A View of the River

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CHAPTER ONE

The River Channel

The Grand Circle

We live on the surface of a planet that is in slow but constant change. The processes accomplishing that change operate because the planet is very special—special in position in the solar system and special in size. Earth moves in an orbit nearer to the sun than Mars, but more distant than Mercury or Venus. If Earth were appreciably closer to the sun, liquid water would not exist; it would occur only as vapor. And if Earth were much farther from the sun, water would be forever frozen. Moreover, Earth is just the right size, large enough to have a semimolten mantle from which volcanoes can erupt, bringing water vapor to the surface. Through this mechanism, it is believed, the oceans of the Earth were slowly developed. The moon is too small to have such volcanic activity and cannot form or hold either oceans or atmosphere.

Thus, by coincidence of favorable size and location in the solar system, Earth alone among the planets has oceans, an atmosphere, and thus a hydrologic cycle. The grand circle of movement of water from ocean to atmosphere to continent and back to ocean is the essential mechanism that allows organisms—including humans—to emerge, to develop, and to live on Earth.

Water plays a part in all physical and biological processes. It is essential to the actions that have developed the Earth's surface as we now observe it. Mountains are forced up by the collision of the great plates that make up the Earth's crust. But mountains on the continental surfaces are gradually worn away by the ubiquitous weathering of their rocks, and the transport of weathered products downhill by the action of water, wind, and gravity. The weathering processes that change hard rocks to erodible material incorporate water at every stage. Furthermore, water is the principal agent of movement of the weathered material that makes up the soil and supports vegetation, of the sedimentary rocks formed by the accumulation of the weathering products, and of the channels along which they are carried.

All the water presently on and in the surface of the Earth was brought there by volcanic action. What we see and use is derived from precipitation. That grand pattern of circulation of water called the hydrologic cycle describes in general terms what happens through time as water evaporates from ocean, plants, and soil, moves in the atmospheric circulation, and reprecipitates locally or far from its point of evaporation.

When precipitation falls on a continent, it separates into that which infiltrates the ground, that which immediately evaporates, and that which runs off the ground surface. The runoff carves or maintains channels of rill, stream, and river. This water on the surface may infiltrate, evaporate, or somewhere else be augmented by emerging groundwater. The terms "groundwater" and "surface water" refer merely to the location of water at a given moment. Water often moves between surface and subsurface depending on local conditions.

Rivers are both the means and the routes by which the products of continental weathering are carried to the oceans of the world. More water falls as precipitation than is lost by evaporation and transpiration from the land surface to the atmosphere. Thus there is an excess of water, which must flow to the oceans. Rivers, then, are the routes by which this extra water flows to the ultimate base level. The excess of precipitation over evaporation and transpiration provides the flow of rivers and springs, recharges groundwater storage, and is the supply from which humans draw to meet their needs.

A good deal of the water that appears as river flow is not transmitted into the river channels immediately after falling as precipitation. A large percentage is infiltrated into the ground and flows underground to the river channels. This process provides, then, a form of storage and thus regulation that sustains the flow of streams during nonstorm or dry periods of bright, sunny weather. The discharge represents water that has fallen during previous storm periods and has been stored in the rocks and in the soils of the drainage basin.

The excess of precipitation over evapo-transpiration loss to the atmosphere is a surprisingly small percentage of the average precipitation. The average amount of water that falls as precipitation over the United States annually is 30 inches. Of this total, 21 inches are returned to the atmosphere in the form of water vapor through the processes of evaporation and transpiration from plants. The balance of 9 inches contributes to the maintenance of groundwater and the flow of rivers.

About 40 percent of the runoff from the continental United States is carried by the Mississippi River system alone. The amount of deep seepage from groundwater to ocean is not known but is believed to be quite small, probably much less than 0.1 inch per year.

For the land area of the continent the water cycle balances: credit, 30 inches of precipitation; debit, 9 inches of runoff plus 21 inches transferred to the atmosphere. In the atmosphere, however, the budget appears out of balance because 30 inches are delivered to the land as rain and snow, but only 21 inches are received back as vapor by evaporation and transpiration. Accordingly, 9 inches of moisture must be transported by the air from the oceans to the continent, to balance the discharge of rivers to the sea. It is estimated that each year the atmosphere brings about 150 inches from the oceans over the land area of the United States and carries back 141 inches.

The precipitation represented by surface runoff, about one-third, flows from the hillslope or valley bottom to definite channels—usually to small channels that join to form larger ones, which in turn meet to form still larger channels. By convention, the smallest of these are called rills; they meet to form creeks, runs, or streams; then, at some undefined size, they are termed rivers. Each is fed water from two sources, overland flow to a channel and groundwater emerging at the channel boundary. In nonstorm periods, all the flow in channels derives from emerging groundwater.

Hillslope to Rill Head

Only recently has the change from overland flow on unrilled hillslope to definite rill or channel been studied. The distance from headwater divide to the upper end of the first rill may be great or small. The hydraulic conditions that lead to rill formation involve raindrop impact, erosion by raindrop splash, and depth of the overland flow. Rainfall impact on a film of water flowing overland splashes up sediment, which tends to fill and obliterate incipient rills or channels, a concept developed and measured by Thomas Dunne. Downslope, where the depth of overland flow is sufficient to shield the soil surface from the direct impact of falling rain, and where the intensity of sediment transport in the flow may be high, rills or small channels begin. This subject is elaborated by Dietrich and Dunne.

A few data are available from some areas of mixed grass and trees, including oak-grassland associations in the San Francisco Bay region,

Table 1.1 Distance from watershed divide to upstream tip of identifiable channel

Location	Distance to closest divide (ft)	Drainage area (sq mi)	Mean slope from rill head to divide
Contra Costa and Marin counties, California			
Briones No. 1	400	0.00074	0.20
Olema No. 1	65	.00003	1.00
Olema No. 2	50	.00033	0.27
Olema No. 3	90	.00033	.27
Sublette County, Wyoming			
Cora Hill A1	70	.00017	.17
Cora Hill A2	115	.00030	.16
Cora Hill A3	200	.00045	.12
Cora Hill A4	275	.00056	.11
Cora Hill A5	365	.00064	.08
Cora Hill B1	200	.00012	.09
Cora Hill C1	200	.00018	.11
Arroyo del los Frijoles, Santa Fe, New Mexico			
Caliente Arroyo	N.A.	.00006	.14 (approx)
Big Sweat	30	.0057	.045
Big Sweat	200	.0057	.045
Big Sweat	190	.0057	.045
Big Sweat	60	.0057	.045
Little Sweat	220	.0013	.045

piñon-juniper woodland in New Mexico, and grass areas in west central Wyoming. They are shown in Table 1.1. In central Kenya, however, unrilled hillslopes in grasslands can be as long as 1,500 feet. In such cases drainage area is difficult to define.

Shape of the Channel

The shape of the cross section of any river channel is a function of the flow, the quantity and character of the sediment in motion through the section, and the character or composition of the materials (including the vegetation) that make up the bed and banks of the channel. Because the flow exerts an eroding force per unit area, or shear stress, on the bed and banks, the stable form the channel can assume is one in which the shear stress at every point on the perimeter of the channel is approximately balanced by the resisting stress of the bed or bank.

A natural channel migrates laterally by erosion of one bank, maintaining on the average a constant channel cross section by deposition on the opposite bank. In other words, there is an equilibrium between erosion and deposition. The form of the cross section is stable, meaning more or less constant, but the position of the channel is not.

The effect of changes in bank material on channel form depends on the relative resistance of bed and bank material. As the threshold of erosion of the bank material increases, whether by the addition of coarse or cohesive sediments or by the presence of vegetation or bedrock, with no change in the bed material or discharge, the channel will be narrower. Thus channels with cohesive silty banks and beds will be narrower than comparable ones with sandy banks and beds.

Most rivers in cross section are not parabolic and they certainly are not semicircular. They tend more to be generally trapezoidal in straight reaches, but asymmetric at curves or bends. The appearance of rectangularity increases somewhat as the river gets larger downstream, since width increases downstream faster than does depth. Some typical cross sections are shown in Figure 1.1, where they have been drawn to different scales so that their widths on the page are the same. When cross sections are drawn without vertical exaggeration, the shapes tend to resemble channels in cohesionless materials. The relatively large width-to-depth ratio for the biggest river in Figure 1.1 is apparent. The asymmetrical cross sections at curves and bends are described in a later section.

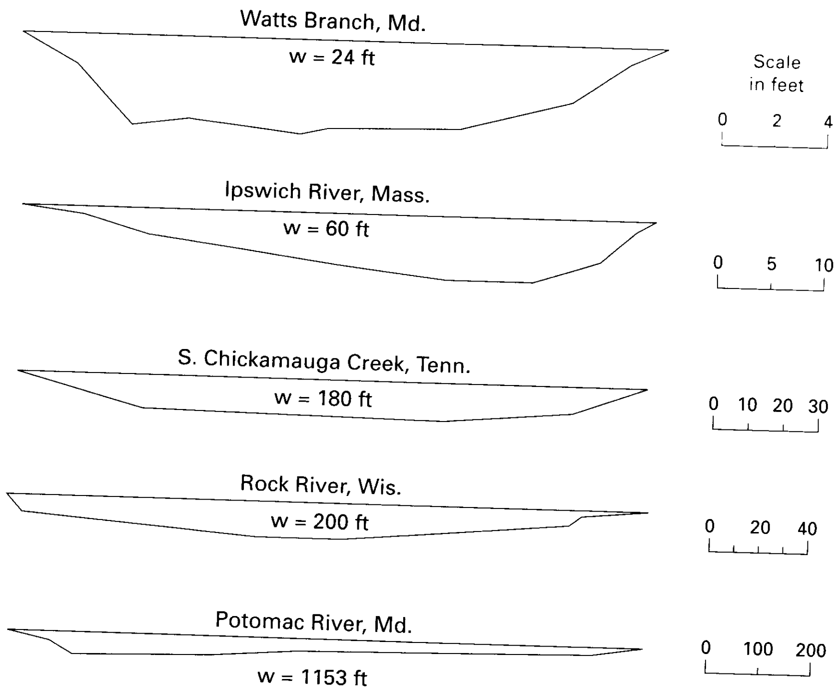


Figure 1.1 Cross sections of some natural rivers scaled so that the width appears to be the same.

The Floodplain

River channels are seldom straight except over short distances. A straight reach as long as 20 times channel width is a rarity. Curves, however slight, promote the tendency for erosion of the concave bank balanced by deposition near the convex bank. This tendency for erosion and deposition increases with the tightness of the curve; that is, with total angular deflection.

As the concave bank recedes due to erosion and the point bar builds outward from the convex bank, the channel width remains the same. The progressive growth of a point bar forms a flat surface or floodplain, the top of which is the level of the bankfull stage as indicated in Figures 1.2 and 1.3. The mean depth of a channel is computed as the cross-sectional area at bankfull divided by the water-surface width. The mean depth, then, is the height of a rectangle having the same area and the same width as the channel section.

A floodplain is built primarily by point-bar extension, as shown by measurements over several years at Watts Branch in Maryland (Figure

Figure 1.2 A typical floodplain, built by extension of the point bar at the right, as the concave bank at the left is eroded. Seneca Creek, Maryland.

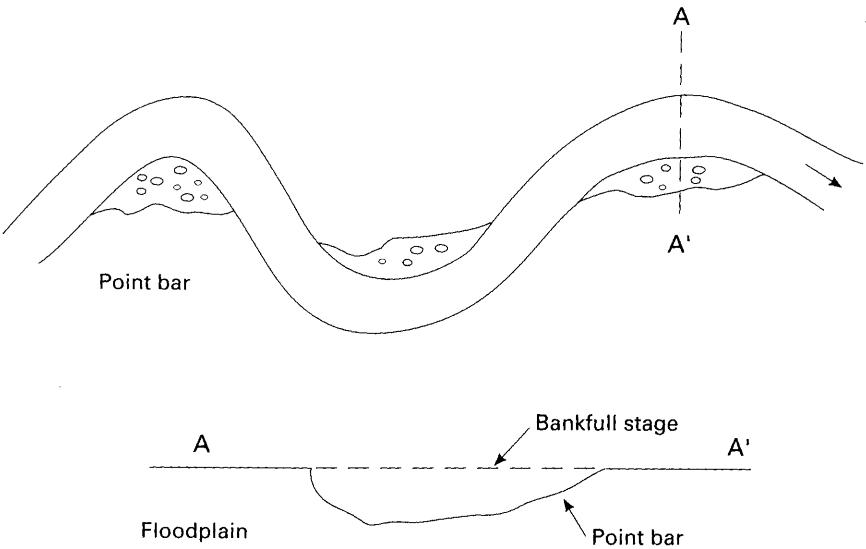


Figure 1.3 Diagrammatic plan view and cross section indicating that the retreat of a concave bank permits the extension of a building point bar. The bankfull condition shows that the level of the floodplain is the same as the top of the point bar.

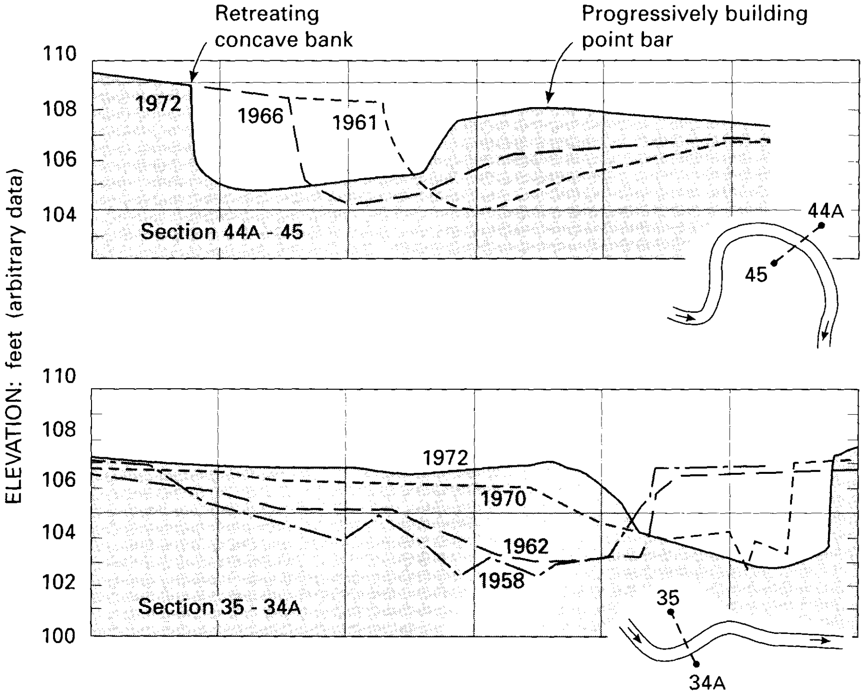


Figure 1.4 Data obtained from successive surveys of Watts Branch near Rockville, Maryland, show lateral migration of a river channel by the building of a point bar into the stream and concurrent erosion of the opposite bank. Continuation of such point-bar building results in the development of a floodplain. The diagrams at the lower right indicate the positions of the cross sections relative to the channel bends. (From Leopold 1973.)

