Geomorphic Hydrology

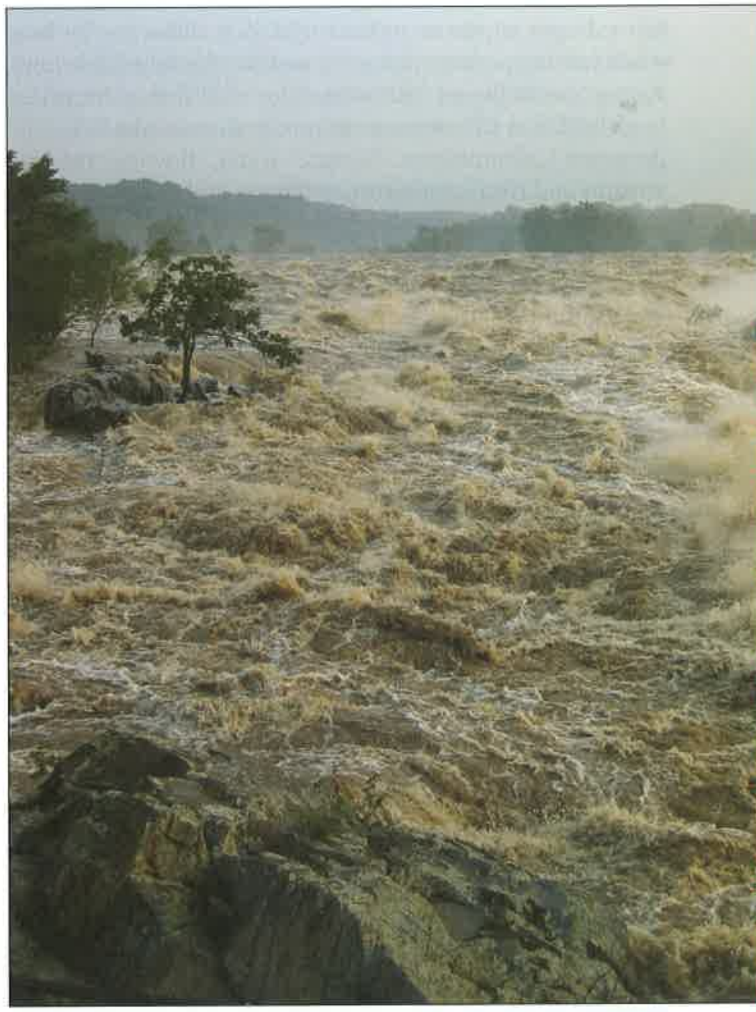
4

Introduction

Without water, Earth's surface would be a vastly different, inhospitable place—rather like present-day Mars. Sediment would move only when meteorites struck, stirring broken rock at the planet's surface, or when sand-charged winds blasted the land. There would be little weathering, no rivers, no lakes, and no streams. Our planet would be dry and barren, devoid of plants and animal life. But Earth is not a dry planet; we live on a planet of water—about 70 percent of its surface is covered by oceans. The temperature range on Earth appears to be unique in our solar system, allowing water to exist at the surface in three phases (liquid, solid, and vapor) simultaneously. All three phases are geomorphically important. Ice and liquid water are effective mass transport agents. Water vapor is also a critical part of the **hydrologic cycle.**

Much of the water that falls on Earth's surface as precipitation eventually ends up in streams and rivers as it moves toward the ocean. Water is constantly on the move, evaporating from the ocean, entering the atmosphere, and then precipitating back onto the land. Once on land, water moves under the influence of gravity (and inertia), flowing down rivers and percolating through the ground until, once again, it enters the atmosphere or the sea. Plants short-circuit this cycle, using solar energy to pump water directly back into the atmosphere through the process of **transpiration,** the uptake of water through roots and the release of water vapor to the atmosphere through leaves.

Most landscapes are shaped either directly or indirectly by water. Understanding how water moves and where it is



P. Bierman

Turbid floodwaters fill the bedrock channel of the Potomac River just upstream of Washington, DC, after the passage of Hurricane Isabel in 2003. More than 4500 cubic meters per second of water are flowing over Great Falls, which is completely submerged.

IN THIS CHAPTER

found are fundamental to understanding how Earth's surface changes. Raindrops dislodge soil grains, sending material downslope in hillslope runoff. Runoff eventually collects, slowly incising valleys or rapidly eroding disturbed surfaces like construction sites and plowed agricultural fields. **Groundwater,** flowing slowly through surficial deposits or deeper in porous or fractured rock, dissolves elements while leaving pore space, caves, and weathered rock behind. Rising groundwater can destabilize hillslopes, triggering landslides that can degrade mountain slopes, take lives, and devastate communities. Surface water, flowing through streams and rivers, transports sediment and shapes channels used for both recreation and transport.

Meltwater moving through and under glaciers sustains river flow through dry summer months, providing clean and abundant drinking water. Yet, when glacial dams burst, areas downstream can be devastated and cubic kilometers of freshwater pour into the world's oceans. Ancient catastrophic megafloods have raised sea level and altered global climate many times in the past. Hydrologic processes have shaped landscapes around the world and the hydrologic response to climate change affects society and may do so with increasing intensity over the coming century.

The geomorphic influence of water on Earth's surface is by no means uniform. In hyperarid regions, such as the deserts of southern Africa, rain may fall only once a decade and channels may flow only several times a century. When rain does fall, massive amounts of sediment can move because most of the rainfall generates runoff and there is little vegetation to hold the landscape together. Conversely, in warm humid climates, abundant water supports dense forests that largely shield the ground surface from the immediate hydrologic effects of heavy precipitation [**Photograph 4.1**]. At the same time, beneath the surface, continually moist conditions and organic acids generated by decaying vegetation dismember rocks and weaken underlying slopes. Reduced erosion and greater weathering promote soil development.

Both the amount of water and the rate at which it is supplied to the landscape are important when trying to understand the geomorphological behavior and ecological diversity of Earth's surface. In addition, temperature is of critical importance because it determines where and when liquid and frozen precipitation will fall and greatly affects the rate at which water-mineral weathering reactions take place. The same amount of water will do more weathering in the hot tropics than if it falls as precipitation in a cooler climate. In this chapter, we follow the water as it moves across, into, and through the landscape and consider how it shapes the planet on which we live.

Precipitation

Precipitation is a key link in the **hydrologic cycle,** as it returns water from the atmosphere to the land surface (see Figure 1.7). Water can fall from the sky either as rain or as various forms of frozen precipitation, including snow. Frozen precipitation is mostly a winter phenomenon (except at high latitudes and altitudes), and in spring and summer, the ice and snow melt and move as liquid water over, into, and through the ground.

Precipitation has long been measured using rain gauges, but such gauges are rare; on average, there is about one official gauge per 1000 km^2 in the United States. Furthermore, dense networks of such gauges are required for accurate delivery estimates because precipitation amounts can vary greatly over the scale of kilometers. In the 1990s, advanced weather radars, called NEXRAD (Next Generation Weather Radar), were installed across the United States. These systems, when used in concert with interpretive mathematical algorithms, allow forecasters to estimate rainfall intensity remotely and at high resolution (km^2). Other satellite-based observations are being used to estimate precipitation amounts around the world and, in some cases, drive geomorphic models.

(a)

P. Bierman

(b)

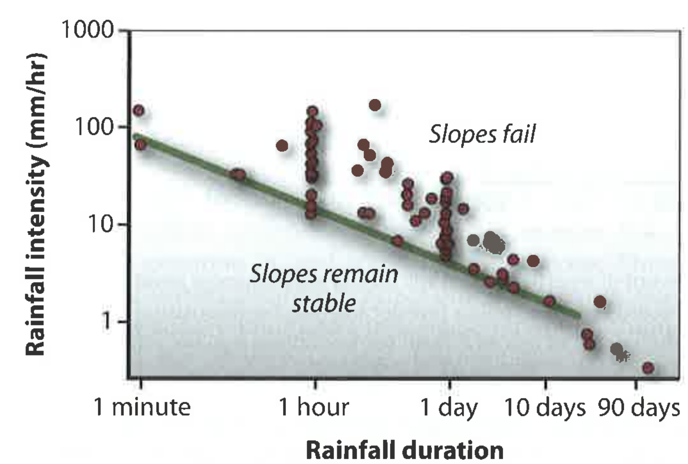
A. Codilean

PHOTOGRAPH 4.1 Water Shapes Earth's Surface. The amount of water on the landscape dramatically affects its appearance. (a) Water-sculpted bedrock channel flowing off the wet tropical escarpment in Queensland, Australia. This area receives around 2 m of rain yearly. (b) The Gaub River in Namibia receives only several centimeters of precipitation a year and both the channel and surrounding area are largely devoid of vegetation. Most of the year, the stream channel is dry.

Duration and Intensity

Precipitation events have two fundamental characteristics: **duration** (how long the storm lasted) and **intensity** (the volume of water that came down, expressed as length per unit time, such as mm/hr). Both of these characteristics are geomorphically important. Long-duration events can saturate the ground, thereby triggering landslides and raising river levels. High-intensity events also trigger landslides and can deliver large volumes of water to landscapes so quickly that some of it runs off, eroding soils and carving rills or deep gullies.

Commonly, precipitation is only geomorphically effective when an event exceeds a combined intensity-duration **threshold** that triggers a discrete change—like a landslide. For example, landslides that produce downstream mudflows can be triggered by specific combinations of rainfall intensity and duration [**Figure 4.1**]—heavy rains lasting a short time or less intense rains that last a long time. If the combination of intensity and duration does not exceed the threshold, slopes will remain stable. Note that the data underlying these threshold curves can be quite variable—the curves describe the average behavior of a landscape. If the threshold values are exceeded, some hillslopes will fail and others will not. While it is important to consider that individual hillslopes behave quite differently, once the threshold is exceeded, more rainfall would be expected to cause more landslides in more locations across the landscape.



Landslides that trigger debris flows are a geomorphic process that occurs when a **threshold** is exceeded. In this case, rainfall is the trigger, saturating the ground and causing slopes to fail. Both rainfall intensity and duration, not just cumulative rainfall, are important in determining slope stability. The combinations of rainfall intensities and durations that trigger slope failure vary regionally, reflecting differences in slope strength, topography, and near-surface hydrology.

FIGURE 4.1 Intensity-Duration Thresholds for Geomorphic Processes. Many geomorphic processes are triggered only when a force or process, such as rainfall intensity and duration, exceeds certain limits. [Adapted from Caine (1980).]

Recurrence Intervals

Experience tells us that extreme events, such as long-duration or intense storms, are rare and that short, low-intensity events are more common. This relationship can be quantified as a **recurrence interval,** the probability of an event occurring (e.g., rainfall over a 24-hour period) based on the average amount of time that passes between events of at least a certain magnitude. Calculating a recurrence interval (RI) for an event of a particular **magnitude** (size) requires many observations (N) of different annual maximum events (the largest event in each year) collected over time. Each annual maximum event in the record is assigned a rank (r) based on its magnitude. The largest event is given a rank of 1. The smallest event is given a rank of N. Rank and recurrence interval (in years) are inversely related:

$$RI = (N + 1)/r \qquad \text{eq. 4.1}$$

Large, rare events have long recurrence intervals and small, common events have short recurrence intervals. In other words, events with long recurrence intervals have a low annual probability of recurring while those with short recurrence intervals occur much more frequently [**Figure 4.2**]. The probability, p, of an event (a rainstorm, flood, or landslide) occurring in any one year is the inverse of its recurrence interval:

$$p = 1/RI \qquad \text{eq. 4.2}$$

Although the recurrence interval formulation is simple, its application is fraught with assumptions and misconceptions. Chief among the latter is that a 100-year event is something that only happens once every hundred years. The 100-year event actually describes an event the size of which has a 1-in-100 chance of occurring in any given year (1 percent chance per year). Similarly, the 50-year event has a 1-in-50 chance of occurring in any year (2 percent chance per year), and so on for the 1-in-25-year (4 percent chance per year) and 1-in-10-year events (10 percent chance per year).

Estimates of event likelihood often require extrapolating beyond the period of record because most stream or precipitation gauges have only a few decades of data. Although estimating recurrence intervals based on short records results in great uncertainty, it is routinely done. Anyone reporting that an area has been devastated by the 500-year storm (i.e., a storm which on average occurs about once every 500 years), when only 30 years of weather data

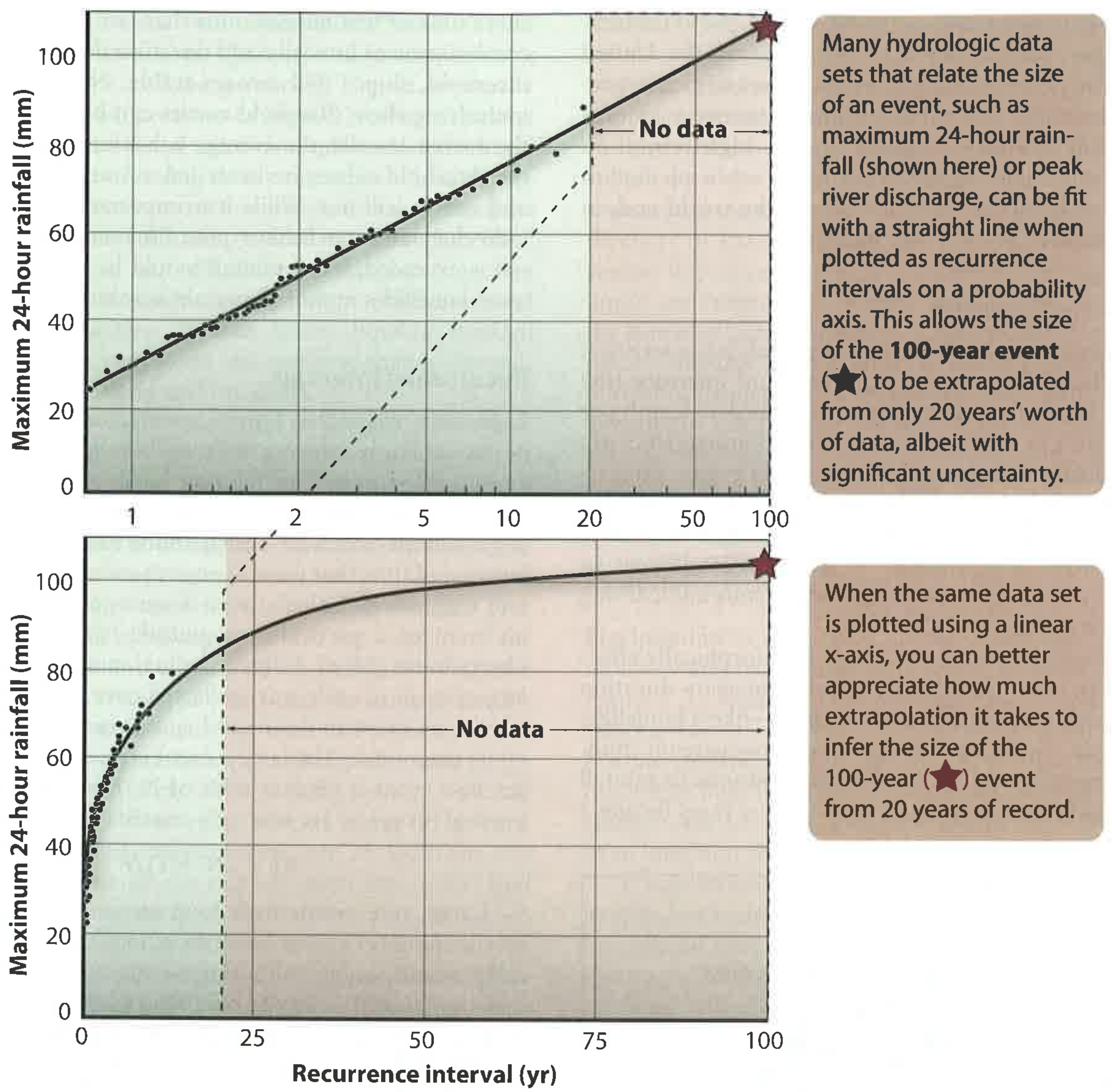


FIGURE 4.2 Precipitation Recurrence Interval with Extrapolation. Extreme hydrologic events are rare. Small events are common. Probability plots are a useful means of estimating how often an event of a particular size will recur. However, most periods of record are short, only a few decades, and extrapolation beyond the period of record is both commonplace and uncertain.

exist, is extrapolating. Fundamental to the application of recurrence intervals is the idea that weather events are random, implying there is no long-term trend in the weather and thus no trend in climate over time. This assumption, termed **stationarity**, is growing less valid given the reality of climate change. Although the forecasts of global climate models differ in details, all predict significant rises in global average temperatures and changes in the distribution, frequency, and amount of precipitation over the next century—a time frame relevant both to geomorphology and human interactions with Earth's dynamic surface.

Precipitation Delivery

Precipitation is delivered to Earth's surface in discrete weather events—each having unique geomorphic impacts because of differing spatial extent and intensity [Photograph 4.2]. For example, many areas near the coast at low and middle latitudes can be affected in summer and fall by intense storms born in the tropics; they are called **hurricanes** in the Atlantic Ocean, **typhoons** in the Pacific and Indian oceans, and **cyclones** in Australasia. These tropical low-pressure systems can rapidly release large amounts of precipitation and trigger floods over thousands of square kilometers.

