the average cross-sectional hydraulics; in fact, local effects may be different from those that occur over a long reach of the channel (Colby 1964).

Fluvial deposition is important in geomorphology in several ways. On a long-term basis, continued deposition, called *aggradation,* results in landforms that reflect distinct periods of geomorphic history. The sedimentology and stratigraphy of the associated deposits indicate the types of rivers involved in the aggradational phase (Schumm 1977) and provide clues to the environmental conditions present at the time of the aggradational event. On a short-term basis, deposition creates bed forms and other microtopographic features such as dunes, bars, and riffle-pool sequences that are closely related to channel pattern and the character of flow within the channel (for example, see Schumm et al. 1982). Finally, you should recognize that the short- and long-term mechanics of deposition have implications beyond the boundaries of geomorphology. They are clearly basic to sedimentology and stratigraphy and, interestingly, may be key factors in subdisciplines of economic geology such as the exploration for valuable placer deposits (Schumm 1977).

The Frequency and Magnitude of River Work

At this juncture we can logically ask when and how fluvial work is done. Is it the super event of very high discharge that happens once in a millennium that causes rivers to do what they do, or is it the normal flow that is repeated time and time again? The answer to this question rests firmly on the concept of geomorphic work.

Geomorphic work is usually estimated in one of two ways. Wolman and Miller (1960) suggest that the work done by a river can be estimated by the amount of sediment it transports during any given flow. They concluded that in most basins 90 percent of the total sediment load (i.e., 90 percent of the work) is removed from the watershed by the sum of rather ordinary discharges that recur at least once every five or ten years. While megafloods transport an abnormally high sediment load, they occur so infrequently that their contribution to the total amount of sediment that is transported out of the basin over a period of years is minimal. In contrast, flows of limited magnitude are incapable of transporting significant loads. Thus, the discharges that transport the most sediment are those that are able to move debris at a moderate rate and that occur relatively frequently.

The Wolman and Miller hypothesis has been extensively examined for a wide range of river systems. These studies demonstrate that the most effective transporting discharges vary significantly from one region to another; considerable variability may even exist within any given region (Ashmore and Day 1988; Nash 1994). Andrews and Nankervis (1995), for example, examined 17 gravel bed rivers in the western United States and found that the most effective flows for transporting bed material load over a period of years ranged from 0.8 to 1.6 times the bankfull discharge. Nonetheless, the basic tenet of the Wolman-Miller hypothesis—that most of the sediment transported by rivers is performed by moderate, relatively frequent discharges—appears to hold true for the majority of the rivers investigated.

The second way to estimate geomorphic work, with perhaps greater implications, is to assess the conditions under which rivers make adjustments to or maintain their channel morphologies. Wolman and Miller (1960) suggest that river channels form and reform within a narrow range of flows. The lower flow limit is set by the demands of competence. Clearly, the shape of the channel cannot be modified by erosional processes if the flows are incapable of transporting the bed and bank material. The upper limit is defined by the flow that exceeds bankfull and is no longer confined to the channel. From this perspective, channel configuration is presumed to be a direct indication of river work, and its precise form is perceived to be the product of high-frequency events. This hypothesis has also received considerable support and, indeed, was reinforced by studies that suggested that channel morphologies are adjusted during flows having a recurrence interval of 1.1 to 2 years, and that approximate bankfull discharge (Kilpatrick and Barnes 1964; Dury 1973). Therefore, the discharge that determines the characteristics and dimensions of a channel, known as the **dominant discharge,** has been implicitly accepted to have a frequency and magnitude equivalent to the bankfull condition.

It seems justified to say that river channel morphology is maintained in all environmental settings by geomorphic work done during a dominant discharge or within a distinct range of flows. However, it should be recognized that the recurrence interval of the bankfull discharge can vary significantly, potentially exceeding 1–2 years by an order of magnitude (Williams 1978). Moreover, it is now questionable as to whether bankfull discharge is the dominant discharge for all rivers. For example, Harvey and his colleagues (1979) found that river flows in northwest England redistributed bed material between 14 and 30 times a year and changed overall channel form from 0.5 to 4 times a year. In coarse-grained rivers, low to moderate flows may be incapable of entraining the bed and bank material. Thus, only rare, high-magnitude events may be able to effect a change in channel form (Baker 1977).

