Drainage Basins

7

Introduction

A **drainage basin** is the entire area drained by a stream and its tributaries. Drainage basins, also known as **watersheds,** are the fundamental unit for geomorphic analysis of fluvial systems because they are the land surfaces over which water and sediment move down topographic gradients. Drainage basins are separated by the elevated land, or ridges, between them that define topographic **drainage divides.** Drainage basins are thus defined by both hydrology and topography **[Figure 7.1]**.

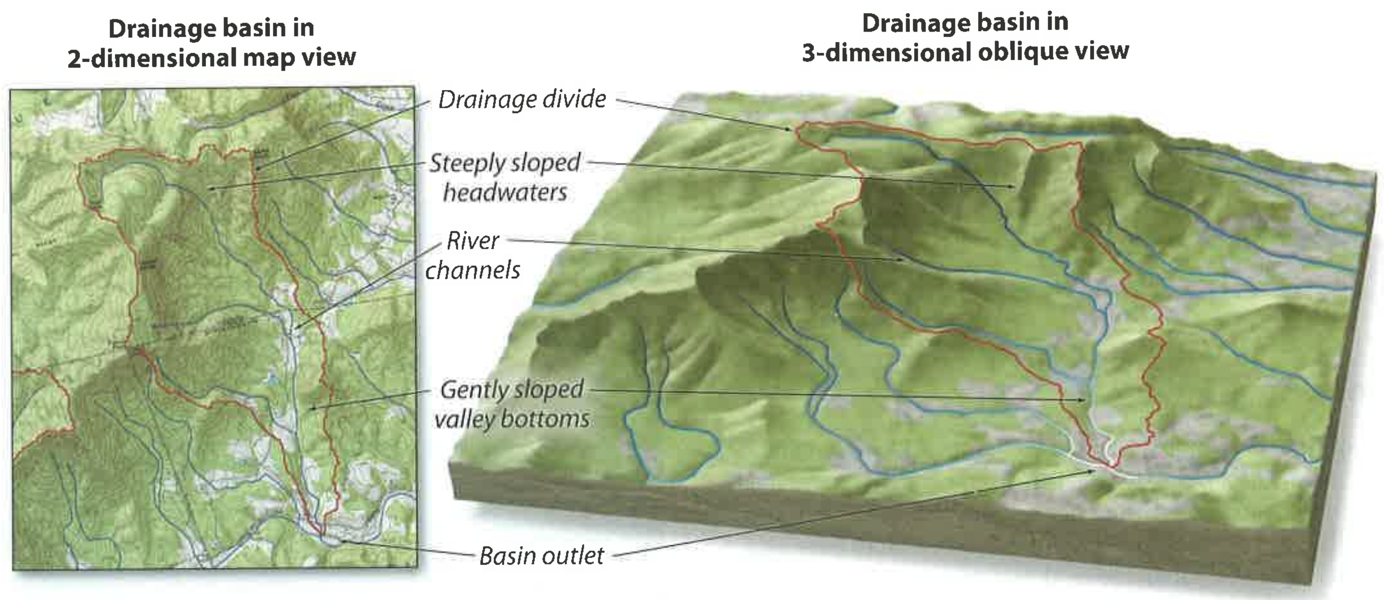
Drainage basins come in all shapes and sizes, from the upper reaches of a small headwater valley to the Amazon River that drains more than one-third of South America. Most humid regions have **perennial streams** and rivers that flow continually. In arid regions, stream channels may stay dry for long periods of time, carrying water perhaps only once a year or even once a decade. These **ephemeral streams** flow only in response to rainfall events. Drainage basins are composed of nested subbasins that geomorphologists analyze as a hierarchical system in which the smallest basins (most of which lie at the head of the channel network) are located at higher elevations along the basin's drainage divide and drain into successively larger watersheds.

Most drainage basins contain a predictable set of components. The upper reaches of the basin generally have steeper slopes and supply sediment to the lower basin. Moving down from the watershed-bounding slopes and ridges, sediment and water flows begin to concentrate in channels until eventually, where slope and/or lateral confinement have diminished sufficiently, there are areas

P. Bierman

Rakaia River, South Island, New Zealand, meanders through spectacular flat-topped alluvial terraces, former floodplains of the river. Along the river's course are well-defined gravel point bars and steep cutbanks. In the distance are the rapidly rising Southern Alps.

IN THIS CHAPTER



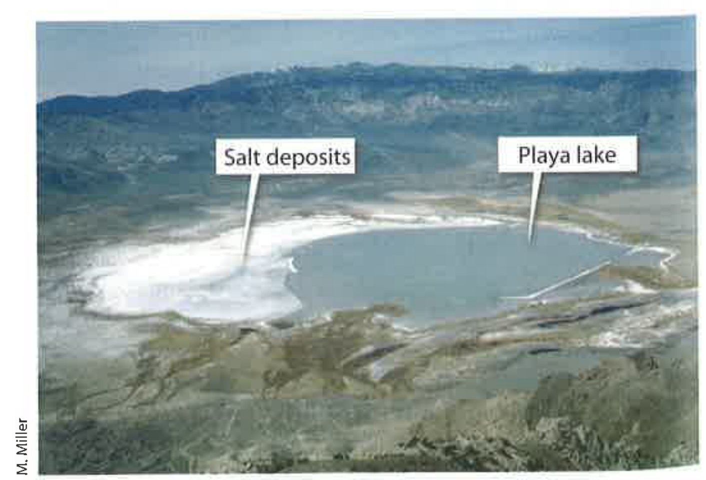
Drainage basins are the upslope area draining to a point along a stream and are a primary way by which geomorphologists subdivide landscapes. Separated by **drainage divides,** rivers and streams in drainage basins convey sediment from generally steep uplands to generally less steep lowlands and then onto an outlet defined as the end of the basin. Drainage basins contain streams of various sizes as well as smaller tributary drainage basins.

FIGURE 7.1 The Drainage Basin. Drainage basins, separated by drainage divides, are a fundamental unit of geomorphic analysis.

where sediment is deposited and stored at least temporarily alongside lowland streams and rivers. In the largest drainage basins, such as the Amazon or the Ganges-Brahmaputra, much of the sediment delivered from the mountains never makes it to the ocean; rather, it is deposited and trapped in isostatically and tectonically subsiding basins where rivers leave upland valleys to flow across broad lowlands (see Chapter 12). In general, steep uplands are zones of erosion, and relatively flat lowlands are zones of sediment deposition. Most sediment that reaches the ocean originates in mountainous headwaters, although it may have spent thousands of years in storage along the way.

Major physiographic, climatic, and geologic features impose broadly comparable controls on channels and their watersheds within a geomorphic (or physiographic) province. Neighboring watersheds in a geomorphic province tend to have similar relief, climate, and rock types. Consequently, broad relationships between drainage area, discharge, sediment supply, and bed-material grain size characterize channels and watersheds in a geomorphic province. In addition, regional similarities in vegetation, climate, and climate history (such as legacies of past glaciation) may impose similar general constraints on channels and thus watersheds.

Drainage basins may be closed or open. Most drainage basins are termed open because mass (water, sediment, and the dissolved load) is transported through and then out of the watershed either to larger watersheds or to the ocean. Drainage basins that are closed (also known as internally drained) terminate in lowlands, where water is lost only by evaporation and seepage into the groundwater system. Closed drainage basins are most commonly found in arid regions where drainage networks are not well developed, in areas where the local tectonic regime is extensional, and in glaciated areas where glacial erosion has overdeepened valley bottoms. Arid lowlands are often occupied by ephemeral lakes that dry out and become salt flats or **playas** (from the Spanish for "beach") between runoff events [**Photograph 7.1**]. These lakes are salty because the



PHOTOGRAPH 7.1 A Salty Playa. The white area is covered by salt left as water evaporates from this closed basin in Deep Springs Valley, California.

dissolved load delivered to them is left behind and concentrated by evaporation. Famous examples of closed drainage basins include those supplying water and sediment to the Great Salt Lake in Utah and to the Dead Sea in the Middle East.

Basin-Scale Processes

At the basin scale, one can investigate the production, transport, storage, and delivery of material. The specific process will vary from place to place; however, there is a generalized framework (sediment budgets, routing, and storage) one can apply to gain a better understanding of landscape behavior.

Sediment Budgets

The fluxes of sediment through a drainage basin can be described in terms of a **sediment budget** that considers **sources** from which mass enters the system (sediment derived from eroding slopes), and **sinks** where mass leaves the system. Sediment budgets allow one to understand where sediment comes from and where it goes. Such budgets are useful for predicting the probability that mass will move into and out of storage, as well as the response of watersheds to disturbances such as the installation or removal of dams that retain sediment. A sediment budget must also account for **storage**, the retention of mass in the system as sediment is deposited in floodplains and other depositional environments. The line between storage and sinks can be a blurry one and often depends on the timescale under consideration. For example, on the timescale of decades, floodplains can be considered as sinks, but over millennia, floodplains are eroded and reworked, functioning as temporary storage for sediment.

A generalized equation for water and sediment budgets considers the flux of mass in (Q_{in}), the flux of mass out (Q_{out}), and the change in storage (ΔST) all considered per unit time:

$$Q_{in} - Q_{out} = \Delta ST \qquad \text{eq. 7.1}$$

Often, the change in storage is assumed to be zero, meaning that $Q_{in} = Q_{out}$ and the basin is in steady state in terms of mass flux.

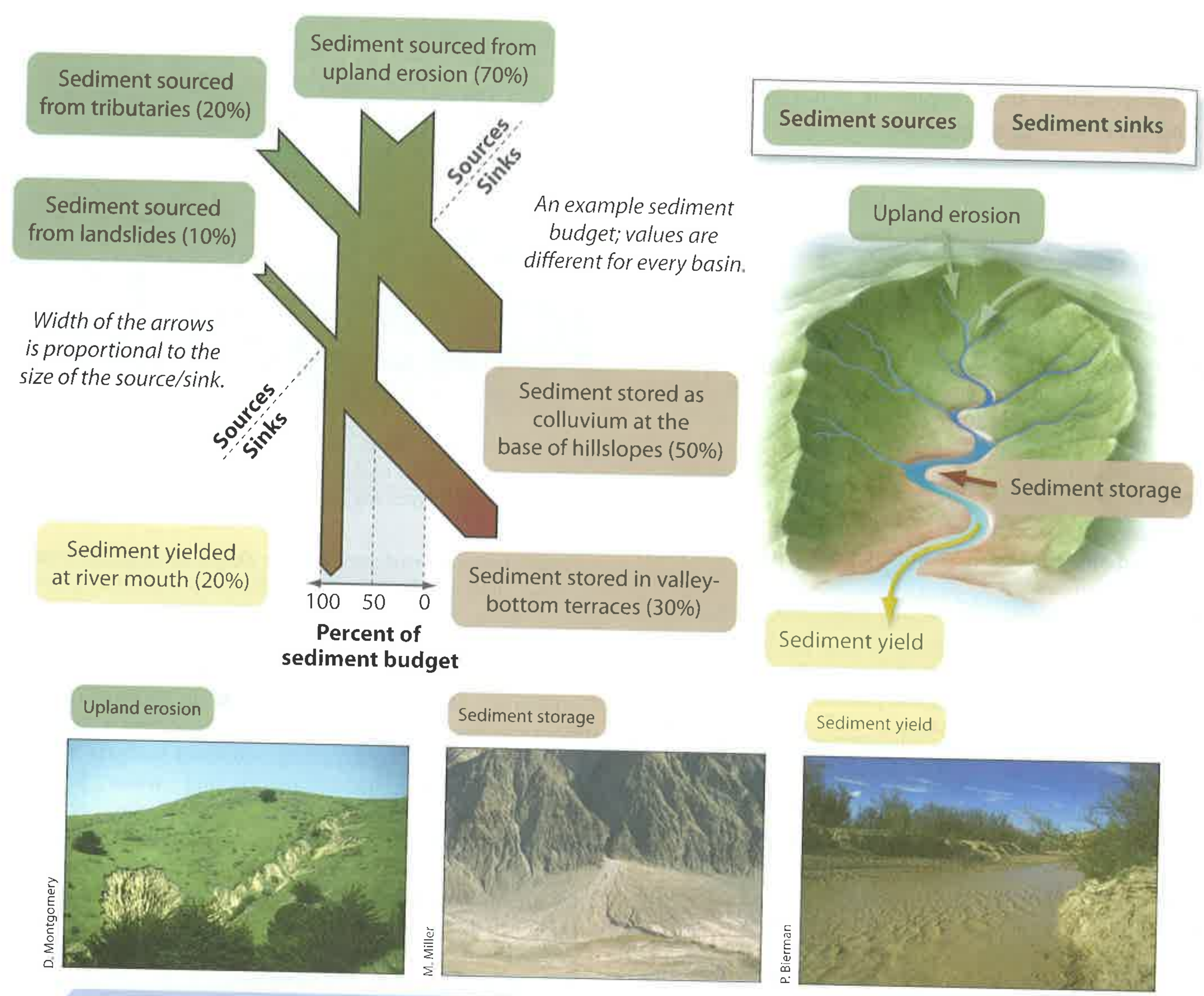
Sediment budgets can be simple or complex. A simple sediment budget might consider only the erosion rate of the sediment-producing hillslopes and the export rate of sediment from the watershed. More complex sediment budgets include terms describing the likelihood of, and thus the flux of material delivered by, dynamic exchanges between sediment storage and active transport—for example, the reworking of floodplain sediment as rivers erode their banks. These kinds of interactions can be assigned probabilities. Sediment stored in higher **terraces** (former floodplains of the river abandoned when the river incised) [**Photograph 7.2**] along valley walls is assigned a

PHOTOGRAPH 7.2 River Terraces. Sequence of river terraces storing large amounts of sediment in a wide, intermontane valley, Kadjerte River, Kyrgyzstan.

lower probability of entering the active channel than sediment stored in lower terraces abutting a river, which are more likely to be reworked by bank erosion. Sediment budgets can be constructed on a variety of different timescales (years, centuries, millennia), and the results can be very different because processes moving and storing sediment can be quite episodic.

Until recently, constructing **sediment budgets** was far more challenging than constructing water budgets, because there were typically few relevant data available [**Figure 7.2**]. The increased use of cosmogenic nuclides (see Chapter 2) to determine basin-scale rates of erosion (sediment supply) over the long term as well as repeat flights of high-resolution LiDAR to estimate short-term rates of surface change have greatly facilitated sediment budget construction. The development of optical sensors for measuring stream-water turbidity, and thus better assessment of suspended sediment load, has improved estimates of mass export from watersheds.

Sediment sources are dominated by the erosion of hillslopes through various processes, all of which can be characterized in terms of rates of mass delivery to the river network. The complication is that many slope processes, such as landslides and gullying, are episodic, and short-term and long-term mass flux rates often differ dramatically. Several different approaches have been used to address temporal variability. It is common to measure the mass of sediment accumulated and stored in depositional landforms in order to estimate average rates of sediment delivery over time and thus constrain rates of sediment generation (and erosion rates) for parts of basins. For example, alluvial fans or colluvial deposits, such as those in hollows, can be dated and the volume of the deposit measured. Defining the area contributing sediment allows calculation of the average rate of erosion and sediment supply from upland slopes. Sediment production rates can also be established through cosmogenic nuclide analysis of soil and sediment and the use of fallout nuclides such as ^{137}Cs (see Chapter 2).

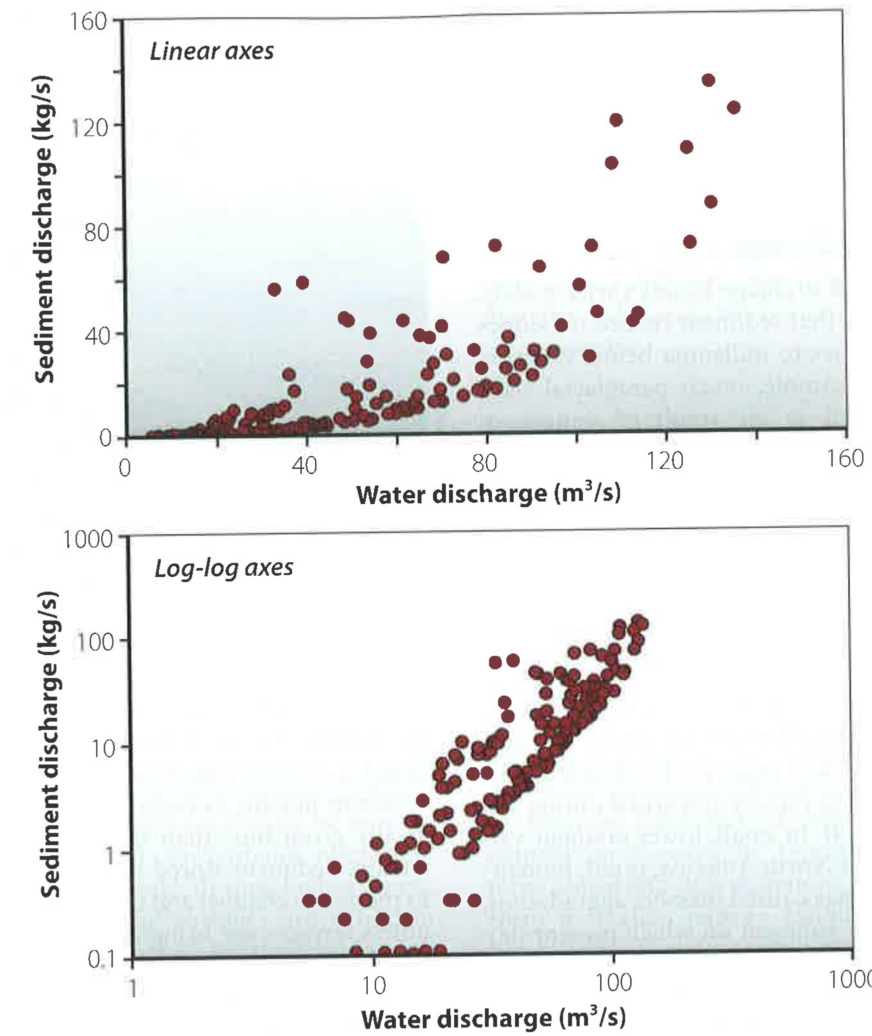


Sediment budgets describe where sediment comes from, where it goes, and how quickly it moves through the landscape. In many basins, sediment is sourced primarily in the uplands where steep slopes erode rapidly. That sediment is often stored for varying amounts of time at the base of slopes as **colluvium** and along stream and river channels as alluvium. Over time, for small basins in equilibrium, sediment generation in the uplands will match sediment yield coming out of the lowlands. However, large amounts of sediment can be stored for decades to millennia in disturbed landscapes. In basins large enough to contain extensive depositional lowlands (like the Amazon), sediment generation rates will exceed sediment yields as some sediment is trapped in the subsiding basin.

FIGURE 7.2 Sediment Budget. Sediment budgets quantify sources and sinks of sediment at the scale of drainage basins. [Adapted from Trimble (1999).]

Sediment evacuation rates are usually established by estimating flow over time and measuring suspended load during a variety of different discharge events, creating what is known as a suspended **sediment rating curve** [Figure 7.3]. This approach accounts for most of the sediment leaving the basin in suspension; however, in some rivers, dissolved load and bedload can be significant and need to be accounted for separately in the total mass budget. Some sediment rating curves display large amounts of scatter due to a phenomenon known as **hysteresis** or path dependence (see Figure 4.11).