Some storms are large and move slowly. In northwestern North America, cold, mid-latitude winter storms moving in from the Pacific Ocean are often long-lasting, gently soaking temperate rainforests with low-intensity rainfall for days on end but leaving little geomorphic evidence of their passage. In eastern North America, sprawling nor'easters draw in Atlantic Ocean moisture and can pile snow meters deep in the Appalachian Mountains while battering the Atlantic coast with rain and high surf for days on end. Compare such large, regional weather systems with the intense summer thunderstorms common in arid and semi-arid regions. These convective storms commonly affect only small areas, perhaps a few square kilometers in a single mountain drainage basin, but deliver intense downpours capable of quickly triggering debris flows and devastating flash floods. At an intermediate scale, frontal lifting of moist air can cause condensation and precipitation over large areas. **Warm fronts** tend to produce soaking rains of low intensity that can last many hours. **Cold fronts** are sharper boundaries, producing high rainfall rates and strong winds that can be geomorphically effective, even though they last for only a few hours.

Mountains and large lakes can create local weather. Lake effect snows caused by cold air moving over relatively warm water can quickly bury downwind locations, such as Buffalo, New York, on the eastern shore of Lake Erie, under a meter of snow. Mountains force air masses to rise, wringing out precipitation as the air cools and water condenses. This **orographic** effect can be significant. Burlington, Vermont, and Seattle, Washington, both receive on average a little less than a meter of precipitation each year. The Green and Cascade Mountains, ranges just to the east of these cities, receive 2 and 3 times the mean annual precipitation of the adjacent lowlands, respectively.

(a)

NOAA

(b)

A. Matmon

(c)

P. Bierman

PHOTOGRAPH 4.2 Storms. Storms come in a variety of shapes and sizes. (a) Hurricanes impact southern and eastern North America every summer and fall; this is Hurricane Katrina, August 2005. (b) Small but intense convective thunderstorms pour down rain on isolated desert valleys in Namibia. (c) Heavy rain from large, mid-latitude, low-pressure system triggered small landslide along a road near Nelson, New Zealand.

In some areas, fog and clouds contribute significant amounts of moisture to the water budget, water that is critical for rock weathering and the survival of life. For example, in hyperarid coastal Namibia, more water is delivered to the land surface by frequent and persistent coastal fogs than by rainfall, which occurs only rarely. The large amount of fog-derived condensation in the Namib Desert may be responsible for enhanced salt weathering of rock surfaces, resulting in rock erosion rates comparable to those observed in topographically similar areas of Australia that receive 10 times more rainfall. Similarly, at high elevation and along humid, temperate coastal zones, large amounts of water may be delivered to trees by condensation from clouds and fog and then flow down stems and tree trunks to the ground. Conversely, leaves, particularly the broad leaves of deciduous trees, can intercept precipitation. If the precipitation is of low intensity and short duration, it will never reach the ground but rather will evaporate directly from the leaves. The effect is greatly lessened for deciduous trees during winter when leaves are off; however, conifers can intercept and retain snowfall on their leaves and branches. Some of this snowfall can sublimate and never reach the ground.

Climate Effects on Hydrology and Geomorphology

Climate affects hydrology, and thus the processes that shape Earth's dynamic surface, in a variety of ways. The primary climatic control on hydrology is the amount and seasonal distribution of precipitation. But the primary hydrologic control on geomorphology is the frequency and timing of high-magnitude events such as spring floods or hurricanes. The rest of the year, precipitation and the weathering that it catalyzes prepares material on hillslopes for evacuation by the big storm events.

In arid regions, rainfall is sporadic and spatially disjunct, often affecting only small parts of a drainage basin as discrete storm cells move across the landscape. Many arid-region storm cells are small relative to the size of drainage basins. Thus, specific geomorphic processes (flooding, landslides, debris flows) may be active in one subcatchment or even one part of a subcatchment while neighboring subcatchments remain unaffected. In humid regions, precipitation events are more frequent, but rain gauge data also show significant spatial variability that helps explain observed spatial differences in active geomorphic processes and process rates. Rainfall data clearly show that the highest rainfall intensities typically cover only small portions of the landscape.

Some climates are strongly seasonal and that seasonality is reflected in regional hydrology and geomorphology. **Monsoonal** climates have warm, wet summers and cooler, drier winters. These climates result from the geographic juxtaposition of an elevated landmass and an adjacent tropical ocean. The landmass is heated by the Sun during the summer, causing the air above it to rise and entrain moisture from the ocean. Strong lift causes the moist air to condense and produce particularly intense rainfall for several months each year. River and stream channel dimensions are adjusted to this regime of heavy rain and significant runoff. The months after monsoon rains pass may be quite dry. During the dry season, channels appear grossly oversized for the small flows they carry. The Himalaya, northern Australia, and the Colorado Plateau are all affected by summer monsoons.

Cold regions, such as the Arctic and Antarctic, are characterized by long periods of hydrologic inactivity punctuated by periods where water comes out of frozen storage and runoff is dramatic. Spring snowmelt quickly releases water from storage in the snowpack and 30 to 50 percent of the yearly runoff can leave the landscape in just a few weeks. Similar to monsoonal climates, channels in cold regions are shaped by and sized for flows that occur during only part of the year.

Evapotranspiration

Much of the water delivered to Earth's surface by precipitation returns directly to the atmosphere without ever entering a stream or river. The portion of the precipitation that is evaporated and transpired does not contribute to runoff.

Evapotranspiration Rates

Evapotranspiration is a broad term describing the flux of moisture into the atmosphere both through the physical process of **evaporation** and the biological process of **transpiration** [**Figure 4.3**]. **Evaporation** removes water directly from Earth's surface and from the stems and leaves of vegetation. **Transpiration** is distinguished from evaporation by the active role that plants play in both facilitating water transfer to the atmosphere and pulling water from the soil. The rate of evaporation is controlled by temperature, the relative humidity of the air, wind speed, and ultimately by the intensity of solar radiation. These same factors, although biologically driven, control transpiration. The effects of temperature on both evaporation and transpiration are indirect, through temperature's effect on relative humidity.

Evapotranspiration rates vary over space and time. Rates are high during the summer and during daytime hours when the input of solar radiation is greatest. Rates of evapotranspiration drop at night in the summer and for deciduous trees in winter, when the leaves are absent and photosynthesis is minimal. In contrast, coniferous trees, which keep their leaves all winter, continue to transpire water all year long. The importance of winter transpiration by conifers is confirmed by dozens of studies showing that the increase in **water yield** (the amount of water leaving the basin) following removal of conifer forests substantially exceeds that when hardwood forests are removed.

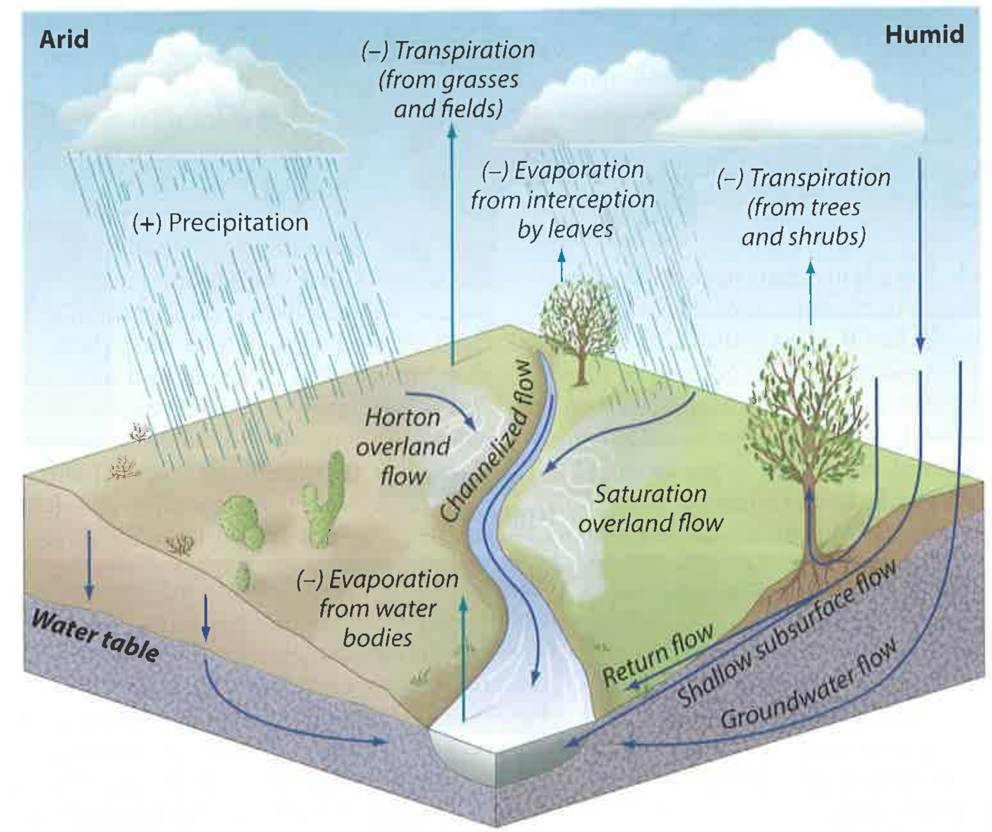
Actual Versus Potential Evapotranspiration

It is critical to make the distinction between actual and potential rates of evapotranspiration (ET) [**Figure 4.4**]. Low and mid-latitude deserts have very high potential rates of ET because of low relative humidity and large inputs of solar energy. However, actual rates of ET are low in arid regions because little water is available to evaporate or transpire. Conversely, in humid regions, where rainfall is common and soils are often moist, actual and potential rates of ET are more closely matched. Actual ET is difficult to measure and is often calculated on a yearly basis as the difference between water supplied to a watershed by precipitation and that running off through rivers. The implicit assumption underlying this calculation—that the flux of groundwater is small by comparison—is reasonable in most terrain.

Geomorphic Importance of Evapotranspiration

Evapotranspiration can be a geomorphically important process, removing large amounts of water from near Earth's surface. In many North American drainage basins, more than half the precipitation is returned to the atmosphere by ET. Movement of water from the soil to the atmosphere by transpiring plants is sufficient to reduce soil moisture levels and draw down **water tables** (the level below the surface where void spaces within soil, in rock fractures, or between sediment particles are completely filled with extractable water, also known as the **phreatic zone**) during the growing season. Trees exert a direct control on water levels in streams by reducing stream flow during storm events (**peak flow**) and flow between storms (**baseflow**). Diminished baseflow results from water table lowering, which occurs during extended periods of evapotranspiration. Reduced flood peaks result from reduced soil moisture levels because tree roots remove water held by **capillary** forces, or surface tension, from the unsaturated zone. Stream flow records in forested, temperate regions show baseflow discharge and groundwater table

Water, falling as precipitation, takes various paths on, below, and away from Earth's surface. Some precipitation never reaches the ground because it is **intercepted** and evaporates from vegetation. Some water **infiltrates** and flows through the ground, some runs off the land surface, and some is returned to the atmosphere by the pumping action of vegetation, **transpiration.**



In arid regions, the groundwater table beneath hillslopes is deep, and precipitation mainly comes as high-intensity, short-duration storms. Because the dry climate supports scant vegetation, little precipitation is intercepted. **Infiltration rates** can be low, especially for A-horizons composed of wind-blown dust and clay. Large amounts of **Horton overland flow** occur because the rate of precipitation exceeds that of infiltration. Runoff also occurs through direct precipitation onto stream channels.

In humid regions, the water table can be close to the surface. Large amounts of vegetation intercept precipitation, make soil more permeable, and transpire water. Where and when the **water table** intercepts the ground surface, there is **return flow** and rainfall generates **saturation overland flow.** Some precipitation infiltrates and moves through the shallow, permeable, near-surface soil as **shallow subsurface flow** before emerging as return flow and entering streams. Precipitation that infiltrates deeply becomes **groundwater.**

FIGURE 4.3 Schematic Diagram of Evapotranspiration and Flowpaths. Water falling as precipitation can take many different paths. Some water is intercepted by vegetation and evaporates before it ever reaches the ground. Other water is taken up by plants and transpired back to the atmosphere. Much of the remaining water either runs off or enters the groundwater system.

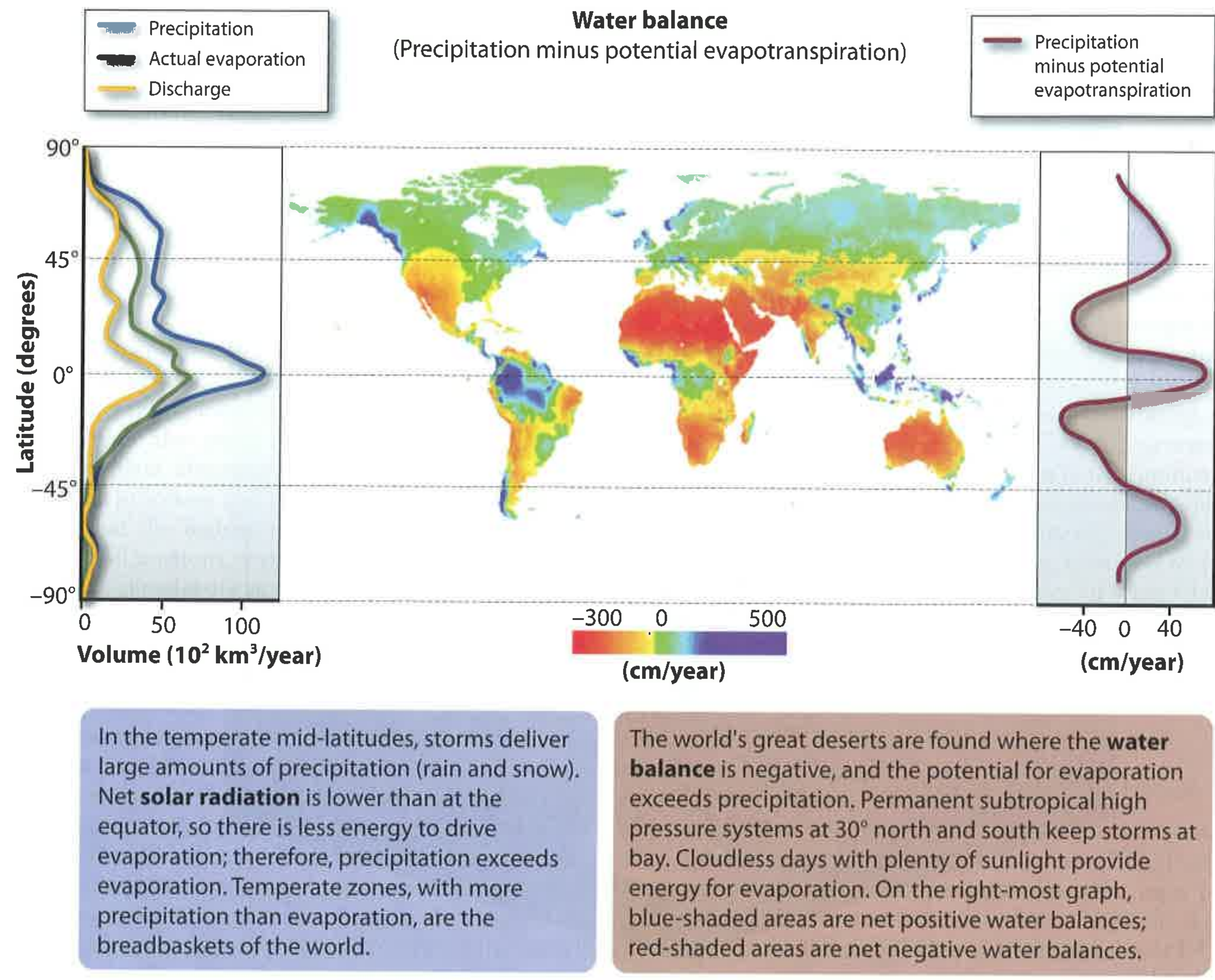


FIGURE 4.4 Water Balance by Latitude. Water balance varies greatly over Earth's surface. Over deserts, potential rates of evapotranspiration far exceed available precipitation. In temperate high and low latitudes, there is an excess of water. The volume of precipitation, evaporation, and discharge are controlled by the location of land areas and by high temperatures near the equator. Differences in land area lead to asymmetry between the northern and southern hemispheres.

elevations increasing in the fall when deciduous trees stop pumping water.

The power of trees to pump water from the ground has led people to deforest watersheds in regions where water is scarce so that water previously transpired by trees would be available for human use. Such modifications have been done in New Zealand and South Africa, as well as in arid southwestern, humid southeastern, and wet northwestern North America. Indeed, **water yields** went up by hundreds of millimeters a year in some cleared watersheds.

There are significant geomorphic implications to such ecosystem manipulation. Erosion, landsliding, and the flux of sediment tend to increase in deforested watersheds because along with removal of the trees comes loss of the binding effect of roots that otherwise help hold hillslope soils in place. Rising water tables on cleared hillslopes have the potential to further exacerbate landslide and erosion hazards. Virtually every paired-watershed study with more than a decade of data postharvesting shows that increases in water yields following vegetation removal are ephemeral, declining to pretreatment levels generally within a decade after cutting as vegetation regrows. Thus, to maintain the increased water yield, the landscape must be kept open and unforested, increasing the likelihood and intensity of erosion.

Groundwater Hydrology

Groundwater, although unseen, maintains baseflows in streams and rivers and has important geomorphic effects, including weathering of rocks. Soils saturated with groundwater are more likely to fail in mass movements. The degree and extent of soil saturation by groundwater influence surface-water flowpaths and the type and distribution of vegetation, fundamental controls on the rate and distribution of surface erosion.

Infiltration: Moving Water into the Ground

Soil, because it covers much of Earth's surface, is the critical link between precipitation, runoff, and geomorphology. The linkage is expressed by a soil hydrologic characteristic termed the **infiltration rate.** Infiltration rates describe how quickly water moves into soils and are calculated as a volume infiltrated over an area per unit time; often, they are expressed in units of length per unit time such as mm/hr (do the unit analysis and you will see that the two units are equivalent). Soil infiltration rates are critical values because they determine the fate of precipitation—how much rainfall or snowmelt soaks into the ground and how much runs over the land surface.