The concept of a dominant discharge is further complicated by the realization that the effect of major floods on channel configuration, referred to as *geomorphic effectiveness,* varies with the environmental setting (Costa 1974b; Gupta and Fox 1974; Baker 1977; Moss and Kochel 1978). This prompted the suggestion that the Wolman-Miller principle should be modified to include factors that control the work of floods in different environments (Wolman and Gerson 1978).

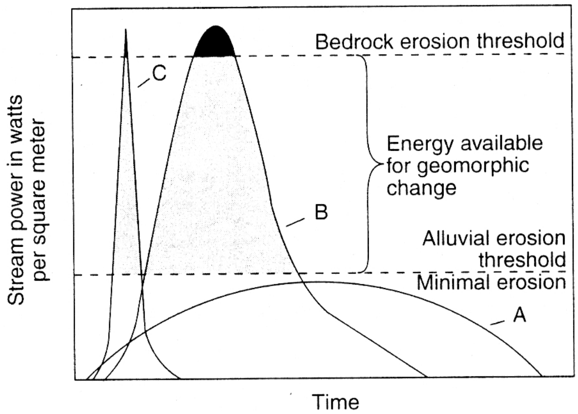

Figure 6.15
Hypothetical stream-power graphs associated with different kinds of floods. The most geomorphically effective floods are those characterized by curve B that exceed the threshold of erosion for significant periods of time.
(From Costa and O'Connor 1995)

Historically, the impact of floods on channel morphology has been related to flood magnitude, a parameter that varies greatly with basin morphometry and climate (see chapter 5). Kochel (1988) points out, however, that the most significant effects of flooding have been associated with peak discharges that are several times greater than the mean annual discharge. Thus, the *difference* between peak flood discharge and the discharge that is normally experienced by the channel may be more important in controlling the extent to which the channel is modified than the *absolute* magnitude of the event. In addition, Costa and O'Connor (1995) found that some dam-burst floods generated exceptionally high instantaneous stream powers, but produced few geomorphic effects. They argue that the limited effects of these events are related to the fact that flow duration was relatively short and the total energy expended was minimal. Long-duration flows may be necessary to wet and disaggregate the soils, thereby reducing the shear strength of the bank materials. Thus, the maximum effects of flooding may be associated with some optimal combination of flow magnitude (stream power), duration, and total energy expenditure above the initiation of particle transport (Costa and O'Connor 1995)(fig. 6.15). The geomorphic response will also be influenced by the erosional resistance of the materials that comprise the channel perimeter.

In light of the above, it should be clear that the magnitude of channel modification during an event is dependent upon the complex interplay between a large number of parameters. Kochel (1988) has subdivided these controlling parameters into two categories, which he refers to as drainage basin factors and channel factors. Figure 6.16 shows that these factors interact in such a way that the most significant effects are generally concentrated along high gradient, coarse-grained channels in headwater areas, particularly those characterized by abundant bedload.

There is a growing realization that an individual basin having constant physical/biological properties can experience different geomorphic responses in successive floods of similar magnitude (Newson 1980; Beven 1981; Kochel et al. 1987). This indicates that effectiveness is partly controlled by factors other than the nature of the flow and the channel characteristics. The most important factor seems to be recovery time (Wolman and Gerson 1978). *Recovery time* is essentially the time needed for a river to recover its equilibrium form after a major flow event has disrupted the channel configuration (for alternative definitions, see Pitlick 1993). Implicit in this perception is that major hydrologic events may be able to affect the form of a channel, and that changes produced may be long-lived or may be quickly erased as the system reverts to its pre-event condition. Thus, the effectiveness must be related to the time needed to obscure the impacts of the event on the river. Moreover, the effects of any given event may be dependent on whether the channel has fully recovered from the impact of the previous flood. Kochel (1988), for instance, documented the responses of the Pecos River of west Texas to catastrophic floods in 1954 and 1974. He found that the 1954 flood resulted in the massive redistribution of channel bed gravels and the severe erosion of the channel margins. In contrast, the 1974 event resulted in few channel changes. Presumably, the recovery times in this area were sufficiently long that the channel was still largely adjusted to the high discharges of the 1954 flood. Kochel's conclusions indicate that the effectiveness of any event is dependent upon both the actual time between successive floods and the time required for the system to recover. The healing interval is generally thought to be climatically controlled. In humid areas, recovery times appear to be short, whereas arid and semiarid regions usually have much longer recovery times (Wolman and Gerson 1978).

