CHAPTER 5 DRAINAGE BASINS
From Ritter et al., 1995

Introduction

Two basic generalizations about rivers were realized long before geomorphology emerged as an organized science: (1) streams form the valleys in which they flow, and (2) every river consists of a major trunk segment fed by a number of mutually adjusted branches that diminish in size away from the main stem. The many tributaries define a network of channels that drain water from a discernible, finite area which is the **drainage basin,** or **watershed,** of the trunk river.

The drainage basin is the fundamental landscape unit concerned with the collection and distribution of water and sediment. Each basin is separated from its neighbor by a **divide,** or **interfluve.** Thus, the basin can be viewed as a geomorphic system or unit. As we will soon see, the basin is inexorably linked with hillslope processes that contribute water and sediment to the channel network in accord with the regional climate, underlying bedrock and tectonic regime, and land use by humans (fig. 5.1). Any feature or portion of the basin can be considered a subsystem having its own unique set of processes, geology, and energy gains and losses. Furthermore, because it is possible to measure the amount of water entering the basin as precipitation and the volume leaving the basin as stream discharge, hydrologic events can be readily analyzed on a basinal scale. Likewise, much of the sediment produced within the basin is ultimately exported from the basin through the trunk river. Thus, considered on a long temporal scale, the rate of lowering of the basin surface can be estimated.

The output from a given basin compartment serves as input to the master channel and influences downstream channel characteristics and hydrologic processes in rivers. The mechanics of fluvial processes usually reflect some balance between the amount of sediment supplied for transport and the water available to accomplish this task. Throughout the discussion of drainage basins and fluvial systems, we will frequently refer to the concepts illustrated in figure 5.1 as we describe the interrelationships between various components of the fluvial system and the regulatory influence of the external variables of water and sediment in the adjustment and evolution of basins and channels.

Most Earth scientists are introduced to watersheds when they learn that drainage patterns or individual stream patterns commonly mirror certain traits of the underlying geology, described in figure 5.2 and table 5.1. Because the gross character of these patterns is evident on topographic maps and aerial photos, the patterns are useful for structural interpretation (Howard 1967) and for approximating lithology in a study of regional geology.

In a hydrologic sense, however, prior to World War II most basins were described in qualitative terms such as well-drained or poorly-drained, or they were connoted descriptively in the Davisian scheme as youthful, mature, or old. The mechanics of how river channels or networks actually form and how water gets into a channel was poorly understood by geologists and

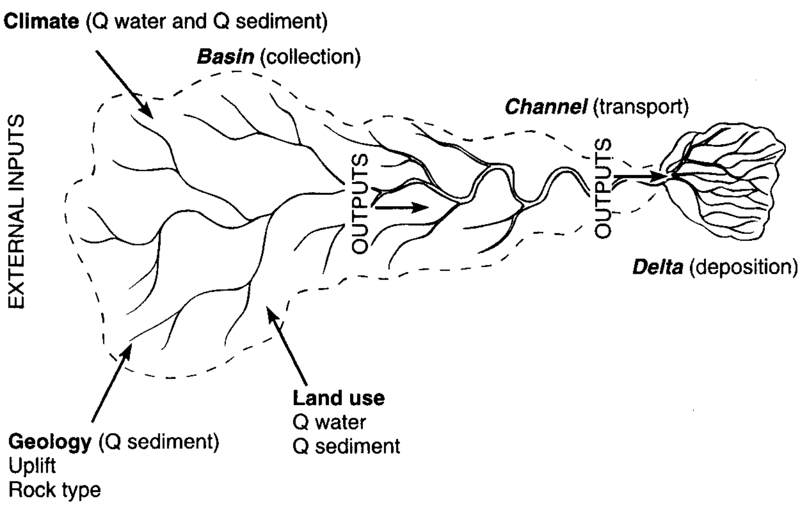


FIGURE 5.1
Schematic surface components of the fluvial system. The tributaries provide links between lithology and climate and are adjusted to both. Channel characteristics vary in response to the external variables of sediment and water discharge (Q), which are influenced naturally from climate, tectonic, and lithologic factors. Human influence also modifies these variables through land use alterations.

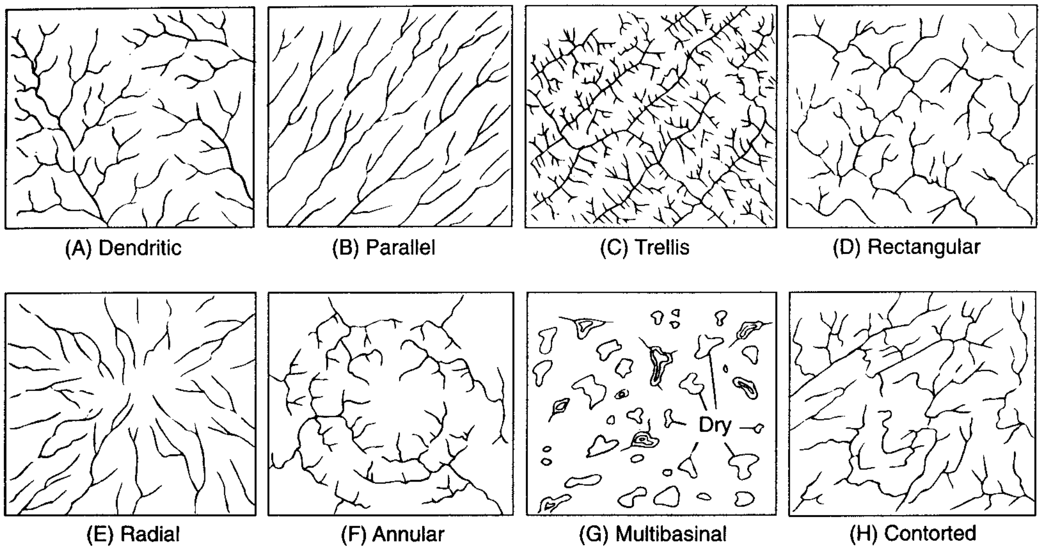


FIGURE 5.2
Basic drainage patterns. Descriptions are given in table 5.1.
(Howard 1967, reprinted by permission)

TABLE 5.1 Descriptions and characteristics of basic drainage patterns illustrated in figure 5.2.

Pattern	Geological Significance
Dendritic	Horizontal sediments or beveled, uniformly resistant, crystalline rocks. Gentle regional slope at present or at time of drainage inception. Type pattern resembles spreading oak or chestnut tree.
Parallel	Generally indicates moderate to steep slopes but also found in areas of parallel, elongate landforms. All transitions possible between this pattern and dendritic and trellis patterns.
Trellis	Dipping or folded sedimentary, volcanic, or low-grade metasedimentary rocks; areas of parallel fractures; exposed lake or seafloors ribbed by beach ridges. All transitions to parallel pattern. Pattern is regarded here as one in which small tributaries are essentially same size on opposite sides of long parallel subsequent streams.
Rectangular	Joints and/or faults at right angles. Lacks orderly repetitive quality of trellis pattern; streams and divides lack regional continuity.
Radial	Volcanoes, domes, and erosion residuals. A complex of radial patterns in a volcanic field might be called multiradial.
Annular	Structural domes and basins, diatremes, and possibly stocks.
Multibasinal	Hummocky surficial deposits; differentially scoured or deflated bedrock; areas of recent volcanism, limestone solution, and permafrost. This descriptive term is suggested for all multiple-depression patterns whose exact origins are unknown.
Contorted	Contorted, coarsely layered metamorphic rocks. Dikes, veins, and migmatized bands provide the resistant layers in some areas. Pattern differs from recurved trellis in lack of regional orderliness, discontinuity of ridges and valleys, and generally smaller scale.

From Howard, 1967, reprinted by permission.

hydrologists alike. This early twentieth-century view of streams and drainages contrasts markedly with the avant-garde approach presented by R. E. Horton during the latter part of this period (Horton 1933, 1945). His attempt to explain stream origins in mathematical terms and to describe basin hydrology as a function of statistical laws marked the birth of quantitative geomorphology. We now know that many of Horton's original ideas are only partially correct. Still, modern geomorphic analysis of drainage basins has its roots in Horton's original work, and his thinking has been instrumental in the development of modern geomorphology.

Slope Hydrology and Runoff Generation

The ultimate source of river flow is, of course, precipitation, which represents the major influx of water to any drainage basin. Precisely how much of that precipitation actually becomes part of the streamflow and what route a particular drop of water follows to reach a channel are topics of great concern to hydrologists. The components of the slope hydrological cycle (fig. 5.3) are the pathways of water to the streams.

Rainfall seldom makes direct contact with the bedrock or soil surface except in arid regions characterized by sparse vegetal cover. Most raindrops are impeded by leaves and trunks of the vegetal cover, in a process known as **interception.** Interception substantially reduces not only the erosive potential of raindrop strike but also the volume of water reaching the surface. Interception losses are quite variable because they depend on numerous hydrometeorological factors, vegetation type, land use, and seasonality. In addition, the loss is dependent on storm duration, being initially high and decreasing as the storm continues. During long storms, the vegetation may become saturated and all additional rainfall is passed on to the surface. Nonetheless, interception typically removes 10 to 20 percent of precipitation where grasses and crops are the dominant vegetation and up to 50 percent under a forest canopy (Selby 1982). Rainwater that does reach the ground may still be prevented from contributing to streamflow by vegetation because some of it is consumed and lost by **evapotranspiration.**

Infiltration

Water on the ground surface may follow different routes to a stream channel. Water flowing into a channel in direct response to a precipitation event is called **storm runoff** or **direct runoff,** whereas the process of water entering the soil is considered **infiltration.** In 1933 Horton suggested that the rate of infiltration into soil on a sloping surface actually governs the amount of storm runoff generated by rainwater. This infiltration theory of runoff can be described as follows. Imagine a hypothetical slope with a surficial mantle of regolith formed by weathering and mass wasting processes. As precipitation falls on the slope surface, water is absorbed into the ground at a rate called the **infiltration capacity.** Infiltration capacity, typically measured in mm/hr is a function of regolith or soil thickness, soil texture, soil structure, vegetation, and the antecedent condition of the soil moisture (which depends on the recent history of precipitation). The rate of infiltration depends primarily on the interaction between three processes: (1) absorption or entry of water into the soil, (2) storage of water in the pore spaces, and (3) transmission of water downward through the soil. Infiltration rates are generally maximized when slopes are gentle with permeable soils and when rainfall intensity is low.

Recent studies of basins within the same climatic setting have shown that runoff processes, and thereby the surface hydrologic parameters such as discharge of water and sediment discharge by channels, are significantly impacted by variations in basin lithology. Kelson and Wells (1989) observed that unit runoff was consistently higher from basins underlain by resistant crystalline rocks than from neighboring watersheds underlain by less resistant sedimentary rocks in the

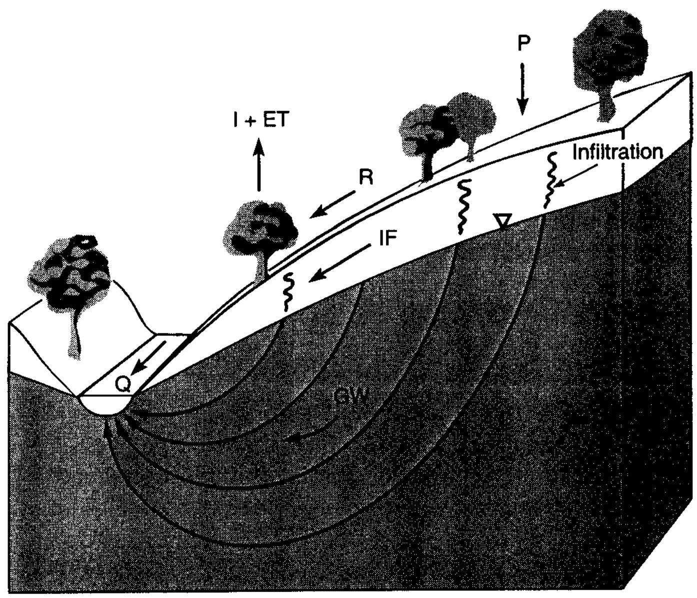


FIGURE 5.3
Slope hydrologic cycle. Some of the precipitation (*P*) is intercepted by vegetation (*I*) or lost by evapotranspiration (*ET*). Upon reaching the ground surface, it becomes part of stream discharge (Q) by direct runoff (*R*), interflow (*IF*), or groundwater flow (*GW*) after it reaches the water table (Δ).

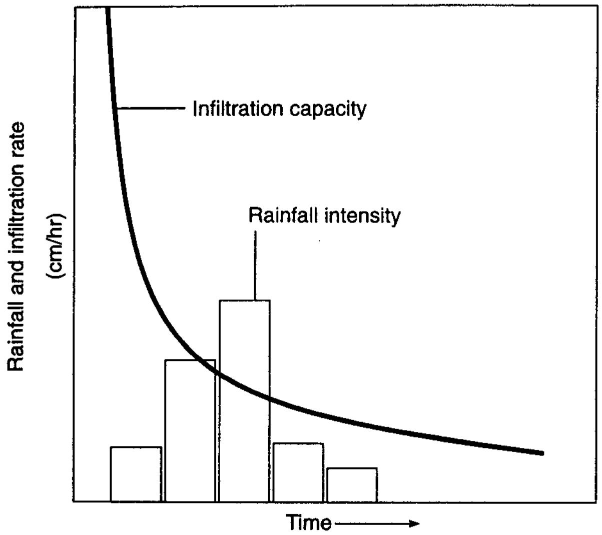


FIGURE 5.4
Infiltration capacity and rainfall intensity plotted against time. Infiltration capacity decreases with duration of storm. Runoff occurs only when rainfall intensity is greater than infiltration capacity.

mountains of northern New Mexico. Differences in regolith production from weathering of the parent lithology and from earlier geomorphic activity (i.e., glaciation) contributed to differences in the timing of water inputs into the basin, variations in subsurface slope, infiltration capacity, and the sediment yield and erosive power of the basin streams. Sala (1988), in a comparative study of watersheds underlain by slate and crystalline rocks in Spain, found that regolith variations attributed to the lithology not only resulted in notable differences in runoff but also produced general differences in process dominance on basin hillslopes. Slate watersheds generated large amounts of coarse debris and were dominated by creep, leading to high erosion rates. On the other hand, crystalline basins were characterized by well-integrated runoff systems with perennial flow due to thicker regolith composed of grus. Therefore, infiltration capacity varies spatially on a large scale with regional geology, but may also change locally (even along the same slope) if the controlling factors vary.

Infiltration capacity normally changes at a site during any precipitation event (fig. 5.4). It usually starts with a high value that decreases rapidly during the first few hours of the storm and then more slowly as rainfall continues, until it finally attains a reasonably constant minimum value. Typically, the ultimate infiltration rate is established by a limiting subsurface horizon with a low water transmission rate, such as a zone of textural B horizon, caliche, or the bedrock surface. The infiltration capacity changes because surface conditions change during a storm, especially as aggregated soil clumps are broken apart into smaller particles which clog some of the pores. In the time interval between rains, the infiltration capacity rises again as the surface dries and reestablishes its aggregated structure, and storage space within the soil increases again as soil water gradually drains downward. The frequency of rainfall, expressed as time between successive rains, can also become a significant factor affecting infiltration. Rains of relatively minor intensity may trigger disastrous floods if the infiltration capacity does not have enough time to recover between storms.

In Horton's model, as long as the infiltration capacity exceeds the rate at which rainfall strikes the surface (known as *rainfall intensity*), all incoming water will be infiltrated and none will run off. However, in those periods when rainfall intensity exceeds the infiltration capacity, runoff will occur as water moving down the slope surface; this runoff is known as **Hortonian overland flow.** Hortonian overland flow has been considered to be the primary determinant of peak flow and total direct runoff in a stream channel from a storm because its flow velocity generally ranges between 10 and 500 m/hr (Dunne 1978). The basic assumption here is that all infiltrated water is delayed greatly in its procession to the stream channels because it must percolate downward to the groundwater table and then move toward the stream by comparably slow groundwater flow velocities (see fig. 5.3). Overland flow, which occurs as broad shallow sheets or in linear depressions called *rills* is uncommon, however, except in sparsely vegetated semiarid and arid regions. Hortonian flow is virtually nonexistent in humid-temperate regions characterized by dense vegetation and thick soils (Kirkby and Chorley 1967; Dunne and Black 1970a, 1970b).

In light of the above, we must ask the logical question as to where the water comes from to produce the rapid peak flows observed in small basins of humid regions if all rainwater reaching the surface is subject to infiltration.

Subsurface Stormflow and Saturated Overland Flow

A major flaw in Horton's original model was the belief that infiltrated water moved directly downward under the influence of gravity. Research subsequent to Horton's analyses has clearly demonstrated a variety of flow within the unsaturated zone, due to the anisotropic character of regolith in most areas with respect to its ability to transmit water. Thus, significant lateral and subhorizontal flow can occur in some areas. Infiltrating water moves downward from the surface as a wetting front to the water table. In doing so, the front often encounters a soil horizon or other zone of low permeability that not only limits infiltration capacity but also tends to divert water downslope (often laterally) along its surface. This occurs because a local saturated condition develops above the low permeability zone, and lateral flow is initiated parallel to the barrier. This type of movement, called **throughflow** or **interflow** (Kirkby and Chorley 1967), can occur above the water table and permits water to take a more direct path to the stream channel than normal groundwater does. Throughflow is probably most common in the more permeable A horizon.

Where no barrier exists, infiltrated water will reach the water table and elevate its position. The water table rises rapidly adjacent to the channel, where antecedent soil moisture is greatest, and slowly in the upper-slope zone. As a result, the water table steepens immediately next to the channel and generates accelerated groundwater flow in that area. The combination of throughflow and accelerated groundwater movement is called **subsurface stormflow.** It was first recognized by Hursh (1936; Hursh and Brater 1941), but the significance of the concept was not fully realized until much later.

Subsurface stormflow from the lower parts of slopes may actually produce runoff in the form of bank seepage early in a storm. Therefore, it is possible for some of the subsurface stormflow to contribute to peak discharge. In other situations, storm runoff is generated from the movement of subsurface water through **macropores,** which are linear openings, having a much greater permeability than the surrounding regolith (Mosley 1979; Bevan and Germann 1982). Macropores can originate along root channels by **piping** (flow concentration and erosion along the rigid root) or in voids left by decayed roots, burrows, and other cavities made by plants and animals. Flow through such pipes can be rapid and turbulent, accounting for as much as one-fifth of the annual contribution to runoff from hillslopes (Roberge and Plamondon 1987). Although some water may enter channels rapidly by these processes, most studies show subsurface flow to be a more important contributor to streamflow after the discharge peak, because measured velocities of subsurface flow range between 0.003 and 1.0 cm/hr (Dunne 1978), too slow to effectively contribute to most peak flows in small basins. This slow response requires yet another source of runoff to help explain the rapid production of peak flood flow. A field experimental study in a forested Appalachian watershed reported distinguishable inputs into surface runoff from throughflow sources (Hornberger et al. 1991). Significant contributions to storm runoff may occur where soil conductivity (with respect to its ability to transmit water) is high due to coarse texture and/or the presence of large macropores (Dunne 1983).

Detailed studies in Vermont (Ragan 1968; Dunne and Black 1970a, 1970b) showed that in many storms the level of the water table rises until it actually intersects the ground surface near the stream channel (fig. 5.5). In addition, zones of throughflow partially fed by laterally

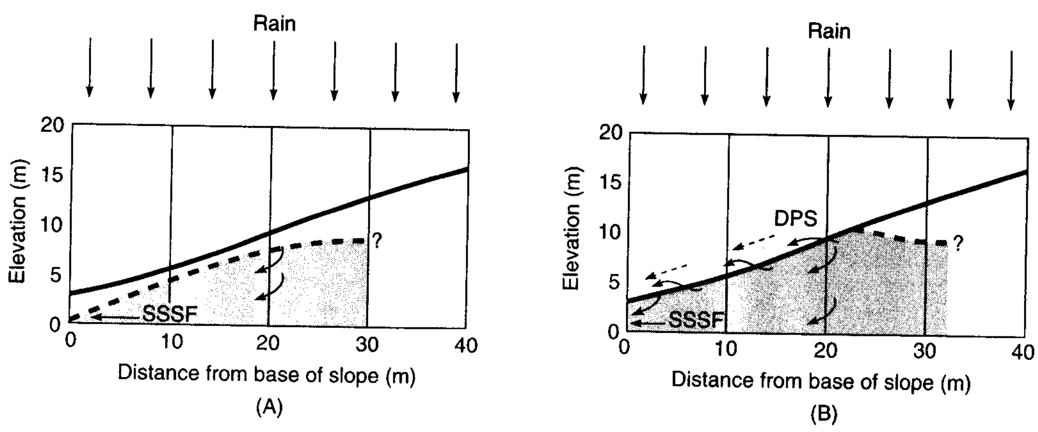


FIGURE 5.5

Runoff and interflow in a steep, well-drained Vermont hillslope. (A) Early in a storm the saturated zone (shaded) yields a small amount of subsurface stormflow (SSSF). (B) Late in the storm the water table rises to the surface as return flow (RF). Precipitation on the saturated area (DPS) adds to the return flow.