Consider what happens to a streambed as flow rises during a storm. Until the flow rises sufficiently to mobilize the bed, sediment transport rates are low because large clasts generally protect smaller material from erosion. Once flow is sufficient to mobilize the bed, then large amounts of sediment move. After the flood peak passes, the flow begins to wane and the large clasts drop out, but the fine material



Sediment rating curves describe sediment discharge as a function of water discharge. Sediment discharge data are noisy—varying greatly over time and from event to event. For example, when 40 m^3/s of water is pouring down this river, it could be carrying between 2 and 60 kg/s of sediment, a 30-fold difference due to variations in sediment delivery. This leads many people to plot sediment data on logarithmic axes that clearly show the systematic, although highly variable, increase in sediment discharge with water discharge.

FIGURE 7.3 Sediment Rating Curves. Sediment rating curves quantify the relationship between the flux of water and the flux of sediment through drainage basins.

continues to move in suspension. Thus, at the same discharge level during flood rise and subsequent recession, very different amounts of sediment may be moving. Measurements of sediment accumulation in lakes and reservoirs can be used to understand sediment flux over a wide range of basin areas and timescales. For example, this approach has been applied to small watersheds feeding farm ponds as well as the entire eastern United States by calculating the volume of sediment deposited offshore over the past 100 million years.

Understanding the magnitude and duration of sediment storage on the landscape is a particularly difficult and time-consuming part of constructing basin-scale sediment budgets. Quantifying storage requires both estimating the volume of sediment stored in depositional landforms (floodplains, terraces, fans, and colluvial deposits) as well as estimating the ages of these landforms and the sediment they contain. Both fieldwork and remote sensing data (LiDAR) are used to identify deposits and quantify their volume. Dating deposits requires techniques introduced in Chapter 2 and often relies on radiocarbon analyses of buried organic material in humid-temperate regions and luminescence or cosmogenic techniques in arid regions (due to the lack of preserved organic matter).

Sediment Routing and Storage

Sediment residence time in drainage basins varies widely. Numerous studies suggest that sediment eroded off slopes can be trapped for centuries to millennia before entering the fluvial system. For example, much **paraglacial** sediment (sediment deposited as the result of nonglacial processes in areas altered by glaciation) still remains on the landscape, and most sediment eroded by postcolonial land clearance and agriculture has yet to enter a stream channel. In both of these cases, the majority of recently eroded sediment remains in colluvial storage at the base of the hillslopes from which it eroded—isolated from the channel network **[Photograph 7.3]**.

Once sediment enters stream channels, its residence time is controlled by grain size and valley morphology. In narrow, steep upland valleys, there are few places for sediment to be stored; there is little space for floodplain deposition, and most sediment is rapidly reworked during the next flood **[Photograph 7.4]**. In small, lower gradient valleys, typical of midwestern North America, rapid, human-induced hillslope erosion has caused massive aggradation, filling narrow valleys with sediment on which present-day streams now flow.

In large, lowland alluvial river valleys, floodplains, once formed, can persist for millennia and longer, isolating at least some sediment from fluvial reworking **[Photograph 7.5]**. For example, in the Amazon River Basin, sediment resides on the floodplain for an average of thousands of years. Floodplain sediment residence times along the Amazon increase downstream, and there is more sediment moving between the channel and the floodplain at any given time than there is sediment moving downstream. Sediment stored in terraces is even less accessible to the active channel and can be stored for many millennia, unless terraces are being actively undercut and eroded by the channel.

For sediment in humid-temperate river channels, transit times are grain-size dependent. The smallest grain sizes, those carried in suspension, generally move through drainage basins at the pace of river flow, except for those grains that settle out on the floodplain as overbank deposits **[Photograph 7.6]**. These grains will eventually be reintroduced to the channel by point bar/cutbank migration. Larger particles move episodically during higher flows. During low-flow periods, coarse particles are stored in the channel bed and on point bars where they reside until the next significant flood **[Photograph 7.7]**. In areas where flows are episodic, such as arid regions and parts of the tropics where precipitation amounts are strongly seasonal, bed material often moves all together, suggesting much less grain-size dependence of residence times.

J. Weller

PHOTOGRAPH 7.4 Lack of Sediment Storage in Narrow Bedrock Valleys. The turbid Salween River stores little sediment in its narrow bedrock valley, Three Rivers Region, Eastern Tibet. The river is running between steep, strength-limited slopes with little soil cover.

M. Miller

PHOTOGRAPH 7.3 Colluvial Storage. Here colluvium is stored at the base of a slope. This small fan-form feature, in the Tien Shan mountains, Kyrgyzstan, is fed by rock falls and debris flows from steep, bare rock slopes above.

P. Bierman

PHOTOGRAPH 7.5 Sediment Storage on Floodplains. In the rapidly eroding North Island of New Zealand, large amounts of sediment are stored along the flat floodplains of the main stem Waipaoa River to the left side of the image and its tributary to the right. The headwaters of the main stem have been heavily logged, triggering erosion and filling the channel and much of the floodplain with sediment.

D. Montgomery

PHOTOGRAPH 7.6 Suspended Silt and Sand. The very turbid Colorado River carries large amounts of silt and sand as suspended load through the Grand Canyon, Arizona. Here, steep hillsides are incised into the Vishnu Schist.

D. Thompson

PHOTOGRAPH 7.7 Gravel Bar. Deposition during a flood created a large gravel bar on the floodplain of the Swift River, New Hampshire. The bar may eventually become vegetated and incorporated into the floodplain.

Channel Networks and Basin Morphology

Channel networks carry water and sediment through drainage basins and are integral both to basin form and function. The form of the drainage network is controlled in part by the earth materials underlying the basin, in part by the tectonic setting of the watershed, and in part by the history of the channel network itself. Over the past century, several different approaches have been developed to describe and classify the pattern and organization of channel networks.

Drainage Patterns

Drainage network development reflects the integrated action of erosional processes as rivers incise over geologic time. The common idea that river courses develop on the blank slate of an initial surface is usually an oversimplification, although this does occur where newly exposed marine terraces that lack channels are first elevated above sea level or on extensive low-relief plains of volcanic flows and ash deposits. More typically, river networks inherit features from preexisting topography and develop on rocks that vary spatially in their resistance to erosion. Tectonic processes can also modify channel network patterns over time.

Channel networks in regions with little structural control on topography typically exhibit **dendritic** patterns consisting of a system of branching tributaries that form a treelike pattern fanning out toward the basin headwaters. There is no dominant orientation of channels in dendritic channel networks, and **junction angles** at which tributaries come together are typically much less than 90°. Dendritic channel networks tend to form in the absence of structural or lithologic controls; thus, they generally occur in relatively flat-lying sediments or in homogenous crystalline rocks. Nondendritic channel networks suggest that geology, topography, or tectonics are influencing network shape.

Qualitative analysis of drainage patterns can reveal much about the materials underlying the basin and about channel history, including structure, rock type, landform shape, and human impact **[Figure 7.4]**. For example, naturally straight channels are rare in nature. Where they are found, they often follow fault lines or joints in rock, exploiting brecciated (broken) zones of weakness or areas of preferential groundwater flow and enhanced chemical weathering. Sharp bends in channels can reflect structural control or they can reflect channels offset by strike-slip faulting (see Figure 2.11).

Geological structures can influence drainage patterns so much that the courses of rivers and streams can be used to interpret or map the underlying structural geology. Such structurally controlled drainage exhibits **trellis** and **rectangular drainage patterns.** Trellis drainage is characterized by two dominant channel orientations in which primary tributaries join main channels at roughly right angles, with secondary tributaries running parallel to main

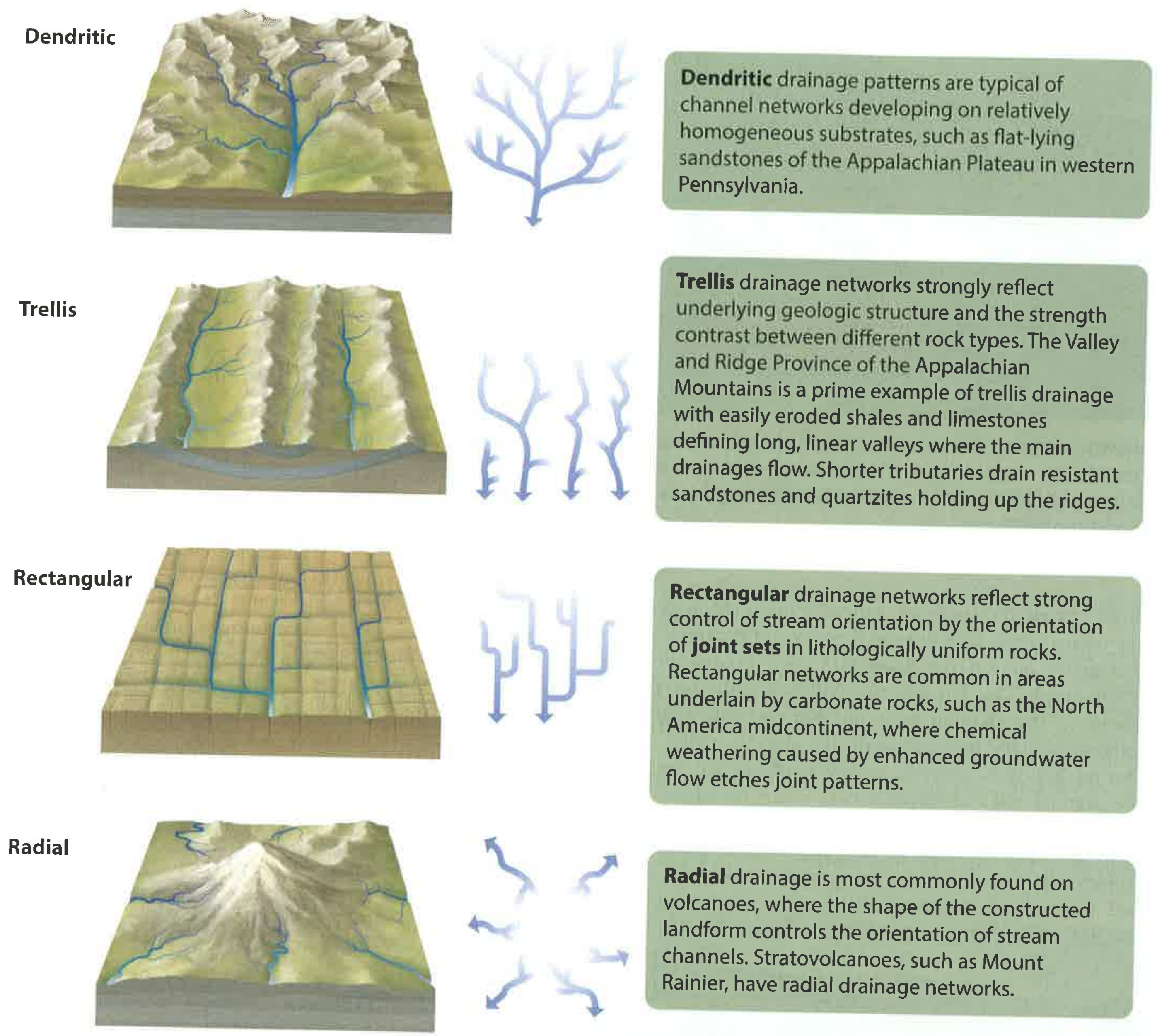


FIGURE 7.4 Drainage Patterns. Qualitative analysis of drainage patterns reveals the influence of underlying earth materials on the orientation of stream channels.

channels. Trellis drainage patterns form in areas underlain by tilted or folded beds of alternately weak and resistant sedimentary rocks, where preferential erosion along weak beds results in development of bedding-parallel strike valleys with short, steep, orthogonally oriented dip and antidip streams incised into the more resistant strata. Rectangular drainage is similar to, but more symmetric than, trellis drainage, with the two dominant drainage directions more equally developed. Rectangular drainage characterizes areas in which jointing or faults govern drainage patterns by producing linear zones that are more susceptible to weathering and erosion; such weaknesses are particularly common in landscapes underlain by carbonate rocks, such as limestones.

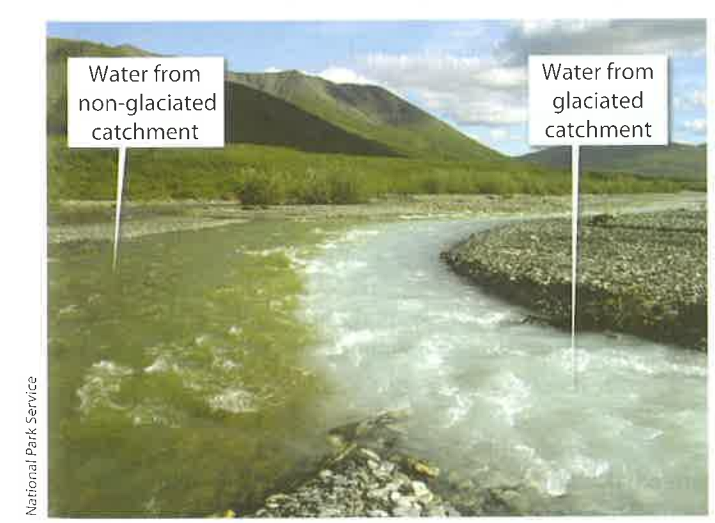
Topography can also influence channel networks. **Radial drainage patterns** consist of channel networks that flow away from or toward a central point. Radial drainage networks typically reflect flow off symmetrical landforms such as **volcanoes** or **structural domes.** Drainage patterns that consist of channels flowing toward a common point typically form in closed depressions, like volcanic calderas, craters, and down-dropped tectonic basins in extensional terrain.

The relationship of channel courses to the underlying geologic structure is another fundamental attribute of drainage patterns. In areas where rocks have been folded (for example, the Valley and Ridge Province in Pennsylvania), tributaries follow long linear courses between ridges

controlled by the underlying structural (geologic) fabric (see Photograph 12.13). Rivers and streams that follow geologic structure, as described above, are readily explained by preferential erosion focused on zones of weakness and therefore greater erodibility. **Consequent** streams are those that generally follow the regional geologic structure, such as by flowing down tilted beds.

More curious is the development of **transverse drainage** that cuts across geological structures. The most common explanation for transverse drainage is that the rivers predate the deformation that accompanied development of a mountain range. In this view, erosion by the **antecedent,** or preexisting, drainage was able to keep pace with uplift and maintain the river's prior course as the mountains rose around it. This is the favored explanation for the development of transverse drainage in many mountain ranges, such as the Himalaya and the Cascade Range in the Pacific Northwest. Another explanation for transverse drainage is **drainage superposition** during which an alluvial drainage pattern or river course developed on sedimentary cover erodes into and is imposed on the underlying bedrock as the cover gradually erodes. Major rivers crossing the Appalachian Mountains are thought to be **superposed.** However, the superposition mechanism is difficult to test because it inherently involves removal of the evidence for it (the initial overlying deposit).

Basin shape and the associated network pattern also affect the likelihood of important confluence effects, changes that occur where a tributary enters the main stem stream. Tributary–main stem confluences are often places where bed-particle size or water geochemistry changes as material from two different basins mixes [Photograph 7.8]. Stream junctions frequently represent important sites of



PHOTOGRAPH 7.8 Stream Junction. Cloudy, turbid stream water, originating from a glaciated catchment (right), merges with a clear stream (left) at a confluence in the Noatak River drainage, at the Gates of the Arctic National Park and Preserve, northern Alaska.

repeated disturbance and sedimentological heterogeneity, often leading to greater ecological diversity. The relative size difference between a tributary and the main stem determines whether a geomorphically effective confluence effect occurs. In general, major geomorphic change occurs at confluences where tributary basin size is ≥0.6 times the size of the main stem basin. These confluence effects are most likely to occur, for example, in oval basins having dendritic network patterns because such networks have large tributary basins. Rectangular basins with trellis network patterns have the least number of geomorphically important tributary inputs because, in general, trunk streams are large and tributaries are small.

Channel Ordering

Because river channels form networks, a variety of schemes have been developed that describe a channel's position and rank in the drainage network. Key to any network analysis is definition of the lowest-order or smallest channel. First-order streams are typically defined as a channel that regularly carries flow. While such a definition makes sense, implementation is more difficult because stream ordering is usually done from maps and is not based on extensive fieldwork; thus, the scale at which the basin is considered is critical. Stream orders calculated from ultra-high resolution LiDAR data will be quite different from those generated using large-scale topographic maps.

The dominant schemes used for channel ordering are those of Strahler and Shreve [**Figure 7.5**]. In the Strahler **stream ordering** scheme, when two first-order channels join, the channel downstream is designated as a second-order channel. When two second-order channels join, the result is a third-order channel, and so it goes down the network. Lower-order streams entering higher-order streams do not influence stream order in Strahler's approach. In contrast, Shreve's alternative approach defines **stream magnitude** as the total number of first-order streams contributing to the reach in question. Both stream order and magnitude are related positively to discharge, stream length, channel width, depth, and cross-sectional area, as well as sinuosity. Usually, stream order is inversely related to channel and basin slope. Some researchers argue that such relationships are the inevitable result of analyzing unidirectional branching hierarchical networks. The Strahler system is more commonly used to describe channel networks.

