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 riff, 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 uncultivated 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

<table>
<thead>
<tr>
<th>Location</th>
<th>Distance to closest divide (ft)</th>
<th>Drainage area (sq mi)</th>
<th>Mean slope from sill head to divide</th>
</tr>
</thead>
<tbody>
<tr>
<td>Contra Costa and Marin counties, California</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Briones No. 1</td>
<td>400</td>
<td>.000074</td>
<td>.20</td>
</tr>
<tr>
<td>Olema No. 1</td>
<td>65</td>
<td>.00003</td>
<td>1.07</td>
</tr>
<tr>
<td>Olema No. 2</td>
<td>50</td>
<td>.00003</td>
<td>.27</td>
</tr>
<tr>
<td>Olema No. 3</td>
<td>90</td>
<td>.00003</td>
<td>.27</td>
</tr>
<tr>
<td>Sublette County, Wyoming</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cora Hill A1</td>
<td>70</td>
<td>.00017</td>
<td>.17</td>
</tr>
<tr>
<td>Cora Hill A2</td>
<td>115</td>
<td>.00030</td>
<td>.16</td>
</tr>
<tr>
<td>Cora Hill A3</td>
<td>200</td>
<td>.00045</td>
<td>.12</td>
</tr>
<tr>
<td>Cora Hill A4</td>
<td>275</td>
<td>.00056</td>
<td>.13</td>
</tr>
<tr>
<td>Cora Hill A5</td>
<td>365</td>
<td>.00064</td>
<td>.08</td>
</tr>
<tr>
<td>Cora Hill B1</td>
<td>200</td>
<td>.00012</td>
<td>.09</td>
</tr>
<tr>
<td>Cora Hill C1</td>
<td>200</td>
<td>.00018</td>
<td>.11</td>
</tr>
<tr>
<td>Arroyo del los Frijoles, Santa Fe, New Mexico</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Caliente Arroyo</td>
<td>N.A.</td>
<td>.00006</td>
<td>.14 (approx)</td>
</tr>
<tr>
<td>Big Sweat</td>
<td>30</td>
<td>.0057</td>
<td>.045</td>
</tr>
<tr>
<td>Big Sweat</td>
<td>200</td>
<td>.0057</td>
<td>.045</td>
</tr>
<tr>
<td>Big Sweat</td>
<td>190</td>
<td>.0057</td>
<td>.045</td>
</tr>
<tr>
<td>Big Sweat</td>
<td>60</td>
<td>.0057</td>
<td>.045</td>
</tr>
<tr>
<td>Little Sweat</td>
<td>220</td>
<td>.0076</td>
<td>.045</td>
</tr>
</tbody>
</table>

pitojuniper woodland in New Mexico, and grass areas in west central Wyoming. They are shown in Table 1.1. In central Kenya, however, unripped 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.
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 1.2). The concave bank at the left is eroded. Seneca Creek, Maryland.

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.
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 ninetenth, 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.

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.

Figure 1.4 Data obtained from successive surveys of Watte 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.)
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 gullying 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 downslope. 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
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.

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.

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
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
Table 1.2 Alluvial chronology of western valleys, United States

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

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 box-like, 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-
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 subtypes that describe the size or coarseness of the bed material. Subclass 1 is bedrock, 2 is boulder, 3 is cobbles, 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,
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 flat 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 downstream 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 Rood, the difference in appearance of the
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 0.44 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 medium 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 17.2 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 alternations 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 layer, 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...
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 gravely 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 scour 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.