THE QUASI-EQUILIBRIUM CONDITION

Every river strives to establish an equilibrium relationship between the dominant discharge and load by adjusting its hydraulic variables (e.g., channel width and depth, velocity, roughness, and water slope). This normal fluvial condition has been aptly referred to as a "quasi-equilibrium" state (Leopold and Maddock 1953; Wolman 1955) because the flow variables are mutually interdependent, meaning that a change in any single parameter requires a response in one or more of the others. The difficulty involved in understanding rivers becomes evident when you consider that discharge and load are in continuous flux, and so all the hydraulic variables must always be adjusting. Obviously a river cannot attain equilibrium as a steady-state condition; thus the term quasi-equilibrium.

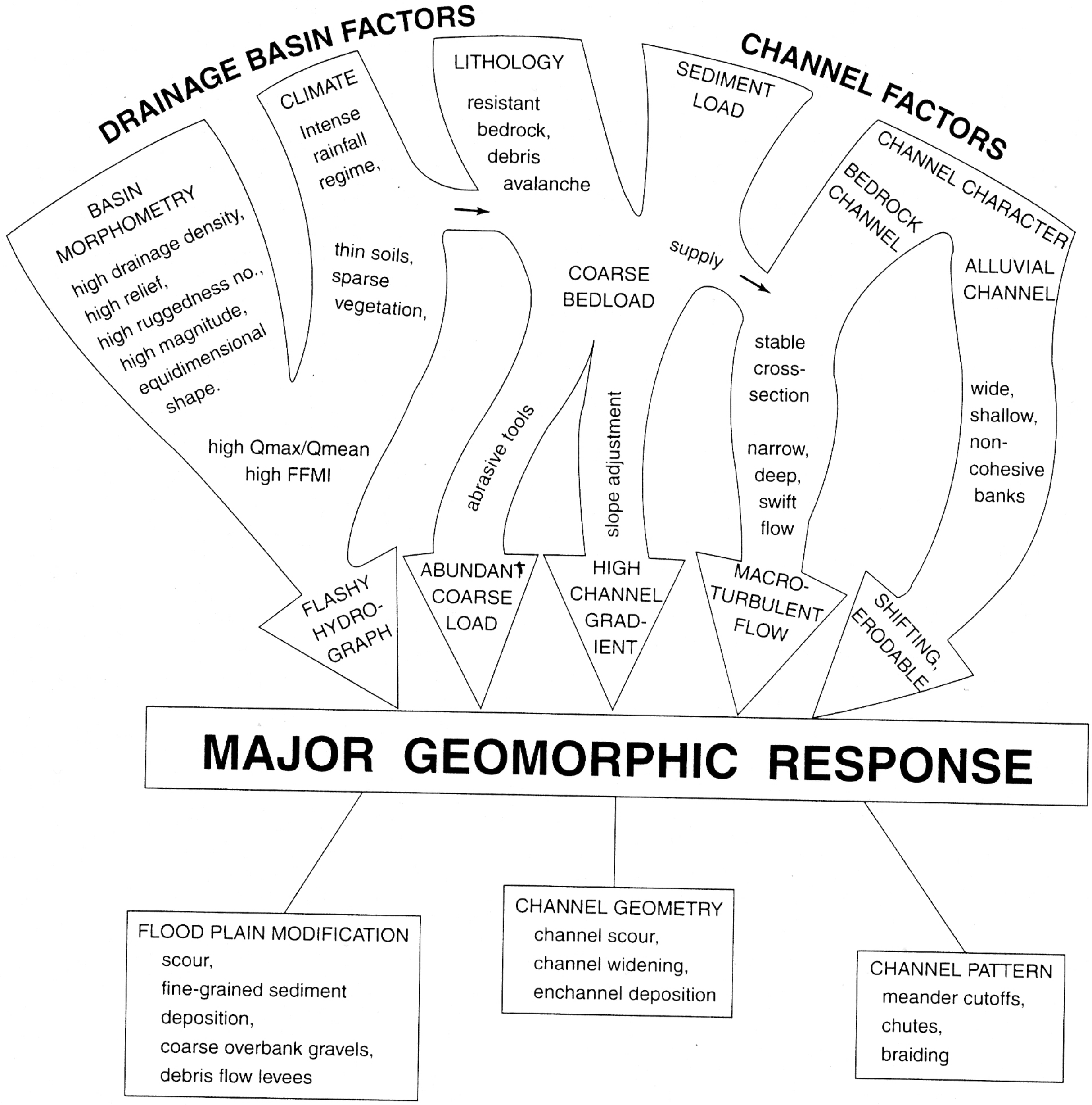

Figure 6.16

Summary of the factors controlling channel and floodplain response to large-magnitude floods.
(From Kochel 1988)

