Channels

6

Introduction

Stream channels drain landscapes and carry material from erosional uplands to depositional lowlands, estuaries, and coastal environments. Channels range in size from small, headwater rivulets to large, continent-draining rivers. Channels are shaped by the flows and sediment loads they carry, the cohesive strength of the material in their banks, the slopes they flow down, and the rocks they cut into. Without rivers and streams to carry debris away, upland valleys would gradually become choked with sediment eroded off hillslopes, reducing ridgetop to valley-bottom relief and leveling the land. Fluvial processes control local deposition and erosion of sediment, allowing rivers to migrate across valley bottoms and form floodplains. These processes also govern the way that rivers incise into bedrock and generate the topography of upland valleys. An understanding of **fluvial** (river) processes and dynamics is thus central to studies of landscape evolution.

The interaction of flowing water and sediment shapes channels, so the supply of water, sediment, and large organic debris (such as logs) greatly influences channel morphology and dynamics. These factors vary within drainage basins, across regions, and over time. Channels exhibit tremendous variability in morphology and can respond quickly and significantly to changes in discharge and sediment load. The physical processes that determine channel morphology and dynamics are similar across different regions, but the importance of local controls and the influence of the downstream routing of water and sediment lead to a wide variety of channel

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A braided river in central Alaska cuts through boreal forest and tundra; its gravel-rich, light-colored active channel stands out from the dark vegetation.

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types. Over long periods of geologic time, channels respond to tectonic uplift and subsidence, sea-level rise and fall, erosion of the landscape, and changes in climate and vegetation. Over shorter timescales, channels adjust to seasonal, annual, and decadal cycles of discharge and sediment supply as well as to changes in land use and extreme events like droughts and floods.

Regional climate and tectonic setting together broadly determine whether streams flow through steep or gentle terrain, and whether channels carry low or high loads of fine or coarse sediment. Climate establishes the amount, type, and seasonal pattern of precipitation and runoff, and controls the dominant runoff-generating mechanisms and thereby streamflow magnitude and variability. The depth to the water table influences whether channels carry flow only during storms or whether they also carry **baseflow** during periods between storms. Channels in arid and semi-arid regions may be **ephemeral** and flow for only part of the year, while channels typical of humid or temperate regions are **perennial** and flow year-round.

The potential for intense thunderstorms over small areas makes flash floods an important (and dangerous) fluvial process not only in semi-arid landscapes like much of Australia and the U.S. Southwest, but also in subhumid landscapes of the North American Midwest and in the humid southern United States and northeastern Australia, where hurricanes and cyclones strike. Ice jams that form in floods during spring ice breakup are an important fluvial process (eroding banks, removing riparian vegetation, and spreading sediment over floodplains) in places where rivers annually freeze over, like New England, subarctic Canada, and the northern Great Plains of North America [**Photograph 6.1**]. In general, regional climate is the dominant influence on the timing, magnitude, and regularity of the streamflows that control channel processes.

Vermont Historical Society

PHOTOGRAPH 6.1 Cleared Ice Jam. In the late 1800s, a spring ice jam along the White River in Sharon, Vermont, had cleared—but not before rising water levels, sufficient to move large ice blocks, spilled out of the channel and onto the floodplain, covering a road and damaging several buildings. Such ice jams recur every spring as river ice breaks up.

Regional tectonic forces generate uplift that elevates the land surface as well as subsidence that lowers the topography over which streams flow and into which they incise channels. Over geologic time, streams **grade** their channels, adjusting channel slope in response to changes in uplift and erosion and the balance between sediment supply and transport capacity. Over shorter time frames, however, channel slope is an externally imposed factor that controls channel processes, particularly the flow of water and the transport of sediment. Regional geologic and tectonic history determine the rock types (lithologies) over which streams flow. Rock type influences the processes of bedrock erosion, as well as the durability of the sediment that streamflow carries.

Steep channels incised into bedrock in tectonically active mountains differ greatly from lowland rivers that flow across depositional basins where sediment accumulates and can be stored over long periods of time. Upland channels tend to have little sediment cover, typically have rocky beds and/or banks, and have little **alluvial** storage in their valley bottoms. In contrast, lowland alluvial channels typically have bed and banks composed of material transported by the channel and store substantial amounts of sediment in their valley bottoms. Mountain (bedrock) streams generally have a **transport capacity** that exceeds their sediment supply, whereas lowland (alluvial) streams have a sediment supply that equals or exceeds their transport capacity.

Long-term and short-term controls on fluvial morphology and processes both act to shape habitats for in-channel and near-channel organisms. River dynamics influence the distribution of species along river corridors, and **riparian** (near-stream) areas are the most biologically diverse parts of many landscapes. Not only do rivers shape riparian zones, but riparian zones also shape rivers by contributing woody debris and slowing flood flows.

This chapter addresses the influences of regional landscape context on stream systems, explains the flow and transport processes that move mass through channel networks, introduces the fluvial landforms that result from the action and interaction of these processes, and describes channel response to common changes in these processes. Rivers around the world have been dammed, straightened, pinned between levees, and cleared of logs and logjams. Modern channel restoration, rehabilitation, and flood-control efforts are based on an understanding of the ways in which environmental controls, the history of human modifications, and physics interact in particular rivers. This chapter provides context for such work by presenting the key concepts in fluvial geomorphology.

External Controls on Fluvial Processes and Form

Some factors that influence streams, including regional climate and tectonics, are externally imposed controls to which channel processes must respond. Other factors, like the pattern of erosion and deposition within a stream valley, are themselves influenced by channel processes and response. As an example of the feedback between external controls and channel processes, consider how net differences in erosion or deposition lead to changes in channel slope. If sediment supply exceeds a stream's transport capacity, the additional deposition will cause channel aggradation that increases channel slope downstream of the point of deposition, thereby increasing the local transport capacity, which will facilitate moving the deposited material downstream. Conversely, if a channel's transport capacity exceeds its sediment supply, erosion will scour the channel bed (potentially down to bedrock), thereby reducing the channel bed slope. In this context, a **graded stream** has a profile that is adjusted to carry its sediment load. A graded stream profile is concave up, with steeper channels in headwaters declining progressively downstream toward the outlet of the river network.

Upland slopes deliver water and sediment to stream channel networks in varying ways and amounts and over different periods of time. Streamflow sorts, breaks down, and transports this material, and thus it shapes channels in response to external and self-limiting controls. Fluctuations in water flow and sediment supply give rise to spatial and temporal variability in channel morphology, processes, and response. The shape and behavior of a channel is governed primarily by its sediments, its discharge, the composition of its bed and banks, and the vegetation growing in and immediately adjacent to the stream. Many of these factors interact in complex ways with human modifications like dams and levees that impose new external controls on a channel or stream system.

Discharge

Discharge is the volume of water flowing past a point on a stream per unit time [**Figure 6.1**]. Stream discharge (Q) is typically measured in cubic meters per second (m^3/s) and is equal to the product of the channel's cross-sectional area (A_{cs}) and the flow velocity (U), or to the product of the average stream width (W), flow depth (D), and flow velocity:

$$Q = A_{cs}U = WDU \qquad \text{eq. 6.1}$$

Discharge stays the same downstream unless water is added or lost, an expression of continuity and the conservation of mass. In humid-temperate regions, discharge systematically increases downstream within channel networks because small channels converge to form larger ones, and because perennial streams in wet climates typically gain water from groundwater aquifers. In arid regions, discharge often decreases downstream as water infiltrates through the streambed.

The discharge in a channel varies over both event and seasonal timescales as individual storms and annual weather patterns deliver precipitation. High-discharge events mobilize large volumes of sediment and typically govern the processes that dramatically reshape channels. The frequency, magnitude, and duration of high flows vary with seasonal patterns of precipitation and temperature. Some stream systems experience regular, predictable flood events, including those in regions with monsoon climates, streams dominated by seasonal melting of snowpacks, and high-latitude rivers that flow poleward and experience massive ice jams when they thaw upstream while downstream reaches remain frozen. In most regions, however, irregular storm-driven rainfall events like hurricanes and local intense thunderstorms produce flood flows.

The way stream channels and their valleys carry high discharges greatly affects channel morphology and valley-bottom landforms. High-discharge events in mountain channels confined between bedrock valley walls have greater flow depths, for the same flood discharge, than do channels flowing through wide valley bottoms across which floodwaters disperse. This fundamental difference in how channels convey high flows leads to the common observation that very little sediment is stored in mountain valleys and there is generally extensive sediment deposition in broad alluvial lowlands.

Sediment Supply

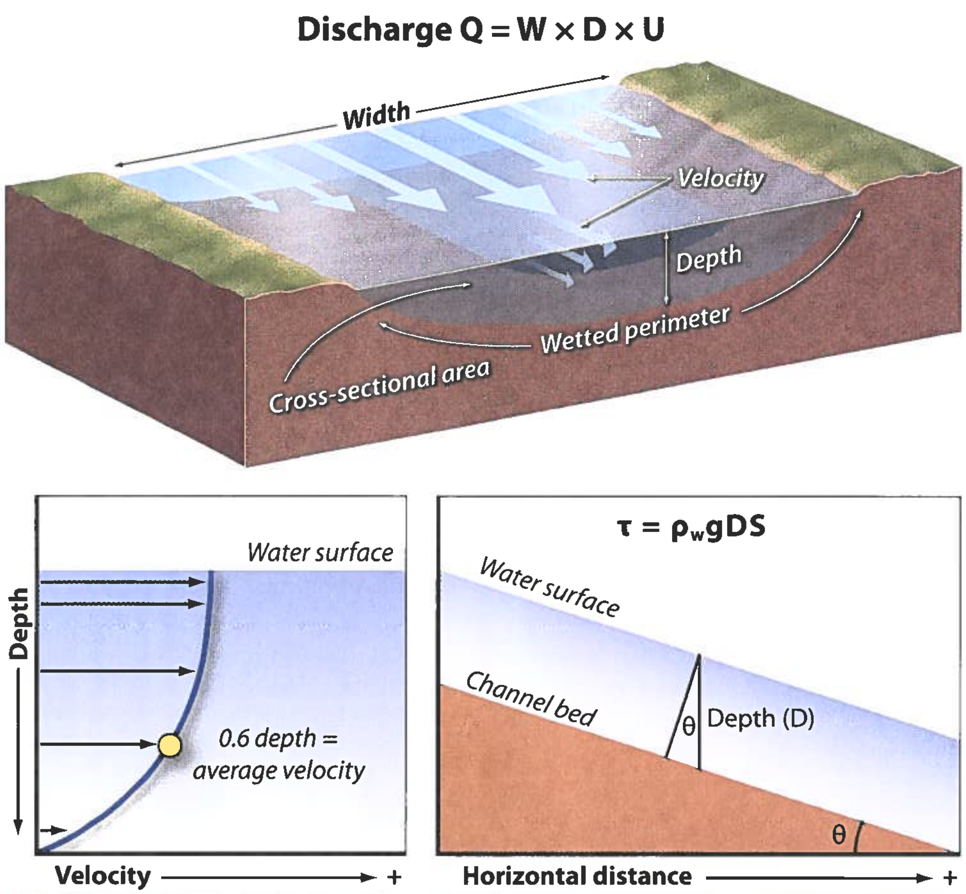
The sediment supply to a channel is imposed by the rate at which material is delivered from upstream channel reaches, from neighboring hillsides, and from bed and bank erosion during channel incision and migration. The volume, grain size, and degree of sorting of sediment supplied to a channel influence how much of the material subsequent streamflow is able to transport and sort, and how much the stream is unable to move and must flow around. Sediment supply often varies tremendously over time. Hillslope processes generally deliver much greater volumes of sediment during abrupt events, such as landslides and intense erosion by overland flow during large storms, than they do through slower, steadier processes like soil creep. In mountain drainage basins, sediment supplied by catastrophic mass wasting typically dominates the sediment supply to stream channels [**Photograph 6.2**]. In contrast, the sediment supplied to lowland channels generally comes from upstream channels and local bank erosion [**Photograph 6.3**].

Grain size and degree of sorting of the sediment supplied to a channel often profoundly influence sediment transport patterns, channel dynamics, and channel morphology of the stream. For example, steep channels that receive only sand from upstream bank erosion may rapidly transport their full sediment load and expose the underlying bedrock, while comparable channels that receive a wide range of particle sizes from a landslide off of a valley wall may develop a coarse surface layer of large clasts that move only during the highest flows, if ever.

Bed and Bank Material

The materials that form the bed and banks of a river control the frequency and rate at which the channel mobilizes sediment, erodes into its banks, and incises its bed. The

The **discharge** (Q) of a river is equal to the product of the average channel width (W), average depth (D), and average flow speed (U), which is commonly referred to as flow velocity, implicitly meaning the net speed of water flowing downstream. The portion of the channel bed in contact with the flow, and thus providing frictional resistance, is the **wetted perimeter** (P_w), which is approximately equal to the channel width plus twice the flow depth ($P_w = W + 2D$). The channel cross-sectional area is W × D.



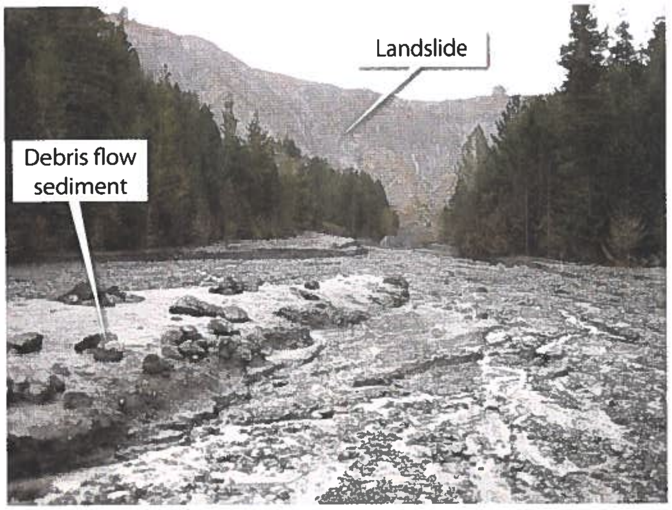
The downstream velocity of water flowing in a river increases in a logarithmic profile from the channel bed toward the surface, with the average downstream velocity at about 0.6 times the total flow depth.

The **shear stress** (τ) exerted on the channel bed by the flow is equal to the downslope component of the weight of the overlying water $\tau = \rho_w gDS$, where ρ_w is the density of water and g is the acceleration due to gravity. The small angle approximation, where $S \sim \sin\theta$, is often used.

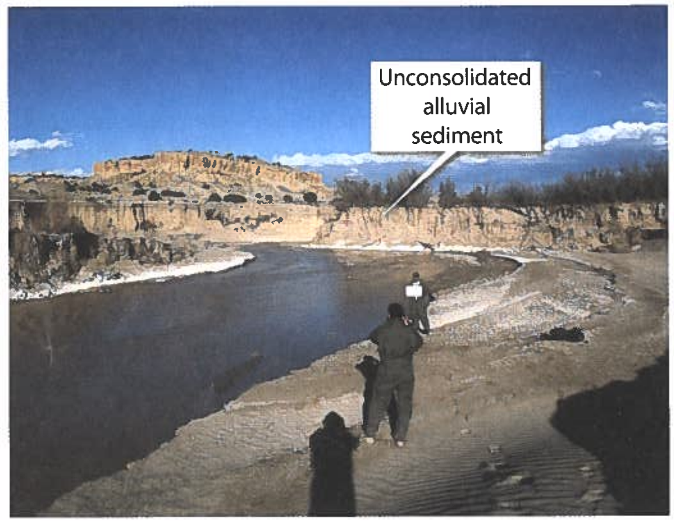
FIGURE 6.1 Channel Cross Section. Channel characteristics, including depth, width, slope, and water velocity, are measured for a variety of reasons, including calculating discharge (volume/time) and shear stress on the bed and bed sediments.

character of the bed and bank material also helps to set the tempo and style of channel migration. Some channels are composed of cohesive materials that resist bank erosion, such as bedrock or clay, whereas some are made of more erodible, noncohesive materials, such as sand and gravel, that are easily and frequently remobilized by streamflows.

Alluvial channels, most common in lower-gradient stream valleys that are not confined by valley walls, have a thick cover of sediment that shields the bedrock from active channel incision and allows channels to migrate laterally across valley bottoms. The beds and banks of alluvial channels are predominantly made of **alluvium,** unconsolidated material that is transported and sorted by the



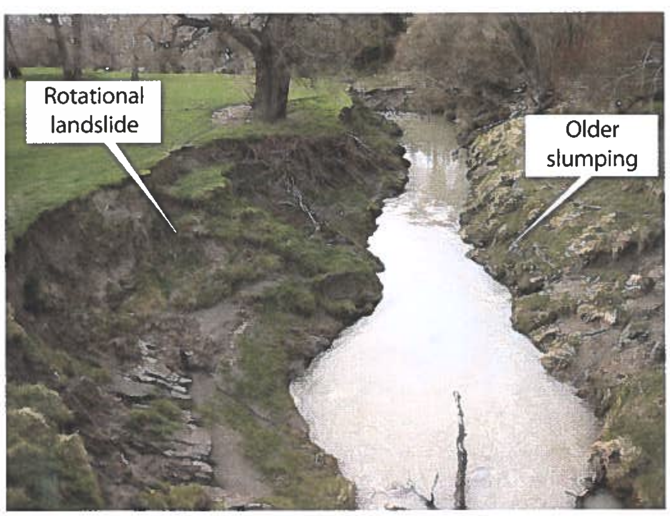
PHOTOGRAPH 6.2 Landslide and Debris Flow Sediment. Channel eroded into debris flow sediment issuing from the massive Tarndale landslide in the steep uplands of the Waipaoa drainage basin, located in northern New Zealand.



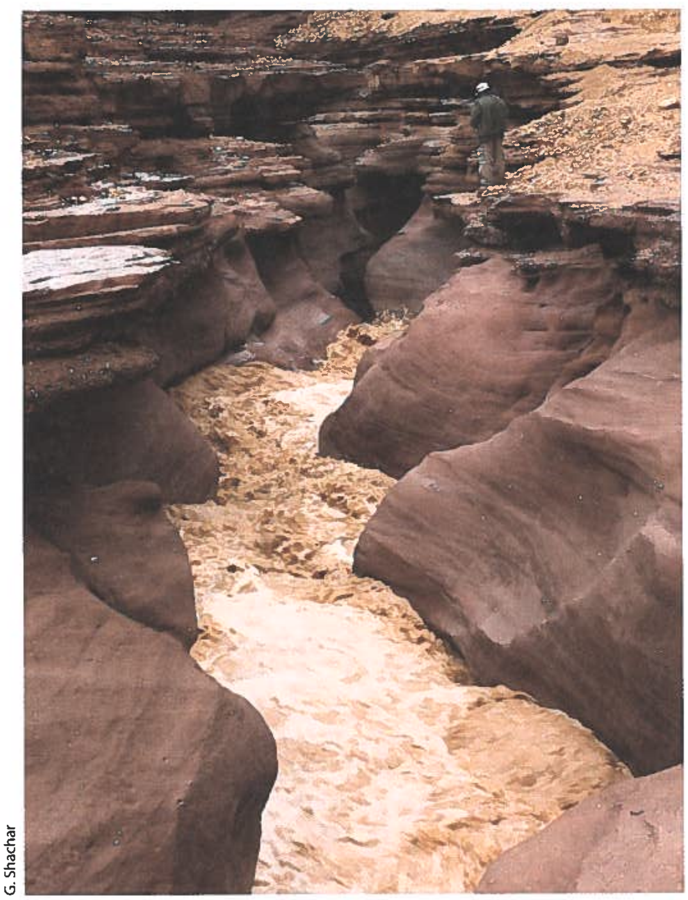
PHOTOGRAPH 6.4 Alluvial Channel. The Rio Puerco flows through banks composed of unconsolidated alluvial sediments in northern New Mexico.

flow [**Photograph 6.4**]. In contrast, streams with **bedrock channels** actively cut into rock and flow directly over bedrock or over a thin layer of alluvium [**Photograph 6.5**]. They typically occupy narrow valleys with rocky walls. Most bedrock channels are found in uplands and hilly or mountainous terrain, but some occur in relatively gentle landscapes that have undergone deglaciation, recent uplift or downcutting, or in places that have dramatic lithologic contrasts or a limited sediment supply.

