe is stable, upper bank retreat has continued unabated for many years. However, closer inspection of the relevant bank profiles indicates that the reason that the toe has been stable is that, in this regulated river, the toe is well below pool level and is thus morphologically inactive. Piping in sand layers at about the elevation of the standard low water plane has produced a bench that represents the toe of the morphologically active bank. At this elevation, bank retreat may still be considered to be covered by the concept of basal endpoint control, with the bank profile above pool elevation almost continually in a state of unimpeded removal, due to the ability of current and wave action to remove the fine debris supplied by piping. Creation of the bench and control of the profile thus depend on the piping process in supplying debris that can easily be removed by waves and currents that would not otherwise be able to erode intact bank material.

Hagerty et al.'s detailed treatment highlights the subtlety of interactions between fluvial and mechanical processes responsible for bank retreat, and it illustrates that great care must be taken when interpreting bank processes from bank form, especially in regulated rivers.

## Vegetation Effects

The role of vegetation in affecting bank erosion and width adjustment is complex and poorly understood. Although vegetation generally reduces soil erodibility, its impact on bank stability with respect to mass failure may be either positive or negative. Hence, depending on the geomorphic context and dominance of either fluvial processes or mass failure, vegetation may produce either a net increase or a decrease in the rate of bankline shifting.

Vegetation can play an important role in limiting the effectiveness of bank erosion by detachment and entrainment of individual grains or aggregates of bank material. Compared to unvegetated banks, erosion of well-vegetated banks is reduced by one to two orders of magnitude (Carson and Kirkby 1972; Smith 1976; Kirkby and Morgan 1980). Gray and Leiser (1982) and Coppin and Richards (1990) have reviewed the effects of herbaceous and, to a lesser extent, woody vegetation in reducing flow erosivity and bank erodibility and concluded that major effects include the following:

- Foliage and plant residues intercept and absorb rainfall energy and prevent soil compaction by raindrop impact.
- Root systems physically restrain soil particles.
- Near-bank velocities are retarded by increased roughness.
- Plant stems dampen turbulence to reduce instantaneous peak shear stresses.
- Roots and humus increase permeability and reduce excess pore water pressures.
- Depletion of soil moisture reduces water-logging.

Gray and Leiser (1982) and Coppin and Richards (1990) also reviewed the ways that woody vegetation may affect the balance of forces promoting and resisting mass failure. Roots mechanically reinforce soil by transferring shear stresses in the soil to tensile stresses in the roots, which root strength is able to resist. However, this effect operates only to the rooting depth of the vegetation, and it does not reinforce potential failure planes that pass beneath the plant rootballs. Hence, root reinforcement is negated when bank height significantly exceeds rooting depth.

Soil moisture levels are decreased by interception on the canopy and evapotranspiration from the foliage, reducing the frequency of occurrence of the saturated conditions conducive to bank collapse. Anchored and embedded stems can act as buttress piles or arch abutments in a slope, counteracting downslope shear stresses and increasing bank stability. However, roots may also invade cracks and fissures in a soil or rock mass and thereby cause local instability by their wedging or prying action. The surcharge weight of vegetation may significantly increase motivating forces, causing destabilization of the bank, and wind loading of tall vegetation may exert an additional and potentially critical destabilizing moment on the bank.

These few examples illustrate the complexity of vegetation impacts on flow erosivity, soil erodibility, and mass stability. A recent scoping study on bank vegetation and bank protection reached the conclusion that vegetation may be either a positive or negative influence on bankline stability and retreat rate (Thorne et al. 1997). This may explain the apparently contradictory conclusions regarding the effect of bank vegetation on equilibrium channel width of, for example, Hey and Thorne (1986), who reported that stable channel width decreases as the density and stiffness of bank vegetation increase, and Murgertroyd and Ternan (1983), who found the opposite in a study of the effects of afforestation on channel form. Also, they may explain why the notable increases in the shear strength of root-permeated soils found in laboratory test soils by Waldron (1977) are not always replicated in strength measurements made in real river banks (Amarasinghe 1992).

As pointed out by Thorne and Osman (1988a), Darby and Thorne (1996a) and, most recently, Thorne et al. (1997) a great deal of further research is necessary before vegetation effects can be properly understood and incorporated into the technical description of bank material characteristics under conditions representative of the range of environments encountered along natural streams and waterways.

## Seepage Effects

In addition to fluvial activity causing scour at the toe of the slope, grain-by-grain detachment, and mass wasting, Parola and Hagerty (1993) have identified a general class of failure mechanisms that is often very important to bank stability. This class of mechanisms is driven by seepage within the bank.