Hydraulic Geometry

The quasi-equilibrium condition was first demonstrated in a landmark study by Leopold and Maddock (1953). Using abundant flow records compiled at gaging stations throughout the western United States, they set out to determine the statistical relationships between discharge and other variables of open channel flow; these relationships are known as *hydraulic geometry* of river channels. Because every river has wide fluctuations in discharge, any given channel cross-section must transport the range of flows that comes to it from the adjacent upstream reach. Discharge, therefore, serves as an independent variable at any station, and the changes in width, depth, velocity, or other variables can be observed over a wide spectrum of discharge conditions (fig. 6.17). At a station each of the factors (w, d, v) increases as a power function such that

$$w = aQ^b$$

$$d = cQ^f$$

$$v = kQ^m$$

where a, c, k, b, f, and m are constants. The exponents b, f, and m indicate the rate of increase in the

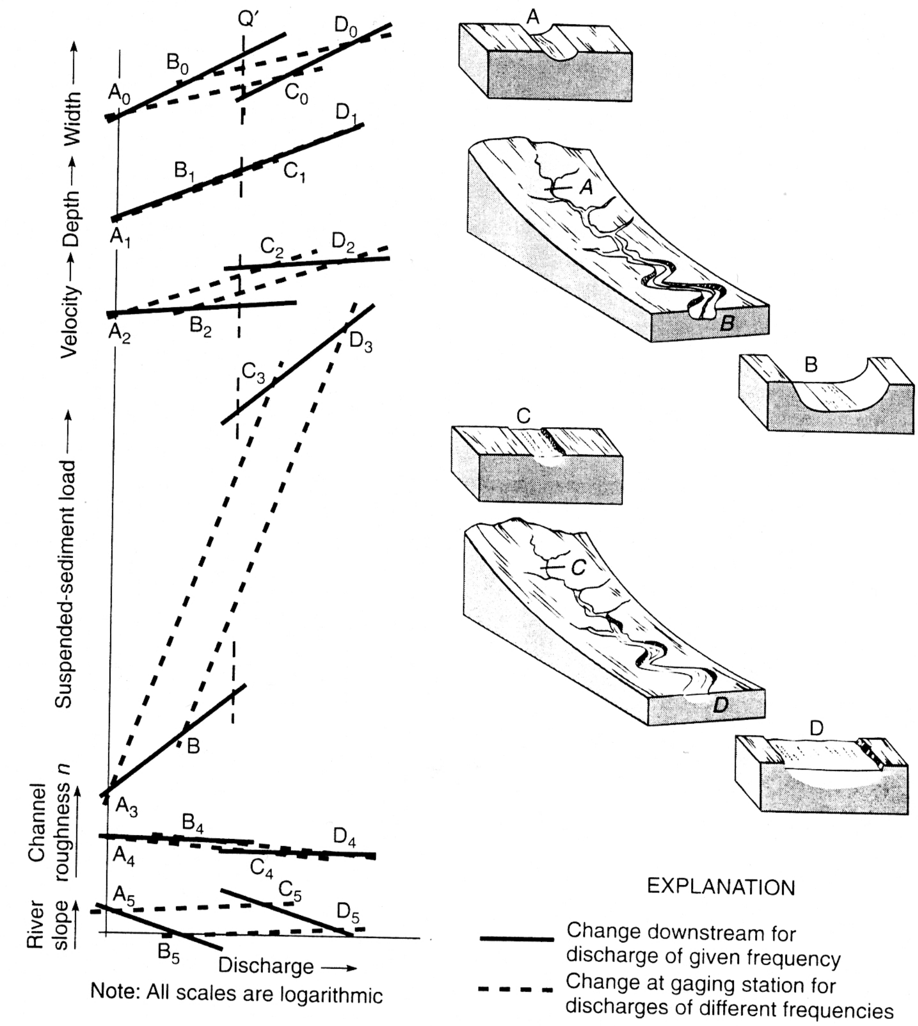

Figure 6.17
Hydraulic geometry relationships of river channels comparing variations of width, depth, velocity, suspended load, roughness, and slope to discharge at a station and downstream.
(Leopold and Maddock 1953)