1.4). In some instances, deposition by overbank flow adds to the floodplain level. During a climatic regime when active aggradation is occurring, overland deposition can be a primary process of floodplain construction.

Only a few definitions are really necessary to an understanding of morphologic processes in rivers. This is one: A floodplain is a level area near a river channel, constructed by the river in the present climate and overflowed during moderate flow events.

Note the phrase "in the present climate," because a floodplain can be abandoned and at least partly destroyed when climate becomes drier. An abandoned floodplain is called a terrace.

Channels and Climate

The consistency with which rivers of all sizes maintain the morphology typical of that climate is an indication that their channel is sensitive to the particular combination of discharge and load contributed from upstream. The river constructs and maintains its channel. The channel at any place is of such a size that the most sediment will be carried over a long period of time during those short periods when the flow is near bankfull.

The river channel responds quickly and sensitively to any change. Indeed, my own observations of channels in western states showed that streams in the semiarid areas changed from a state of erosion and instability during the first quarter of the twentieth century to a state of healing by vegetation in midcentury. Raw and unvegetated channels became stable and were gradually recolonized by vegetation beginning in about 1950. I found this to be the case in many western states. The change in upstream channels was apparently a response to the climatic shift at that time in the United States, England, and elsewhere. The British climatologist Hubert Lamb documented an increasing frequency and severity of storms after 1950 in the British Isles and northwest Europe. In the western United States the change brought on a slight cooling and a decrease in the number and frequency of intense short-lived rainfalls. Variability from place to place and from one season to another also increased.

This natural trend seems to have changed in the 1980s and 1990s, possibly because of the anthropogenic introduction of various gases into the atmosphere, and the poorly understood changes of ocean temperature in the equatorial region. The present outlook is for increasing air temperatures worldwide. The climatic changes of the past suggest that if the trend toward a warmer and more arid climate actually continues in the coming decades, the erosion of alluvial valleys seen in the thirteenth century, and again in the nineteenth, will be repeated in many of the semiarid areas of the planet where the rainfall is primarily of the thunderstorm type.

To understand how climate affects river channels, it is essential to perceive the primary difference between humid and arid locations in the semiarid parts of the globe. Figure 1.5 shows that the difference in the character of rainfall between a location with 14 inches annual rainfall (Santa Fe, New Mexico) and one with 8 inches (Las Cruces) is the number of small rains each year. The frequency of heavy rains is identical. Small rains foster vegetation and do not cause great discharges in channels.

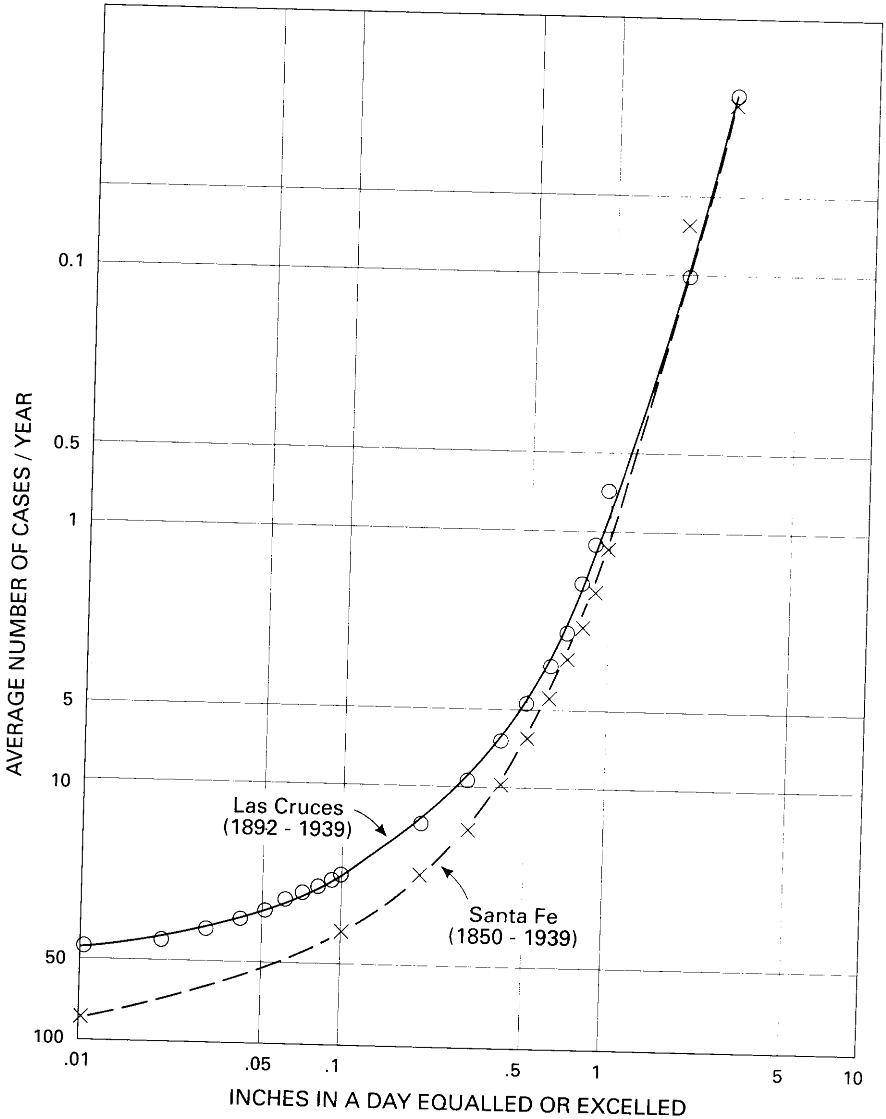


Figure 1.5 Cumulative frequency curve of daily rains, Las Cruces and Santa Fe, New Mexico. (From Leopold 1951.)

Several important relations are apparent from this graph. Las Cruces gets only 60 percent of the mean annual rainfall of Santa Fe, yet the frequency with which the two stations receive large rains (more than an inch a day) is about the same. The difference in total annual rainfall is determined by the larger number of small rains in Santa Fe.

In a semiarid region the mean annual rainfall of a given station is higher than that of another station because of the larger number of rains, more especially of small rains. A corollary to this fact is that at a given station a large percentage of total annual fall is contributed by the small rains.

The period of valley erosion in the western states, 1880–1920, was not characterized by a change in annual rainfall but by a change in rainfall intensity. The period of gully erosion saw many heavy storms and few light rainfalls. It was a period that can be described as more arid than others.

These types of changes have occurred several times in the Holocene period, the 10,000 years since the retreat of Wisconsin ice. Gradual warming occurred for the first 4,000 or 5,000 years of the Holocene, culminating in the Altithermal period of temperatures higher than today. This period, in Europe called the Climatic Optimum, was characterized by warm temperatures and boreal vegetation. In the U.S. Southwest it was the end of a long warming period, and in geologic terms it was a time of valley aggradation or deposition. At the end of this depositional period, a calcareous soil developed and is now seen as paleosol, marked by deposition of calcium carbonate in the B horizon.

The geologic evidence leads to the generalization that valley alluviation or deposition occurs during periods of relative humidity, except perhaps in regions of very high precipitation. Erosion and valley evacuation or degradation take place in periods of climatic aridity, owing to the prevalence of sporadic heavy rains and the infrequency of small, light rainstorms. These changes profoundly influence river channels. During periods of aggradation, widespread deposition increases the elevation of the valley floor, resulting in a floodplain built at a relatively high elevation.

When such a period is followed by relative aridity, channels cut down; a previously constructed floodplain is not only abandoned but dissected, leaving only fragments standing above the valley floor. Terraces, the remnants of previous floodplains, are mute evidence of changes in previous conditions, either in climate or in tectonic activity.

Terraces stand above floodplains in many areas of the world and can be seen nearly anywhere in the United States. I have studied such terraces

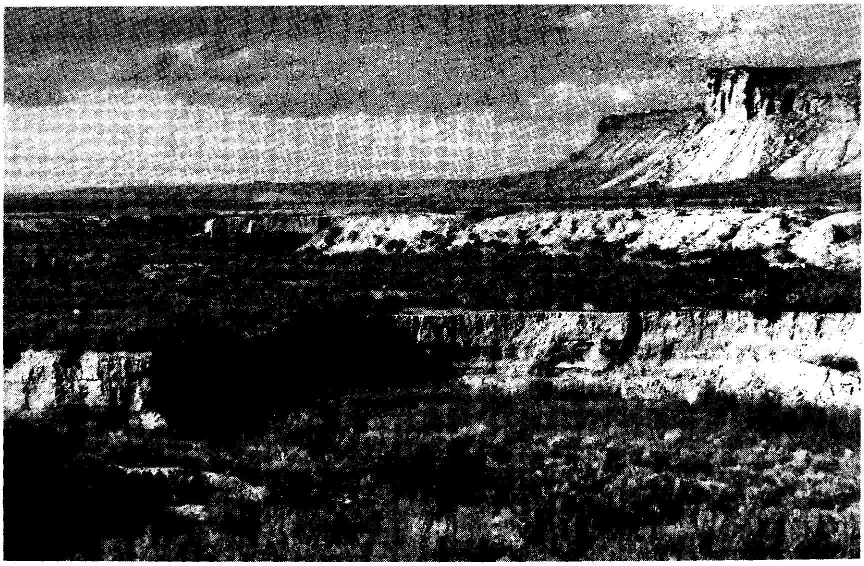


Figure 1.6A Three terraces in the alluvial valley of Salt Wells Creek, south of Rock Springs, Wyoming. Toward the back is the scarp of the high terrace, about 30 feet above the present creek; in the mid-foreground is the sage-covered middle terrace, 15 feet above the creek; at the front is the low terrace vegetated with tumbleweed, about 5 feet above the creek.

Figure 1.6B A terrace composed of materials of two different ages. The flat top, where growing plants can be seen, was once a floodplain. Later the stream lowered, cutting a wide channel. Subsequent climatic changes caused this channel to fill with the dark red silt seen in the center of the photograph. A more recent downcutting cut even deeper, exposing older material below the red silt. Rio Puerco near Gallup, New Mexico.

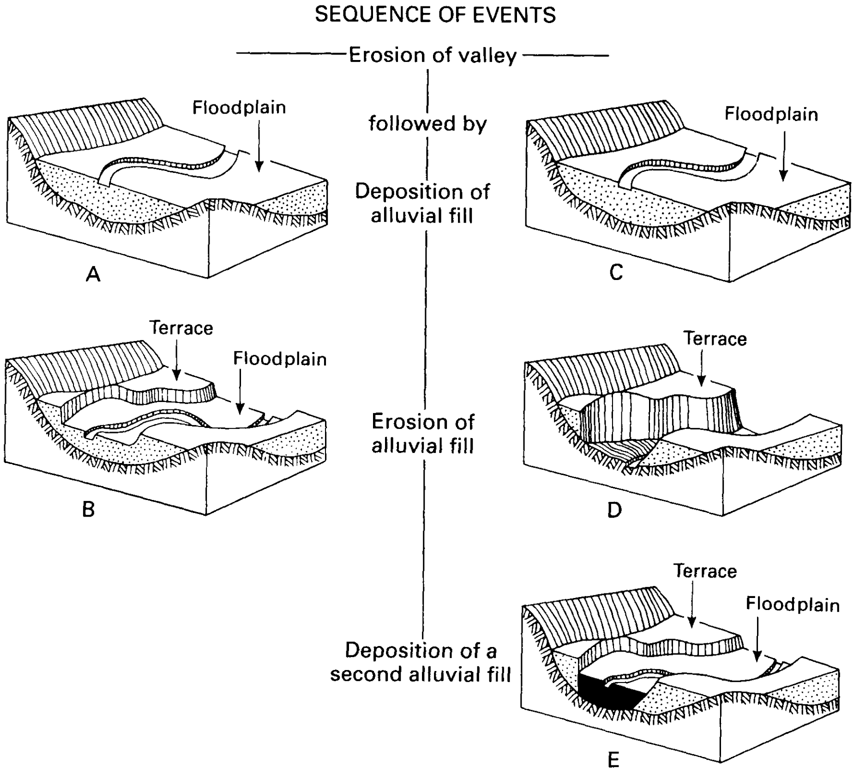


Figure 1.7 The stages in development of a terrace. Two sequences of events leading to the same surface geometry are shown in diagrams A-B and C-D-E.

in New Mexico, California, Colorado, Wyoming, Maryland and elsewhere. Particularly on the channels of small rivers and creeks, three levels of terraces are readily apparent. Figure 1.6 shows typical examples.

The sequence of depositional and degradational events is depicted in Figure 1.7. The difference between a cut terrace and a fill terrace is that the former results from interrupted downcutting of a floodplain with no intermediate period of aggradation. That sequence is shown in diagrams A-B of the figure. If a period of aggradation follows the downcutting, that surface, when abandoned, is called a fill terrace. The filling process is shown in diagrams C-D-E of Figure 1.7. Such sequences of cutting and filling can lead to a variety of valley cross sections, as illustrated in Figure 1.8, which shows sequences that lead to no terrace, one terrace, or two

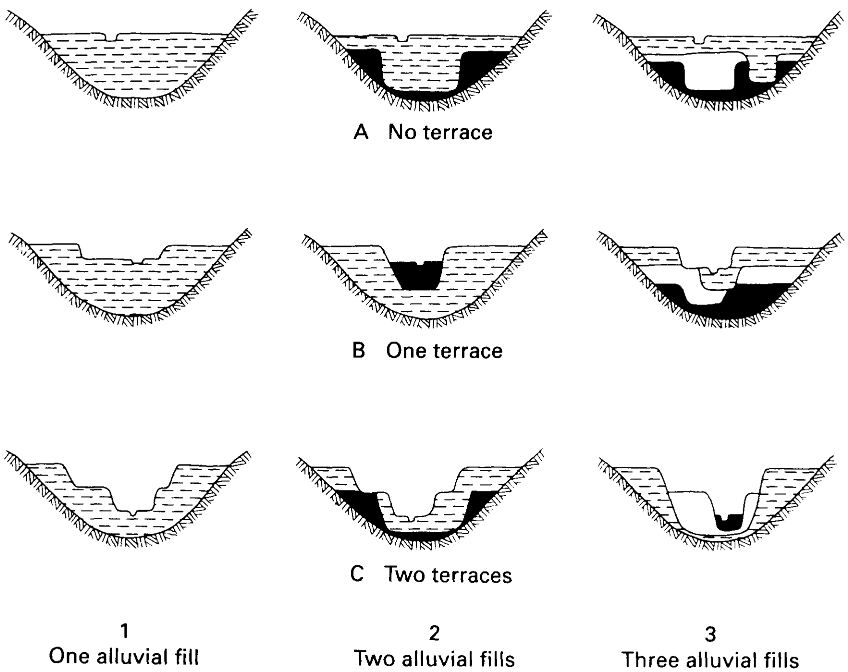


Figure 1.8 Valley cross sections showing some of the possible stratigraphic relations in valley alluvium.

terraces. Actual examples from various parts of the western United States are shown in Figure 1.9.