Not all soils are similar, and infiltration rates vary widely among different materials (Table 4.1). Compacted, clay-rich soils might infiltrate only a millimeter or two per hour while loose, sandy soils could easily pass 30 times more water in the same time interval. Initially, dry soils infiltrate water rapidly as water fills empty pores [Figure 4.5]. As a rainstorm continues and the soil wets up or saturates, the rate of infiltration falls and then levels off at a steady value. In some clay-rich soils, apparent infiltration rates decline as clays hydrate (take on water) and swell, closing cracks and reducing the rate at which water can enter the soil.

Biological activity dramatically affects soil infiltration rates. Earthworms and burrowing animals loosen soil and create large open spaces in the soil, known as **macropores**, some of which are centimeters to decimeters across. Similarly, plant roots disturb soils. When plants die and their roots begin to decompose, the cavities left behind provide additional conduits through which water can easily move. Macropores intersecting the surface increase infiltration rates.

Human activity tends to decrease the rate of infiltration. Both agriculture and urbanization typically reduce or remove the permeable, organic-rich cover typical of naturally vegetated areas. Soil is easily and quickly compacted by only a few passes of foot or vehicle traffic, both of which can reduce infiltration capacity by an order of

TABLE 4.1

Typical infiltration rates

Material	Infiltration rate (mm/hr)
Clay soil	1–5
Clay loam soil	5–10
Loam soil	10–20
Sandy loam soil	20–30
Sandy soil	>30

From United Nations Food and Agriculture Organization, http://www.fao.org/docrep/S8684E/S8684E00.htm

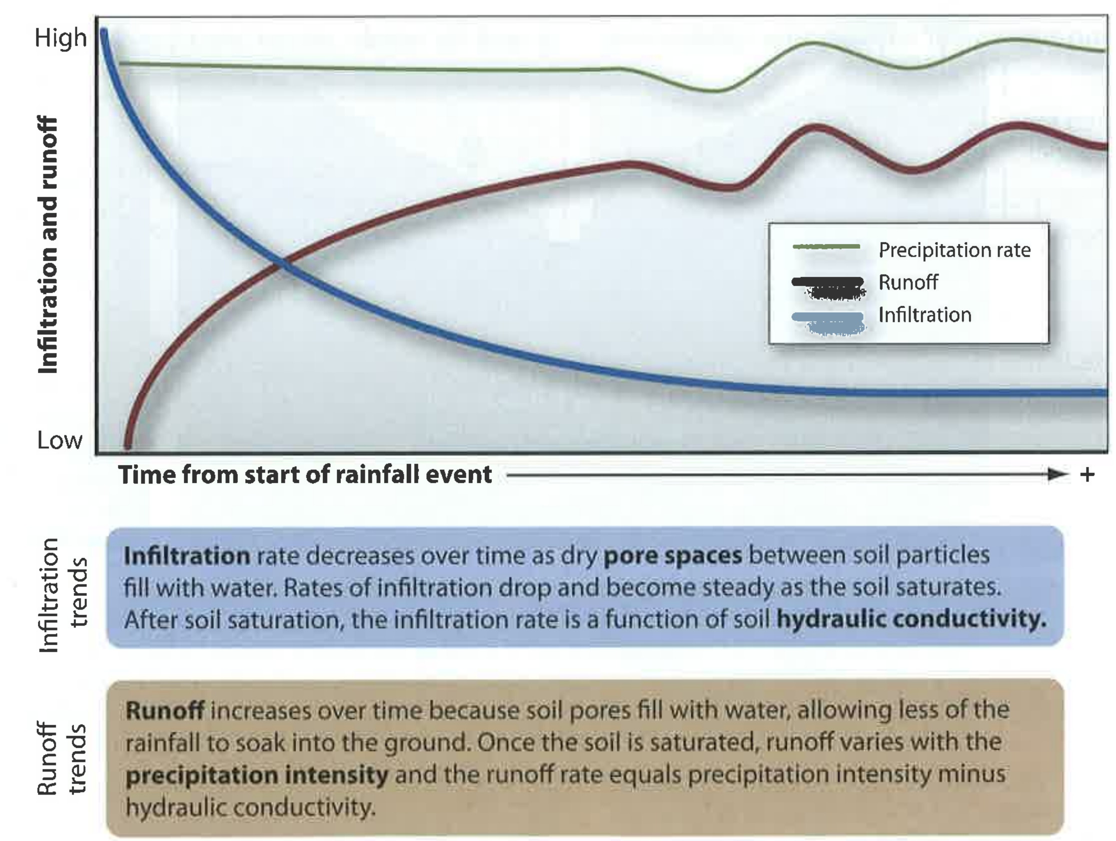


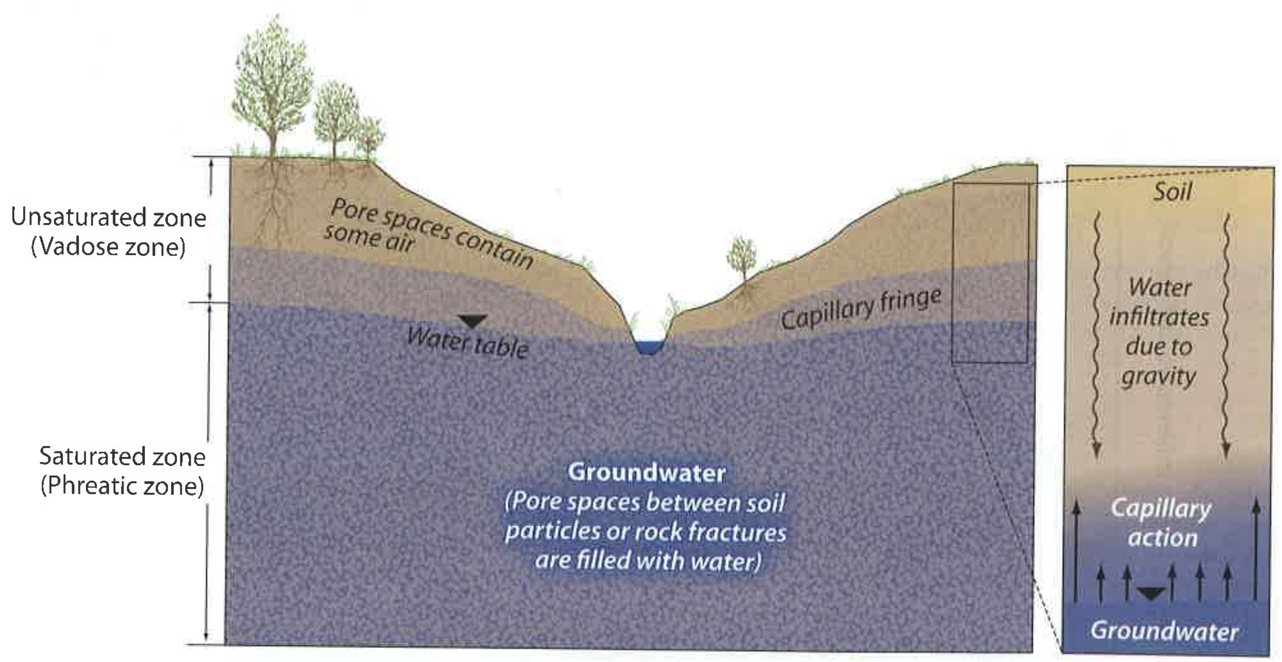
FIGURE 4.5 Infiltration and Runoff over Time. When rainfall begins, infiltration rates are high. As the ground saturates, infiltration rates drop off, and as soon as the rate of precipitation exceeds the rate of infiltration, runoff begins.

magnitude or more. One only needs to hike in a rainstorm to see water running down a compacted trail; less than a meter away, precipitation will be infiltrating into the leaf-covered mineral soil [Photograph 4.3].

Moving Water Through Earth Materials

Infiltrating water begins what could be a long or a short journey through the solid Earth and its cover of loose, **unconsolidated** materials. As water enters the ground, it moves first through the unsaturated or **vadose zone** defined by the presence of both air and water in the pore spaces between grains. There, water movement is controlled by the force of gravity, which promotes soil drainage, and the **capillary force** that holds water between grain surfaces. After sufficient water infiltrates the soil, the gravitational force overcomes the capillary force and the water moves deeper into the soil. At some depth below the surface, all the pore spaces become filled with water. If a well were drilled and lined with a perforated pipe, water would flow freely from the soil into the pipe and come to rest at a certain level; this is the **water table** [Figure 4.6]. The area just above the water table, termed the **capillary fringe**, contains enough water held in tension between grains to saturate pores. But in this zone just above the water table, the water is held tightly enough by capillary forces that it is not free to drain. The capillary effect is most significant in fine-grain sediments.

PHOTOGRAPH 4.3 Overland Flow. Rainwater, unable to infiltrate into the compacted soil of an informal footpath, runs off the University of Vermont campus in Burlington, Vermont. This is an example of Horton Overland Flow—rainfall intensity exceeding the infiltration rate of the compacted soil.



Below the surface, Earth materials can be classified by the degree to which they are saturated by water. In the **unsaturated** or **vadose** zone, pore spaces between grains contain at least some air. In the **saturated** or **phreatic** zone, all pore spaces are water-filled. Pore spaces in the **capillary fringe** are water-filled; however, the water is held tightly to and between grains by capillary forces and cannot freely drain. This diagram shows a typical situation in a humid region where rainfall is frequent.

FIGURE 4.6 Water Table and Definitions. The water table marks the boundary between zones where the pores in rock and sediment are saturated with water that can freely drain (the phreatic zone) and the vadose or unsaturated zone. The water level in the stream reflects the local water table.

The concept of **hydraulic head,** a measure of the total energy of water at a point (the sum of its potential and pressure energy), is particularly important in geomorphology, because it defines the energy gradient and thus determines the flow direction of groundwater. Head can be measured in reference to a variety of different datums, including the local ground surface, sea level, and geologic contacts [Figure 4.7]. Water moves from locations of high head to locations of lower head, down the energy gradient. A familiar example is how water flows into a well when the well is pumped and the level of water in the well (its head) drops beneath that of the adjacent groundwater, causing water to flow into and refill the well.

The ability of Earth materials to pass water depends on two physical properties, **porosity** and **permeability** (**Tables 4.2** and **4.3**). Porosity quantifies the amount of a material that is void space, open volumes that water could potentially occupy. **Primary porosity** is a function of the pore sizes within the material itself. Porosity can also be secondary, the result of biological, chemical, and physical processes altering soil and sediment. Biologically-induced, **secondary porosity** includes soil openings created by roots and animal burrows. Joints and fractures create secondary porosity through physical means; chemical dissolution can enlarge fractures, opening water-bearing conduits, particularly in soluble rocks such as limestone. The amount of groundwater moved in primary and secondary porosity varies widely and depends on the type of material. In coarse gravel, most flow passes through the primary pores. In well-developed karst terrain, almost all flow is through secondary conduits, fractures widened by dissolution of bedrock.

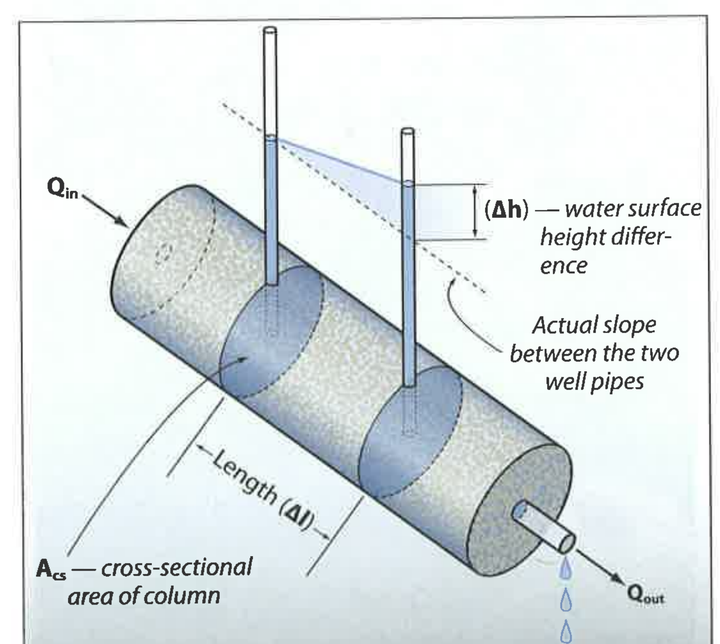
TABLE 4.2

Typical porosity of selected Earth materials

Material	Porosity (%)
Soil	30–50
Gravel and sand	20–35
Clay	45–55
Sandstone	15
Shale or limestone	5
Granite	1

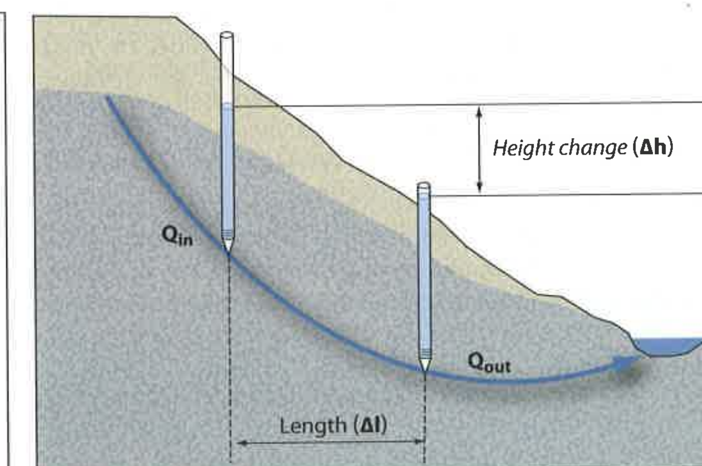
From http://geology.er.usgs.gov/eespteam/brass/aquifers/aquifersintro2.htm

Permeability describes the ability of a fluid to move through any porous medium, including soil, sediment, and rock, and is a way to quantify the integrated effect of the size of pores through which the fluid moves and the degree to which these pores are interconnected. When the fluid is water, the permeability is expressed in terms of **hydraulic conductivity** (see Table 4.3), a value that incorporates both the **density** and **viscosity,** or resistance to flow, of water. Hydraulically conductive or permeable soils allow large amounts of rainfall and/or snowmelt to quickly infiltrate or soak into and then move through the ground, sustaining groundwater flow and minimizing surface runoff. Low permeability soils generate more rapid runoff at or near



$$Q_{out} = KA_{cs}(\Delta h/\Delta l)$$

K = Hydraulic conductivity of the soil
A_{cs} = Cross-sectional area of the soil column
$\Delta h/\Delta l$ = Hydraulic gradient or head difference



Darcy's Law describes the flow, Q, of a fluid, in this case water, through **porous material** as a function of the energy or **hydraulic gradient** and the ability of the material to pass water, the **hydraulic conductivity.**

FIGURE 4.7 Darcy's Law. Darcy's Law explains the flux of groundwater as a function of energy gradient (change in head), the cross-sectional area through which the groundwater flows, and the ability of earth materials to pass water (their hydraulic conductivity).

TABLE 4.3

Saturated Hydraulic Conductivity (cm/s)

	10^{-11}	10^{-10}	10^{-9}	10^{-8}	10^{-7}	10^{-6}	10^{-5}	10^{-4}	10^{-3}	10^{-2}	10^{-1}	10^{0}	10^{1}	10^{2}
Permeability	Impervious					Semipervious				Pervious				
Aquifer	None				Poor				Good					
Soils	Clay					Very fine sand, silt, loess, and loam			Clean sand or sand and gravel			Clean gravel		
Rocks	Granite		Limestone, dolomite		Sandstone		Oil-bearing rocks			Fractured rock, karst				

Adapted from Bear, J., 1972, Dynamics of fluids in porous media, Dover, New York.

the surface than do high permeability soils, promoting erosion and starting the movement of sediment downslope. Permeability and porosity are not always directly related. For example, some soils or rocks have high porosity but low permeability because the pores, although numerous, are not well interconnected.

Once infiltrating water reaches the water table, it becomes part of the groundwater system and its movement can be described quantitatively by **Darcy's Law**, a mathematical formulation proposed by a French engineer in the nineteenth century (see Figure 4.7). Based on a series of experiments, this expression relates the volumetric rate of fluid flow (Q) through a porous medium (like sand) to the cross-sectional area (A_{cs}), the material's hydraulic conductivity (K), and the energy or head gradient ($\Delta h/\Delta l$) available to move the fluid (the hydraulic head difference divided by the distance between the observation points):

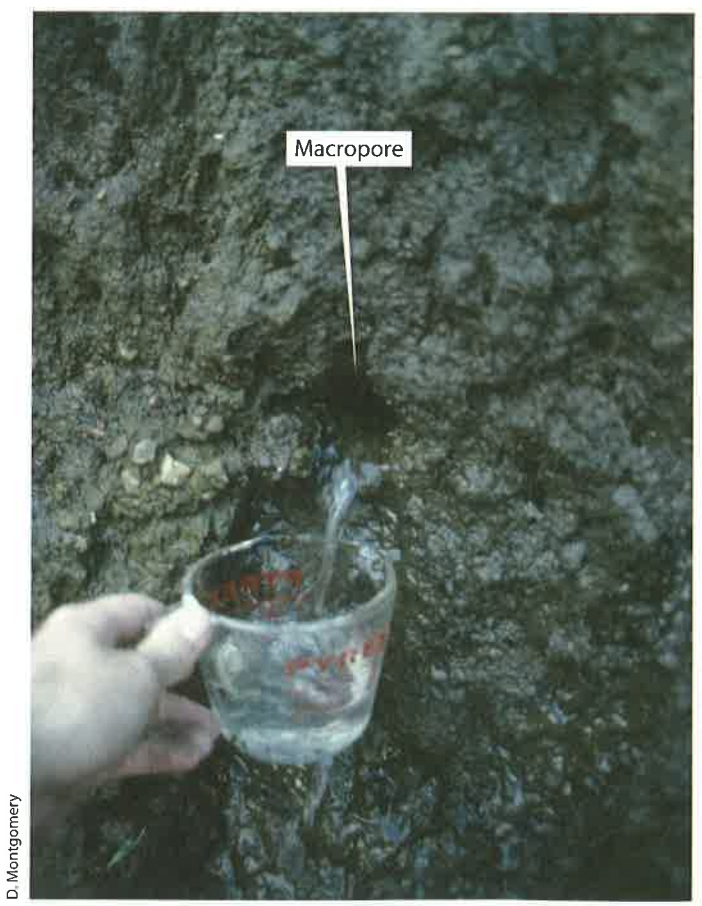
$$Q = A_{cs} K \Delta h/\Delta l \qquad \text{eq. 4.3}$$

In other words, the groundwater discharge (Q) equals the product of the cross-sectional area (A_{cs}), the ability to pass water (K, the hydraulic conductivity), and the hydraulic gradient ($\Delta h/\Delta l$) driving the flow.