hydraulic variable (w, d, v) with increasing discharge. Because discharge (Q) equals the product of width, depth, and velocity, the relationship can be expressed as

$$Q = aQ^b \times cQ^f \times kQ^m$$

or

$$Q = ackQ^{b+f+m}$$

and it follows that $(a \cdot c \cdot k)$ and $(b + f + m)$ must equal 1. Leopold and Maddock found that the average at-a-station values of b, f, and m for a large number of midwestern and western streams were 0.26, 0.40, and 0.34, respectively. Essentially the at-a-station exponents tell us what portion of the increase in discharge will be caused by an increase in each of the component variables. It should be recognized, however, that the exponent values represent average values and, thus, will not fit any particular stream. In fact, Phillips (1990) suggested that the proportion of the increase in discharge accounted for by each of the component variables may not even be consistent from one flow to the next.

Discharge also increases with the expansion of drainage area, and so on most rivers it must increase downstream. The question is how much of the downstream increase in discharge results from width, depth, and velocity. To make this analysis, care must be taken to ensure that the variables are measured during the same flow conditions. On a given day, for example, a disastrous flood with high w, d, and v values may be occurring in an upstream reach whereas flow conditions far downstream are normal. A comparison of the hydraulic variables in these two widely divergent frequencies of flow would be misleading. Obviously the frequency of the discharge must be considered for any observations of downstream hydraulic geometry to be valid.

In sum, then at-a-station and downstream hydraulic geometry differ in that one (at-a-station) compares flows of vastly different frequencies whereas the other (downstream) analyzes variables at the same frequency of Q even though the absolute values of Q differ between downstream stations.

Leopold and Maddock (1953) found that width, depth, and velocity increase downstream with increasing mean annual discharge (fig. 6.17). The average values of b, f, and m for western streams are 0.5, 0.4, 0.1, respectively. In general the rate of change in depth (f) is relatively consistent in both downstream or at-a-station geometry, whereas width usually increases much more rapidly and with more consistent values downstream than at a station (Knighton 1974; Williams 1978). Velocity increases more rapidly at a station than it does downstream.

The suggestion that mean velocity increases downstream came as a shock to most geologists, who intuitively "knew" that water in small tributaries flowing on steep slopes must be traveling faster than water in the low-gradient trunk rivers. Their surprise at this new interpretation of velocity probably resulted from geologists' inclination to consider slope as the major, if not the overriding, control of velocity. Nonetheless, the possibility of a downstream increase in velocity should have been suspected because Manning's equation tells us that depth plays a greater role than slope in determining velocity. In a stream with a constant roughness, increased depth can overcompensate for the loss of velocity resulting from a decrease in slope.

While it is clear that velocity increases downstream along some rivers, the exponents derived for downstream analysis vary from region to region and even for any particular stream within a region.

Carlston (1969), for example demonstrated that on large rivers downstream velocity is probably constant, but on smaller streams it may increase or decrease according to local controls. Thus, the downstream analysis of hydraulic geometry represents the exposition of a general trend which may not be applicable to all rivers. The utility of hydraulic geometry in geomorphic studies has yet to be satisfactorily documented. In fact, Park (1977) found that variations in sets of *b-f-m* values do not even distinguish between rivers in diverse climates. Nonetheless, Rhodes (1977) argues that hydraulic geometry may provide a relatively simple means of describing local variations in geomorphic process. He believes that all rivers can be categorized on the basis of various ratios of exponent values or other river properties (e.g., Froude number, roughness) that are controlled by the exponential values. The groups proposed by Rhodes reflect basic fluvial mechanics, and this approach suggests that hydraulic geometry should be useful in predicting how any particular river will work. For example, one group includes rivers in which the rate of increase in velocity (m) exceeds the combined changes in width and depth ($b + f$). Such rivers should experience a rapid increase in competence with rising discharge, a condition that is probably needed to entrain coarse bedload (Wilcock 1971).

In a slightly different approach, Williams (1987) was able to define temporal changes in the unit hydraulic geometry (i.e., geometry based on discharge per unit channel width) of selected stream reaches in the western United States. He argued that these changes were the direct result of channel adjustments to natural or anthropogenic disturbances. Thus, the analysis and comparison of hydraulic geometry for different time periods may represent a valuable means of assessing the impact of environmental change on river systems.