(Dunne and Leopold 1978)

moving, infiltrated water may become saturated to the surface (Kirkby and Chorley 1967). In this situation some of the infiltrated water is returned once again to the surface and begins to flow downslope toward the channel; this flow is aptly called return flow. **Return flow** is obviously a form of overland flow, but its origin and location is distinctly different from Hortonian overland flow. The velocity of return flow is much greater than subsurface stormflow, attaining speeds of 3 to 15 cm/sec (Dunne 1978). In addition, direct precipitation on the saturated area marked by return flow or zones of saturated throughflow adds significant amounts of water to the volumes of direct runoff.

The combination of return flow and direct precipitation on saturated areas, called **saturation overland flow,** is now documented in many humid areas as the major contributor to direct runoff to stream channels. For example, Ragan (1968) estimated that on the average it supplied 55 to 62 percent of the total storm runoff in a small watershed near Burlington, Vermont, and was predominant in determining peak discharge. Subsurface storm flow then provided 36 to 43 percent of the total flow but exerted its greatest influence after the peak during the recessional phase of the runoff.

There are many sources of direct runoff other than Hortonian overland flow. Moreover, it is becoming apparent that the area of a watershed actually providing runoff during a storm is not temporally constant. This perception, known as the *variable source concept* or *partial area concept*, indicates that the area over which quick runoff occurs varies seasonally and during any given storm (fig. 5.6). This change is fundamentally controlled by topography, soil characteristics, antecedent moisture, and rainfall properties. Areas with moderate to poorly drained soils, gentle slopes, and concave recessions along the valley walls are prone to have the greatest expansion of contributing areas during storms and on a seasonal basis.

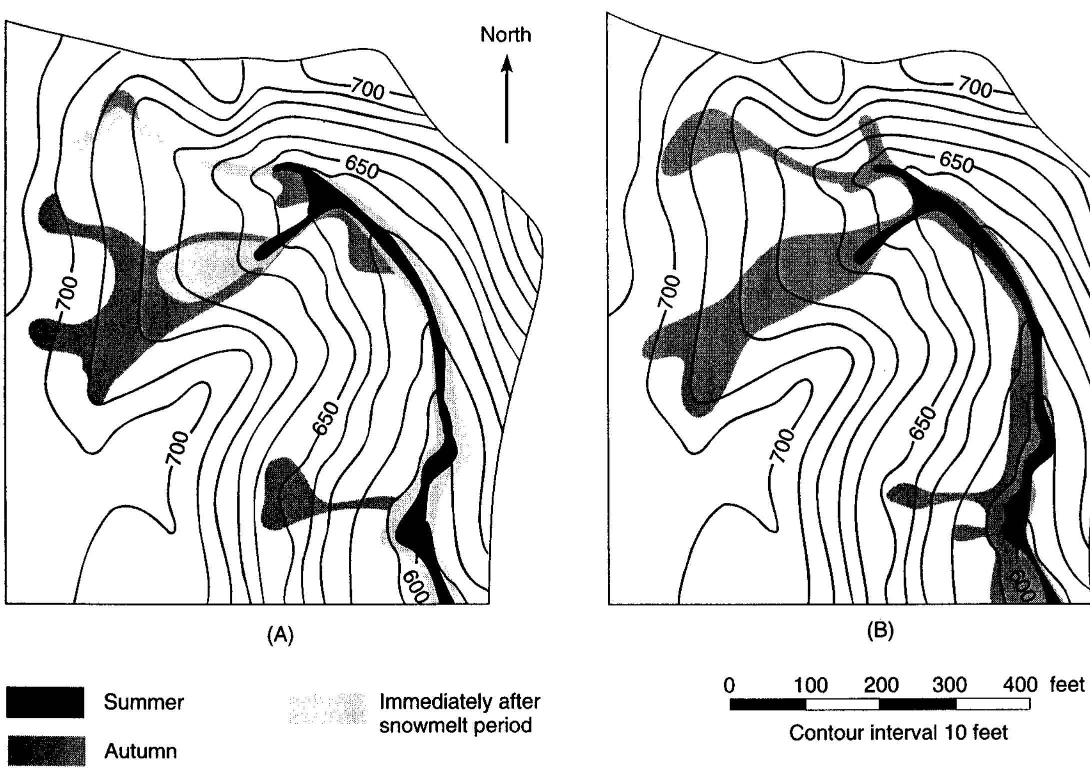


FIGURE 5.6

Variations in saturated areas on well-drained hillslopes near Danville, Vt. (A) Seasonal changes of prestorm saturated area. (B) Expansion of saturated area during a single 46 mm rainstorm. Solid black represents beginning of storm. Light shade represents saturated area at end of storm where water table has risen to the surface.

(Dunne and Leopold 1978)

Although the generalizations discussed above seem reasonable, they are based on a limited number of studies, and much more work is needed to refine the model (Selby 1982). Nonetheless, the observations about runoff generation have an important bearing on the geomorphic processes operating within a drainage basin, and further research in this field should be vigorously pursued (Freeze 1980).

The Stream Hydrograph and Response to Basin Characteristics

Hydrologists have always been concerned about how much runoff will issue from any precipitation event and how quickly the runoff will enter the stream channels. These factors determine if and when a flood will occur and what height the river will reach at peak flow. Figure 5.7 compartmentalizes the hydrologic processes that contribute to flood runoff in a stream channel. The response of the stream to a storm is depicted graphically as a **flood hydrograph** (or storm hydrograph) which shows the passage of flood flow volume, or *discharge*, with time (fig. 5.8). Discharge can be separated, using one of numerous separation techniques (see Chow 1964), into **direct runoff** and **base flow.** Runoff represents the sum of numerous types of overland flow, interflow, and stormflow, whereas base flow arises from contributions of groundwater spread out over longer periods of time. Whereas discharge may contain elements of both if measured shortly after a storm, base flow is the sole source of water sustaining river flow in dry periods between storms.

The maximum or peak flow usually develops soon after the precipitation ends, separating the hydrograph into two distinct segments, the *rising limb* and the *recession limb.* **Lag time** is the time between the center of mass of the rainfall and the center of mass of the runoff. The rising limb generally reflects the input from direct runoff; hence, the increase in discharge with time is relatively rapid. The recessional phase, however, is controlled more by the gradual depletion of water temporarily stored in the shallow subsurface

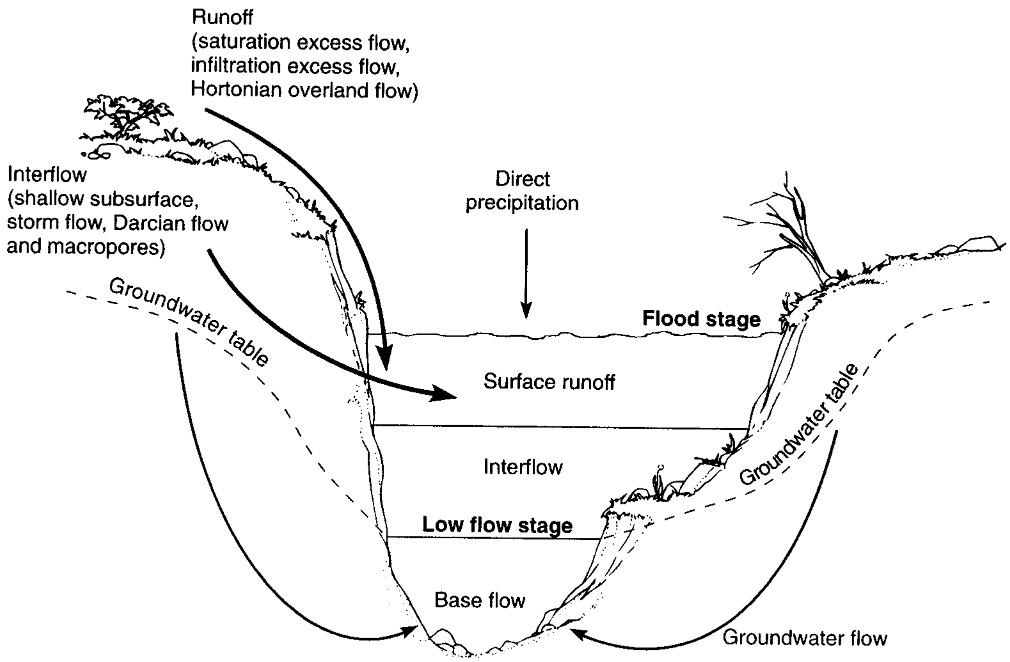


FIGURE 5.7
Schematic of stream channel showing major kinds of water contributions. Arrival of water from any given precipitation event is progressively delayed from runoff to interflow to groundwater flow.
(Based on Kochel 1992)

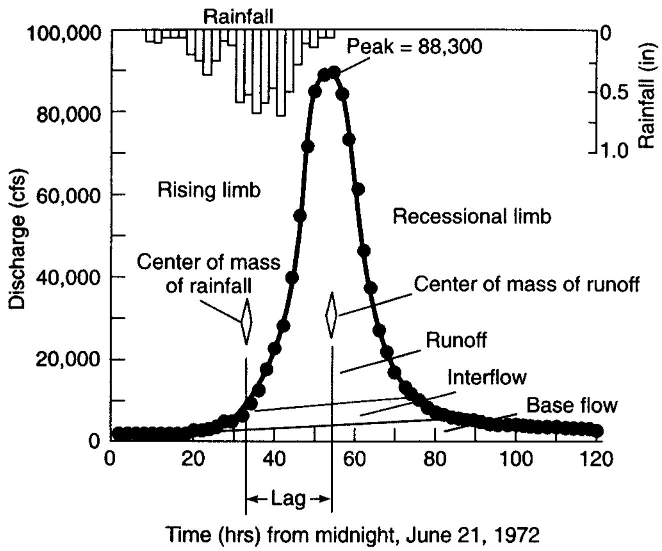


FIGURE 5.8
Flood hydrograph of the Hurricane Agnes flood of June 1972 on the Conestoga River at Lancaster, Pa. Although the curve is rather symmetrical, most hydrographs show significant skewness with a broader recessional limb reflecting interflow and groundwater inputs after a storm.

system, and thus the decrease is less pronounced with time. This phenomenon normally results in an asymmetrical hydrograph, unlike the one shown in figure 5.8. In addition, the shape of the recessional limb is not influenced by the properties of the storm causing the increased discharge but is more closely related to the physical character of the basin.

Effect of Physical Basin Characteristics

The physical characteristics of a basin contribute to the magnitude of the flood peak. The shape of the flood hydrograph is influenced not only by the temporal and spatial distribution of the rainfall, but in large manner by the physical character of the drainage basin. The flood hydrograph represents the integrated effects of basin area, channel density and geometry (basin morphometry), soils, and land use and thus is the summation of the physical processing of precipitation from the divides to the site of measurement. Sophisticated models have been developed to predict how water is collected, stored, routed, and summed from all parts of a basin to achieve a final output hydrograph (for example, U.S. Army Corps Engineers 1985). If geology and topography are alike throughout an area, then rainfalls having similar properties should generate hydrographs with the same shape. On this premise, a *type* hydrograph for a basin, called a **unit hydrograph,** has been developed, in which the runoff volume is adjusted to the same unit value (i.e., one inch of rainfall spread evenly over the basin over one day). The unit hydrograph has been used as a connecting link in many studies attempting to relate basin morphometry to hydrology. By comparing the shapes of unit hydrographs from different basins, we can see the effects of differences in physical attributes of the basin.

Basin lag is the time needed for a unit mass of rain falling on the basin to be discharged through the measuring point (or basin outlet) as streamflow. Basin lag is normally estimated as the time interval between the centroid of rainfall and the centroid of the flood hydrograph (fig. 5.8) and is comprised of two distinct parts, the time involved in overland flow and the channel-transit time. Lag time for any basin is fairly consistent, with minor variations sometimes caused by the position and movement of the storm center relative to the gaging station. It seems intuitive that lag should increase with the size of the basin, but comparisons of similar-sized basins can show lag times up to three times as *sluggish* as basins with short lag times, which are referred to as *flashy* streams. Lag, therefore, is influenced by parameters other than drainage area alone, such as basin slope, the density of channels in the basin, and soils.

The time elements of basin hydrology can have a monumental influence on the magnitude of peak discharge during floods. Take, for example, the progression of flood crests recorded at several gaging stations in the Susquehanna River basin during the Hurricane Agnes flood in June 1972 (fig. 5.9). Most of the precipitation entered the basin during the period between June 19 and June 22. In the minor tributary Bald Eagle Creek, the flow peaked at 143 cms (5050 cfs) on the night of June 22, indicating a relatively quick response to the storm. In the larger Juniata River, the flood crested about 12 hours later with a considerably higher discharge value of 3538 cms (125,000 cfs). Far downstream on the main Susquehanna River, the flood peak did not occur until 12 hours after the Juniata peak, when discharge rose to 30,564 cms (1,080,000 cfs). Discharge peaked later on the main stream, and at a significantly higher magnitude than in the tributaries.

Lag time and peak discharge are positively correlated with basin size. However, simple expansion in

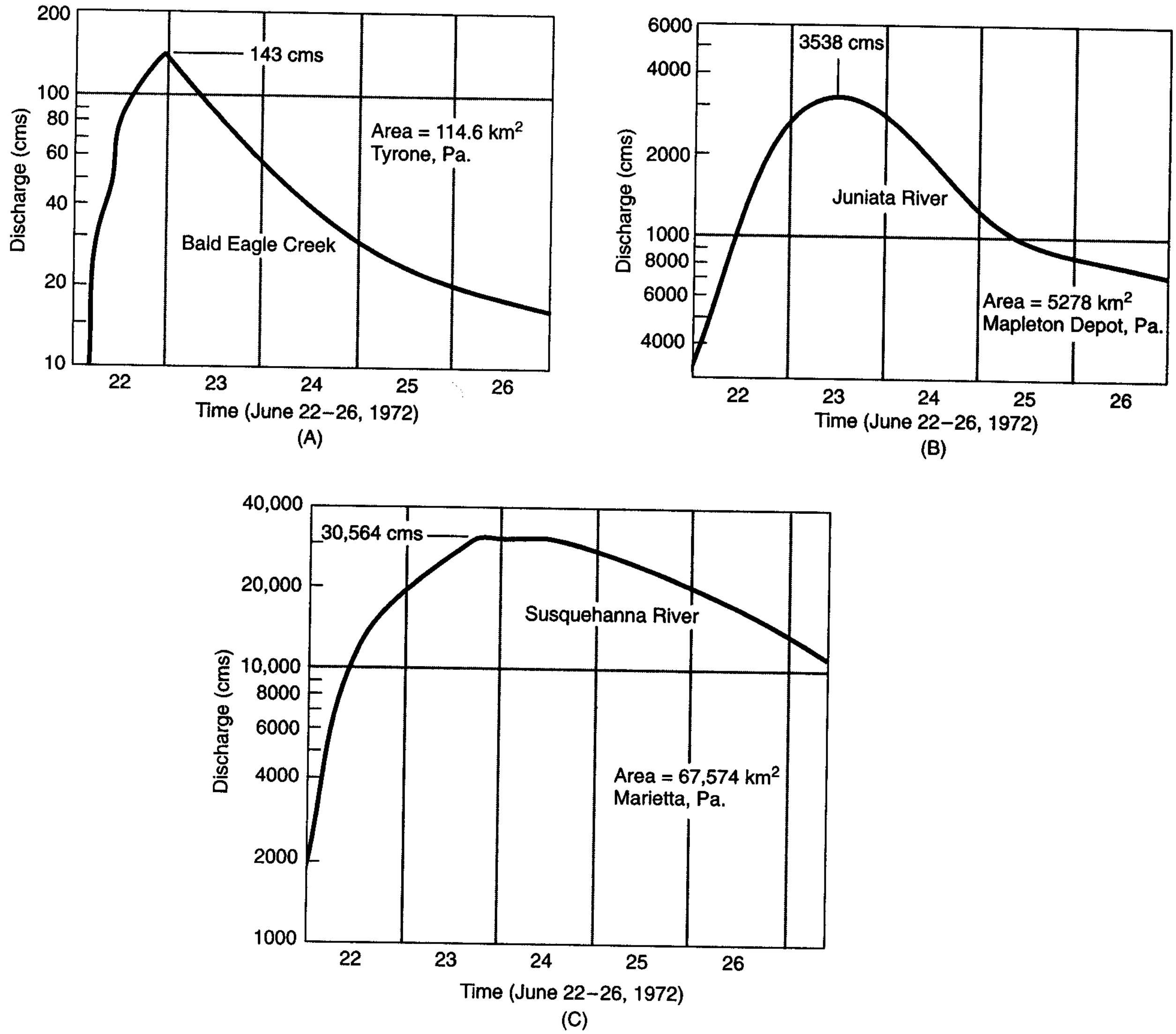
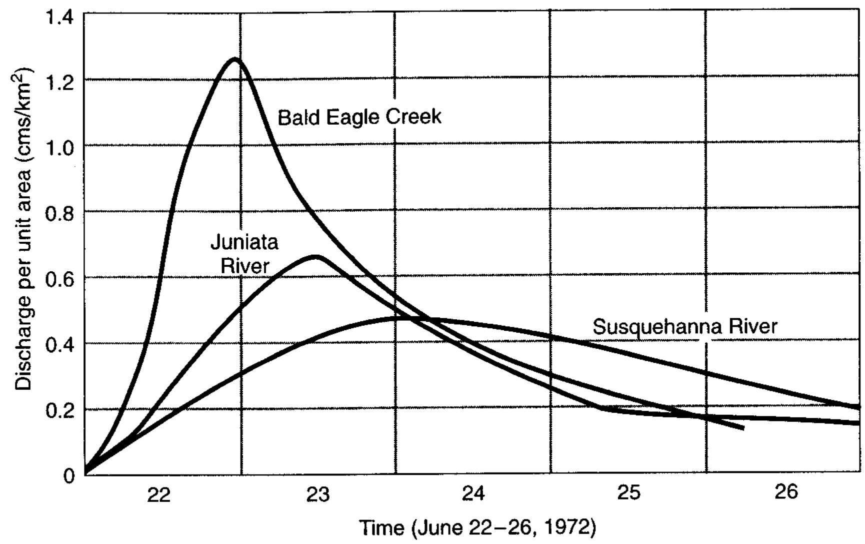


FIGURE 5.9

Progress of flood crests caused by Hurricane Agnes in the Susquehanna River basin, June 1972. Basins of increasing size: (A) Bald Eagle Creek; (B) Juniata River; (C) Susquehanna River.

basin size cannot fully explain the observed flow characteristics during a flood. Using data from figure 5.9 and replotting the peak discharge as Q/km^2, figure 5.10 reveals the interesting hydrologic property that discharge per unit area is much higher for the smallest tributary than for the massive basin of the main stem. Figure 5.11 shows how a similar relationship between basin area and discharge can be useful as a guide for anticipating the maximum flow likely for a basin of a given size. Had the increase in discharge during the 1972 Susquehanna flood been only a function of increased basin area, the peak flow at the Marietta station should have been a simple product of the Q/km^2 value of Bald Eagle Creek times the drainage area above Marietta, or 1.25 cms/km^2 $\times$ 67,574 km^2 = 84,468 cms (2,978,454 cfs), but this is almost three times the actual measured value. Somewhere within the basin, a built-in flood control mechanism exists that not only holds down the magnitude of the peak Q but simultaneously maintains abnormally high flow over a longer period of time. Most of this ability to lower peak flow can be attributed to the storage of water on floodplains and riverine wetlands. Floodplains store or retard large volumes of water until the flood crest

FIGURE 5.10
Discharge per unit area during flood caused by Hurricane Agnes in the Susquehanna River basin, June 1972. Note that the main river has considerably less discharge/area than tributaries. Main stem also peaks later and discharges floodwater over a longer time period, showing the effect of storing water on floodplains.