The geologic evidence of terrace levels in many western valleys validates the theory that a period of valley aggradation occurred from the end of the Ice Age up to the Altithermal period, 4,000 to 6,000 years ago. During the arid conditions of the Altithermal, erosion carried away much of the accumulated valley fill, leaving a terrace standing 20 to 30 feet above the present channel in many valleys. There followed another period of aggradation that ended in 200 years of drought, approximately A.D. 1200–1400. In this dry period, again the erosion carried away previous deposits of valley alluvium. Humid and cool conditions dominated the continent in the well-known cold period called the Little Ice Age, ending in about 1860. A turn toward aridity caused widespread erosion of western valleys between 1880 and 1920. These alternate periods of erosion and deposition left indelible indicators of past climates, and

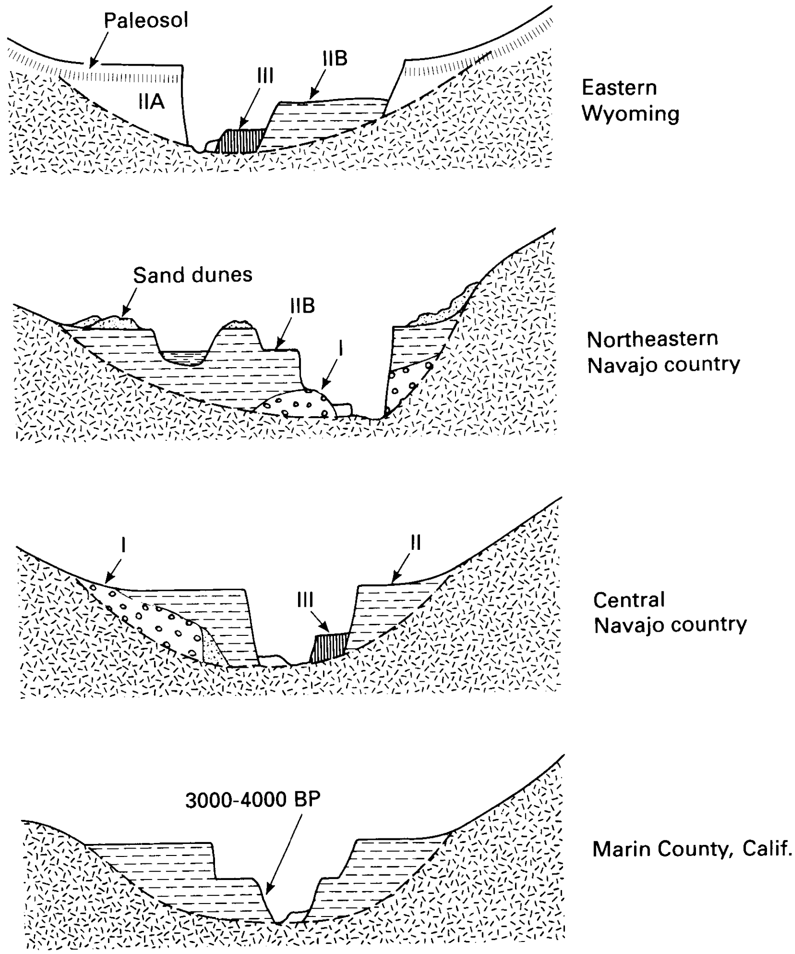


Figure 1.9 Generalized cross sections of alluvial valleys in four areas of the western United States. In each, previous epicycles of erosion have cut down to bedrock, as has the modern gully. The stratigraphic units are briefly described in Table 1.2.

Table 1.2 Alluvial chronology of western valleys, United States

Period and name (after J. T. Hack)	Character and date	Inferred climate
Deposition I, Jeddito	Included extinct animals; Pleistocene	More arid at end of period
Erosion	Sand dunes locally; paleosol of more arid type	More arid
Deposition II, A and B, Tsegi	Often subdivided into parts; paleosols at end of first phase; second phase contained artifacts dated as late as A.D. 1200	More humid
Erosion	1200–1400	Warmer; more arid
Deposition III, Naha	Ended with nineteenth-century erosion	More humid; colder near end
Modern gully erosion beginning about A.D. 1880	Generally intensified by overgrazing	Summer rainfalls exceptionally intense
Initial aggradation or cessation of gully extension	Began 1940–1960	Trend toward cooler; more precipitation in most but not all regions

knowledge of the sequence provides some indication of what may be expected as climate changes in the next centuries.

The chronology of erosion and deposition illustrated in Figure 1.9 is shown in simplified form in Table 1.2.

The four generalized valley cross sections of Figure 1.9 typify wide areas in the states included in the figure—Wyoming, Arizona, and central California. In each area the erosion preceding Deposition I cut down to bedrock. The sediments of Deposition I contain a well-developed caliche horizon and in some places contain extinct fauna such as camel and extinct bison, indicating a late Pleistocene age (that is, prior to 10,000 B.P.). Most of that alluvial fill was eroded before the subsequent Deposition II. Some of the oldest portions of Deposition II are marked at the top by a calcium-carbonate-rich paleosol attributed to the more arid climate of the Altithermal period, about 5000 B.P. The younger alluvium of Deposition

II contains Paleoindian artifacts and pottery dated in the interval A.D. 950–1300.

All the sections have certain characteristics in common: the wide valley floor, or most extensive level, is underlain by Deposition II material. This is true in the California section (as well as in the others), where W. W. Haible, whose observations furnished the section shown, obtained material near the base of the major alluvial fill dated at 3000–4000 B.P.

In all the examples the modern gully, as well as most previous periods of downcutting, eroded down to bedrock at least in some places. The deep and steep-walled, or boxlike, character typified previous gullies as well as the modern gully. The high terrace, constituting in many places the valley flat, stands 15 to 30 feet above the present or recent channel bed; and a middle terrace, where one exists, stands 6 to 10 feet above the streambed.

Modern gullies began downcutting in the period 1880–1900 as a result of climatic factors, especially an increase in intensity of summer storms, and exceptionally heavy grazing by stock.

An example of Deposition II overlying bedrock is seen in Figure 1.10. An example of valley trenching in the 1880–1900 period is shown in Figure 1.11, on the Rio Puerco del Oeste at Manuelito, New Mexico.

The generalized sequence shown in Table 1.2 and illustrated in Figure 1.9 is common, but not universal. Interest in the chronology and its dating has spurred research that shows examples of other sequences, but substantiates the effect of changes in rainfall intensity. With R. C. Balling, Stephen G. Wells showed that in a given drainage basin the sequence of filling and erosion may differ among small headwater tributaries and trunk channels downstream. The erosion process is not necessarily contemporaneous in all parts of a drainage basin.

The terrace sequence in any valley is important because a stream impinging on and eroding a terrace deposit produces a large addition to the sediment load for each lateral unit of erosion. Terrace remnants are often lateral constraints on stream movement and thus control the width of valley floor that can be utilized in flood periods for amelioration of flood peaks by channel storage.

Some Practical Insights Drawn from Alluvial History

The history of river cut and fill revealed in the stratigraphic relations in valley alluvium leads to some valuable insights into channel maintenance. The alluvial history shows that deposition leading to valley ag-

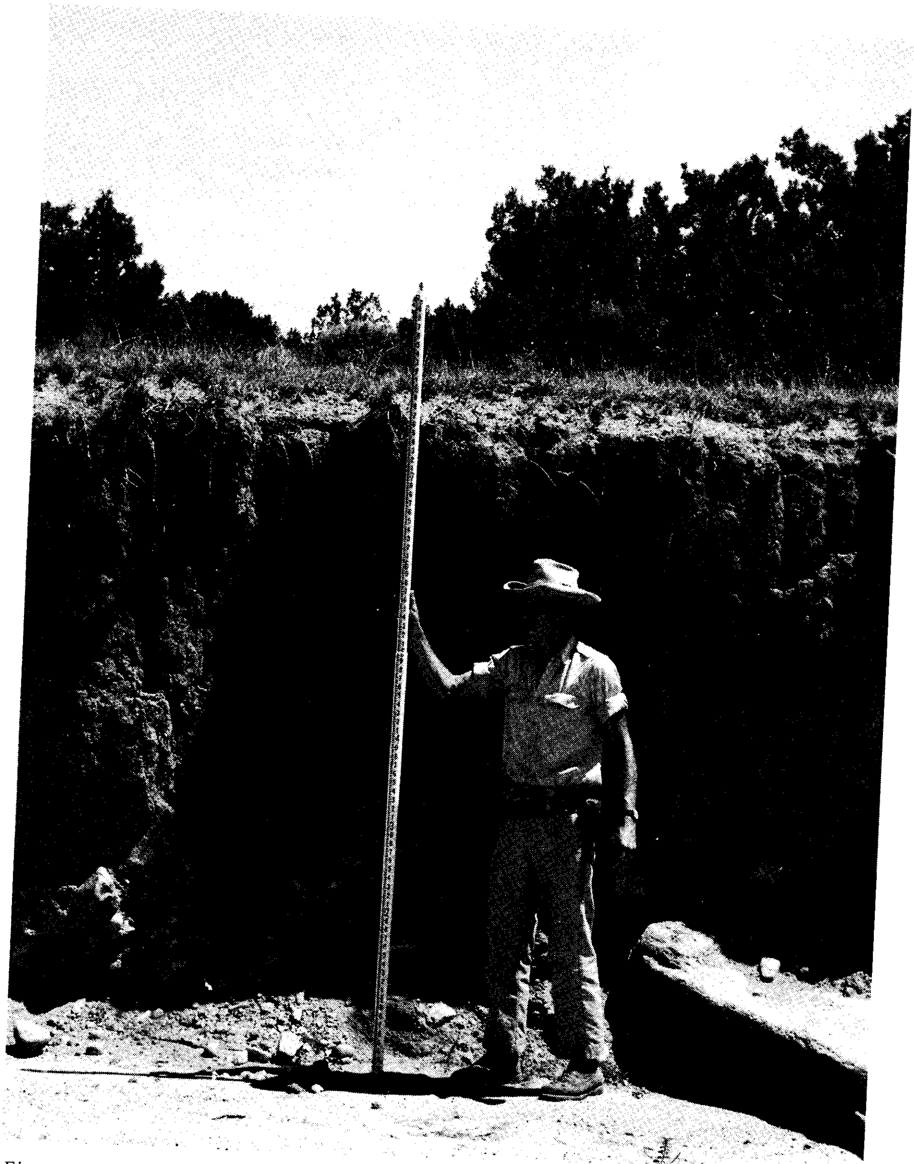


Figure 1.10 William W. Emmett at a base of the 8-foot terrace in Arroyo de los Frijoles near Santa Fe, New Mexico. Note the bedrock near his feet, overlain by gravel, then 6 feet of silty alluvium. Deposition of the alluvium was in progress in 2200 B.P. in Deposition II and probably began at the end of the Altithermal, about 5000 B.P.

Figure 1.11 The great trench of the Rio Puerco del Oeste at Manuelito, New Mexico, typical of arroyo cutting in southwestern valleys during the last decade of the nineteenth century.

gradation or alluviation is a slow process, whereas erosion is a rapid process. The valley fill accumulating in the early half of the Holocene took 4,000 to 5,000 years to fill western valleys with 30 to 100 feet of alluvium, a process of aggradation that ended about 5,000 years ago. In contrast, the deep gullies cut by erosion in pre-Columbian time took less than 200 years to evacuate a large part of the early Holocene fill. In the period 1880–1920, overgrazing and climatic change repeated the events of A.D. 1200–1400 in a period of less than 50 years.

In an effort to combat the erosion at the turn of the century in western states, government agencies including the Forest Service, the Soil Conservation Service, and experiment stations built thousands of small check dams in gullies. The results have been not only useless, but in some cases conducive to more erosion. Check dams can be useful only if a gully is deepening. Then they may provide a local base level to prevent further deepening. Check dams cannot store sediment because the volume to be stored is so small. The gradient of deposition behind a check dam is about half the gradient of the original valley, so the wedge of deposition extends upstream only a short distance. Check dams often fail by cutting around the dam; this lateral cutting enhances the erosion process and widens the gully in the vicinity of the failed dam.

As shown in Figure 1.10, many gullies immediately cut down to or near bedrock, so progressive deepening is not possible. Gullies cut by ancestral streams in the period A.D. 1200–1400 cut down to about the same depth as modern gullies.

Bank stabilization by vegetation is the best treatment of gullies in valley alluvium. It can often be facilitated by sloping the gully wall so that it no longer stands vertically. When such treatment is used, livestock should be kept out to protect the new vegetation.

It might be hoped that aggradation would fill the gullies cut during the 40 years between 1880 and 1920. Such filling has occurred in the geologic past, but the process is slow and takes many hundreds of years, and occurs only if climatic conditions are appropriate. At this point in time, the proper climate cannot be either forecast or influenced.

Classes of Channels

Channels differ in shape depending not only on size of river but also on climatic-geologic setting. As indicated in Figure 1.1, the width increases downstream faster than the depth. Large rivers are very wide and may even resemble a lake. The Mississippi at high flow is 60 to 80 feet deep, but a mile or more wide.

If we compare rivers of the same size, those in a coastal plain setting such as in Alabama or Georgia are relatively deep, and they are muddy from the suspended sediment. In contrast, rivers in semiarid regions are relatively wide and shallow, examples being the upper Rio Grande in New Mexico and the Platte in Nebraska. Such rivers tend to be wide because the bed and banks are sandy and thus erodible, having only small amounts of fine-grained sediment load. These differences are also reflected in the slope or gradient, the size of material on the bed, the sinuosity or extent of meandering, and the nature of the bank material. Because a river channel can be characterized by a particular combination of these shapes and pattern parameters, a channel classification system is possible.

David L. Rosgen has proposed, tested, and explained a river classification system that is currently the most widely accepted manner of describing a channel. The classification is based on parameters of form and pattern but has the advantage of implying channel behavior. It also can indicate how restoration might be approached if a reach of river becomes aberrant or different from its normal condition. The Rosgen system describes an individual reach—that is, a short length of channel—a few hundred feet or a quarter of a mile. The system does not describe a whole drainage system. Under natural conditions a given river may vary in character and thus in class, even through short distances downstream, as a result of passage from one lithologic type to another, tributary entrance, or change in landscape character.