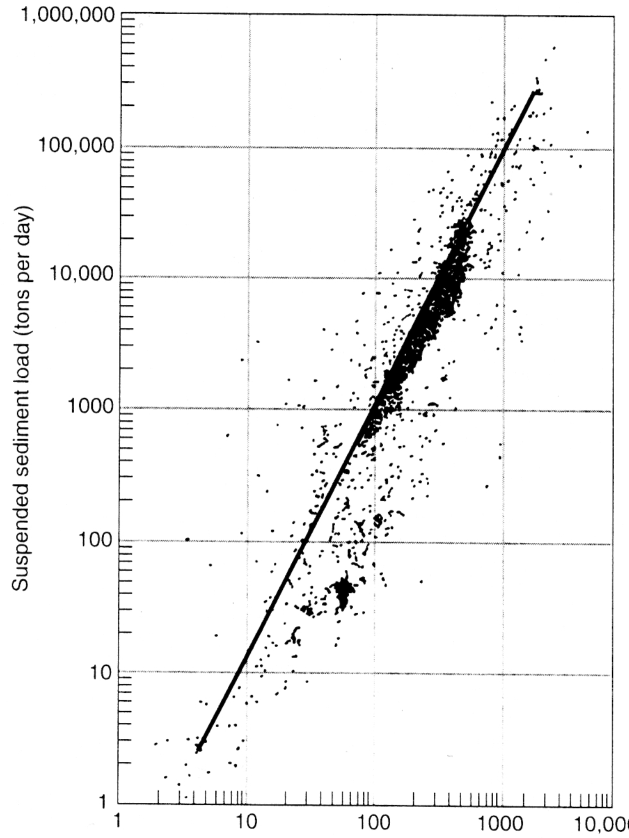

Figure 6.18
Relation of suspended load to discharge in Powder River at Arvada, Wyo.
(Leopold and Maddock 1953)

In addition to examining the variations in width, depth, and velocity with increasing discharge, considerable attention has been given to the changes in suspended sediment loads as discharges fluctuate at a site. Within most rivers, the amount of suspended sediment at a station increases directly with discharge (fig. 6.18) and can be expressed as the simple power function in which

$$L = pQ^j$$

where L is suspended load and p and j are constants. The at-a-station value of j is often > 1, indicating that the influx of sediment to the river is greater than the addition of water. Interestingly, the dramatic increase in sediment content is not necessarily caused by scouring of the channel floor. Several studies have shown that scouring can occur at peak flow, during rising flow, or in the waning part of the flood (Leopold and Maddock 1953; Foley 1978; Andrews 1979).