Examining Darcy's Law, it is clear that large volumes of groundwater can move rapidly through highly permeable material (with large hydraulic conductivity) since K and Q are directly related. Steep hydraulic gradients (large change in h for a unit change in l), such as those formed when the groundwater table is steepened just after a large storm, will also move large amounts of water through the ground. Note that the rate at which rainwater steadily infiltrates into surface soil after the initial wetting period is also governed by the saturated hydraulic conductivity of the subsurface material (see Figure 4.5). Hydraulic conductivities are derived experimentally either in the lab, or more usefully in the field; they range over many orders of magnitude (see Table 4.3) and depend on void size, including joint dimensions, the space between grains, and the size and connectivity of **macropores** (large voids) [**Photograph 4.4**]. Head gradients are measured in the field by comparing the level to which water rises in standpipes and monitoring wells (see Figure 4.7).

Hydrologic Flowpaths

Water reaching Earth's surface can follow a variety of **flowpaths**—the route water takes down the hydraulic gradient. Different flowpaths have different geomorphic implications (see Figure 4.3). Water flowing over the ground



PHOTOGRAPH 4.4 Macropores. Macropore flow through otherwise low-permeability soil in Tennessee Valley, California.

surface moves rapidly, offering the potential to erode and deposit sediment, thereby doing geomorphic work and shaping landforms. Subsurface flow is considerably slower but can help weather subsurface materials, weakening regolith over much longer time frames. Rock weathering lowers hillslope strength over time and can change hillslope hydrology by concentrating flow in what are already high permeability zones of weakness. There are substantial and important feedbacks here. For example, more effective subsurface flow can increase rates of rock weathering, further increasing the permeability of preferential flowpaths.

Much of what we know about flowpaths has been learned using a combination of physical field experiments, chemical and isotopic tracers, and numerical models. Physically based field experiments typically involve placing instruments on a hillslope to understand the distribution of flow over time at various depths and locations. Chemical tracing of flowpaths can involve both natural and artificially introduced tracers, including major elements such as chlorine; noble gases including helium, neon, argon, and krypton; and chlorofluorocarbons (CFCs) introduced into the atmosphere by industrial activity. Ratios of stable isotopes ($^{16}O/^{18}O$, $^{1}H/^{2}H$) and the abundance of radioactive isotopes such as ^{3}H (tritium) in water have been used to trace the source, flowpaths, and age of groundwater.

When rainfall intensity exceeds the soil infiltration rate, water runs over the land surface. Runoff generated in this fashion is termed **Horton overland flow (HOF)**, after the hydrologist Robert Horton who first studied this process. HOF is common in arid regions where precipitation intensity often exceeds the infiltration capacity of low permeability, clay-rich soils, and on disturbed ground where the pounding of boots, hooves, or vehicles have lowered infiltration rates. In humid temperate regions, HOF is rare because dense vegetation and its root zones promote high infiltration rates. HOF in humid regions is largely restricted to areas where development or disturbance have compacted soils (see Photograph 4.3).

Most precipitation infiltrates and begins to move through the shallow subsurface in response to gravity. In the unsaturated zone, water can flow through connected pores such as animal burrows or root casts, spilling out farther downslope if the pore encounters the surface (see Photograph 4.4). Much water moves laterally in the subsurface along flowpaths largely controlled by contrasts in hydraulic conductivity, such as the interface between the looser, more permeable soil near the ground surface and more clay-rich, less permeable lower soil horizons or the interface between the soil or surficial deposits and underlying, solid bedrock. This **shallow subsurface flow** or **interflow** is largely disconnected from deep groundwater flow and is often ephemeral because the wetting front moves downward only temporarily after storms as water percolates into the soil.

Some of the near-surface groundwater moves downward and recharges deep groundwater flow systems where it can remain isolated from Earth's surface for days to millennia. Groundwater can discharge into rivers, which function as drains if they are lower than the water table along their banks. This groundwater discharge sets the **baseflow** that supports aquatic life between rainstorms. Fossil groundwater, recharged during wetter glacial times before the last glacial maximum more than 20,000 years ago, is today mined in the central United States, pumped from **aquifers** such as the famous Ogallala, and is used to irrigate the dry lands of the western Great Plains. The rate of water removal is much greater than the current recharge rate. Continued pumping of this and other aquifers has led to land-level **subsidence** as pore spaces dewatered and collapsed.

Saturation overland flow is generated when shallow subsurface flow returns to the surface, such as at the base of a hillslope, or where rainfall on already saturated soil cannot infiltrate and runs off. Saturation overland flow is common just after snowmelt or winter rainstorms, when water tables are high [**Photographs 4.5 and 4.6**]. The

PHOTOGRAPH 4.5 Seaonally Saturated Ground. Snowmelt saturates the ground in Greensboro, Vermont. A rapid April melt due to strong, moist southerly winds followed heavy warm-frontal rains, saturated the ground and caused water to run over the pastures. The wet areas, which are low sections of the landscape, will be dry by late spring.

PHOTOGRAPH 4.6 Saturation Overland Flow. Saturation overland flow on grassy surface after a March rainstorm in Tennessee Valley, California.

observation that saturated areas change seasonally led to the **variable source area** concept, the idea that source areas for rapid runoff expand and contract over time. In wet seasons, large amounts of runoff may reach the channel by saturation overland flow whereas during drier times of the year, most rainfall infiltrates and moves to channels through subsurface flowpaths [**Figure 4.8**].

Not only do flowpaths depend on time (variable source area), but they are also controlled by surface and subsurface heterogeneities. Transitions in slope topography, such as from convex to concave portions of the landscape, influence both surface and subsurface flow and the resulting sediment transport. For example, **hollows** or **swales**, which are concave portions of the landscape, concentrate flow and sediment transport [**Photographs 4.7** and **4.8**] and are often the sites where stream channels begin. In contrast, **noses** (topographic convexities) cause flow and sediment to diverge or spread out. Longitudinal, lateral, and vertical (with depth) heterogeneities in soil properties (such as hydraulic conductivity and grain size) control hillslope hydrology and sediment movement by influencing the amount of water moving at and near the soil surface.

Not all the water associated with a stream enters it over the ground surface. The subsurface component of flow along stream corridors, called **hyporheic flow**, sometimes accounts for a substantial portion of stream flow and can be ecologically important for riparian ecosystems [**Photograph 4.9**]. Channels often exchange substantial flow with permeable sediments within stream valleys, a common feature of gravel-rich mountain streams and floodplains in general.

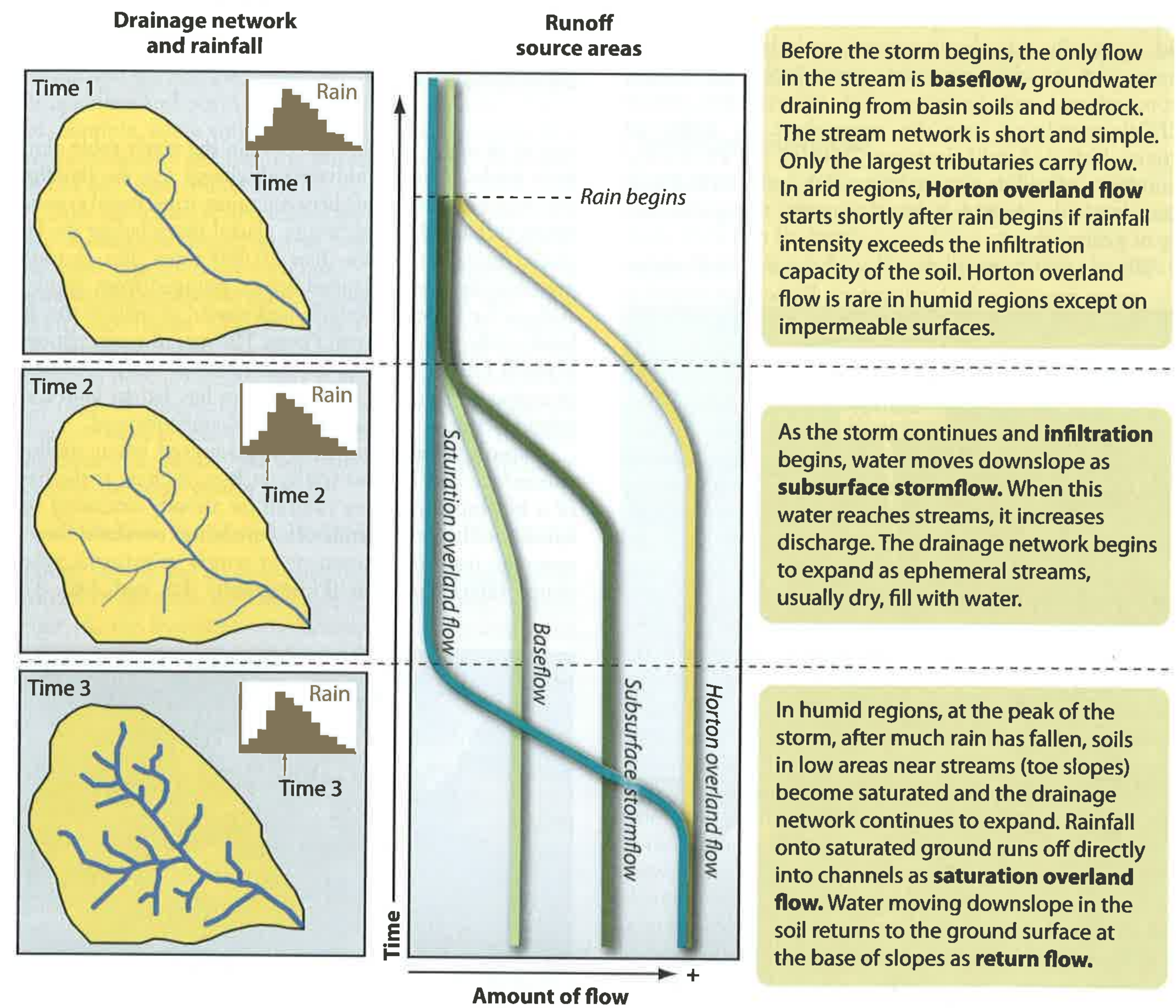


FIGURE 4.8 Variable Source Areas. Watershed flowpaths change during a storm. Areas of saturated ground expand away from channels, and the channel network itself grows larger as ephemeral streams fill with water. Baseflow increases as water tables rise. If rainfall intensity exceeds the infiltration rate or when the ground saturates near stream channels, overland flow will begin.

D. Montgomery

PHOTOGRAPH 4.7 Unchanneled Valley. Hollow or unchanneled valley, in southern Sierra Nevada, California. Here, the convergent topography focuses sediment transport, groundwater movement, and surface water flow.

P. Bierman

PHOTOGRAPH 4.9 Hyporheic Flow. Rio Puerco (New Mexico) has incised its channel. The deep channel acts as a drain, lowering the local water table. Along the incised channel, tamarisk trees thrive as their deep roots tap groundwater below this ephemeral river.

Soil characteristics can be used to identify areas of the landscape that are alternately wet and dry, seasonally saturated zones (see Photograph 4.5). Soils that are always wet are gray-green, the result of reducing conditions (the lack of oxygen). Soils that are well-drained and usually dry are well-oxygenated, allowing minerals to oxidize and leaving the soil reddish-brown. Topography also has a large effect on drainage, including the water table position and its seasonal variation, and thereby the location of perennially saturated **gley** soils (gray-green from reduction of iron).

When a geomorphologist finds red **mottles** (evidence of oxygen) in an otherwise gray (anoxic) soil, that's a hint that the water table (and the amount of oxygen) fluctuates over time—red reflecting iron oxidation during the dry season and gray reflecting reducing conditions during the wet season [**Photograph 4.10**]. Such seasonally varying saturation of the ground is often used as a criterion for wetlands

D. Montgomery

PHOTOGRAPH 4.8 Channel Head. Channel head and initiation of surface flow at the base of a hollow in Tennessee Valley, California.

J. Turenne

PHOTOGRAPH 4.10 Soil Mottling due to Variable Water Table Position. The red and gray features shown in this soil are mottles that form at and below the seasonal high water table, indicating alternation between oxidizing and reducing conditions.

delineation and subsequent protection. Soil mottles imply a varying water table. Finding mottled soils suggests that the ground has been saturated during wet periods, and thus saturation overland flow likely occurred during rainfall events.

Surface-Water Hydrology

Water moving over Earth's surface is a potent geomorphic agent, eroding slopes and riverbanks, moving sediment, and carrying dissolved loads. Surface water and groundwater flows are closely linked, with water moving rapidly and continually between the two systems.

Hydrographs

A fundamental means of characterizing surface water flow is to measure the volume of water passing a point in the channel over time. A common visualization tool used for describing this flux is the **hydrograph**, a graph that charts the volume of water moving through a channel (termed **discharge**) through time [Figure 4.9]. Hydrographs are

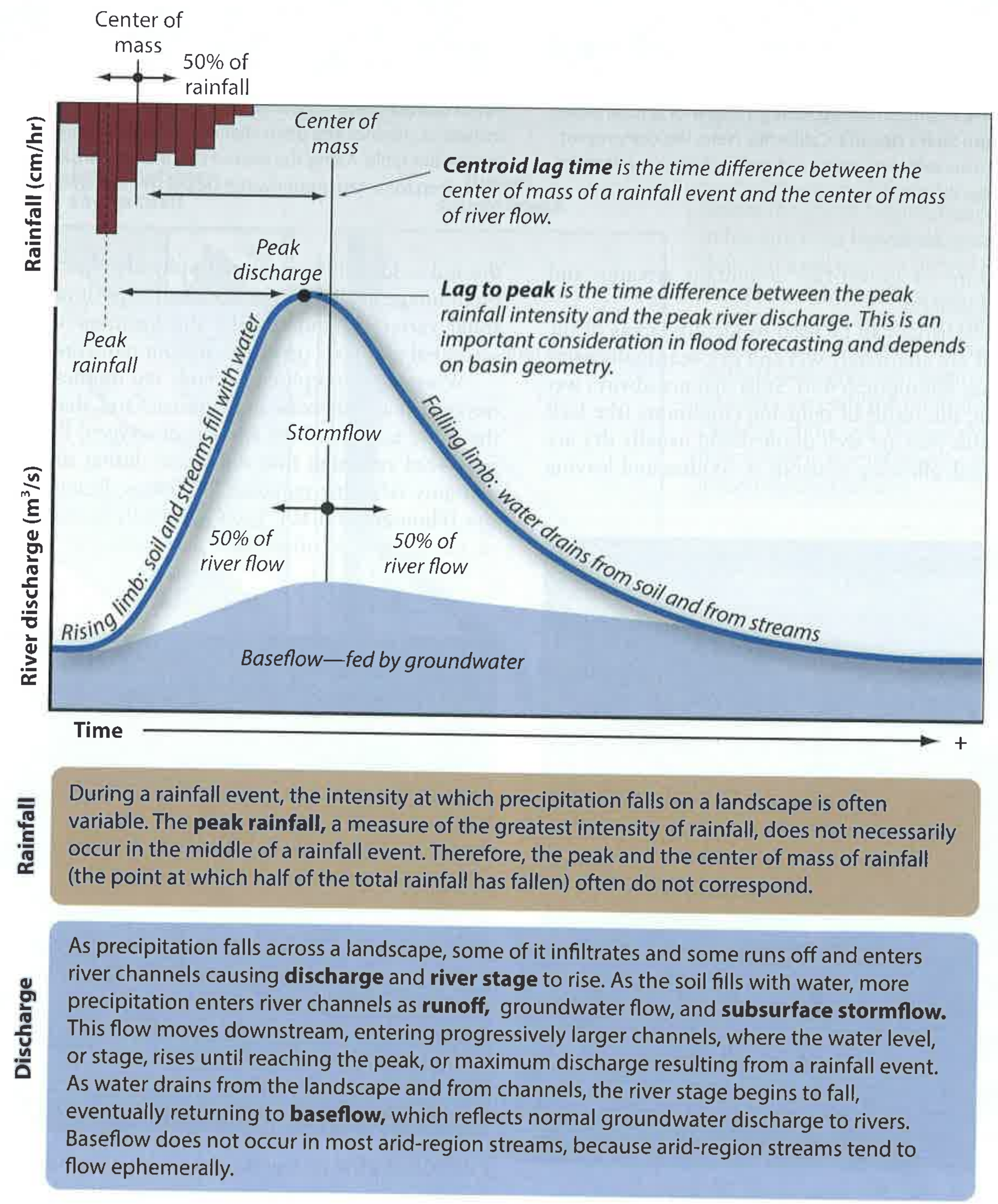


FIGURE 4.9 Hydrograph Definitions. Hydrographs describe flow through streams over time. Stream flow reflects the timing and volume of precipitation. The hydrograph can thus be interpreted through the lens of relevant runoff processes and pathways.

important because they can be interpreted to infer when geomorphically important events happen, such as when a river overflows its banks.