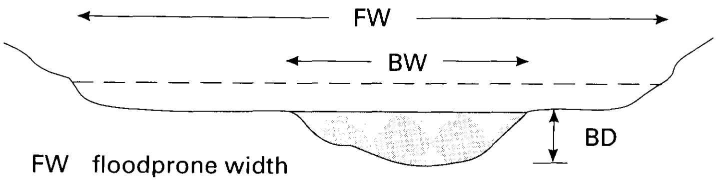
A river type according to Rosgen is defined by a particular combination of the following parameters: channel slope (gradient), bed material, ratio of width to depth, amount or degree of meandering as defined by the value of sinuosity, and degree of confinement or constraint to lateral movement. The classification system has seven types, A to G. In the simplified version considered here, each type has six subclasses that describe the size or coarseness of the bed material. Subclass 1 is bedrock, 2 is boulder, 3 is cobble, 4 is gravel, and so on. The total number of combinations is 42, but by far the largest number of channels fall in types A to D and subtypes 2 to 6, for a total of 20 most common field conditions. Figure 1.12 is an abbreviated explanation of the classification system and does not purport to include all possible types.

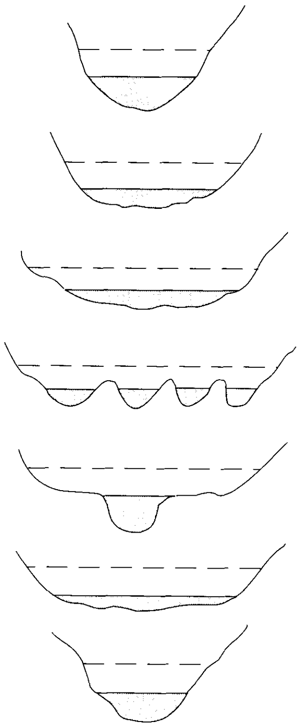
The preceding brief description of the Rosgen system cannot do it justice. It includes less common types omitted here. This summary merely indicates the variety of channel types that exist in nature and directs the reader to the Rosgen publication for a complete discussion of the implications.

Riffles and Bars

There are characteristics of river channels that are so general that they must be recognized in any discussion of morphology. A straight or nonmeandering channel characteristically has an undulating bed and alternates along its length between deeps and shallows, spaced more or less regularly at a repeating distance of 5 to 7 widths. The same can be said of meandering channels, but this seems more to be expected because the pool or deep is associated with the bend, where there is an obvious tendency to erode the concave bank. The similarity in spacing of the riffles in both straight and meandering channels suggests that the mechanism which creates the tendency for meandering is present even in the straight channel.

The alternating pool and riffle arrangement is present in virtually all perennial channels in which the bed material is larger than coarse sand,





channel type	slope percent	width to depth ratio, BW / BD	entrenchment ratio, FW / BW
A	4 - 10	< 12	1 - 1.4
B	1.5 - 4	> 12	1.4 - 2
C	1.5 - 2	> 12	> 2.2
D	± 1.5	> 50	> 2.2
E	< 2	< 12	> 2.2
F	< 1.5	> 12	1 - 1.4
G	2 - 4	< 12	1 - 1.4

Subtype number based on channel bed material

1	2	3	4	5	6
bedrock	boulder	cobble	gravel	sand	clay

Figure 1.12 The Rosgen system of channel classification.

but it appears to be most characteristic of gravel-bed streams—whether the gravel is the size of a pea or of a human hand. There appears to be a latent tendency for the development of pools and riffles even in boulder-bed channels.

Another longitudinal morphology exists in very steep channels in mountainous areas, particularly where the bed material consists of boulders and large rocks. In step-pool morphology the steps are often nearly vertical, the pools short and either deep or shallow. The spacing of the steps is much shorter than that of the pool-riffle channel, often 2 to 4 widths, but far less regular than the riffle spacing. The step-pool is also caused by hydraulic factors—that is, it is a natural phenomenon resulting from the transport and deposition of rocks by flood discharges. The causative mechanisms are not well known, especially the reason for the observed spacing.

A diagrammatic sketch of the plan and profile of the pool-riffle sequence and its relation to alternate bars is shown in Figure 4.14. The flow over a bar at high stage involves water actually being forced upward to rise over an obstruction, because the riffle is a mound on the streambed. At low flow some water sinks into the obstructing bar and flows through the gravel to emerge from the bed on the downstream end of the riffle. This sinking of water into the gravel is part of the reason that trout and salmon thrash their redd, or nest, out of the upstream part of a riffle; the downward flow into the gravel keeps the fish eggs from washing downstream.

Measurement of lengths of individual pools and riffles is not only a matter of judgment but is subject to considerable variation along a particular reach. A riffle bar in Seneca Creek at Dawsonville, Maryland, drainage area 100 square miles, is shown in Figure 1.13. A prominent riffle may be a bar adjoining one bank and sloping off to a deep hole at the opposite bank; it may be a central bar not flanked by deep holes; or it may be a low, off-center mound. In Seneca Creek the average length of one repeating distance is 324 feet, which is 5.1 times the mean channel width. The average length of pool in Seneca Creek is 1.6 times the length of riffle. The comparable figure for Watts Branch, a smaller stream a few miles to the south, is 1.1.

At low flow the water surface over a pool and riffle sequence tends to consist of alternating flat reaches of low gradient and steeper reaches often involving white water. This appearance of smooth water over the pool and riffles over the bar—terms well known to trout fishermen—led me to use these terms in describing the feature.

As the water rises during flood, the difference in appearance of the

Figure 1.13 A riffle bar on Seneca Creek near Dawsonville, Maryland, looking downstream at low flow. The head of the next pool can be seen in the background.

water over pool and riffle tends to disappear. At sufficiently high flow, about bankfull, the longitudinal profile of the water surface tends to become less stepped. Still, some difference in slope over pool and riffle remains. The riffle is then said to be "drowned out," a process that apparently occurs at smaller discharge in a meandering river than in an otherwise comparable straight reach of channel. The significance of the stepped or nonlinear profile at bankfull stage will be discussed when we compare it to the profile of a meandering reach.

In a study of the effect of diversion and realignment of certain gravel streams in Scotland on their ability to maintain trout, Tom Stuart noted that newly constructed streambeds dredged by a dragline were of uniform depth and without pools and riffles. With the aim of producing the usual pool and riffle sequence, in one river he directed the operator of the dragline to leave piles of gravel on the streambed at intervals appropriate to riffles—that is, 5 to 7 widths apart. After a few flood seasons these piles had smoothed out and in all respects appeared natural for a pool and riffle sequence. Moreover, the riffles so formed have been stable over a number of years.

Pole Creek, a mountain stream that has incised itself into a moraine of

Wisconsin age near Pinedale, Wyoming, has a coarse gravel bed derived from the moraine. Through this reach the stream averages 80 feet wide and 3 to 4 feet deep at bankfull stage. It exhibits alternating deeps and shallows, which in form are typical pools and riffles, but their spacing is variable and not clearly related to any function of width.

In many of the pools in Pole Creek boulders were conspicuously absent and the bed material was fine enough to be counted by measuring individual pebbles. To obtain a quantitative measurement of the concentration of boulders in the rapids or riffles, I counted the number of boulders equal to or greater than 3 feet in diameter in sample reaches of pool and riffle. The average number per 100 square feet of stream was zero in the pools sampled. In the nearby riffles the averages were 0.18, 0.27, 0.38, and 0.65.

Median grain diameter in pools varied between 0.04 and 0.4 foot, and local channel gradient from 0.002 to 0.013. The rapids, by contrast, were composed principally of boulders, which were measured individually in place on a sampling grid. The comparative median diameter was 1 to 2 feet, and the average slope through the rapids was 0.02. This same sampling method was applied to boulders seen on the surface of the moraine into which the stream was incised. The average number of boulders per 100 square feet on the moraine was 0.24.

Although these measurements are crude, they tend to support the conclusion that in Pole Creek the pools have a relative dearth of large boulders compared with the source material, and that boulders have been concentrated in the riffles by stream action. Thus, boulders must have been swept out of incipient pools and collected in incipient riffles.

At Seneca Creek, Maryland, I painted the individual pieces of gravel (0.25 to 6 inches in diameter) lying at the surface of a gravel bar during low flow when the bar was exposed. During subsequent high flows all the painted particles progressively moved, but the bar itself was the same height and topography as before. Some of the painted pebbles were found on the next riffle downstream. In these studies the movement of gravel of median size on the riffle requires a discharge that fills the channel about three-quarters full (depth equal to 0.75 bankfull depth), which has a recurrence interval of about one year.

In gravel-bed channels during periods of observation extending up to 7 years, we found no indication that the bars comprising riffles move downstream with time. Movement of gravel bars or riffles appears to be relatively slow.

One of the requirements for the existence of pools and riffles in nonmeandering streams is apparently some degree of heterogeneity of bed-material size. Channels that carry uniform sand or uniform silt have little tendency to form pools and riffles.

Alternate deeps and shallows also occur in rivers that are incised into deep canyons. The Colorado River in the Grand Canyon is one example. This river, as well as many others flowing between rock-walled cliffs, are noted for their rapids where there is a local steep gradient of the water surface. But rapids belie the existence of numerous deeps not apparent on the water surface. Figure 1.14 gives a profile of two reaches of the Colorado River upstream of the Grand Canyon.

In the Grand Canyon the steep rapids sections are spaced an average of 1.6 miles apart. Their locations are dictated by tributary entrances and faults. But local deeps are much more numerous, and this periodicity is probably due to the same factors that cause riffles and pools in more common river types. In the reaches illustrated in Figure 1.14, the average spacing of deeps in the 7-mile reach 187–194 miles, is 2,500 feet or 11.2 channel widths. In the reach of miles 199–205, the spacing is 2,100 feet or 9.6 channel widths. Although these spacings are larger than the expected 5 to 7 channel widths in successive pools of many streams, the occurrence of these alterations does not depend on the rapids and is dictated by quite different factors. It is my conjecture that the deeps and shallows of the Grand Canyon result from the same basic causes as pools and riffles in nonincised rivers, but in rock-walled canyons the channels are influenced by additional factors.

The Coarse Surface Layer

The extensive study of rivers worldwide in the last several decades has shown that most gravel-bed channels have larger cobbles or pebbles at the surface of the bed than in the layer immediately below the surface. This layer of coarse surface material has been called armoring or pavement. For example, Paul Komar reports that Oak Creek, a mountain river in Oregon, has a surface layer of 5 centimeters median size (50 percent of the material equal or finer), with the subsurface or next lower level, the subpavement, at 1.8 centimeters.

It is generally assumed that armoring results from the winnowing away of fine particles to leave a lag deposit of the coarse fraction at the surface. This is probably the process that causes surface pavement immediately below a dam where a channel is exposed to the discharge of clear water, the sediment having been trapped above the dam. Where pavement occurs in channels carrying the natural sediment load, other factors

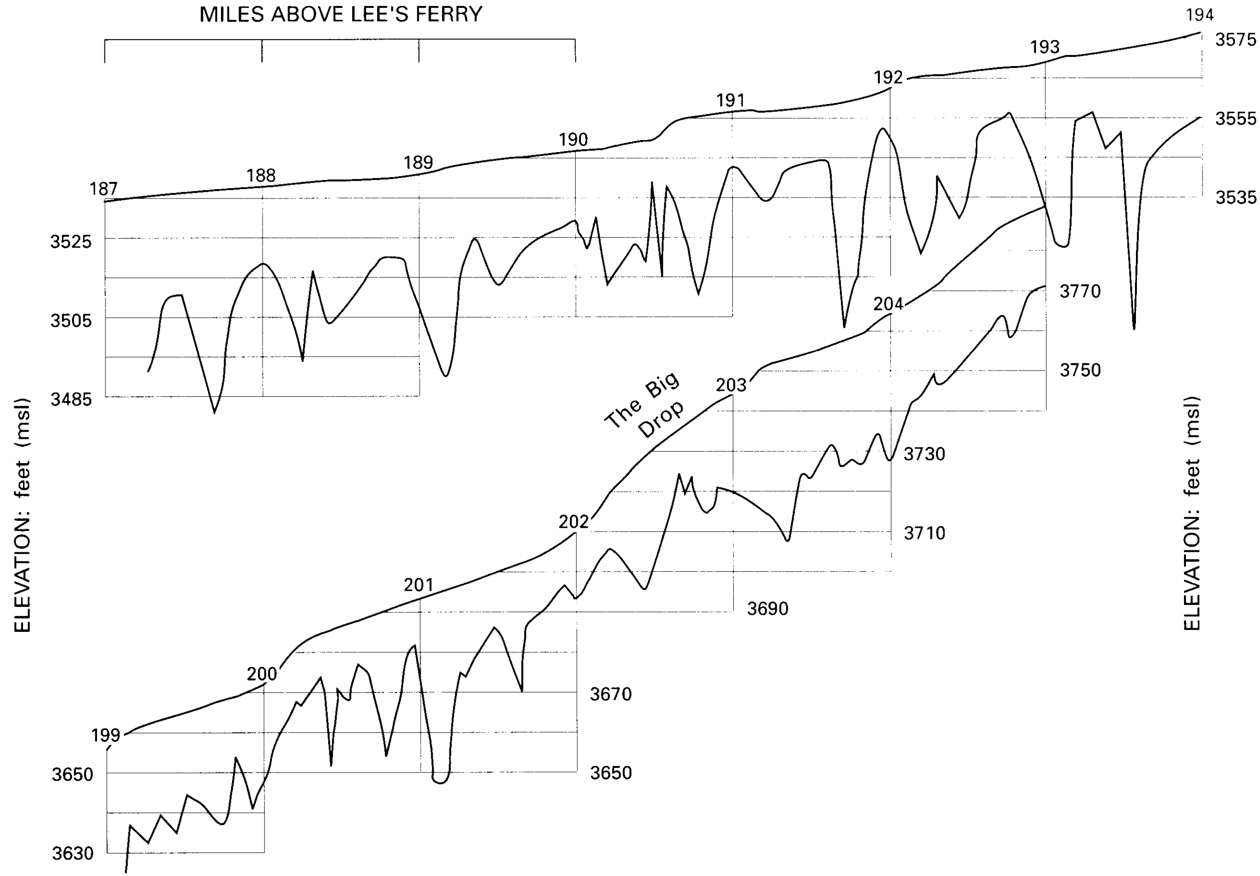


Figure 1.14 Profile of the water surface and bed of the Colorado River in Cataract Canyon, Utah. The depths were measured by the author during the U.S. Geological Survey expedition of June 1967 at a discharge of 38,900 cfs.

must be considered. At least in some relatively uniform environments, as in a laboratory flume, an increasing discharge flowing over a bed of heterogeneous size does not put in motion the smallest particles, then the somewhat larger particles, and finally the largest. Rather, when motion begins, nearly all sizes move at the same time.

Natural channels are characterized by their nonuniformity in topography and distribution of sediment sizes. T. E. Lisle and M. A. Madaj emphasized local variability in both the direction of sediment transport and the magnitude of the flow-induced stress. They found that degree of armoring is different in aggrading and degrading reaches. Armored locations appeared to be depleted of fine material rather than being enriched by coarse particles.