Downstream Trends

In addition to stream order, one can consider downstream changes in stream channel characteristics as a function of drainage basin area. Such area-based relationships have the advantage of being reproducible, independent of ordering scheme, and easily implemented in **Geographic Information Systems (GIS)**. Relationships between upstream basin area and channel characteristics depend critically on climate. In humid basins, discharge increases downbasin. In arid

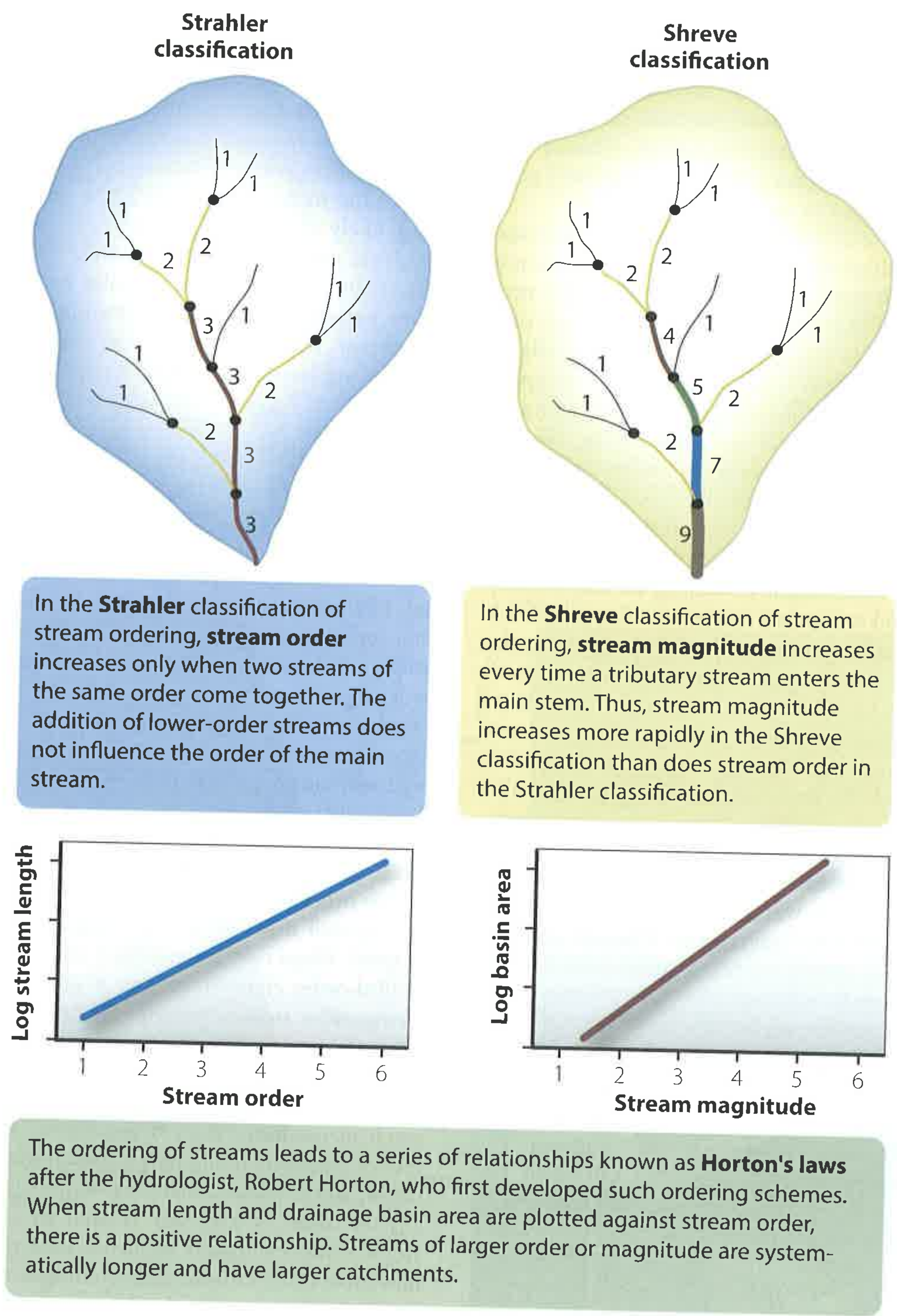


FIGURE 7.5 Stream-Order Classifications. Stream-order classification is a quantitative analysis procedure that assigns a value to a stream reach based on its place in the stream network. The two most common ordering schemes, those of Strahler and of Shreve, use different classification algorithms.

basins and in areas underlain by soluble rocks and karst, the relationships are often more complex because losing streams are common (see Chapter 4); thus, channels need not convey increasing discharges downstream.

In humid regions, there is usually a robust, positive correlation between basin area and both bankfull and mean annual discharge. Indeed, numerous studies have found that in humid-region drainage basins, basin area is positively related to many different hydraulic geometry variables, including channel width, depth, velocity, and cross-sectional area **[Figure 7.6]**. In general, basin area is inversely related to both channel and basin average slope.

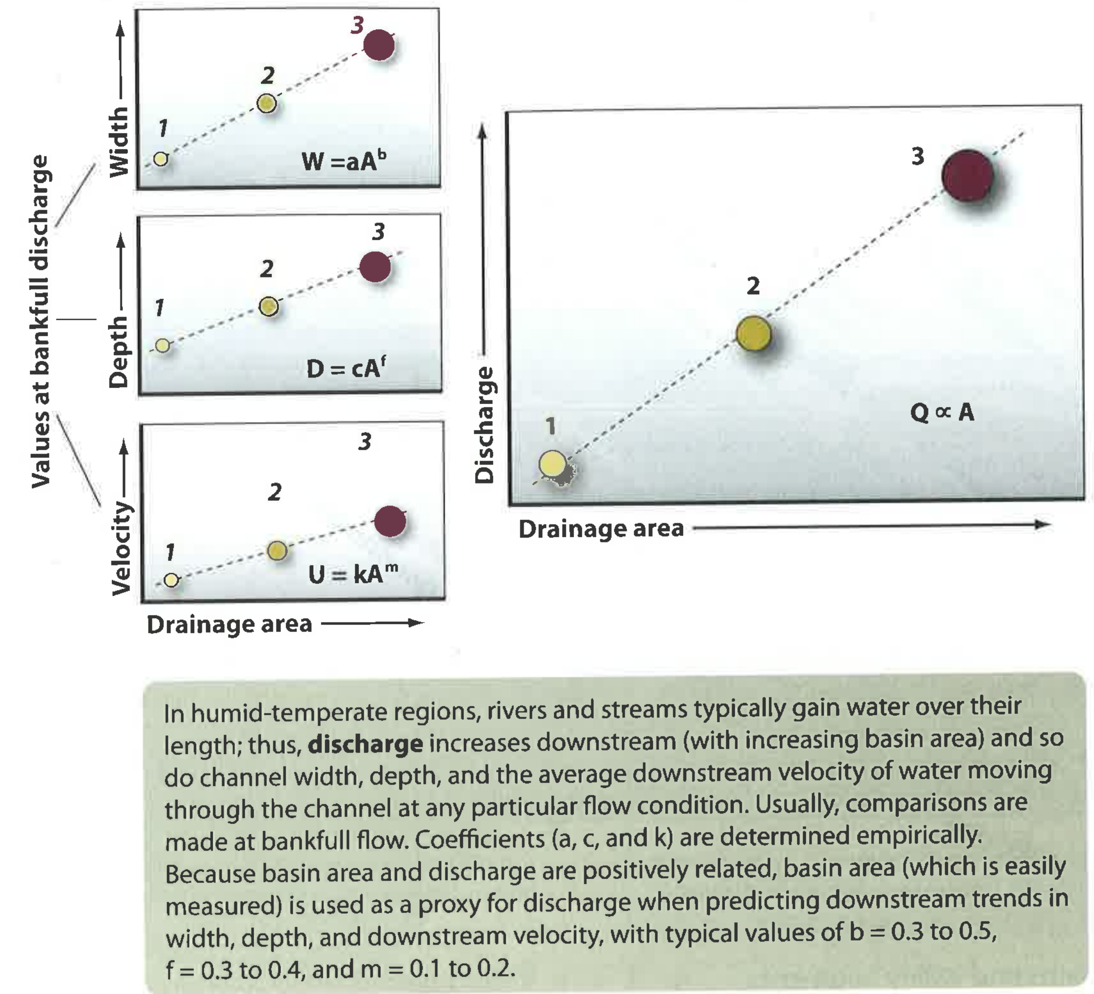


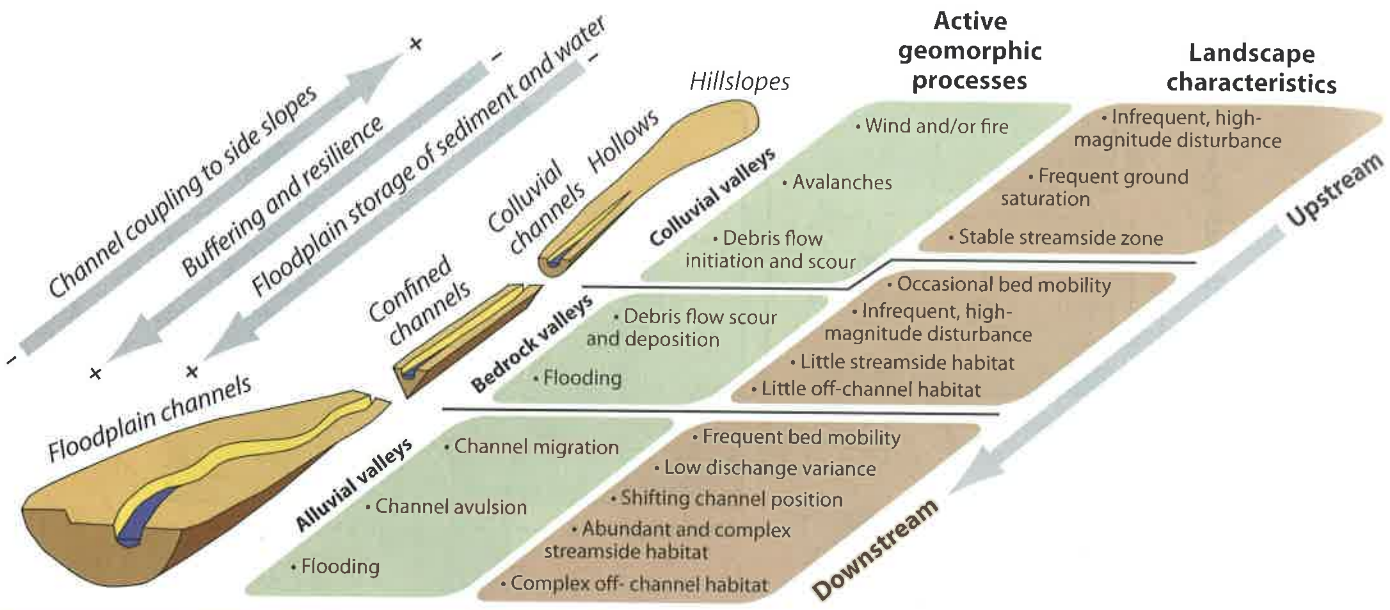
FIGURE 7.6 Downstream Changes in Channel Geometry. Downstream hydraulic geometry relationships quantify the observation that in humid-temperate drainage basins, channel width, depth, and water velocity all increase with basin area for a specified river level, such as bankfull.

Such correlation between basin area and geomorphically important channel characteristics means that by amalgamating hydraulic geometry data for multiple stations along a drainage, one can predict channel characteristics downstream at specific, geomorphically meaningful discharges, such as bankfull, as a function of basin area. Equations of the same form as at-a-station relationships (see Chapter 6, eq. 6.4) are used for downstream (basin area) relationships. When downstream relationships are plotted, the exponents represent the downstream rate of change in width, depth, and cross-sectional area with increasing basin area.

Finding that flow velocity generally increases downstream comes as a surprise to many people accustomed to thinking that the rushing, turbulent waters of mountain streams flow faster than the apparently calm water of large lowland rivers. However, the roughness of mountain streams and their shallow depth together lead to energy dissipation that more than makes up for their steeper slopes. Deep, wide lowland rivers flowing through relatively smooth channels have little friction along the bed and banks, allowing the water to move quickly despite the low slope (see Chapter 6).

Uplands to Lowlands

Drainage basins encompass a continuum of elevation, climate, and biota from their headwaters in the uplands to their outlets downslope. In large drainage basins, such as the Amazon, this continuum extends from frigid, lofty mountain peaks to sweltering tropical lowlands. Moving down the main stem river in a drainage basin, one can make predictions about changes in the dominant geomorphic process and form, changes indicative of slope, temperature, and weathering intensity **[Figure 7.7]**. In general, upland areas are sediment sources and lowland areas are sediment sinks.



Drainage basins are composed of hillslopes and channels, including unchanneled slopes high in the basin uplands and large floodplain channels in the lowland. In between are colluvial channels, just downstream of channel heads, and there can be confined channels in steep bedrock valleys. The active geomorphic processes that shape and disturb the landscape change predictably downstream, and result in a suite of landscape characteristics. Not all landscapes include all of the landforms illustrated here.

FIGURE 7.7 Hillslope-Channel Continuum. Active geomorphic processes, and thus the landscape character and landform distribution, change predictably from hillslopes to downslope valley bottoms. [Adapted from Montgomery (1999).]

Process Domains and Valley Segments

Different portions of a drainage basin are shaped by different processes. Hillslopes, hollows, channels, and valley bottoms define different process domains that may be distinguished on the basis of different relationships between drainage area and slope.

Near drainage divides on convex hillslopes (see Chapter 5), slope angle increases with distance downslope and thus with drainage area [Figure 7.8, middle right panel]. In contrast, drainage area-slope relations along the valley network typically exhibit an inverse relationship between slope and area:

$$S = K_s A^{-\theta_c} \qquad \text{eq. 7.2}$$

where K_s is the steepness index and θ_c is the concavity index. Spatial variations of the steepness index (K_s) have been attributed to regional and along-channel differences in precipitation, rock strength, sediment supply, and uplift rates. Different values of θ_c are typical for hollows, debris flow–dominated channels, and fluvial channels. Concavity index values range from 0.1 to >1 and are thought to reflect the influence of different surface processes, thus allowing recognition of different process domains within a landscape.

In mountain drainage basins, values of $\theta_c < 0.3$ characterize steep headwaters influenced by debris flows and channels with downstream increases in incision rate or rock strength. Bedrock channels dominated by fluvial incision typically exhibit $0.3 < \theta_c < 0.7$, and high concavities ($\theta_c > 0.7$) are associated with alluvial rivers. These downstream variations in river profile concavity are associated with identifiable **valley segments** defining portions of the channel network that have similar valley-scale morphologies and in which specific geomorphic processes operate (see Figure 7.8).

Colluvial, bedrock, and **alluvial** valley segments reflect different relative balances between sediment transport capacity and sediment supply along fluvial systems. Depositional **estuarine** valley segments define the transition from terrestrial to marine environments. In many landscapes, and mountain drainage basins in particular, fluvial processes in headwater valleys are relatively ineffective at transporting sediment delivered from surrounding hillslopes. Consequently, colluvial valley fills, composed of material that has undergone little to no transport by flowing water, accumulate in valley bottoms. Material stored in **colluvial valleys,** including the unchanneled valleys or hollows at the head of the channel network, is then transported to downstream channels when entrained by debris flows or rare high flow events [Photograph 7.9]. Colluvial valleys are net sediment sources. Colluvial channels that dominate the headwaters of steep, landslide-prone drainage basins typically give way to fluvial channels, such as those discussed below, at drainage areas of 1–10 km^2.

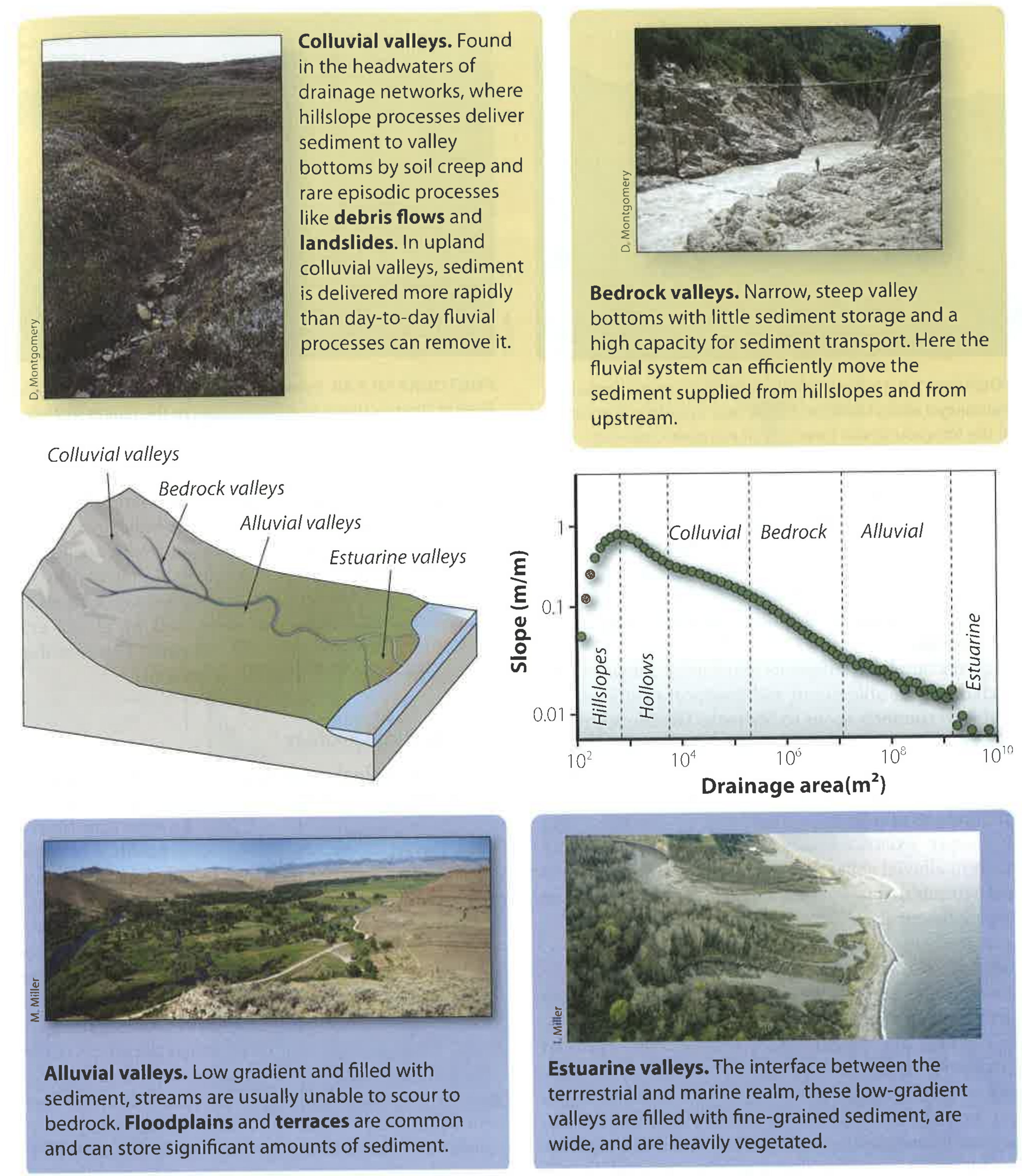
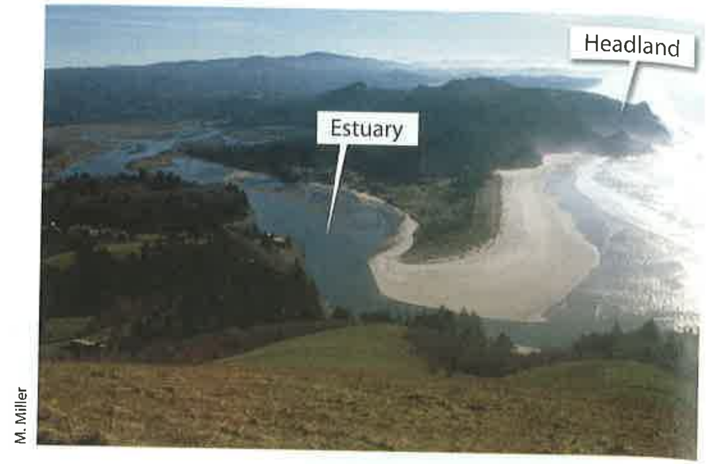


FIGURE 7.8 Valley Segment Types. Characteristic valley segment types are found in different parts of the drainage network: colluvial valleys in the uplands, bedrock valleys where sediment transport capability exceeds supply, alluvial valleys in low-gradient lowlands, and estuarine valley segments where a river meets the ocean, or a local base level. [Adapted from Montgomery (2001).]

Bedrock valleys generally lack significant valley fill and have bedrock valley walls that typically confine channel migration to a narrow, often V-shaped valley bottom (see Photograph 7.4). Narrow valley bottoms favor development of relatively straight channels, although some regions have deeply entrenched bedrock meanders. Channel floors in bedrock valleys generally consist of either patchy accumulations of alluvium, exposed bedrock, or a mixed morphology of alternating alluvial patches and bedrock outcrops. The geologically insignificant sediment storage

PHOTOGRAPH 7.9 Hollows. Hillslope hollows (unchanneled colluvial valleys) along Mettman Ridge, near Coos Bay, Oregon. Area in the foreground was forested but has been clear-cut.