The grain-size distribution of material carried by a stream is an important control on channel morphology, but it is often not reflected in the sediments that accumulate temporarily in and form the channel bed. Perennial streams generally sort the material they carry; coarser sediment collects in the channel bed beneath fast-moving



PHOTOGRAPH 6.3 Bank Failure. Rotational landslide and bank erosion on a small stream in the Waipaoa basin, North Island, New Zealand. Note the termination of the slump where the bank is stabilized by the large root system of the tree at upper left. The right-hand bank shows older slumping.

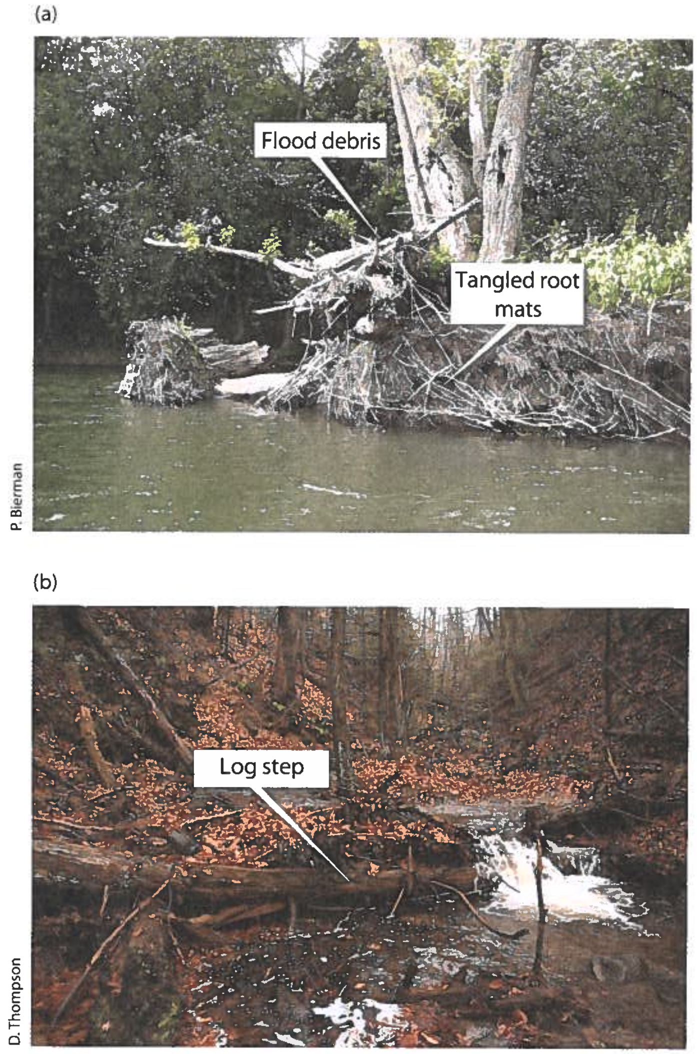


PHOTOGRAPH 6.5 Bedrock Channel. An ephemeral stream in flood is confined to a bedrock channel as it flows through Red Canyon in the southern Negev Desert of Israel.

currents, and finer grains travel farther downstream until deposited in calmer water. Further sorting of the channel bed occurs during high-discharge flows that mobilize larger sediment grains. Channel beds can be armored by coarse clasts that overlie finer material. Only when high flows mobilize the armor layer, can the fine sediment beneath become entrained in the flow.

Boulders introduced into channels by landslides can be too large for even rare high flows to transport. Accumulations of boulders sometimes form persistent obstructions to the flow, such as the unmovable lag deposits from debris flows where tributaries enter the Colorado River in the Grand Canyon. Debris from these flows creates the steep rapids for which the river is famous. Since the damming of the Colorado River, reduced flood flows have allowed additional coarse debris to accumulate at such tributary junctions. Streambed material in the ephemeral streams common in arid regions and at the upstream tips of most channel networks is generally poorly sorted, lacks a surface coarse layer, and closely matches the composition of the sediment supply. In arid regions, flows are usually of such short duration and are so uncommon that there is little transit time for grain sorting.

Different sizes of sediment move at different rates and have different fates in fluvial systems. Coarse materials moving slowly as **bedload** (which rolls along the channel bed) and **saltating** load (which leaves the bed in short hops before settling back to the bed) remain in the channel and can be incorporated into floodplains and terraces by channel migration. Finer material, suspended in the flow, moves downstream at the speed of the water. Some of the fine material is evacuated from the basin, some is deposited on floodplains if the flow overtops the bank, and some is deposited in the channel as flow wanes and velocities decrease.



PHOTOGRAPH 6.6 Woody Debris. (a) Tangled root mats from silver maple trees protect the upstream margin of an alluvial mid-channel island in the Winooski River, Vermont. Flood debris (mostly wood) is jammed on the upstream side of the island. (b) Large woody debris in a small stream in Cobalt, Connecticut, creates a log step, changing the local stream gradient and impounding water and sediment.

Vegetation

Riparian vegetation that grows along the banks of stream channels and **large woody debris** (logs) within channels influence channel processes and dynamics. Bank vegetation strongly affects channel shape and bank stability [**Photograph 6.6**]. Large trees growing in floodplains or along channels provide substantial root reinforcement (apparent cohesion) that stabilizes channel banks and reduces bank erosion. The stabilizing effects are most apparent along small, shallow channels where roots penetrate the full depth of the channel bank. Even in larger channels, where roots only penetrate a portion of the bank depth, root reinforcement still slows the pace of bank erosion. Deep rivers that undercut their banks can overwhelm the stabilizing contribution of roots.

Riparian vegetation also provides a source of in-channel woody debris (logs or mats of sticks, leaves, and roots) that acts as both mobile sediment and forms stable logjams. Accumulations of woody debris that slow, block, or divert streamflow influence patterns of streambed scour, deposition, and sediment transport at scales that range from a single pool scoured from a streambed where flow is forced around a logjam, to long-term storage of alluvium behind log dams in steep, bedrock-floored mountain channels. Fallen trees may redirect stream currents into channel banks, setting off a new round of channel shifting and sediment remobilization. In forested regions, the amount of organic debris supplied to a channel can be as important a control on channel morphology, processes, and dynamics as the stream's sediment or discharge regime. For example, forest stream channels vary greatly in width locally along their courses as a result of flow deflection from wood debris. In contrast, grassland stream channels are more uniform in width.

Fluvial Processes

Understanding the discharge of water and the transport of sediment is crucial for understanding river dynamics because these are the most important processes that interact to form channels. The ability of a stream to erode, transport, and deposit sediment reflects the balance between driving and resisting forces. The shear stress (drag force per unit area) exerted by water flowing over the channel bed provides the impetus for sediment entrainment and transport. The shear strength of the bed and bank-forming materials resists erosion. The total load of a stream includes fine-grained sediment suspended within the flow; coarser sediment that slides, rolls, and bounces along the channel bed; and the material dissolved in the flow itself. Together, these characteristics influence the ability of channels to incise into bedrock and migrate laterally.

Flow Velocity

The flow velocity (U) in a channel depends on the gravity-impelled fluid driving force that is controlled by the flow depth and the slope of the channel (S), as well as the resisting force generated by the frictional resistance provided by the channel bed and banks. This frictional resistance is characterized by the channel **roughness.** Flow velocity is measured in meters per second, but channel roughness cannot be measured directly because it includes the integrated resistance to flow from the viscosity of the water and the irregularities of the channel bed and banks. The overall roughness of a channel reflects the resistance caused by obstructions that protrude into and impede the flow, including individual sedimentary particles, clusters of particles, and larger-scale **bedforms.**

Manning's equation relates the velocity of flow in a stream (U) to characteristics of the channel,

$$U = [R^{2/3}\, S^{1/2}]/n \qquad \text{eq. 6.2}$$

where S is water surface slope, R is called the **hydraulic radius** and is defined as the cross-sectional area of the flow (A_{cs}) divided by the wetted perimeter (P_w), and n is the **Manning roughness coefficient.** As the cross-sectional area of a rectangular channel is the product of width and depth ($A_{cs} = WD$) and the wetted perimeter is approximately equal to $W + 2D$, the hydraulic radius R may be approximated as the flow depth ($R \approx D$) in wide channels, where W is much greater than D (i.e., $W >> D$).

The Manning roughness coefficient (n) is one commonly used empirical assessment of frictional resistance to flow. The coefficient can be back-calculated if the depth, water-surface slope, and downstream water velocity are known, but it is usually estimated from channel characteristics or using comparisons with channels of known roughness value. It is typical to assign a single value of n to a channel reach, but roughness values actually change with discharge because obstacles become submerged at high flow. For natural channels, high-flow values of the Manning roughness coefficient range from 0.01 in smooth, sandy channels to ~0.2 in wood-clogged, bouldery channels. Note that when the equation is in feet rather than meters, the numerator on the right-hand side needs to be multiplied by 1.49 to account for unit conversion, because the equation is not dimensionless.

Channel roughness generally decreases as the flow depth increases in channels. At greater flow depths, friction on channel boundaries influences a smaller proportion of the flow, and deeper flows submerge large roughness elements, such as boulders and logs. Flow velocities in channels thus typically increase during floods. However, when the flood grows large enough that water spills over the bank, roughness greatly increases as shallow floodwaters move through valley bottom vegetation, such as riparian forests where trees and thick underbrush impede flow, slowing the water, and causing sediment to drop out on the floodplain. For example, Manning n values for overbank flows through woody vegetation commonly range from 0.07 to 0.2, many times higher than in most channels.

Hydrologists have developed a number of methods to estimate flow roughness from channel characteristics. One popular technique is to compare a channel with pictures of channels of known flow depth, slope, and velocity for which the roughness has already been calculated. For example, U.S. Geological Survey Water Supply Paper 1849 shows an array of channels with measured roughness values to which a channel of interest can be compared in order to arrive at a quick estimate of an appropriate roughness coefficient **[Photograph 6.7]**. Another method is to estimate roughness by consulting a table of channel characteristics (**Table 6.1**).

Bed and bank roughness that slows flow along the margins of a channel usually results in maximum flow velocity near the surface in the middle of the channel (see Figure 6.1). Flow velocity increases from zero at the bed to maximum velocity directly below the water surface. Flow at the surface is somewhat slower because of the drag associated with the air–water interface and because of vortices of more slowly flowing water shed from the channel margins. Flow immediately above the channel bed is typically characterized by a thin layer of water, sometimes called the **laminar sublayer,** where water molecules travel smoothly along parallel flowpaths. Higher in the water column, the velocity increases logarithmically toward the maximum near the surface. Because of this nonlinear increase, average flow velocity generally occurs at about 60 percent of the flow depth, and the average speed of the flow is usually about 80 to 90 percent of the surface velocity.

Discharge is not uniform across a channel, so determining an average velocity in the channel is best done by measuring flow depth and velocity at a number of locations (typically at least 10) across a channel and then calculating an average discharge value. This can be done either by averaging

PHOTOGRAPH 6.7 Channel Roughness. Channels have different roughness values, expressed as Manning n values. Images taken from the USGS Water Supply Paper 1849, a compendium of channel roughness values with photographs and cross sections of channels in which the n values were measured. (a) n = 0.026, Indian Fork below Atwood Dam, near New Cumberland, Ohio. (b) n = 0.036, West Fork Bitterroot River near Conner, Montana. (c) n = 0.057, Mission Creek near Cashmere, Washington. (d) n = 0.073, Boundary Creek near Porthill, Idaho.

many measurements of depth and the downstream component of the flow velocity and multiplying the averages by the stream width, or it can be done making a series of discharge measurements across the channel. The latter method requires sectioning the channel into virtual flow tubes. Each flow tube defines a width increment of stream that appears to have similar depth and velocity. Starting at either the right or left bank, as seen when facing downstream looking in the direction of the flow, one measures the width, average depth, and downstream water velocity (using a current meter) in the first increment and calculates a discharge. This process is continued

TABLE 6.1

Typical Manning coefficient values (n)

Channel-bed material	n =
Straight canal/concrete banks	0.01–0.02
Straight canal/earthen banks	0.02–0.03
Sand	0.01–0.03
Sand/gravel	0.03–0.05
Cobble/boulder	0.04–0.08
Timber/vegetation-choked	0.07–0.16

across the entire channel width and the resulting incremental discharges are then summed.

Measurements of discharge and **stage** (water-level elevation) taken at a range of flows over time define a **rating curve** (a plot of discharge versus stage) that allows for estimating discharge at different water levels (see Figure 2.8). Flow stage is readily converted to flow depth if one also knows the elevation of the streambed. Once a rating curve has been constructed for a channel, discharge can then be estimated by simply measuring the flow stage. This can be done manually by reading a **staff gauge** or can be automated using digital pressure transducers.

Most natural channels exhibit **turbulent flow** in which the velocity continuously fluctuates, producing eddies that mix the flow and greatly increase the flow resistance. Different types of turbulent flow (defined below) determine what kind of bedforms develop on the channel bed. The specific type of flow that occurs can be calculated using the **Froude number** (Fr), defined as the ratio of the flow velocity (U) to the speed at which a surface wave will propagate, which is given by the square root of the product of flow depth (D) and gravitational acceleration (g):

$$\mathrm{Fr} = U/(D\,g)^{0.5} \qquad \text{eq. 6.3}$$

Tranquil (or subcritical) flow happens when $\mathrm{Fr} < 1$, during **critical** flow $\mathrm{Fr} \approx 1$, and **supercritical** flow is when $\mathrm{Fr} > 1$. The Froude number indicates whether the water is moving faster or slower than its own wake, so a simple test for the difference between subcritical and supercritical flow is to toss a rock into the stream. If the ripple from the splash travels upstream against the current, then the current is subcritical. But if the current sweeps the expanding splash downstream, then the flow is supercritical. At critical flow, there are standing (or stationary) waves. These three flow regimes are geomorphically important because they transport and deposit sediment differently and are thus related to development of distinctive bedforms on channel beds.

Discharge Variability

Flows that fill a channel to the point of overflowing are called **bankfull flows [Photograph 6.8]**. The frequency with which a channel experiences bankfull flow varies from several times a year in humid environments to once every few decades in arid regions. When discharge exceeds bankfull stage, flood flows spill out over the channel banks and inundate valley bottoms, as discussed in Chapter 4.

Flows of different size have different velocities and thus transport differing amounts and sizes of sediment. The discharge that transports the most sediment over a period of years to decades is called the **effective discharge.** More extreme events (such as the flow that occurs on average once every 50 years) move more sediment per event than flows that recur more frequently. However, such extreme events are too rare to dominate alluvial channel development, although extremely large flows may cause **avulsions** or cutoffs, which change the path of channels, altering their form and the direction and location of flow.

P. Bierman

PHOTOGRAPH 6.8 Bankfull Flow. Bankfull flow of high turbidity water after a very heavy rain, shown at Small Creek near Nelson, North Island, New Zealand.

In perennial humid-temperate streams, effective discharge often corresponds to the bankfull flow with a one-to-two-year recurrence interval. In monsoon-driven stream systems, the effective discharge corresponds to the monsoon-season baseflow because that flow is large and lasts long enough to carry a lot of sediment. In arid channel systems, the effective discharge usually has a long recurrence interval, reflecting the rarity of precipitation events long enough and intense enough to generate significant runoff.

In channels with alluvial banks, the bankfull flow is important geomorphically because it generally represents the discharge to which channel width and depth are adjusted. In humid regions, channel dimensions thus reflect flows that recur on average every year or two. In arid regions, channels are shaped by longer recurrence interval events because bankfull flows are much less common than in humid regions.

A channel's **"at-a-station hydraulic geometry"** describes how its width (W), depth (D), and velocity (U) vary as discharge (Q) rises and falls **[Figure 6.2]** over the course of many different flow events. Hydraulic geometry is defined using three linked equations that describe how these factors increase with discharge through a given channel cross section:

$$W = a\,Q^{b} \qquad D = c\,Q^{f} \qquad U = k\,Q^{m} \qquad \text{eq. 6.4}$$

where a, c, and k are empirically determined constants, and b, f, and m are empirically determined exponents. Because discharge is the product of the flow width, depth, and velocity ($Q = W D U$), the product of the constants must equal 1 ($ack = 1$), and the sum of the exponents must also equal 1 ($b + f + m = 1$). The wide range of b, f, and m values empirically determined for at-a-station hydraulic discharge relationships (0–0.5, 0.3–0.6, and 0.2–0.6, respectively) reflects differences in channel cross-sectional form that reflect, for example, whether the stream flows between resistant bedrock outcrops or

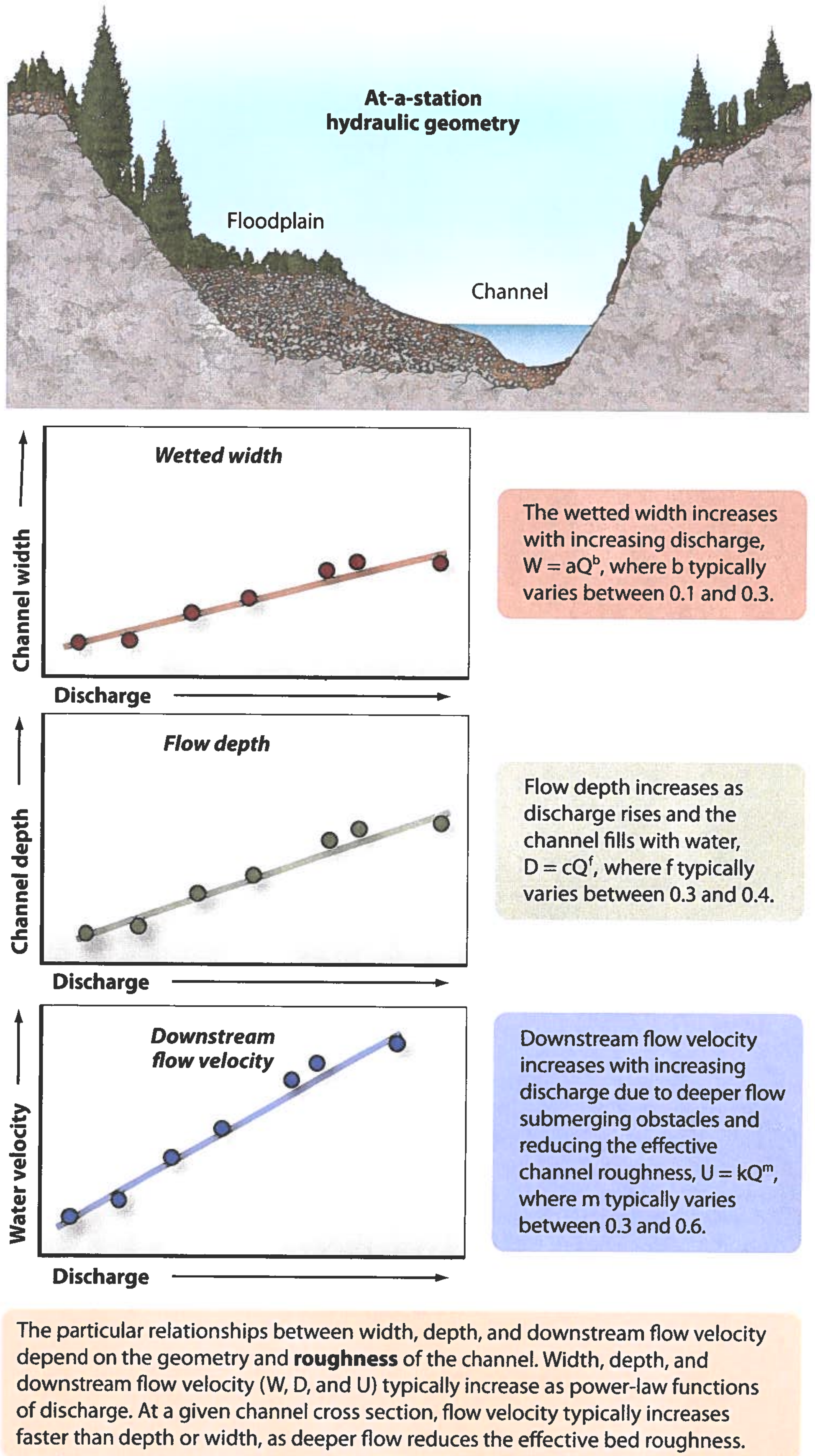


FIGURE 6.2 At-A-Station Hydraulic Geometry. The hydraulic geometry of a channel cross section (at a station) describes the relationships between increasing discharge and the flow width, depth, and velocity.

erodible alluvial banks, which may be scoured back during high flows.