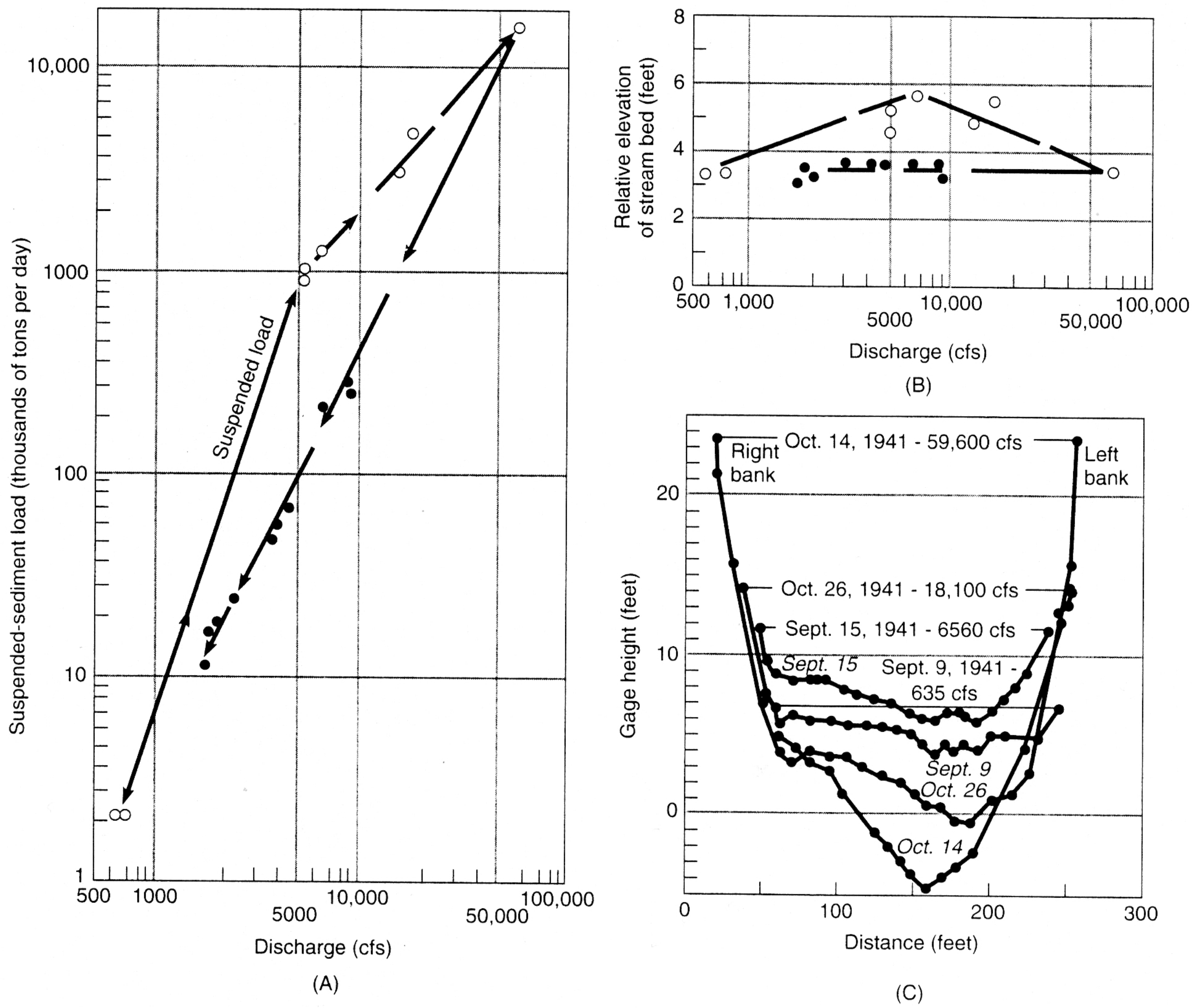

Figure 6.19
Changes in (A) suspended load, (B) streambed elevation, and (C) water-surface elevation with discharge during the September–December 1941 flood of the San Juan River near Bluff, Utah.
(Leopold and Maddock 1953)

In those situations where scouring occurs at peak flow or during the recession phase, deposition may take place in the rising stages of the flood, precisely when the suspended load is increasing rapidly (fig. 6.19). Given the channel bed deposition is occurring, the suspended sediment cannot be derived from the erosion of the channel floor. This observation means that the bulk of sediment added to a river during a flood is derived from the valley-side slopes of the watershed, channel bank erosion, or tributary input. Moreover, figure 6.19A shows that for a given discharge, the suspended sediment load is greater during the rising limb of the flood than when the flood waters are receding. Similar observations have been made for the relations between sediment concentration (suspended load/unit volume H_2O) and discharge. This phenomenon, referred to as *hysteresis*, helps explain the notable variation in suspended load at any given discharge that is evident in figure 6.18.

Not all rivers exhibit higher suspended sediment loads during rising flood stages. Williams (1989) has identified five relationships that may exist between suspended sediment concentration and discharge for any given river (fig. 6.20). The differences in these relations between rivers, or even between reaches of the same river, have been attributed to the interaction of a large number of factors. These include the intensity and areal distribution of precipitation within the basin, the amount and rate of runoff, distance of the gaging station from sources of water and sediment production, differences in the transport rates between water and sediment, the amount of sediment stored within and along the channel, and the depletion of easily eroded debris within the channel or the surrounding uplands (Williams 1989). Because many of these variables change seasonally, it is possible that the relationships between sediment load and discharge for an event will also change during the year (although such trends have not been adequately documented).