Pore-water movement within the bank is most vigorous during and following a high flow event. As flood waters rise in the stream, the increased hydraulic head drives seepage into the bed and banks, resulting in ground-water recharge. As the flood stage recedes hydraulic gradients reverse, driving seepage into the stream from the banks. The distribution of inflow, movement, and outflow through the bank is seldom uniform but is, in fact, strongly influenced by the layered stratigraphy that is a characteristic of alluvial banks.

Alluvial banks consisting of sand, silt, and clay layers typically have hydraulic conductivity that is much greater in the horizontal direction than the vertical. Consequently, groundwater flow occurs principally by horizontal seepage into and out of sandy layers. During bank drainage, outflowing water may entrain and remove grains from a sand layer—a process termed piping by Hagerty (1991a,b). Piping erosion leads to undermining of overlying, less pervious layers causing those layers to deflect and distort. The most common result of this undermining is the formation of cracks in the undermined layer, where the soil is unable to support the tensile stresses created by deflection (Parola and Hagerty 1993). Mass wasting then occurs as cracking reduces the operational strength of the bank.

Another type of bank failure associated with strong seepage is gully development. While gully development is usually regarded as resulting from surface erosion, subsurface erosion by piping may lead to subsequent collapse of the pipes to form gullies along streambanks (Harvey et al. 1985). This mode of gully formation is particularly likely in loess and alluvially redeposited loess.

Bank weakening and erosion by seepage is often overlooked by river engineers. Failure to identify subsurface piping erosion can lead to misclassification of the erosion problem and subsequent problems with bank stabilization works that are adequate to armor the bank against fluvial attack, but which are likely to fail due to internal erosion driven by piping.

## Bank Advance

Bank advance occurs through sediment deposition and may result in channel narrowing. For example, in an analysis of East Fork River, Wyo., Andrews (1982) documented sedimentation in cross sections wider than the mean bank-full width, with a significant portion of the accumulated sediment deposited against the channel banks. Andrews noted that if the material deposited against the banks during one spring flood remained through the next year, the channel would become narrower and would gradually approach the mean bank-full width of the reach.

Conversely, channel narrowing as the result of channel metamorphosis can drive bank advance. Metamorphosis is the result of local morphological response of the channel to changes in runoff, sediment supply, or management regime (Schumm 1977).

For example, in their channel evolution models for incised channels, Schumm et al. (1984) and Simon (1989) described how the recovery of stability was characterized by aggradation of the channel bed and fluvial deposition on the banks. Schumm et al. (1984) referred to the bank depositionary features as berms. Harvey and Watson (1988) observed that in Muddy Creek bank berms grew to a relatively constant elevation above the bed of the channel. This implies that the berms formed by bank accretion form in response to some relatively systematic feature of flow and sediment regime (Taylor and Woodyer 1978), most likely the dominant or effective channel-forming flow.

A hypothesis accounting for the processes responsible for berm development has been proposed by Harvey and Watson (1988). According to this hypothesis, bed material sediments that are transported in dunes during higher flows are left in remnant bed forms along the channel margins as the stage falls. The remnant dunes are then draped with fine-grained silts and clays from the suspended load as the flow diminishes, stabilizing the deposits. Repetition of this process eventually produces a stable, sedimentary berm, permanently advancing the bankline.

Evidence for the cyclical deposition of dune sands and wash load silts is preserved in the accretionary bank stratigraphy as sand-mud couplets that increase in thickness and grain size with depth below the berm surface. It should be noted that, according to this hypothesis, the bank advance process may be accelerated through sediment trapping in pioneer vegetation that invades the berms, although vegetation invasion is not responsible for the initial deposition of the berm sediments (Taylor and Woodyer 1978).

## CONCLUSIONS

The conclusions of the TC findings are as follows:

1. Width adjustments take place within a wide range of geomorphic contexts. Adjustment may take place as part of the natural evolution of channel morphology, it may be caused by disruption of the long-term equilibrium morphology due to an extreme event, or it may be a morphological response to river engineering or management.
2. To understand, predict, and manage changing channel width it is essential that civil engineers are aware of the geomorphic context within which width adjustment is occurring.
3. Analysis of equilibrium width in stable channels can be approached using (1) empirical, regime methods; (2) extremal hypotheses; and (3) rational, tractive force methods. These approaches are strictly limited to prediction of time-invariant width in graded or regime channels. They can be used with care to predict asymptotic values of width following disturbance of the graded or regime condition, but they cannot predict either the rate of change or intermediate widths attained during dynamic adjustment of channel morphology. Despite these limitations, these methods have many useful engineering applications.
4. The time and spatially averaged boundary shear stress is an important parameter in predicting both equilibrium width and width adjustment. However, the lateral distribution of local values of boundary shear stress is poorly understood, especially for channels with nonuniform cross sections.
5. Improved understanding of the effects of over-bank flows on river width adjustment processes is needed.
6. A variety of mass failure mechanisms may be involved in bankline retreat. Care must be taken to match the slope stability analysis used to check bank stability to the critical failure mechanisms observed in the field. It is essential that engineers identify actual and potential instability mechanisms prior to selecting an engineering or management strategy for dealing with bank retreat and width adjustment.
7. The long-term rate of bank retreat or advance of the bank toe can be explained using the concept of basal endpoint control. However, seepage-driven procedures operating within the bank can lead to serious bank instability due to piping even when wave and current action at the toe is not excessive.
8. Bank advance takes place through sediment accumulation as a berm or bench in the channel, often accelerated by invasion of pioneer riparian vegetation.