Flow depth at a particular channel cross section (at a station) typically exhibits significant proportional change as discharge increases, but channel widths are generally constrained by the streambanks. Consequently, increased discharge is mostly accommodated by increased flow depth and velocity. Knowing the hydraulic discharge relationships for a location allows predictions of flow depths and velocities at a range of discharge values.

Stream Power

The ability of a stream to transport sediment and carve into the underlying bedrock is related to its **stream power** (Ω), the rate of potential energy loss per unit channel length, which is defined as

$$\Omega = \rho_w g Q S \qquad \text{eq. 6.5a}$$

where ρ_w is water density, g is gravitational acceleration, Q is discharge, and S is channel slope ($\tan\theta$); see **Box 6.1** for derivation.

Unit stream power (ω) is the rate per unit area of channel bed at which the river expends or dissipates its potential energy in the process of flowing downstream,

$$\omega = \Omega / W \qquad \text{eq. 6.5b}$$

where W is the channel width. Unit stream power is equivalent to the product of shear stress and flow velocity (i.e., $\omega = \tau U$). If the same discharge flows through a narrower channel, the deeper and likely faster flow will have more stream power per unit channel bed and will therefore be capable of greater sediment transport or bedrock incision. Similarly, steep channels will have higher unit stream power than channels with low slope. The geomorphic implications of unit stream power are significant. For example, mountain stream channels, steepened by uplift, can have sufficient stream power to incise at high rates and thus keep up with uplift.

Bedrock Incision

Channel incision into bedrock requires mobilization of the alluvial cover to expose the underlying bedrock to erosion. In general, it takes more stream power (slope, discharge) to cut down into bedrock than to move sediment, so channel incision into bedrock occurs only after sediment is in motion [**Figure 6.3**]. In many cases, nonflood flows and frequent, smaller floods drive sediment transport processes, whereas larger, less frequent events control channel incision into bedrock. Thick sedimentary cover serves to protect the bedrock beneath a streambed from erosion. In contrast, a very low sediment load provides few clasts that can act as abrasive tools entrained in the flow to erode exposed bedrock. The highest rates of bedrock incision are thus expected to occur in channels with intermediate sediment loads (see Digging Deeper).

Streamflow incises into bedrock through a combination of **abrasion, plucking,** and **dissolution,** and the relative importance of each process depends on the bedrock type and stream morphology. Abrasion sandblasts bedrock with material

BOX 6.1 Derivation of Stream Power

Water flowing down a river channel loses gravitational potential energy as it drops in elevation. Conservation of energy requires that potential energy is transformed into other forms of energy, including frictional heating and performing work to transport sediment or erode the channel bed. The rate of potential energy loss, defined as the stream power, is a measure of the ability of the channel to move sediment and incise into rock. The potential energy (PE) of the water per unit channel length is given by the product of the mass of the volume of water, itself the product of water density (ρ_w) and gravitational acceleration (g), and the channel width (W), flow depth (D), and elevation (z):

$$PE = \rho_w g W D z \qquad \text{eq. 6.A}$$

The downstream rate of potential energy loss (dPE/dx) is given by how fast the water drops in elevation, the channel slope (S). Thus, stream power (Ω) is given by:

$$\Omega = \rho_w g W D U S \qquad \text{eq. 6.B}$$

where U is the flow velocity and S is the channel slope ($\tan\theta$ or dz/dx). Noting that the product of WDU is equal to the flow discharge (Q), this reduces to:

$$\Omega = \rho_w g Q S \qquad \text{eq. 6.C}$$

If we consider the amount of work that the channel is able to do on its bed, it is reasonable to formulate the downstream rate of power loss in terms of the rate of power loss per unit bed area, termed the unit stream power (ω). In this case, we divide eq. 6.C by the channel width, yielding

$$\omega = \rho_w g Q S / W \qquad \text{eq. 6.D}$$

an expression that suggests that the rate of channel incision in upland bedrock channels is a function of the local water discharge, slope of the channel, and channel width.

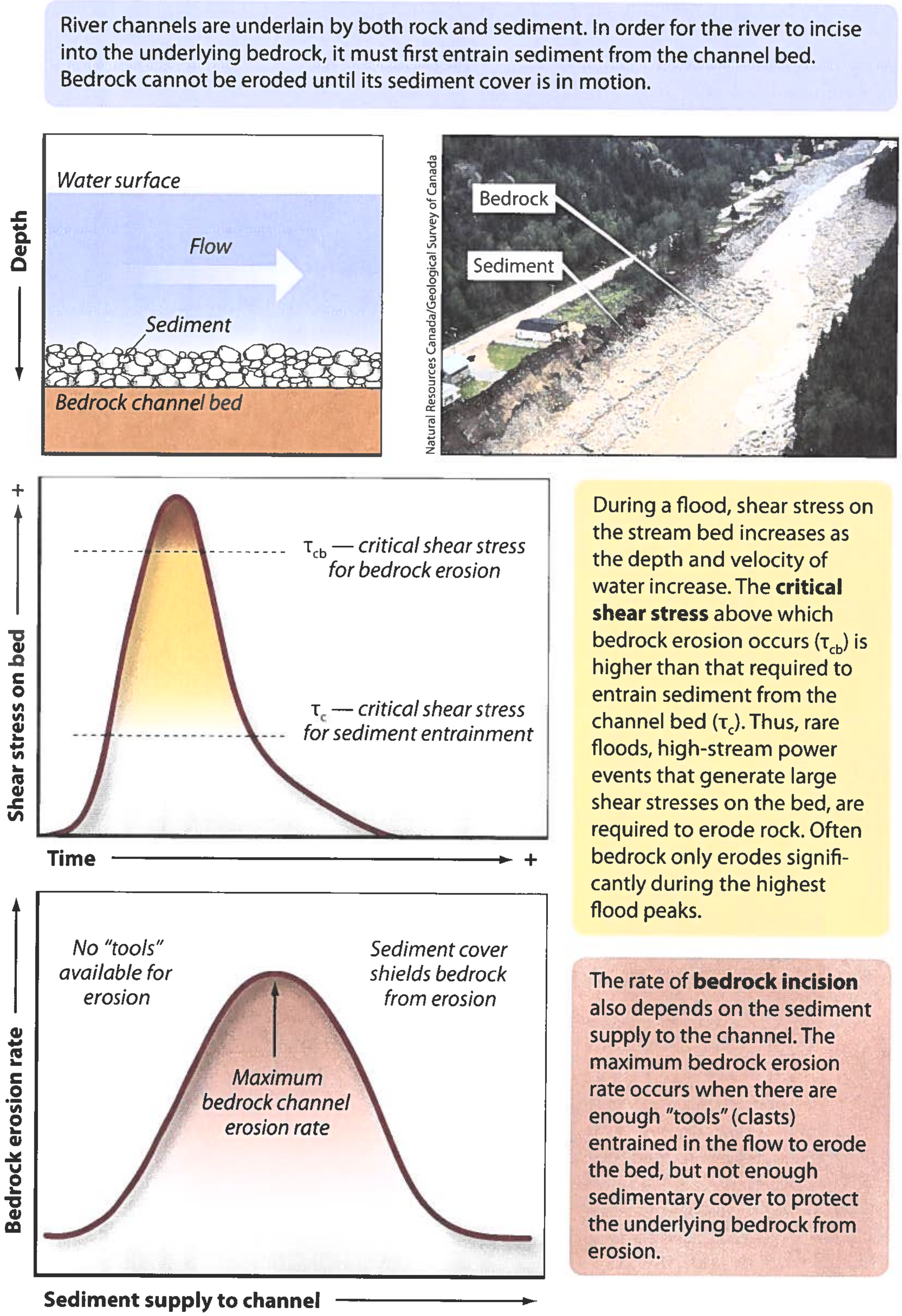
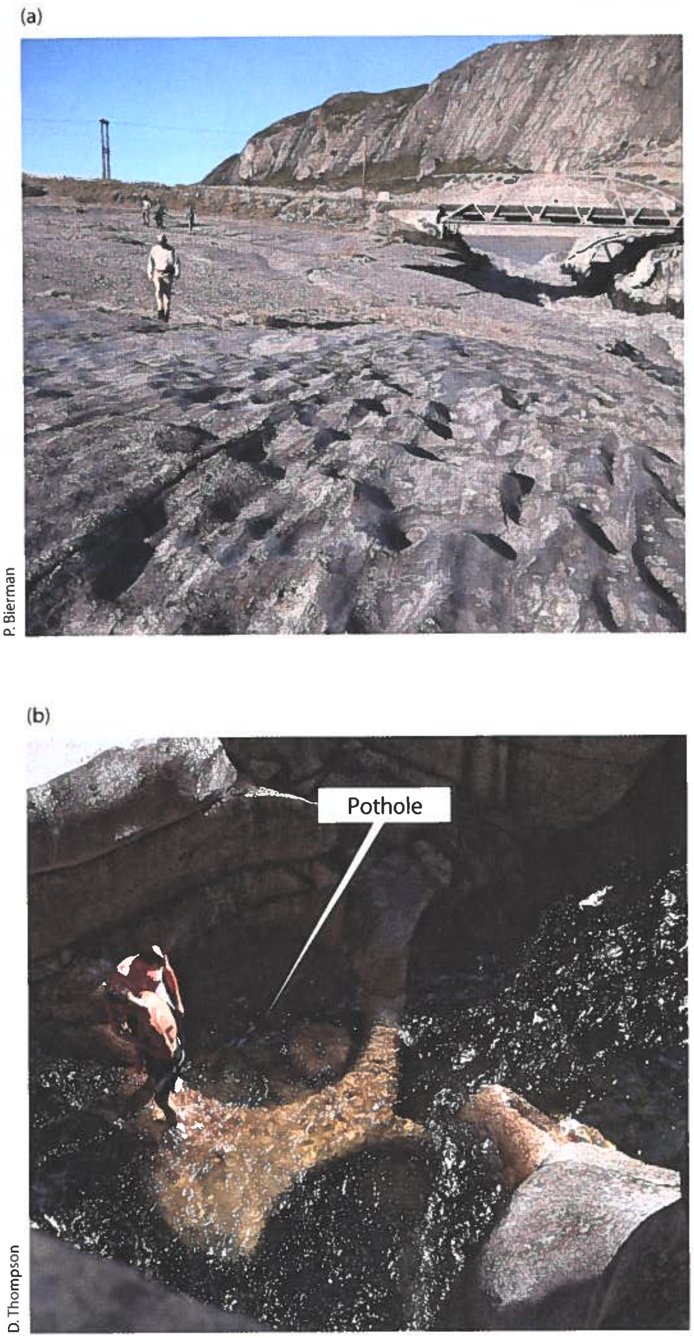


FIGURE 6.3 Bedrock Erosion Thresholds—Geomorphically Effective Events. Bedrock erosion requires that shear stress exceeds a critical threshold. The rate of bedrock erosion depends on the competing effects of the availability of sediment to provide tools for erosion and its effect on shielding the bedrock surface from impacting grains.

entrained in the flow, carving channels and sculpting rock into polished bedforms like potholes and flutes on cohesive, erosion-resistant channel beds [Photograph 6.9]. **Potholes** are cylindrical forms eroded into rock by rapidly moving vortices carrying abrasive, sand-sized sediment. Potholes can form vertically or horizontally. Once formed, potholes significantly lower rock-mass strength and therefore catalyze plucking, which is the removal of rock slabs



PHOTOGRAPH 6.9 **Bedrock Incision.** (a) Fluted bedrock surface on a former bedrock riverbed above the Watson River near Kangerlussuaq, western Greenland. (b) Large, round symmetrical pothole next to swimmers on the Ammonoosuc River, New Hampshire.

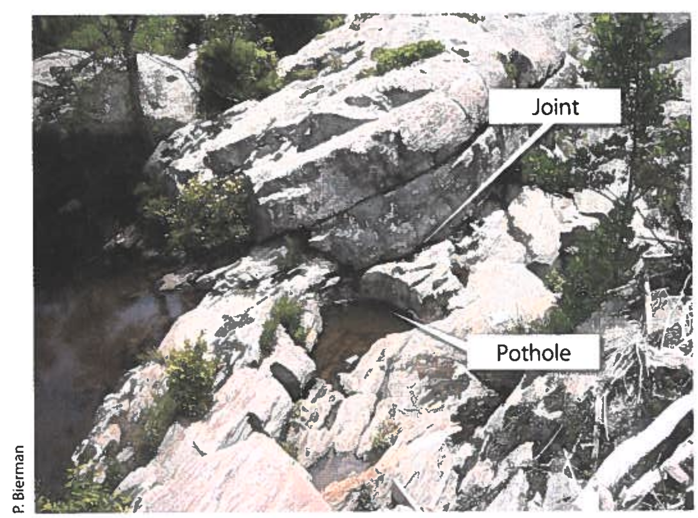
along joints or other fractures in the channel walls and bed. Adjacent potholes can erode into one another, undercutting whole sections of gorge walls.

Flowing water also loosens and plucks material from the beds of channels. Loose or broken rocks along fractures, joints, or bedding planes are particularly susceptible to plucking during floods **[Photograph 6.10]**. **Hydraulic wedging** of sand and gravel into openings helps loosen bedrock and make it susceptible to plucking. Streams also dissolve their way down through bedrock in regions with soluble rock, particularly limestone landscapes in humid climates where extensive dissolution leads to development of karst terrain (see Chapter 4). Dissolution often preferentially enlarges fractures or jointing in channel-bed rocks, which in turn promotes plucking.

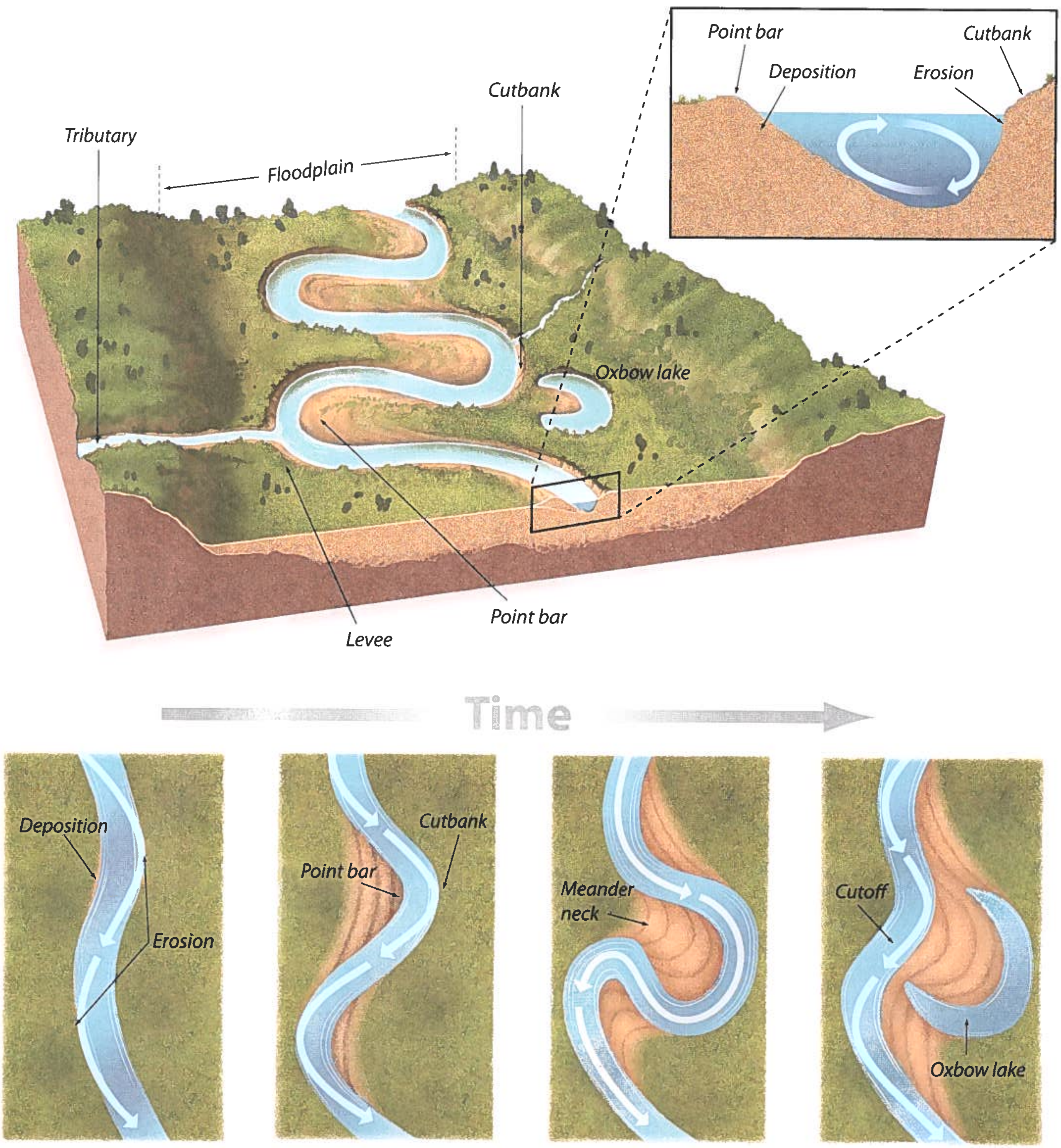
Channel Migration

The flow of water around channel bends determines in-channel patterns of sediment erosion and deposition and drives channel migration in alluvial channels. As flow enters a bend, the centrifugal force elevates the water surface on the outside of the curve. This superelevation sets up a cross-channel component to the water surface slope that drives flow down the outer bank and back across the channel, creating helical flow that spirals downstream in a roughly corkscrewlike pattern **[Figure 6.4]**. This secondary circulation results in a zone of converging flow that scours out pools along the outer bank of a bend, erodes the bank, and deflects flow back across the channel. Divergent flow on the inside of the channel bend likewise results in local deposition that builds up a crescent-shaped **point bar** from deposition of bedload, creating a topographic obstruction that steers flow back across the channel into the outer bank. **Cutbank** erosion on the outside of channel bends and point-bar deposition on the inner bank drive progressive channel migration laterally and downstream **[Photograph 6.11]**.

In almost all channels (not just meandering ones), the deepest part of the flow, called the **thalweg** (from the German, meaning "valley way"), follows a path through shallow riffles (areas of fast, shallow, turbulent flow) that connect deeper pools. During periods of low

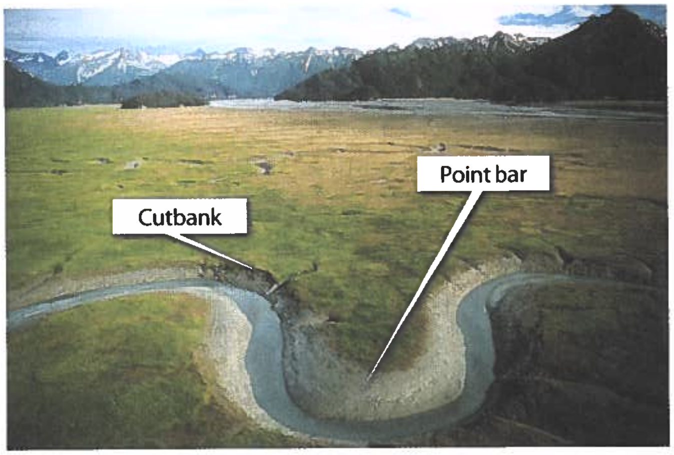


PHOTOGRAPH 6.10 **Plucking Jointed Blocks.** Joints in schist on the bed of the Potomac River, northern Virginia, provide weaknesses that allow tractive and lift forces, developed during flood flows, to pluck and move large pieces of rock. Note the pothole drilled on a gently dipping joint face in center of the image.