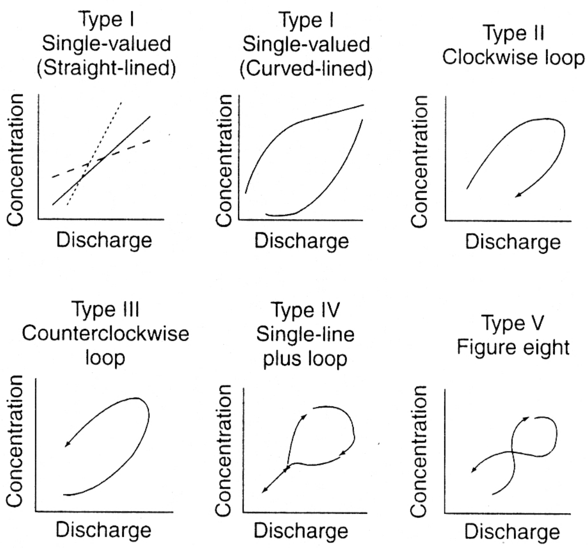

Figure 6.20
Types of suspended sediment concentration-discharge relationships recognized in natural channels.
(Adapted from Williams 1989)

The Influence of Slope

Channel slope has always been recognized as a prime adjustable property of rivers, and there is abundant evidence to substantiate the importance of slope in a river striving to maintain balance. Nevertheless, it is clear that gradient represents only one of the many variables that may be altered to maintain the quasi-equilibrium condition as changes in sediment load and discharge occur. In fact, observations at gaging stations show that the slope of the water surface remains relatively constant during flows of different magnitudes. Therefore, we cannot call on a dramatic increase in slope to produce the relatively high rate of increase in velocity (m). The increasing velocity must be generated by an increase in depth, a decrease in roughness, or both. Downstream the channel gradient does exert an influence, because in most rivers there is a notable decrease in slope. Roughness, however, usually remains fairly constant (Leopold and Maddock 1953) because of the offsetting effects of a decrease in particle size (decreased n) and a decrease in sediment concentration (increased n). As a result, any increase in velocity downstream can best be justified by the increase in depth, explaining the low exponent values of m in that direction.

The relationships between slope and other hydraulic parameters reveal the complexities of quasi-equilibrium, but they do not explain why slope usually decreases downstream or what external factors may control the form of the longitudinal profile. As early as 1877, G. K. Gilbert concluded that slope was inversely related to discharge, and because Q increases with basin area and stream length, it is axiomatic that slope should decrease downstream. However, in most rivers particle size generally diminishes downstream, prompting many observers to suggest that channel gradient adjusts to the size of the bed material. Actually both factors are probably involved. Rubey (1952) demonstrated that if channel shape is constant the slope will decrease with (1) a decrease in particle size, (2) a decrease in total load, and (3) an increase in discharge. Rubey concludes that the channel gradient at any point along the river is a function of both sediment and discharge. If Rubey is correct, then slope is dependent, or partially so, on all hydraulic variables because they are also related to discharge.

Many studies subsequently have shown the correctness of Rubey's analysis. In one of these studies, comparing stream profiles in areas of differing geology, Hack (1957) found no consistent correlation between slope and bed-material size when all sample localities from a geologically divergent region were considered together. Only after Hack added a third variable, drainage area, to the analysis did a significant relationship become apparent (fig. 6.21), and slope could then be defined by the equation

$$S = 18(^{M}/_{A})^{0.6}$$

where M is the median size of the bed material in millimeters, A is area in mi^2, and S is slope in ft/mi. Because basin area can normally be used as an index of discharge (Leopold et al. 1964), Hack's study reinforces Rubey's contention that both Q and sediment are determinants of slope. It does not indicate which factor is the principal determinant; indeed, one would expect the relationship to be defined by different mathematical equations in different physical settings. In addition, there is some evidence to suggest that the factors controlling channel gradients may be scale dependent. Prestegaard (1983b) found that within some gravel bed streams, particle size and bed configuration (topography) were the major determinants of water-surface slope. Particle size was the most important determinant over a range of distances from local (one to three times the channel width) to an entire reach (100–300 m). Bed configuration exerted an influence only on a scale relating to the entire reach length.

The interaction of the factors controlling channel gradient ultimately results in the river's longitudinal profile (change in elevation with increasing length). Within alluvial channels it is generally accepted that a concave-up longitudinal profile is associated with rivers in equilibrium (fig. 6.22). However, the concave-up form is not a necessary requirement of the equilibrium condition (Sinha and Parker 1996), and recent studies have demonstrated that along rivers formed in bedrock, diverse longitudinal profiles can be maintained over graded time (Pazzaglia et al. 1998). In contrast to rivers developed in alluvium, bedrock channels are capable of transporting more sediment than is available. This suggests that for bedrock rivers, sediment may not be the most critical factor controlling the shape of the