Various aspects of armoring are under study by investigators and remain a subject needing clarification. Another process, the action of dispersive stress, can lead to concentration of larger particles at the surface of a streambed. An example of the action of dispersive stress appears in sandy ephemeral channels viewed in the usual dry state. Pools and riffles are generally absent, though careful observation or detailed mapping discloses an analogous feature—thin surface accumulations of coarse material in the form of gravel bars. The distribution of these bars is strikingly reminiscent of the occurrence of pools and riffles in gravelly perennial streams, for they tend to be spaced at 5 to 7 widths along the channel length and remain there with only minor change from year to year. In these bars by far the majority of the cobbles are at or very near the surface; the sand below is quite free of rocks and cobbles. Such gravel bars, then, are mere surface features that we presume are caused by the same general process that accounts for riffles and bars in perennial rivers.

That large rocks accumulate at the surface of sandy ephemeral washes is particularly surprising in view of the fact that the channel bed scours at high flow and fills again to approximately the same level when the high flow ceases. You can observe the same phenomenon in the kitchen: place white flour in a cake pan and add some whole wheat flour. When you shake the two together, they will not mix. Rather, the shaking will separate the coarse whole wheat grain from the fine-grained white flour, with the larger grains accumulating at the surface.

In rotating-drum experiments conducted by various people to test the rate of abrasion of particles, the large particles were generally on top of the mixture. Ralph A. Bagnold explains this phenomenon as the effect of the intergranular dispersive stress. The knocking of particles against one another increases as the square of the particle diameter; hence differential stress on the larger particles may be enough to force them to the surface, where the dispersive stress is zero.

The same phenomenon occurs in dry granular material that flows under gravity. When a truck dumps dry gravel in a pile, the largest particles come to the surface, roll down the face of the conical pile, and tend to segregate themselves at the base.

The full significance of this phenomenon is not known. The concentration of the largest movable particles near the surface of the streambed seems to occur in a variety of channels in quasi-equilibrium. Whether or not the phenomenon contributes to the armoring observed in gravel rivers is unknown.

CHAPTER TWO

River Measurement

Need for River Discharge Data

The United States is unique in that more of its citizens are supplied with potable water than those of any other country. There is still a rural population that obtains water from individual wells, but even small towns have a water supply system. About 80 percent of the water used in this country comes from surface sources, the rest from groundwater. Nearly all public supplies are treated, with the treatments ranging from simple chlorination to full-scale filtration and chemical treatment. Furthermore, all public surface-water sources have been developed subsequent to hydrologic analysis of available streamflow data, thanks to far-sighted engineers.

Analysis of streamflow data can give good estimates of the types of information needed for water development and management. These include the amount of water available, the frequency of its deficiencies and excesses, the volume of storage needed for specific conditions, the frequency and magnitude of floods, the chemical and biological quality of supplies, and the duration of various low-flow condition.

All of these purposes are served by a network of river gaging stations and the published data obtained by their operation. The importance of data publication cannot be overemphasized; fortunately, there is essentially no proprietary withholding of river flow data. Credit for this laudable state of affairs is due the United States Geological Survey, one of the most important data collection agencies in the world. In the last part of the nineteenth century the engineers of this remarkable organization not only saw the need for a widespread uniform network of observation stations, they developed standard operating procedures, organized a training program for hydrographers, and instituted a publication program for producing and disseminating the data obtained.

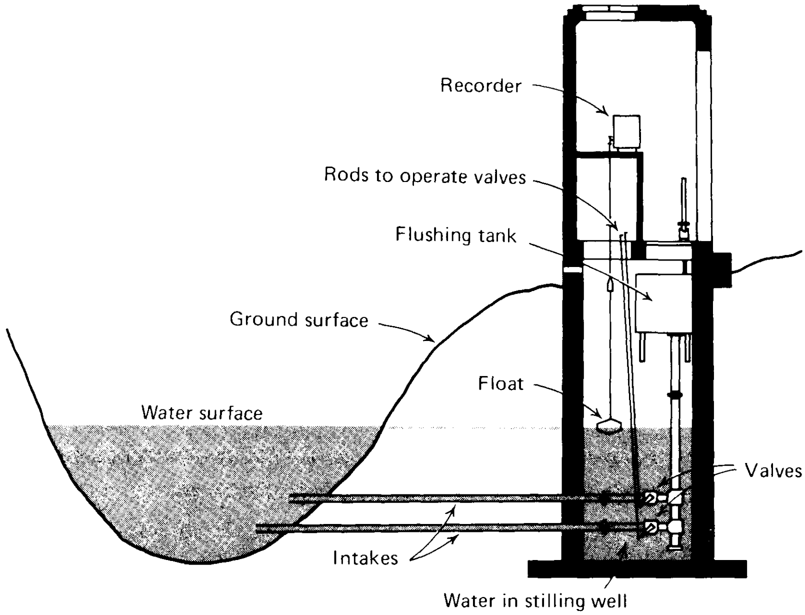


Figure 2.1 Diagram of a gaging station, showing the relation of the water in the stilling well to the river.

The Gaging Station

Driving along roads in city or country, beside a stream channel you will often see a vertical round corrugated tube with a conical roof. There are other types of gage structures as well—square brick ones, tall tubes, or square wood houses. The contents are virtually the same. The gaging station holds a device that records the changing height of water in the adjacent channel. The vertical housing is a water reservoir connected by a pipe to the stream, and the water level in the little house goes up and down in concert with the rise and fall of level in the channel. As shown in Figure 2.1, a float on the water surface within the house is connected to a device whereby the water level is recorded as a function of time. The recorder may be an ink trace on a paper chart moved by a mechanical clock. It may be a paper tape on which the water level information is punched, and it may also include radio or telegraph transmission to some distant office. There are other ingenious devices for countering the effect of ice in winter, for substituting water pressure for the usual float, and

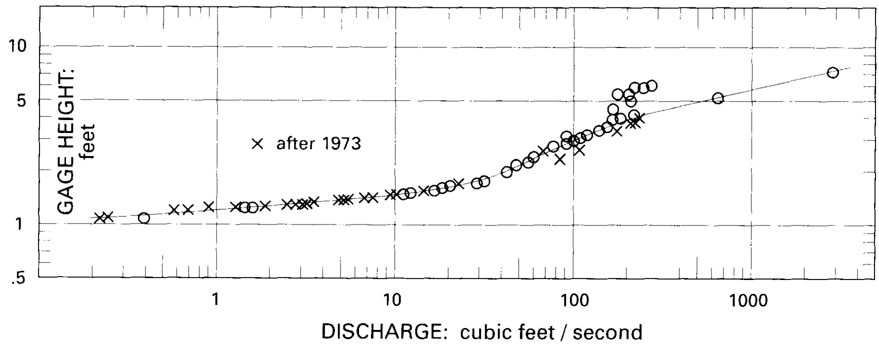


Figure 2.2 The discharge rating curve at Watts Branch, Maryland. The observations that deviate from the main curve near a discharge of 250 cfs were influenced by accumulations of organic debris at the gaging station near bankfull stage.

for obtaining samples of the water for later analysis. The essential item present in all stations is the recorder of water surface levels.

Water surface level is not equivalent to discharge, however. The cross-sectional area of flowing water and its mean velocity must be known. These are obtained separately, near the recording gage, from a bridge, from a trolley suspended from a cable, or by wading at low flow. The cross section and velocity are obtained simultaneously by measurement of depth and velocity when a current meter is lowered into the water.

In the United States the standard current meter has cups like a wind anemometer, whereas in Europe a propeller-type rotor is employed. In both situations the velocity is measured at about 20 to 30 positions across the channel and the discharge is computed for each position.

Associated with every gaging station is a gage plate or vertical scale on which the water level is measured. The height of water on this gage plate is called stage or gage height. When the total discharge for the cross section is computed, its value is plotted against the gage height. Successive measurements of stage and discharge are plotted in what is called the discharge rating curve (Figure 2.2). The measurements at this gage are very consistent over a period of nearly 26 years, except for 7 points close to bankfull stage at discharges between 180 and 250 cfs. These aberrant values were caused by logs and brush clogging the concrete measurement section.

Each plotted point in Figure 2.2 is a measurement of velocity and gage height, and thus discharge. At most gaging stations a measurement is made about once a month—except during storm periods, when the busy hydrographer makes as many measurements as he can at many gaging stations. Over a period of time the rating curve is developed, with the hope that it will include a wide range of discharge conditions.

The mean velocity of rivers in flood varies from 4 to 10 feet per second. The mean velocity attained in large rivers tends to be slightly higher than that in small rivers. There are, of course, many local situations where, owing to constrictions or rapids, velocity attains greater values. The figures cited above include a large majority of river channels in reaches that have no unusual features. For rivers of moderate size (2 to 100 square miles of drainage area), the flow at bankfull stage will ordinarily have a mean velocity on the order of 4 feet per second. If one had to make a guess without any measurement data, that figure would be a usable approximation.

The U.S. Geological Survey has analyzed individual velocity measurements made by current meter at the point of maximum velocity in river cross sections. The data were from routine measurements at 48 gaging stations on 27 large rivers throughout the country. A frequency table of 2,950 maximum values was compiled. Analysis showed the mean to be 4.84 feet per second, the median 4.11, and the mode 2.76 feet per second. Data on the Mississippi River constituted 13 percent of the sample and had a median value of 8.0 feet per second.

Less than 1 percent of the total measurements exceeded 13 feet per second. The highest velocity known to have been recorded with a current meter by the U.S. Geological Survey was 22.4 feet per second in a rock-bound section of the Potomac River at Chain Bridge near Washington, D.C., on May 14, 1932. Velocities of 30 feet per second (20 miles per hour) have been reported but were not measured by current meter. No greater values are known.

The Role of the Hydrographer

Each state now has a district hydrologist, whose office is usually located in the state capital. There may be several subdistricts that act as offices for portions of the state, because field personnel must travel to gaging locations within those subdistricts. Until the last few decades the hydrographer was a graduate engineer, but now individuals with technical training are employed. Hydrographers continue to bear a great responsibility, because the measurements they obtain are the basis for design and construction of projects costing millions of dollars. During flood periods these individuals are often at great risk under conditions of rain, snow, flood, or darkness. During storm periods the measurements made at high discharge are the unusual but highly important events contributing to the rating curve. In flood times where brush, trees, and even houses are floating down a high-velocity river, to be the sole occupant of a swaying open sled or cable car that hangs from a cable stretched across the channel is perilous. Yet these demanding conditions are common for all the hydrographic staff and are among the most important circumstances for data-collection purposes.

Discharge Records

In quieter times these persons are rotated into central offices to check each and every detail of the field notes taken during measurement duty. They also contribute to the compilation of the recorded data for publication and wide distribution.

A summary of each field discharge measurement is compiled on a form crucial to the study of rivers. This summary form, the 9-207, includes for each river measurement the date, width, cross-sectional area, mean velocity, gage height, discharge, and name of hydrographer. These forms are not published but can be obtained for any gaging station by application to the Geological Survey district office in the state.

The public record of streamflow is published by the Geological Survey in annual volumes. These volumes can be found in the government documents section of all large libraries, under the title *Surface Water Records* for each state. They can also be retrieved from computer data bases. Each currently operating gaging station is represented in the annual volume, and the main tabulation consists of the value of mean discharge for each day of the year at that gage. Additional descriptive data include geographic location of the gage, the drainage area, a brief history of the gage, the periods of record taking in the past, notes on diversion, and reservoir storage upstream of the gage. Also specified are the momentary peak discharges experienced during the year, data that are valuable for studying flood frequency.

With the advent of computers, modern record collection and publication has been much improved. Analyses formerly made by intensive hand computation are now routinely made and in many states are published in compact form. A typical example is the book titled *Streamflow Characteristics for California,* compiled by the U.S. Geological Survey. Especially useful are published values for the duration curve of each station, which indicate the percentage of time given discharge values are equaled or exceeded at the station. For many states the annual volumes now include summary data on suspended sediment and chemical quality of water for stations where such measurements are made.

Other highly valuable analyses have been published for various river basins in the United States. The series of volumes entitled *Flood Frequency* appear in water supply papers 1671–1688 of the U.S. Geological Survey.

The first gaging station was established on the Rio Grande in 1895, so the longest record is just short of 100 years. The number of gaging stations has gradually increased; today there are 7,590 operating stations in the United States. An early gage is shown in Figure 2.3.

The length of time to maintain a gaging station has always been debatable, because the number of possible gaging sites is so large. It could be argued that no one place deserves operation forever while other locations go unmeasured. Network design for attainment of maximum usefulness within the constraints of budget and personnel has long been a preoccupation of the Geological Survey. No perfect solution is possible. When the record of a specific station is long enough so that it can be correlated with other stations, and thus an estimate of flow is available by statistical means, then that station may be discontinued. The matter is complicated and will not be treated in detail here. But it is important to realize that many river gaging stations are so affected by the works of man that they do not provide a representation of the hydrology of the river basin. The flow of most rivers in the United States is now outflow from a reservoir, and because water is held over from one season to another the discharge record is not a record of the natural flow.

The Geological Survey has in place a network of chemical quality sampling carried out at some gaging stations, and a developing program of assessing biologic materials at selected stations. Somewhat similar sampling is done by the Environmental Protection Agency, but those programs are outside the scope of this book.

Some physical measurements not made by these agencies are needed at gaging stations that would be useful for monitoring long-term changes as well as for describing the morphology of the river channel. Extensions of present descriptions of the gaging site, they need be made only once, or perhaps once every 20 years. They include a permanently monumented cross section of the channel, a description of bed material made by the pebble counting method, a survey of the longitudinal profile of bed and water surface in the reach of channel, and the determination of bankfull stage by a standard uniform method. We call such a procedure a channel geometry survey (Figure 2.4).