PHOTOGRAPH 7.10 Estuary. Low-energy estuary of the Salmon River in Oregon stands in stark contrast to the raging surf off the rocky headland.

in bedrock valleys, which can amount to just a few years' worth of sediment in active transport through the channel network, indicates rapid downstream export of the sediment supplied from hillslopes. Rates of bedrock valley incision limit the lowering of adjacent hillslopes and thus set the pace of landscape evolution in many unglaciated upland settings.

Thick, unconsolidated deposits characterize **alluvial valleys** where channels are able to sort and transport their loads but are unable to routinely scour to bedrock. The transport and deposition of sediment in alluvial valleys leads to the formation of floodplains and valley bottoms across which channels can migrate unconfined by bedrock valley walls. In mountainous terrain, postglacial fluvial sedimentation can fill in deeply excavated glacial or preglacial valleys. Channels in alluvial valley segments exhibit a wide range of channel patterns that reflect differences in sediment transport capacity, sediment supply, and vegetation type.

Low gradients and extensive deposits of fine-grained sediment characterize estuaries where saltwater and freshwater mix, and there can be a strong tidal influence moving sediment both up and down the estuary **[Photograph 7.10]**. Valley bottoms are typically wide and heavily vegetated with extensive areas of land that are occasionally submerged, either by exceptionally high tides or flood flows. During generally rising Holocene sea levels, estuaries were usually net sediment sinks, but during times of lower sea level in the past (such as at glacial maxima around 21,000 years ago when sea level was ~130 m lower), former estuaries would have become sediment sources as they were incised by rivers responding to lower sea level.

Valleys are formed in different ways and their shapes reflect both how they formed and how they were later modified. The formation of some valleys is tectonically influenced, such as those valleys running along the trend of faults where weak, brecciated (broken) rock is common. The shape of other valleys is directly controlled by the underlying structure, such as the linear valleys of the Valley and Ridge Province of the Appalachian Mountains. The shape of valley segments in cross section can be indicative of the processes that both formed the valley and subsequently modified it. For example, valley segments dominated by fluvial erosion tend to be V-shaped in cross section with relatively planar slopes. Valley segments shaped by glacial erosion tend to be U-shaped with wide, flat valley bottoms abutting steep, sometimes near-vertical valley walls.

Longitudinal Profiles

River **longitudinal profiles,** plots of channel elevation versus downstream distance, are generally concave-up, reflecting a systematic decrease in slope downstream from their headwaters **[Figure 7.9]**. When such a profile is smooth and concave-up, the river and its drainage basin are termed **graded,** implying that the river's slope has adjusted to a steepness that allows all the sediment supplied to the river to be transported **[Box 7.1]**. At the downstream end of the profile is the **base level.** The ocean, a lake, another stream, or a bedrock obstruction that controls the elevation of the channel can provide base-level control. Over geologic time, base level often moves up and down as the effects of climate change cause sea level and the water surface of closed-basin lakes to rise and fall. Uplift driven by tectonics and isostatic rebound alters the relative elevation of different parts of the landscape and can change base level. In general, rivers respond to falling base levels by incising their bed and to rising base levels by depositing sediment and aggrading.

Relief is a widely used description of drainage basins and can be considered and calculated in various ways. Total **basin relief** is simply the lowest elevation in the basin subtracted from the highest elevation. **Fluvial relief** is the elevation difference between the highest and lowest points on the river network and is less than the basin relief because of the slopes that connect the highest points in the basin to the channel heads where streams initiate. In mountainous terrain, wider mountain ranges tend to have

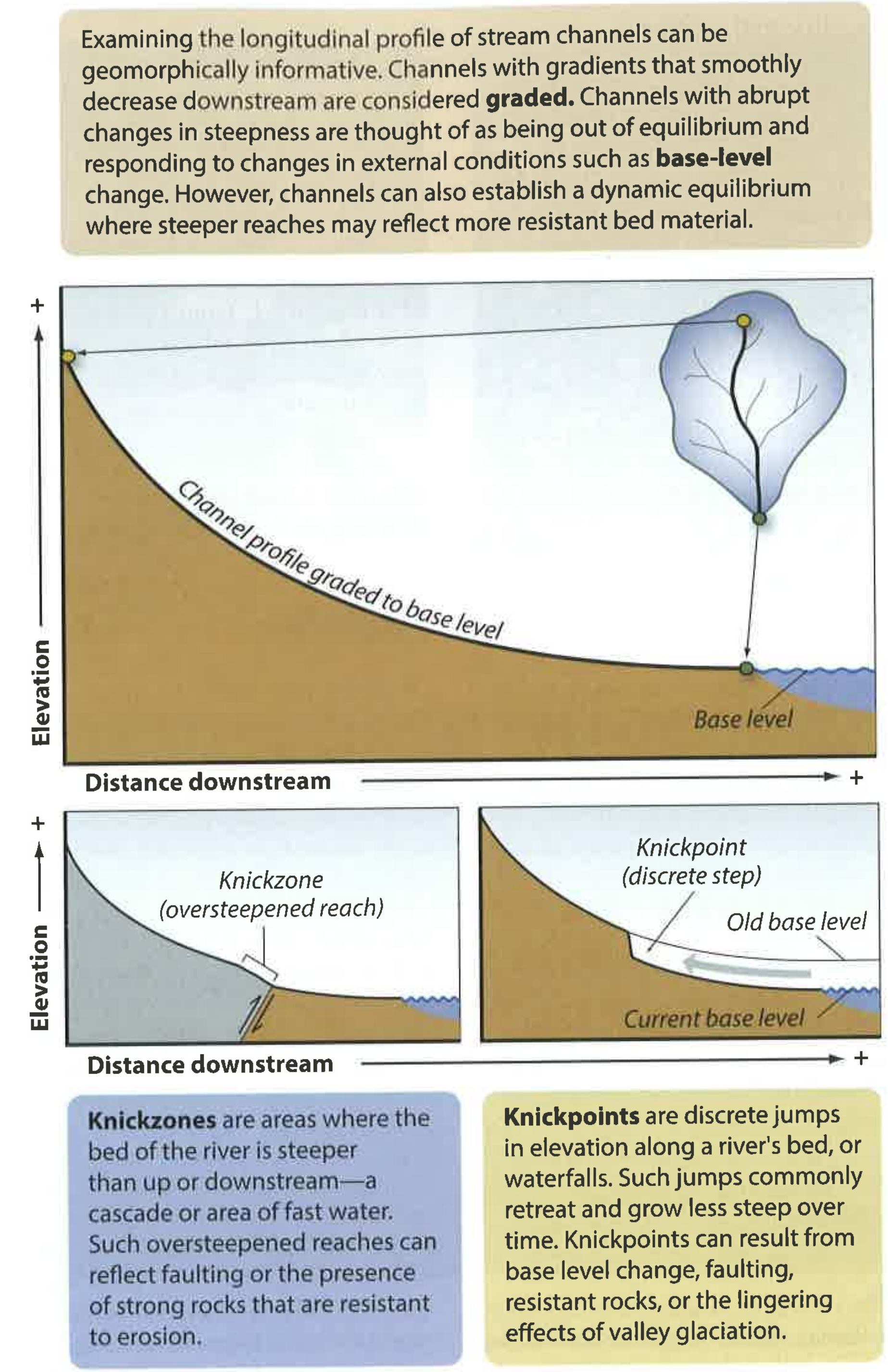


FIGURE 7.9 Longitudinal Profiles. Departures of river longitudinal profiles from the ideal, smoothly graded case reflect the influence of rock type, structure, climate/glaciation, tectonic forcing, and/or the effects of changing base level. Discontinuities in longitudinal profiles are referred to as knickzones and knickpoints.

greater relief because the main stem rivers cutting through them are longer and thus have a greater channel length, generating greater fluvial relief. In part to normalize for this effect, the **local relief** can also be calculated as the minimum and maximum elevations within an area, generally a circle of set radius, typically 5 or 10 km, or a grid cell of comparable dimensions.

Channel Confinement and Floodplain Connectivity

From the headwaters downstream to the channel mouth, there are significant differences in the lateral extent of floodplains and their connectivity with the channel. In the uplands, floodplains, if they exist, are narrow and channel

BOX 7.1 River Longitudinal Profiles

A simple prediction for the shape of a river longitudinal profile may be arrived at by assuming that the rate of energy loss, or stream power ($\Omega = \rho_w g Q S$), is uniform along the length of a river system. Because discharge generally increases with distance downslope along river systems, we can consider the predicted form of a river profile if there is a constant value of the stream gradient index (SL) defined by the product of channel slope (S, dz/dx) and distance from the drainage divide (x_d):

$$SL = (d_z/d_x)\, x \qquad \text{eq. 7.A}$$

Rearranging this expression and adopting a sign convention consistent with elevation dropping downslope results in an expression for channel slope as a function of SL and x:

$$d_z/d_x = -SL/x_d \qquad \text{eq. 7.B}$$

Integrating this expression with respect to x yields an expression for the riverbed elevation as a function of distance downstream:

$$z = -SL \ln x + c_1 \qquad \text{eq. 7.C}$$

where c_1 is a constant of integration. Considering the case where elevation declines to a base level of zero at a distance L from the drainage divide implies that $c_1 = SL \ln L$, in which case the full profile of the river may be described as a logarithmic function of distance downstream:

$$z = -SL \ln(x_d/L) \qquad \text{eq. 7.D}$$

Different values of the slope-gradient index (SL) lead to differing degrees of concavity in the overall river profile.

migration is confined by valley walls. Many steep upland basins have no floodplains at all; steep rocky hillslopes directly border the channels. In the lowlands, floodplains are wide and provide extensive areas for both temporary and long-term sediment storage. Wide floodplains, and the riparian vegetation communities they support, provide significant roughness that slows flood flows, dissipating energy and encouraging the deposition of additional fine sediment.

Floodplain connectivity, the ability of floodwaters to reach the floodplain, is easily changed by human activity; for example, dredging channels or building levees prevents floodwater from leaving a river. This increases the velocity and the depth of floodwaters in the river channel, thereby exacerbating the potential for damage downstream. Flow regulation by dams can also have significant effects on channel/floodplain connectivity by preventing or limiting the occurrence of overbank flows.

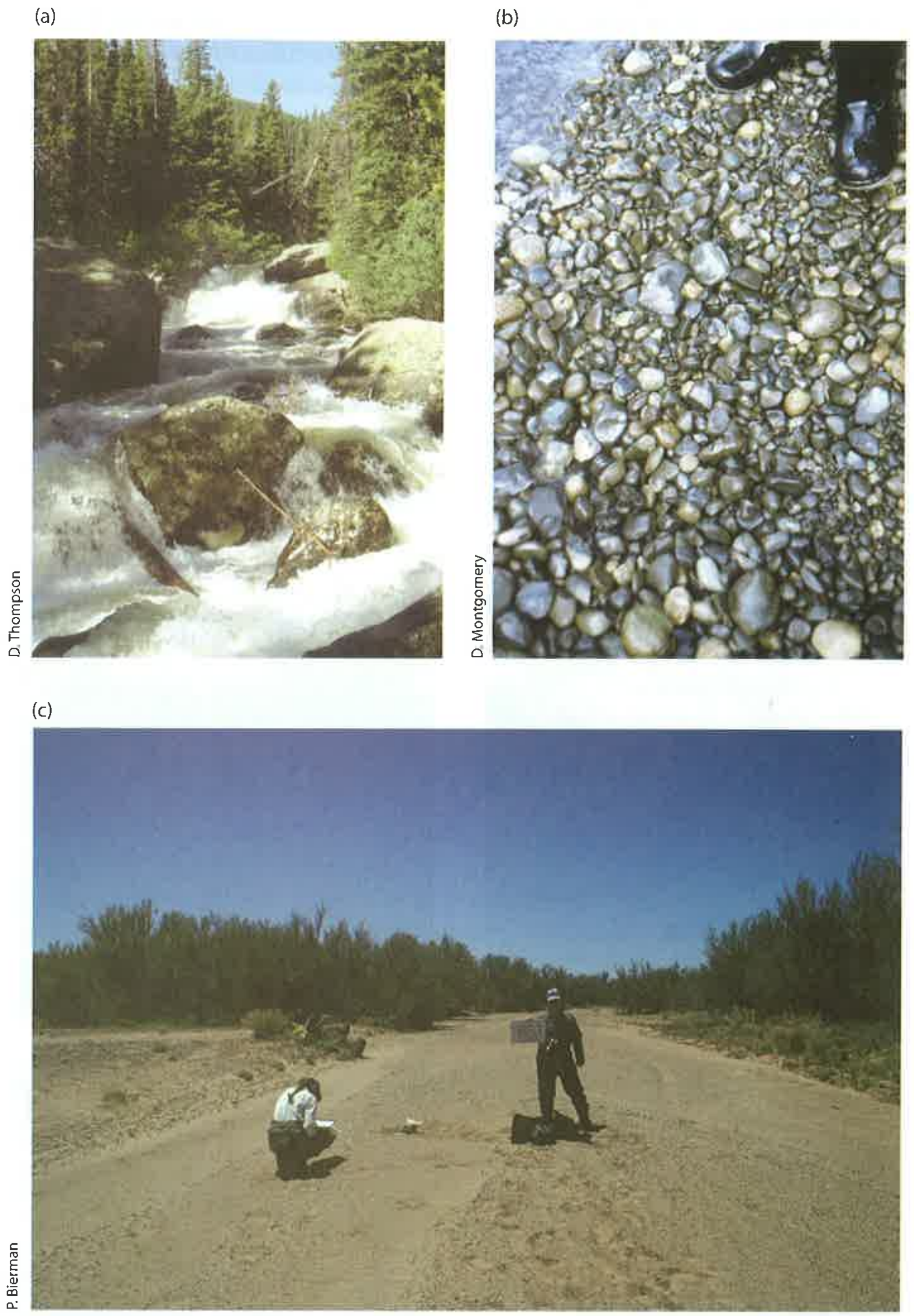
Downstream Trends

The characteristics of fluvial sediment (grain size, shape, and composition) systematically change downstream in channel networks. Bed material grain sizes typically decrease downstream in a drainage basin **[Photograph 7.11]**. Some of this fining is due to abrasion of grains as they are transported, some is due to weathering and breakdown of clasts during storage in terraces and bars, and some is due to selective transport through which finer grains are preferentially conveyed to downstream reaches.

In the small headwater channels of mountain drainage basins, mean grain size typically increases downstream from the channel head to a maximum before the onset of downstream fining that is typical for most of the channel network. The initial downstream increase is the result of slope processes (including rock fall, debris flows, and soil creep) delivering a range of grain sizes, including clasts large enough that flows in the relatively small channel are unable to move them. As the channel grows larger downstream and discharge increases, a larger fraction of the finer material is removed and fluvial processes become increasingly effective at transporting larger material. In addition, as floodplains widen downstream, it becomes less likely that a valley-margin landslide will deliver coarse material directly to the stream. Downstream of the zone where hillslope processes dominate sediment delivery to channels, bed grain size begins to fine downstream as particles abrade or shatter during fluvial transport and weathering during periods of storage along river systems.

Local inputs of coarse material to the channel, such as from steep tributaries, rock fall, and landslides, can reset downstream grain-size patterns. Similarly, downstream patterns of bed surface fining can be reversed where channel slopes steepen. Moving downstream in a drainage basin, one typically encounters boulder-bed, gravel-bed, and then sand-bed channels. Although the downstream transition from boulder-bed to gravel-bed channels is gradual, the transition from gravel-bed to sand-bed channels is typically rather abrupt. Sand-bed channels are actively mobile at most (if not all) flows, whereas gravel-bed and boulder-bed channels exhibit threshold mobility in which the bed is stable at flows lower than a critical discharge.

Particle size and shape change downstream as abrasion increasingly rounds clasts during transport. In basins with heterogeneous geology, the composition of particles in fluvial sediment also changes with distance downstream, as more readily weathered grains are altered. Weathering-resistant grains, such as feldspar and quartz, form an increasing proportion of the clast sediment population with distance downstream.



D. Thompson

D. Montgomery

P. Bierman

PHOTOGRAPH 7.11 Channel Bed Material. (a) Boulder-bed mountain channel with a step-pool morphology, North Saint Vrain Creek, Colorado. (b) Gravel stream-bed, Washington State. (c) Dry sand-bed channel of the Rio Puerco, New Mexico.

Drainage Basin Landforms

Drainage basins contain a variety of landforms that result from the complex interaction between rivers and both solid and unconsolidated earth materials. These landforms can be used in some cases to interpret landscape history and in others to infer tectonic and climatic change. The distribution of drainage basin landforms is closely tied to the history and location of base-level controls (see Longitudinal Profiles section above).

Knickpoints

Deviations from smooth concavity in a river's longitudinal profile are often interpreted as the influence of tectonics, climate history, and/or lithology. For example, uplift along a fault over which a river is flowing will often result in an unusually steep reach along a channel, known as a **knickzone** (see Figure 7.9). Some steepened stream reaches reflect the presence of immobile debris deposited into the channel by landslides and debris flows, and are not related to base-level change. If the steepened area is a discrete jump in bed elevation (a waterfall), it is known as a **knickpoint** [**Photograph 7.12a**]. Longitudinal profiles of streams in glaciated landscapes often have steps related to hanging valleys inherited from glacial times, steps that fluvial erosion over the Holocene Epoch has not been able to erase [**Photograph 7.12b**]. Steep channel reaches can also reflect the presence of strong rocks, such as quartzite or volcanic dikes, in the midst of otherwise weaker lithologies such as shale or mudstone [**Photograph 7.12c**]. The change in slope downstream affects important geomorphic processes, especially sediment transport and the energy available for channel bed incision.

Some knickpoints, once established, retreat upstream over time, leaving in their wake incised channels and abandoned floodplains, which then become terraces. Knickpoint retreat is easily demonstrated in some locations, for example, at Niagara Falls, on the border between Ontario, Canada, and upstate New York and in many small catchments affected by postglacial isostatic response and relative sea-level drop such as in Scotland. In other areas, such as the Great Falls of the Potomac River, cosmogenic dating of exposed bedrock terraces is more consistent with spatially uniform incision and persistence of the knickzone at about the same location. At Great Falls, it appears that the river incised downward into rock along a several-kilometers-long reach of the channel at the same time.

Gorges

Gorges are deeply incised reaches along rivers, the result of river channels cutting into competent material, usually rock, that can hold steep to near-vertical slopes [**Photograph 7.13**]. Gorges form in steep channel reaches

PHOTOGRAPH 7.12 Knickpoints. Knickpoints come in different sizes and reflect different conditions and processes. (a) Skogafoss, a large waterfall fed by glacial meltwater on the southern coast of Iceland, acts as a migrating knickpoint in response to changes in base level imposed by the eruption of basaltic lava flows. (b) Hanging valleys are common in once-glaciated terrain. Here, Bridalveil Creek hangs above the deeper Yosemite Valley, California, forming the famous falls. (c) This waterfall in Bryce Canyon National Park, Utah, reflects a resistant layer of bedrock. Undermining at the base of the falls allows upstream migration of the knickpoint.

where sufficient energy is available to erode rock and prevent sediment accumulation. Oversteepening can be the result of tectonic uplift; for example, deep gorges are found on Himalayan rivers that flow across localized areas of active rock uplift. Diversion of rivers by glaciers or large rock slides can steepen gradients and lead to localized incision. Many gorges in northeastern North America are the result of postglacial rivers reestablishing their courses over bedrock ridges and then cutting down through these obstacles.