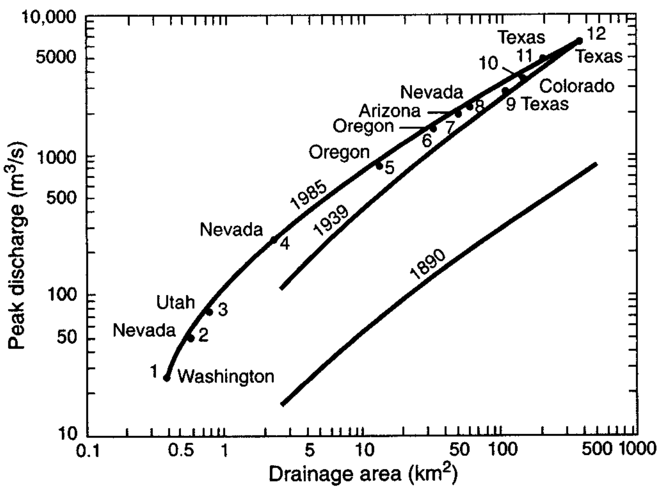


FIGURE 5.11
Peak discharge per unit from the largest flash floods in U.S. history. Note how the relationship indicates an increase in discharge coincident with development and land use alterations over the past 100 years.
(From Costa 1987)

passes a given channel locality. Additionally, differences in flood-crest travel time on the numerous tributaries in the basin and vagaries introduced by variable sources of runoff may help retard the downstream peak flow. Thus, a portion of runoff never contributes to the increasing of peak flow downstream but shows up in the record after the crest has passed as the extended part of the recessional limb of the flood hydrograph.

Initiation of Channels and the Drainage Network

In the Horton model, when rainfall intensity exceeds the infiltration capacity, overland flow occurs and only then does erosion become possible (see fig. 5.4). For a stream channel to develop, the erosive force (F) of the overland flow must surpass the resistance (R) of the surface being eroded. According to Horton (1945), as overland flow begins to traverse the slope, the force it exerts on a soil particle depends on the slope angle, the depth of water, and the specific weight of the water. Actually, F represents a shear stress exerted parallel to the surface by the water. The stress progressively increases downslope because the depth rises as more and more water is added to the volume of overland flow. Depending on the size of the slope material, a threshold stress is eventually reached at some point on the slope where $F > R$, and particles are dislodged or entrained. The processes and magnitude of slope erosion by water will be discussed later in the chapter when we examine sediment yield from drainage basins. Here we will look at the initiation of channels on sloping surfaces.

Surface resistance (R) is affected by the type and density of the vegetal cover (referred to in a combined factor called biomass by Graf 1979). When vegetation intercepts raindrops before they can strike the surface, it preserves cohesion in the aggregated-clay soil structures. In addition, rootlets tend to bind soil particles, and litter often serves as a protective mat above the surface material. Vegetation also inhibits the free flow of water and retards its velocity. In areas devoid of vegetal cover, soil surfaces typically develop a hard crust upon desiccation, which provides high resistance during the initial portion of a precipitation event. This resistance is often destroyed as the storm progresses, and thus is apt to vary with time during rainfall events. The factors affecting resistance are similar to those controlling infiltration capacity.

Erosion by overland flow can, therefore, be considered a threshold process, occurring only after resistance forces are exceeded. Flowing water produces erosional forces commonly referred to as *tractive force* (τ_o), which is dependent on the character of the water, the flow depth, and the slope or gradient of the channel (Graf 1979):

$$\tau_o = \gamma_f D \, \theta c$$

where, γ_f is the specific weight of the fluid, D is flow depth, and θc is the channel gradient. Erosion typically begins as a series of subparallel rills parallel to the slope gradient. Slight variations in surface topography produce greater depth of flow, resulting in increased erosive forces in low spots. Erosion is accelerated at those points. Once water is focused into channels, this positive feedback process promotes continued evolution of channel networks at the expense of unconfined sheet flow. The actual point where rill formation begins depends on how efficiently the force of overland flow increases as the water moves downhill. This ultimately relates to infiltration capacity, rainfall intensity, and the resulting rate of runoff, or *runoff intensity.*

Assuming a constant slope and runoff intensity, the distance between the watershed divide and the upper boundary of the rills is a measurable segment called the critical length (X_c) (fig. 5.12). Horton considered the critical length to be the most important single factor in the development of stream networks, but its significance can perhaps be appreciated more fully on a local scale because it is highly sensitive to changes in the factors that control erosion. Assume, for example, that a farmer wishes to plant additional row crops in an area near the divide. As the land is cleared of its original vegetal cover and replaced by the crops, resistance and infiltration capacity are drastically lowered. Under prevailing precipitation more runoff occurs, the force of overland flow increases, X_c is shortened, and rills and gullies form in the newly cropped region. Depending on their depth, these gullies may make the area impassable for the heavy equipment needed for plowing and harvesting. A useful review of gully erosion processes and models is found in Bocco (1991).

The analysis just presented is probably applicable in areas that are prone to the development of Hortonian overland flow. In humid-temperate regions, however, overland flow emanating from upper slope

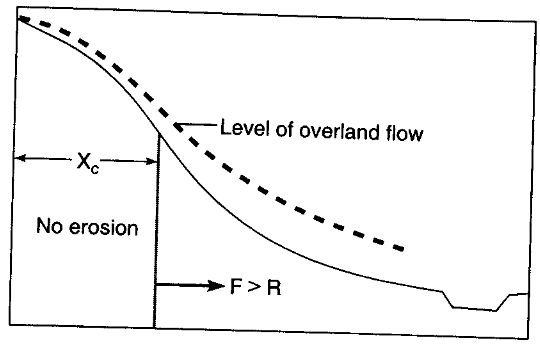


FIGURE 5.12

Hypothetical slope showing overland flow. No erosion occurs until the force of overland flow (F) exceeds the resistance of the surface material (R). Upslope from that point no erosion occurs. X_c is the distance from the divide to point where erosion begins.

(After Horton 1945)

areas is rare, and therefore the concept of a critical length has no particular significance. In cases where saturated overland flow dominates, the erosive action probably begins closer to the slope base (Kirkby and Chorley 1967; Kirkby 1969), and rills are gradually extended upslope. This headward growth most likely occurs because the headcut in any rill exposes a substratum that has a lower resistance to erosion than the slope surface. The formation of rills does not explain stream channels because a rilled surface is still part of a slope system. Rill development, however, is the first necessary step toward a true river. It is now generally recognized that rills initiated on a slope cannot long remain as parallel unconnected channels. Deeper and wider rills develop. These master rills carry more water, and because of their greater depth, undergo downcutting until all the flow is contained within the channel and the rill becomes a tiny stream. Because they become slightly entrenched, master rills capture adjacent rills when bank caving or overtopping during high flow destroys the narrow divides between them. The repeated diversion of rills, a process called **micropiracy,** tends to obliterate the original rill distribution, and gradually the initial slope parallel to the master channel is replaced by slopes on each side that slant toward the main drainage line.

The development of new slope direction in accordance with the master channel was called **cross-grading** by Horton (1945). In the final stage of micropiracy, only one stream, confined in the master rill channel, crosses the slope. The side slopes presumably develop a new rill system sloping to the position of the initial stream, and the process repeats itself, culminating in a secondary master rill serving as an incipient

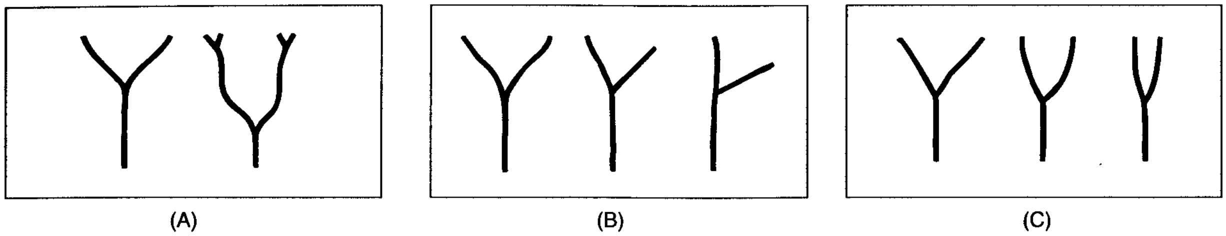


FIGURE 5.13
Development of bifurcation angles. (A) The original angle is preserved. (B) One branch becomes dominant. (C) Angle decreases and branches merge into one channel; occurs on steep slopes.
(After Schumm 1956)

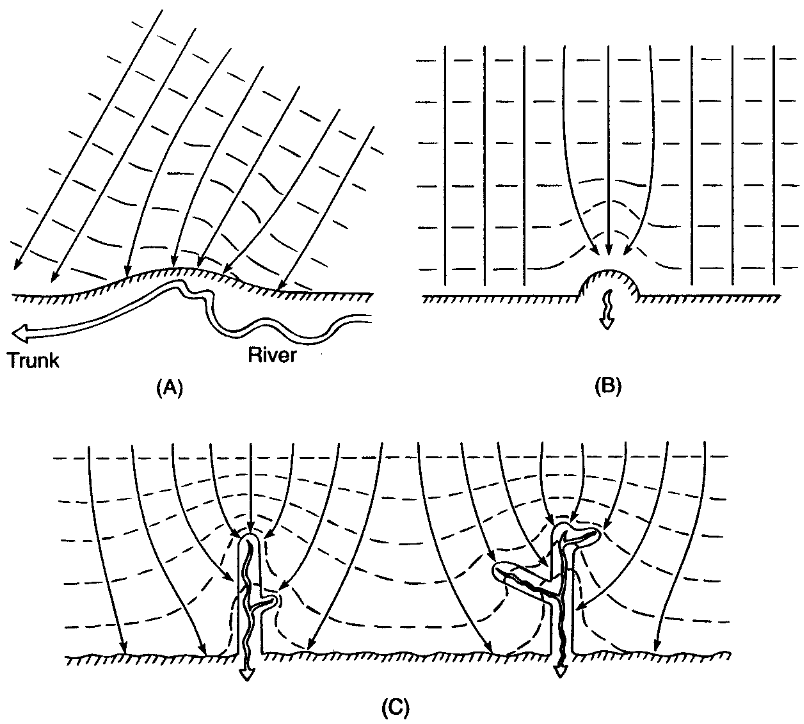


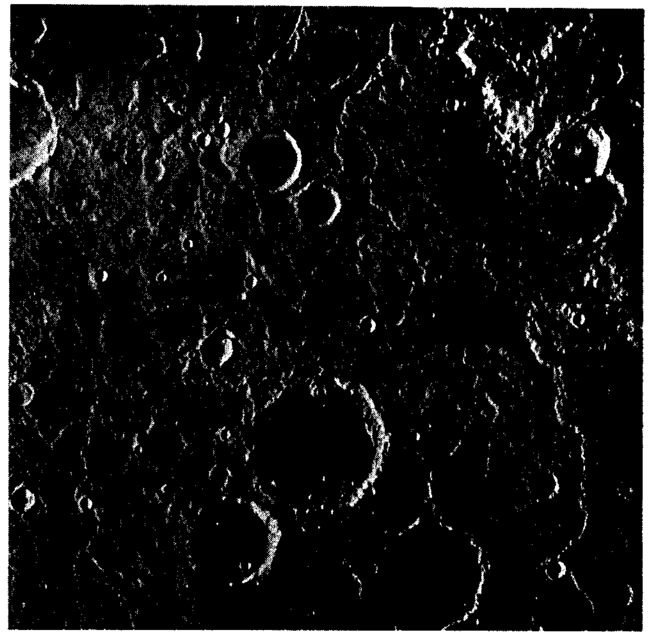
FIGURE 5.14
Evolution of groundwater-sapping system. Seepage begins along a stream bank at a local irregularity. Groundwater flow lines (solid) converge at the point of emergence and progressively accelerate seepage erosion. Dashed lines are lines of equal potential. This positive feedback system extends headwardly until bifurcation takes place, which is followed by development of a network system. The amphitheater-headed valley in (C) is similar to the forms shown on Mars in figure 5.15 and on Earth in figure 5.16.
(Dunne 1990)

tributary. Each smaller tributary evolves in a similar way until the network of streams takes form.

The network pattern develops by repeated division of single channel segments into two branches, a process known as **bifurcation.** Schumm (1956) suggests that the angle between the limbs of a bifurcated channel probably evolves in one of three ways (fig. 5.13): (1) both limbs grow headward while preserving the original angle at their juncture; (2) one branch straightens its course and becomes dominant; or (3) the angle on steep slopes progressively decreases until the branches reunite into a single channel. Any or all of these procedures might occur in the evolution of a network, constrained only by the fundamental erosive controls and the geologic framework. Divides between adjacent basins are predetermined by the extent to which streams can expand headward. Because critical length varies with resistance, the areal extent of the uneroded uplands partially reflects the geology and its history. Within the basin itself, smaller interfluves may be present where cross-grading has not operated. These small areas parallel the main stream and preserve the slope of the original surface.

Although the discussion above demonstrates the importance of overland flow in channel initiation, many recent studies have highlighted the additional role of subsurface water as an agent of surface erosion and channel initiation (for example, Dunne 1980; Kochel et al. 1985; Howard et al. 1988; Dunne 1990; Higgins and Coates 1990). Groundwater seepage zones along hillslopes and major escarpments commonly results in groundwater flow convergence, which produces forces capable of eroding materials in the vicinity of the groundwater emergence (fig. 5.14).

(A)



(C)

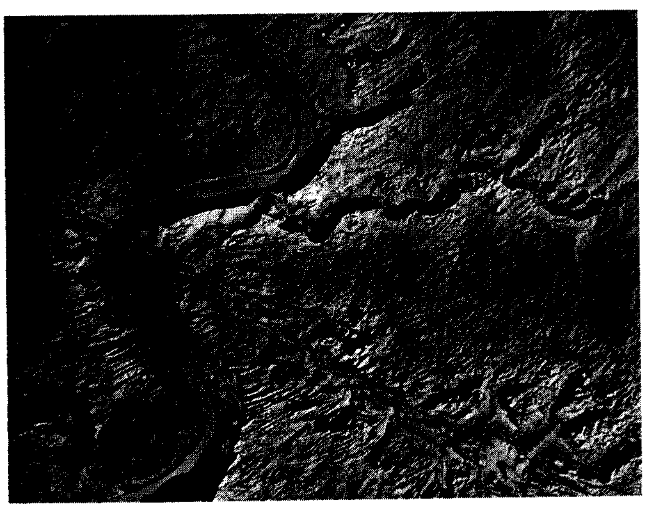
(B)

FIGURE 5.15

Drainage networks on Mars where groundwater-sapping processes are believed to have been the major formative process. (A) Nirgal Vallis, a longitudinal valley system with short, amphitheater-headed tributaries; channel is about 2–3 km wide. (B) Portion of Valles Marineris chasm showing stubby tributary complex (image is about 80 km across). Note how many tributaries have advanced headward along grabens and fractures. (C) Complex of smaller channel networks formed in the ancient hilly and cratered terrain.

This process, commonly referred to as *groundwater sapping,* is important in the formation and maintenance of valleys in regions as lithologically diverse as the Colorado Plateau (Howard et al. 1988) and Hawaii (Kochel and Piper 1986). It has also been proposed as a major valley-forming process on Mars (fig. 5.15). Valleys strongly influenced by sapping processes tend to have distinctive morphological characteristics such as amphitheater valley heads, stubby tributaries, and low channel densities (fig. 5.16). It is unlikely in a terrestrial environment that any valleys are initiated or maintained solely by runoff or groundwater seepage processes. Most channels are probably composite forms (Schumm and Phillips 1986), but the dominance of surface or subsurface processes may be reflected in the evolution of channel and basin morphology.

(A)



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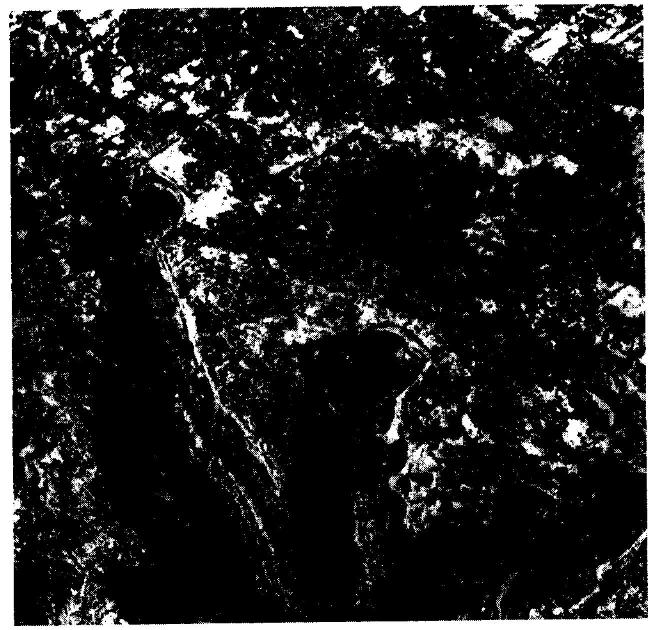


(C)

FIGURE 5.16
Earth examples where groundwater seepage and sapping processes have played a major formative role in valley development. (A) Northeast Kohala coast of Hawaii. The large, amphitheater-headed valleys have major springs at their head, fed from high-level aquifers (see Kochel and Piper 1986). The small, less-incised valleys in between are fed only by runoff. (B) Tributaries up-dip from the Colorado River have been significantly enlarged by groundwater-sapping processes in the permeable Navajo Sandstone. Note the lack of tributaries down-dip (to the bottom left). Runoff-dominated drainage systems typically show less influence on structural control. (C) Headward end of tributaries in the Navajo Sandstone of the Colorado Plateau in northern Arizona. Note the extension of valley heads along major joints where groundwater flow is enhanced. Compare to the right-angle junctions of valley heads on Mars in figure 5.15B.

Basin Morphometry

The drainage basin, the fundamental unit of the fluvial landscape, has been the focus of research aimed at understanding the geometric characteristics of the master channel and its tributary network. This geometry is referred to as the **basin morphometry** and is nicely reviewed by Abrahams (1984). Increasingly, studies have used the patterns of basin morphometry to predict or describe geomorphic processes; for example, it has been used to predict flood peaks, to assess sediment yield, and to estimate erosion rates (for example, Baumgardner 1987; Gardiner 1990). Some researchers believe that basin morphometric studies may ultimately be extended to show the influence of basin characteristics on channel cross-sections and channel attributes.

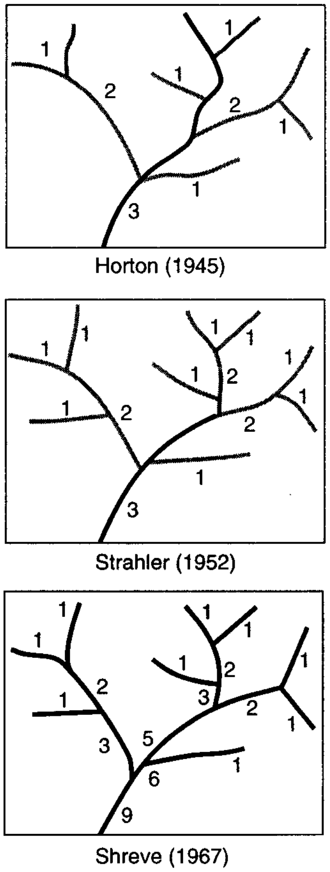


FIGURE 5.17
Methods of ordering streams within a drainage basin.

One of Horton's greatest contributions was to demonstrate that stream networks have a distinct fabric, called the *drainage composition,* in which the relationship between streams of different magnitude can be expressed in mathematical terms. Each stream within a basin is assigned to a particular order indicating its relative importance in the network, the lowest order streams being the most minor tributaries and the highest order, the main trunk river.

Figure 5.17 shows several methods of ordering steams. Horton's cumbersome method was refined by Strahler (1952a) so that stream segments rather than entire streams become the ordered units. In Strahler's system, a segment with no tributaries is designated as a first-order stream. Where two first-order segments join they form a second-order segment; two second-order segments join to form a third-order segment, and so forth. Any segment may be joined by a channel of lower order without causing an increase in its order. Only where two segments of equal magnitude join is an increase in order required. The Strahler method created an apparent omission in accounting of low-order tributaries that was later accommodated in another network ordering scheme proposed by Shreve (1966a, 1967). The Shreve Magnitude, as it is called, considers streams as links within the network, with the magnitude of each link representing the sum of the link numbers of all the tributaries that feed it; that is, networks in which the downstream segments are of the same magnitude have equal numbers of links within the basins. Shreve's designations thereby express the number of first-order streams upstream from a given point. Geomorphologists investigating relationships between rainfall and runoff find the Shreve Magnitude system useful. Because the first-order streams serve as the primary collectors of rainfall within a basin, they are better flood flow predictors than the Strahler ordering system (Patton and Baker 1976). Shreve's system appears in many of the sophisticated runoff modeling packages which are beyond the scope of this discussion (for example, see Smart and Wallis 1971; Abrahams 1980; Abrahams and Miller 1982).