In addition to the discharge and suspended sediment network, a small

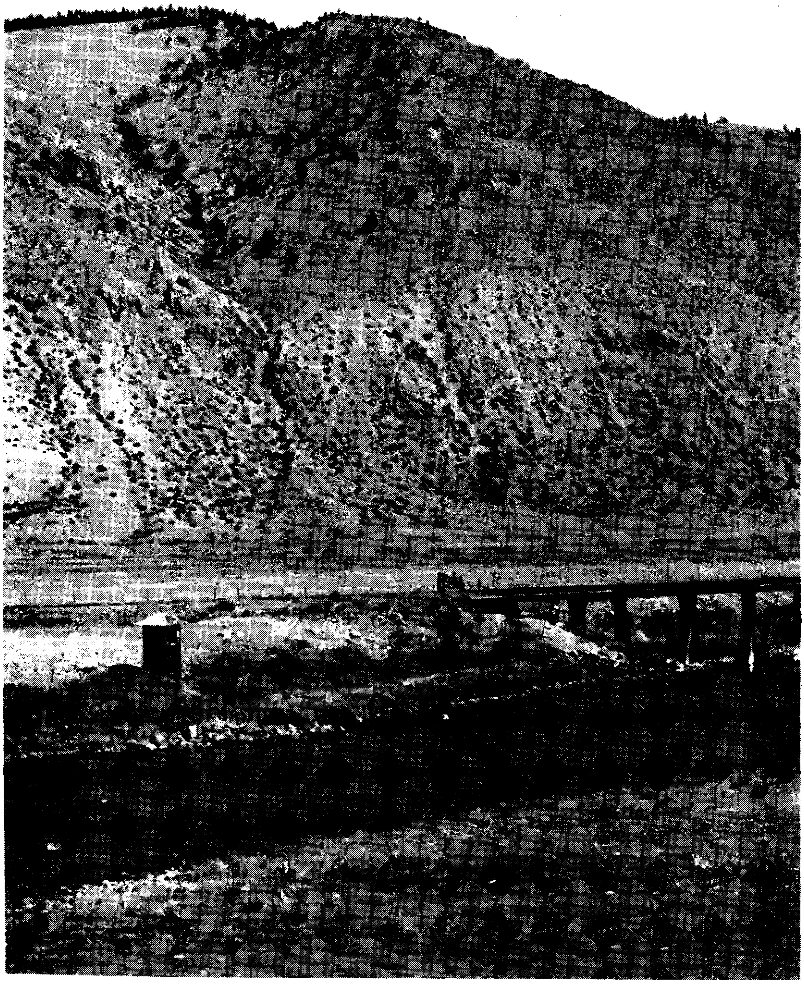


Figure 2.3 A patriarch of the gaging network: Rio Grande at Wason, below Creede, Colorado. The station was established in 1907 as a staff gage. The house seen here was built in 1910. A recorder was installed, and a cable was built at the gage house. When the station was discontinued in 1954, Lisle Alsdough of Del Norte moved the cable to another location.

net of bedload measurement sites is needed to include representative river channels in various climatic and geologic regions. Until recent decades such a network was impossible because no sampling equipment was available. But we now have a tested, easily available sampler called the Helley-Smith, which is simple to operate and practical for field use.

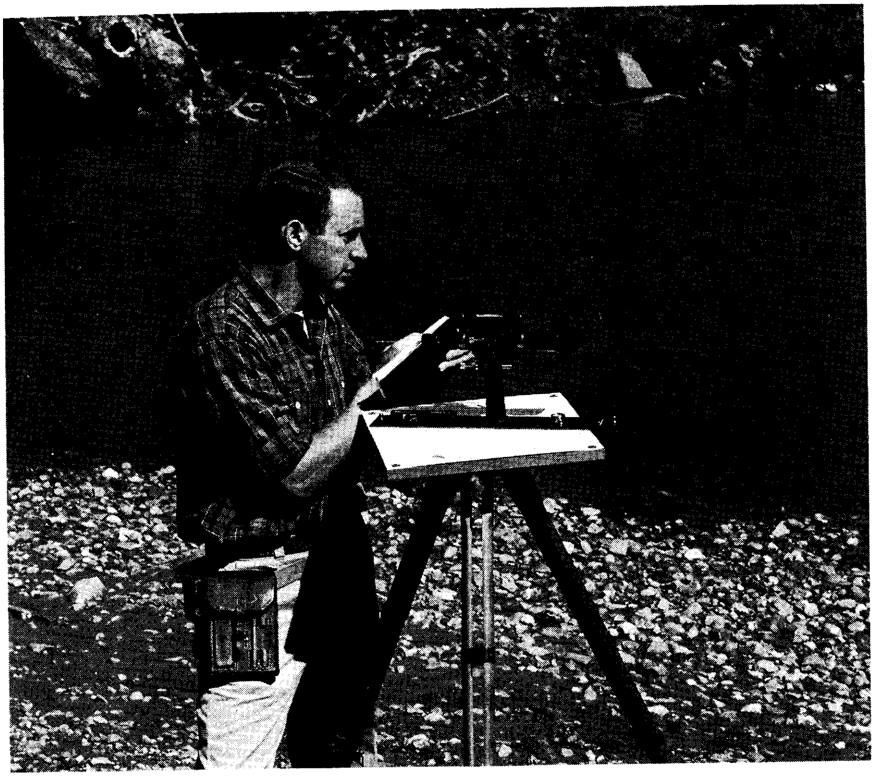


Figure 2.4 M. Gordon Wolman surveying on a typical point bar. Seneca Creek, Maryland.

The Bench-Mark Stations

In mid-century the Water Resources Division of the U.S. Geological Survey established a small number of gaging stations carefully located in basins protected from human activities. These sites included national and state parks, monuments, and other small areas without roads, clear-cutting of forests, or mining. The purpose of these stations is to provide a minimum number of stream locations that reflect the natural hydrologic reality, and to record the effects of any climatic changes that occur. It is hoped that these few stations will be operated without interruption for a very long time.

Fortunately, successive administrators of the Geological Survey have realized the importance of records uninfluenced by man-made alterations of the drainage basin. The number of such bench-mark stations remains small, only 57, but has not decreased even in times of budget stringency.

Backyard Hydrology

You can learn a great deal about hydrology by engaging in the interesting and rewarding activity of direct observation. Hydrology is a discipline that requires hands-on collection of data. Observation, the very essence of the hydrologic business, can be as simple as placing a thermometer on the back porch and reading it every morning. Of higher value and sophistication is buying a plastic rain gage for a few dollars and putting it on the front lawn, to be inspected each time it rains. A rain gage can also be made from a coffee can. In this case, buy a baby bottle from the dime store; most such bottles have gradations in cubic centimeters. You can pour the rainwater from the can into the bottle, record the volume in cubic centimeters, then convert the volume into inches of rain.

A still more sophisticated procedure is to place a staff gage in the nearest creek, gully, or channel and read the gage at certain times. If you are fortunate enough to live near a creek or have your office near one, it might be possible to place the gage where it can be seen from your window and read through binoculars. You can make a staff gage out of any old piece of wood, lettering or burning into it a scale of gage height in feet or centimeters.

Once you are reading a staff gage, you are into the crux of hydrologic science, because it is then necessary to construct a rating curve and measure stream velocity. Obviously this requires that a cross section be surveyed, a bench mark established, and velocity measurements made at various stages of flow. The easiest way to measure velocity is by floats, and the best float is an orange peel. It has just the right specific gravity to float nicely at the surface, it is brightly colored and thus easily seen, and it is readily available.

Choose a fairly straight reach of channel. Measure a distance along the bank of at least one width—but two or three widths would be better. Mark the beginning and end with stakes or colored markers. Pace along the bank or measure with a tape the distance between the markers. A stride (right footprint to right footprint) is for most people close to 5 feet.

You will use at least five floats, so have a few extra. Throw one upstream of the first marker and start counting seconds as it passes the marker. Measure time in seconds with a stopwatch, with the second hand of your wristwatch, or by counting out loud—one, and two, and three . . .

Note the time at which the float passes the lower marker and write that time in the field book. Throw your floats at various distances from the bank, so that you will measure velocity in different places in the channel width. Average the times in seconds. Divide the distance by the time in seconds to get a mean surface velocity in feet per second. Multiply this by 0.8 to get the mean velocity in the cross section.

This velocity must be multiplied by a cross-sectional area to get the discharge. To get the cross-sectional area, take a rule or meterstick or rod marked in appropriate units, preferably tenths of feet, and at uniform distances measure the depth of the water as you wade across. Take no less than 10 measurements; 15 would be better. The depth should be in even centimeters or to 0.05 foot. That is, readings might be 0.05, 0.55, 0.65, 0.80, 1.15, and so forth in feet; in centimeters they might be 6, 10, 14, and the like.

Measure the water surface width with care, using a tape or pacing carefully. Average the depths, multiply by the width to get the area in square feet. Then multiply by the mean velocity to get the discharge in cfs. Record all the figures as you compute them, writing them in your notebook not on scraps of paper.

Record the discharge with the staff gage reading taken at the same time. The gage height plotted as ordinate against the discharge plotted as abscissa on double-log (log-log) paper is the beginning of the discharge rating curve. For double log paper the line through the plotted points usually has a slope of about 0.34.

An excellent way to begin this kind of field observation is to make some preparations in advance of rainstorms. Install a staff gage, survey the cross section by wading, then wait for a storm. When rain appears imminent, put the plastic rain gage out on the lawn in an open space. From under an umbrella, watch the water at the staff gage. Read the rain gage every 10 minutes. When the water begins to rise, read the staff gage every 2 minutes. Keep careful note of the clock time of every reading. If you are lucky enough to get a burst of rain and a resulting rise and fall of the river, a great deal can be learned from even a single storm. Your observed hydrograph can be plotted and analyzed as discussed in the next chapter.

(I observe the staff gage in my yard by looking at it through a telescope. During a storm I step outside under my umbrella every 10 minutes to read my plastic rain gage.)

Individual observation sharpens awareness of the basic processes in nature that provide us with the necessities of life. Our personal measurements can add important details to the knowledge of our resource base.

CHAPTER THREE

Down the Channel System

The Hydrograph

Rivers drain water from the continents to the oceans and are the principal routes of transport for the products of weathering. Gravity provides the force by which both excess water and movable debris are brought from higher to lower elevations.

In accomplishing this transfer, the water that flows off the land toward the oceans forms and maintains a highly organized system of physical and hydraulic features. So complex are the interrelations that to focus on any single portion tends to make one lose sight of other equally important features. In the natural environment, the interrelationships make it difficult to visualize all of the system simultaneously. Yet it is precisely these interrelationships that constitute the most distinctive and pervasive characteristic of natural systems including rivers.

One reason why the interrelationships are difficult to visualize in rivers is that large variations occur through short periods of time. A storm that lasts a few hours or a few days produces runoff that appears in the river channel and subsequently drains away. The water level rises as the storm flow arrives at a given point and falls as it passes downstream. This rise and fall may be graphed as a function of time. Each location in a river system displays characteristic features in such a graph.

Discharge is defined as volume per unit of time. It is usually expressed as cubic meters per second (cms) or cubic feet per second (cfs). A plot of stream discharge as a function of time is, by definition, a hydrograph. An important aspect of hydrology involves the analysis of hydrographs to derive quantitative characteristics of the basin and its channels. A given basin will characteristically produce nearly the same hydrograph from different storms of equal magnitude and distribution. Hydrograph analysis, the use of the unit hydrograph, and analysis of rainfall-runoff relations are detailed in standard texts on hydrology. Of concern in this volume are the principles important to the fluvial geomorphologist, the student of rivers.

The morphology of channels involves more than rainfall-runoff relations. But an understanding of those relations must begin with the hydrology of a basin and the runoff in its channels. An example of such interrelations is presented to illustrate several important characteristics of a runoff hydrograph. In Figure 3.1 is a plot of rainfall and the resulting hydrograph measured at the U.S. Department of Agriculture Walnut Gulch Experiment Station, Tombstone, Arizona, on August 18, 1961. The basin is called W-3, its drainage area is 8.99 square kilometers (3.47 square miles), and its mean gradient from headwater to gaging station is 0.019. The vegetational cover is grass-woodland.

A graph of the rate of rainfall as a function of time is called the hyetograph. Note that the scale for plotting rainfall intensity is different than that used for runoff rate because the runoff rate is always smaller than the rate of the rainfall causing the runoff. It is smaller because of the infiltration that takes up some of the precipitation and because of the storage during the process of runoff generation.

In this example the maximum rainfall rate was 3.3 inches per hour (84 mm/hr), whereas the maximum runoff rate was 0.31 inch per hour (7.9 mm/hr).

The runoff began some time after the beginning of rainfall, 18 minutes in this case. Local depressions in the ground must be filled before overland flow can begin (depression storage). There is also a delay known as lag between the center of mass of rainfall and the center of mass of runoff. The former occurred at about 9:45 and the latter at 10:28, a lag of 43 minutes. The lag time is a hydrologic quantity of great significance.

Early in the history of hydrograph analysis it was recognized that the time distribution of accumulated percentage of runoff volume, called the distribution graph, was a useful way to compare hydrographs. W. B. Langbein found that if the time scale of the distribution graph was expressed in terms of lag rather than hours, the resulting dimensionless graph fitted most hydrographs for large as well as for small basins. His graph is shown in Figure 3.2.

The graph is presented in two forms. The S-shaped line is the usual distribution graph showing the percentage of runoff accumulated with passage of time. Time in this graph is expressed not in hours but in lags—that is, the time period between the center of mass of the rainfall

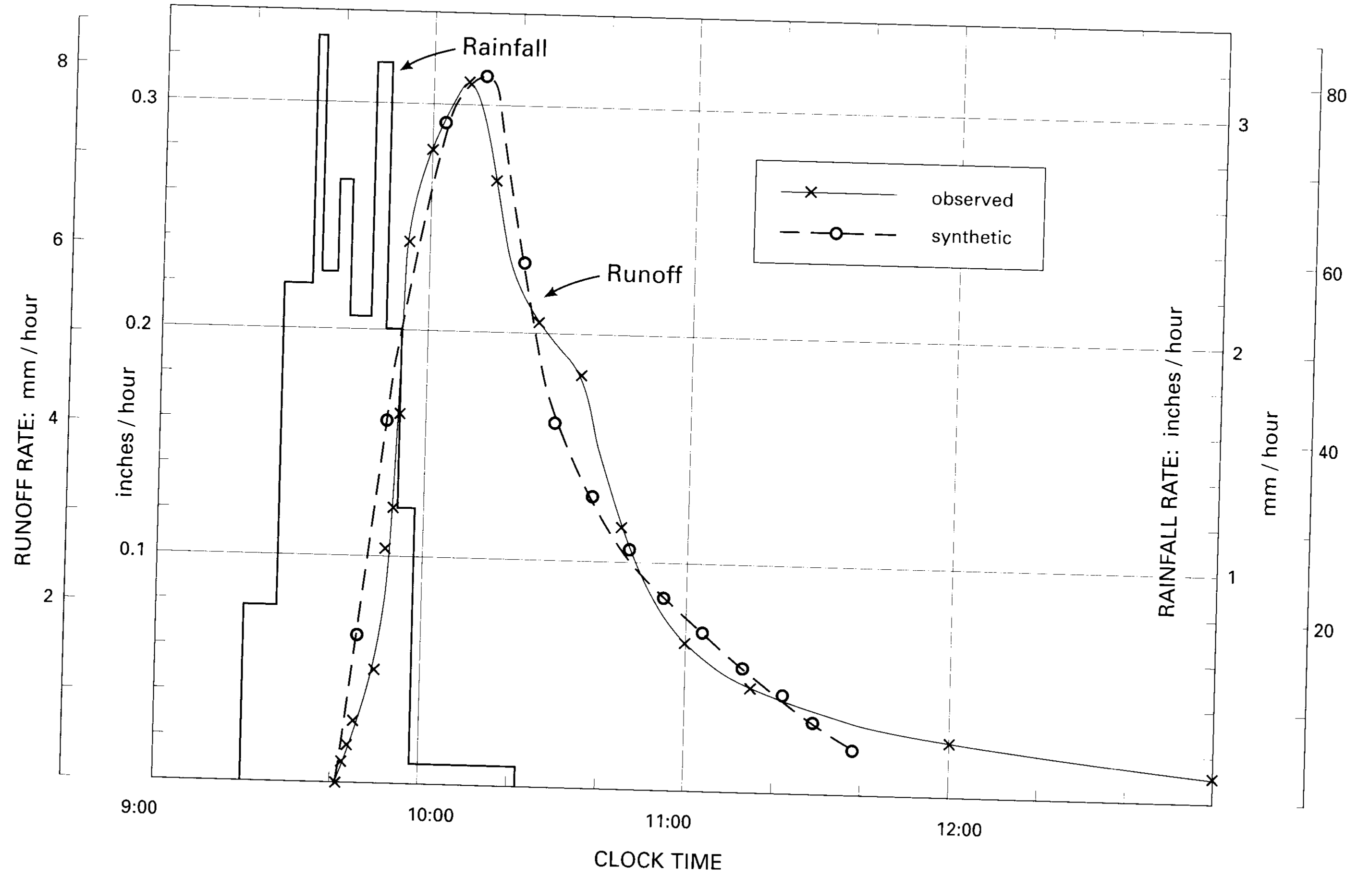


Figure 3.1 A hyetograph of rainfall (rain intensity as a function of time) and the resulting runoff hydrograph for a storm of August 18, 1961, at the W-3 basin of Walnut Gulch Experiment Station, Tombstone, Arizona. (Data from U.S. Department of Agriculture.)