All hydrographs share common characteristics. **Baseflow,** derived from both deep and shallow groundwater drainage, is the amount of water flowing into and moving through a stream between storms. During a runoff event, flow rises from baseflow to a **peak discharge** defined by the intensity and duration of the precipitation or snowmelt that is providing water to the system. The **rising limb,** during which flow is increasing, is typically steep as water moves quickly to the channel over land and through shallow subsurface flowpaths. The **falling limb,** during which flow is decreasing, is generally less steep and persists longer than the rising limb as water slowly drains from the subsurface into the stream.

There is a delay between rainfall and runoff because it takes time for water to move across and through the landscape and into and down channels. This delay is referred to as the **lag-to-peak** (L_p) and is a function of basin shape and size [Figure 4.10]. L_p typically increases with basin size and varies by runoff generation mechanism, with short lag times between rainfall and peak runoff for overland flow and longer lag times for shallow subsurface storm flow and groundwater flow. Small, urbanized basins can have lag-to-peak times of minutes whereas the lag-to-peak in a large drainage basin, such as the Mississippi River Basin, may span weeks to months as water moves downstream through the river system. The lag-to-peak can be defined either as the time difference between the peak rainfall and peak runoff or as the difference between the center of mass of rainfall and the center of mass of runoff.

Hydrograph shapes reflect both the geomorphology and hydrology of the drainage basin as well as the antecedent weather and hydrologic conditions that preceded the runoff event. Steep narrow watersheds, saturated ground, and urbanization all generate hydrographs with especially steep rising limbs; basins with such characteristics are known as **flashy.** Headwater streams have steeper rises and more peaked hydrographs than main-stem, lowland rivers because these large rivers are fed by many tributaries, all contributing water at different rates and at different times. For small basins, the shape of the

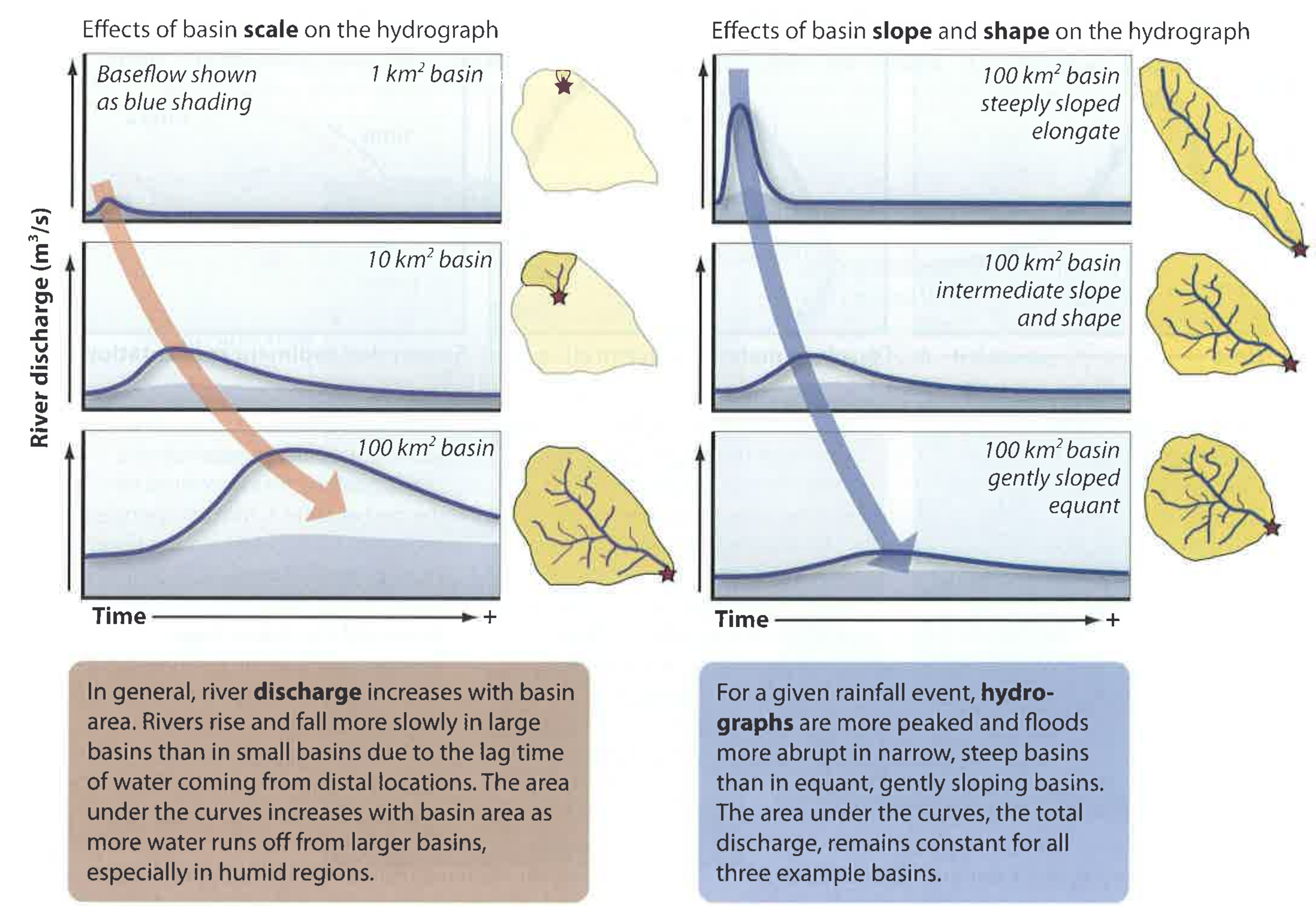


FIGURE 4.10 Hydrograph: Basin Scale and Shape. The shape, size, and timing of the hydrograph are directly related to the size and the geomorphology of the drainage basin.

hydrograph's receding limb is controlled by the rate at which both surface and subsurface processes deliver water to the channel after the storm ceases.

Elevated groundwater tables and long drainage networks lead to long, gradual **flow recession.** In contrast, hydrographs from small urban streams fall quickly because these basins are largely paved and thus little water infiltrates to sustain baseflow. In large basins, the shape and duration of the receding limb predominantly reflect the time it takes for water to move through the drainage network.

The **recession constant,** K_r, empirically describes the rate at which discharge decreases after the hydrograph peaks

$$Q_t = Q_p K_r^{\,t} = Q_p e^{-\alpha t} \qquad \text{eq. 4.4}$$

where Q_t is the discharge at time t, and Q_p is the peak discharge at the start of uninterrupted periods of declining discharge. By defining $\alpha = \ln K_r$, equation 4.4 may be expressed as

$$\ln Q_t = \ln Q_p - \alpha t \qquad \text{eq. 4.5}$$

and α can be calculated from the slope of semilogarithmic plots of discharge recession, where time is on the x-axis and lnQ is on the y-axis. K_r values cluster in different ranges for different runoff generation mechanisms, with low K_r values for Horton Overland Flow (<0.3), reflecting rapid discharge recession, whereas higher K_r values for subsurface stormflow (>0.3) imply sustained drainage and prolonged discharge after the storm ends. Saturation overland flow hydrographs exhibit a wide range of K_r values due to the influence of subsurface stormflow on sustaining return flow during discharge recession.

Hydrograph shape integrates the effects of rainfall patterns, runoff generation processes, and hydrologic and topographic properties of the catchment. The lag-to-peak (L_p) and discharge recession constant (K_r) provide simple measures of the timescale of runoff response. These two measures, L_p and K_r, quantify differences in drainage basin hydrologic processes, can change through a storm event or hydrologic year as different runoff generation mechanisms become active, and can be measured for basins from the scale of individual headwater channels to large rivers.

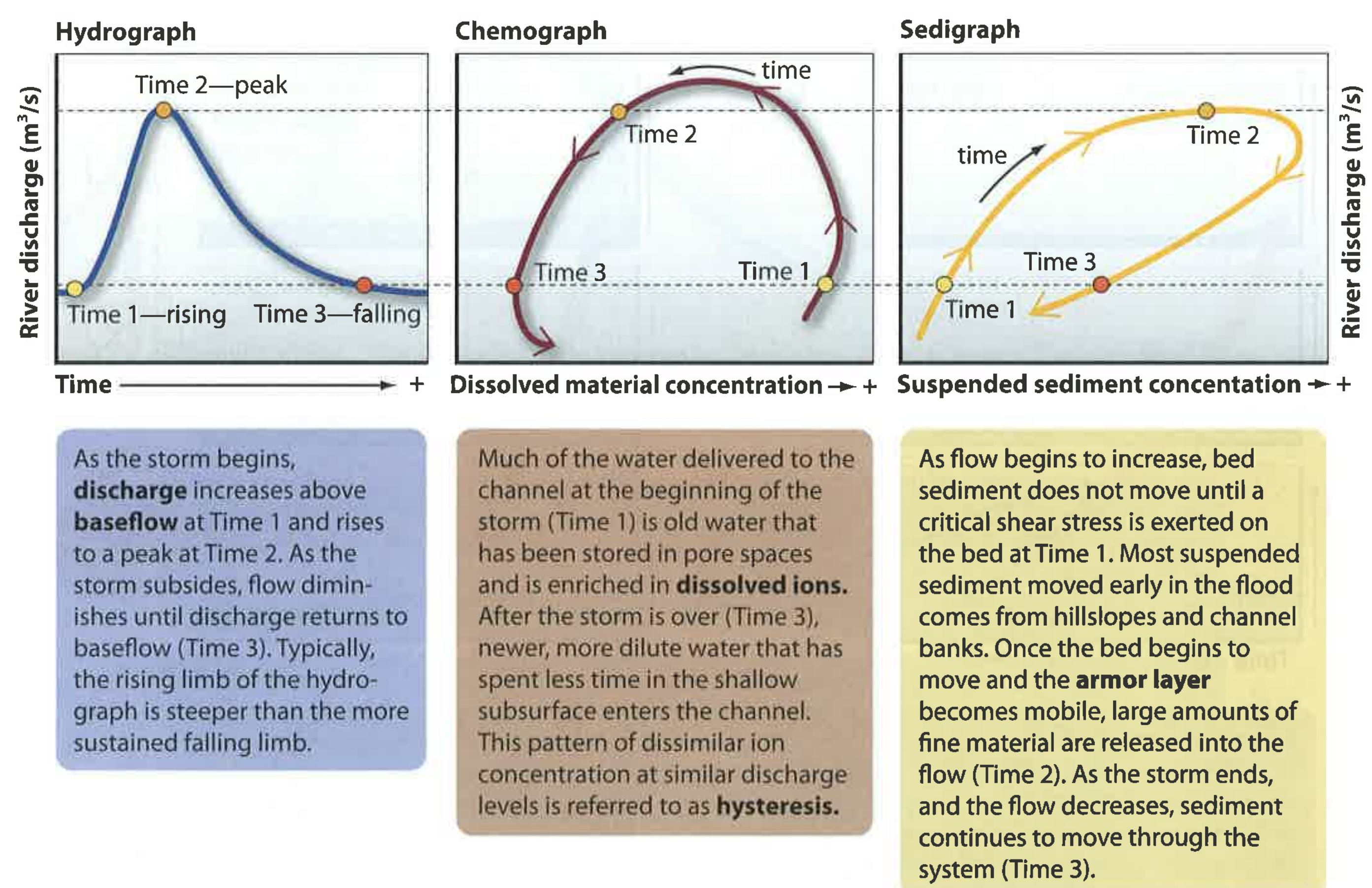


FIGURE 4.11 Hydrographs, Chemographs, and Sedigraphs. Streams and rivers move mass across Earth's surface, both as sediment and as material dissolved in solution. Due to the complex nature of sediment transport and flushing of pore water from soil by rainstorms, sediment and solute concentrations vary through the hydrograph and are not directly related to discharge—the phenomenon referred to as hysteresis.

Closely related to the hydrograph are the **chemograph** and the **sedigraph.** The chemograph relates the dissolved mass loading of an element to discharge; the sedigraph plots the concentration of suspended sediment in transport as a function of discharge. Both chemographs (dissolved material) and sedigraphs (solid material) tend to exhibit a phenomenon known as **hysteresis,** in which dissolved load and sediment load values are path- and time-dependent rather than depending only on discharge. For example, sediment and dissolved load concentrations at a discharge value on the rising limb of a hydrograph are different from those at the same discharge value on the falling limb **[Figure 4.11]**.

Thinking about the physical processes active in and near the channel can explain hysteresis. For both sediment and dissolved load, hysteresis reflects the mobilization and flushing of material from the system. As the discharge increases, geomorphic action in the channel begins. As the velocity and the depth of flow rise, so does the shear stress on the bed. When bed material begins to move, the concentration of sediment in the flow increases. Larger amounts of sediment move after the flow becomes strong enough to mobilize the coarse layer of clasts that armors the surface of most channels, exposing the more easily moved, finer-grained material below.

Once liberated from the bed, sediment moves downstream rapidly so that sediment concentration is often higher during the receding limb than during the rising limb of the sedigraph. Interpreting chemographs is also complex. They reflect the addition to the channel both of rainfall precipitating onto the basin and the addition of "old" water previously stored in the groundwater system (and thus enriched in dissolved mineral constituents such as calcium and sodium) but now displaced and forced out by newly infiltrated stormwater.

Interactions Between Groundwater and Surface Flow

Although many people and many legal theories consider surface flow and groundwater to be separate systems, they are intrinsically connected. Water moves freely from streams, rivers, and lakes into the ground and then comes to the surface again in other locations. An informative way to understand this linkage and interaction is to consider the connections between streams and groundwater.

Streams and rivers can be characterized in different ways. One taxonomy places streams in one of two categories, **gaining streams** and **losing streams [Figure 4.12]**. Discharge in gaining streams tends to increase downstream because those streams act as drains for groundwater. This occurs because the water level or head in the stream is lower than that of the adjacent groundwater table, establishing a head or energy gradient so that groundwater moves into the stream. The result is baseflow. Losing

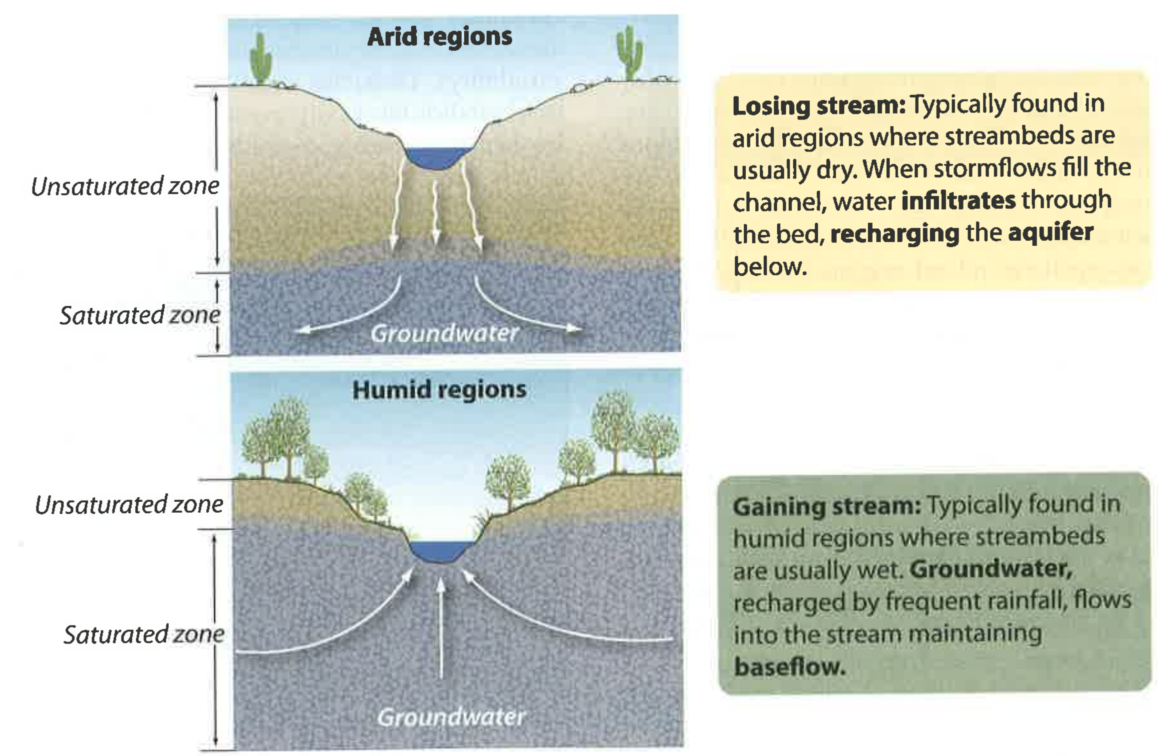


FIGURE 4.12 Gaining and Losing Streams. Streams and groundwater interact. In arid regions, streams lose flow through their beds, recharging aquifers; thus, discharge decreases downstream. In humid regions, streams usually function as drains, gaining flow from groundwater; thus, discharge increases downstream.

streams are most common in arid regions where groundwater tables are typically well below the bottom of the stream channel; thus, discharge in losing streams tends to decrease downstream. In the desert, these losing streams serve as intermittent sources of recharge to the groundwater system operating whenever water from storms or snowmelt in adjacent highlands flows through their channels. Losing streams often run dry between storm events, all their flow having infiltrated.