Gorge incision occurs through a variety of bedrock incision processes including plucking, pothole formation, and abrasion. Because flow in gorges is deeper and narrower than in reaches that typify streams outside of gorges, unit stream power exerted on the bed is higher. This provides a positive feedback that enhances incision within gorges.

Very narrow, steep-sided **inner gorges** are often found incised into the bottom of wider gorges or valleys. Inner gorges in nonglaciated regions of the Himalaya and northern California have been attributed to recently increased uplift rates and/or concomitant base-level fall, leading to renewed and focused incision. In some locations, inner gorge formation has been attributed to greater soil moisture, producing a higher landslide frequency on the lower portions of valley walls. The origin of V-shaped gorges at the base of glaciated U-shaped valleys invites a different question, especially in tectonically quiescent landscapes. Do these gorges record postglacial incision in response to lowered base level or have they survived under ice through multiple glaciations?

Terraces

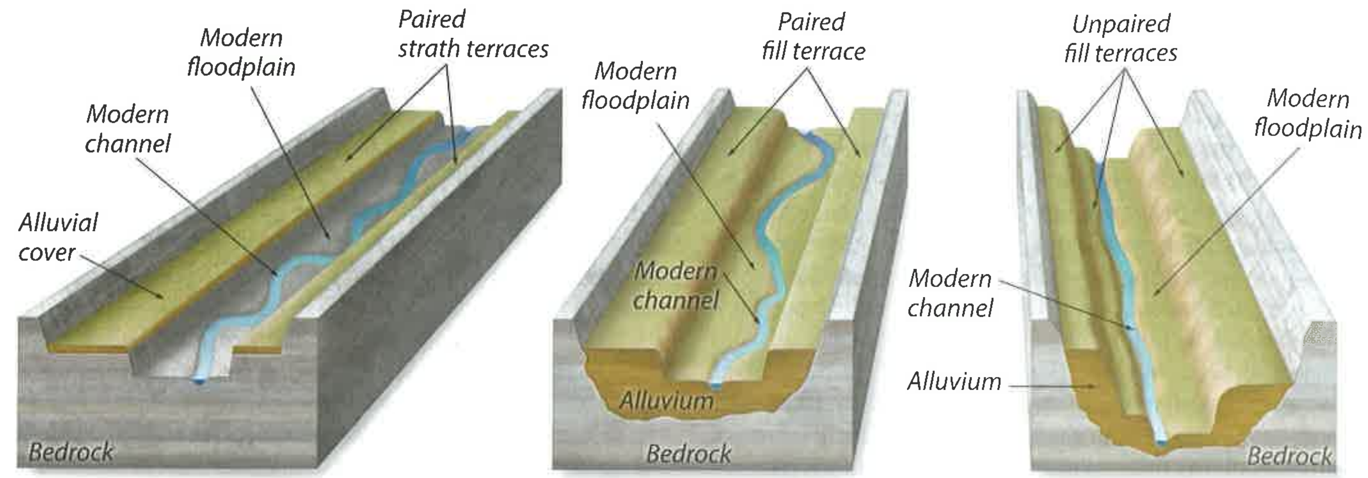
Terraces are important drainage-basin landforms that preserve sediment and provide a record of the past behavior of the fluvial network, as well as of the supply and nature of sediment delivery to the river system. Over time, as rivers shift position and erode their banks, terrace remnants are removed so that eventually only isolated parts of any originally continuous terraces remain.

The existence of terraces provides strong evidence that thresholds are important in geomorphology. For terraces to form, floodplain abandonment must be an episodic rather than gradual process. Terrace formation can be caused by changes in many different boundary conditions. Changes in sediment supply, often linked to changes in climate over the Quaternary (the Pleistocene and Holocene Epochs), have caused many rivers to aggrade and then abandon and incise their floodplains, leaving flights of terraces behind. In areas that have been uplifted, terraces, if they can be dated, are a means by which to characterize rates of uplift and surface deformation over time.

Terraces can be divided into erosional and depositional terraces [**Figure 7.10**]. Terraces eroded into stable substrates are known as **strath terraces** [**Photographs 7.14 and 7.15**]. Aggradational terraces [**Photograph 7.16**] that represent filling of an already incised valley are known as **fill terraces** and consist of fluvially deposited unconsolidated sand and gravel. They form when, after a period of incision and valley formation, a river does not have the transport capacity to move the sediment load supplied to it from drainage basin slopes. When formed, fill terraces were laterally continuous floodplains across the valley and downstream. Fill terraces are common in valleys just outside glacial margins. For example, in the Teton Range near the Grand Teton in Wyoming, aggradational terraces reflect the very large sediment loads supplied to nearby rivers by glacial meltwater during glacial periods. Fill terraces can also reflect increasing sediment supply off hillslopes when the climate changes. Isolated terrace remnants perched high on gorge walls indicate that thick stacks of sediment once filled deep valleys in the Himalayan Mountains. The terraces, of which these sediments are the last surviving evidence, are thought to reflect increased monsoonal activity and sediment delivery in the mid-Holocene, about 7000 years ago.

Strath terraces have thin deposits of unconsolidated alluvium overlying beveled and better consolidated material beneath. The beveled material can be bedrock or consolidated

PHOTOGRAPH 7.13 River Gorge. Steep river gorge cut into rock along a tributary to the Salween River in eastern Tibet. Note sediment deposition and temporary storage in the foreground where the canyon widens (along with the washed-out pier of a bridge).

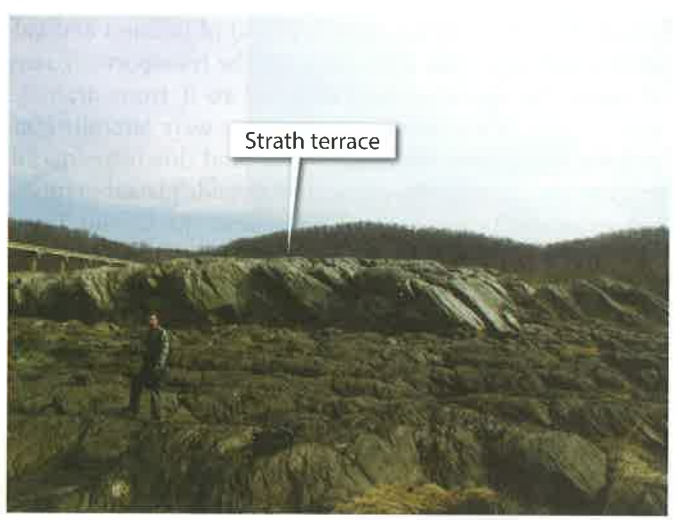


Erosional fluvial terraces are indicative of a geomorphic regime in which the river has sufficient energy not only to move the sediment load supplied to it, but to cut into the material that makes up the channel bed. Terraces formed by erosion are referred to as **straths.** Such terraces are frequently found in areas where active uplift or tilting provides the potential energy for incision. Straths can be covered by a thin layer of alluvial sediment.

Depositional fluvial terraces are indicative of river systems where sediment supply once exceeded the capacity of the river to transport sediment. The excess sediment was deposited in valley bottoms, filling them. At later times, if the sediment transport capacity increases (from more water, or steepening of the river gradient by tectonic tilting) or the sediment supply decreases, then the depositional surface can be incised.

Paired terraces are found at the same elevation across the width of a valley. Paired terraces form when river migration rates across the valley are rapid in comparison to incision rates. **Unpaired terraces** are found on only one side of the valley and result when rivers incise much more rapidly than they migrate across the valley bottom. Unpaired terraces also can result from stream erosion removing the terrace from one side of the valley but not the other.

FIGURE 7.10 Terrace Types. Depositional (fill) and erosional (strath) terraces form in response to different forcings. Erosional terraces imply incision capable of eroding underlying rock or sediment. Depositional terraces indicate changing sediment-transport capacity and/or sediment supply for the channel.

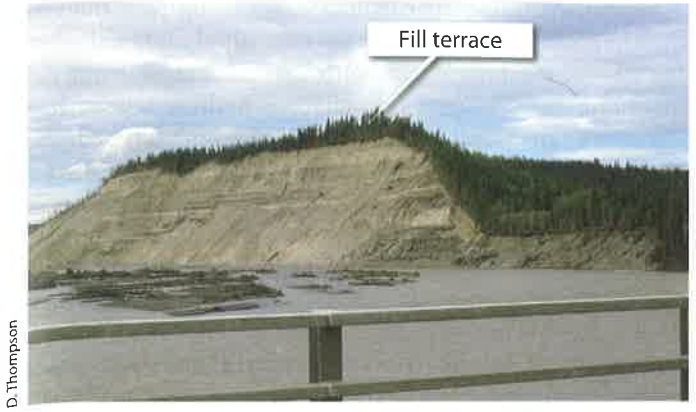


L. Reusser

PHOTOGRAPH 7.14 Strath Terrace. Series of bedrock strath terraces along the Susquehanna River in Holtwood Gorge, Pennsylvania. Any alluvium that once capped these terraces has been removed by later river flows.

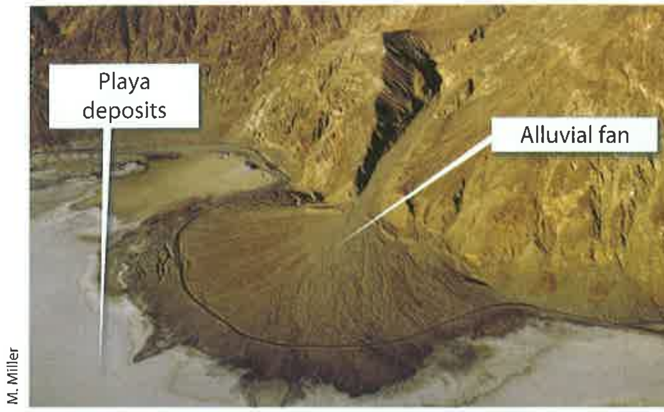
D. Montgomery

PHOTOGRAPH 7.15 Strath Terrace and Alluvial Mantle. Strath terrace eroded across steeply dipping sedimentary rock and covered with a thin mantle of alluvium, Nenana River, Alaska.



D. Thompson

PHOTOGRAPH 7.16 Fill Terrace. This fill terrace along the Copper River in Alaska shows that the river once aggraded significantly before incising.



M. Miller

PHOTOGRAPH 7.17 Alluvial Fan. Dark alluvial fan abutting white playa deposits in Death Valley, California. Note the road around the toe of the fan.

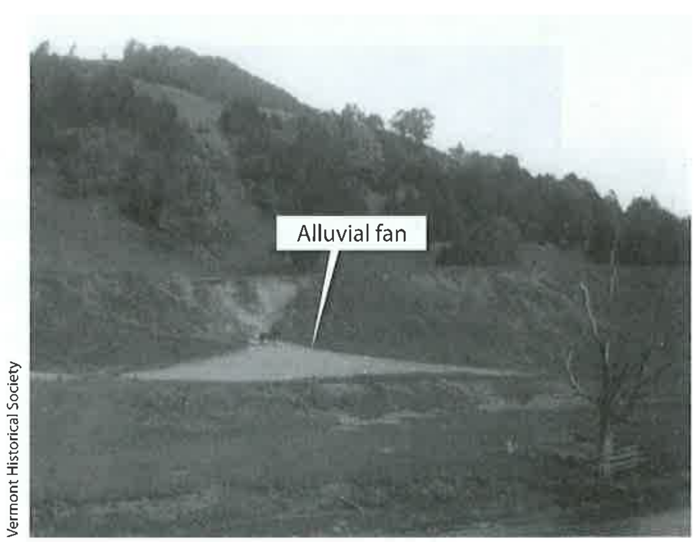
sediment. Current hypotheses suggest that strath terraces form when a river with a bedrock-floored channel meanders across a valley bottom and that strath formation is favored by, but not restricted to, weak rocks—especially those that lose strength when desiccated (such as micaceous siltstone). Beveling bedrock is widely thought to require a thin, mobile layer of sediment, which provides the tools for cutting the bed (see Chapter 6, Figure 6.3, and Figure DD6.4). If the river carries too much sediment, it will aggrade rather then bevel the rock beneath; if the river carries too little sediment, it will incise. Strath terraces are abandoned when boundary conditions change, allowing the river to incise. The incision leaves the strath terrace and its thin blanket of alluvial sediment high and dry (see Photograph 7.15).

The spatial distribution of terraces reflects the ratio between vertical and lateral rates of channel migration. Rapid downcutting, with little lateral channel migration, leads to **paired terraces** at similar elevations on both sides of the channel. More rapid lateral migration over greater distances provides the opportunity for **unpaired terraces** to form in close proximity to one another but at different elevations. Such unpaired terraces form because by the time a laterally migrating river returns to its original position, the river has incised enough that it forms a floodplain at a new, lower level (see Figure 7.10).

There are numerous reasons why rivers form terraces. In rapidly uplifting, tectonically active areas, strath terraces are left behind as rivers incise into rock. Changes in base level, such as those resulting from the draining of ice marginal lakes in the mountainous terrain of New England, have resulted in strath terrace sequences as newly reestablished rivers cut through large accumulations of glacial and postglacial sediment. These are not bedrock-floored strath terraces but rather terraces cut into cohesive glacial lake sediments and the underlying glacial till.

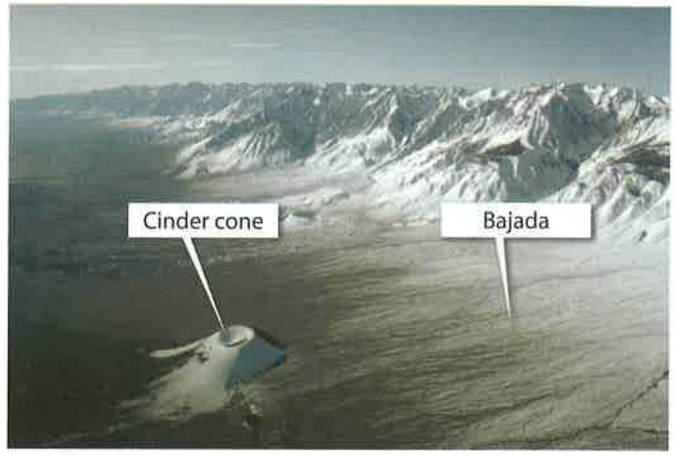
Fans

Fans are cone-shaped depositional features that form in both arid and humid regions where generally steep channels emerge from confined canyons and narrow valleys onto lower gradient surfaces where channels and flows are less confined. Sediment supplied from the highlands builds up on fans from which it cannot easily be removed. Fans are a dominant and conspicuous landform in arid regions where fluvial sediment transport processes in valley bottoms are limited by the paucity of water. Fans are also present in humid temperate zones, although they are often hard to find under forest cover. Large fans are commonly found at the base of slopes in tectonically active regions where rates of hillslope erosion are very high and sediments fill the valleys of the Basin and Range in western North America **[Photograph 7.17]**. Small fans can form in places that receive more sediment from upslope than can be removed by surface processes **[Photograph 7.18]**.



Vermont Historical Society

PHOTOGRAPH 7.18 Historical Alluvial Fan. A horse stands at the apex of a small alluvial fan near Tunbridge, Vermont, in the late 1800s. This fan, an accumulation of material eroded from the river terrace above, testifies to the ease with which soil moves after trees are stripped from the landscape.



M. Miller

PHOTOGRAPH 7.19 Bajada. Fault-controlled mountain front and recently erupted cinder cone on a bajada (the surface of coalescing fans sloping gently away from the mountain front) draped in snow. Fans here, in the Owens Valley below the Sierra Nevada, California, are dominated by debris-flow deposits that accumulated primarily during glacial periods.

The largest fans generally form where channels cross a mountain front and flow out onto a structural valley. Along many mountains fronts, drainage basins are spaced closely enough and sediment loads are sufficiently high that individual fans merge into a broad, low-gradient piedmont termed a **bajada,** from the Spanish for "slope" **[Photograph 7.19]**. Fan surfaces typically decrease in slope, and fan sediments typically decrease in grain size, from the **fan head** to the distal end of the fan, the **toe.** Changes in sediment load due to climate change and/or tectonic tilting can result in sediment redistribution on fans with **trenching** (incision) at the head of fans and deposition at the toe.

Sediment transport capacity is lower on fans than in the channels of drainage basins that supply fan sediment for several reasons. When source streams cross from the mountain front onto a fan, channel hydraulic geometry changes as the streams are no longer confined in bedrock channels but rather can adjust their geometries in unconsolidated fan materials. In arid regions, where fans are a common landform, there can be significant infiltration losses from the channel, reducing discharge and sediment transport capacity down fan. In addition, channel slope declines down fan, reducing stream power. Streams flowing on fans commonly split into several channels resulting in less stream power per unit area (see eq. 6.5) on the broad fan toe than at the fan apex.

Geomorphologists distinguish **alluvial fans** and **debris fans** on the basis of the dominant sediment-transport process, as inferred from surface morphology and the sedimentology of fan deposits. Although some fans are composed entirely of fluvially transported material and others are entirely composed of debris flow–delivered sediment, many fans are composed of a mix of water-borne sediments and debris-flow deposits. Nevertheless, the distinction persists in the literature based on the observation of depositional processes and deduction from fan stratigraphy that debris fans are built primarily by debris flows, while alluvial fans are built primarily by stream flows **[Photograph 7.20]**.

Alluvial sediments tend to be better sorted than debris-flow deposits, and many alluvial sediments are **clast-supported** (one clast resting on another with no intervening fine-grain matrix). Active alluvial fan surfaces are commonly dominated by channels with bars and swales formed by flowing water. In cross section, the alluvial sediments are sorted by grain size and stratified, indicating

(a) (b)

P. Bierman

PHOTOGRAPH 7.20 Alluvial and Debris Fan Formation. Fans can result from both alluvial and debris flow deposition. (a) Trench into small fan in Huntington, Vermont, reveals stratified layers of water-lain, alluvial sediment demarcated here by flags. A standard soil color chart (known by its trade name, Munsell) is being used to describe the color of the fan sediments. (b) Nearly 2-meter-high geologist attempts to climb an exposure in a fan dominated by debris-flow deposition, as indicated by the unstratified, clast-supported nature of the deposit, in Coso Range, southern California.

fluvial transport. Alluvial deposits typically lack very fine material, at least close to the range front.

In contrast, debris-flow deposits are typically unsorted and **matrix supported** (there is a fine-grained matrix between the clasts). The surface morphology of fans dominated by debris flows is varied but can be distinctive. Common landforms include channels bordered by debris-flow levees, channels plugged by debris flows that ran out far enough that the shear stress imparted by gravity acting on the fan slope was no longer sufficient to overcome the **yield strength** of the flowing debris, "freezing" it in place **[Photograph 7.21]**, and lobes of debris that spilled out of channels and onto the broader fan surfaces.