Flow through river bends focuses erosion on the outer side, creating **cutbanks,** and deposition on the inside of bends, creating **point bars** composed of bedload sediment. The combination of erosion on the outer side of bends and deposition on the inner side results in channel migration across the **floodplain** (toward the outside of bends and downstream). Helical flow, shown in the channel cross section inset, results in water flowing down along the cutbank and up over the point bar.

FIGURE 6.4 Meander Migration—Erosion and Deposition. Flow through meanders results in a predictable pattern of erosion and deposition as well as downstream translation of channel bends over time.



National Park Service

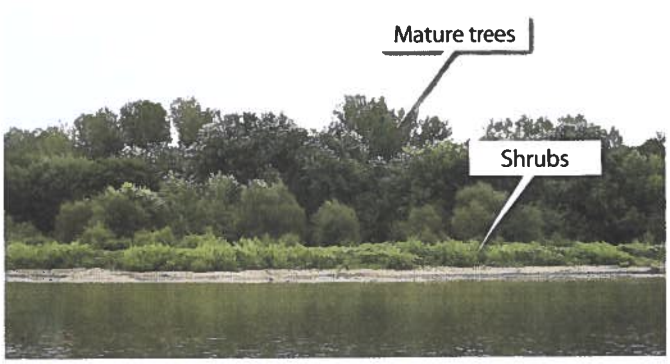
PHOTOGRAPH 6.11 **Point Bar and Cutbank.** Paired point bar and cutbank along a well-developed meander bend in Tuxedni Bay, Alaska.

flow, riffles have steep water-surface slopes and rapid flow, whereas pools have relatively flat water-surface slopes and slower flow. So how are pools scoured out? In contrast to conditions at low flow, high-flow velocities and water-surface slopes in the pools increase more rapidly than in the riffles, leading to a velocity reversal in which the water speed at high flow is greater in the pools than in the riffles. At high flows, shear stress increases toward the downstream ends of the pools, scouring and transporting coarse bed material to be deposited in lower velocity riffles where shear stresses decline downstream.

The combination of focused bank erosion at cutbanks and deposition of point bars on the inner bank causes channel migration. **Meanders,** the winding curves or bends in a river, migrate toward the eroding outer banks, and point-bar deposition fills the inside of the curve. In most cases, the rate of erosion on cutbanks matches the deposition rate on point bars, which then become buried by overbank deposition over time. The pace of bank erosion is controlled by bank erodibility, stream size, and flow velocity. In many channels, one can readily see evidence for this process of gradual channel migration in the progression of vegetation height (and age) away from the channel on the inner (point bar) side of meander bends [**Photograph 6.12**].

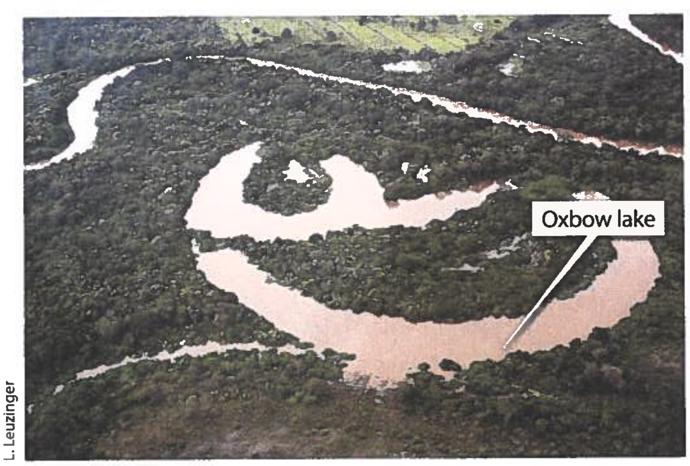
The point of maximum bank erosion is often just downstream of the meander apex, which leads to the downstream migration of meanders. However, obstructions like bedrock outcrops in alluvial floodplains or along valley walls can block downstream migration and force upstream meanders to bunch up. In-channel obstructions like boulders, ice dams, and logjams also influence meander patterns and locations.

In addition to gradual channel migration by cutbank erosion and point-bar deposition, channels may suddenly move to a new position, a process called **avulsion.** Typically, an avulsion happens in places where ongoing lateral migration brings the outer banks of two adjacent meanders so close that continued erosion allows them to intersect. When the upstream side of a meander captures the downstream side, it cuts off the intervening loop and creates a loop-shaped slough, called an **oxbow lake** [**Photograph 6.13**]. In addition to meander cutoff events, channels sometimes avulse when a flow obstruction like a logjam blocks the channel and causes flow to spill over the streambank and carve a new channel across the floodplain or shift flow into a secondary side channel or to an inactive, abandoned channel. Streams with multiple side channels are thus common in forested terrain where woody debris is large enough to create logjams capable of diverting flow. Stable logjams at the heads of diverted channels are often porous enough to allow some flow to enter, creating high-quality, low-disturbance fish habitat.



P. Bierman

PHOTOGRAPH 6.12 **Vegetation Height and Channel Migration.** Point bar on the Winooski River, Vermont, with "layers" of vegetation indicating time since disturbance. Closest to the river, only annual grasses are present; farther back are shrubs, then small trees, and, at a distance, mature trees.



L. Leuzinger

PHOTOGRAPH 6.13 **Oxbow Lake.** This oxbow lake formed from an abandoned river meander and is now isolated from the channel of the Rio Aquidauana, Mato Grosso do Sul, Brazil.

Sediment Transport

Streams dissipate most of their energy by friction and turbulence. Only a small fraction is available to erode and transport sediment, but even so, streamflow is a key agent of sediment erosion, transport, and deposition on the continents. The physics involved in rigorously analyzing the movement of sediment in rivers is complex enough to have reportedly discouraged a young Albert Einstein from pursuing a career in physical geography. Consideration of the forces acting on particles within the flow and on the streambed provides substantial insight into how rivers transport sediment. Such insight is geomorphically important because the processes of sediment entrainment, transport, and deposition govern the sorting of material in transport, the development of systematic patterns of channel morphology, and the formation of channel bedforms.

Initiation of Transport

Water flows in a channel under the influence of gravity (see Figure 6.1). The ability of streamflow to displace or erode the material in its bed, whether to transport sediment or incise bedrock, is due to the **shear stress** (τ), the force per unit area that a river exerts on its bed:

$$\tau = \rho_w g D \sin\theta \qquad \text{eq. 6.6}$$

where ρ_w is the density of water, g is gravitational acceleration, D is flow depth, and θ is the water surface (channel) slope in degrees. The flow depth is location-specific and is typically an average value for the area over which the channel slope is determined. For the small slopes typical of riverbeds, $\sin\theta \approx \tan\theta$, and geomorphologists commonly adopt this small-angle approximation for applications in low-gradient channels. This means that $\tau \approx \rho_w g D \tan\theta$, where $\tan\theta$ is the drop in water surface elevation, divided by downstream distance for a given channel segment. A second common assumption is that the water-surface slope may be approximated by the channel slope.

The flow of water over sediment particles exerts both lift and drag forces that act to drive sediment transport (see Figure 1.11). The submerged weight of the particles and the frictional resistance between neighboring grains act to resist particle movement. The degree to which a grain protrudes into the flow and its surface exposure to the

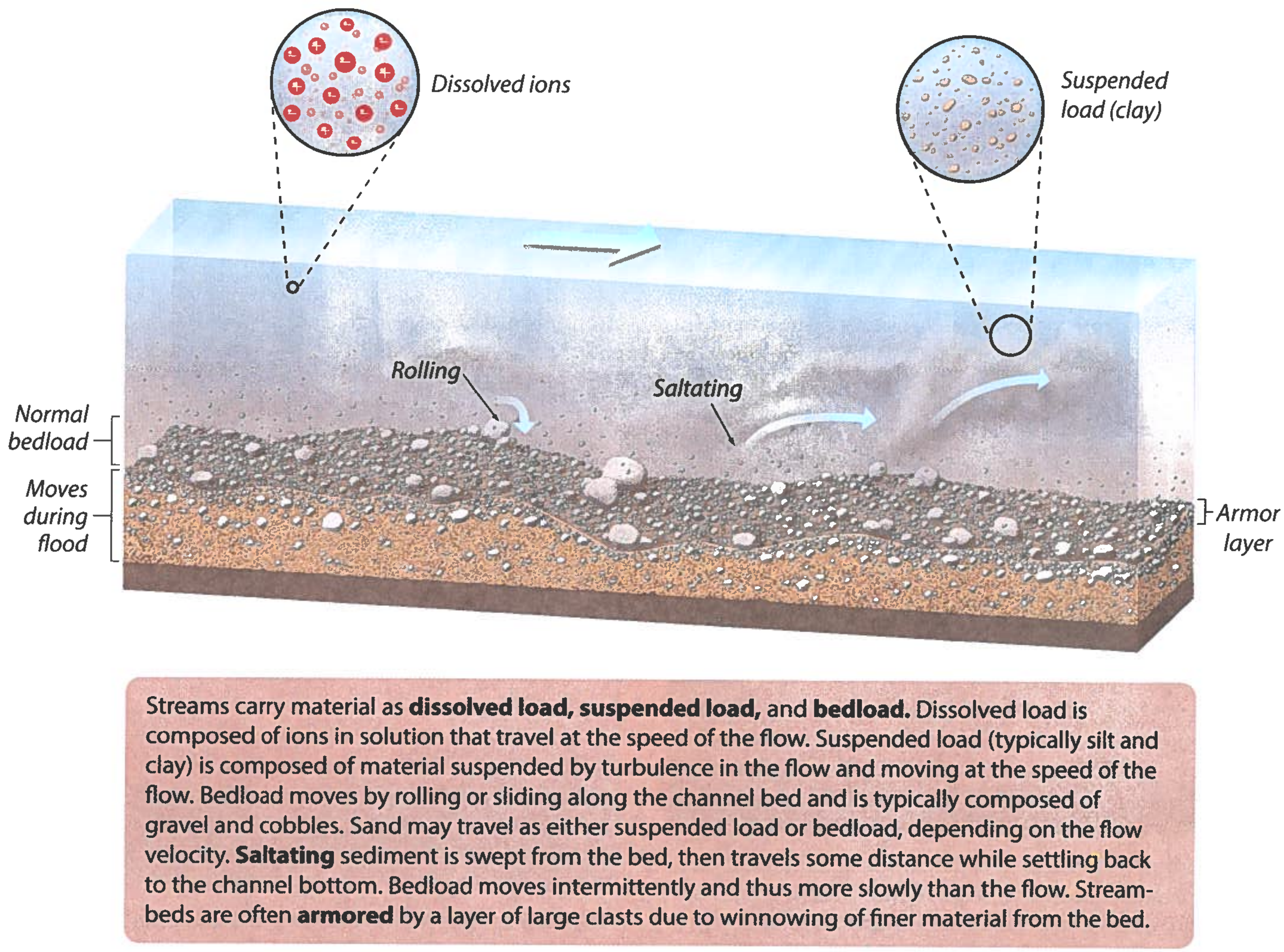
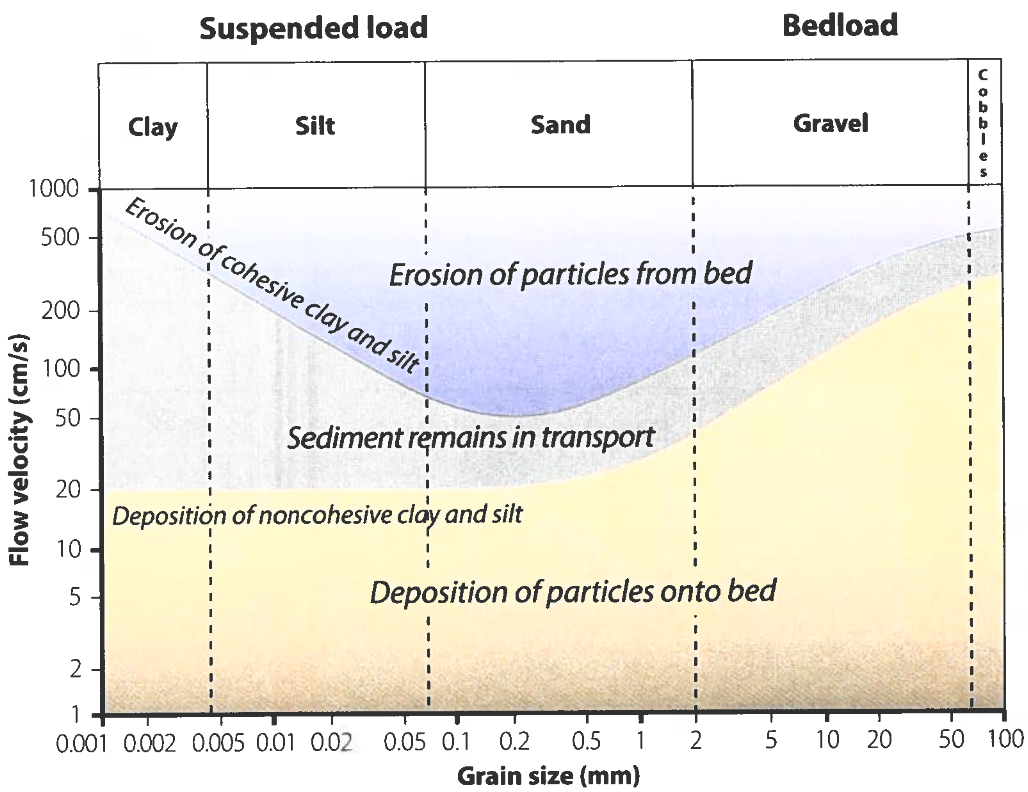


FIGURE 6.5 Stream Load. Streams and rivers carry sediment as bedload, suspended load, and dissolved load.

flow influence the drag forces acting on the grain. When the driving forces exceed the resisting forces, sediment begins to roll along the streambed or is thrust up into the flow (**saltation**), where it is carried along and settles at a velocity that depends on clast size and density. Upward-directed velocity fluctuations keep small particles suspended in the flow while larger sediment rolls, slides along the bed, or bounces before it settles back to the bed [**Figure 6.5**]. Flow velocity increases rapidly above the bed (see Figure 6.1), so sediment tends to move downstream once it is entrained in the flow.

Entrainment of material from the streambed is necessary to initiate bedload transport [**Figure 6.6**]. Resistance to motion depends on the size, shape, and density of sediment particles; their interlocking relationship with neighboring grains; and their exposure to the flow. Although not all particles on a streambed are mobilized at the same time, streambed gravels generally mobilize at flow velocities that exceed a **critical shear stress** (τ_c), characterized by

$$\tau_c = \tau_c{}^* g(\rho_s - \rho_w) d_{50} \qquad \text{eq. 6.7}$$



Erosion: The flow velocity required to erode material from a channel bed is a function of grain size. Sand is eroded at lower flow velocities than both coarser material (gravel and cobbles) and finer-grained material (silt and clay). To erode silt and clay, water must be moving quickly enough to overcome the **cohesive** strength of the material. The greater velocity needed to erode larger particles reflects their greater mass.

Transport: For large grain sizes that travel as **bedload,** there is little difference in the flow velocity required for erosion and deposition. In contrast, smaller particles that travel as **suspended load** can remain in motion at flow velocities well below those required to erode them.

Deposition: Fine-grained material (silt and clay) settles out in very still water, whereas coarse-grained material settles out even in swift water.

FIGURE 6.6 Hjulström Curve. Different sediment grain sizes are entrained, transported, and deposited in streams at different but characteristic flow velocities.

where τ_c^* is the **Shields parameter** (also known as the dimensionless critical shear stress), ρ_s is the density of the sediment, ρ_w is the density of water, and d_{50} is the median diameter of the bed-forming grains. Because individual particles are locally mobilized when bursts of high-velocity, turbulent flow sweep the bed and kick clasts up into the flow, initial motion is generally defined not by the motion of a single clast but rather by a less precise observation of general bedload mobility.

Streambed mobility is usually analyzed using the median grain size because larger clasts that protrude above the bed protect their smaller-than-average neighbors from the flow. Shields parameter values are generally about 0.06 for gravel streambeds, but the value varies among different channels, a reflection of the relative degree of sorting and packing of the streambed material, and ranges from 0.03 for loosely packed gravel to 0.1 for coarse, well-packed cobbles. Shields parameter values are generally high for cohesive, fine-grained sediments, like clay, that require large shear stress (velocity) to initiate bed mobility. Once the critical shear stress threshold is reached and streambed sediment is mobilized, bedload transport rates generally increase with increasing shear stress or stream power **[Figure 6.7]**.

Sediment grain size determines the flow velocity at which sediment erodes and is deposited, in part because grain mass scales nonlinearly with grain diameter and in part because cohesive forces between grains become important at small grain sizes. For grain sizes larger than 1 to 2 mm, the critical-flow velocity required to erode material from a riverbed increases with grain size and is similar to the velocity below which sediment will be deposited. For cohesive materials smaller than about 0.2 mm in diameter, the flow velocity required to erode material from the bed increases with decreasing grain size because of the cohesion characteristic of fine-grained material like silt and clay (see Figure 6.6) and because fine particles do not extend above the laminar sublayer into the turbulent flow above.

Sediment mobility is a function of grain size. Once mobilized, fine-grained particles tend to remain suspended in the flow, even at very low flow velocities. Coarse grains rapidly fall back to the stream bottom and transport of coarse material ceases at flow velocities close to those at which it was entrained. Sediment finer than sand tends to travel in suspension, while gravel and coarser sediment travels along the bed, as bedload. Sand, however, can

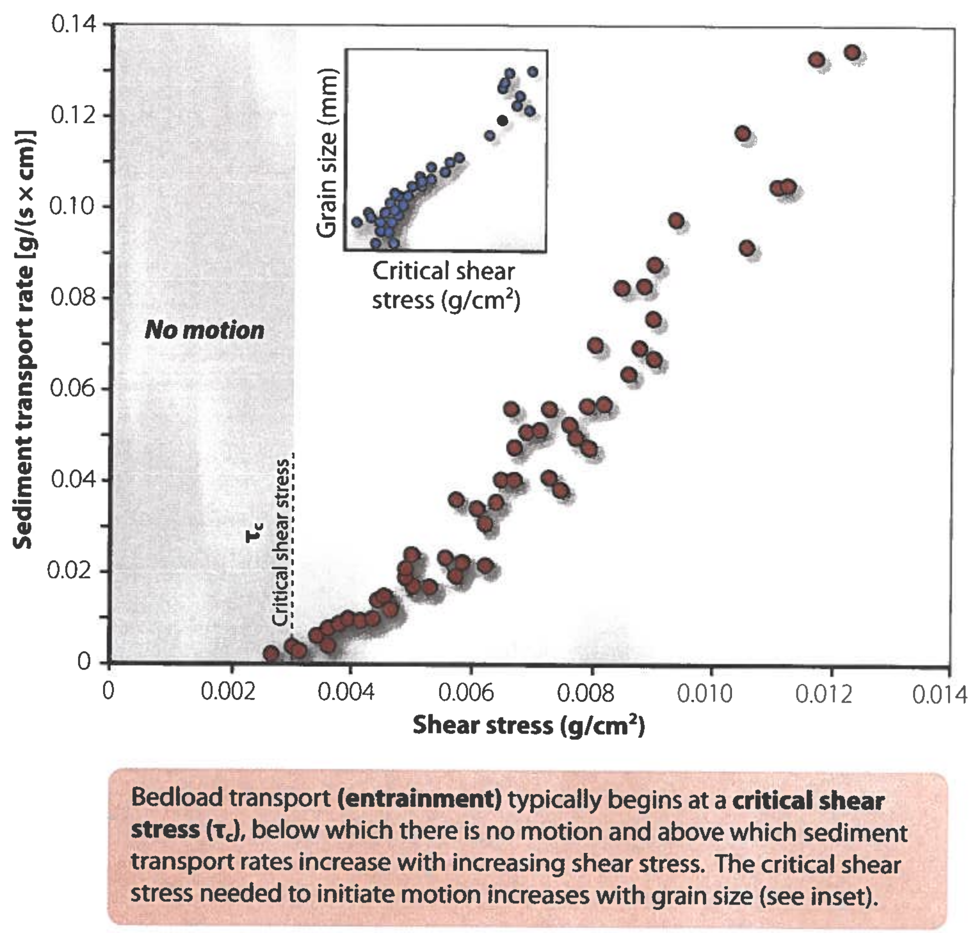


FIGURE 6.7 **Shear Stress and Transport Rates.** A critical shear stress must be exerted on the channel bed by flowing water before sediment can move. The critical shear stress needed to move sediment increases with sediment grain size. As shear stress increases, the sediment transport rate also increases. [Adapted from Leopold et al. (1964).]

travel either as suspended load or bedload in many river systems.