A single stream may gain or lose water at different times and at different places along its channel. When groundwater levels are high, such as during the winter or immediately after storms and snowmelt events, the head difference is great and large amounts of water flow into streams. However, in large channels that receive substantial discharge from upstream, river stage can rise more rapidly than adjacent groundwater levels. When the head in the stream exceeds that in the adjacent groundwater, river water flows into the bank material and the stream loses water. If flood levels drop quickly, pore pressures in the bank can remain elevated because soil takes longer to drain than river stage takes to drop. The resulting gradient in pore pressures can trigger bank collapse.

Another taxonomy classifies streams as perennial, ephemeral, seasonal, and intermittent. **Perennial streams** always contain flowing water and are typically found in humid regions where precipitation is distributed throughout the year and where groundwater tables are at or near the land surface during all seasons. The size of the drainage basin needed to support a perennial stream depends on climate and lithology; in moist climates and on generally impermeable lithologies, small drainage basins can support perennial streams. **Ephemeral streams** have flow only during large or intense precipitation events or during snowmelt. Ephemeral streams are common in arid regions and in upland drainage basins with small catchment areas. In many cases, such streams are disconnected from the regional groundwater table. During storms, shallow subsurface flow enters these upland streams and they fill with water. As a storm subsides, groundwater levels lower and water in ephemeral streams infiltrates into the bed.

Seasonal streams have flow only during certain, predictable parts of the year, for example, during spring snowmelt. Flow in **intermittent streams** is discontinuous in space, usually as a result of changing hydraulic conductivity in the subsurface and thus differences in the ability of the subsurface material to transmit water. For example, where a stream flowing over rock enters a valley bottom reach filled with large cobbles, water may sink into the subsurface and flow between the cobbles [Photograph 4.11].

Flood Frequency

Floods occur when rivers overtop their banks. Floods are natural and normal events punctuating the otherwise tranquil baseflow of many rivers and streams [Photographs 4.12 and 4.13]. Because of the geomorphic influence of regular high flows, contemporary restoration efforts on large river systems often aim to recreate at least some of the effects of regular, high flows associated with snowmelt or other hydroclimatic drivers. The level or discharge of the annual high flow can have a great deal of interannual variability (which tends to increase as precipitation decreases—i.e., dry regions have the greatest interannual variability). Thus, the word "flood" can refer to a regularly predictable yearly event as well as to unusual, less frequently occurring, storm-driven high flows.

P. Bierman

PHOTOGRAPH 4.11 Intermittent Stream Flow. This intermittent stream drains the west side of the Appalachian Mountains in Shenandoah National Park, Virginia. The streambed, being highly permeable (because it is composed of large, quartzite clasts), carries flow beneath the surface in most conditions—except during floods, when the volume of water increases so much that some water flows on the surface. Upstream and downstream, where the channel bed is finer and less permeable, water flows on the surface.

S. Haberman/Angonia

PHOTOGRAPH 4.12 Suspended Sediment. The Kuiseb River in Namibia rarely floods, but when it does, it carries large loads of suspended sediment. The blue sky suggests that the silty water pouring down the river likely fell on the more humid highlands rather than on the arid lowland region where this image was taken.

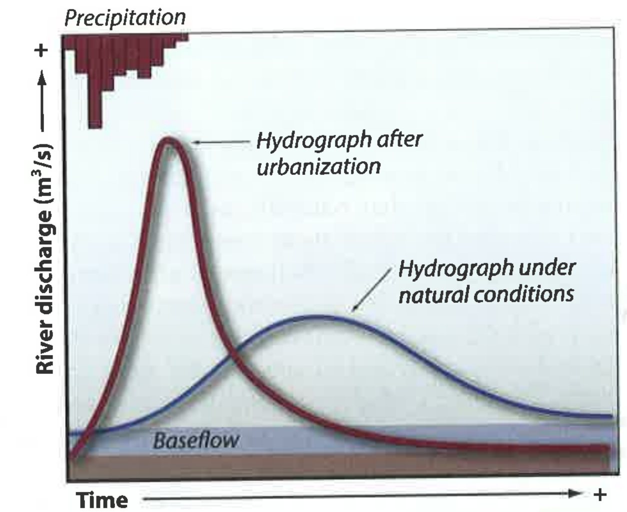
Fletcher Free Library, Burlington, VT

PHOTOGRAPH 4.13 Connecticut River in Flood. In 1927, the largest flood on record hit the Connecticut River. Heavy rain fell for two days in early November onto ground saturated by record October rainfall. The rain quickly ran off, driving rivers and streams out of their banks. At St. Johnsbury, Vermont, this bridge was swept away as the townspeople watched.

Similar to precipitation events, floods can be described in terms of recurrence intervals using equation 4.1 and substituting peak annual **flood stage** (height) or discharge for precipitation intensity [Figure 4.13]. Flood recurrence intervals are established by analyzing discharge records typically maintained by government agencies. In the United States, many such records are kept by the Water Resources Division of the U.S. Geological Survey and are available online. Other records are kept by the Army Corps of Engineers, National Oceanic and Atmospheric Administration (NOAA), and state agencies.

Additional flood data, stretching back hundreds to thousands of years, can sometimes be determined geologically. **Paleoflood hydrology,** the study of ancient floods predating gauge records, relies on evidence for past flood heights by dating flood debris preserved above the modern-day channel. Often such material is preserved in caves or bedrock alcoves. Such an approach has been of particular utility for identifying the size of extreme floods in arid regions.

Bankfull flow is a commonly used datum to describe the size of floods that fill the channel but are not high enough to inundate the floodplain. In humid regions, many streams fill their banks every year or two, on average. Geomorphologists tend to pay particular attention to the bankfull flow, based on the idea that this flow shapes alluvial channels because larger floods may do more work but are so rare that they are relatively unimportant in setting channel form. Lower flows, although more common, do not have the energy to cause significant changes to the channel. The recurrence interval for bankfull flows varies depending on the environment. In arid regions, bankfull flows may recur as infrequently as every 50 years, on average. In bedrock canyons and channels without well-developed floodplains, defining the bankfull stage is not possible. In



Under natural conditions, rainfall follows convoluted paths through the landscape. Water is held in **detention storage** by irregular pit and mound topography, infiltrating into organic-rich forest soil, and moving slowly to the channel. The infiltrating water feeds **baseflow** during times when it is not raining. **Flood peaks** are delayed because natural landscape characteristics slow the rising limb of the hydrograph, lower the peak flow, and extend the flood duration.

After urbanization, rainfall moves rapidly to the channel with little chance to infiltrate; thus, baseflow is reduced. Flowing directly off **impervious surfaces,** such as parking lots, and into storm sewers, runoff rapidly enters streams, raising their level quickly. Flood peaks now come sooner and are higher, increasing flood hazards and the tempo of geomorphic change. In the example below, the natural 25-year flow becomes the much more frequent urbanized 2-year flow.

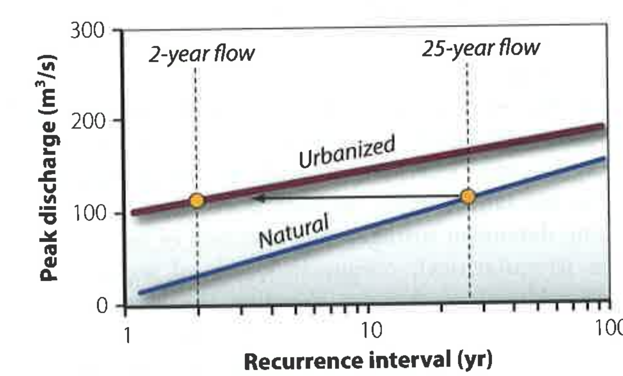


FIGURE 4.13 Pre- and Post-Development Hydrographs. Development, and the increase in impermeable surfaces including pavement and roofs, changes runoff processes and the shape of the hydrograph. As development increases, rising limbs steepen, peak flow increases, and baseflow diminishes. Return intervals shorten for floods of the same magnitude. In the case illustrated, a flow of ~125 m^3/s occurred under natural conditions on average once every 25 years, but under urbanized conditions the same flow now occurs about once every 2 years.

channels responding to disturbance, especially those that are incising, the bankfull flow level may be nothing more than a discontinuous low terrace at the base of steep banks.

Because the concept of bankfull flow is commonly applied in stream management and restoration, it is important to realize that bankfull, as defined based on channel morphology, is not always associated with a particular recurrence interval discharge. Furthermore, the concept is most applicable to environments where a relatively frequent discharge is most likely to fill the channel banks and do a great deal of geomorphic work (humid-temperate regions, snowmelt systems). In arid environments, where infrequent discharges shape the channel and transport most sediment, bankfull flow is not likely to be the annual flood flow but rather a flood that occurs more rarely. Still, the bankfull flow is important for setting the scale of alluvial channels even in arid regions.

Of broad societal importance is the **100-year flood,** the flood with a yearly return probability of 1 percent. There is nothing scientifically special about the event that has a 1 percent chance of occurring each year. Rather, the 100-year flood was an arbitrary risk level chosen by U.S. government agencies and Congress when they wrote the 1968 Flood Control Act. Even though the annual probability of a 100-year flood is very low, there is no physical reason preventing a 100-year flood from recurring in two successive years. Characteristics of the 100-year flood (discharge and stage) are used extensively for planning and zoning purposes and often form the basis for defining floodplains in terms of insurance and development regulation. In addition to planning that considers hazards of inundation from rising floodwaters, "smart" river corridor planning also needs to consider lateral movement of the channel driven by bank erosion. Such lateral migration is typical of most streams with erodible banks and can be particularly rapid in arid-region streams where riparian vegetation is sparse and there is little, if any, root reinforcement.

Land-use and climate changes (both natural and human-induced) can alter flood recurrence intervals. Clearing forest or grassland to build houses or businesses smooths the land surface so that lawns can be planted (reducing **detention storage,** the collection of rainwater in surface irregularities), compacts the land surface, and reduces infiltration, thereby speeding the movement of water to channels (see Figure 4.13). The addition of impermeable surfaces and storm sewers further increases the speed of runoff and thus the peak discharge, stage (height), and erosion potential of floods downstream. It is common to find that peak discharges that occurred, on average, once every decade before development, occur every year or two after development (see Digging Deeper). The geomorphic and societal effects of development, and its attendant hydrologic alterations, can be wide ranging and include channel incision, flooding, and reduced baseflow between runoff-producing events.

Water Budgets

Basin-scale **water budgets** are important tools for understanding basin behavior and managing water resources. They can be used to answer such practical questions as, "How much water can be withdrawn from a stream for snowmaking in the winter before flow will drop below levels needed to support overwintering aquatic life?" or "How much water can be withdrawn from a stream for irrigation before withdrawals exceed the annual water supply?" Water budgets are useful for predicting how much water might be available for uses and for partitioning that usage when demand exceeds supply.

Such budgets consider inputs from precipitation **[Photograph 4.14a]** and direct condensation from fog or clouds in locations where this is important (mountaintops, hyperarid seacoasts). Outputs of water from drainage basins include direct evaporation from water surfaces and soils, transpiration by plants, groundwater flow, and channelized surface flow **[Photograph 4.14b]**. Constructing accurate water budgets, especially for small basins, is not straightforward because basin-specific data are usually hard to come by.

Precipitation, the primary input of water to most watersheds, can be estimated in a variety of ways. Established weather stations record precipitation directly, but such stations are few and far between. Where dense networks of precipitation collectors have been installed, the resulting data clearly show large variability in precipitation amounts and rates over small areas, especially in rugged terrain where topography affects precipitation patterns.

Recently, the application of basin-scale precipitation models **[Photograph 4.14c]** has become more widespread; these models rely both on the observational record and known relationships between elevation and rainfall/snowmelt to produce spatial estimates of precipitation. Weather radar systems have been calibrated to provide real-time precipitation estimates during storms. Such data can be integrated to estimate precipitation inputs over time. Flood forecasting combines water balances with computer models that include precipitation estimates and route stormwater downstream to predict the timing and magnitude of high flows. In areas where the winter precipitation falls as snow, transfer of snow between watersheds by wind can complicate water budgeting. In small, high-elevation catchments, large amounts of water (in the form of blown snow) can move from one basin to another.

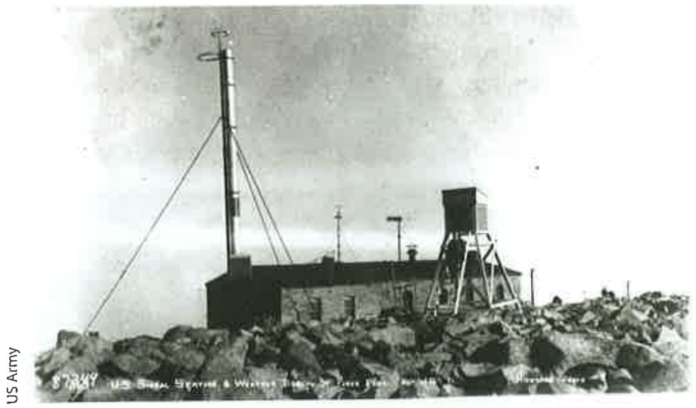
Losses of water from a watershed can be measured or estimated. Evaporation and transpiration are usually modeled using established approximations that consider temperature, wind speed, and vegetation type. Because most watersheds, even those in developed countries, do not have stream gauges that provide continuous discharge records, generalized regional water discharge/basin area relationships are useful. Such relationships are created by regressing basin area and mean annual discharge for basins of different sizes that are gauged. These relationships are

most useful when the ungauged basin has similar shape, topography, and orientation as the calibration basins.

Groundwater losses are even more difficult to quantify and are usually calculated by differencing after the other terms in the water budget have been estimated. While this approach has been very useful in the past, temporal variability, especially the nonstationary behavior of the climate and thus the hydrological system in response to human-induced climate change, may introduce greater uncertainties and render the approach less useful in the future.

Surface water moves rapidly through basins; in-channel storage times are short (minutes to days) and scale with channel length and inversely with slope and velocity. Groundwater **residence times** are much longer. Shallow subsurface groundwater might reside in the basin for days to months. Deeper groundwater can remain in large basins over glacial/interglacial timescales and is essentially fossil; the residence time of water in such deep aquifers can be many millennia. Indeed, groundwater extracted from the major irrigation aquifers in the central United States, such as the Ogallala, and in the Great Artesian Basin of Australia, was likely recharged during the Pleistocene Epoch under a completely different climate regime. Once withdrawn, such ancient water will not be replaced on human timescales, making fossil groundwater a nonrenewable resource.

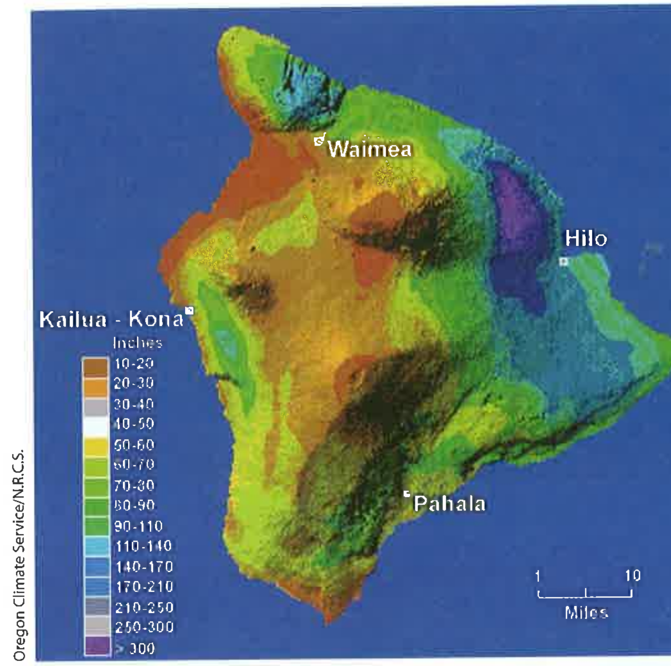
(a)


US Army

(b)

USGS

(c)

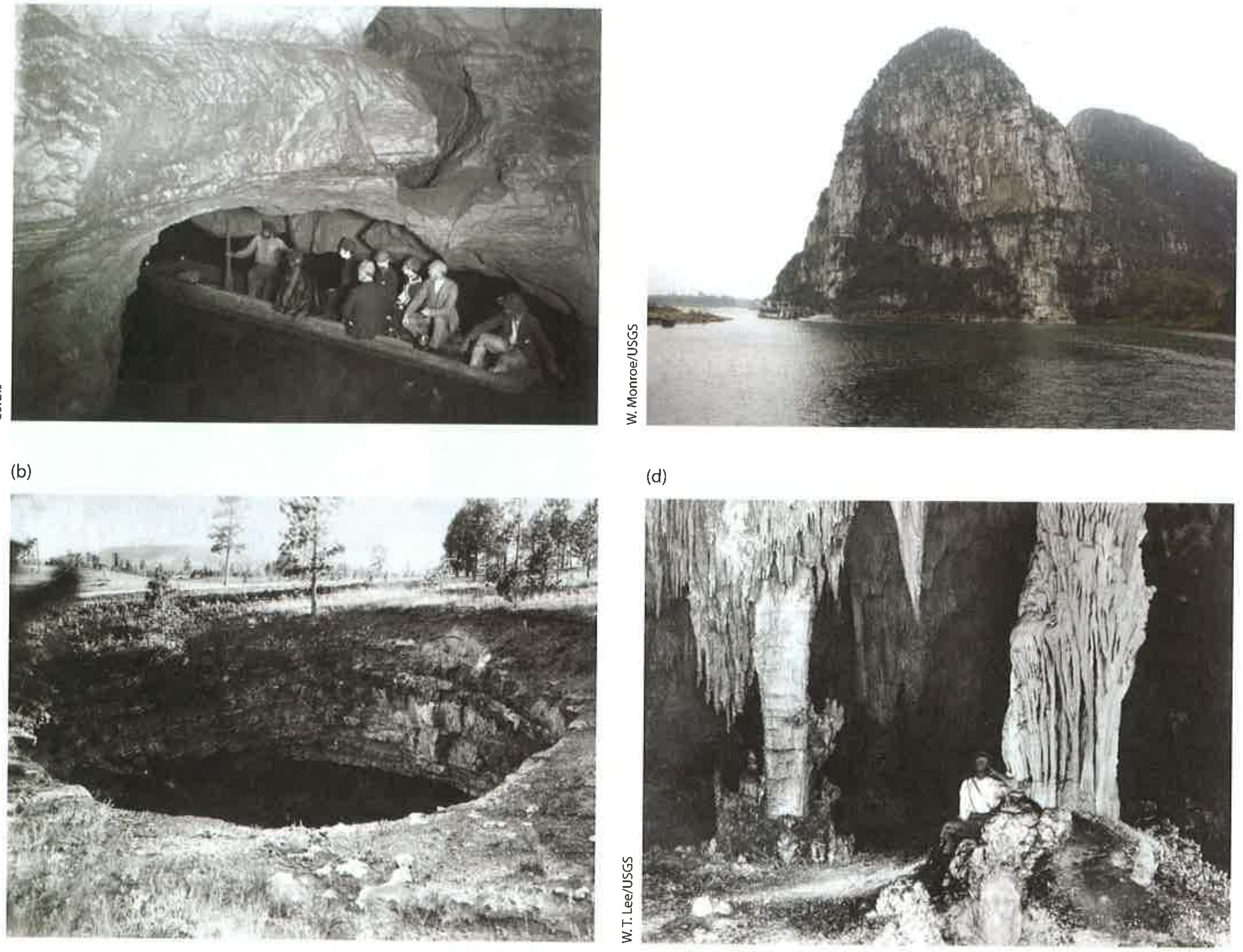


Oregon Climate Service/N.R.C.S.