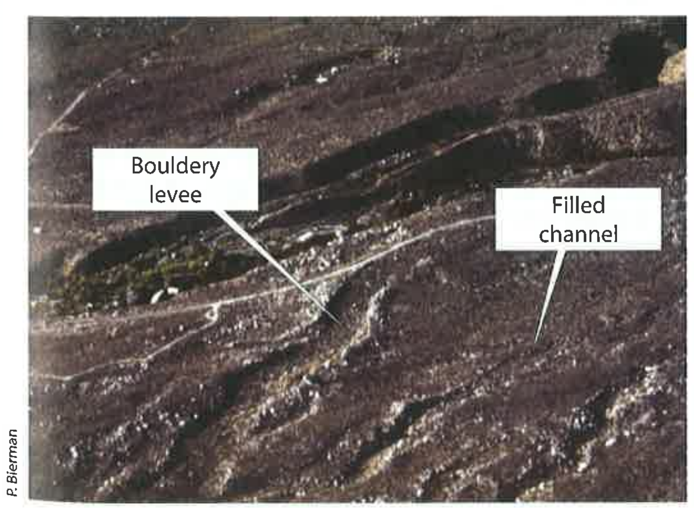
Road cuts into such fans, which are common along the White Mountains and Sierra Nevada of southern California, reveal mostly matrix-supported material in which clasts appear suspended in a finer-grained matrix (a diamicton)—a telltale signature of debris flows (see Photograph 5.14). However, there are also lenses of well-sorted sediment moved by flowing water, an indication that more than just debris flows moved sediment onto and across these fans in the past.

Development and urbanization of fans causes numerous and significant geologic hazards. Such development is widespread in the arid southwestern United States but is also occurring in humid regions, such as northeastern North America and parts of Europe. Because of the rapid hydrologic response of steep drainage basins, flash flooding on fans is common in both humid and arid regions. Channels on fans may shift episodically and unpredictably during such events through lateral migration, plugging, and **avulsion** (channel shifting). Large storms and the runoff they generate can cause sediment deposition on fans both from stream flows and debris flows. This sediment and the water that carries it can disrupt infrastructure, destroy homes, and take lives.

Lakes

Lakes are found in many different landscapes and result from a variety of solid-Earth and surficial processes. In humid regions, lakes are usually permanent landscape features on human timescales, although their levels can change in response to climate variability. In arid regions, lakes are commonly ephemeral, reflecting the balance between precipitation delivered mostly to the highlands and evaporation rates that can be extremely high in valleys where the lakes are located. Arid region lakes tend to be shallow and saline, drying to salt and mud flats known as **playas** (see Photograph 7.1).

Lakes are common in glaciated terrain. Glacial erosion can overdeepen valleys, creating bedrock basins that fill with water when the ice melts. Such **tarns** are often arranged in linear strings and are referred to as **pater noster lakes,** after the rosary beads they resemble **[Photograph 7.22]**. Moraines also dam lakes in glacial terrain. These dams can fail catastrophically, triggering massive and hazardous outburst floods, especially during rapid glacial retreat. Advancing or retreating glacial ice can dam rivers and streams, forming ice-marginal lakes that drain when the ice retreats or the lake grows deep enough that the ice-tongue dam floats (see Chapter 13, Digging Deeper). Other ice-marginal lakes formed where isostatic depression, caused by the weight of continental ice sheets, lowered land surfaces around the ice margin, creating depressions that filled with meltwater. These lakes later drained as the ice sheet melted and Earth's surface rebounded.



P. Bierman

PHOTOGRAPH 7.21 Abandoned Channels on Alluvial Fan. Here, in an aerial view of the Lone Pine Creek fan, there are many abandoned stream channels, filled in places by debris-flow lobes and lined by bouldery debris-flow levees. The channel of Lone Pine Creek is deeply incised and runs diagonally across the photograph. Two-lane road provides scale.

E. Portenga

PHOTOGRAPH 7.22 Tarns. A series of tarns, formed by glacial erosion, are visible from the summit of Sunset Peak at the Brighton Ski Resort in Utah.

PHOTOGRAPH 7.23 Crater Lake. This photo of Crater Lake, Oregon, shows Wizard Island, a small cone in the center of the lake, which occupies the caldera. The lake formed about 7700 years ago when Mount Mazama exploded, blowing off the upper 1600 meters of the volcano.

Lakes also commonly form in the craters of stratovolcanoes [**Photograph 7.23**] and in the centers of explosive calderas such as Yellowstone. Fed by orographic precipitation, these lakes become prime sources of water to feed subsequent volcanic explosions (when hot lava and lake water mix) and **lahars** (volcanic mudflows). **Maars,** craters formed explosively when magma interacts with groundwater and flashes it to steam, are often filled with water, creating lakes with extremely small drainage basins [**Photograph 7.24**]. Sediment cores from maar lakes have been used to assess geochemical inputs from the atmosphere over time because runoff contributions from the very small drainage basin (only the crater rim) are minimal.

PHOTOGRAPH 7.24 Maar. Laguna de Armenia is a maar near the San Miguel Volcano in El Salvador.

Applications

Drainage basins respond on various timescales to external forcings, including climate, tectonics, and human activity. On geologic timescales, tectonic forcing, including isostatic response to erosion, is the dominant control on the size, shape, and slope of drainage basins. On intermediate timescales, climatic forcing is important, and on decadal timescales, the influence of human actions can be dominant.

Land use affects drainage-basin dynamics, changing sediment budgets and altering geomorphological processes and zones of process dominance. Major human impacts on watersheds in Europe began with the advent of deforestation and intensive agriculture in the Bronze Age between 4000 and 5000 years ago. The clearance of forests, for agriculture and for wood, changed sediment budgets as soils began to erode faster than they were being formed—a reversal of the preceding postglacial millennia, during which soils formed more rapidly than they eroded. At a drainage-basin scale, much of the soil and sediment eroded from exposed uplands as the result of human activity never made it to main-stem river channels. Rather, it still resides on the colluvial footslopes, fills upland hollows, or builds terraces in low-order channels. The sediment that did enter the fluvial system often caused aggradation downstream, building bars and raising floodplains. The most glaring effect in the classical world of human-landscape interaction and drainage-basin response was rapid sedimentation in estuarine valleys and siltation of harbors. One famous example is the Roman port of Ostia at the mouth of the Tiber River, which is now several kilometers from the sea.

In eastern North America, New Zealand, China, parts of Canada, and elsewhere, forests have regrown after clear-cutting either naturally, after cleared land was abandoned, or where replanted as part of soil-conservation or reforestation efforts. As a result, sediment loads have diminished. In many places, rivers are incising through decades' to centuries' worth of sediment deposited when the landscape was cleared, the legacy of earlier land use and abuse [**Figure 7.11**]. Such **legacy sediments** are found nearly worldwide, and in many places form a low terrace at or just above the modern floodplain (see Digging Deeper).

Development, both land clearance for agriculture and subsequent urbanization, has specific impacts on drainage basins, changing water and sediment yields and channel behavior over time [**Figure 7.12**]. Removing forests, smoothing the land, and adding impermeable surfaces all affect the volume and speed of runoff (see Chapter 4), changing basin-scale water and sediment budgets. As impervious cover increases, floods become far more common as rainfall cannot infiltrate and greater runoff results. Flood peak magnitude increases and flood duration shortens because less water is stored in the vadose zone and on the landscape.

The removal of wood from streams, either by intentional clearance or indirectly through clearing of streamside forests and the trapping and extirpation of beavers that drop trees into streams to build dams, appears to

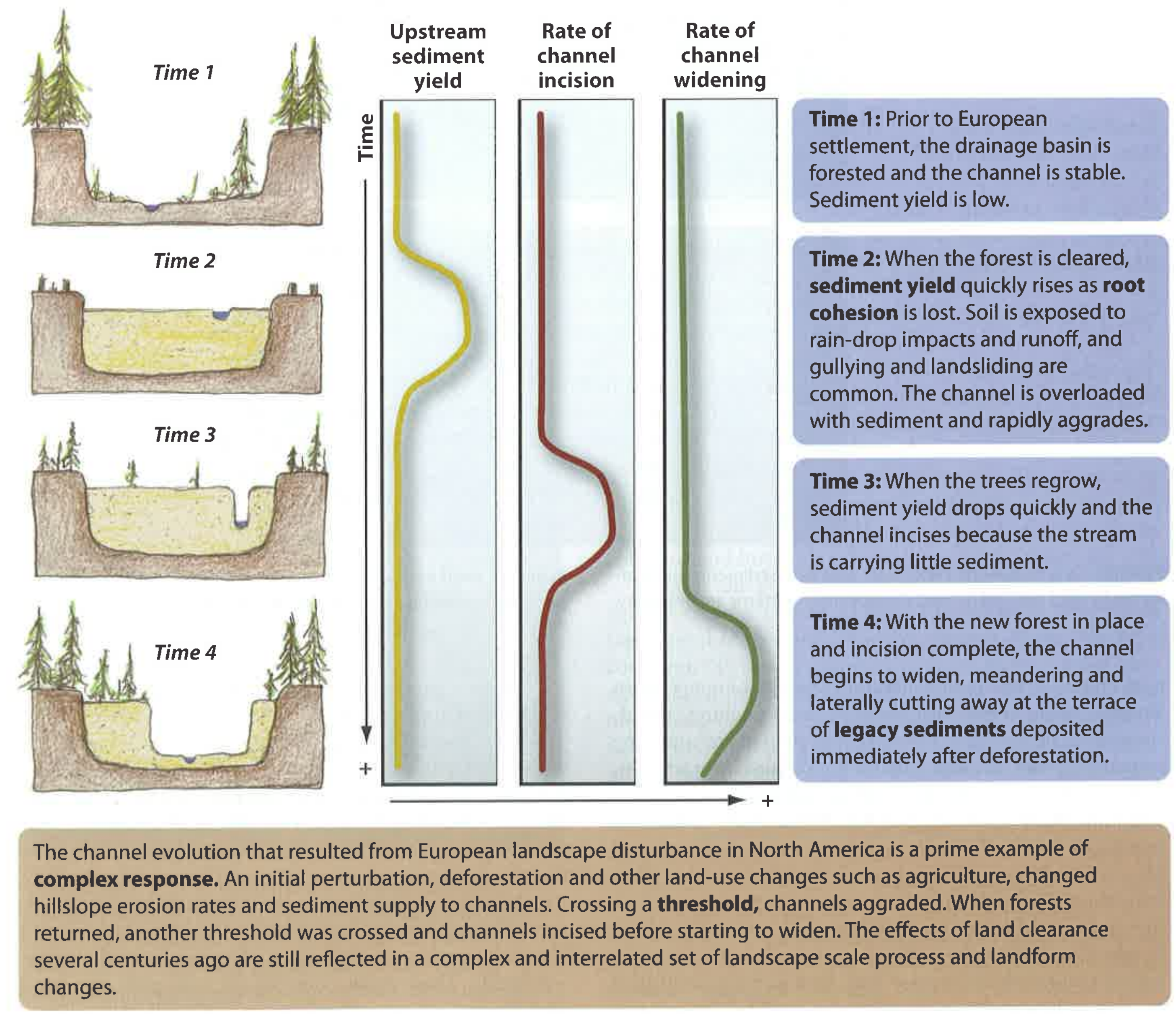


FIGURE 7.11 Channel Change. Channel change resulting from disturbance often follows a trajectory that includes aggradation, then incision followed by widening. [Adapted from Schumm and Rea (1995).]

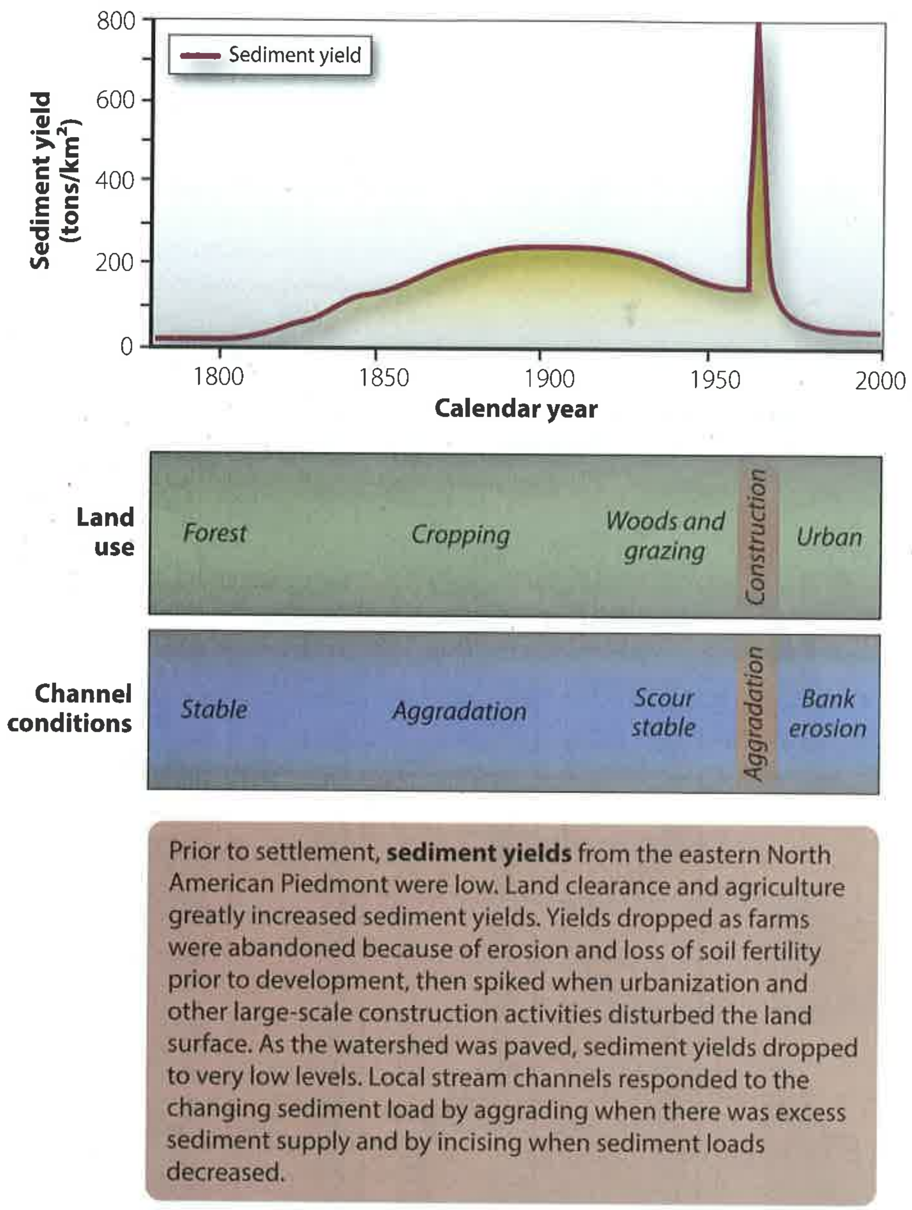


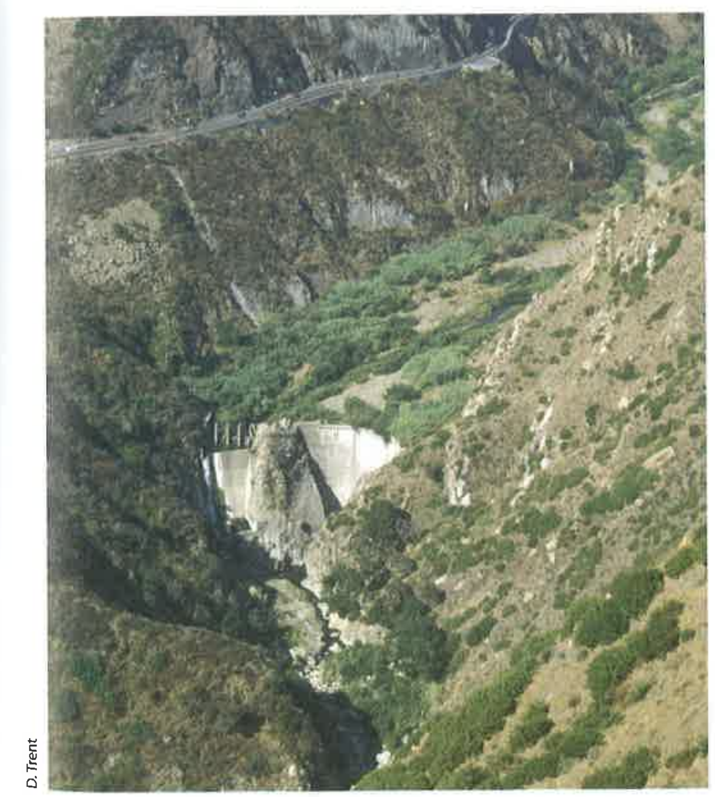
FIGURE 7.12 Sediment Yield and Land Use. Sediment yield is affected by land use and by land development patterns and intensity. Sediment yield peaks during active construction and the accompanying land-surface disturbance. [Adapted from Wolman (1967).]

have changed the fundamental morphology of many forest streams. What were once shallow, anabranching, wood-clogged channels have now become incised, meandering systems. Channel incision isolates floodplains from streams, reducing connectivity, preventing sediment deposition, and reducing the potential of floodplain water storage to attenuate floods. In cases where streams have incised deeply into legacy sediments and are isolated from their floodplains, one stream restoration strategy is to remove the legacy sediment to reconnect the stream and its floodplain. Nature does this over time, but the result of letting nature take its course may be decades to centuries of high-sediment loads downstream and failing streambanks (see Figure 7.11).

Forestry practices have significant, long-lasting, and widespread impacts on drainage-basin geomorphology. Timber harvesting affects watersheds primarily in two ways: (1) Removal of trees from basin hillslopes reduces the apparent cohesion from roots (see Chapter 5, Digging Deeper), encouraging mass movements, decreasing evapotranspiration, and raising water tables (see Chapter 4), thereby decreasing slope stability and increasing runoff. (2) Roads built to remove logs from mountain drainage basins run along or are cut into steep slopes and bridge stream channels, creating constrictions that can act as dams during floods. These roads compact the soil, increase runoff, and channel it onto potentially unstable hillsides, where it can cause gully erosion and mass wasting. Road fill failure

is also common. In areas as far apart as Oregon and New Zealand, forestry impacts on drainage basins have produced similar hillslope erosion, high sediment loads, and downstream channel aggradation. These effects are not unique to modern industrial forestry. Similar drainage basin–scale effects were observed in response to deforestation in the French Alps in the late eighteenth and early nineteenth centuries.

Dam construction and removal both have major impacts on drainage basins and their geomorphology. Where dams block river channels, they raise local base levels, triggering aggradation upstream. Over time, reservoirs fill with sediment that would otherwise have been transported downstream **[Photograph 7.25]**. Water released from dams typically carries little sediment and thus incision results downstream. Recently, efforts to remove outdated dams have increased as regulators of fisheries, hydropower, and water-dependent industries reconsider the economic and environmental impacts of dams, as well as safety concerns reflecting the age and location of older dams. Understanding and predicting the impact of dam removal



PHOTOGRAPH 7.25 Sediment Stored Behind Dam. The 30-meter-tall Rindge Dam is located in Malibu, California. The main arch of the dam was completed in 1924 but now the reservoir is completely filled with sediment from the steep, rapidly eroding watershed. Fish advocates have called for the dam's removal because it is blocking steelhead trout from accessing the upper reaches of Malibu Creek.

is a relatively new but very important area of research for geomorphologists.

Large dams that block many of the world's major rivers, such as Glen Canyon Dam in Arizona or the Three Rivers Dam in China, clearly and dramatically affect the geomorphology of the downstream river channels. Yet geomorphologists have documented that even small dams, such as those used to power colonial-era mills, can have major effects at a basin scale **[Figure 7.13]** (see Digging Deeper).