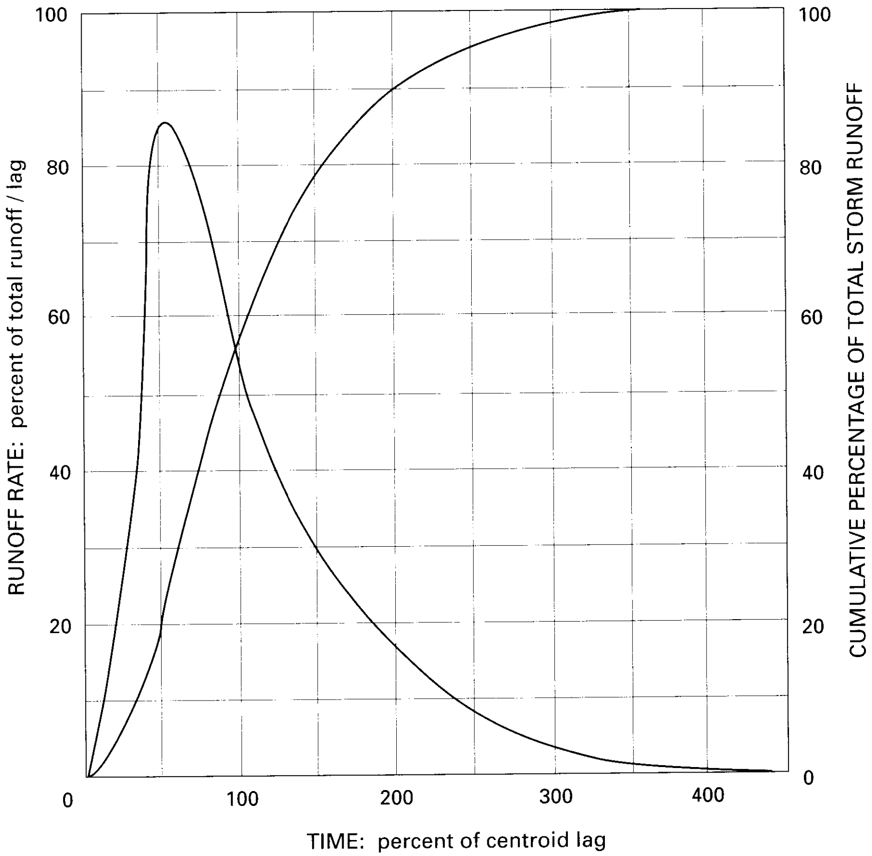


Figure 3.2 Summation graph and derived unit hydrograph. (From Langbein 1940.)

and the center of mass of the runoff. The peaked curve is the slope of the distribution graph and is in the form of a hydrograph. On that graph several important traits of hydrography are shown. Practically all of the water has run off after 3.5 times the lag; that is, nearly all runoff ends after a time equal to 3.5 lags. In the Tombstone example runoff began at 9:42 A.M. and after 3.5 lags or 150 minutes (at 12:12 P.M.) the runoff rate was small, although it did not cease until about 4.8 lags. Figure 3.2 shows that the peak rate of runoff occurs about 0.55 lag after the beginning of runoff. In the Tombstone example the peak occurred at 10:08, or about 26 minutes, 0.60 lag, after runoff began.

Note the way the Langbein hydrograph of Figure 3.2 is labeled. To construct a graph of the slope of the distribution curve, the slope must be expressed as percentage of runoff for some time period. A logical choice is one lag. Where the distribution curve is steepest, thus where the hydrograph peaks, the slope of the tangent to the distribution graph extended to a period of time equal to one lag gives a runoff rate of 87 percent of total runoff per lag.

The distribution curve can be used to compute a synthetic hydrograph that will agree closely with the observed hydrograph. The computation is described in Leopold 1991. The computed synthetic hydrograph using a lag of 43 minutes (0.72 hour) is shown as the dashed line in Figure 3.1.

Average Discharge

The average discharge is defined as that flow rate which, if continued every day of a year, would yield the observed annual volume of water. The average discharge usually fills a channel to about one-third the channel depth, and this flow rate is equaled or exceeded about 25 percent of the days in a year. That is, the river flows at a discharge less than average about 75 percent of the time. This figure varies among rivers between 60 and 75 percent.

Eight river channels in the upper Green River, Wyoming, were analyzed to determine the percentage of depth of channel filled when the discharge is equal to its mean annual value. The average was 43 percent. A study of 21 streams in the west central part of California showed that the mean annual discharge filled the channel to 0.28 of its bankfull depth on the average.

The discharges that are less than the average value contribute about 25 percent of the total yearly volume of runoff; those discharges less than half the average value contribute about 15 percent of the total volume. In most rivers these low values derive from the emergence of groundwater. Average discharge is, of course, important in analysis of water supply, but the average depends on the particular period of record over which the average was compiled. Serious errors in water management have been made by too heavy reliance on the average runoff value. A much more useful analysis is a frequency study of the whole array of available data so that the probability of various levels of excess or deficiency may be considered.

Because average values of discharge depend on the drainage area, rivers of different sizes are not comparable using average values. Comparison is facilitated by expressing the average as discharge per square mile of area or in inches of runoff. One inch of runoff from one square mile is 2.32 million cubic feet or 53 acre-feet. This would be produced by 0.074 cfs flowing continuously for one year.

A map of the United States showing average annual runoff may be found in the *National Atlas of the United States,* published by the U.S. Geological Survey in 1970.

Channel Storage

When you go outside to water the garden and turn on the faucet, water does not immediately come out at the far end of the hose. The hose must be filled with water before the far end begins to flow. The volume of water needed to fill the hose is called channel storage. By the same token, when you are finished and turn off the faucet, water flows out of the hose at a rapidly decreasing rate. The water in channel storage is draining out.

The river acts in a similar manner. Storm water inflow at some upstream point must at least partly fill the volume of the channel in order for outflow to feel the storm inflow. The time distribution of the discharge as measured at the downstream location changes.

Published works in engineering hydrology usually concentrate on dams, reservoirs, and other works on rivers large enough to be represented adequately in the river measurement network. These are basins having drainage areas of 25 square miles or larger. Environmentalists and even planners are often involved in the hydrology of creeks and small streams. They may be faced with assessing the impact of development on a basin of a square mile or less. For such small basins, river measurement stations are few; seldom are there two gaging stations in tandem along the length of such a small stream. Therefore hydrograph changes along the channel cannot be ascertained by analysis of changes between successive measuring points. Procedures applicable to such analyses are available in principle but not in specific example. In Chapter 2 a field procedure was outlined that would result in a hydrograph of discharge for a single storm event. Such a hydrograph developed by the most simple means—a stick for a staff gage, a measuring tape, orange peel for floats, and graph paper—can be just as satisfactory for many purposes as the W-3 basin of the USDA Experiment Station. Most interested observers who are not professional hydrologists are more likely to be dealing with hydrographs developed by simple means, not by instrumented basins such as W-3.

Once field observations are in hand, procedures explaining how the observed hydrograph of a small basin affects downstream locations are not available in the literature. The hydrograph will change shape and peak discharge as channel storage affects it in its downstream travel. A method of computing the effect follows.

Imagine water flowing from a faucet into a tub that has a drainpipe in the bottom. No appreciable outflow occurs down the drain until some water has accumulated in the tub to build up a depth over the drain orifice. Thus, in the early stages of putting water into the tub, the rate of inflow is greater than the rate of outflow. If you then turn off the faucet and the inflow ceases, the outflow down the drain continues until the water temporarily stored in the tub has drained out. During this time the outflow exceeds the inflow. Total outflow must equal total inflow. The relationships in any increment of time are expressed by the storage equation that states, "Outflow equals inflow plus or minus the rate of change of storage."

Think of a reach of channel as if it were a reservoir, or tub. The flow into the upper end of the reach is the inflow, the outflow is the water passing the downstream cross section, and the volume of the channel in between is the reservoir. The channel reach acts as if it were indeed a reservoir having the same relation of outflow rate to water level or storage volume, as if the long reach of channel were a lake just upstream of the outflow point. The storage volume in the channel thus acts like the bathtub storage. Channel storage in a river alters the outflow hydrograph, as would a reservoir.

The same kind of storage or reservoir action translated the hyetograph (precipitation versus time) into a hydrograph (stream flow versus time) that has a more rounded shape and smaller peak. Similar changes occur as the inflow hydrograph is compared with the outflow graph at the upper and lower ends of a reach of river.

In the left-hand portion of Figure 3.3, a storm hyetograph of precipitation as a function of time is represented by the tall rectangle. The precipitation started abruptly, lasted a short period of time at a constant intensity (inches per hour) and ended abruptly. The sketch of the discharge in an upstream channel reach as a function of time is the hydrograph (labeled "Discharge at 2"). It is drawn so that the area enclosed under the curve is smaller than the area under the precipitation rectangle. These areas represent volumes of water (flow rate multiplied by time equals volume). The difference in volume of precipitation and the consequent volume of runoff is what has been infiltrated into the soil during the time surface runoff collected and has become channel flow. Note also that the runoff hydrograph began slightly later than the beginning of the precipitation.

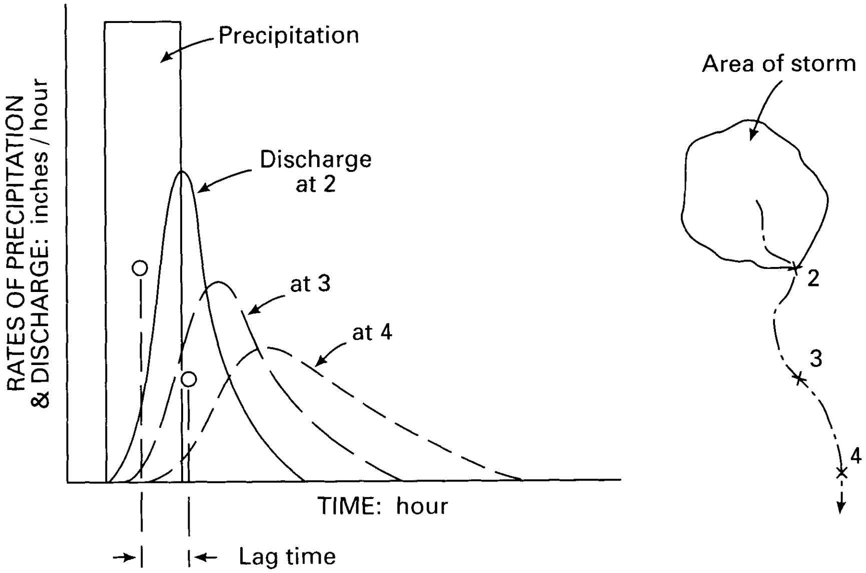


Figure 3.3 A diagram of the rainfall hyetograph (rectangular area) and the resulting hydrograph at location 2. The small open circles have been placed at the center of the mass of rainfall and runoff; the time difference between these circles is the centroid lag. The storm rain fell on the circular basin above location 2, and the flood wave moved downchannel to 3 and 4. Hydrographs at locations 3 and 4 are shown.

Let the right-hand diagram represent a plan or map of the area under consideration. The balloon-shaped area received the precipitation. Location 2 is the point on the channel where hydrograph 2 was observed.

The figure shows that no rain fell in the zone between points 2 and 3, so all the discharge experienced at 3 flowed earlier past point 2. Because it takes time to flow that distance, the beginning of flow at 3 is later in time than at 2, and at point 4 it is even later.

The first part of the hydrograph (discharge is rising) is called the rising limb, and that part after the peak is the recession limb. The point on the hydrograph where the curvature of the recession limb changes—that is, the point of inflection—indicates the time when inflow ceases. All of the recession limb of the hydrograph later than the point of inflection is water draining out from channel storage. Therefore, the volume of water in the channel for various values of outflow rate can be compiled by measuring the area under the recession limb at different times. For each time chosen, the outflow rate is shown by the ordinate value of the hydrograph.

The point of inflection of the recession limb marks the transition from that part of the hydrograph maintained by inflow into channels and that portion representing drainage from channel storage. This is true for small basins, but may not be true for large ones. The reason is that large basins are likely to have appreciable inflow from groundwater, which alters the base flow. In addition, numerous tributaries may enter a long reach, complicating the form of the hydrograph. Simple hydrographs such as the one in Figure 3.3 are caused by discrete storms having a sharply defined beginning and end. Storms over large areas usually have complicated patterns of time and area, resulting in complex hydrographs.

Flood Routing

The process of determining the timing and shape of a flood wave at downstream points is called flood routing.

In natural channel networks, a storm produces a runoff hydrograph in each small basin. At the junction of tributaries these respective hydrographs augment one another, and thus the flow below the junction is the sum of the rates of flow of the tributaries at the junction point. One such tributary may be on the rising stage while the joining one may be at peak or on the recession limb. The magnitude of a flood downstream depends strongly on the coincidence or lack of coincidence of the contributing basins.

Along the length of a drainage system, channel size and thus channel storage increase downstream. The flood peak from a given storm decreases downstream through the action of channel storage. The strongest tendency for overflow or for flood conditions is immediately downstream of that part of the basin experiencing the greatest runoff. Therefore, flooding of out-of-channel banks usually occurs just downstream from the storm location. The size, distribution, duration, and placement of a storm within a basin materially affect the location and distance along the major valley where overflow and the resulting flood damage will occur.