PHOTOGRAPH 4.14 Instrumentation Needed for Water Budgeting (a) An early weather station (1880) with rain gauge on Pike's Peak, Colorado, at an elevation of ~4200 meters. (b) Dam and rectangular weir for measuring stream flow in the Los Angeles River in southern California, 1904. (c) Geography strongly controls the distribution of annual precipitation on the island of Hawaii. This map shows how wet the east side of the island is due to orographic lifting of moist tropical ocean air carried by northeasterly trade winds. The data are from the PRISM model, an algorithm that takes scarce observational data and combines these with terrain models to estimate the spatial variability of precipitation. [*Source:* 4 × 4 km PRISM Model Grid. Based on 1961–90 annual precipitation data from NOAA Cooperative Stations. Hillshade relief derived from U.S. Geological Survey DEMs.]

Hydrologic Landforms

Weathering and erosion of carbonate rocks and evaporites produce unique topography dominated by dissolution and groundwater hydrologic processes; landscapes produced this way are referred to as **karst,** a German word for a limestone region in Slovenia. Karst landscapes are both morphologically and hydrologically distinct systems in which large amounts of water move between Earth's surface and interconnected cavities below ground **[Photograph 4.15]**. Chemical weathering and dissolution produce distinctive types of karst landforms, and drainage of karst terrain primarily occurs below ground, producing unique and often very limited



PHOTOGRAPH 4.15 Karst Landscapes. Limestone, and the karst landscapes developed on it, can be spectacular. (a) Touring watery passages of Mammoth Cave, Kentucky, by boat. The National Park Service discontinued tours such as this in the 1990s. (b) Sinkholes form as rock below collapses from dissolution. Here is a sinkhole in the Minnehakta limestone southeast of Boyd, Weston County, Wyoming, circa 1900. (c) Dramatic view of tower karst topography at river's edge, taken from a boat on a trip down the Li River south of Guilin, China. (d) Pillars of dripstone or flowstone deposited from calcium carbonate–saturated solutions at Carlsbad Caverns National Park, New Mexico.

surface drainage patterns. Most of the world's great cave systems are developed in limestone and occur in karst terrain.

The solubility of limestone is primarily regulated by groundwater flux, acidity (pH), and CO_2 concentration (eq. 3.3). Consequently, abundant vegetation cover and rainfall favor karst development. As calcite dissolution occurs, the largest, fastest-growing voids tend to capture flow and grow at the expense of smaller, slower-growing voids, resulting in preferential development of a limited number of conduits. This favors development of distinct caverns and subsurface drainage. Karst topography in silicate rocks, referred to as **silicate karst,** is rare, can take many millions of years to develop, and is found only on unusually stable, ancient land surfaces in some parts of the equatorial tropics, such as the stable cratons within the Amazon basin.

Although karst landforms may be found in any environment, they are most common in humid-temperate and tropical regions. In the United States, karst terrain is concentrated in the eastern and southern regions, principally in Florida, the Appalachian Mountains, southern Indiana, New Mexico, and western Kentucky. Other important karst landscapes include areas in China, Australia, Malaysia, Jamaica, and Spain. This distribution reflects the location of both carbonate rocks and sufficient precipitation to dissolve those rocks.

Most extensive regions of karst topography are developed on limestone consisting of more than 50 percent $CaCO_3$, although dolomite (magnesium-rich carbonate) and evaporites such as gypsum may also develop karst topography. In general, karst formation potential increases

with limestone purity. In particular, the development and nature of porosity along joints, faults, fractures, and bedding planes promote dissolution if coupled with high permeability (connectivity) that allows free circulation of water.

Tufa and **travertine** are deposits that result from the evaporation of or degassing of CO_2-enriched waters, which causes precipitation of calcium carbonate that can coat silicate rocks or accumulate to form thick deposits [**Photographs 4.16** and **4.17**]. These deposits mark the prior location of springs, groundwater flow, and surface water flow; thus, mapping their locations allows for the reconstruction of past flowpaths.

The most common karst landforms are **sinkholes,** also known as **dolines.** Typically wider than they are deep, dolines are generally circular to elliptical and narrow with depth, producing a funnellike shape. Dolines can reach sizes of up to 1 km across and hundreds of meters deep. Areas with abundant dolines can have a distinctive pitted appearance and lack typical valley development. Dolines can form by either solution or collapse into subsurface caverns that themselves result from solution [**Figure 4.14**].

Hydrologically, karst is characterized by disrupted surface flow with interrupted stream valleys and closed depressions, some of which are filled by groundwater, indicating that they intersect the groundwater table. There are frequent diversions of surface water underground as streams disappear into subterranean conduits only to reemerge as **springs** (places where groundwater emerges at the surface) of all sizes. Springs are often found at topographic, stratigraphic, and structural discontinuities.

Karst landscapes have particularly high secondary porosity, with most groundwater moving through conduits of differing dimensions dissolved into the rock. Differences in nearby flowpaths (distinct conduits) can lead to significant groundwater level changes over short length scales. Groundwater moving through karst may move underground between watersheds (crossing below surface watershed boundaries), because the flow of groundwater responds to head gradients rather than topography, which controls the flow of surface water.

Caves and springs are common in karst terrain. Cave formation appears to be controlled by rock structure, fracture patterns, and lithology with caves developing as conduits along zones of preferential groundwater flow and dissolution. The extensive cave networks common in karst terrain exhibit a variety of morphologies based on their pattern as viewed from above, their branching characteristics, and their cross-sectional shape. In a positive feedback loop, fractures that capture the most flow grow most rapidly into caves at the expense of others.

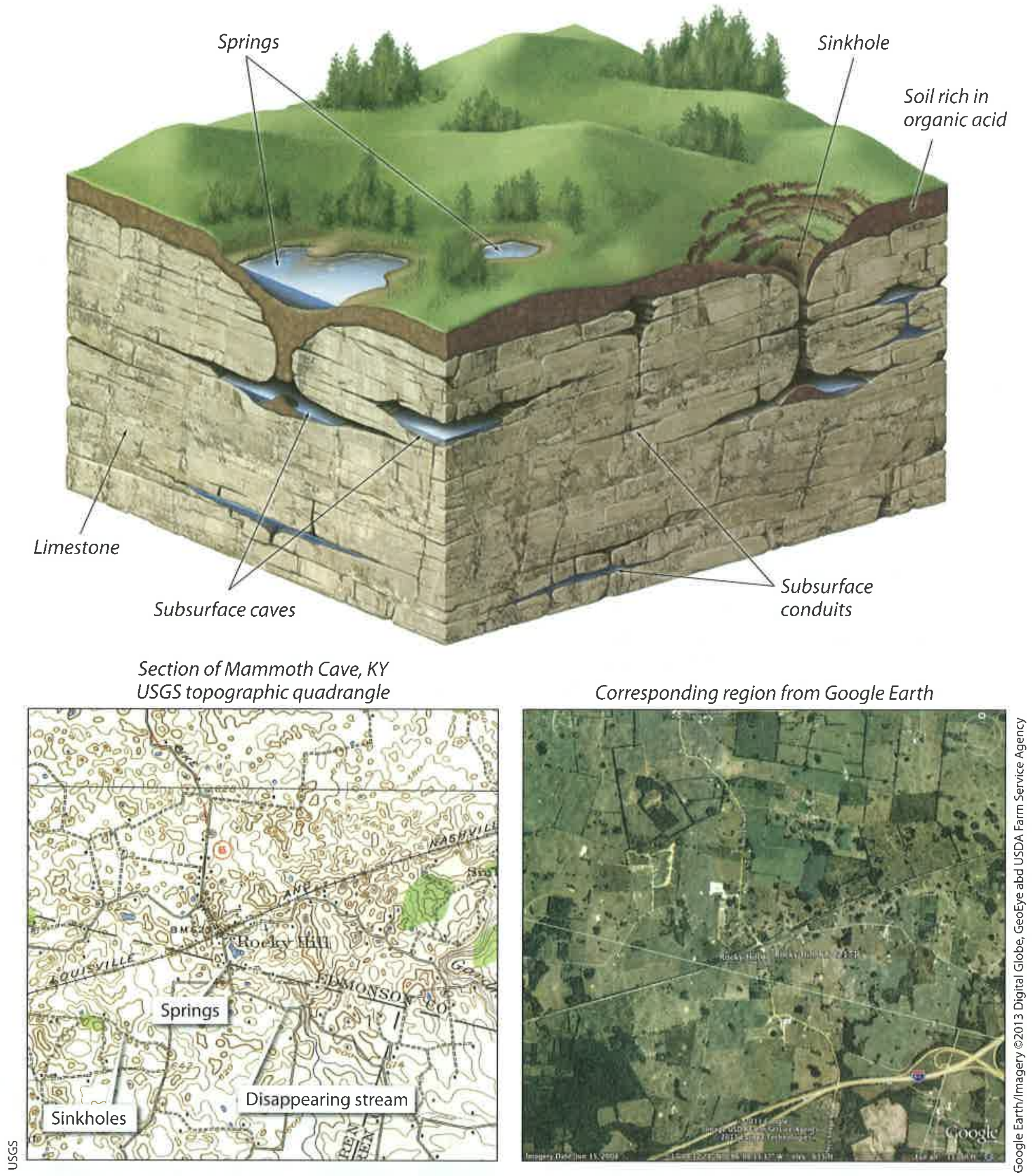
Caves that form at the groundwater table can be used to estimate regional river incision rates if deposits in a vertical series of caves can be dated. Such dating can be done using U/Th on carbonate dripstone deposits in the cave; cosmogenic methods (using two isotopes with different half-lives) can be used to date how long clasts, washed into the cave, have been isolated from cosmic radiation.

PHOTOGRAPH 4.16 Tufa Deposits. Stream near Ruapehu volcano on the North Island of New Zealand, encrusting a waterfall with tufa, a deposit of calcium carbonate.

In some karst terrain, rivers flow into **blind valleys** with no outlet, from which the only exit for water is underground. **Dry valleys** are common in karst terrain; these are a kind of intermittent stream that results when a river, flowing across a valley floor, disappears into a sinkhole and flows below ground, leaving its former valley high and dry. Karst terrain often consists of closed depressions that are not connected by an integrated surface drainage network.

Tower karst is found most often in the tropics and describes a landscape where steep-sided hills of limestone rise above a low-lying alluvial plain. Some suggest that these towers are the more resistant rock and that recent stripping has removed the intervening weaker material, leaving the rugged, high-relief landscape.

PHOTOGRAPH 4.17 Tufa Towers. The "man and woman" tufa towers were deposited by springs entering Mono Lake in southern California when it was deeper, before the streams feeding the lake were diverted as water supplies for the city of Los Angeles. Now, they stand as dry monuments to the lowering of the lake level so that the Los Angeles population and farm fields could have more water.



Karst landscapes are very distinctive; topographic maps show many closed contours and airphotos show pits, some filled with water. Karst landforms are the result of rock dissolution and include subsurface caves and conduits through which groundwater flows before emerging in springs and flowing into rivers. There are **sinkholes,** which can be caused by cave collapse. **Disappearing streams** appear to vanish; in fact they flow for some distance underground in connected passages. On the surface, karst terrain is often rough, pockmarked by sinkholes. There can be large areas with no surface drainage. The lack of streams in karst terrain, even in humid, temperate environments with high mean annual rates of precipitation, reflects the large amount of water moving underground.

FIGURE 4.14 Karst Landscape. Karst landscapes, such as the one shown on the Mammoth Cave, Kentucky, topographic quadrangle, are complex, with numerous closed depressions and intermittent surface drainage as water moves between the surface and the ground. Much of the subsurface flow in karst is carried by enlarged fractures, secondary porosity in otherwise nearly impermeable limestone. [Upper image adapted from Panno and Wolf (1997).]

Applications

There are complex and ongoing interactions and feedbacks between water and the landscape. Although surface water and groundwater shape our planet's surface by eroding earth materials physically and chemically, it is the landscape that determines hydrologic flowpaths. Topography controls the direction of surface water flow and in many cases the direction of subsurface, groundwater flow. Water flowing over slopes does geomorphic work, moving sediment downslope both in diffuse overland flow on surfaces with low infiltration rates and in concentrated flow in rills and small channels. As water erodes the landscape, it reshapes topography, which in turn controls future flows and the flow of groundwater. Together, surface water and groundwater are geomorphic agents important in shaping landscapes.

Vegetation plays a role in surface and subsurface hydrology and thus in the geomorphology of many landscapes. By transpiring water, trees change the hydrologic balance of watersheds, reducing the water available for transporting materials and reducing the frequency and extent of saturated soils during growing-season rainstorms. People have tried various schemes to increase water yield by harvesting trees, but unless the slopes are kept clear, young trees regrow and use water. In addition, when slopes are cleared of vegetation, the root systems rot, weakening the slope and eventually closing off macropores that can rapidly transmit large amounts of water through the subsurface.

Not all landscapes were once forested and changes in vegetation other than deforestation, such as the conversion of grassland to agriculture, also affect the hydrologic cycle, erosion, and sedimentation. One of the most important hydrologic effects of modern, mechanized agriculture is the installation of tile drains to lower groundwater levels in flat-lying fields, particularly in the U.S. Midwest. These drains improve crop yields and extend the growing season by drying the fields earlier in the spring and preventing waterlogging during rainy spells, but the drains speed the transport of both water and nutrients to streams and rivers. The results are increased flood peaks and eutrophication of water bodies (including the Gulf of Mexico) as nutrients, applied as fertilizers, move downstream.

Humans both affect and are affected by the hydrologic system. Landscape modification, including removal of forests, agriculture, and urbanization, change the rate and means by which water leaves the landscape through their effects on infiltration rate and detention storage. Not only do such changes affect groundwater and surface water response to precipitation at the local or hillslope scale, but they change the size and frequency of surface water discharges downstream. Floods that used to occur only rarely can become commonplace, and their effects can ripple through the landscape, destabilizing stream banks, changing the sediment transport capacity of streams, and degrading habitat for valuable aquatic organisms, such as salmon.

The intimate connection between groundwater, surface water, and plants is clear to geomorphologists. Extracting large amounts of groundwater can reduce baseflow, drying out streams and changing their ability to transport sediment and provide habitat. Groundwater levels go up and down with the seasons, in part reflecting seasonal changes in precipitation, but these levels are also driven by the ability of trees and other vegetation to pump water from the ground by transpiration. Measuring shallow groundwater levels in the winter often gives very different results than measuring them in the summer. Such changing levels could be key when modeling slope stability—low summer groundwater levels might suggest a slope is stable, whereas high winter groundwater levels might indicate that the slope is near failure.

Climate change will affect the hydrologic system, dramatically in some places. With a warming climate and increasingly active hydrologic cycle, recurrence intervals for precipitation and stream flows are changing. Over the next century, the warming climate will likely affect the distribution of vegetation and the frequency of droughts, forest fires, and storms. Together, these changes will affect slope stability, runoff, and the amount of water and sediment shed from landscapes. One of the most dramatic hydrologic changes will be in streams issuing from high-elevation, glaciated mountain basins. Many of these streams are critical water supplies for villages in the valleys below. If the climate warms sufficiently that glaciers melt away, then the natural reservoirs of ice and snow, which provide water all summer from melting ice, will vanish. Many towns in the Andes and the Himalaya will face the need to develop alternative water supplies they can ill afford.

Geomorphic hydrology and natural hazards are closely related. In karst terrain, sinkholes can open unexpectedly, swallowing homes and, in one famous case, a Porsche dealership—cars and all. Around the world, floods cause billions of dollars of damage yearly, often killing or displacing large numbers of people as channel locations change and water and sediment flow overbank. In the United States, floods are the number one geologic hazard in terms of dollars lost. Water supply has been and continues to be a critical geopolitical issue. Geomorphologists, with their long-term, broad-scale view on the landscape, are well equipped to advise society about issues related to reducing the hazard from large-magnitude, infrequent natural events.