Every basin possesses a quantifiable set of geometric properties that define the linear, areal, and relief characteristics of the watershed (table 5.2), known as the basin morphometry. These variables correlate with stream order, and various combinations of the parameters obey statistical relationships that hold for a large number of basins. Two general types of numbers have been used to describe basin morphometry or network characteristics (Strahler 1957, 1964, 1968). *Linear scale* measurements allow size comparisons of topographic units. The parameters may include the length of streams of any order, the relief, the length of basin perimeter, and other measurements. The second type of measurement consists of *dimensionless numbers,* often derived as ratios of length parameters, that permit comparisons of basins or networks. Length ratios, bifurcation ratios, and relief ratios are common examples. Table 5.2 shows the most commonly used linear, areal, and relief equations, but numerous others have been derived from these.

Linear Morphometric Relationships The establishment of stream ordering led Horton to realize that certain linear parameters of the basin are proportionately related to the stream order and that these could be expressed as basic relationships of the drainage

TABLE 5.2 Common morphometric relationships.

Linear Morphometry	
Stream number in each order (N_o)	$No = R_b^{s-o}$
Total stream numbers in basin (N)	$N = \dfrac{R_b^{s} - 1}{R_b - 1}$
Average stream length	$\bar{L}_o = \bar{L}_1 R_L^{o-1}$
Total stream length	$L_o = \bar{L}_1 R_b^{s-1} \left(\dfrac{u^s - 1}{u - 1}\right)$ where $u = R_L/R_B$
Bifurcation ratio	$R_b = N_o/N_{o+1}$
Length ratio	$R_L = \bar{L}_o/\bar{L}_{o+1}$
Length of overland flow	$\ell_o = \dfrac{1}{2D}$
Areal Morphometry	
Stream areas in each order	$\bar{A}_o = \bar{A}_1 R_a^{o-1}$
Length-area	$L = 1.4A^{0.6}$
Basin shape	$R_F = \dfrac{A_o}{L_b^2}$
Drainage density	$D = \dfrac{\Sigma L}{A}$
Stream frequency	$F_s = \dfrac{N}{A}$
Constant of channel maintenance	$C = \dfrac{1}{D}$
Relief Morphometry	
Relief ratio	$R_h = H/L_o$
Relative relief	$R_{hp} = H/p$
Relative basin height	$y = h/H$
Relative basin area	$x = a/A$
Ruggedness number (Melton 1957)	$R = DH$

Adapted from Strahler 1958.
s = order of master stream, o = any given stream order, H = basin relief, P = basin perimeter.

composition. Much of linear morphometry is a function of the *bifurcation ratio* (R_b), which is defined as the ratio of the number of streams of a given order to the number in the next higher order (using Strahler ordering). The bifurcation ratio allows rapid estimates of the number of streams of any given order and the total number of streams within the basin. Although the ratio value will not be constant between each set of adjacent orders, its variation from order to order will be small, and a mean value can be used. Also, as Horton pointed out, the number of streams in the second highest order is a good approximation of R_b. When geology is reasonably homogeneous throughout a basin, R_b values usually range from 3.0 to 5.0.

The *length ratio* (R_L), similar in context to the bifurcation ratio, is the ratio of the average length of streams of a given order to those of the next higher order. The length ratio can be used to determine the average length of streams in an unmeasured given order (L_o) and their total length. The combined length of all streams in a given basin is simply the sum of the lengths in each order. For most basin networks, stream lengths of different orders plot as a straight line on semilogarithmic paper (fig. 5.18), as do stream numbers. The re-

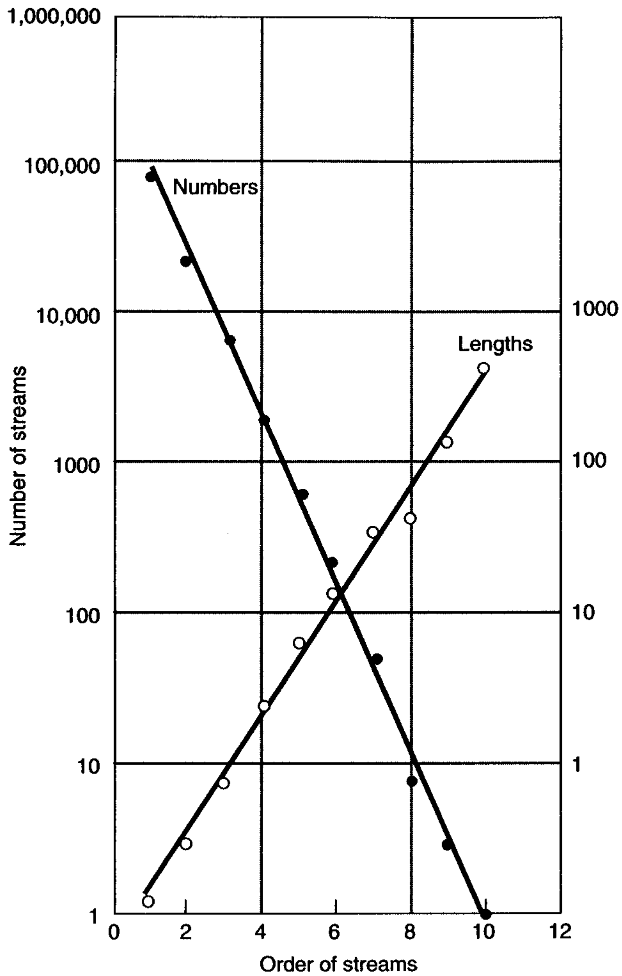


FIGURE 5.18

Relation of stream order to the number and mean lengths of streams in the Susquehanna River basin.

(After Brush 1961)

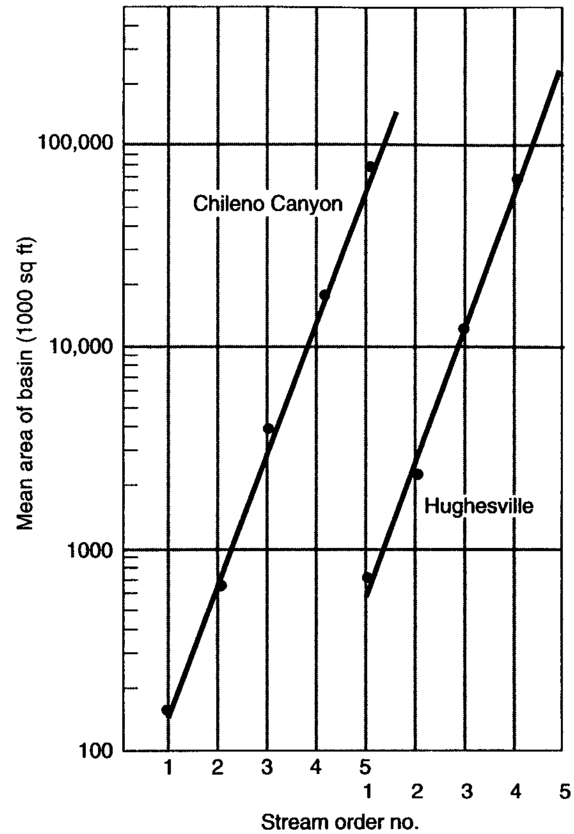


FIGURE 5.19

Relationship between stream order and mean basin area in two drainage basins.

(After Schumm 1956)

lationships between stream order and the number and length of segments in that order have been repeatedly verified and are now firmly established (Schumm 1956; Chorley 1957; Morisawa 1962; and many others).

Areal Morphometric Relationships The equity among linear elements within a drainage system suggests that areal components should also possess a consistent morphometry, because dimensional area is simply the product of linear factors. The fundamental unit of areal elements is the area contained within the basin of any given order (A_o). It encompasses all the area that provides runoff to streams of the given order, including all the areas of tributary basins of a lower order as well as interfluve regions. Schumm (1956) demonstrated (fig. 5.19) that basin areas, like stream numbers and lengths, are related to stream order in a geometric series.

Although area by itself is an important independent variable (Murphy et al. 1977), it has also been employed to manifest a variety of other parameters (see table 5.2), each of which has a particular significance in basin geomorphology, especially in regard to the collection of rainfall and concentration of runoff. Numerous studies have been successful in formulating relationships between basin area and discharge. One of the more important areal factors is *drainage density* (D), which is essentially the average length of streams per unit area and as such reflects the spacing of the drainageways. Drainage density reflects the interaction between geology and climate. As these two

TABLE 5.3 Summary of sample morphometric data for drainage basins in various regions.

Parameters[a]	Central Texas	Utah Wasatch	South California	Indiana	West Pennsylvania	Virginia	West Texas
	(19)[b]	*(11)*	*(12)*	*(10)*	*(12)*	*(1)*	*(25)*
A (km^2)	12.4	29.7	2.3	156.8	122	34.1	32.8
L (km)	9.2	9.8	2.0	26.8	26.8	8.3	9.1
W (km)						4.7	3.5
R (km)	.11	1.24	.44	.05	.28	.50	.71
S	4.5	4.5	4.6	5.9	5.4	4.0	5.0
D (km/km^2)	4.05	5.58	13.7	3.83	2.31	2.3	4.9
M						80	253
R	.55	6.25	5.78	.25	0.59	1.1	3.5
F_s (per km^2)	28.7	12.4	133.3	11.0	8.4	13.8	44.2
R_b						4.4	4.5
K						1.6	2.0
Geology	Carbonate	Mixed sedimentary	Mixed metamorphic and igneous	Sandstone and shale	Sandstone and shale	Metamorphic	Carbonate

(Patton and Baker 1976)
[a]S = average Strahler order, K = shape based on lemniscate, M = Shreve magnitude.
[b]Number of basins.

factors vary from region to region, large variations in D can be expected (table 5.3). In general, resistant surface materials and those with high infiltration capacities exhibit widely spaced streams, consequently yielding low D. As resistance or surface permeability decreases, runoff is usually accentuated by the development of a greater number of more closely spaced channels, and thus D tends to be higher. As a rule of thumb, where geology and slope angles are the same, humid regions develop thick vegetal cover that increases resistance and infiltration, thereby perpetuating drainage density lower than would otherwise be expected in more arid basins. Thus, drainage density not only reflects the geologic framework, but it may serve as a useful parameter in climatic geomorphology (Daniel 1981). Methods for rapid estimation of drainage density have been devised (McCoy 1971; Mark 1974; Richards 1979; Bauer 1980).

Drainage density has also been used as an independent variable in the framing of other morphometric parameters. For example, the *constant of channel maintenance* and the length of overland flow (see table 5.2) both utilize a reciprocal relationship with density to demonstrate the link between factors that control surface erosion and those that describe the drainage net (Schumm 1956). The constant of channel maintenance indicates the minimum area required for the development and maintenance of a channel; that is, the ratio represents the amount of basin area needed to maintain one linear unit of channel length. As Schumm points out (1956, p. 607) this relationship requires that drainage networks develop in an orderly way because the meter-by-meter growth of a drainage system is possible only if sufficient area is available to maintain the expanding channels.

Relief Morphometric Relationships A third group of parameters shown in table 5.2 indicates the vertical dimension of a drainage basin; it includes factors of gradient and elevation. Like stream numbers, length, and area, the average slope of stream segments in any order approximates a geometric series in which the first term is the mean slope of the first-order streams. This relationship is reasonably valid as long as the geologic framework is homogeneous. Channel slopes and surface slopes are closely akin to the parameters for length. Horton suggested, for example, that the length of overland flow as a function of only the drainage density is at best an approximation because overland flow also depends on slope parameters.

As relief refers to elevation differences between two points, slopes that connect the points are the integral factors affecting the flow of runoff. The most useful relief parameters are the *maximum basin relief* (highest elevation on the basin divide minus the elevation of the mouth of the trunk river) and the *divide-averaged relief* (the average divide elevation minus the mouth elevation). The *relief ratio* (Schumm 1956), the maximum

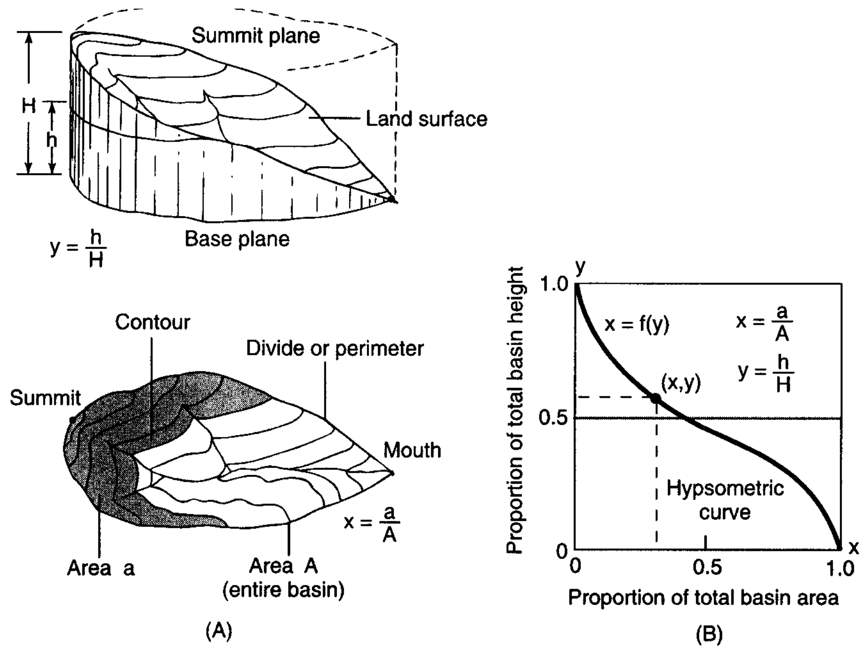


FIGURE 5.20
Ingredients of a hyposometric analysis. (A) Diagram showing how dimensionless parameters used in analysis are derived. (B) Plot of the parameters to produce the hyposometric curve.
(Strahler 1952b)

basin relief divided by the longest horizontal distance of the basin measured parallel to the major stream, indicates the overall steepness of the basin.

A different relief relationship is found by *hypsometric analysis* (Strahler 1952b), which relates elevation and basin area. As figure 5.20A shows, the basin is assumed to have vertical sides rising from a horizontal plane passing through the basin mouth and under the entire basin. Essentially, a hypsometric analysis reveals how much of the basin occurs within cross-sectional segments bounded by specified elevations. The relative height (y) is the ratio of the height (h) of a given contour above the horizontal datum plane to the total relief (H). The relative area (x) equals the ratio a/A, where a is the area of the basin above the given contour and A is the total basin area. The hypsometric curve (fig. 5.20B) represents the plot of the relationship between y and x and simply indicates the distribution of mass above the datum. The form of the curve is produced by the *hypsometric integral (HI)*, which expresses, as a percentage, the volume of the original basin that remains. In natural basins most *HI* values range from 20 to 80 percent, higher values indicating that large areas of the original basin have not been altered into slopes. Although computing the hypsometric integral can be tedious, methods have been introduced which streamline the procedure (Chorley and Morley 1959; Haan and Johnson 1966; Pike and Wilson 1971). Some researchers have found it to be an effective means of describing successive phases of landscape evolution (for example, Miller et al. 1990).

Basin Morphometry and the Flood Hydrograph

The application of geomorphic principles to environmental hazards, such as flood potential, has led to a significant amount of research attempting to identify relationships between basin morphometry and stream flooding (see review in Patton 1988). Clearly, the shape and character of a stream flood hydrograph should be affected greatly by the manner in which a basin collects

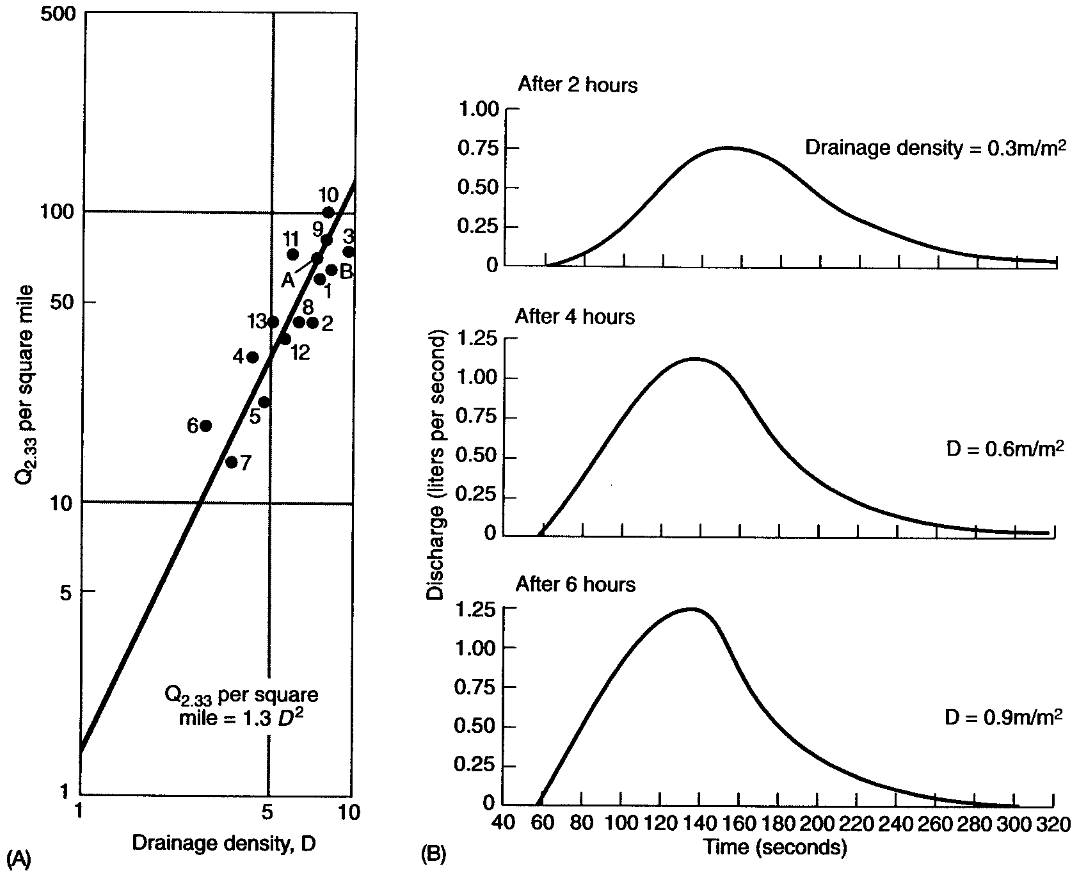


FIGURE 5.21

(A) Discharge (mean annual flood, $Q_{2.33}$) controlled by drainage density in 13 basins. (B) Effect of increasing drainage density on flood hydrograph in an experimental drainage system.

(A): (Carlston 1963), (B): (Zimpfer 1982)

and routes water through its network. Stream hydrology, as defined by the flood hydrograph and by time elements such as flood frequency and lag, is significantly related to many components of basin and network morphometry. The interdependence of morphometry and hydrology is statistically real but does not necessarily indicate a cause and effect relationship; given two apparently related factors, one factor is not necessarily the cause of changes in the other. The high correlation probably exists because both factors vary in a consistent way with the same underlying climatic and geologic controls. In general, area and relief factors are closely related to flow magnitude, and length elements to the timing of hydrologic events. All morphometric types, however, are themselves so complexly woven together that no single factor can be isolated as a completely independent variable (Murphey et al. 1977).

Because basin area and peak discharge are highly correlative, we could expect that many other areal parameters will be similarly related to discharge. Every factor involving area differs in its success as a predictor of discharge, but one parameter, drainage density, seems to have considerable value as a gage of peak flow. In a study of 15 small basins in the southern and central Appalachians and the Interior Lowland Plateau region, Carlston (1963) demonstrated a very close relationship between drainage density and mean annual flood (fig. 5.21A). Notably, the basins in his sample have wide variations in relief, valley-side and channel slopes, and precipitation characteristics; yet none of these factors disrupts the flood magnitude–drainage density relationship. Similar relationships have been observed in experimental studies (fig. 5.21B). Carlston suggests that the general capacity of a terrain to infiltrate precipitated water and transmit it through the underground system is the prime controlling factor of the density–mean annual flood relationship in basins up to 260 km² in area. In larger basins, channel transit time plays the dominant role in the flow character. The rate of base flow, found to be inversely related to

TABLE 5.4 Regression formulas for predicting flood magnitudes from drainage basin morphometry in diverse hydrogeomorphic regions.