Sediment Loads

The total load of a stream consists of material dissolved in, carried within, or pushed along by the water flowing in a channel (see Figure 6.5). Loads vary between watersheds and over time. Most mass is carried by streams when they are in flood. Stream load can be partitioned into the **clastic load** (grains of sediment) and the **dissolved load** (ions and molecules in solution). Stream systems that drain steep catchments in rapidly eroding, tectonically active settings, like the Himalaya, carry most of their load as sediment grains, the clastic load. The relative amount of clastic load varies greatly among rivers, from less than 10 percent to more than 90 percent of the total load and averages about 75 percent for the world's largest rivers. The Saint Lawrence River, which drains to the North Atlantic, is a notable example of a river that carries a high percentage of its total load as dissolved material. Most of the clastic sediment is trapped by the Great Lakes, which are in the basin's headwaters.

The dissolved load consists of material contributed by chemical weathering and therefore reflects the weathering regime and solubility of the rocks within the drainage basin. Dissolved load concentration varies greatly, from ~40 parts per million (ppm) for the Amazon to ~850 ppm for the Colorado River, with a global average of about 120 ppm. A stream's total dissolved load is directly related to the discharge volume and the average concentration of dissolved material. The percentage of the total sediment load that is carried as dissolved matter varies greatly among streams in different environments. Channels that drain low-gradient catchments dominated by chemical weathering, or that have large lakes acting as sediment traps, carry most of their loads as dissolved matter. The dissolved load has little effect on fluvial processes and channel morphology; however, it is often critical for stream organisms and the health of downstream water bodies.

The **suspended load** is the part of the clastic load that consists of material fine enough to remain suspended by turbulence; it is carried along at a velocity similar to that of the water. For a grain to be held in suspension, its **settling velocity** (the rate at which it moves downward through still water) must be lower than the upward component of the velocity field created by turbulent eddies. The suspended load of most rivers consists of silt and clay because the settling velocity of a particle depends on its density and the square of its radius—large particles settle out faster than finer ones. The finer component of the suspended load is likely to stay suspended until it either ends up on a floodplain, river delta, or lake floor or settles out during waning discharge as a readily remobilized layer that blankets the streambed until the next high flow. This can be seen in patches of sand on bar tops along many gravel riverbeds and in clay drapes on sand ripples. Material fine enough that it never settles to the bed, even during low flow, is called the **washload.** Like the dissolved load, washload is transported downstream at the velocity of the flow with no geomorphic effect.

Bedload is clastic material that is transported by rolling, **saltating** (bouncing), and sliding along the channel bed. The material that makes up a streambed generally consists of bedload material that is periodically remobilized and deposited by fluctuating streamflow. Although bedload generally accounts for a minor portion of the sediment moved by rivers, typically amounting to 10 percent (and up to as much as 30 percent) of the total load, it is an important factor in determining channel morphology as it forms the bed and some banks.

The size of material transported as suspended load or bedload changes as the discharge volume and flow velocity rise and fall. For example, in many channels, sand moves as bedload at low flows but becomes part of the saltating and then suspended load during high flows. Material moving as suspended load during high-discharge events likewise settles to the streambed as a flood recedes. Gravel and coarser material generally travel as bedload in most rivers and streams, but even boulders can be temporarily suspended and moved downstream in large enough floods on big, steep rivers.

In contrast to the dependence of washload and suspended load on the supply of material delivered to the channel (most channels are capable of carrying far more suspendable material than they actually receive), bedload conveyance in alluvial channels is limited by the transport capacity of the channel. Bedload sediment transport rates (Q_b) generally increase nonlinearly with increasing discharge above the flow required to initiate bed mobility. However, prediction of bedload transport rates from channel characteristics like slope, depth, or stream power is complicated by the observation that rates of bedload transport vary greatly for similar hydraulic conditions in different channels. Consequently, the accurate prediction of bedload transport rates often requires calibration against field measurements.

Direct measurement of representative bedload transport rates is difficult because of temporal and spatial variability and because placing a sample collection device on the streambed disturbs stream flow and bedload transport. Also, it is usually neither safe nor feasible to collect samples during flood flows that mobilize coarse riverbeds. Clever ways to measure bedload transport include recessing open troughs in small streambeds and tracking the movement of magnetically tagged particles as they move downstream. Where conditions and access allow, bedload can also be measured directly with a sampling device that resembles a sturdy butterfly net, tipped on its side at the end of a long pole (called an Elwha sampler). Placed flush with the streambed, material rolling into the net is retained in a trailing mesh bag.

Wood and other floating organic matter sometimes constitute a substantial component of the overall load carried by a stream. Unsaturated wood is less dense than water and floats downstream as washload. Saturated logs can sink and become part of the bed material. In forested regions, extensive logjams can partially obstruct, dam, and even divert streams. The presence of a root ball enhances the stability of a log because the heavy, waterlogged wad of roots acts like an anchor as a log grounds out and begins to deflect flow. Logs that are large and waterlogged enough form stable **key members**, which are obstructions to flow that can capture additional wood and anchor logjams. The transport and storage of logs and organic debris within stream systems can greatly influence channel morphology and dynamics in forested regions.

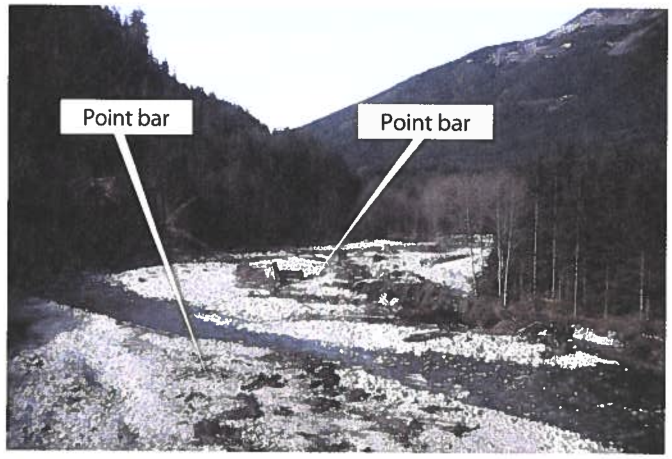
Bedforms

Streambeds are rarely smooth because erosion, transport, and deposition of sediment shape a variety of **bedforms** on channel floors that result from feedback between flow and sediment transport. **Bars** are large-scale, elongate bedforms that are often longer than the channel width and occur in many shapes and positions within channels. Some bars are relatively permanent features that form in local sediment storage zones where sediment accumulates during high flows [**Photograph 6.14**]. Persistent bars commonly develop at channel bends, confluences, or logjams along mobile alluvial channel beds. In floods, bedload particles generally move from one sediment bar to the next and are temporarily stored there between bed-mobilizing flows. Gravel-bed and sand-bed channels have similar types of bars, but their suites of finer-scale bedforms differ. Some bedforms are stable features while others are mobile and progressively advance up- or downstream.

Flow over a channel floor of readily deformable material, like loose sand, produces a range of bedforms [**Photograph 6.15**] that change with the flow depth and contribute substantial flow resistance. In such channels, the Froude number is geomorphically important because it determines the bed geometry. For example, plane beds and low-amplitude ripples commonly form in sandy channels at low-flow velocities or when deep flow ($Fr << 1$) generates a flow regime near the channel bed with minimal bedform roughness. Dunes with wavelengths of 4 to 8 times the flow depth and heights up to one-third of the flow depth form at higher flow velocities and/or shallower flow depths ($Fr < 1$). When flow approaches critical velocities ($Fr \approx 1$), dunes wash out and form an upper flow-regime plane bed devoid of bedforms. Supercritical flow ($Fr > 1$) at even higher velocities and/or shallower depths produces anti-dunes that are like dunes but face upstream and their form migrates upstream if the current is not too fast. Similar suites of bedforms develop in coarser-grained channels at high discharges.

In gravel-bed channels, bars are the dominant bedform, although other distinctive bedforms and erosional features can develop. Streamlined particle tails and clusters of coarse clasts aligned with the flow direction accumulate downstream of flow obstructions. Transverse-to-flow

(a)



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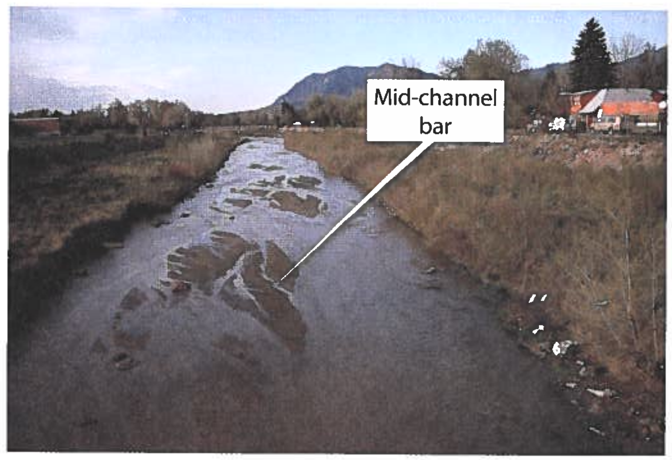
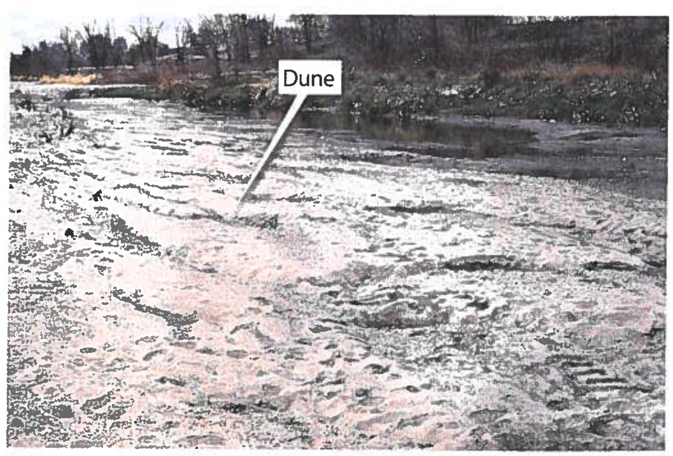


PHOTOGRAPH 6.14 Bars. Bars are common bedforms along channels carrying sediment. (a) Alternating point bars along the gravel-bedded, meandering Tolt River in western Washington State. (b) Mid-channel bars form in a sandy stream reach, outside of Colorado Springs, that has been artificially straightened.



PHOTOGRAPH 6.15 Dunes. These bedforms in the South Platte River, Colorado, are large sediment waves called dunes. Dunes form in subcritical flow with high sediment-transport rates.

ribs, which consist of repeated ridges of coarse clasts at a spacing determined by the channel width and the largest clasts, develop in areas of supercritical flow. Larger, stair-like gravel steps sometimes floor steep gravel channels, and channel-spanning steps of coarse clasts are common in cobble-boulder channels.

In perennial channels, a coarse-grained surface layer generally forms from winnowing of small particles from the bed surface during nonflood flows between full bed mobilizing events. This coarse surface layer (armor) typically extends down about twice the median surface particle diameter and overlies finer material that is more characteristic of the total bedload grain-size distribution once the bed surface is mobilized. Ephemeral channel beds in semi-arid regions generally do not have this coarse surface layer because they lack day-to-day flows that are capable of sorting and winnowing particles between flood events, and because abundant sand precludes formation of a gravel armor layer in most locations.

Channel Patterns

Channel morphology reflects the interplay of fluvial processes and the routing of material through drainage basins, stream-valley segments, and individual channel reaches. Distinct channel patterns and the morphologies of individual reaches arise from different balances between sediment supply and transport capacity, as well as from the bedrock structure, climate, and supply of large organic debris along the stream valley.

Stream channels are either single-thread or multi-thread and follow either relatively straight or tortuous paths. The **sinuosity** of a river can be defined in various ways, but a simple definition is the ratio of the channel length measured along the center of the channel to the straight-line distance measured down the valley axis [**Figure 6.8**]. High-sinuosity channels follow convoluted, twisting paths, and low-sinuosity channels follow relatively straight paths. Sinuosity and channel pattern change

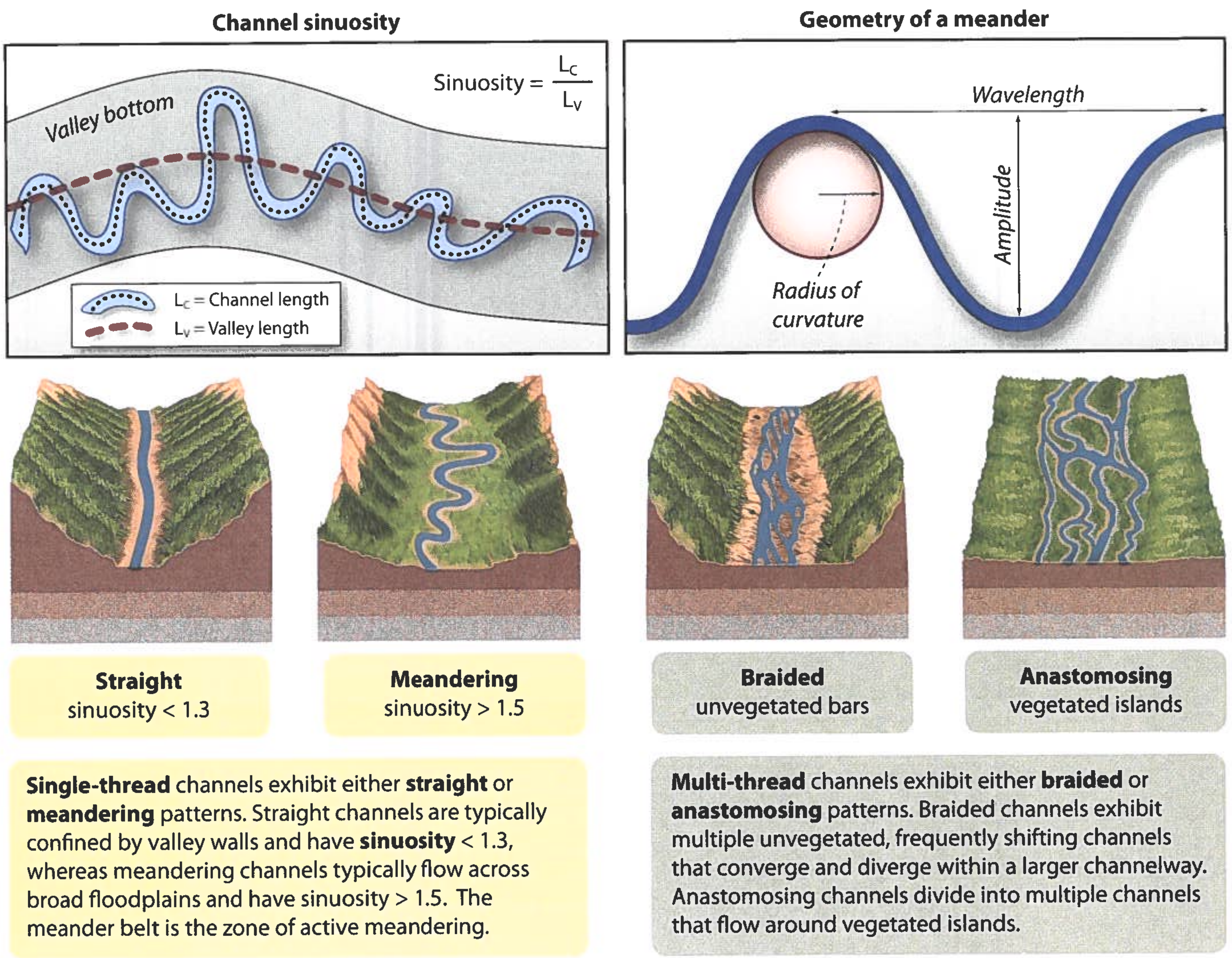


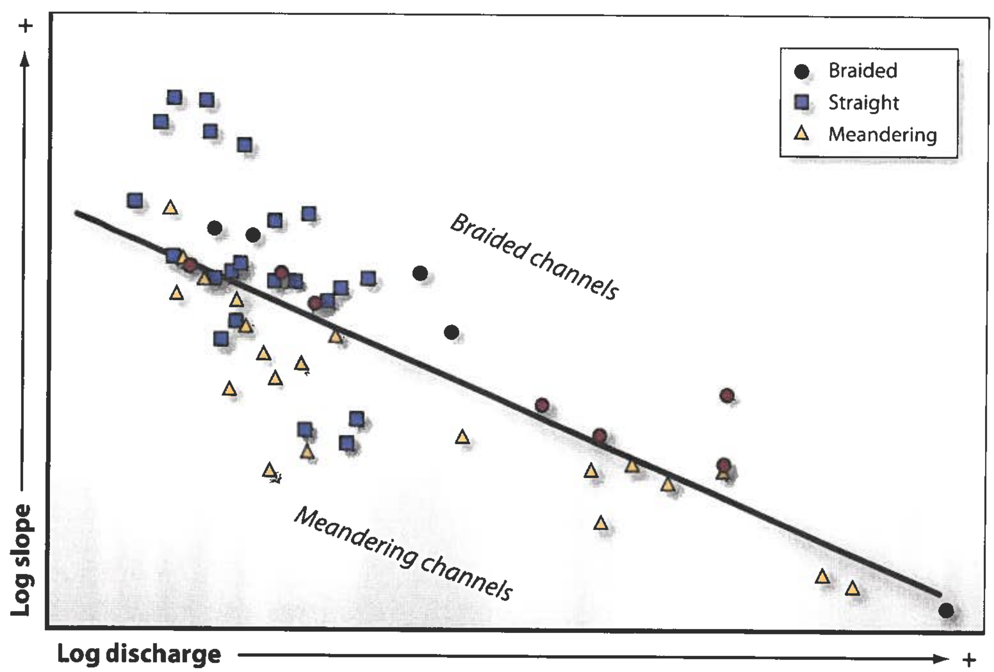
FIGURE 6.8 Channel Patterns. Channels can have a variety of forms, which are classified as single-thread or multi-thread. Both sinuosity and meander wavelength can be quantified.

from one reach to another downstream within a channel network as a result of local influences on sediment supply and discharge. Natural channels vary within a wide spectrum of patterns, but the simple distinction of **straight, meandering, braided,** and **anastomosing** channel patterns provides a useful framework for understanding the processes that control channel morphology (see Figure 6.8).

Different channel patterns arise from differences in bankfull discharge, gradient, sediment supply, and bank material. Braided channels occur on steeper slopes and with greater discharges and sediment loads than single-thread meandering channels [**Figure 6.9**]. Cohesive banks favor development of meandering channels, whereas weak, noncohesive banks favor development of braided channels. Local stabilization by vegetation or large organic debris leads to development of anastomosing channels that are split into multiple individual channels separated by stable, vegetated islands. Streams with highly erosion-resistant banks, such as bedrock canyons, often follow structural weaknesses like faults, joints, or sedimentary bedding planes. Although it has long been argued that meandering channels within deeply incised bedrock valleys inherited their courses from past channels cut in higher (now eroded) alluvial surfaces, it now appears that meandering channels can form over long periods of time as bedrock channels migrate laterally when carving down into competent bedrock.