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DIGGING DEEPER Humans, Hydrology, and Landscape Change—What's the Connection?

Seven billion people occupy Earth and their actions are changing our planet's landscapes [**Figure DD4.1**] (Hooke, 2000). Using a series of model-based calculations and some data, Hooke (1994) argued that humans have become the most effective geomorphic agent on Earth. Not only are people active geomorphic agents, but their actions alter both hydrology and sediment transport at local and global scales. Clearing native vegetation for agriculture and

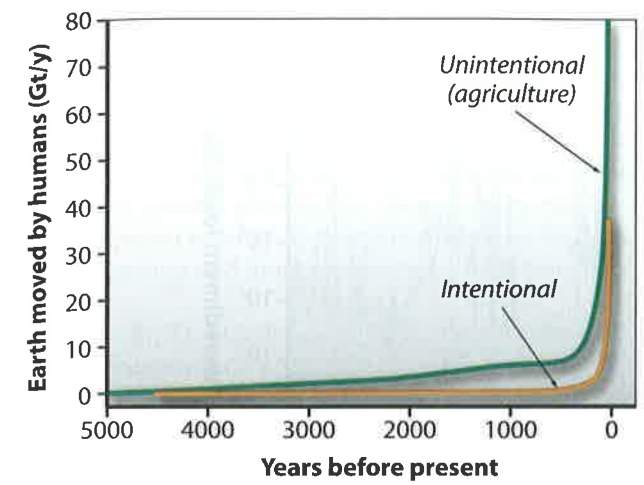

FIGURE DD4.1 Using a variety of assumptions and different data sets, Hooke estimated the amount of earth moved by humans over the past several thousand years in units of gigatons per year (a gigaton is 10^{12} kg). He differentiated movement of soil for agriculture (tillage) with that moved for construction (intentional). The dramatic increase in earth moving corresponds to the Industrial Revolution and the advent of cheap, easily available energy from fossil fuels. [From Hooke (2000).]

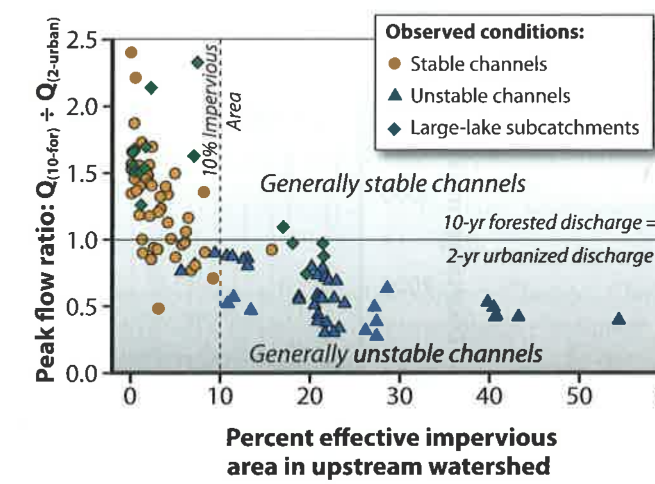

FIGURE DD4.2 Examining stream morphology, Booth et al. (2002) found that in watersheds where more than 10 percent of the land had been converted to impervious cover (such as roads, buildings, and parking lots), stream channels were unstable, and streams were incising, eroding their banks, and migrating laterally. Channel instability increased dramatically when the 2-year recurrence interval peak discharge after development (2-urban) exceeded the 10-year recurrence interval discharge (10-for) before development. [From Booth et al. (2002).]

through urbanization reduces infiltration rates and changes sediment yields as well as the locus of erosion, all in a predictable fashion. Damming rivers influences the distribution and timing of water and sediment discharge. Changing climate affects the volume, intensity, and spatial distribution of storms, precipitation, and runoff (IPCC, 2007).

How do we know the effects of landscape change on rivers, their channels, and the sediment loads they carry? Wolman (1967) documented the hydrologic and resulting geomorphic effects of urban land-use change by surveying channels and collecting sediment yield data before, during, and after development. In a now-classic schematic diagram, he showed how the yield of sediment changes over time as progressive development of a humid, vegetated landscape alters hydrologic flowpaths and the availability of sediment for transport (see Figure 7.12).

Initially, sediment yields are low under native forest cover as rainfall infiltrates and stream banks are stable. Sediment yields rise as agriculture disturbs the land surface and reduces infiltration. More sediment pours off the landscape during urbanization as construction exposes easily eroded soil and paving decreases infiltration and increases runoff. Finally, with the landscape urbanized, most sediment sources are paved and impermeable or otherwise stabilized with plantings; stormwater flows quickly into remaining channels, which, carrying lots of fast-moving water, rapidly scour their banks, deepening and widening as the channel cross section enlarges (Trimble, 1997). In many cases, the flood that occurred on average once every 10 years under native vegetation occurs after urbanization at least every 2 years, if not more frequently [**Figure DD4.2**] (Booth et al., 2002).

Damming rivers changes the frequency and magnitude of water flows, and thus the movement of sediment through river systems, by attenuating flood flows and, in some cases, augmenting baseflows. In arid regions, the most obvious hydrologic and geomorphic effect of dams on rivers is the creation of lakes in the desert, but there are other, more subtle effects. Doing fieldwork and collecting flow data, Grams et al. (2007) documented major changes that occurred downstream after the Glen Canyon Dam on the Colorado River was closed in 1963; the size and duration of spring floods decreased, baseflow increased, and sediment was trapped behind the dam [**Figure DD4.3**]. There were major geomorphic changes downstream. Starved for sediment, the alluvial channel downstream incised and narrowed; gravel and sandbars disappeared from the channel and the average size of riverbed material increased 80-fold. Floodplains were abandoned and became terraces, providing ideal habitat for nonnative species.

The hydrologic and geomorphic effects of dams in humid regions can also be substantial. Magilligan and Nislow (2001) used flow records from seven rivers in northeastern North America that had been dammed and created two flood frequency-magnitude curves for each flow record, one before and one after damming. The predam record included, on average, 30 years of flow data, and the post-dam record included about 40 years. They found that peak flows decreased an average of 32 percent on impounded rivers, which was not a surprise since most of these were flood control dams. They found major

DIGGING DEEPER Humans, Hydrology, and Landscape Change—What's the Connection? *(continued)*

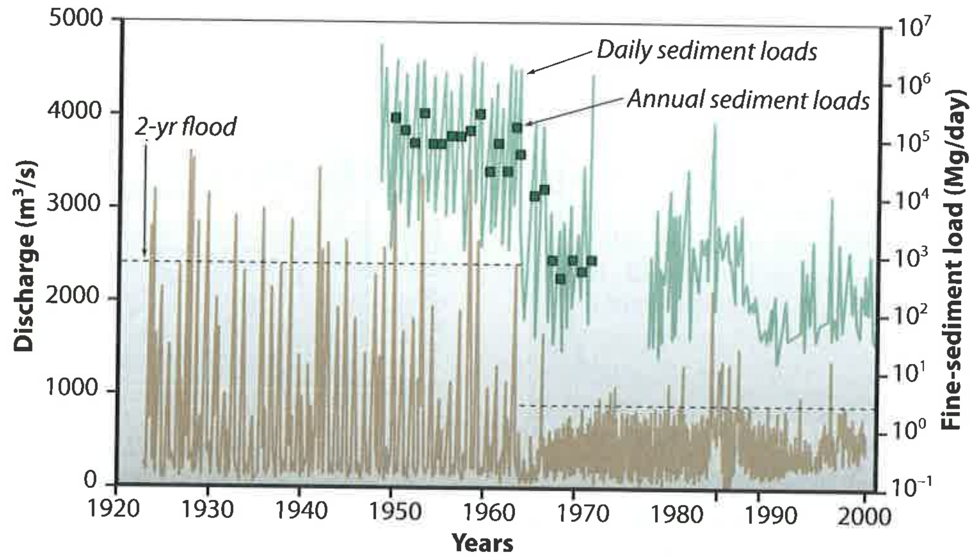

FIGURE DD4.3 When the Glen Canyon Dam closed in 1963, the flow regime and the sediment load of the Colorado River changed dramatically. Flow variability (the brown curve) dropped and sediment load (the light green curve) was diminished by orders of magnitude. The green boxes are the annual fine sediment loads (expressed in Mg/day). The horizontal dashed lines show that the magnitude of the 2-year recurrence flood dropped from 2400 m^3/s to about 800 m^3/s after damming. Look carefully at the two vertical axes (left and right). They are different—discharge is linear, and sediment load is logarithmic. [From Grams et al. (2007).]

changes in the frequency of bankfull discharge, the geomorphically important, channel-shaping flow. In the 40 years since damming, four of the seven channels did not experience a single bankfull flow; on these rivers, water and sediment have not reached the floodplain for decades.

There is overwhelming scientific consensus that humans have changed global climate and that climate change has altered the hydrologic cycle (IPCC, 2007). How might these changes in hydrology affect the rate and distribution of Earth surface processes? Over the past 150 years, data show that rivers and streams are freezing later and that the ice is breaking up sooner **[Figure DD4.4]** (Magnuson et al., 2000). In New England, the timing of spring runoff has advanced between one and two weeks in the last 30 years and is predicted to become even earlier (Hayhoe et al., 2007) as snowpacks thin and melt earlier than they used to. Hayhoe et al. (2007) predict that if warming continues, by the end of this century there will no longer be winter snowpacks in southern New England and therefore spring snowmelt floods will be a thing of the past. Warmer temperatures increase the saturation vapor pressure of water, putting more water into the atmosphere and making for more

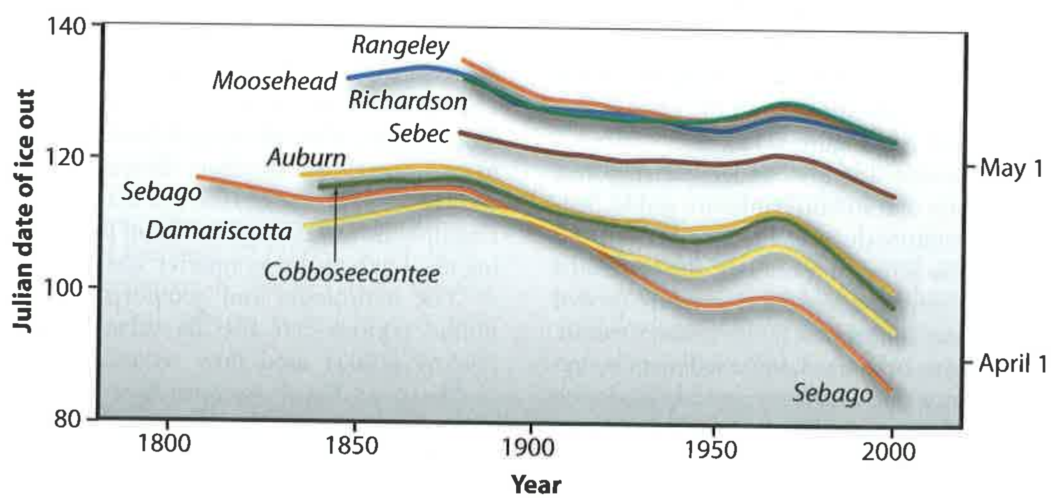

FIGURE DD4.4 Today, ice on lakes in New England and in other regions in the Northern Hemisphere (Magnuson et al., 2000) is melting and breaking up earlier than it did a century ago. These curves show the number of days after January 1 (the Julian date) that ice left each of these lakes in Maine. The yearly data are shown as smooth curves. [From Hodgkins et al. (2002).]

powerful storms. Indeed, Emanuel (2005) compiled meteorological data and showed that over the past 30 years, hurricanes are releasing increasingly more energy (wind speed integrated over time) and have greater maximum power release (the cube of the storms' maximum wind speeds). These enhanced storms now have the potential to cause more coastal geomorphic change; in particular, the increased power of tropical storms may increase the amount of beach erosion and longshore sediment transport.

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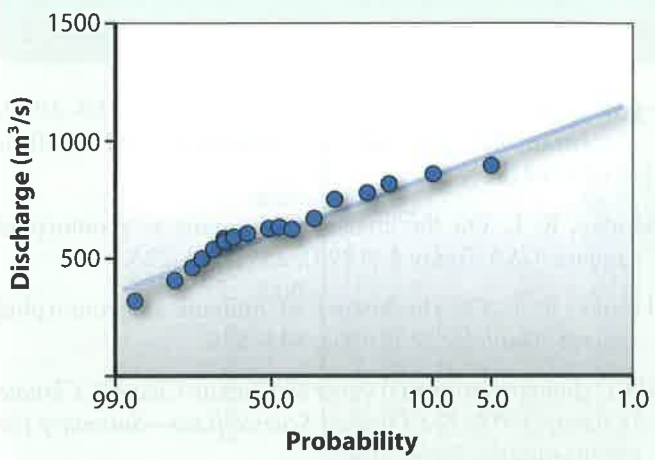
WORKED PROBLEM

Question: The table below lists the mean annual flood discharges between 1970 and 1989 for the Winooski River, which drains 2700 km^2 of northern Vermont. Based on these data, how large is the flood that occurs on average every 10 years (the flood with the 10 percent annual chance of recurrence)? The 1927 flood, the flood of record for the Winooski River, had an estimated discharge of 3200 cubic meters per second (m^3/s). Extrapolating the short flood record, what is the estimated discharge of the 100-year flood (1 percent recurrence probability) and how does it compare to the discharge recorded in 1927? In what season does the annual maximum flood most often occur? Why?

Annual maximum flood discharges, Winooski River at Essex, Vermont

Year	Date	Discharge (m^3/s)
1970	25 Feb	581
1971	4 May	586
1972	5 May	776
1973	1 Jul	813
1974	22 Dec	651
1975	2 Apr	629
1976	2 Apr	790
1977	14 Mar	906
1978	17 Oct	612
1979	6 Mar	702
1980	27 Nov	331
1981	25 Feb	521
1982	18 Apr	858
1983	25 Apr	493
1984	14 Dec	603
1985	3 Dec	396
1986	31 Mar	572
1987	1 Apr	790
1988	29 Apr	473
1989	7 May	575



For the Winooski River at Essex, Vermont, the annual maximum flood series for 1970 to 1989 is plotted on probability axes that make the record linear and allows for extrapolation. Extrapolating from the 19-year record to the 1:100 event (the 100-year return flood; 1 percent annual probability) suggests that flood would be about 1200 m^3/second.

Answer: First, rank the flood data. Then calculate a probability and recurrence interval for each flood using equations 4.1 and 4.2. The discharge of the 10 percent annual probability flood is ~860 m^3/s and it occurred in 1982. Use probability graph paper and plot the flood data, connecting them with a line. If you do this, you find that the 100-year flood (probability of 1 percent) would be estimated, on the basis of this short record, to have a flow of about 1200 m^3/s. The 1927 flood, with its discharge of 3200 m^3/s, is more than 3 times higher than any flood experienced in the short, 10-year record. This analysis clearly demonstrates the uncertainty of extrapolating a short discharge record to estimate the magnitude of extreme events. The annual maximum floods on the Winooski River tend to happen in late winter and early spring; this is typical for rivers draining areas that receive significant winter snowfall. These winter and spring floods are often driven by rain-on-snow events. The runoff is a mix of snowmelt and rainfall onto saturated ground.

KNOWLEDGE ASSESSMENT Chapter 4

1. Explain why each of water's three phases is geomorphically important.
2. Explain the difference between the intensity and duration of precipitation.
3. Give an example of how thresholds are important in geomorphology.
4. Explain a recurrence interval in words stating the underlying assumptions.
5. Give the formula for calculating recurrence intervals.
6. What is stationarity?
7. List several types of weather systems that can deliver precipitation.
8. Explain why the geomorphic effects differ depending on the type of weather system delivering the precipitation.
9. Define evapotranspiration.
10. Provide examples of how climate affects the geomorphology of streams and rivers.
11. Explain why the seasonal distribution of runoff differs between monsoonal climates and cold region climates.
12. Predict how evapotranspiration will change with climate.
13. Explain why evapotranspiration is geomorphically important.
14. Explain the difference between actual and potential rates of evapotranspiration.
15. Give an example of a place where potential and actual rates of evapotranspiration are different.
16. Explain the hydrologic effects of removing trees from a drainage basin.
17. List factors that can affect the infiltration rate.
18. Explain how macropores affect the movement of water across a landscape.
19. How do the vadose zone and phreatic zone differ?
20. Explain why some very porous materials have low permeability.
21. Contrast the causes of primary and secondary porosity.
22. How is Darcy's Law used?
23. Provide the formula for Darcy's Law.
24. Define hydraulic conductivity and explain how it varies between different Earth materials.
25. Explain how Horton overland flow differs from saturation overland flow.
26. Explain the variable source area concept.
27. Using the variable source area concept, explain how and why flowpaths will differ by season.

28. What are mottles and what do they tell geomorphologists about local hydrology?
29. Draw a labeled diagram of a hydrograph.
30. Predict how lag-to-peak will differ as a function of basin geomorphology.
31. Predict how the hydrograph will change as a basin is urbanized.
32. What is hysteresis?
33. What is a chemograph?
34. Why do sedigraphs exhibit hysteresis?
35. Where would you go to find a losing stream?
36. How does detention storage influence flood peaks?
37. Define bankfull flow and explain why it is important in shaping channels.
38. What is a water budget, how would you create one, and why could it be geomorphically important?
39. How do surface water and groundwater systems interact?
40. How does the flow system in a karst terrain differ from the flow system in a terrain underlain by granite rock?
41. List four common karst landforms and explain how each forms.