Region	Equation	R^2	Probability
Central Texas	$Q_{max} = 17{,}369M^{0.43}(R)^{0.54}F^{-0.96}$	0.85	0.001
	$Q_{max} = 36{,}650M^{0.64}(R_h)^{0.54}(D)^{-1.68}$	0.74	0.01
Southern California	$Q_{max} = 155M^{1.04}(R)^{-0.83}F^{-0.73}$	0.85	0.001
	$Q_{max} = 380M^{0.89}(D)^{-1.87}$	0.86	0.0001
North-Central Utah	$Q_{max} = 23M^{0.90}(R)^{1.19}F^{-1.58}$	0.72	0.005
	$Q_{max} = 38{,}618M^{2.20}(R_h)^{2.51}F_1{}^{-3.73}$	0.83	0.005
Indiana	$Q_{max} = 424M^{0.46}(R)^{0.73}F^{0.21}$	0.67	0.01
	$Q_{max} = 424M^{0.82}(R_h)^{0.67}(D)^{0.56}$	0.66	0.05
Appalachian Plateau	$Q_{max} = 100M^{0.79}(R)^{0.19}F^{-0.29}$	0.92	0.0001
	$Q_{max} = 38M^{0.89}(D)^{-0.50}$	0.91	0.0001

Source: Patton and Baker 1976.
M = basin magnitude, R = ruggedness number, F_1 = first-order channel frequency; D = drainage density, R_h = relief ratio, Q_{max} = maximum peak discharge.

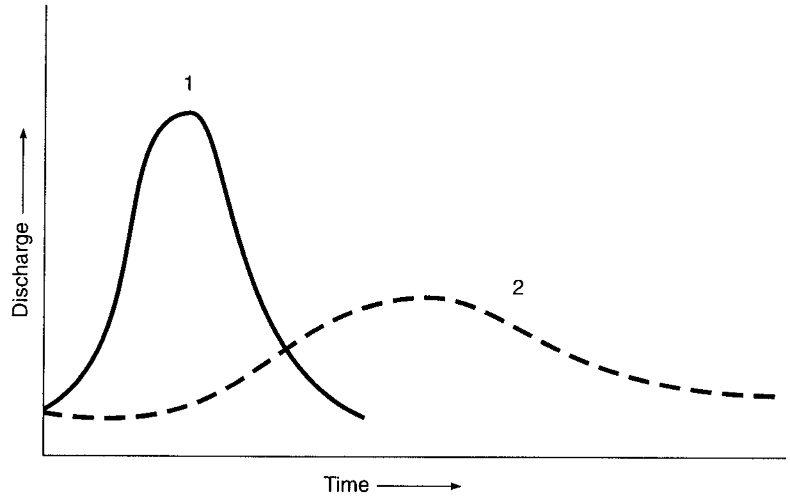
drainage density, is also dependent on terrain transmissibility. Thus, as Horton suspected earlier, high transmissibility (as evidenced by infiltration capacity) spawns low drainage density, high base flow, and a resultant low-magnitude peak flood. In contrast, an impermeable surface will generate high drainage density and efficiently carry away the abundant runoff; base flow will be low and peak discharge high.

Patton and Baker (1976) demonstrated predictive relationships between several morphometric parameters and peak flood discharges for streams in several physiographic regions of the United States (tables 5.3, 5.4). They found that areal morphometric parameters such as drainage density and stream frequency accounted for much of a model's ability to predict peak discharge, along with the relief measure known as *ruggedness number* (*R*) which is the product of relief and drainage density. These data were used to develop an index of flash flood potential (Beard 1975). Patton and Baker found that basins with high flash flood potential had greater ruggedness numbers than low-potential watersheds. Dingman (1978), however, warned that the relationship between drainage density and flow can be overridden by other effects in the basin such as floodplain or channel storage. In addition, where saturated overland flow is the major source of runoff, drainage density may not be related to the efficiency at which a basin is drained. Costa (1987) investigated the morphometry of basins associated with the largest historic floods in the United States. Although these flash flood basins did not uniformly possess the basin attributes expected from studies like that of Patton and Baker (1976), Costa was able to find some commonalties. Basins with flashy or peaked flood hydrographs generally contained significant area of exposed bedrock, occurred in semiarid to arid climates, were short, and had high relief.

In recent years a large amount of research has focused on the development of more sophisticated models of runoff that are linked closely with geomorphic attributes of the basin and their impact on the production of floods. One of the predominant models is the **geomorphic unit hydrograph** (Rodriquez-Iturbe and Valdez 1979). The success of modeling efforts have been mixed (Patton 1988), partly because of our incomplete understanding of the complex interrelationships between rainfall-runoff events and the contributing basin networks. Further work using small, instrumented watersheds, as well as numerical analytical approaches that explore relationships between the geomorphic unit hydrograph and basin parameters (Chutha and Doodge 1990), will refine our understanding and perhaps lead to more reliable models for predicting floods using basin parameters.

Abrahams (1984) aptly summed up the difficulties in elucidating quantitative relationships in basin networks by noting that the apparent randomness arises largely from independent variation of a large number of factors such as lithology and microclimate. The possible interrelationships between hydrology and morphometry are seemingly infinite, and the parameters are so complexly related that equations will not explain all the variability. Still, the hydrogeomorphic approach has some validity and should not be abandoned in future research. The hydrogeomorphic approach is especially applicable in determining regional flood hazards (Baker 1976). Figure 5.22 is a schematic model showing the expected influence of variations in

FIGURE 5.22
Idealized flood hydrograph and generalized responses to drainage basin characteristics.
The effect of an individual characteristic is shown assuming the other characteristics are held constant.



Characteristic	Flashy (hydrograph 1)	Sluggish (hydrograph 2)
Basin area	Small	Large
Drainage density	High	Low
Basin magnitude	High	Low
Relief	High	Low
Ruggedness number	High	Low
Basin shape	Equidimensional	Elongate
Soils	Thin	Thick
Vegetation	Dense	Sparse
Storm track	Down the basin	Up the basin

basin morphometry on the flood hydrograph based on generalizations from a large number of studies. In each case, the influence of a specified morphometric variable is displayed assuming that all other morphometric, geologic, and climatic variables remain constant. Although the relationships summarized here are somewhat qualitative until more conclusive research is completed, they offer a general guide for planners making an initial assessment of expected flood character in ungaged basins.

Basin Evolution

Although morphometric values differ from basin to basin, each network still obeys the statistical relationships discussed above. Many authors have suggested that morphometry reflects an adjustment of geomorphic variables that is established under the constraints of the prevailing climate and geology (for example, Chorley 1962; Leopold and Langbein 1962; Strahler 1964; Doornkamp and King 1971; Woldenberg 1969). Essentially, once a network is established, the basinal characteristics can be defined by the same quantitative terms at any time during the drainage growth. As the basins and networks evolve, an equilibrium is eventually produced by the interplay of climate and geology and maintained as a time-independent phenomenon. Once the components within a basin become balanced, any changes in climate or geology will be compensated for by adjustments of the basin parameters in such a way that the relationships of drainage composition will be preserved. As originally conceived, however, these relationships issued from well-developed stream systems, and the measurements needed to derive the equations were made on topographic maps of these basins. Such an approach provides no insight as to how quickly morphometric balance is attained or what changes in its character occur as the basin ages. It seems appropriate, therefore, to consider the influence of time on the morphometry of a basin.

Some studies have touched on the question of how rapidly morphometry is established and what changes occur in its nature as the basin evolves. These studies found that a quantitatively balanced drainage net forms rapidly in erodible material. This was clearly demonstrated by Schumm (1956) in the Perth Amboy, N.J., badlands, by Morisawa (1964) on the uplifted floor of Hebgen Lake, Mont., and by Kirkby and

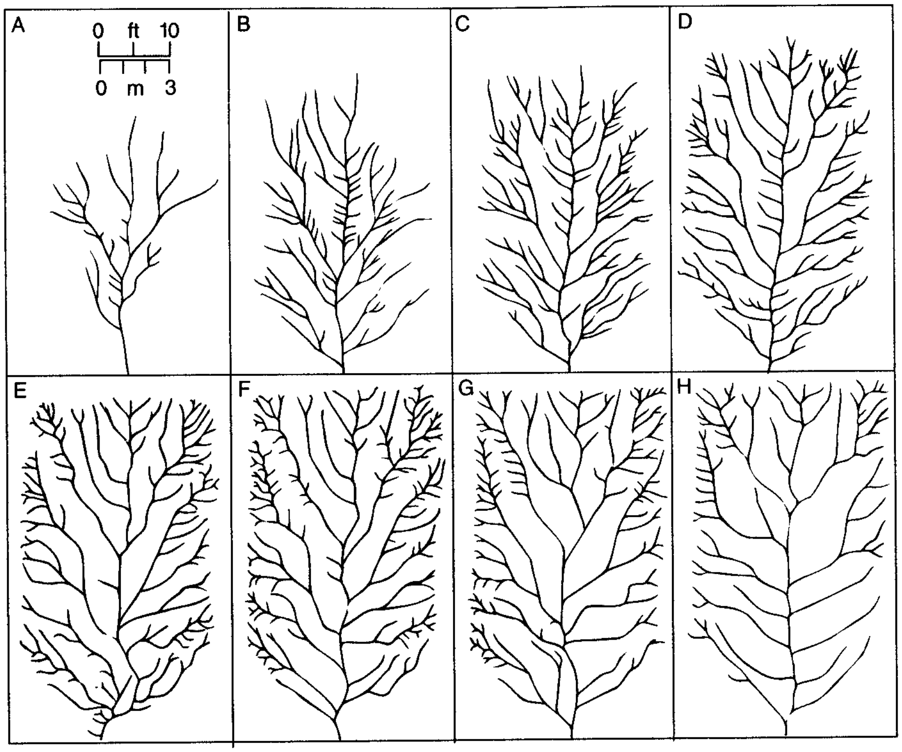


FIGURE 5.23
Chronologic development of an experimental drainage system in response to simulated rainfall.
(Parker 1977)

Kirkby (1969) on the raised beach and harbor floor around Montague Island, Ak. These studies showed that a balanced drainage composition evolved in a period no longer than a few years (for the Alaska case, in only a few days). Certainly, the time needed for drainage development to occur in erodible material is insignificant in terms of geological time. However, the amount of time needed in absolute terms probably varies according to the resistance of the material, the climate, and the initial slope angles. Unfortunately, examples of drainage establishment are documented only in areas where the least resistant materials underlie the system. Precisely how long it takes to form a balanced network in regions underlain by resistant crystalline rocks is rather conjectural.

Drainage evolution has been examined extensively by experimental means at the Rainfall Erosion Facility at Colorado State University by S. A. Schumm and his students and colleagues. Although experimental studies suffer some drawbacks in comparison to natural systems, such as measures of scale, they are exceedingly useful in providing conceptual models for processes that cannot be monitored in the field over the life span of an observer (Schumm et al. 1987).

Parker (discussed in Schumm 1977 and Schumm et al. 1987) found that patterns grow by headward extension until dissection reaches the watershed margin. The total drainage density and the mode of growth depended on the slope of the basin. Phillips and Schumm (1987) demonstrated a significant influence of slope on the form of drainage networks. They found that as slopes steepen the pattern changed from dendritic to parallel. Dendritic patterns predominated on slopes less than 1 percent, semiparallel patterns prevailed on 3 percent slopes, and parallel patterns formed once slopes attained 5 percent—clearly demonstrating the influence of slope angle on junction angle. The basin with a lower slope had a higher density because tributaries formed inside the basin during early extension growth, whereas in the steeper basin streams extended to the margins before interior links were added (Parker 1976). In general, Parker's work showed that the network will develop as many streams and as much length as is needed to efficiently drain the basin. However, distinctive changes in drainage density were commonly observed in different parts of the basin during its network evolution (fig. 5.23).

The experiments summarized by Schumm et al. (1987) outline predictable sequences of drainage network evolution which highlight processes such as stream piracy in facilitating the growth process. Similarly, this experimental work was successful in establishing strong linkages between variations in sediment yield and activity in the evolving channel network. One significant cautionary note arising from their work is that different portions of the basin

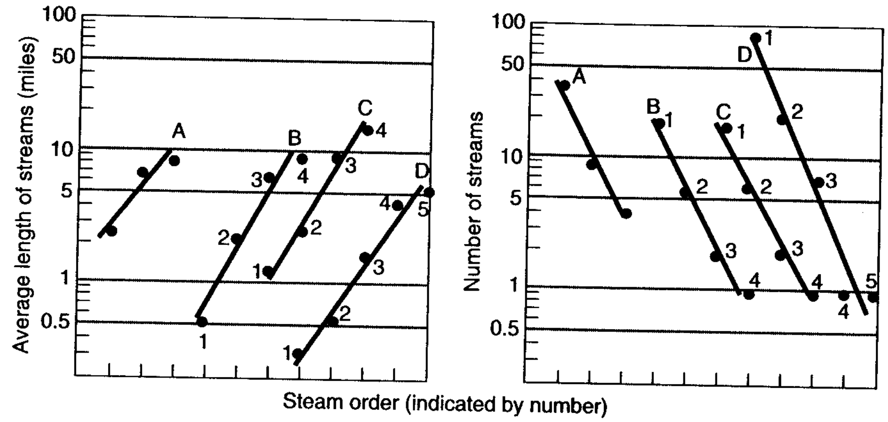


FIGURE 5.24

Relation of stream order to stream length and number in drainage basins of the Ontonagon Plain, Mich. *A* = streams between Mineral and Cranberry rivers; data from maps; stream orders not known. *B* = Little Cranberry River, *C* = Weigel Creek, *D* = Mill Creek.

(Hack 1965)

typically evolve disharmoniously, reemphasizing the concept of complex response often cited in Schumm's field studies. Thus, sediment yield and storage will operate in a complex manner, and making correlations between valleys and even between different sites along the same valley is difficult.

Once basin elements attain a statistical balance, further changes in morphometry are usually revealed in the shape, drainage texture, or hypsometry of the basin. The area limits of any basin presumably are determined by the hydrophysical controls denoted by Horton and by the competition for space between adjacent basins. During the period of expansion to its peripheral limits, however, each parameter should change in such a manner as to maintain the original quantitative relationships among the factors. Hack (1957) showed that for a large number of basins, the stream length and basin area are related by the simple power function

$$L = 1.4\,A^{0.6}$$

where L is the distance from any locality on a stream to the divide at the head of the longest segment above the locality, and A is the basin area above the locality. Many investigations of basin growth indicate a tendency toward elongation (for example, Hack 1965; Miller 1958; Jarvis and Sham 1981), while others suggest that widening occurs as rapidly as elongation, particularly in large basins (Mueller 1972; Shreve 1974; Moon 1980). Figure 5.24 demonstrates that although many local variables influence basin shape and development, distinct relationships exist between stream order and morphometric parameters such as stream length and stream number.

Part of the controversy regarding trends in basin evolution through time relates to a concept suggested by Schumm (1956), that the locus of intense erosion migrates with time from the basin mouth vicinity early in basin development to the headward region in later stages. In addition, Ruhe (1952) demonstrated that both drainage density and stream frequency increase systematically with time in areas underlain by glacial deposits of known age. The rate of textural change is not constant; it was probably greatest during the first 20,000 years and then decelerated. The marked transition in the rate of textural evolution perhaps represents the time at which complete equilibrium was established and the basins were filled with as many streams as possible (Leopold et al. 1964).

Throughout the period of growth in Ruhe's study, the channel lengths and numbers seem to obey Horton's geometric laws, suggesting that texture may be a surrogate for time in the analysis of basin evolution. This proposition was given added credence when Melton (1958) found stream frequency (F_1) and drainage density (D) to be related by a simple equation:

$$F_1 = 0.694\,D^2$$

In deriving this equation, Melton used as his sample 156 basins with widely different geology, erosional history, and probable age. Because all stages of drainage development were thrown into the statistical pot, the significant empirical relationship between the variables most likely reflects a general trend followed by basins as they grow. The dimensionless ratio F_1/D^2, called *relative density*, should indicate how completely the stream network fills the basin (Melton 1958). High relative-density values suggest that stream lengths are short and the basin outline is not yet completely filled with the stream network. As the drainage evolves and expands into each basin niche, the ratio decreases until equilibrium is established.

Melton's growth law is one in which spatial parameters can be substituted for time. However, the universal applicability of the growth equation has been questioned. Abrahams (1972) suggested that time and space are interchangeable in morphometry only if the basins analyzed are environmentally similar and are of the same order. Further studies, however, did not completely substantiate this criticism (Wilcock 1975), and a reasonable argument can be made that all factors—linear, areal, and relief—are probably involved in the statistical relationships affected by time. An interesting addendum to the controversy over basin adjustments is provided by a simulation study modeling long-term basin morphologic development (Wilgoose et al. 1991). This study suggests that once a basin attains dynamic equilibrium in a setting where fluvial sediment transport is dominant, all basins will be similar except for variations in drainage density, in spite of differences in tectonics and lithologic resistance.

Alterations of basin morphometry do occur with time. These changes do not violate the steady-state concept because parameters vary with one another in a systematic way. It is tempting to assume, therefore, that a well-balanced network with discernible morphometry evolved in an orderly manner from infancy to its present state. Such an assumption, however, may be totally erroneous because basin morphometry may tell us little, if anything, about *basin history.*

A cogent example of this is the drainage system of Volney Creek, a small basin in southern Montana within the watershed of the Yellowstone River. Volney Creek heads in the piedmont region of the Beartooth Mountains, rising approximately 3 km from the mountain front. This basin, covering an area of 60 km^2, is underlain entirely by Mesozoic and Tertiary clastic sedimentary rocks. The basin exhibits all the normal morphometric characteristics of neighboring piedmont basins, and it is presumed to be in equilibrium because its morphometry is balanced statistically, as discussed above. The geomorphic history of the basin, however, shows that its evolution was anything but orderly (Ritter 1972). Two terraces standing well above the present level of Volney Creek are capped by gravel containing crystalline clasts that had to be derived from the Beartooth Mountains. In fact, upstream tracing of the terraces demonstrates that they cross through the present basin divide and continue mountainward in the neighboring valley of West Red Lodge Creek (fig. 5.25).

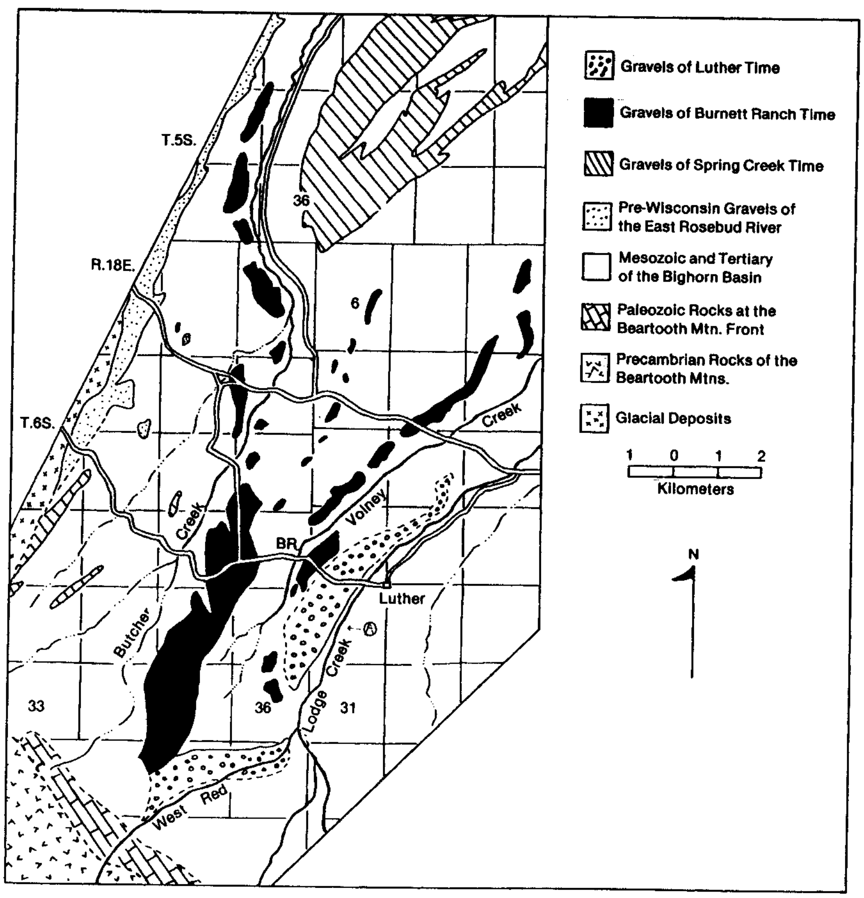


FIGURE 5.25

Map of terrace deposits in a small piedmont area of the Beartooth Mountains, Mont. Major stream piracy occurred near *A* between deposition of Burnett Ranch gravels and deposition of Luther gravels.