Throughout eastern and central North America, small streams and rivers were harnessed for power to cut wood, mill grain, and eventually to generate electricity. In Pennsylvania alone, fieldwork and historical documents suggest there were at least 16,000 mill dams trapping sediment and altering stream morphology at a density of about one dam every 6 or 7 km^2. Now, as many of these dams are failing or being breached, this sediment, left from years of hillslope land clearance and stored for decades to centuries, is being remobilized and is once again entering river channels.

Of particular concern is the release and subsequent transport of this sediment to Chesapeake Bay and other rich but sensitive estuaries in eastern North America. A sustained increase in fine-grained sediment delivery is injurious to estuaries because increased water turbidity decreases light transmission, smothers benthic organisms (like oysters and crabs), and carries sediment-associated nutrients like phosphorus, which cause eutrophication and unwanted algal blooms.

Channel restoration, rehabilitation, and drainage-basin management have become widespread over the past several decades as societal awareness regarding the importance of rivers as both ecological and hydrological systems has increased. Two questions come up when channel or watershed restoration or rehabilitation work is proposed; both require specific understanding of drainage-basin behavior and history.

First, is the postdevelopment hydrology of the drainage basin today similar to that of the basin at the time of the desired restoration objective, known as the **reference condition**? If runoff volumes, timing, and intensity have all changed from development, climate change, or a mix of causes, the historical characteristics of the channel before development (including its slope, width, depth, sinuosity, and pattern) may no longer be stable under contemporary conditions. Thus, it may not be possible to reconstruct the historical or predisturbance river and have such a reconstruction persist.

The second question is pertinent when the watershed has changed enough to preclude reestablishing or achieving the reference condition: How can the channel be rehabilitated so as to meet the objectives of the community in terms of aesthetics and river behavior? Managing drainage basins effectively requires a deep understanding of the complex interactions between water and sediment

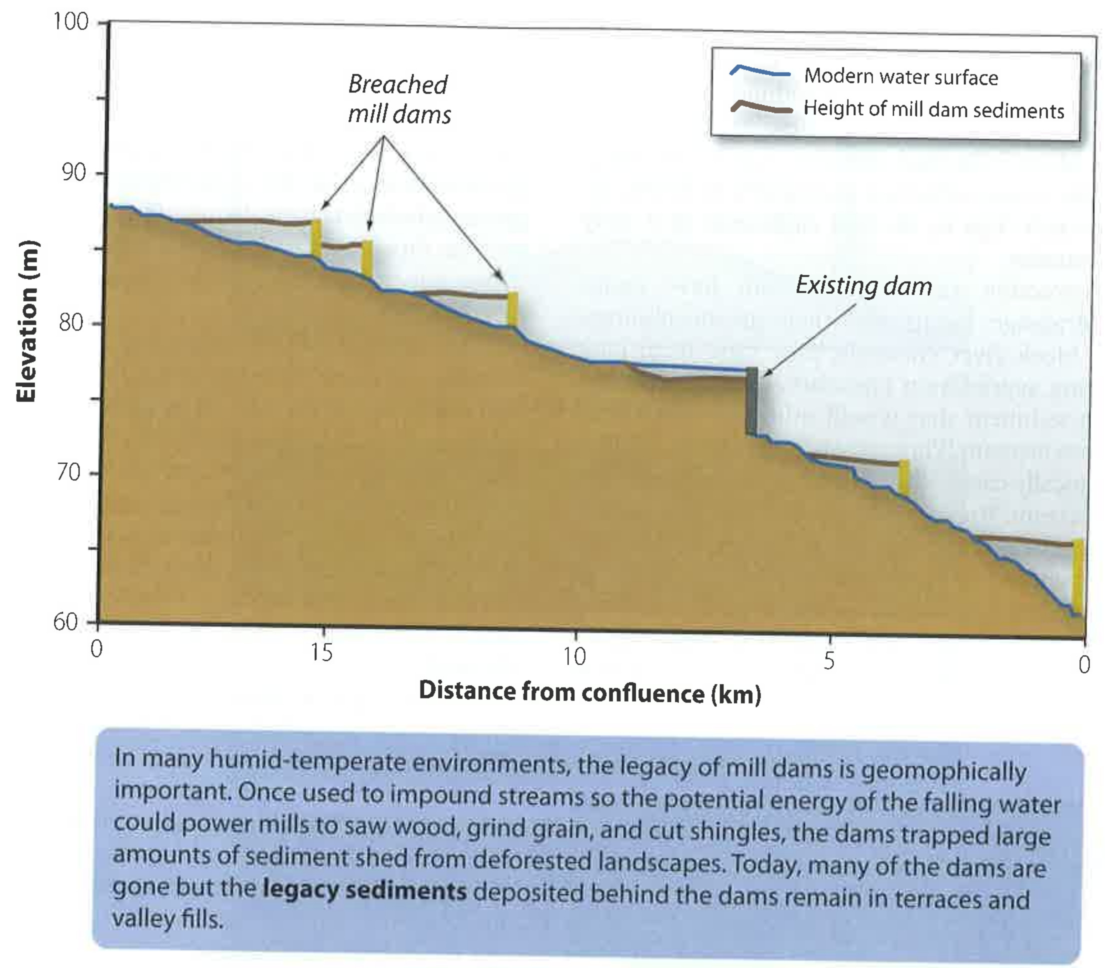


FIGURE 7.13 Mill Dam Sediments. In hilly eastern North America, mill dams and the head they created on small streams were the primary power source during and after colonial settlement. These dams, most long since abandoned, trapped sediments shed from denuded hillsides. [Adapted from Walter and Merrits (2008).]

over time. Geomorphologists, with their understanding of process, form, and landscape history, are particularly well suited to tackling this increasingly important task.

Selected References and Further Reading

Abrahams, A. D. Channel networks: A geomorphological perspective. *Water Resources Research* 20 (1984): 161–188.

Ambers, R. K. R., and B. C. Wemple. Reservoir sedimentation dynamics: Interplay and implications of human and geologic processes. *Northeastern Geology & Environmental Sciences* 30 (2008): 49–60.

Beach, T. The fate of eroded soil: Sediment sinks and sediment budgets of agrarian landscapes in southern Minnesota, 1851–1988. *Annals of the Association of American Geographers* 84 (1994): 5–28.

Bishop, P., T. B. Hoey, J. D. Jansen, and I. L. Artza. Knickpoint recession rate and catchment area: The case of uplifted rivers in Eastern Scotland. *Earth Surface Processes and Landforms* 30 (2005): 767–778.

Brummer, C. J., and D. R. Montgomery. Downstream coarsening in headwater channels. *Water Resources Research* 39 (2003): 1294, doi: 10.1029/2003 WR001981.

Burbank, D. W., J. Leland, E. Fielding, et al. Bedrock incision, rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature* 379 (1996): 505–510.

Church, M., and J. M. Ryder. Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. *Geological Society of America Bulletin* 83 (1972): 3059–3072.

Costa, J. E. Effects of agriculture on erosion and sedimentation in the Piedmont province, Maryland. *Geological Society of America Bulletin* 86 (1975): 1281–1286.

Dietrich, W. E., and T. Dunne. Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie, Supplementbände* 29 (1978): 191–206.

Gardner T. W. Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogenous material. *Geological Society of America Bulletin* 94 (1983): 664–667.

Hack, J. T. *Studies of Longitudinal Profiles in Virginia and Maryland.* U.S. Geological Survey Professional Paper 294-B. Washington, DC: Government Printing Office, 1957.

Jennings, K. V., P. R. Bierman, and J. Southon. Timing and style of deposition on humid-temperate fans, Vermont, United States. *Geological Society of America Bulletin* 115 (2003): 182–199.

Kelsey, H. M., R. Lamberson, and M. A. Madej. Stochastic model for the long-term transport of stored sediment in a river channel. *Water Resources Research* 23 (1987): 1738–1750.

Leopold, L. B. Downstream change of velocity in rivers. *American Journal of Science* 251 (1953): 606–624.

Montgomery, D. R. Process domains and the river continuum. *Journal of the American Water Resources Association* 35 (1999): 397–410.

Montgomery, D. R. Slope distributions, threshold hillslopes, and steady-state topography. *American Journal of Science* 310 (2001): 432–454.

Montgomery, D. R. Observations on the role of lithology in strath terrace formation and bedrock channel width. *American Journal of Science* 304 (2004): 454–476.

Montgomery, D. R., G. E. Grant, and K. Sullivan. Watershed analysis as a framework for implementing ecosystem management. *Journal of the American Water Resources Association* 31 (1995): 369–386.

Nichols, K. K., P. R. Bierman, M. Caffee, et al. Cosmogenically enabled sediment budgeting. *Geology* 33 (2005): 133–136.

Pazzaglia, F. J., and M. T. Brandon. Macrogeomorphic evolution of the post-Triassic Appalachian mountains determined by deconvolution of the offshore basin sedimentary record. *Basin Research* 8 (1996): 255–278.

Pratt, B., D. W. Burbank, A. M. Heimsath, and T. Ojha. Impulsive alluviation during early Holocene strengthened monsoons, central Nepal Himalaya. *Geology* 30 (2002): 911–914.

Reid, L. M., and T. Dunne. *Rapid Evaluation of Sediment Budgets.* Reiskirchen, Germany: Catena Verlag, 1996.

Reusser, L., P. Bierman, M. Pavich, et al. An episode of rapid bedrock channel incision during the last glacial cycle, measured with ^{10}Be. *American Journal of Science* 306 (2006): 69–102.

Schumm, S. A. *The Fluvial System.* New York: Wiley-Interscience, 1977.

Schumm, S. A., and D. K. Rea. Sediment yield from disturbed earth systems. *Geology* 23 (1995): 391–394.

Shreve, R. L. Statistical law of stream numbers. *Journal of Geology* 74 (1966): 17–37.

Strahler, A. N. Quantitative analysis of watershed geomorphology. *Transactions, American Geophysical Union* 38 (1957): 913–920.

Trimble, S. W. Decreased rates of alluvial sediment storage in the Coon Creek Basin, Wisconsin, 1975–93. *Science* 285 (1999): 1244–1246.

Walling, D. E., and B. W. Webb. *Erosion and Sediment Yield: Global and Regional Perspectives,* Publication no. 236. Wallingford, UK: International Association of Hydrological Sciences, 1996.

Walter, R. C., and D. J. Merritts. Natural streams and the legacy of water-powered mills. *Science* 319 (2008): 299–304.

Whipple, K. X. Bedrock rivers and the geomorphology of active orogens, *Annual Review of Earth and Planetary Science* 32 (2004):151–185.

Whipple, K. X., and T. Dunne. The influence of debris-flow rheology on fan morphology, Owens Valley, California. *Geological Society of America Bulletin* 104 (1992): 887–900.

Wolman, M. G. A cycle of sedimentation and erosion in urban river channels. *Geografiska Annaler* 49A (1967): 385–395.

Wolman, M. G., and A. P. Schick. Effects of construction on fluvial sediment, urban and suburban areas of Maryland. *Water Resources Research* 3 (1967): 451–464.

DIGGING DEEPER When Erosion Happens, Where Does the Sediment Go?

Geomorphologists commonly analyze the connection between erosion, sediment transport, and sediment export to understand the history and dynamics of particular regions or drainage basins. Early workers (e.g., Dole and Stabler, 1909) generally assumed that the volume of sediment issuing from a catchment was equal to that eroded from the catchment slopes. This style of thinking continued until scientists and engineers discovered that much of the sediment eroded by human disturbance of basin hillslopes was not making it into or through river systems. This insight was gained through creating sediment budgets, the construction of which challenged those involved to figure out where the sediment came from and at what rates it was stored within and exported from catchments. Dietrich and Dunne (1978) formalized such budgeting using a small catchment in the Oregon Coast Range. Their work pointed out the need to determine rates of processes and volumes of sediment in motion and in storage.

The **sediment delivery ratio** quantifies the ratio between the mass delivered at the drainage basin outlet and the mass eroded from basin hillslopes [**Figure DD7.1**]. For very small basins (<km^2) considered over human timescales, one-quarter to perhaps one-half of the eroded sediment makes it to the outlet. For large basins, the sediment delivery ratio on multidecadal timescales is often much lower, invalidating the assumption of equality between erosion and sediment export (Trimble, 1977). For example, by analyzing flow records, using soil-erosion models, and

DIGGING DEEPER When Erosion Happens, Where Does the Sediment Go? *(continued)*

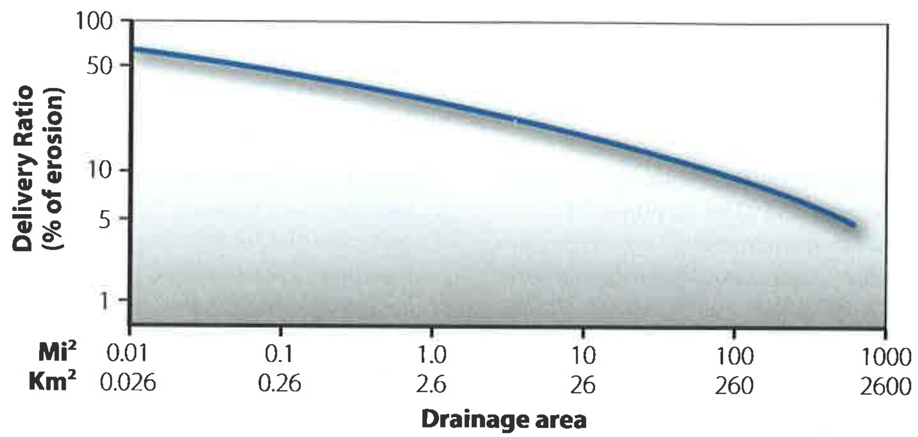


FIGURE DD7.1 The sediment delivery ratio (sediment yield/eroded mass) falls dramatically as basin area increases, suggesting the importance of sediment storage in the lower-gradient reaches of large drainage basins. Note logarithmic axes. [From Trimble (1977).]

mapping deposits in the Coon Creek Basin of Wisconsin, Trimble (1981) created sediment budgets for two time intervals since western settlement of the basin and learned that only 7 percent of the sediment generated by human impact to the basin had been exported **[Figure DD7.2]**. Where was all this recently eroded sediment hiding?

The first clues came from careful examination of soil profiles (Haggett, 1961; Trimble, 1974). Working in the piedmont of eastern North America, an area intensively cultivated for tobacco and other crops between the late 1600s and the mid-1900s, Costa (1975) noted that the upper horizons of hillslope and ridgetop soil profiles were truncated by erosion. Much of the A and O horizons was gone. On the basis of fieldwork, he and others estimated that about 15 cm of soil (on average) had been eroded from the upland during the several-hundred-years-long agricultural period. From this depth estimate, he could calculate a total volume of eroded soil. Then, working on the foot slopes and along river channels, Costa identified accumulations of colluvium and alluvium and estimated their volume. Of the upland soil eroded in his 155 km^2 study basin by colonial and post-colonial agriculture, about half was still sitting at the base of hillslopes and about 15 percent was retained in river terraces. Streams transported the rest (about 35 percent) out of the basin. The low sediment delivery ratio was now explained. Over human time frames, years to decades, a large proportion of the sediment eroded from slopes was trapped and stored in the basin and thus largely inaccessible to the river system for centuries to millennia.

Starting in the 1960s, extensive fieldwork in east-central North America led to the development of a qualitative model for river and floodplain behavior **[Figure DD7.3]** (Jacobson and Coleman, 1986). Geomorphologists suggested that after European colonists cleared the trees from much of eastern North America, rivers and their floodplains responded to the high loads of sediment moving down channels by aggrading. As agricultural practices changed and degraded farmland was abandoned and reforested, sediment loads dropped and rivers carried less sediment (see Figure 7.12). The response to lowered sediment loads was incision at first, then channel widening (see Figure 7.11). The idea was that the disturbed channels were working to reestablish the pre-settlement equilibrium conditions that included meandering planforms, extensive floodplains, and pool-riffle sequences with coarse-grained gravel bars.

Recent geologic work in east-central North America has provided an alternative explanation for the field observations. Walter and Merritts (2008) suggest that the morphology of today's streams bears little resemblance to the streams in precolonial times. They argue that the tens of thousands of mill dams constructed along the East Coast to harness water power, before the widespread use of fossil fuels began in the late 1800s **[Figure DD7.4]**, changed how sediment moved and was stored in eastern North America. Much of the sediment eroded from slopes during the colonial period remains along stream channels and in river valleys where it was trapped behind dams. Although some sediment is stored on floodplains that aggraded when sediment loads were high during the peak of deforestation and land disturbance, much of this sediment was trapped behind the mill dams, creating extensive deposits of fine-grained sediment (over coarser stream gravel) similar to the stratigraphy one would expect to result from overbank sedimentation. If the new interpretation is right, much of the human-induced overbank sedimentation of Jacobson and Coleman (1986) is now the dam-trapped legacy sediment of Walter and Merritts (2008).

Now that upland erosion has been sharply reduced by improved soil conservation practices and natural afforestation resulting from land-use change, streams are carrying less sediment, incising recent sediments, and causing accelerated bank erosion as they adapt to new boundary

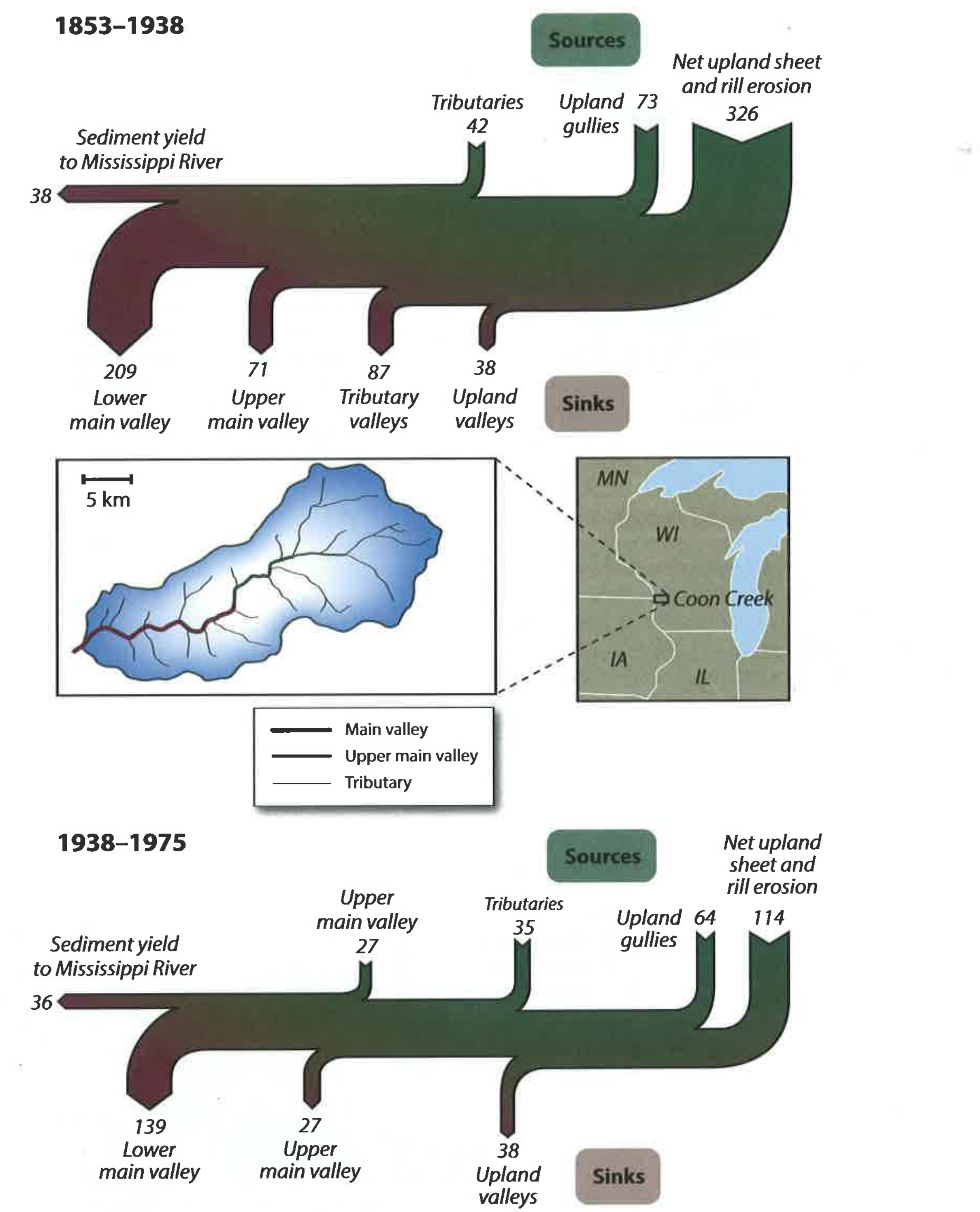


FIGURE DD7.2 Sediment budgets for the Coon Creek drainage basin in Wisconsin for periods before and after 1938. The thickness of the arrows is proportional to the sediment flux and the numbers are annual sediment volumes (10^3 m^3/year). Most of the soil that eroded during and after European settlement never made it to the Mississippi River. Sediment generated by upland sheetwash and rill erosion from cleared lands did not move far; it remains as colluvium on nearby footslopes and as alluvium in the valleys. Soil conservation practices implemented in the twentieth century have reduced the flux of sediment off hillslopes and caused a greater percentage of what was eroded to be retained as colluvium nearby. The flux of sediment to the Mississippi River has remained constant despite large changes in the Coon Creek tributary drainage basin. [From Trimble (1981 and 1999).]