Straight and Sinuous Channels

Straight channels with sinuosities < 1.3 are relatively rare in natural streams because even very slight flow irregularities lead to deposition and accumulation of sediment in **alternate bars** that are successively positioned on opposite sides of the channel. Sinuous channels (those with sinuosities of 1.3–1.5) are quite common because as soon as a subtle bar forms on one side of a channel, it steers flow coming from upstream toward the opposite bank, where the flow begins to excavate a pool as it impinges on and erodes the far bank. Flow returning across the channel is, in turn, directed into the opposite bank, leading to the development of another pool and bar downstream, and so on down the channel. This process leads to the development of a sequence of alternate bars and pools that promotes lateral channel migration and the growth of meanders through erosion of the cutbank. Most straight channels have erosion-resistant banks, follow structural controls like faults, or have been confined by engineered levees that prevent natural lateral migration (see Photograph 6.14b).



Discharge and slope influence channel planform patterns. At a particular slope, higher discharges are likely to produce **braided channels.** Likewise, for a particular discharge (or stream size), **meandering channels** tend to have lower slopes than do braided channels. **Straight channels** occur at low discharges over a variety of slopes. Sediment supply and the variability in discharge also play a role in determining channel form.

FIGURE 6.9 Channel Patterns as a Function of Discharge and Slope. Channel patterns are determined by slope and discharge. Braided channels are more common at higher slopes and higher discharges. [Adapted from Leopold and Wolman (1957).]

PHOTOGRAPH 6.16 Meandering Channel. Aerial view of meanders on floodplain of the Owens River, California.

Meandering Channels

Meandering channels have sinuosity values ≥ 1.5 and generally have a single, deep, narrow channel with few islands **[Photograph 6.16]**. Unconfined meandering channels typically develop a meander wavelength (the distance from the apex of one bend to the next on the same side of the channel) equivalent to about 10 to 12 channel widths, a relationship that is related to the physics of thalweg oscillation as water flows along an undulating channel. Pools are generally located on the outer banks of each bend and are thus typically separated by an average distance of 5 to 6 times the channel width.

The radius of curvature, the radius of a circle that fills the arc of a meander, describes the tightness of a river bend (see Figure 6.8). A small radius of curvature describes a tight bend. Feedback between flow through the bend and bank erosion typically leads to a radius of curvature between 2 and 3 times the channel width in meandering channels. In broad bends with a larger radius of curvature, the greatest shear stress on the outer bank occurs upstream of the meander apex, leading to preferential erosion that tightens the bend and decreases the radius of curvature. Conversely, in tight bends, where the radius of curvature is small relative to the channel width, the greatest shear stress and most intense bank erosion are downstream of the meander apex, which causes lateral migration that broadens the bend and increases the radius of curvature. Through such feedbacks, the typical form of meandering rivers represents an equilibrium adjustment to coupled flow and erosion through channel bends.

Meandering streams are common today, but we have found no evidence in the rock record that they existed before the evolution of land plants about 400 million years ago. Fluvial sediments from older periods of geologic time record braided channel morphologies. Conventional wisdom holds that this is because of the effect of root strength on bank stability, and recent flume experiments that demonstrate the role of streamside vegetation in stabilizing meander formation support this hypothesis. One would not expect meandering bedrock channels to be preserved in the geologic record for the simple reason that they occur in eroding upland environments.

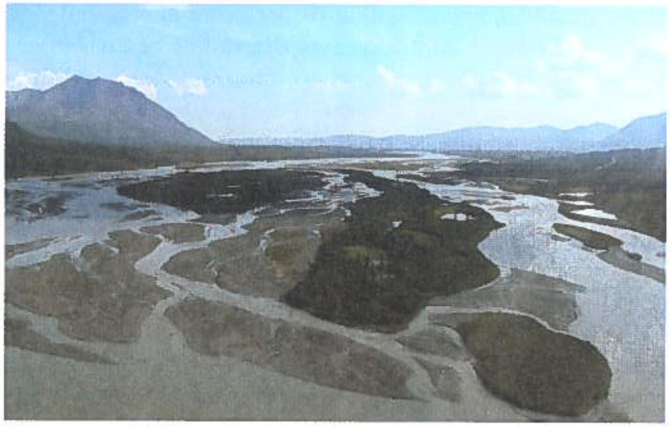
PHOTOGRAPH 6.17 Braided Channel. The braided Resurrection River in Alaska flows between steep valley walls.

Braided Channels

Braided channels are made of multiple, active threads within a broad, low-sinuosity, high-flow channel. A series of shallow, wide, low-flow channel strands that branch, diverge, and converge again form a distinctive braided pattern within the banks of a typical braided channel **[Photograph 6.17]**. The sediment bars that divide the flow into multiple strands are called **braid bars.** Braided channels generally are quite dynamic, and individual strands shift positions within the main channel, sometimes on a daily basis, as ephemeral patterns of deposition and erosion shift the sediment that makes up the braid bars. Erodible banks and a sediment load that exceeds the stream's carrying capacity favor formation of braided channels because these factors force the stream to flow around its own sediment at low flow. High slope, frequent variations in discharge, lack of bank-stabilizing vegetation, and a high load of coarse sediment promote the development of braided channels. Braided channels are commonly found downstream of glaciers and at mountain fronts with high sediment loads and steep channels.

Anastomosing Channels

Anastomosing (or anabranching) channels exhibit a complex pattern of individual channels that bifurcate and rejoin to flow around relatively stable, typically vegetated, islands **[Photograph 6.18]**. Anastomosing channels are generally narrower and deeper, have lower gradients than braided channels, and migrate by discrete avulsions instead of by steady lateral channel migration. Cohesive banks that limit lateral migration, flood-prone discharge regimes, and mechanisms that promote local overbank



PHOTOGRAPH 6.18 Anastomosing Channel. Some of the sandbars in this anastomosing channel in the Alaska Range are stabilized by substantial vegetation growth.

flooding or channel blockage promote formation of anastomosing channels. In forested environments, the presence of large organic debris capable of forming stable logjams that locally split flow into multiple channels can result in an anastomosing channel pattern. Similarly, blockage of individual channels by logjams can trigger avulsions that shift flow from one channel to another or that cause flow to spill overbank and form a new channel.

Channel-Reach Morphology

A **channel reach** is a stretch of a channel that exhibits similar characteristics; reach types often reflect similarities in bed and bank material and position in the landscape. The distribution of typical channel reach types within a stream system reflects differences in relative transport capacity, as determined by the ratio of sediment supply to a stream's transport capacity. At a fine scale, within individual channel reaches, the channel consists of groups of morphologically distinct forms, called **channel units,** which are typically up to several channel widths in length. Unique suites of channel units—bars, steps, pools, and riffles—define different types of channel reaches [**Figure 6.10**].

Colluvial Reaches

Colluvial reaches typically occur in stream valley segments in the headwater portions of channel networks. Hillslope processes of mass wasting, soil creep, tree-throw, and burrowing activity introduce sediment into upland channel reaches and shape these channels. Flow in the channel does not govern the formation of the valley fill because shallow flow and limited fluvial transport capacity are insufficient to alter significantly the patterns of gravity-driven deposition, especially because of the stabilizing role of large woody debris.

Bedload sediment in colluvial stream reaches is typically poorly sorted and includes finer grain sizes than downstream alluvial channels. Average bed-surface grain size generally increases downstream as flow begins to winnow the finest grains; most colluvial reaches exhibit downstream coarsening. Average grain size typically reaches a maximum at or near the downslope transition from colluvial to alluvial reaches. In steep colluvial channels, debris flows that deposit clasts too large for normal flows to move are a dominant sediment transport process, and coarse-bed channels are common. The downstream terminus of a substantial debris flow often coincides with a grain-size maximum in a mountain stream profile because of the role that debris flows play in the delivery of coarse clasts. Because water flows in colluvial channels are insufficient to move large rocks, episodic debris flows are the primary means by which steep headwater channels are cleared of accumulated large debris.

Bedrock Reaches

Bedrock channel reaches are cut mostly into rock. They have little, if any, alluvial bed material or valley fill, generally lack floodplains, and are typically confined by narrow valley walls. Bedrock reaches typically occur on steeper slopes than alluvial reaches within the same drainage basin. In general, bedrock reaches lack an alluvial bed because the stream's transport capacity is greater than its sediment supply, a discrepancy that results from a high slope, and thus high transport capacity, a low sediment supply, or a combination of both.

Channels that are subject to scouring and/or deposition by periodic debris flows may alternate between bedrock, colluvial, and alluvial morphologies as they recover following disturbance; for example, a channel scoured to bedrock by a debris flow may slowly accumulate sediment delivered by creep down steep adjacent slopes. Channels in mountain drainage basins often exhibit mixed alluvial-bedrock reach morphologies that arise from fluctuations in local controls on sediment delivery, accumulation, and storage. For example, softer, more erodible rocks wear down faster and create wider valleys with more local **accommodation space** for storage of alluvium in the valley bottom than do more erosion-resistant rock types.

Alluvial Reaches

Alluvial channel reaches have morphologies that are predominantly formed by the interaction of flowing water and the sediment it carries. Several distinct channel reach morphologies can be identified in a natural continuum of alluvial channel types that reflect variations in relative transport capacity.

Cascade reaches are characterized by longitudinally and laterally disorganized bed material, and their bed typically consists of cobbles and boulders. Flow in cascade reaches diverges and converges around individual large clasts that protrude into the flow, generating large vortices and waves that dissipate a substantial portion of the total

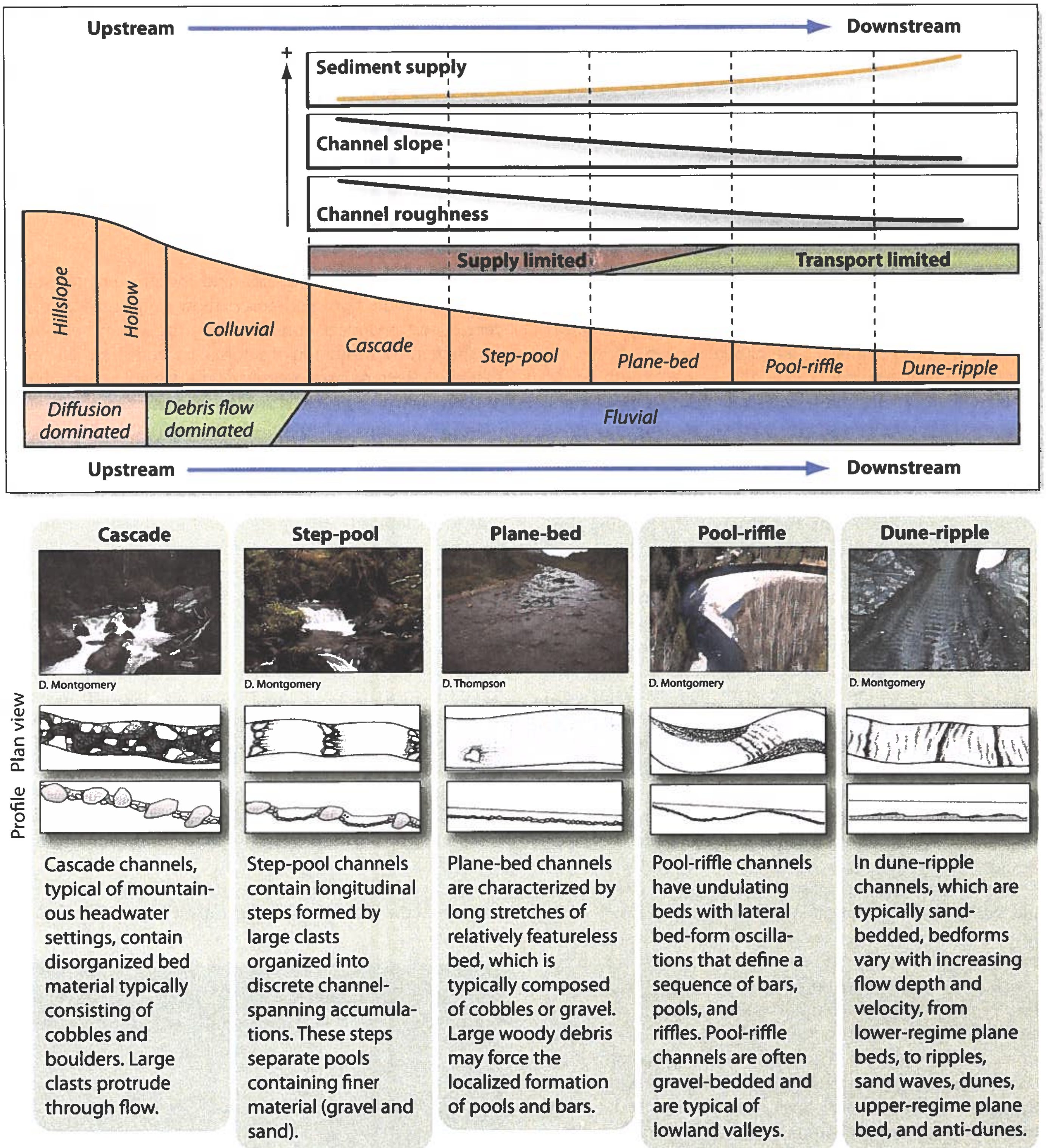


FIGURE 6.10 Downstream Trends in Mountain Stream Channel Types and Characteristics. Schematic illustration of typical downstream trends in sediment supply, channel slope, channel roughness, and channel types on a longitudinal profile from ridge crest to basin outlet. The particular sequence of channel types along any given river system will reflect both local and systematic downstream trends in channel slope.

flow energy; flow tumbles over obstructions. The largest clasts may move only every few decades; smaller clasts are rapidly transported during more frequent flows. Cascade reaches are steep (between 8 and 20 degrees), have large bed-material grain sizes, and have relatively shallow flow depths. Sediment transport in cascade channels tends to be supply limited because the transport capacity generally exceeds the sediment supply.

Step-pool reaches are characterized by a series of discrete channel-spanning accumulations of large clasts that form coarse-grained steps between pools floored by finer-grained sediment. The steps account for much of the elevation drop within step-pool reaches, and the steps provide bed roughness, causing areas of supercritical flow that alternate downstream with tranquil flow through pools. Steps nucleate where large clasts accumulate in congested zones with high local flow resistance, and they grow by trapping additional large clasts. Infrequent flood events that move large clasts typically form the framework of step-pool reaches; finer sediment is transported over the steps and deposited in pools during more frequent, lower-velocity flows. Step-pool reaches typically form at channel slopes between 4 and 8 degrees. Like cascade reaches, they are generally supply limited.

Plane-bed reaches have relatively featureless channel beds that are mainly composed of riffles. Although they share the same name, they are not to be confused with the upper and lower regime plane-bed morphology in sand-bed channels. Plane-bed reaches are distinguished from steeper cascade reaches by the absence of tumbling flow and by smaller grain sizes relative to flow depth. The largest clasts are submerged at all but the lowest flow in plane-bed channels. The few pools in plane-bed channels typically form in eddies where flow has been forced around obstructions. Plane-bed channels are typically straight, have slopes of 1 to 4 degrees, and are floored by a coarse surface layer that becomes mobile at or near bankfull flow. Plane-bed channels represent a transitional morphology between steeper, supply-limited channels and lower-gradient, transport-limited channels. Consequently, they are often found between step-pool reaches and lower-gradient meandering channels. Plane-bed reaches are rare in forested mountain drainage basins because stable obstructions, such as large woody debris, force local pool and bar formation.

Pool-riffle reaches consist of a sequence of bars, pools, and riffles and are typical of meandering channels. The pools and riffles themselves are often relatively stationary features, even though the bed-forming material that composes them is under constant flux. Development of alluvial bars requires a large channel width-to-depth ratio and small grains (relative to flow depths) that are readily mobilized by the flow. Pool-riffle reaches typically have gravel-sized to cobble-sized bedload, slopes less than 1.5 degrees, a coarse surface layer, and general bed-surface mobility at flows approaching bankfull. Pool-riffle reaches tend to be transport limited. Fining of the bed surface and changes in bedform size or amplitude adjust transport capacity to the sediment supply.

Sand-bedded channel segments may have a pool-riffle morphology but are referred to as **dune-ripple** reaches where they exhibit a succession of mobile bedforms that provide the primary flow resistance. Bedform type is determined by flow depth and velocity as expressed by the Froude number (eq. 6.3). Sediment transport in sand-bedded reaches occurs at almost all discharges, and transport rates are strongly dependent on discharge. Sand-bedded reaches are thus transport-limited channels.

Braided channels fit into the lower gradient parts of the classification scheme. The individual threads of most braided channels have pool-riffle morphology, although some consist of individual channel threads that are plane-bed and/or dune-ripple.

Large Organic Debris

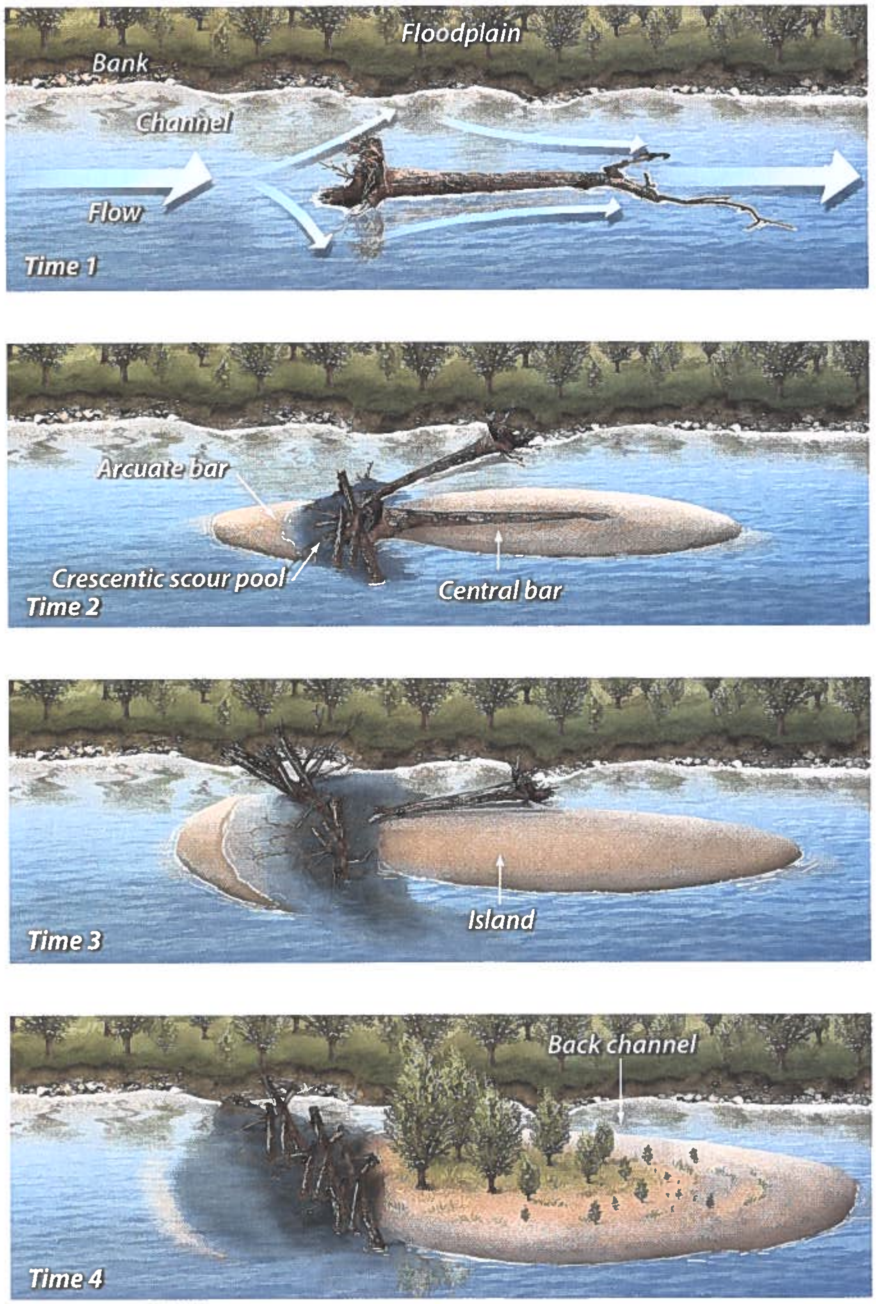
Large organic debris, like logs and logjams, creates stable obstructions to flow and forces flow convergence, divergence, and sediment impoundment in stream channels, resulting in the formation of pools, bars, and steps. The morphologic effects of large organic debris depend on its volume, position, orientation, and size relative to channel dimensions. Large logs in small channels can be stable for decades, even centuries. These obstructions force local, long-term flow convergence that scours out pools and flow divergence where sediment accumulates to form bars. In larger channels, individual logs are more mobile and less likely to obstruct flow, but groups of logs and other organic debris form stable logjams that influence pool formation, channel patterns, and sediment storage [**Photograph 6.19; Figure 6.11**].