(Ritter 1972)

The significance of this is that prior to early Wisconsinan glaciation (Burnett Ranch time in fig. 5.25), the valley now occupied by Volney Creek was the drainage avenue for the master stream of the area. The crystalline-rich gravel capping the terraces indicates that the paleoriver drained from the mountains, and its basinal area must have been vastly greater than that of the modern Volney Creek. Immediately preceding the glaciation, the river was diverted into its present position in the valley of West Red Lodge Creek, leaving the Volney segment abandoned until it was occupied by the very small modern stream.

Diversions of the Volney Creek type are common in piedmont regions and may be important phenomena in the expansion of any drainage system, especially in the early stages (Howard 1971). Such changes, nonetheless, are catastrophic events in basin development because they drastically alter basin properties such as area, relief, and stream length. The internal adjustment to changes spurred by piracy must occur rapidly, for most basins possess a balanced morphometry even though such spasmodic events must be commonplace.

Basin Hydrology

For planning purposes, estimates are needed of how much water exists in a drainage basin and whether it is available for use. Hydrologists usually employ a concept known as the **water balance** or **hydrologic budget** to make such estimates. The water balance simply refers to the balance that must exist between all water entering the basin (input), all water leaving the basin (output), and any changes in the amount of water being stored.

Figure 5.26 illustrates the hydrologic cycle with respect to a drainage basin. The major inputs are rainfall and snow, and the major outputs are evapotranspiration and streamflow. Water is stored in lakes, soil moisture, and groundwater, and changes in these values actually represent losses or gains of available water. Although the basin cannot actually be divorced from the global hydrologic system, we can view the basin hydrologic budget simplistically by the following equation:

$$\Delta S = I_P - O_E - O_T - O_R$$

A change in storage, measured as recharge to the groundwater system (ΔS), is equal to the difference between input as infiltration from precipitation (I_P) and losses or outputs from evaporation (O_E), transpiration (O_T), and runoff (O_R). Although runoff provides the predominant connection between our basin and the global hydrologic system, some groundwater may exit the basin by regional subsurface flow pathways. For example, a positive change in groundwater storage in-

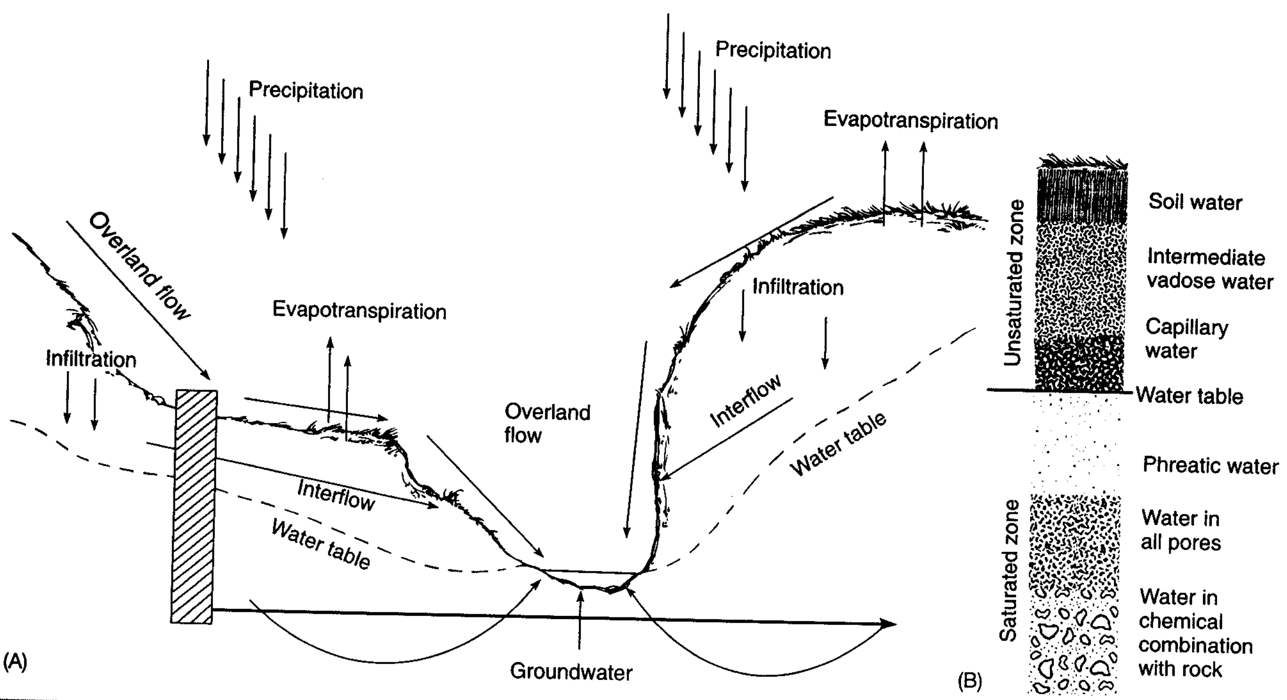


FIGURE 5.26
Surface and shallow subsurface hydrology of a watershed. (A) Schematic of basin showing directions and relative water movements measured in discharge (volume per time). (B) The groundwater profile as it would appear in homogenous material in the vertical slice on the left.

dicates that the underground reservoir is being recharged, but in the budget it represents a loss because the availability of the water is being lowered. On the other hand, groundwater runoff (base flow) represents an input because water availability is increased as it is released from storage.

Subsurface Water

The increased demand for water that has accompanied population growth, industrial expansion, and extended irrigation in the United States has brought with it a marked increase in the utilization of groundwater. Groundwater as a resource is highly sought after because it typically maintains chemical and physical characteristics that fluctuate far less radically than surface water supplies. For example, groundwater maintains a constant year-round temperature, a more constant chemical composition year-round, and is less affected by periodic droughts than surface reservoirs. The volume of water contained in fractures and pores in Earth's underground reservoir is enormous; it is estimated at almost 8 million km^3 in the outer 5 km of the crust (Todd 1970). Unfortunately, this vast resource is not evenly distributed. Some regions are blessed with abundant groundwater whereas others are seriously deficient. Ironically, many of the most rapidly expanding areas of the United States are in regions with low reserves of surface and groundwater. The lack of available water, coupled with the growing demand and ever-present threat of contamination, poses a major challenge to geologists and requires that we continue to expand our knowledge concerning the distribution, movement, and utilization of groundwater. For example, some regions possess groundwater with exceptionally high natural concentrations of dissolved minerals, whereas others are plagued by anthropogenic contaminants. Groundwater geology, or **hydrogeology,** has evolved into a separate discipline in itself; however, because it is an important part of basin hydrology, we will briefly examine how the groundwater system works.

The Groundwater Profile Groundwater in porous and permeable rocks or unconsolidated debris usually has a rather distinct distribution that can be visualized as a vertical zonation known as the *groundwater profile* (fig. 5.26B). The *unsaturated zone* above the water table is variable in water content within its pores and fractures. Air occurring in pores that are only partly water-filled is physically connected to the atmosphere. Slugs of infiltrating water move vertically down, under the influence of gravity, through the unsaturated zone in response to episodic precipitation events at the surface. Flow paths include movement through the interstices of porous sediment and bedrock, flow through bedrock fractures, and accelerated flow along zones of increased permeability referred to as macropores such as root channels and burrows. Interruptions in the complete downward transfer of water through the unsaturated zone commonly result from water being held by surface tension or that which is extracted by plants during evapotranspiration.

The uppermost portion of the unsaturated zone is the *zone of soil moisture* which experiences large fluctuations in water quantity and quality because of variations in infiltration, evapotranspiration, and atmospheric pressure. The *intermediate zone* below the soil moisture zone experiences downward flow toward the water table. This zone, commonly thick in semiarid regions, may be totally absent in many humid areas.

The base of the unsaturated zone is marked by a zone of variable thickness where all available pore space is saturated with water. This area of apparent saturation is known as the *capillary fringe.* Capillary action draws water upward through the pores from the zone of saturation and holds this water in tension. Even though this zone is saturated, water will not drain freely into a well placed within the capillary fringe because the hydrostatic pressure remains less than atmospheric pressure, as it does throughout the unsaturated zone. The thickness of the capillary fringe depends on pore diameter, its thickness being inversely proportional to the diameter of the pores.

All pore spaces below the capillary fringe are filled with water; hence this area is called the *saturated* or *phreatic zone.* Phreatic water will drain freely into a well because the hydrostatic pressure in this zone exceeds that of atmospheric pressure. The *water table* marks the level where the hydrostatic pressure is equal to atmospheric pressure and defines the upper boundary of the saturated zone. This zone extends downward until rocks become dense enough that small occurrences of interstitial water are no longer interconnected. The saturated zone may extend for tens of thousands of feet in sedimentary basins, being less extensive in crystalline rocks.

Movement of Groundwater Water in the saturated zone is not static but can move in any direction depending on the distribution of potential energy fields within the zone. Groundwater flow in the saturated zone is not solely regulated by gravity as it is in the unsaturated zone. Each unit volume of water contained in the saturated (phreatic) zone possesses a certain amount of potential energy, called its *potential* or *head.* The amount of potential varies from droplet to droplet, being dependent on the pressure of each drop

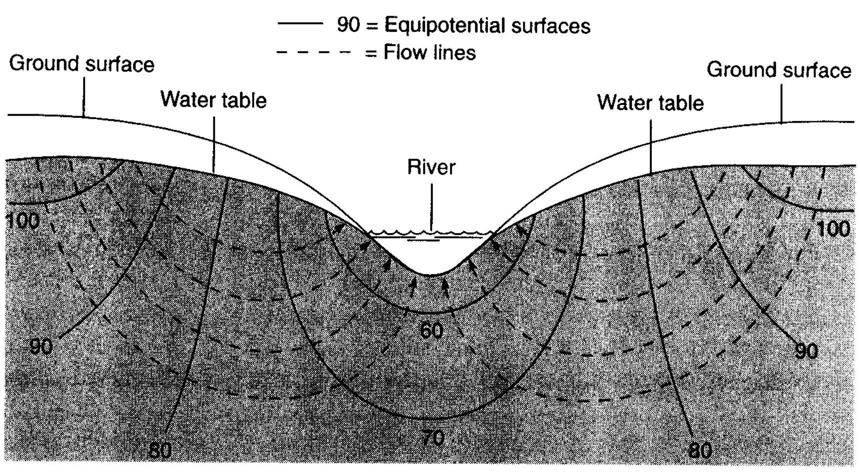


FIGURE 5.27
Movement of groundwater according to distribution of potential in the underground system. Water moves from high to low potential and perpendicular to the equipotential surfaces.

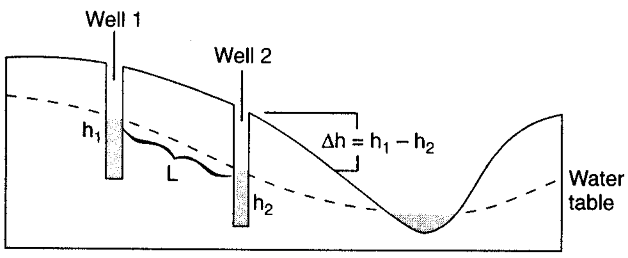


FIGURE 5.28
Water table and loss in head as water moves from well 1 to well 2 in unconfined aquifer.

and its elevation above some datum. When pressure and elevation are known, the potential for each unit volume of water can be calculated, and water particles having the same potential can be contoured along surfaces known as *equipotential surfaces* (fig. 5.27). Although some diffusion occurs, groundwater particles move along paths that are perpendicular to equipotential surfaces.

The movement of groundwater is a mechanical process whereby some of the initial potential energy of the water is lost to friction generated as the water moves. It follows that water moving from one point to another must have more potential energy at the beginning of the transport route than at the end. Thus, groundwater always moves according to the following rules: (1) it moves from zones of higher potential toward zones of lower potential, and (2) it flows perpendicular to the equipotential surfaces, at least in simple homogeneous and isotropic materials (Hubbert 1940).

The velocity and discharge of groundwater flow is directly proportional to the loss of potential (head) that occurs as water moves from one point to another (fig. 5.28). This concept was demonstrated in 1856 by H. Darcy in his famous experiments on flow through porous media. He also observed that flow velocity was inversely proportional to the length along the flow path. Darcy's experiments led to the fundamental law of groundwater flow, expressed as

$$V = K\frac{h_1 - h_2}{L}$$

where V is a volumetric flow rate similar to velocity, h is head, L is the distance between the two points of measurement (i.e., between monitoring wells), and K is a constant of proportionality representing the permeability of the medium, known as the *hydraulic conductivity*. In figure 5.28, the velocity of flow can be calculated from the difference in hydrostatic level (head) between two wells ($h_1 - h_2$) if the permeability of the material is known or can be estimated. Discharge can also be calculated by including the cross-sectional area of the aquifer to the equation so that

$$Q = KIA$$

where Q is discharge, A is the cross-sectional area of an aquifer perpendicular to flow ($w \times d$), and I, called the *hydraulic gradient*, equals $\dfrac{h_1 - h_2}{L}$.

Darcy's Law is applicable to groundwater flow through porous materials, but adjustments are often made to accommodate the effects of variable media and fluid properties on the actual values of K. More complex relationships must be used to estimate flow characteristics through fractures in relatively impermeable host materials.

The water table, at the top of the phreatic zone, is an important hydrologic feature (see fig. 5.26). Because hydrostatic pressure everywhere along the surface of the water table is equal to atmospheric pressure, the potential of the water table is equal to atmospheric pressure. The potential of water there is completely a function of elevation, and water at this surface will always move from higher to lower elevations. This partially explains why the water table is generally a mirror image of surface topography. Whereas surface water and groundwater systems are physically connected, the levels of rivers, lakes, swamps, and other wetlands are merely surface extensions of the underground water table (see fig. 5.27).

Aquifers, Wells, and Groundwater Utilization Problems *Aquifers* are lithologic bodies that store, transmit, and yield water in economic amounts. The most common aquifer is called an *unconfined aquifer* because it is open to the atmosphere and its hydrostatic level (shown by the level at which water will stand in a well) is within the water-bearing unit itself. The hydrostatic level in an unconfined aquifer is the water table. In some other aquifers the water is held in a porous and permeable unit that is not connected vertically to the atmosphere but instead is overlain and underlain by less permeable layers called confining layers (or aquitards). Water in these *confined aquifers* (fig. 5.29) will rise above the top of the aquifer when it is penetrated by a well or open hole. The level to which water will rise is called the *potentiometric surface.* The height of the potentiometric surface above the aquifer depends on the difference in potential at the point where precipitation or infiltration enters the aquifer, called the recharge zone, and the position of the screen at the base of the well or hole (fig. 5.29). If the potentiometric level is above the elevation of the ground, surface water will flow freely out of the well without pumping as *artesian flow.*

The development of an aquifer for water supply requires wells; the larger the demand, the larger and more numerous the wells. As a well is pumped, the hydrostatic level (water table or potentiometric surface) surrounding the well declines and forms an inverted cone known as the *cone of depression* (fig. 5.30). The cone develops because the release of water (or of

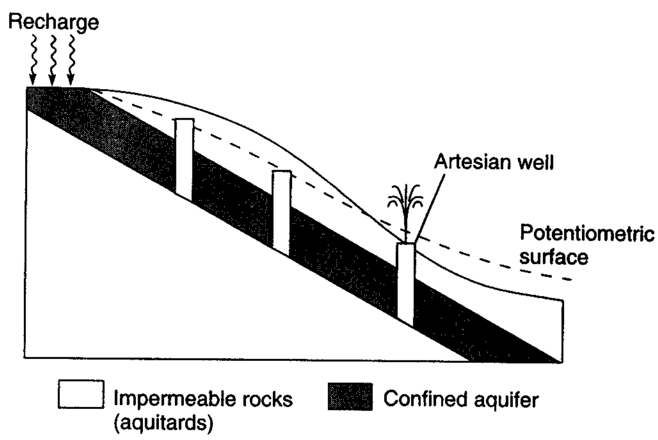


FIGURE 5.29
Confined aquifer and potentiometric surface.

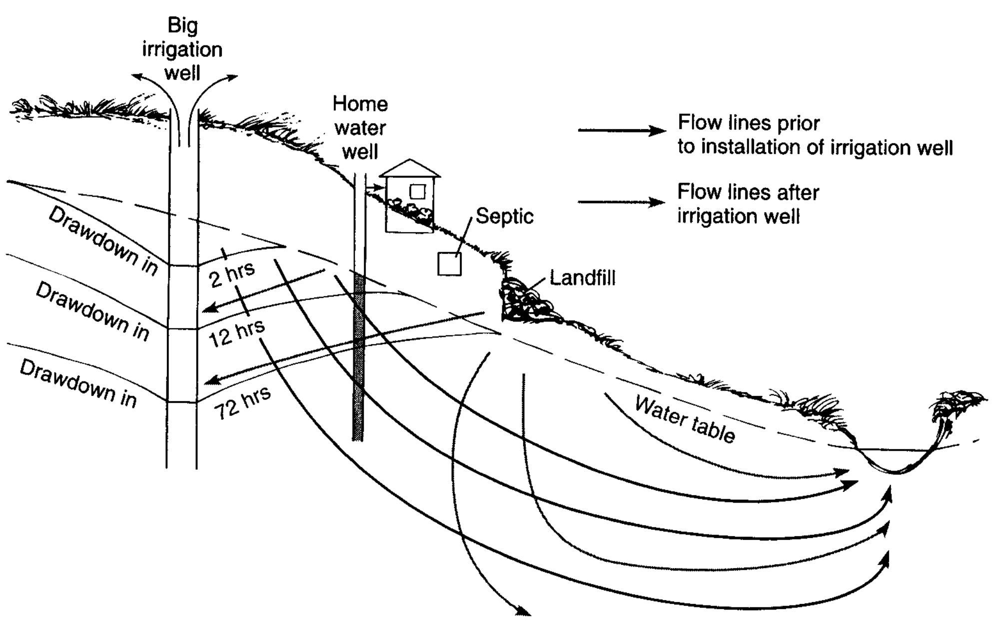


FIGURE 5.30
Development of cone of depression drawing down the water table due to the pumping of a major irrigation well. The cone of depression can alter groundwater flow patterns significantly as indicated here where the leachate from a landfill, which normally would move to the right, would now flow toward the water supply well for the home. This example assumes a simple homogenous system.

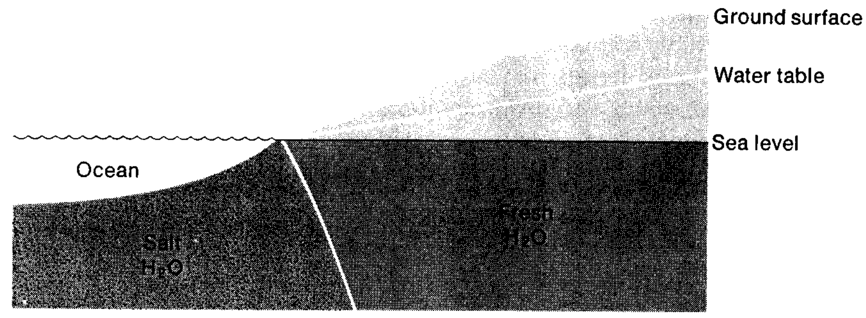


FIGURE 5.31

Relationship of salty and fresh groundwater in an aquifer of a coastal region. Lowering of water table by overdraft requires a 40-fold rise in the boundary between the fresh and salt water and increases the possibility of pollution of the aquifer by salt water.

pressure in a confined aquifer) is greatest near the well. This causes pronounced lowering of the water table or potentiometric surface, called *drawdown*, adjacent to the well. The effect of pumping decreases away from the well, so drawdown is less toward the perimeter of the cone. In the initial phase of pumping, the rate of drawdown is normally high, but as pumping continues the rate gradually decreases until the cone attains a nearly constant form. The dimensions of the quasi-equilibrium cone depend on the rate at which the well is pumped and the hydrologic properties of the aquifer.