DIGGING DEEPER When Erosion Happens, Where Does the Sediment Go? *(continued)*

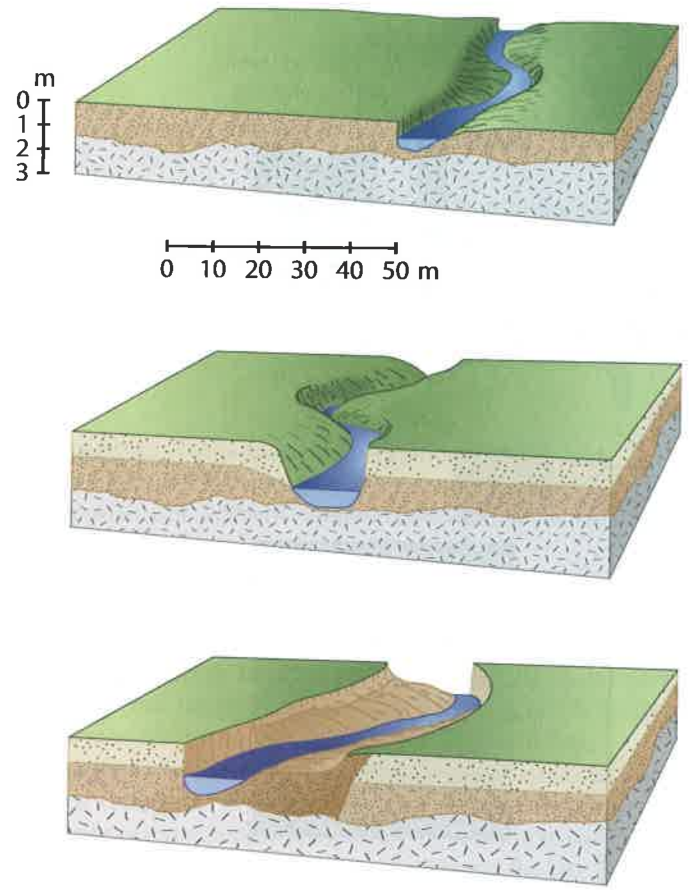


FIGURE DD7.3 Schematic model based on fieldwork in the Maryland Piedmont, showing a narrow meandering stream presettlement with extensive fine-grain, overbank deposits (upper panel). During the period of extensive farming, erosion of the uplands and high sediment loads led to floodplain aggradation (middle panel). Then during very recent time, lowered sediment loads led to incision and the establishment of the floodplain and gravel bars at a new, lower elevation (lower panel). [From Jacobson and Coleman (1986).]

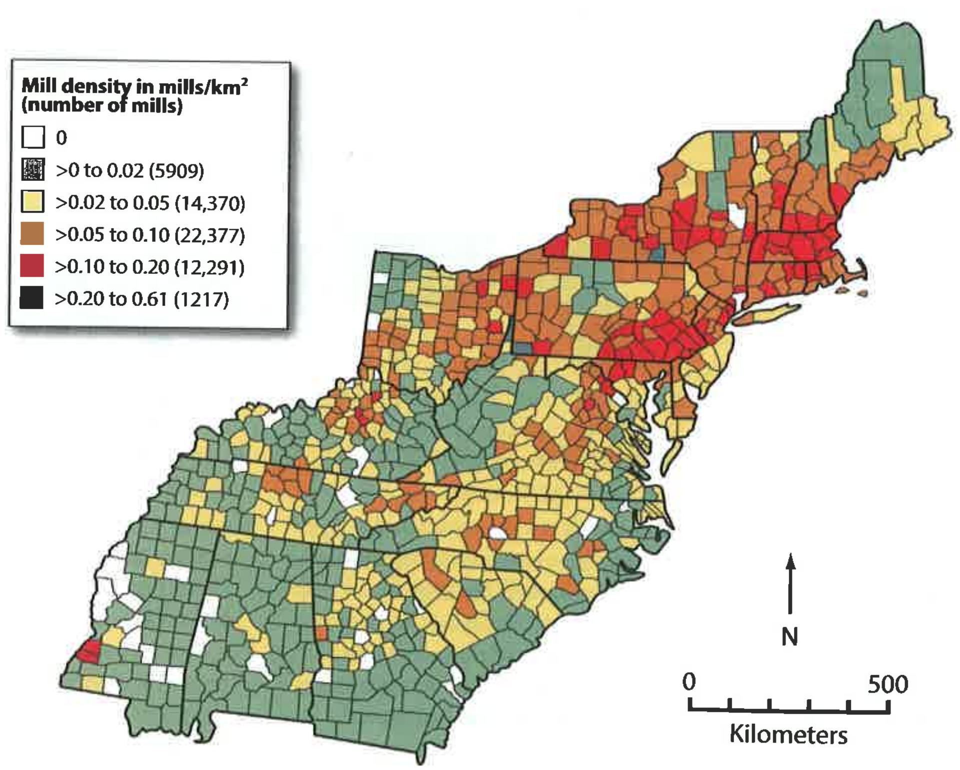


FIGURE DD7.4 This map, made from county-scale U.S. Census data, shows the density of mill dams in the eastern United States, where more than 60,000 mills and mill dams were built in the 1800s. Many areas in the Piedmont and Valley and Ridge provinces, where the density of mill dams was greatest, had one mill dam for every 5–10 km^2 of land area. [From Walter and Merritts (2008).]

conditions by widening and reestablishing floodplains at lower elevations (see Figure 7.11). Such channel responses to changes in drainage-basin land use often provide the rationale and social catalyst for stream channel restoration and rehabilitation efforts.

This change in thinking has implications for the design of projects that aim to return a stream to a reference condition that reflects fluvial process and form that existed prior to disturbance. If the Walter and Merritts (2008) model is correct [**Figure DD7.5**], then many low-order drainages in eastern North America were swampy, low-gradient, anabranching channels carrying little sediment load. This stands in sharp contrast to the gravely headwater streams with overbank sands previously thought to be representative of the pre-European settlement landscape. It also raises the question of whether society considers the characteristics of precolonial streams to be desirable from an aesthetic and environmental perspective.

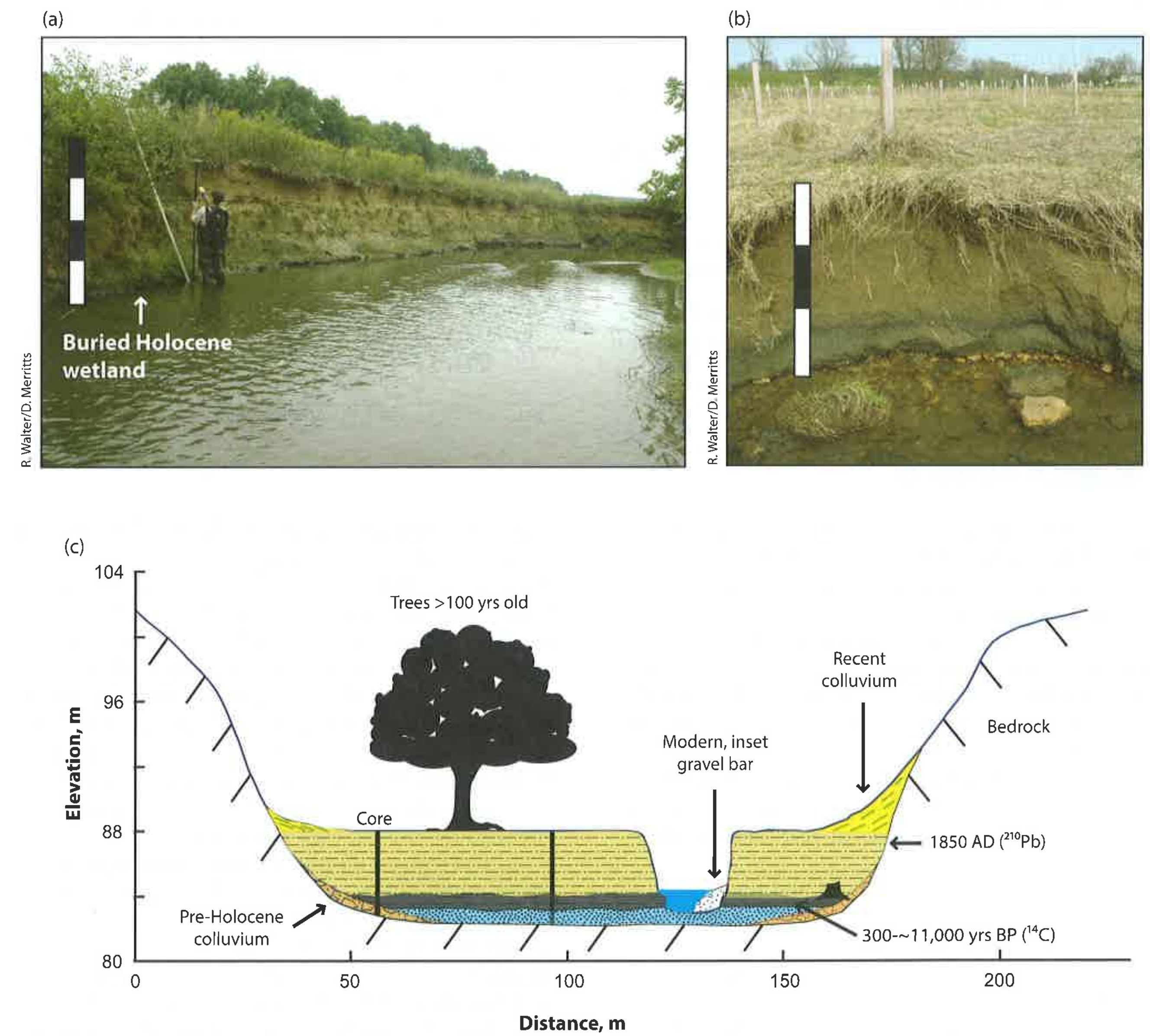


FIGURE DD7.5 Streams along eastern North America often have steep banks of tan fine-grained material overlying dark, organic-rich material, which in turn overlies gravel and bedrock. Photo (a) is from Western Run, Maryland, and photo (b) is from Big Spring Run, Pennsylvania. The scale bars have 0.5-m divisions. Walter and Merritts (2008) propose that fine-grain terraces were deposited in still water caused by mill-pond damming of streams and thus reflect historical land use. They conclude that presettlement streams were not incised into floodplains but rather that the streams carried very little sediment as they moved through wet, marshy valley bottoms (c) Idealized cross section. [From Walter and Merritts (2008).]

DIGGING DEEPER When Erosion Happens, Where Does the Sediment Go? *(continued)*

Understanding the roles of legacy sediment and the present-day response of channels to drainage-basin disturbance over the past centuries has substantial—and usually underappreciated—implications for channel restoration and river-corridor management (Renwick and Rakovan, 2010). In particular, local stream-restoration and rehabilitation projects that do not consider historical changes in the drainage-basin hydrology, sediment supply, and sediment delivery are likely to fail rapidly because the channels they create are not appropriate for contemporary drainage basin conditions.

Costa, J. E. Effects of agriculture on erosion and sedimentation in the Piedmont province, Maryland. *Geological Society of America Bulletin* 86 (1975): 1281–1286.

Dietrich, W. E., and T. Dunne. Sediment budget for a small catchment in mountainous terrain. *Zeitschrift für Geomorphologie Supplementbände* 29 (1978): 191–206.

Dole, R. B., and H. Stabler. Denudation. *United States Geological Survey Water Supply Paper* 234 (1909): 78–93.

Haggett, P. Land use and sediment yield in an old plantation tract of the Serra Do Mar, Brazil. *The Geographical Journal* 127 (1961): 50–59.

Jacobson, R. B., and D. J. Coleman. Stratigraphy and recent evolution of Maryland Piedmont flood plains. *American Journal of Science* 286 (1986): 617–637.

Renwick, W. H., and M. T. Rakovan. Sediment supply limitation and stream restoration. *Journal of Soil and Water Conservation* 65 (2010): 67a.

Trimble, S. W. *Man-induced Soil Erosion on the Southern Piedmont, 1700–1970.* Ankeny, Iowa: Soil Conservation Society of America, 1974.

Trimble, S. W. The fallacy of stream equilibrium in contemporary denudation studies. *American Journal of Science* 277 (1977): 876–887.

Trimble, S. W. Changes in sediment storage in the Coon Creek Basin, Driftless Area, Wisconsin, 1853 to 1975. *Science* 214 (1981): 181–183.

Trimble, S. W. Decreased rates of alluvial sediment storage in the Coon Creek Basin, Wisconsin, 1975–93. *Science* 285 (1999): 1244–1246.

Walter, R. C., and D. J. Merritts. Natural streams and the legacy of water-powered mills. *Science* 319 (2008): 299–304.

WORKED PROBLEM

Question: Describe a conceptual sediment budget (using a flow diagram like that shown in Figure DD7.2) to begin the process of quantification for (1) a nonglaciated 100 km^2 drainage basin in a humid-temperate region, and (2) the same basin in an arid region. First, consider sediment sources and the processes by which sediment is created and delivered to the river network. Then consider areas in the basin where sediment is stored. Finally, consider the rate at which sediment is exported and the processes by which sediment is removed from the basin. Ensure that your sediment budget considers both long and short timescales (thousands of years and decades).

Answer: Sediment budgets for arid and humid regions are distinct because the active geomorphic processes in these regions differ.

In humid regions, weathering converts rock to regolith that is readily eroded to become sediment. Most slopes in humid regions are transport-limited and covered with a mantle of soil. The source terms in a humid-region sediment budget include rock weathering rates and the diffusive transport rates of sediment downslope and advective rates of sediment transport in gullies, rills, and in mass movements. Some sediment is stored in colluvial deposits at the base of hillslopes, some is stored in fans, and other sediment is stored along river and stream valleys as terraces and in floodplains. Sediment is exported from the basin by streams primarily as suspended load during large flow events, such as storms and snowmelt floods. Over thousands of years, the rate at which sediment is supplied from hillslopes is similar to the rate at which it is exported from the watershed. At decadal timescales, these rates are often quite different, with sediment being retained at the base of slopes or along channels and exported by rare but high-magnitude floods.

In arid regions, weathering and sediment production occur more slowly and most slopes are weathering limited and have little if any soil cover. Streams are ephemeral and usually dry. Rare flows can move large amounts of sediment in short times. Source terms in arid regions include advective transport of material off rock slopes by overland flow from gentle slopes, rock fall from steep slopes, and movement of sediment from other parts of the basin (and from other basins) by wind.

In areas where the rocks are weak and there is tectonic activity raising ranges above the valley floors—for example, around Death Valley, California—large amounts of sediment are stored in fans at the base of steep slopes. In contrast, tectonically stable areas underlain by hard rocks produce little sediment and fans are mostly absent. Depending on the tectonic setting, little if any sediment may leave arid-region watersheds; rather, the sediment may accumulate in closed basins slowly filling them, or some material may be carried out of the basin by wind. Because arid-region processes are so episodic, rates of surface processes over short timescales may bear little if any resemblance to rates over longer timescales.

KNOWLEDGE ASSESSMENT Chapter 7

1. Define a drainage basin.
2. Where in drainage basins does sediment tend to be deposited?
3. Where in drainage basins does sediment most likely originate?
4. Describe the difference between open and closed drainage basins.
5. In what climate and tectonic setting are you most likely to find closed drainage basins?
6. Describe three specific challenges geomorphologists face in creating a sediment budget.
7. Define a sediment rating curve and describe how sediment rating curves are created.
8. Explain why data in sediment rating curves are so variable.
9. Sketch four common types of drainage patterns and suggest a location where each type might be found.
10. Explain the factors leading to each of the four common drainage patterns.
11. Draw a series of sketches to illustrate the difference between superposed and antecedent drainages.
12. Describe the issues related to defining a "first-order" channel.
13. Compare and contrast the Shreve and the Strahler stream ordering classifications.
14. Predict how channel width, depth, and cross-sectional area change downstream in a humid-temperate river network.
15. Why and how does discharge change downstream in an arid-region river network?
16. Describe how both average basin slope and channel slope change as a function of basin area.
17. Where is stream velocity greater—in a mountain cascade or a large lowland river? Explain your answer.
18. Discuss the residence time of sediment in a drainage basin.
19. Explain why fine-grained and coarse-grained sediment have different average residence times in humid-temperate drainage basins.
20. List the four valley segment types and describe the dominant processes in each.
21. Sketch a river longitudinal profile and explain (giving three reasons) why the slope of the longitudinal profile changes in the downstream direction.
22. List three different base levels and describe what processes might lead them to change over time.
23. Define both a knickpoint and a knickzone; explain how they are different and what can cause them to occur.
24. Explain how and why floodplain morphology changes downstream.
25. Predict how and explain why the grain size of sediment carried by a river changes downstream.
26. Explain the difference between depositional and strath terraces.
27. List three reasons why rivers leave behind terraces.
28. Predict where alluvial and debris fans are most likely to be found.
29. List several characteristics that would help you differentiate between alluvial and debris fans.
30. Explain the feedback mechanism that encourages gorge formation and deepening.
31. At what timescales do tectonic, climatic, and human forcings affect drainage basins?
32. Define pater noster lakes and explain how they form.
33. Give two examples of climate changes that can affect drainage basins.
34. Draw a diagram showing how sediment yield might change over time as forests are cleared and land is developed.
35. Define "legacy sediment."
36. Explain the effects of forestry practices on drainage basins.
37. Define a reference condition and argue whether or not it is a valid concept.