Stable logjams obstruct and divert flow and can split flow into multiple channels, forming anastomosing channel patterns. In forest channels, large organic debris can force the formation of pool-riffle or step-pool morphologies by retaining sediment in reaches that would otherwise have bedrock or plane-bed morphologies. In such reaches, the abrupt failure or removal of a logjam, or a gradual reduction (due to deforestation) in the supply of wood large enough to form jams, can lead to changes in channel morphology, allowing large volumes of sediment to be scoured and flushed downstream.

The type of trees that supply organic debris to a stream determines the effect of plant material on stream morphology.

D. Montgomery

PHOTOGRAPH 6.19 Logjam. Logjam on the cutbank of a meander on the Queets River, Washington.



Deposition of a large, stable log (shown here with an attached rootwad) causes local flow **convergence** and **divergence** that result in bed scour (erosion) upstream and sediment deposition downstream. Faster flow around the rootwad causes scour.

Scour around a stable, **key member** log creates a crescentic **scour pool** on the upstream side of the obstruction and deposition builds a **central bar** that buries the tree trunk on the downstream side. An arcuate bar forms as flow diverges upstream of the obstruction.

Continued scour and deposition occurs as additional debris racks up on the logjam, enlarging the scour pool. Continued deposition can build up the central bar into an island, and flow deflection can result in localized channel widening due to bank scour.

Eventually the logjam can become partially buried. It then protects the associated island from erosion, providing stable habitat where trees large enough to produce key member logs can grow even in disturbance-prone valley bottoms. The bar can eventually attach to the channel bank and become integrated into the floodplain.

FIGURE 6.11 Effects of Large Woody Debris. Large woody debris in channels changes the distribution of water flow and velocity. The resulting sediment scour and deposition can eventually form a new patch of floodplain. [Adapted from Abbe and Montgomery (1996).]

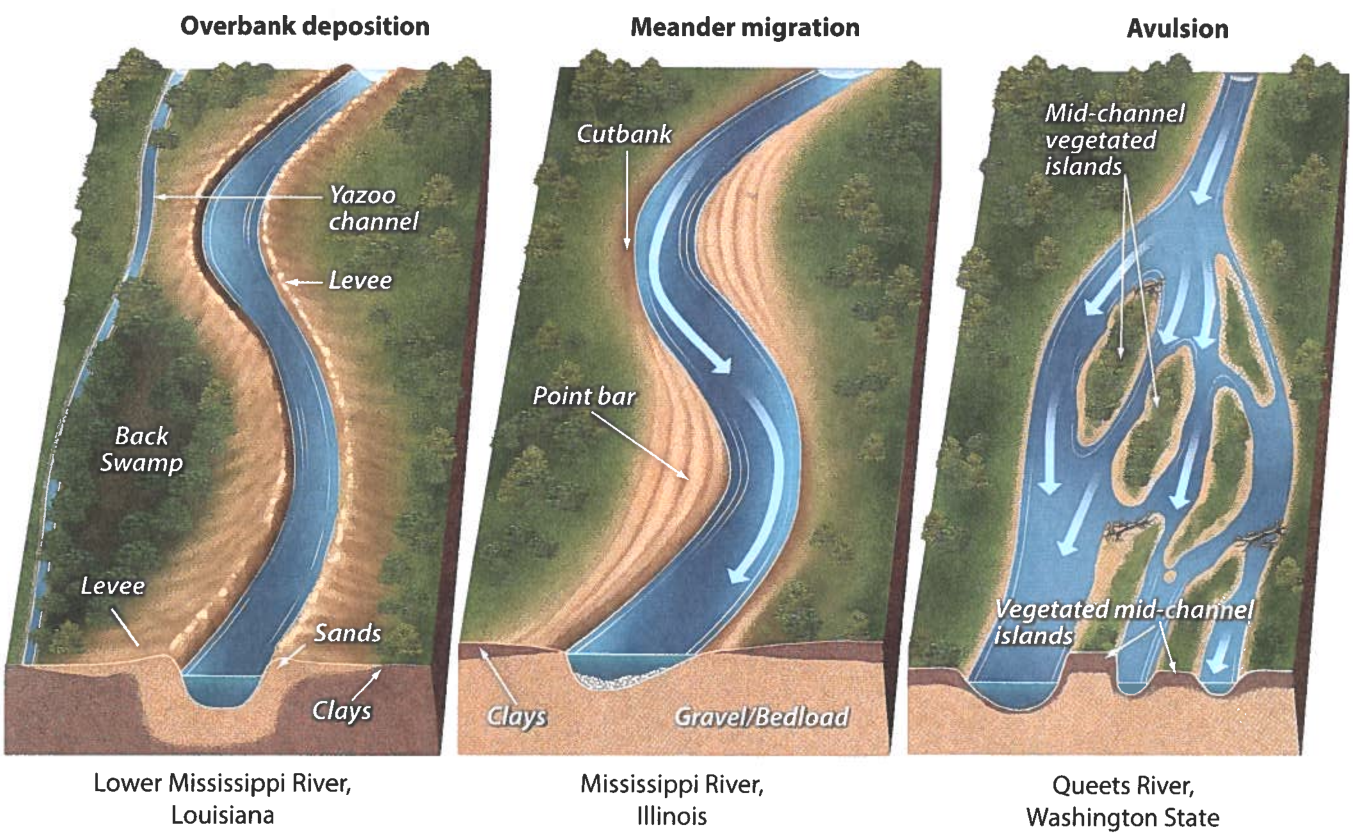
The highly branched trees typical of temperate deciduous forests and tropical rainforests form individual snags. Such snags plagued navigation on the Mississippi River historically and still interfere with boat traffic on tributaries of the Amazon River today. In contrast, the more readily transported, telephone pole-like morphology of coniferous trees leads to extensive development of logjams, where material routed downriver hangs up on large key members, as is common in the Pacific Northwest and in other evergreen forests around the world.

Floodplains

Floodplains, the low ground adjacent to stream channels, are built up by sediment deposited during floods. Alluvial channels typically have laterally extensive floodplains. Bedrock channels generally do not have floodplains, which makes defining bankfull flow difficult in many bedrock channels because they often have no discernible banktop. Floodplains are built by lateral and vertical accretion of sediment that results from (1) deposition of suspended load that settles out from overbank flow, (2) bedload deposition from lateral channel migration, and (3) amalgamation of local surfaces formed by alluvium trapped by blockages like landslides, channel-choking plant growth, and logjams [**Figure 6.12**]. These three processes lead to formation of different types of floodplains, composed of different materials and with different suites of floodplain landforms.

When sediment-laden discharge spills out over channel banks during floods, the velocity of the unconfined flow decreases due to decreased depth and increased roughness. This causes material carried by the flow to settle out onto the floodplain. Floodplains built up by the accumulation of suspended sediment are composed of landforms derived from differential settling of material leaving the river. As flow slows upon spilling out of the channel, the coarsest material settles out close to the channel, building berms parallel to the channel banks known as **natural levees** [**Photograph 6.20**]—sandy deposits decimeters to meters high along channel margins. These levees are breached by **crevasses**, which allow water and sediment out to create **crevasse splays**, fanlike deposits of sediment on the floodplain just outside the crevasse.

Farther from the channel, floodplains built primarily of overbank deposition are composed of fine-grained sediment, typically silts and clays that were carried as suspended

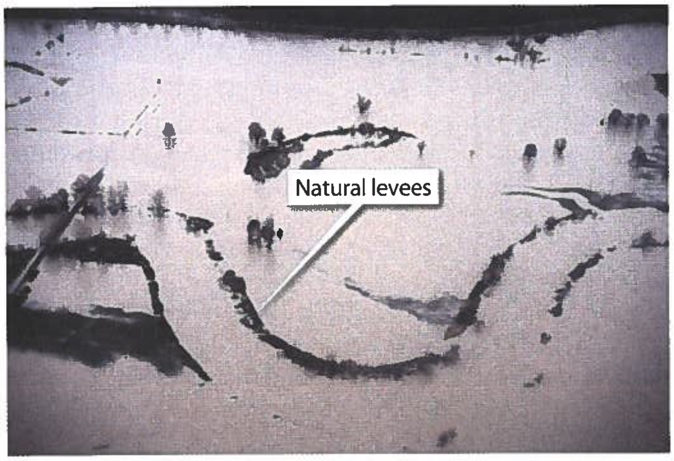


Floodplains dominated by **overbank deposition** develop **natural levees** composed of relatively coarse material (generally sand) with finer material (silt and clay) dominating deposition farther from the main channel. The floodplain surface is typically lower than the elevation of the natural levees, leading to the development of back swamps and **yazoo channels.**

Floodplains formed by lateral channel migration are built as bedload material is deposited on the inside of meander bends. Subsequent deposition of overbank sediments builds the point bars up to the level of the floodplain surface, resulting in a cap of fine-grained material above a floodplain generally composed of bedload material.

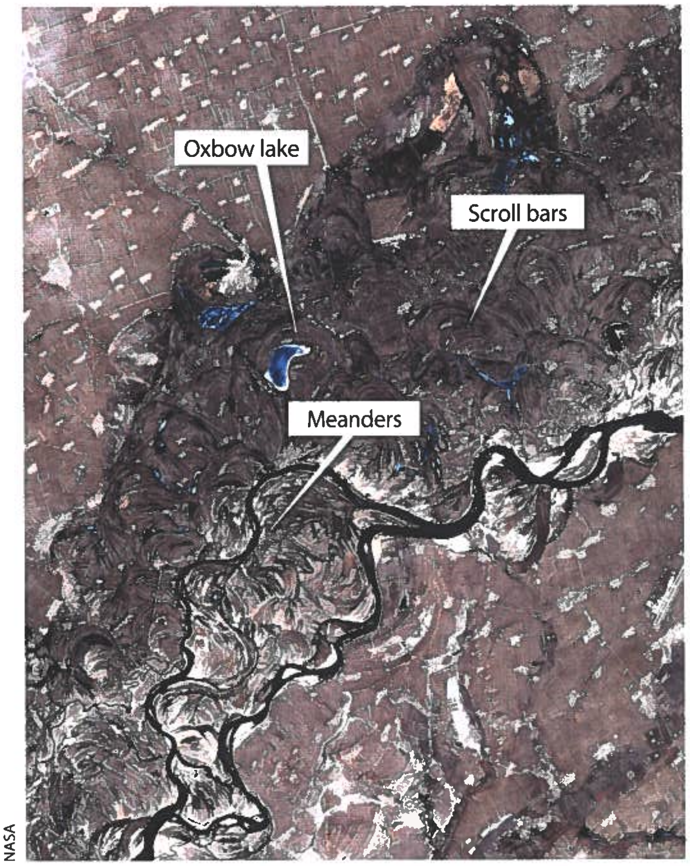
Floodplains formed by integration of deposits formed around individual stable log jams are composed of a patchwork of surfaces, some of which may stand above the general floodplain by as much as several meters due to the effect of local sediment impoundments behind individual logjams.

FIGURE 6.12 Floodplain Development. Floodplain formation processes and landforms differ depending on active channel processes. Floodplains can be built by overbank deposition, channel migration, and avulsion.



D. Montgomery

PHOTOGRAPH 6.20 **Natural Levees.** Natural levees with trees growing on their crests define the channel of the Snohomish River in Washington during a 1996 flood.



NASA

PHOTOGRAPH 6.21 **Scroll Bars and Meanders.** Satellite photograph of the Songhua River, just west of Haerbin, northeast China, showing meanders, scroll bars, and oxbow lakes. The area shown is about 30 km wide.

load. Extensive flooding can occur when rivers overtop their levees because the surrounding floodplain typically sits at a lower elevation; **backswamps** commonly occupy low-lying ground on valley margins along suspended sediment-dominated floodplains. **Yazoo channels** are tributary streams that flow down floodplains parallel to and outside the main channel levees and that serve to drain floodwaters back into the main channel at some distance downstream.

Lateral channel migration by cutbank erosion on the outer side of meanders and point-bar deposition on the inner bank builds floodplains composed of relatively coarse-grained bedload material at depth (old point bars and channel lags) overlain by fine-grained overbank deposits. **Scroll bars** are laterally stacked, abandoned point-bar deposits associated with former positions of the inner margin of channel meanders **[Photograph 6.21]**. They record past positions of the stream channel and can be used to track the history of its trajectory across a floodplain. Channel avulsions leave abandoned meanders that sometimes form **oxbow lakes,** narrow looping lakes that may remain partially connected to the main channel. Oxbow lakes slowly fill with fine-grained material deposited from suspension during subsequent floods. Once filled with relatively strong, silt- and clay-rich cohesive sediment, old oxbows become erosion-resistant masses that can obstruct lateral migration of meanders and modify groundwater flowpaths.

In forested terrain, floodplains are topographically varied and hydraulically rough. Logjams cause local channel damming and avulsions, creating networks of side channels. Local accumulations of sediment are trapped behind stable logjams and other obstacles to flow. These can coalesce into a patchwork floodplain composed of deposits at elevations determined by the depositional contexts of individual logjams. Unlike floodplain surfaces that have slopes similar to the valley slope, in forested floodplains, individual patches of bedload trapped by logjams can form terracelike surfaces at elevations up to several meters above the riverbank. These flat surfaces rise up to twice the diameter of the key member logs and are discontinuous both laterally and longitudinally.

Channel Response

Concern over impacts on human communities that rely on rivers and streams, as well as on aquatic and riparian ecosystems, motivates a desire to understand the ways that stream channel systems respond to disturbances, including both natural events and those that result from land use or climate change. The wide variety of channel types, the complex ways that channels adjust to local and regional factors, and the potential time lags between perturbation and channel response complicate interpretation and make prediction difficult.

Alluvial channel morphology adjusts to variations in sediment supply and discharge. In general, channels respond to fluctuations in (1) the delivery rate, volume, and grain size of supplied sediment; (2) transport capacity as affected by changes in bankfull discharge as well as the frequency, magnitude, and duration of high-discharge events; and (3) vegetation that influences bank stability or the size, amount, and stability of in-channel woody debris. Channels alter their width, depth, bed slope, grain size, and plan view patterns to accommodate changes in the three boundary conditions listed above.

Changes in discharge or sediment supply are often offset by corresponding changes in channel-bed grain size and slope, which can be formalized as a relationship that states that the product of the bedload supply (Q_b) and the bed-surface grain size (d_{50}) is proportional to the product of the water discharge (Q) and channel slope (S):

$$Q_b d_{50} \approx QS \qquad \text{eq. 6.8}$$

The sediment load (Q_b) and discharge regime (Q) are imposed on the channel by upstream conditions. The channel is free to adjust grain size (d_{50}) by sorting sediment and to adjust channel bed slope (S) by incising or aggrading. Different streams and stream segments within a drainage basin may be out of phase in their responses to the same initial disturbance because of the time lags that arise from routing sediment through the channel network.

Changes in the sediment supply to a channel occur naturally when mass movements (such as landslides, rock falls, and debris flows) suddenly deliver sediment from hillslopes into stream channels. Anthropogenic changes, like those that accompany mining, logging, urban development, or land clearing for agriculture can also dramatically alter the amount and grain size of sediment supplied to stream channels (see Figure 7.12). Channels with a high sediment load tend to have finer beds lacking a distinct, coarse surface layer because flows are insufficient to transport and sort all the material delivered to the channel. Conversely, channels with low sediment loads generally have coarser beds and a well-developed coarse surface layer because there is enough energy to flush fine material downstream.

Aggradation occurs when an increased sediment supply overwhelms stream transport capacity and the channel bed builds up. Rapid addition of a large amount of sediment can result in significant channel infilling and loss of flow conveyance, which increases the frequency of overbank flood flows. Channel responses to an abrupt increase in sediment supply may also include fining of the channel bedload, pool infilling, channel widening, and development of channel braiding and narrow braid bars.

Pulses of sediment introduced into channel systems may move down through channel networks as coherent slugs, causing a wave of progressive aggradation and re-incision as the sediment moves downstream. Patterns of sediment storage that are out of equilibrium with the stream system as a whole can persist for decades within a channel network as streamflow gradually mobilizes and redistributes sediment from the reaches where it accumulated. The impact of increased sediment loads can be long-lasting. Channels in parts of the Sierra Nevada are still adjusting to large inputs of mining debris that occurred during California's Gold Rush in the mid-1800s.

Mass movements can introduce material into the channel that is too large for moving water to transport. Rock falls, valley-wall rock slides, and debris flows all convey large volumes of sediment into headwater channels in mountain drainage basins. Like dams engineered by humans, large landslides block rivers and impound lakes, but these natural dams typically fail and erode away over time. As rivers erode down through landslide debris, clasts that are too large to transport sometimes become concentrated and form an immobile lag on the channel bed, retarding river incision.

The effects of debris flows vary with slope and position along the channel network. Although debris flows can scour steep headwater channels to bedrock, they generally deposit material when they reach lower slopes of 3 to 6 degrees or when they lose substantial momentum traversing sharp bends at tributary junctions. Debris flow deposition results in local aggradation. Sometimes debris-flow deposits completely fill in and obliterate a channel. Recovery following debris-flow disturbance differs between steep and low-gradient channels. Steep, high-energy channels (bedrock, cascade, or step-pool reaches) recover relatively quickly and rapidly transmit most debris-flow material to downstream reaches. In contrast, low-gradient channels (pool-riffle and plane-bed reaches) generally take longer to recover from debris-flow impacts because of their lower transport capacity. The cycle of debris flow disturbance and recovery can take decades to centuries and varies with position in the channel network.

Applications

Human actions greatly affect fluvial systems. In addition to direct impacts like channelization by levees and impoundment by dams, human modifications to the land surrounding a stream system (e.g., deforestation, agricultural development, and urbanization) profoundly affect stream systems by changing discharge, sediment supply, and the caliber and amount of woody debris supplied to the stream.

Construction and paving of land in urban and suburban areas increases the area of impervious surfaces, which alters the volume and timing of runoff delivered to stream channels, often triggering channel adjustment. Increased sediment delivery to channels from agriculture and forestry practices, as typically accompanies plowing and clear-cutting, alters sediment loads and causes streams to adjust. Channel responses associated with reduced sediment supply include both changes in the grain-size distribution and sorting of the channel-bed sediments and channel incision, or **entrenchment**, as occurs when people mine substantial amounts of gravel from a channel or when denuded catchments are revegetated.

Alteration of channel-margin vegetation changes the size and species of wood entering a channel, thereby influencing the abundance of pools, bars, and steps generated by accumulation of woody debris. The conversion of channel-margin vegetation from forest to grassland species (e.g., when forest is cleared for pasture land) can lead to systemic channel widening or narrowing, depending on the relative amount of bank reinforcement by roots. In general, however, in-channel logs and logjams recruited from streamside forests tend to promote variability in channel width by locally deflecting flow toward channel banks and by trapping sediment. In forest channels, depletion or reduced size of large organic debris causes pools to be lost, alluvial reaches to erode down to bedrock, and anastomosing channel patterns to simplify and become braided, straight, or meandering. In small channels, where logs and logjams provide significant sediment storage, a decreased supply of large woody debris accelerates sediment transport. Channels in which large wood provides a dominant control on pool formation and sediment storage, like those with a pool-riffle or step-pool morphology forced by logjams, are particularly responsive to changes in the size and amount of woody debris.

Dam construction and large natural stream obstructions like landslide deposits, logjams, lava flows, and glaciers, change both the discharge regime and the sediment supply to channels downstream. An abrupt reduction in sediment supply, such as occurs when a dam is constructed across a stream, typically results in channel incision and coarsening of bedload sediments downstream because the sediment formerly delivered from upstream is impounded behind the dam. Decreased discharge or frequency of high-flow events below a dam cause channel narrowing in downstream reaches. Conversely, dam removal causes channel widening and delivers a pulse of sediment that moves through downstream reaches as the stream redistributes the material that accumulated in the reservoir.