## RECOMMENDATIONS

The following recommendations synthesize the findings of the TC:

1. Improved techniques for stream reconnaissance should be developed to establish the geomorphic context of width adjustments in alluvial channels.
2. Laboratory studies are required to allow detailed investigation of near-bank flow hydraulics and local boundary shear stress distributions in nonuniform channels. Laboratory studies must be verified by field data. Field studies are required to support investigation of the rates and directions of change in channels with actively adjusting widths and to allow measurements of flow fields and bank material characteristics affecting bank erosion/accretion, bank stability, and processes of bankline retreat and advance.
3. Since very few detailed studies of turbulence have been conducted, even in channels with planar boundaries, there is scope for further original work, including studies in natural channels with bed forms, at both laboratory and prototype scales.
4. Further work on the governing equations of flow is needed, in particular studying how terms are spatially averaged or approximated, especially in meandering rivers.
5. To validate morphological models it is important that simultaneous measurements are made of rates of adjustment in width, depth, slope, cross-section shape, longitudinal energy gradient, and sediment transport.
6. Recognizing these challenges, a need exists for a commitment to long-term, strategic monitoring of a small number of river systems undergoing morphological adjustment to provide benchmark data that can never be obtained from piecemeal, project-related studies of individual reaches where an instability problem generates monitoring of local changes without reference to the river system as a whole. There is also a need for some well-focused laboratory work on disequilibrium, mobile-boundary channels with unsteady flows.
7. Numerical tools must be developed to predict the distribution of local friction factors, roughness heights, and boundary shear stresses in nonuniform channels.
8. Research should be undertaken to develop improved methods of predicting the erodibility of cohesive bank materials.
9. Methods capable of predicting the stability of streambanks with respect to a range of possible failure mechanisms must be further developed.
10. A major research effort should be directed at improving knowledge of the impacts of aquatic, emergent, and ri-

parian vegetation on near-bank flow hydraulics, bank erosion, mass stability, and equilibrium channel morphology.

11. Bank advance through sediment accumulation has been relatively neglected in research to date, and further work is urgently needed on this topic.

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## APPENDIX II. NOTATION

*The following symbols are used in this paper:*

$B$ = channel width;
$B_f$ = flatbed channel width;
$B_f^*$ = $B_f/D_c$, dimensionless flatbed width;
$c$ = bank material cohesion;
$D_c$ = depth at center of channel;
$D^*$ = dimensionless bed elevations;
$f$ = local friction factor defined in Eq. (3);
$g$ = acceleration due to gravity;
$H$ = channel depth;
$H_c$ = critical bank height;
$H^*$ = $H/D_c$, dimensionless vertical depth;
$P_b$ = wetted perimeter of bed;
$P_w$ = wetted perimeter of both walls;
$Q$ = volumetric flow rate or discharge;
$R$ = hydraulic radius;
$S_f$ = friction slope;
$S_0$ = channel streamwise slope;
$s$ = local channel side slope of banks;
$U$ = long-stream velocity;
$U_d$ = depth-averaged mean velocity;
$U_0$ = section mean velocity;
$U_*$ = shear velocity;
$V$ = cross-stream velocity;
$y$ = lateral distance from center of channel;
$y^*$ = $y/D_c$, dimensionless lateral distance from center of channel;
$\varepsilon_{yx}$ = depth-averaged eddy viscosity;
$\Gamma$ = secondary flow parameter defined in Eq. (3);
$\gamma$ = bank material unit weight;
$\theta$ = bank angle;
$\lambda$ = dimensionless eddy viscosity defined in Eq. (3);
$\rho$ = water density;
$\tau_a$ = depth-mean apparent shear stress defined by Eq. (4);
$\tau_b$ = local bed shear stress;
$\tau_{bc}$ = critical local bed shear stress for entrainment of bed material;
$\tau_w$ = local wall shear stress;
$\tau_0$ = boundary shear stress;
$\phi$ = bank material friction angle; and
$\omega$ = lateral bed shape.