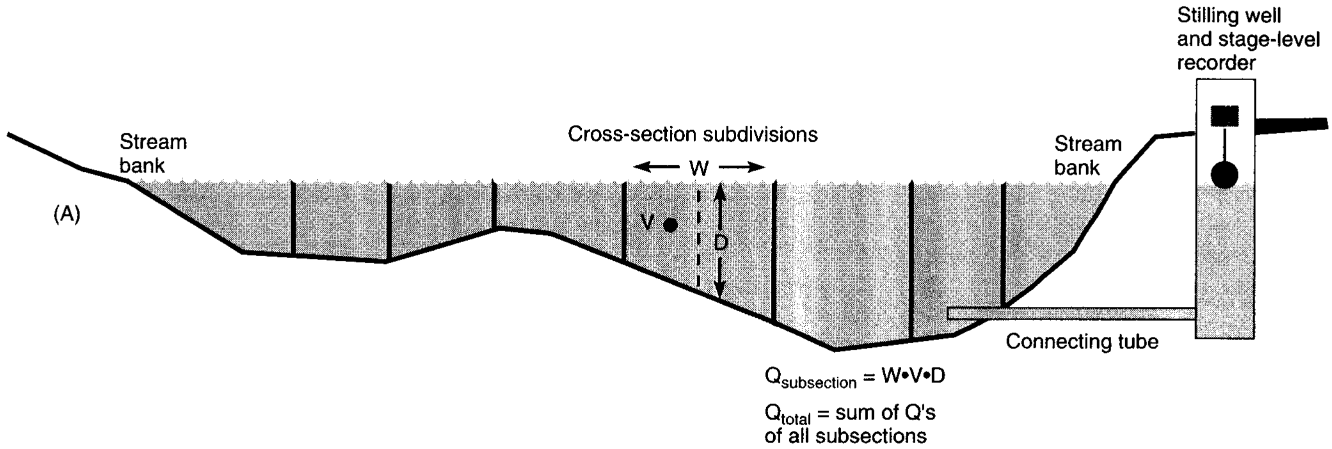
Utilizing groundwater can cause a number of environmental problems if the system is not studied carefully before development. For example, excessive drawdown can occur locally if wells are placed too close to one another and the radius of influence (the maximum diameter of the cone of depression) of adjacent wells overlap. This produces abnormally high drawdown in the zone of overlap because the drawdowns from all interfering wells are cumulative at any point. On a regional scale, most problems arise when more water is pumped from the aquifer over a period of years than is returned to the aquifer by natural or artificial recharge, a practice known as *overdraft*. In such a case the aquifer is actually being mined of its water, and on a long-term basis the normal hydrostatic level may be drastically lowered. The socioeconomic importance of these issues are well illustrated by the recent article in *National Geographic* (Zwingle 1993) which treats the concerns over depletion of groundwater in the major hydrostratigraphic unit of the United States High Plains region, the Ogallala Aquifer.

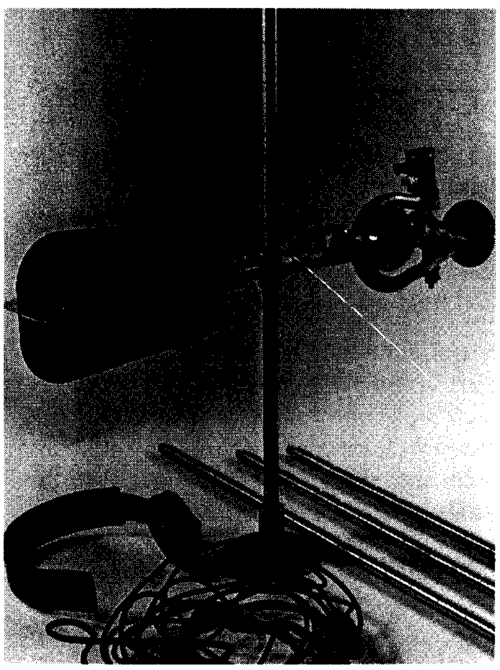
The effects of continued overdraft vary, but two types of responses will demonstrate the problems that can result from the misuse of the groundwater system. First, in confined aquifers, the drawdown of the potentiometric surface reflects the decrease of pressure within the aquifer. Because the pressure is lowered, water from the overlying confining layer seeps downward into the aquifer. As the confining layer drains, the normal load exerted by the weight of the overlying rocks compacts the confining layer and decreases its thickness. Ultimately, the process culminates in measurable subsidence of the ground surface. Some areas, such as Mexico City, Las Vegas, the Central Valley of California, and the Houston-Galveston area, have experienced 1 to 5 meters of overdraft subsidence, creating a variety of annoying and hazardous conditions such as cracking of buildings, strain on buried pipelines, highway subsidence, and destruction of well casings.

A second major effect of overdraft usually occurs in coastal regions where an aquifer is physically connected to the ocean. There, two fluids (ocean water and fresh water) having different densities are separated by a sharp boundary, as shown in figure 5.31. The location of the interface depends on the hydrodynamic balance between fresh water (density = $1.00\ g/cm^3$) and salt water (density = $1.025\ g/cm^3$). In general, the depth below sea level to the saltwater boundary is about 40 times the height of the water table above sea level. In such a situation, continued overdraft can cause pollution of the aquifer because minor lowering of the water table necessitates a much greater rise of the saltwater-freshwater interface, a phenomenon known as *saltwater intrusion*. For example, a 2-meter drawdown of the water table requires a concomitant 80-meter rise of the salt water, and any well extending to a depth greater than the new interface level will be polluted with nonpotable water. Calvache and Pulidd-Bosch (1991), using a numerical modeling approach, eloquently demonstrated how expanding groundwater withdrawal is responsible for the gradual encroachment of saline water into important regional aquifers in coastal Spain.

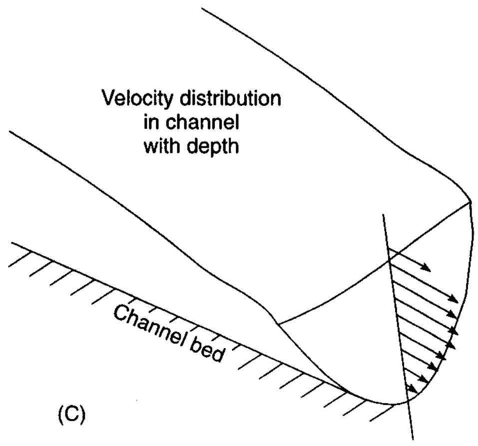
Many cities along the coasts of California, Texas, Florida, New York, and New Jersey have been affected by saltwater intrusion. This phenomenon, however, can occur wherever two fluids of different density exist within the same aquifer. For example, the water supply of Las Vegas, Nevada, is in jeopardy from pollution by high-magnesium groundwater located about 24 km south of the city. The intrusion is a response to the large overdraft from the aquifer beneath Las Vegas.

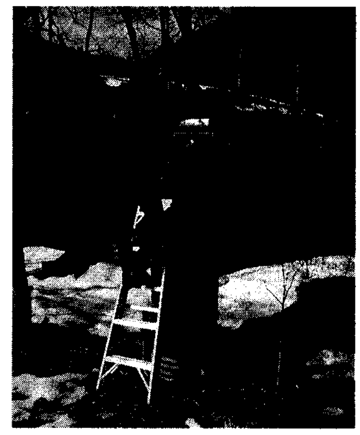
Linkages between groundwater and surface water systems have been the focus of many studies since the late 1970s. These have highlighted interactions related





(B)





(D)

FIGURE 5.32

Stream discharge gaging station. (A) Channel cross-section subdivided into small compartments for discharge measurement in each. (B) Price current meter for measuring velocity. (C) Vertical profile of velocity in the channel. Average velocity can be measured just below half the depth. (D) Stream gaging station on Antes Creek, Pa. A rotating drum chart recorder is positioned at the top of a stilling well to monitor stage changes in the creek. The well is connected with the creek via lateral pipes below the floodplain.

to channel initiation (Dunne 1990; Kochel et al. 1985), particle entrainment, floodplain aquifers and river-stage fluctuations (e.g., Sophocleous 1991), and the effects of shallow groundwater circulation on soil development and chemistry and of soil properties on groundwater (e.g., Richardson 1992). In light of our ever-increasing concern about the environment, increased knowledge of the interactions between subsurface and surface hydrologic systems is becoming very important.

Surface Water

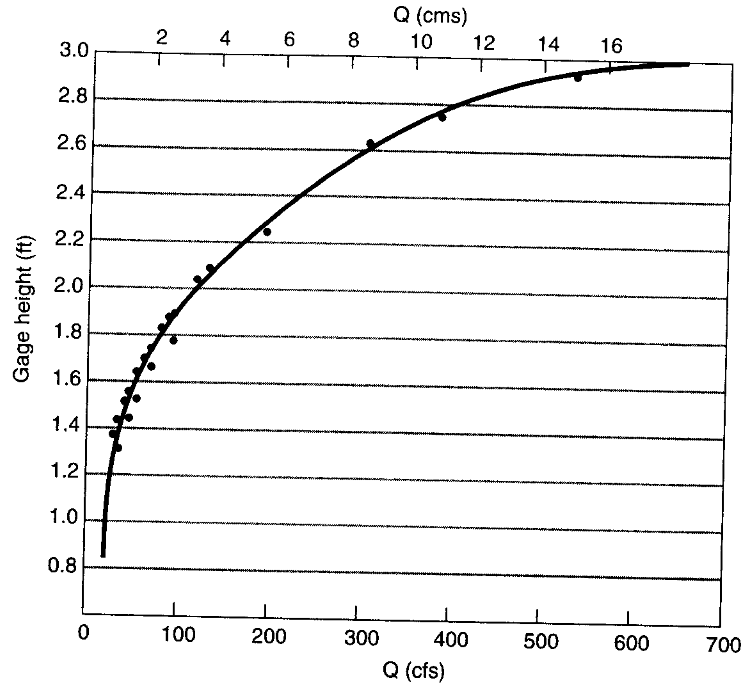
The major export from drainage basins to the global hydrologic system occurs as stream discharge. The volume of water passing a given channel cross-section during a specific time interval, is expressed as

$$Q = wdv = Av$$

where Q is discharge in m^3/s (cms) or ft^3/s (cfs), w is width, d is depth, A is area (wd), and v is velocity. Measurement of discharge is relatively simple by dividing the channel cross-section into segments of even width (fig. 5.32A). Depth and velocity are measured in each segment, and then total discharge is computed by summing the discharges of all of the sections across the channel. The most difficult measurement to obtain is velocity. Velocity varies with depth in open channels (fig. 5.32C), thus surface velocity is not very representative of the mean velocity. Average velocity is

FIGURE 5.33

Rating curve for low flow. Rock Creek near Red Lodge, Mont.



obtained with a current meter (fig. 5.32B) adjusted to measure the flow rate just below the halfway mark on the depth in each channel subsection.

To be of any scientific value, discharge must be measured repeatedly at the same locality; this gives hydrologists a better understanding of how flow varies with time. In the United States these sampling localities, or *gaging stations,* have been in operation at some sites since the late 1800s. Thus, a wealth of flow data for a large number of streams is available from the U.S. Geological Survey and numerous state water surveys. Common measurement frequencies vary from continuous recording to intervals of one hour. The average of these data are published annually as *mean daily discharge.* The *mean annual discharge* is the average of the daily values over the entire period of record, assuming that the station has been maintained for longer than one year. Actually, discharge is not frequently measured by the procedure described above because it is very time-consuming. Instead, discharge is estimated from a *rating curve* (fig. 5.33), or *rating table,* which relates a range of discharge values to the elevation of the river above some datum, called the *river stage* or *gage height.* Once the rating curve for a specific station has been constructed from a series of actual discharge measurements (the data points on fig. 5.33), only the gage height must be observed directly to be able to predict discharge.

Flood Frequency Geomorphologists are interested in the frequency and magnitude of flow events because each has an important bearing on how basins and channels evolve and function. Mayer and Nash (1987) compiled a collection of papers focusing on the geomorphic effects and frequency of catastrophic floods. The frequency of a given discharge is graphed as a **flow duration curve** (fig. 5.34) which relates any discharge value to the percentage of time that it is equaled or exceeded. At any station, then, the lowest daily discharge in the period of record will be equaled or exceeded 100 percent of the time. The largest flow will be equaled or exceeded only once out of the entire number of days in the sample, giving it a percentage value slightly greater than zero.

The most common approach to finding the frequency-magnitude relationship (especially for floodplain planning) is to consider only the peak discharge during each year (*annual series*) or only the discharges above some predetermined censoring value (*partial duration series*). The statistical samples in these time series approaches are much smaller than in the analyses utilizing daily records; however, they are use-

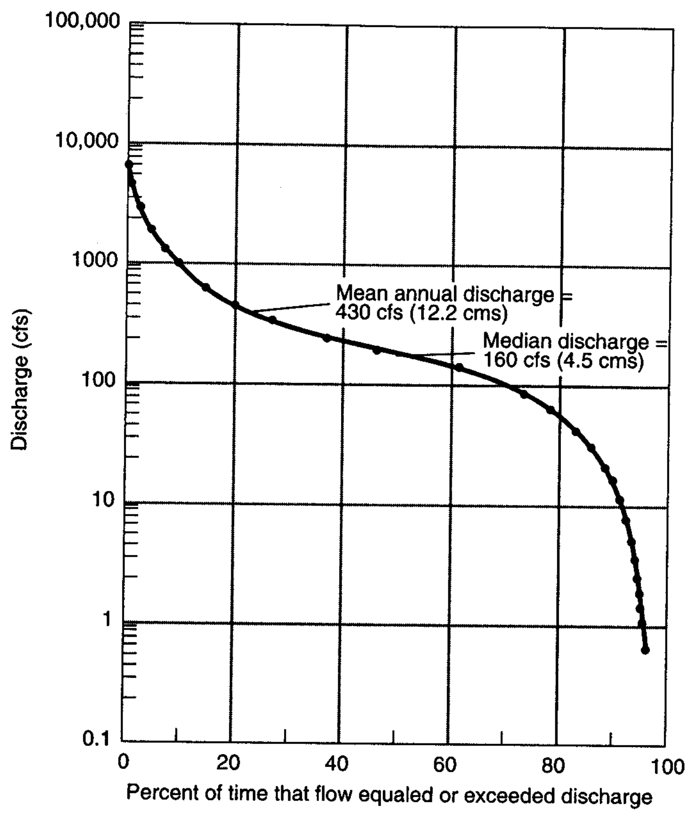


FIGURE 5.34

Flow duration curve for the Powder River near Arvada, Wyo., 1917–1950. (Leopold and Maddock 1953)

ful in studies of major flow events, provided the gaging station has a record of considerable length. The annual discharges shown in table 5.5 have been ranked according to the magnitude during the years of record displayed. The **flood recurrence interval** can be calculated from an annual series like this one using one of a number of formulas. The simplest plotting formula, the Weibull Method, calculates the recurrence interval by taking the average time between two floods of equal or greater magnitude:

$$R = \frac{n + 1}{m}$$

where R is the recurrence interval in years, n is the number of discharge values (the number of years in an annual series), and m is the magnitude rank of a given flood. The results of this type of analysis are then plotted on probability graph paper to show the relationship of discharge to recurrence interval (fig. 5.35), and the curve is used by hydrologists to estimate the magnitude of a flood that can be expected *within a specified period of time.* United States governmental agencies have adopted another method called the Log Pearson Type III (U.S. Water Resources Council 1981) which utilizes three parameters from the annual flood series distribution—the mean, the standard deviation, and the skewness. There is little agreement among the experts as to which of the numerous analysis procedures works best (for a review and discussion of procedures for analyzing flood frequency see Dunne and Leopold 1978; Singh 1987; Klemes 1987; Kumar 1987; Cunnance 1987).

The Pecos River flood frequency curve (fig. 5.35) shows that on the average the flow is expected to equal or exceed 2800 cms (100,000 cfs) once during a 25-year interval. Also from this curve, the 100-year flood would be estimated to have a discharge of about 27,400 cms (970,000 cfs) if the outlier is included. Bear in mind, however, that a flood recurrence interval estimate represents a probability statement and does not indicate that a flood of given size can only occur in time at regularly spaced intervals specified by its frequency. A river, unfortunately, does not understand statistical theory. Thus, there is no reason to believe that 25-year floods or 100-year floods will be distributed evenly over time. In the next 75 years, the Pecos River could experience three successive 25-year floods during the first 5 years, or there could be one now and several toward the end of the time period. Similarly, a flow equaling or exceeding the 25-year magnitude

TABLE 5.5 Annual discharge extremes and averages, 1900–1977 for the Pecos River near Langtry, Texas.

Year	Discharge (m^3/s) Minimum	Average	Maximum	Year	Discharge (m^3/s) Minimum	Average	Maximum
1900	7.3	—	2991	1939	3.9	8.1	162
1901	4.5	17.8	310	1940	4.6	8.8	157
1902	5.8	19.6	936	1941	5.7	51.4	523
1903	5.0	12.9	363	1942	7.4	21.8	264
1904	3.0	13.8	2013	1943	5.4	9.9	313
1905	14.2	40.8	1314	1944	4.4	8.1	250
1906	11.6	27.4	2516	1945	3.9	9.9	774
1907	5.0	12.9	327	1946	4.2	10.3	1817
1908	5.9	20.4	1901	1947	4.2	7.2	171
1909	4.2	10.1	727	1948	3.1	6.4	1434
1910	4.3	10.2	2851	1949	4.0	10.9	2753
1911	5.3	11.6	755	1950	2.8	7.2	1255
1912	3.4	6.4	512	1951	2.6	5.1	229
1913	3.8	15.4	1761	1952	2.3	3.9	100
1914	6.4	10.9	1873	1953	2.0	4.5	414
1915	10.2	32.9	1453	1954	3.0	78.3	27,392
1916	5.7	14.9	2711	1955	2.8	8.0	758
1917	4.1	8.3	176	1956	2.1	4.4	208
1918	2.9	6.1	1448	1957	3.3	15.1	1073
1919	3.5	30.8	243	1958	3.6	10.4	1075
1920	6.3	20.9	1219	1959	3.9	11.3	1328
1921	7.4	17.8	517	1960	2.9	5.7	134
1922	6.2	15.5	2152	1961	2.4	6.6	411
1923	4.2	6.9	358	1962	1.7	5.3	503
1924	3.6	10.0	1816	1963	2.2	4.1	88
1925	3.8	14.5	1705	1964	1.8	12.3	1037
1926	5.0	20.5	553	1965	2.9	6.5	436
1927	3.2	10.0	408	1966	2.9	7.4	433
1928	3.9	11.4	464	1967	2.6	—	60
1929	3.8	8.9	176	1968	2.6	5.0	626
1930	2.7	6.1	785	1969	2.3	5.5	172
1931	5.1	10.3	240	1970	1.7	3.7	62
1932	5.3	32.0	324	1971	2.0	13.6	2490
1933	6.7	11.9	126	1972	3.3	5.6	123
1934	3.8	6.8	229	1973	2.4	4.4	93
1935	3.6	4.4	2359	1974	1.6	41.9	16,128
1936	4.9	11.4	559	1975	6.0	9.3	274
1937	5.0	10.2	78	1976	3.9	10.8	1660
1938	5.3	12.1	881	1977	4.5	6.8	170

Source: Data from US Boundary and Water Commission.

may fail to occur at all. The *probability* that a flow of a given magnitude will occur during any year is the reciprocal of the recurrence interval:

$$P = \frac{1}{R}$$

where P is the probability of flow being equaled or exceeded in any one year, and R is the recurrence interval. Thus, a 50-year flood has a 2 percent chance of occurring in any given year, and a 10-year flood has a 10 percent chance. Over a 100-year period, the 50-year flood has an 86 percent chance of occurring while the 10-year flood has virtually a 100 percent chance of occurrence, and so on. These probabilities are computed using the probability equation

$$q = 1 - \left(1 - \frac{1}{T}\right)^n$$

where q is the probability of a flood with a recurrence interval (T) occurring in the specified number of years (n)

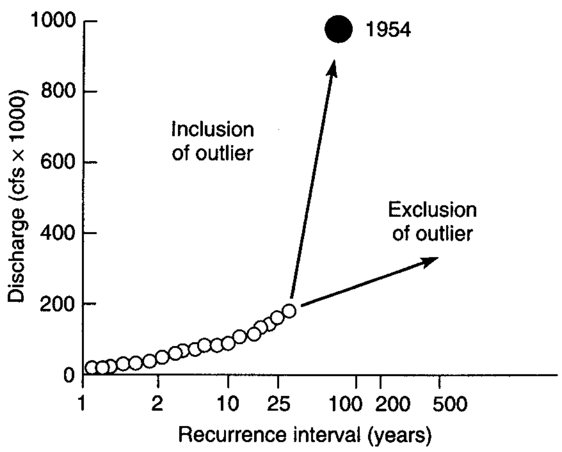


FIGURE 5.35

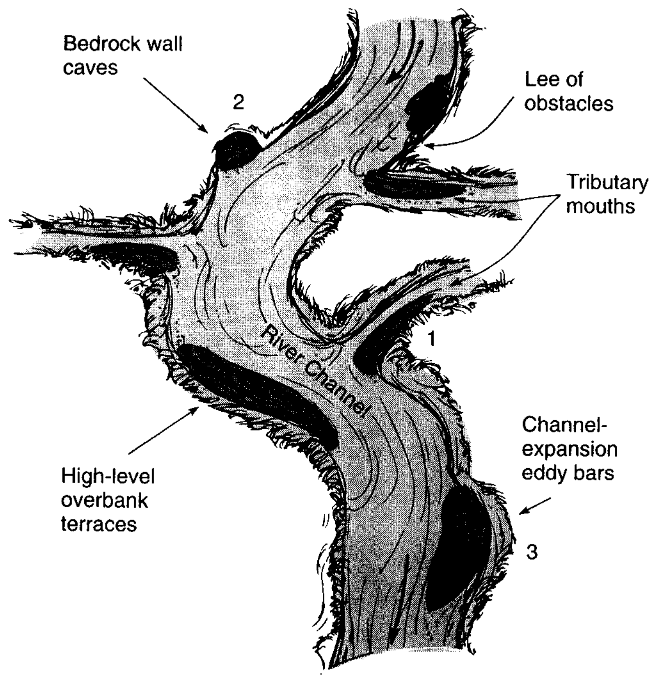
Flood frequency curve for the Pecos River near Langtry, Texas. Note the extreme outlier representing the 1954 flood and the problem for environmental planners of how to interpret this point.