The construction of artificial levees along a river can exacerbate downstream flooding by preventing the spread of floodwaters across the floodplain and speeding flow to downstream reaches. When we look back at the history of many stream systems, it is common to see that once levees were constructed in one part of a basin, levees or other flood-control structures were soon required on stream reaches throughout the basin. In locations where channel gradients decrease downslope, levee construction also results in deposition of material inside the channel because flow is confined within the levees. Such confinement prevents the stream from carrying sediment over the banks and onto the floodplain. This leads to aggradation of the channel bed and deflation of the floodplain surface as loose sediments compact under their own weight. Ongoing levee reinforcement over the past several thousand years along China's Yellow River (Huang He) has elevated the channel more than 30 meters above its floodplain. Today, this ensures a catastrophic disaster every time a levee fails.

Structures in streams, including dams, levees, and other flood-control and flow-modification structures, affect river behavior. The direct effects of such changes are easy to appreciate. Levees prevent channel migration and dams store floodwater and trap sediment. The indirect geomorphic effects of changes to riparian and in-stream vegetation can be more difficult to identify but are no less profound. Consequences include changes in stream gradient, redistribution and destruction of aquatic habitats, and variations in the flux of sediment and water through the stream corridor over time and space.

Understanding how such changes influence particular river systems is fundamental to assessing and mitigating flood hazards, as well as to understanding human impacts on aquatic resources and to designing or evaluating river restoration and rehabilitation projects.

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DIGGING DEEPER What Controls Rates of Bedrock River Incision?

Upland bedrock landscapes are dramatic. Their form, dynamics, and evolution are greatly influenced by the processes and pace of bedrock channel incision. Early workers hypothesized that the rate at which rivers incised into bedrock was a function of rock resistance to erosion, river discharge, and slope (Gilbert, 1877). Indeed, discharge can be a critical variable in the rate of bedrock erosion. Although bedrock channels are typically the result of slow erosion over millennia, megafloods, caused by the release of large amounts of ponded water over short periods, can carve canyons rapidly. When a spillway on the Guadalupe River in Texas was first used to convey floodwaters threatening to overwhelm a dam, the flow cut a 7-meter-deep, several-kilometer-long canyon into limestone in just 3 days (Lamb and Fonstad, 2010). Studying streams in this same area of Texas, where intermittent high flows interrupt typically semi-arid conditions, Baker (1977) concluded that much, perhaps most, geomorphic change including bedrock incision was occasioned by rare, storm-driven, high flows.

Geomorphologists understand the effects of lithology on rock resistance to erosion using a variety of approaches. Geological constraints on rates of long-term channel incision provide context while field observations of bedrock channels are useful for characterizing erosion processes. Field experiments that monitor channel bed changes can measure short-term erosion rates, and computer modeling is useful for predicting future evolution of channels and the landscape of which they are part.

Formal expressions used to predict the rate at which mountain rivers incise into bedrock include terms not only for the discharge, flow depth, and channel slope, but also include the supply of sediment to the channel (Whipple and Tucker, 1999). Such sediment is key because it provides tools for erosion when it is mobile but, if thick enough, can shield the bed from erosion.

Stock and Montgomery (1999) used paleoriver profiles of known age to evaluate rates of bedrock river incision. They found that bedrock erodibility varies by more than five orders of magnitude among different lithologies. Not surprisingly, hard crystalline rocks erode more slowly than softer volcaniclastic rocks, and poorly indurated mudstones erode fastest among the five rock types studied. The wide range of bedrock erodibility between different lithologies implies that rates of river incision into bedrock could vary greatly in different geological settings.

Whipple et al. (2000) studied field evidence from a wide range of environments and concluded that the distribution of fluvial erosion processes (plucking, abrasion, and solution) is strongly influenced by bedrock lithology. Although lithology influences the pace of solution (more soluble rocks will erode faster), field evidence led Whipple et al. to conclude that the spacing and orientation of joints, fractures, and bedding planes exert direct control on whether plucking or abrasion dominates bedrock incision. Rocks that were well jointed or fractured on a submeter scale were most often eroded by plucking [**Figure DD6.1**]. Abrasion by suspended sediment controlled rates of incision into more massive, less fractured rocks. Weathering-limited erosion (such as spalling from disintegration by wetting and drying) tends to control erosion of weak, friable rocks.

Lab experiments can reveal a lot about bedrock erosion. Remember your rock tumbler? Sklar and Dietrich (2001) conducted experiments on the pace of bedrock erosion by loading 6-mm-diameter gravel into a rotating drum with a floor made of different types of rock for each of many experimental runs. They, too, found that the rate of bedrock erosion varied by more than five orders of magnitude (rapid = mudstone, slow = quartzite). Rocks with high tensile strength eroded more slowly than those

DIGGING DEEPER What Controls Rates of Bedrock River Incision? *(continued)*

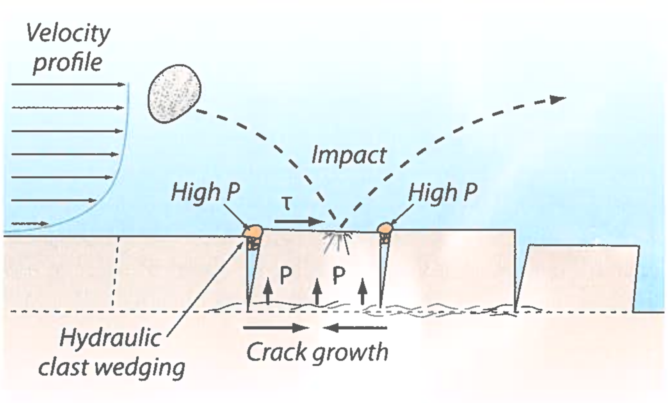

FIGURE DD6.1 Schematic illustration of the processes and forces contributing to bedrock channel erosion by plucking. On the left is the velocity distribution with depth in the channel (arrow length scales relative velocity). Impacts by large saltating grains (shaded) generate stresses that drive the crack propagation necessary to loosen joint blocks. High pressures (P) that develop across crack openings in the bed help drive hydraulic wedging of clasts into fractures that work to further open cracks. Surface drag forces due to shear stress (τ) and differential pressures (high P at the surface versus low P at depth) act to lift loosened blocks from the bed. Where the downstream neighbor of a block has been removed, both rotation and sliding become possible, and it is then much easier for rocks to be plucked out of the riverbed. [From Whipple et al. (2000).]

with low tensile strength [**Figure DD6.2**]. By varying the amount of gravel used in their experiments, they found that the pace of erosion increased initially as more sediment was added. Erosion rates decreased once the sediment covered the majority of the bed; the rate of erosion fell more than an order of magnitude once the bed was fully covered with sediment [**Figure DD6.3**].

The implication was clear. Sediment-covered channels erode slowly because impacting rocks cannot reach the bedrock. Increasing amounts of sediment in transport acted to shield the riverbed from erosion, whereas increasing numbers of sediment clasts (tools) acting as abraders increased the rate of erosion (Sklar and Dietrich, 2006). The combined effect of these two processes makes the pace of bedrock incision greatest when the bed is partially or lightly covered by sediment [**Figure DD6.4**].

Results of landscape evolution models show that bedrock channel incision plays a critical role in the development of mountain landscapes (Whipple and Tucker, 2002). In particular, the rate at which channels erode bedrock communicates changes in uplift rate and base-level lowering through mountain landscapes and governs the timescale of landscape response to such perturbations. The wide range of rates of bedrock channel incision means that the potential pace of landscape response to disturbance also varies widely.

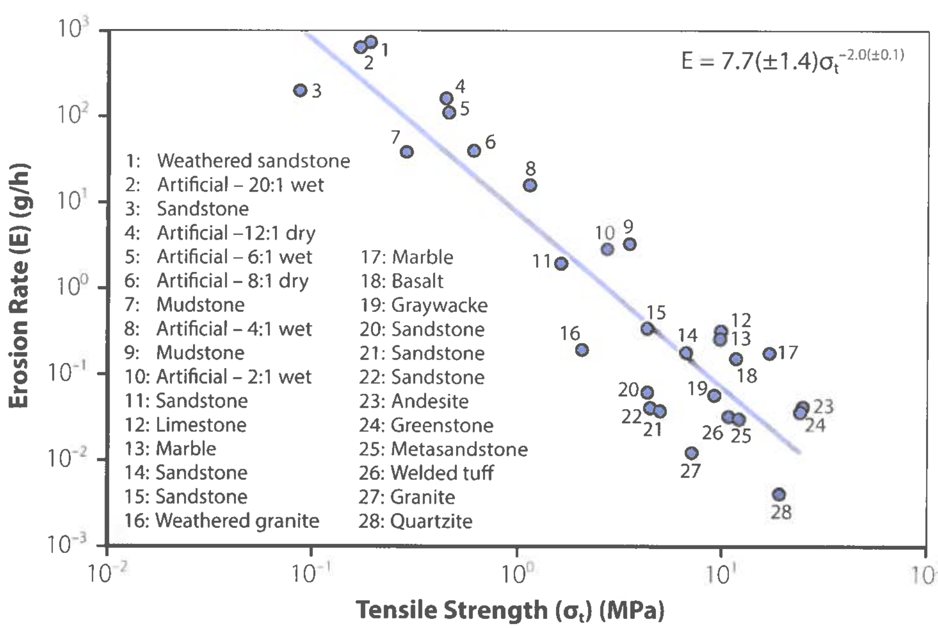

FIGURE DD6.2 Plot of experimental data from the rock-tumbling experiment described in the text. It shows that the variation in measured erosion rate (expressed in grams per hour) decreases with the square of rock tensile strength. Strong rocks (e.g., quartzite, #28) erode more slowly than weak rocks (e.g., sandstone, #3). For this set of runs, 150 g of 6-mm-diameter gravel sediment were loaded into the tumbler. [From Sklar and Dietrich (2001).]

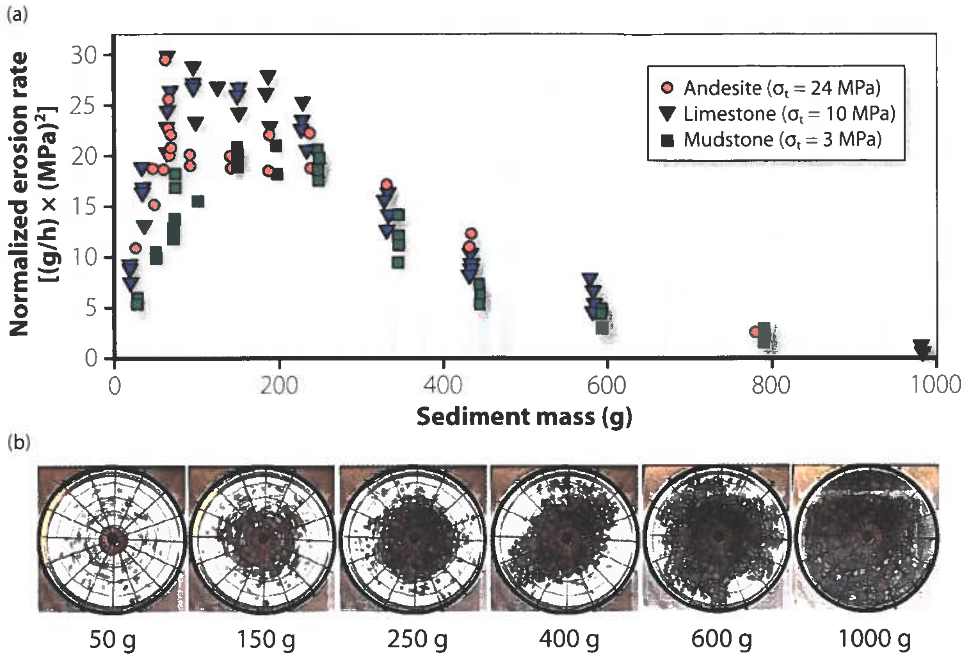

FIGURE DD6.3 (a) The variation in experimental erosion rate (the tumbler, again) with increasing sediment load (mass). Data for three rock types are normalized for rock strength by multiplying erosion rate (expressed in grams per hour) by the square of the rock tensile strength (expressed in megapascals, MPa). (b) A series of photographs taken from below the abrasion mill looking up through the glass bottom. They show how alluvial bed cover increases with increasing sediment loading. [From Sklar and Dietrich (2001).]

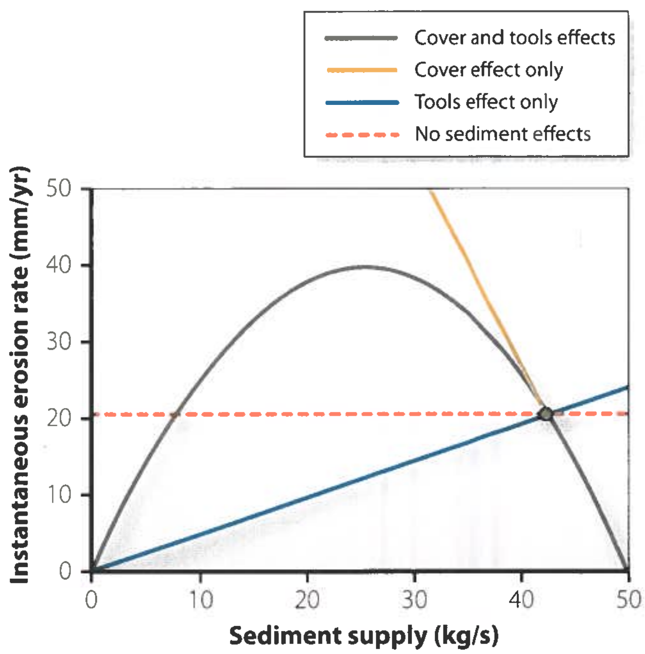

FIGURE DD6.4 Model result for instantaneous bedrock erosion rate as a function of sediment supply. The sloping (yellow) line indicates the effect of bed cover, a reduction in erosion rate with increasing sediment supply. The rising (blue) line indicates the increasing erosion rate caused by more tools abrading the bed. The arched (green) curve represents the combined effect of cover and tools and indicates that the erosion rate peaks when there is some but not too much sediment on the bed. The circle at the intersection of the lines indicates the instantaneous incision rate that corresponds to the assumed long-term bedrock erosion rate at a site on the South Fork Eel River in northern California. The dashed (orange) line represents the case of no sediment effects. [From Sklar and Dietrich (2006).]

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DIGGING DEEPER What Controls Rates of Bedrock River Incision? *(continued)*

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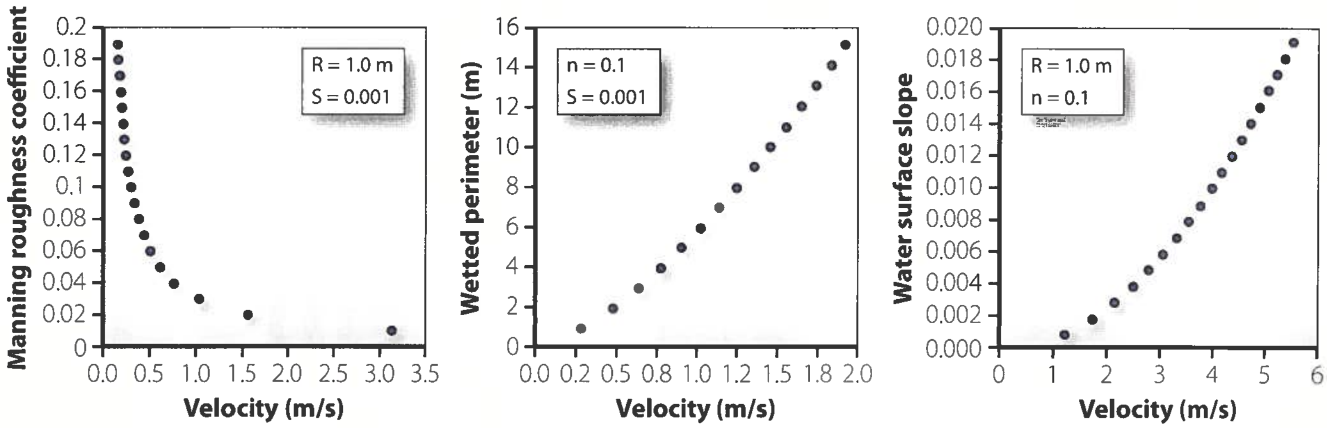
WORKED PROBLEM

Question: Manning's equation (eq. 6.2) describes the relationship between flow velocity and channel hydraulic radius (R), water surface slope (S, or tan θ), and the roughness or energy dissipation characteristics of an open channel (n). Do a sensitivity test so that you can describe how velocity of flow changes if (1) channel roughness increases through the range of Manning n values found in natural channels while holding R and S constant, (2) R (the wetted perimeter) increases while n and S are held constant, and (3) slope increases while R and n are constant. Express your results in words and graphically.

Answer: All three relationships are nonlinear. As n increases, flow velocities decrease. As the wetted perimeter, R, grows larger, flow velocities increase. As the water-surface slope increases, velocities also rise. Note that the difference in the exponents for R and S (2/3 versus 1/2) in Manning's equation control the rate of change in velocity for a unit change in the variables R and S.



Graphs showing the relationship between velocity and changing values of the Manning roughness coefficient, channel wetted perimeter, and water surface slope.

KNOWLEDGE ASSESSMENT Chapter 6

- 1. Explain why understanding channels and their behavior is fundamental to geomorphology.
- 2. Define a graded stream.
- 3. List three factors controlling the shape of channels.
- 4. Describe several ways in which climate can influence channels.
- 5. Contrast the change in discharge downstream in humid-region and arid-region channels.
- 6. List three ways in which sediment reaches river channels.
- 7. Describe how bed and bank material affect the behavior of channels over time.

- 8. Predict where you are most likely to find alluvial channels and where you are most likely to find bedrock channels.
- 9. Explain two very different ways in which riparian vegetation influences channel behavior.
- 10. The balance between what two physical constraints determines flow velocity?
- 11. Write out Manning's equation and explain each of the terms.
- 12. Write the formula for discharge and for the hydraulic radius of a channel.
- 13. Explain what the Manning roughness coefficient represents.
- 14. List several physical properties that determine the roughness of a channel.
- 15. Explain how flow depth affects channel roughness.
- 16. What is the Froude number for streamflow and why is this an important value to know?
- 17. Define effective discharge and explain how its frequency is controlled by climate.
- 18. Explain the velocity distribution along a channel cross section and identify where you would find the highest flow velocity.
- 19. Define bankfull flow and explain why it is important to geomorphologists.
- 20. Sketch graphs of at-a-station hydraulic geometry and explain why they are useful.
- 21. Describe how fluid forces can move and entrain sediment along the bed of a stream.
- 22. List the three types of sediment load carried by a stream. Which type generally represents the greatest amount of transported mass?
- 23. Define critical shear stress and explain why it is relevant to bedload transport.
- 24. Explain the concept of stream power and, in the context of stream power, explain why deeper, quickly flowing water can transport more sediment or incise rock more rapidly than slowly flowing, shallow water.
- 25. Explain why different discharges are required to move sediment than to incise bedrock.
- 26. List the three ways by which rivers incise into rock.
- 27. From the perspective of flow dynamics, explain why alluvial channels migrate across floodplains.
- 28. Contrast the characteristics of pools and riffles at high and low flows.
- 29. Explain why and how channels meander.
- 30. Define river sinuosity.
- 31. Sketch straight, meandering, braided, and anastomosing channel patterns and suggest in what settings each might be found.
- 32. Explain the difference between channel reach types and channel units by providing examples of each.
- 33. Sketch the usual sequence of channel reach types as one moves downstream from mountainous headwaters to lowland river valleys.
- 34. Predict the effects when large woody debris enters a channel by landsliding or bank collapse.
- 35. Describe three different ways in which rivers build floodplains.
- 36. Give three different examples of channel response to external perturbations.
- 37. Predict how alluvial channels will respond to both increases and decreases in sediment supply and discharge.