In a small headwater tributary a severe flood may peak within a few tens of minutes of the time of heaviest rain, and the peak may last only a few minutes before receding. In great rivers such as the Mississippi, the flow takes weeks to build up to the peak and high water may last a month or more.

For a simple problem in which only one hydrograph is available rather than two hydrographs from separate gaging stations, it can be assumed with little probability of error that channel storage acts as a reservoir. Therefore we can use a reservoir storage analysis.

Channel routing involves successive solutions to the storage equation:

outflow = inflow ± rate of change of storage

The computation illustrated follows the procedure outlined in Dunne and Leopold (1978, p. 353). The equation to be solved at each chosen time period is

$$\frac{S_2}{\Delta t} + \frac{O_2}{2} = \frac{S_1}{\Delta t} - \frac{O_1}{2} + \frac{I_1 + I_2}{2}$$

where S is storage volume, I is rate of inflow, O is rate of outflow, subscripts are 1 at beginning of period and 2 at end of period, and Δt is the duration of each step in the computation.

The storage characteristics will be estimated from the recession limb of an observed hydrograph. The problem posed is to compute the hydrograph at points downstream from the measuring station on basin W-3, introduced earlier.

Begin by reading the simultaneous values of discharge and time for the whole hydrograph plotted in Figure 3.1. In Table 3.1 the time and observed discharge are shown in columns 1 and 2, read at intervals of 5 minutes. The 5-minute period was chosen to give a reasonable number of points to represent the full hydrograph.

Column 2 is observed discharge in inches per hour, but because it is a reading for 5 minutes the value is also a volume of water equal to inches per hour for 5 minutes. In inches it is equal to the value in column 2 divided by 12, for there are 12 units of 5 minutes in an hour. We wish to accumulate the volume of runoff in the recession limb.

In the period later than noon, the tail of the hydrograph constituted 0.015 inch. So by accumulating the volumes in column 2 in inches and adding 0.015 inch, column 3 is the accumulated volume in inches for 5-minute intervals backward from noon. For example, at 12:00 the runoff rate was 0.022 inch per hour which in 5 minutes is a volume of 0.022 divided by 12, or 0.002 plus the amount 0.015 that was in storage after 12:00, giving 0.017 inch. At 11:45 the rate was 0.025 inch per hour or 0.002 inch, which added to the previous accumulation of 0.021 inch gives 0.023 inch.

Now compute the values of the outflow relations. Choose a time (distance) for the routing that is the length of channel down to the outflow

Table 3.1 Discharge and volume of observed runoff at Tombstone W-3 on August 17, 1968

1 Time (A.M.)	2 Observed discharge (in/hr)	3 Accumulated storage (in)
9:40	0	
9:45	0.025	
9:50	.050	
9:55	.190	
10:00	.280	
10:05	.300	
10:10	.310	
10:15	.275	
10:20	.245	0.158
10:25	.205	.137
10:30	.195	.120
10:35	.185	.104
10:40	.150	.089
10:45	.117	.076
10:50	.098	.066
10:55	.082	.058
11:00	.065	.051
11:05	.049	.046
11:10	.041	.042
11:15	.038	.038
11:20	.034	.035
11:25	.032	.033
11:30	.030	.030
11:35	.028	.027
11:40	.026	.025
11:45	.025	.023
11:50	.024	.021
11:55	.023	.019
12:00	.022	.017

Table 3.2 Storage outflow relations at Tombstone W-3 for time period 20 minutes (= 0.33 hr) on August 17, 1968

1 Outflow O (in/hr)	2 S (in)	3 $\frac{S}{\Delta t}$ (in/hr)	4 $\frac{S}{\Delta t} - \frac{O}{2}$ (in/hr)	5 $\frac{S}{\Delta t} + \frac{O}{2}$ (in/hr)
0.245	0.158	0.527	0.404	0.772
.205	.137	.457	.355	.560
.195	.120	.400	.303	.498
.185	.104	.347	.255	.440
.150	.089	.297	.222	.372
.117	.076	.253	.195	.312
.098	.066	.220	.171	.269
.082	.058			
.065	.051	.170	.138	.203
.049	.046			
.041	.042	.140	.120	.161
.038	.038			
.034	.035	.117	.102	.151
.032	.033			
.030	.030	.100	.085	.115
.028	.027			
.026	.023	.077	.064	.090
.025	.023			
.024	.021	.070	.058	.082
.023	.019			
.022	.017	.057	.046	.068

station for which the computation is being made. The routing envisions the hydrograph downstream by 20 minutes from the assumed inflow point. To estimate the distance along the channel, we can use the approximation that the flood wave proceeds downstream at about half the speed of the water particles.

In this example 20 minutes was chosen, or 0.33 hour. Table 3.2 repeats in columns 1 and 2 the flow rate and associated storage from columns 2 and 3 of Table 3.1. Then the third column is the storage from column 2 divided by the routing time, 0.33 hour. The units are inches per hour. Column 4 shows the result of subtracting half the outflow rate in column 1 from the storage rate in column 3, as indicated in the storage equation. Column 5 is similar, except that half the flow rate is added to the storage rate.

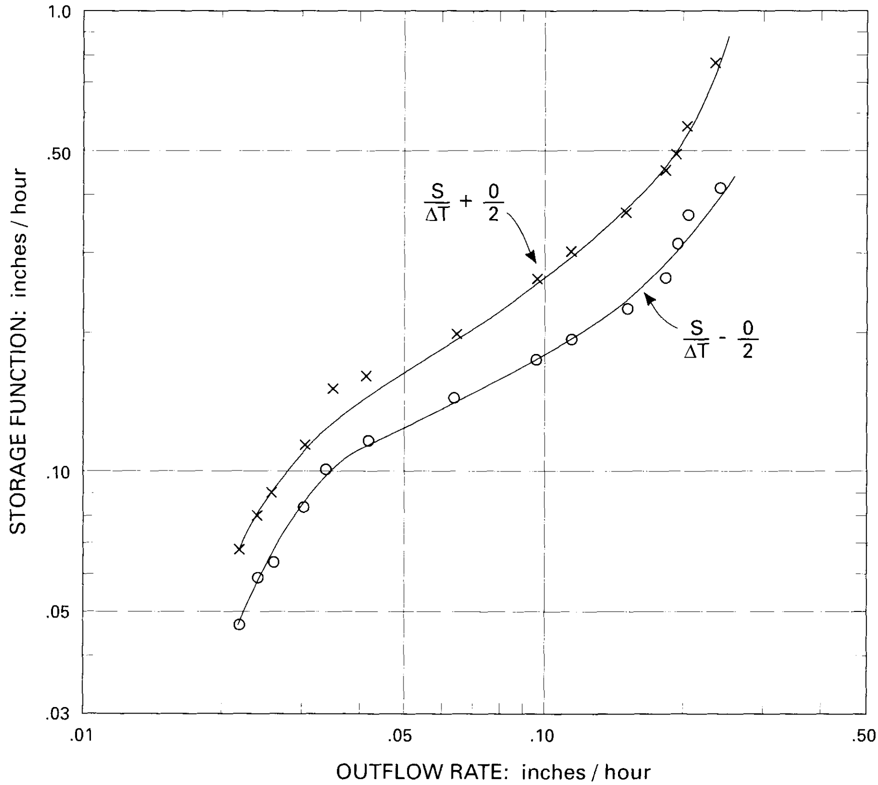


Figure 3.4 Storage function plotted against outflow rate, computed from the recession graph of the storm of August 18, 1961, on basin W-3.

The outflow rate of column 1 is now plotted against the two storage factors of columns 4 and 5, and the curves are shown in Figure 3.4.

The routing is carried out in Table 3.3. The hydrograph data observed are in columns 1 and 2. Column 3 shows the inflow occurring between the beginning of the time period and the end, a period of 20 minutes in this example. For convenience the computation is made at 10-minute intervals, a matter of choice. The first value of the end of the period is chosen as 9:50, but 20 minutes earlier the inflow was zero. So the value in column 3 at the time 9:50 is half of 0.050. This small inflow goes to storage, as will be seen. At 10:00 the average inflow is the average between 0.280 and zero, or 0.140, which entered in the abscissa of Figure 3.4 gives a value on the ordinate of $(S/\Delta t - O/2)$, or 0.220. Add this number to the inflow of 0.025 and get 0.360 which, entered into the ordinate of the other curve of Figure 3.4, on the abscissa gives a value of

Table 3.3 Routing increments of 20 minutes (0.33 hr) at Tombstone W-3 on August 17, 1968

1	2	3	4	5	6	7	8	9	10
		Routed 20 minutes				Routed 40 minutes			
Time (A.M.)	Inflow in/hr	Average inflow $\frac{I_1 + I_2}{2}$	At begin $\frac{S}{\Delta t} - \frac{O}{2}$	At end $\frac{S}{\Delta t} + \frac{O}{2}$	O (in/hr)	Average inflow $\frac{I_1 + I_2}{2}$	At begin $\frac{S}{\Delta t} - \frac{O}{2}$	At end $\frac{S}{\Delta t} + \frac{O}{2}$	O (in/hr)
9:40	0								
9:45	.025								
9:50	.050	0.025	0.060	0.060	0.085	0.012	0.015	0.027	0.025
9:55	.190								
10:00	.280	.140	.220	.360	.150	.075	.145	.220	.075
10:05	.300								
10:10	.310	.180	.275	.455	.180	.115	.190	.305	.118
10:15	.275								
10:20	.245	.263	.420	.683	.240	.195	.300	.495	.195
10:25	.205								
10:30	.195	.253	.400	.653	.230	.205	.315	.520	.200
10:35	.185								
10:40	.150	.198	.300	.498	.195	.218	.330	.548	.205
10:45	.117								
10:50	.098	.147	.225	.372	.150	.190	.290	.480	.190
10:55	.082								
11:00	.065	.108	.185	.293	.110	.153	.235	.388	.160
11:05	.049								
11:10	.041	.070	.140	.210	.072	.111	.180	.281	.105
11:15	.038								
11:20	.034	.050	.123	.173	.054	.082	.155	.237	.086
11:25	.032								
11:30	.030	.036	.103	.139	.037	.055	.128	.183	.058
11:35	.028								
11:40	.026	.030	.083	.113	.030	.042	.115	.157	.045
11:45	.025								
11:50	.024	.027	.060	.087	.026	.032	.088	.120	.032
11:55	.023								
12:00	.022	.024	.056	.080	.024	.027	.068	.095	.027

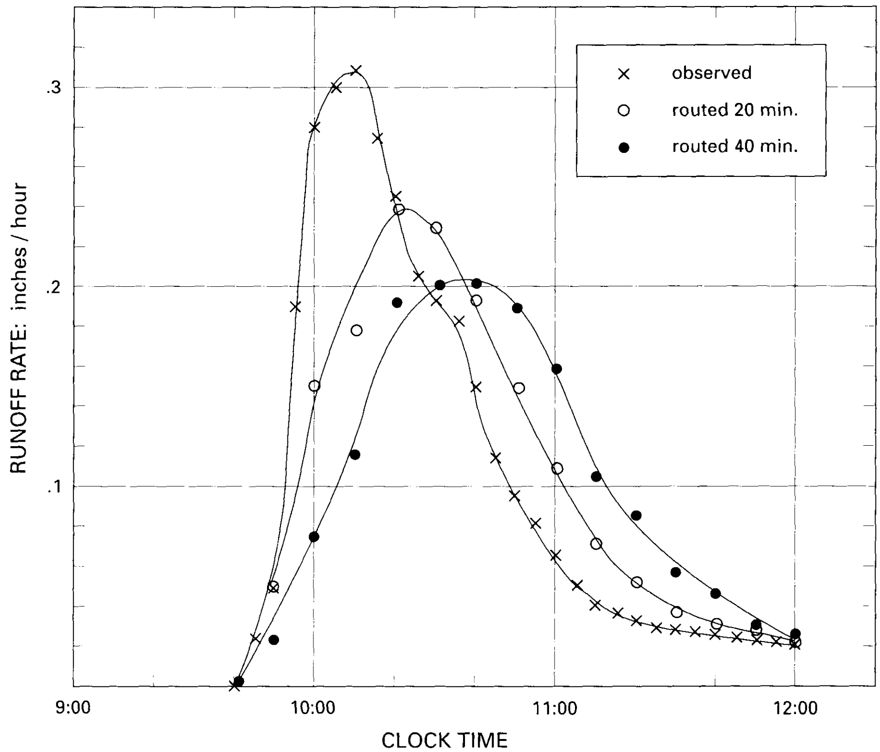


Figure 3.5 Routing the observed storm on the W-3 basin, shown in Figure 3.1. The routing shows two hydrographs 20 minutes apart, which represent the relation of discharge to time 0.4 and 0.8 mile downstream from the original point of measurement.

outflow of 0.150, to be plotted as the outflow of the routed hydrograph at 10:00 on Figure 3.5.

At the next time 10 minutes later, at 10:10, the average inflow between 9:50 and 10:10 is (0.050 + 0.310) ÷ 2 = 0.180. This value entered on the abscissa as flow rate in Figure 3.4 gives a storage function of 0.275, which added to 0.180 gives 0.455. At the ordinate of 0.455 of storage function read an outflow rate of 0.180, which is plotted in Figure 3.5 at time 10:10.

At the next time 10 minutes later, 10:20, the average inflow is (0.245 + 0.280) ÷ 2, or 0.263. Enter this value on the abscissa of Figure 3.4 and read the value of $(S/\Delta t - O/2)$ as 0.420, which added to 0.263 gives a storage function of 0.683. Enter that value in the ordinate of Figure 3.4 and read on $(S/\Delta t + O/2)$ an abscissa value of 0.240, which is plotted on Figure 3.5 at time 10:20. It can be seen that the values of time in column 1 are plotted against the outflow rates in column 6 to produce the outflow hydrograph of Figure 3.5.

To route for the next downstream point, later by 20 minutes, values of outflow of column 6 are now inflow for the new hydrograph. At the end of the computation, time in column 1 is plotted against outflow in column 10.

The routing envisions the hydrograph that would be observed at a location downstream at a distance of 20 minutes from the assumed inflow point. Using the approximation that the flood wave proceeds downstream at about half the speed of the water particles, and assuming that the water velocity in a storm flow is 3 to 4 feet per second, the flood wave may proceed at about 1.75 feet per second, and in 20 minutes has moved $1.75 \times 60 \times 20 = 2{,}100$ feet or 0.40 mile in the 20 minutes.

This procedure allows a quantitative estimate of how a flood wave moves downstream, being continually ameliorated by the reservoir action of channel storage. Visualization of this process of hydrograph alteration as it passes down the channel is essential to an understanding of channel dynamics. A river channel maintains its form even though it is constantly subjected to variation in discharge as weather goes from dry to wet and back to dry. The form that it takes and the processes of adjustment within the channel are the focus of the science of fluvial geomorphology.