(Costa and Baker 1981). One should realize, however, that the chance of experiencing a flood of any specified recurrence interval is never zero; there is always some risk, and predicting the year in which it will happen is virtually impossible.

Even though it may be impossible to predict when a given flood will occur, magnitude-frequency analyses have considerable practical value in river management, especially for higher-frequency floods such as the 20-year (q_{20}), 10-year (q_{10}), or 5-year (q_5) floods. Using the annual series, the *mean annual flood,* which is the arithmetic mean of all the maximum yearly discharges in the sample, is the flow that should recur once every 2.33 years ($q_{2.33}$). In our Pecos River example the mean annual flood has a discharge of 790 cms (28,000 cfs) and plots at a recurrence interval of approximately 2.33 on the flood frequency curve (fig. 5.35).

Paleoflood Hydrology Perhaps the single most difficult problem faced by hydrologic planners is to estimate the magnitude of the maximum probable flood that can be expected to occur within any given basin. Such estimates usually require extension of the flood frequency curve far beyond the limits provided by historically measured flow events. In other cases, estimates are required for basins that have very short gaging records or lack any kind of historical record of flow events. This is very risky business because critical decisions about the type and cost of flood control and protection are based on these estimates. Most hydrologists agree that simple extension of statistical flood frequency relationships beyond about 1.5 times the period of historical record are invalid. In the United States, gaged observations rarely extend beyond 100 years, thus making assessments of flood magnitudes beyond the 100- to 150-year recurrence interval very uncertain. Standard hydrologic methods of estimating peak flood flows from basins lacking flood series data are similarly suspect.

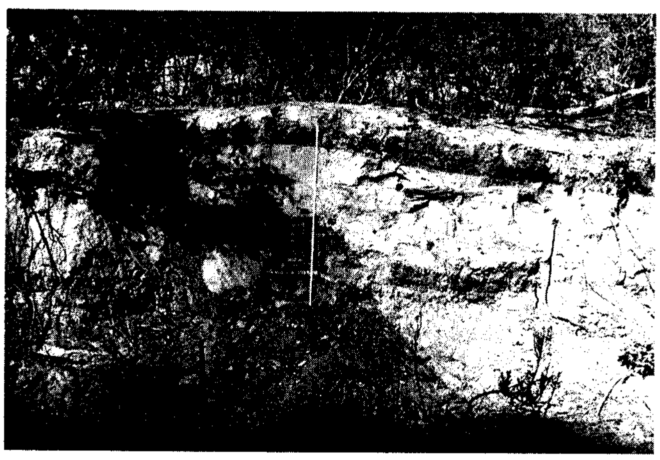
Recently, geomorphologists have developed a suite of techniques that rely on the sedimentary and botanical record for information about floods that occurred along a given stream prior to direct observation or measurement. This approach is called *paleoflood hydrology* (Kochel and Baker 1982). Damage caused by floods to trees growing along the channel and floodplain can be used to establish the time of specific flood events, often down to a single year (Sigafoos 1964; Phipps 1985; Hupp 1988). Sometimes, the height of flood-induced damages on trees, such as corrasion scars, can be used to estimate a minimum flood stage (Yanosky 1982), thereby permitting estimates of paleoflood magnitude using standard indirect techniques to reconstruct flood flows (see Dalrymple 1960; Benson 1971). Costa (1983) reviews a number of flow regime-based techniques for reconstructing paleoflood depth and velocity which use the dimensions of flood-transported boulders in combination with the characteristics of channel cross-section and channel slope. A number of techniques focusing on analysis of Holocene floodplain stratigraphy have also been used to reconstruct paleoflood history. These techniques commonly interpret the stratigraphic sequence of floodplain sediments and the relationships between tributary fan sediments and mainstream sedimentation to bracket time intervals of major flow events (for example, see Costa 1978; Patton, Baker, and Kochel 1979; Blum and Valastro 1989).



(A)

(A1)

(A2)



(B)

(A3)

FIGURE 5.36

Slackwater deposits. (A) Location of common sites for slackwater sediment accumulation and preservation. Locations 1, 2, and 3 are illustrated by photos from the Pecos River in southwest Texas. (B) Stratigraphic package of over 2000 years slackwater sediment in a tributary mouth site along the Pecos River. Some of the flood units contain logs and other organic material useful in dating paleoflood events.

One of the most successful and widely used stratigraphic approaches is based on the analysis of *slack water deposits* (Kochel and Baker 1982; Kochel, Baker, and Patton 1982; Baker et al. 1983), which are relatively fine-grained sediments deposited in areas of backflow or flow separation from the main current. Application of the slackwater method works best in narrow bedrock reaches along streams where large stage increases result from small increases in discharge and where channel cross-sections are not subject to major change during floods. During high-flow conditions, water is backflooded into tributary mouths, shallow caves, eddies in areas of flow expansion, and in the lee of flow obstacles (fig. 5.36). These areas are characterized by relatively low-velocity conditions where sediment is deposited rapidly from suspension, leaving a record of the flood event in an ever-growing stratigraphic section of flood sedimentation units. The time of deposition of individual paleoflood slackwater units can be derived by radiocarbon dating of organic matter deposited with the sediment. Fine-grained organics, commonly found at the top of a unit, provide the most accurate estimate of the flood because larger material such as a log may have been eroded from older flood sediments and redeposited by younger events (Kochel and Baker 1982, 1988).

Slackwater sediments can be used to estimate paleoflood stage by tracing a given unit up a tributary canyon and projecting its highest or terminal elevation back to the main valley (fig. 5.37). Experiments by Kochel and Ritter (1987) and field observations during historical floods indicate that these deposits provide a close approximation of the true paleoflood stage. Paleoflood stage information can be used to compute paleoflood discharge using a number of methods. The most common is combining the paleoflood stage information with channel cross-sections and profiles to model the paleoflood flow characteristics using the step-backwater method (Hydrologic Engineering Center 1982). For examples of this see O'Connor and Webb (1988); Baker (1987); Baker and Pickup (1987); and Jarrett (1991). New applications of statistical methods have made the use of discharge-censored data, such as minimum paleoflood stages, even more attractive (Stedinger and Cohn 1986; Stedinger and Baker 1987).

Slackwater deposits have been used to extend flood frequency curves over periods of 2000 to 10,000 years and have been used in a broad range of climatic zones (Moss and Kochel 1978; Patton and Dibble 1982; Kochel et al. 1982). When used in conjunction with the historical record, paleoflood data can provide realistic estimates of major floods that plot as *outliers* on flood frequency plots (see fig. 5.35) like that for the Pecos River (Kochel 1988). In this case, conventional statistical techniques for estimating flood frequency applied in 1980 provided estimates of the recurrence interval of the 1954 flood on the Pecos River ranging from 81 years

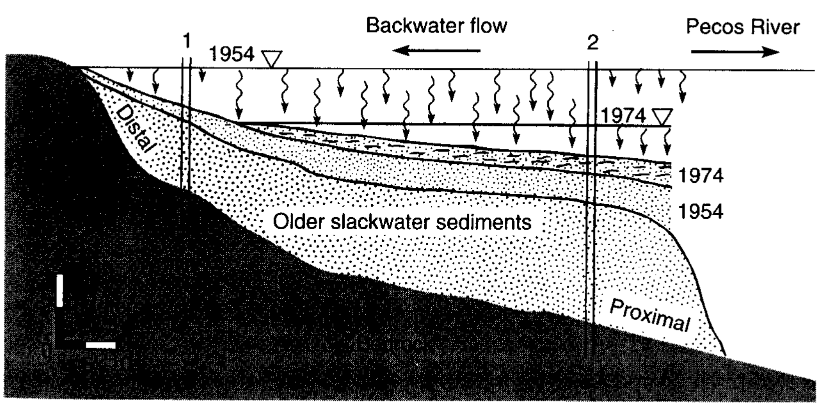


FIGURE 5.37

Schematic of on- and off-lap sequences and peak flood stage in a tributary valley for the 1954 and 1974 floods on the Pecos River, Texas. Sections in the proximal region (area 2) contain both floods, while distal regions (area 1) farther up the tributary record only the larger 1954 flood. Paleostage reconstructions are based on the elevation of the most distal sediments of each flood unit.

(Kochel et al. 1982)

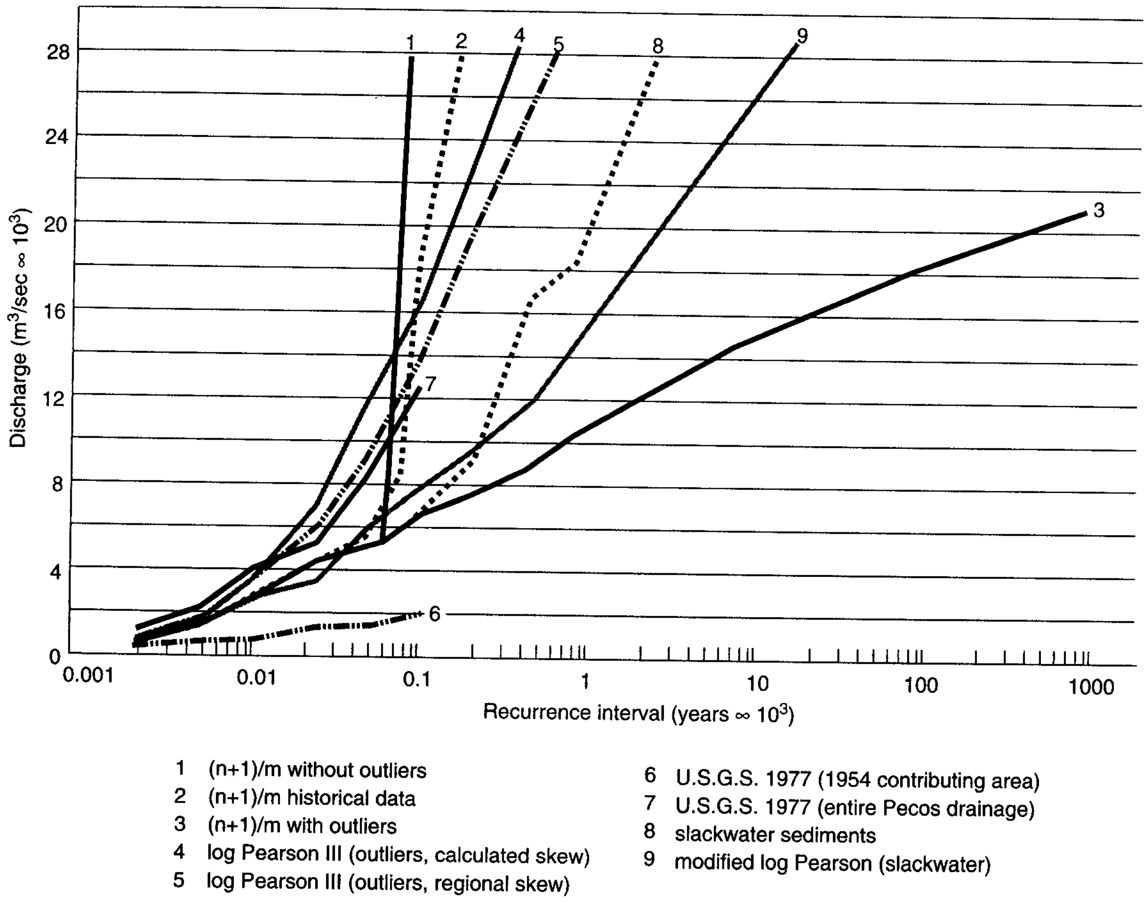


FIGURE 5.38

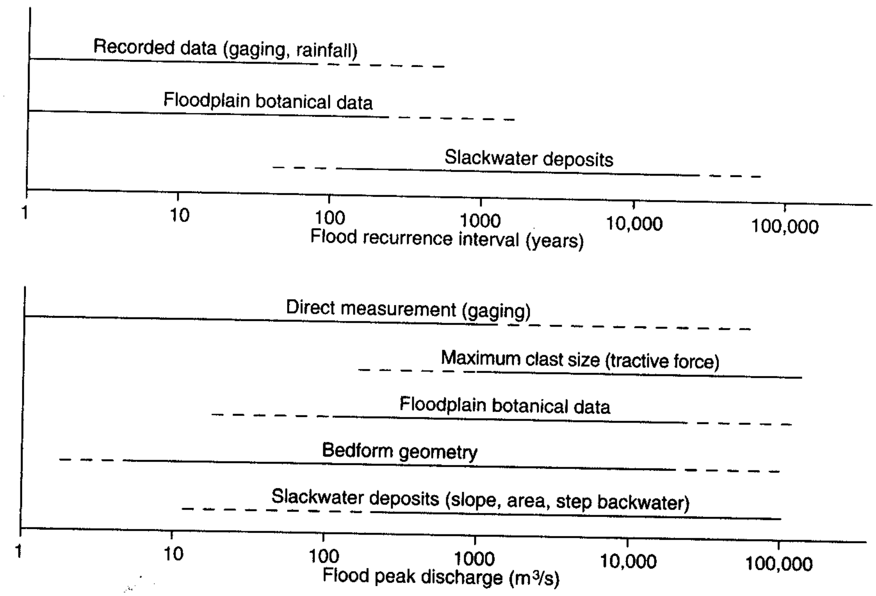
Range of potential flood frequency curves calculated using a variety of common standard techniques applied to the Pecos River flow data. Estimates of flood frequency for the 1954 flood outlier range from less than 100 years to more than 20 million years. Slackwater paleoflood deposits were used to provide a more realistic estimate based on physical flood evidence of around 2000 years.
(Kochel and Baker 1982)

to approximately 10 million years (Kochel and Baker 1982). With the inclusion of slackwater paleoflood data (fig. 5.38) in the analysis the estimated recurrence interval is revised to approximately 2000 years, which is a more reasonable value utilizing both the historical and stratigraphic record of flooding (Kochel et al. 1982).

Although a certain amount of error is associated with all of the paleoflood hydrologic methods (errors can be minimized through careful selection of field sites), they enable hydrologists to make reasonable estimates of the recurrence intervals for large, infrequent floods and are now frequently used as resources in planning decisions. Baker et al. (1987) used slackwater deposits to conclude that overdesign of flood control projects in Arizona's Salt River would be unwise given a record of Holocene flooding spanning thousands of years. Cooley (1990) applied paleoflood techniques to alluvial streams in Wyoming to improve flood frequency curves used in highway design. Successful techniques used in the Wyoming study included study of terraces, slackwater deposits, debris lines, dendrogeomorphology, and soils. Table 5.6 provides a summary of common paleoflood techniques and the range of applications.

A tremendous explosion in research related to paleohydrology has occurred since Schumm's (1965) seminal paper focusing attention on the prospects for Quaternary paleohydrologic studies (see, for example, entire books dealing with paleohydrologic studies such as Gregory 1983; Starkel et al. 1991). Paleohydraulic flood reconstructions have been used to gain perspective on the magnitude of some of the most catastrophic flows experienced in the Quaternary record such as the great Missoula floods responsible for carving the Channeled Scablands of eastern Washington (Baker 1973; Baker and Nummedal 1978; O'Connor and Baker 1992) and similar glacial lake-related floods in Siberia (Baker et al. 1993). Similar techniques have been used to detail the hydrology of Pleistocene lakes and breakout floods associated with the midcontinent portion of the Laurentide ice sheet (Kehew and Lord 1986, 1987).

TABLE 5.6 Common methods for estimating paleofloods.



Paleohydrological techniques also promise to play a major role in assessing the impact of human modifications on global climate, by facilitating the reconstruction of Holocene hydrologic regimes. Recent research has focused on fluvial responses to climatic change. This work is being done in arid bedrock channels (O'Connor et al. 1994) as well as in alluvial channels in semiarid regions (McQueen et al. 1993) and humid climates (Knox 1988; Patton 1988; Martin 1992). Paleoflood studies have been able to elucidate connections between climate and hydrology (Hirschboeck 1987) as well as the spatial variations in flooding between small basins (Martinez-Goytre et al. 1994). Knox (1993) presented a detailed reconstruction of paleofloods for the upper Mississippi River showing distinctive hydrologic variations during the Holocene. His data indicate that a drier and warmer climate prevailed from 5000 to 3300 B.P. Since then it has been cooler and wetter with more frequent large floods, perhaps similar to the devastating high water experienced in the summer of 1993 (fig. 5.39). Regional and even global correlations are now beginning to appear in paleoflood syntheses (Ely and Baker 1990; O'Connor et al. 1994; Smith 1992), indicating that these methods may be able to function as useful tools for reconstructing hydroclimatic variations during the Holocene. The apparent correlation of regional paleoflood events with climatic variations should be anticipated because of the high sensitivity of the hydrologic system to climate change (Barron 1989).

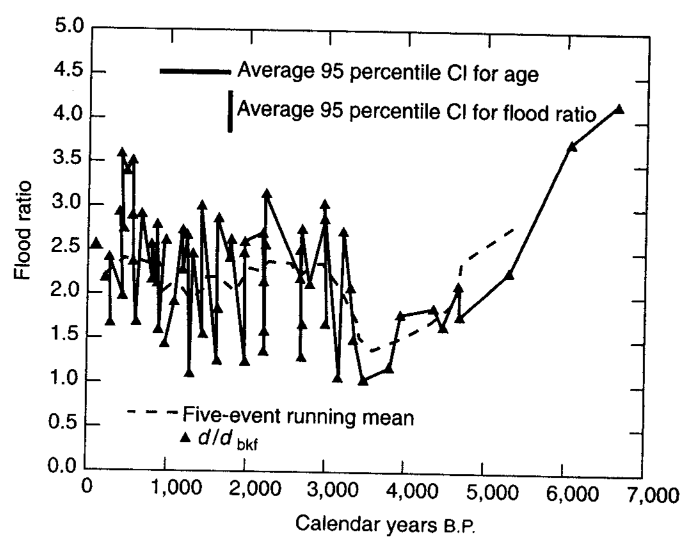


FIGURE 5.39
Holocene flood variations in the upper Mississippi River basin in response to small variations in climate.
(From Knox 1993)

Summary

In this chapter we examined a remarkable statistical balance between the spatial characteristics of river networks and the watersheds that contain them. Because the parameters of this morphometry also relate in a significant way to the hydrologic and erosional properties of most watersheds, drainage basins serve as primary units for systematic analyses of geomorphology. Drainage basins and river networks probably evolve according to fundamental hydrophysical laws, but their ultimate character is conditioned by the geological framework and the external constraints of climate. An equilibrium condition, defined in terms of mathematical balance, is probably attained early in the growth history of most basins. This does not mean, however, that basins evolve in an orderly way with time. Geologic catastrophes that upset equilibrium tend to be filtered out in a morphometric sense because basinal parameters apparently adjust to changes rather quickly.

Water flowing in basin rivers is derived from variable sources, and sediment reaching stream channels is produced and delivered by numerous erosive processes operating on basin slopes. Both water and sediment are amenable to budget analyses, which provide basic data for watershed planning. The amounts of water and sediment entering stream channels are functions of the physical properties of the basin and the effect produced by human activities. Estimates of basin denudation can be made on the basis of total sediment yield, but difficulties are created by sediment storage within the basin.

Suggested Readings

The following references provide greater detail concerning the concepts discussed in this chapter.

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