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History of the Grand Canyon and of the Colorado River in Arizona

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INTRODUCTION

Many desert rivers, including the Nile River of Africa, the Tigris and Euphrates rivers of Asia, and the Colorado River of the western United States, obtain their waters from mountain ranges far removed from the warm and parched lands that border most of their courses. This combination of abundant water and warm climate has made parts of these rivers lifelines, allowing the cultivation of large areas that otherwise are desert.

The Tigris, Euphrates, and Nile led to the establishment of ancient civilizations such as those of Mesopotamia and Egypt—and probably to the development of settled urban living as we know it today. It is no coincidence that sites such as Ur, Babylon, Thebes, and Memphis were sited on the banks of these rivers.

The Colorado River cannot claim to be the cradle of civilization; its virtue, instead, is to support modern cities—such as Los Angeles, Las Vegas, and Phoenix—that could not flourish in its absence. Many of these cities are far removed from the Colorado, whose waters are brought to them by a system of artificial impoundments and aqueducts. The Colorado, therefore, has enabled us to carry out a great experiment, based on contemporary technology, in settling areas that are in themselves inhospitable.

Great demands are made on the waters of the Colorado River, yet the amount of available water is finite. This has resulted in a host of social, political, and engineering problems that have centered chiefly on how many reservoirs should be built and who should get the water. Many of these problems are fertile areas of endeavor for geologists concerned with the practical application of their science.

For other geologists, however, the river has been the source of quite different interests and controversies. These interests are theoretical in nature and have to do with the general problem of how rivers are born and develop. When and how did the Colorado River come into being? How do rivers like the Colorado evolve? When did canyon cutting and correlative uplift occur? How and why has the Colorado cut across the many belts of high ground astride its course? How quickly was the Grand Canyon cut? Might the answers to these questions give us an insight into how rivers in general establish their courses?

This chapter considers the history of the Colorado River and its Grand Canyon (Fig. 15.1). To understand what follows, it is important to remember two contrasting views regarding the history of the Colorado River. The first is that the river from birth has been part of an integrated drainage system with a course approximating the present one.

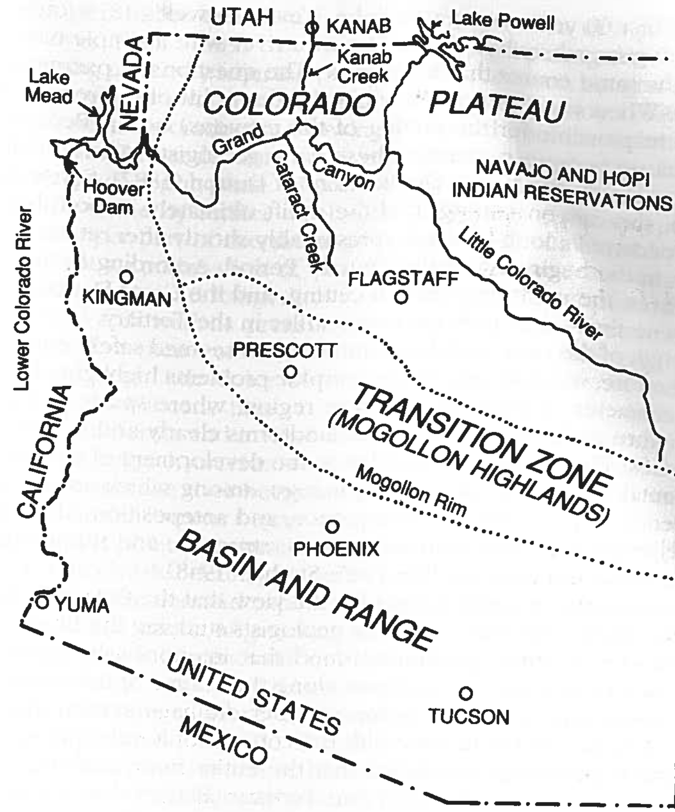

Figure 15.1. Location map, Arizona.

According to the first view, the river evolved pretty much as it is today and at some well-defined time, such as the Eocene. A statement that is made about any part of the river applies to the river as a whole; the entire river is young or old, as the case may be.

The second view is that most rivers are continually changing entities that have evolved from various ancestors and will continue evolving into progeny whose configuration depends on factors such as tectonism and climate. According to this view, the answer to the question "When was the Colorado River born?" can only be another question: "How much departure from the present configuration is one willing to tolerate and still speak of the Colorado?"

It is important to remember that the Colorado traverses two contrasting terrains in Arizona. The first is the canyon country, typified by the Grand Canyon. This is highly dissected terrain, commonly with substantial topographic relief. The second is the plateau country, which is typified by most of the Navajo and Hopi reservations (Fig. 15.1). This landscape is characterized by low relief, wide mature valleys, and scarps that develop on beds of contrasting resistance and retreat down structural slopes. The plateau country is older and more widespread than the canyon county, which is encroaching on it.

For the first 60 years or so after John Wesley Powell's 1875 journey of discovery, geologists subscribed to the idea of a river with a simple history; it was born with the same course that it has now. The questions of paramount importance were: When was it born and when did the uplift of the region (which as considered responsible for the cutting of the canyons) occur? Because erosion of the plateau country is pervasive, these early geologists inferred that the erosion was also deep—the "Great Denudation" of Dutton (1882). Consequently the denudation, the canyon cutting, and the uplift ultimately responsible for both must have occurred a long time ago, presumably shortly after retreat of the great inland seas at the beginning of the Tertiary Period. According to this view, the Colorado River, the uplift, the canyon cutting, and the Great Denudation all began in Eocene time—and perhaps even earlier in the Tertiary.

The origin of the river and the Grand Canyon seemed safely established. Attention, therefore, was focused on geomorphic problems highlighted by the textbook-like character of the Grand Canyon region, where sparse vegetation and simple structure make it possible to see landforms clearly and to trace them for great distances. These characteristics led to the development of several concepts of fundamental importance in geomorphology, among which are the principles of antecedence, superposition, consequence, and anteposition, all having to do with the relations between drainage systems, structure, and topography (Davis 1901, 1903; Babenroth and Strahler 1945; Strahler 1948).

Storm warnings signaling danger for the view that the Colorado River is old were hoisted in the 1930s and 1940s by geologists studying the Basin and Range country (Fig. 15.1). These geologists found that interior-basin deposits of late Miocene and Pliocene age are common along the course of the Colorado River. They also could find no evidence for an older drainage system that could be called the Colorado. In conformity with the concept of a monophase history for the river, these geologists concluded that the entire river, and thus the Grand Canyon as well, was no older than late Tertiary (Blackwelder 1934; Longwell 1936, 1946).

The next development occurred in the plateau country of Arizona, Utah, and Colorado. Here, widespread evidence, eventually summarized by Hunt (1969), showed that drainage systems, locally departing from the present course of the Colorado River but arguably ancestral to it, existed certainly in the Miocene and very probably as early as the Oligocene. They might have existed even earlier, but if so, the evidence is gone. There was now a major paradox: The same river seemed to be at least as old as Miocene–Oligocene in its upper reaches, but no older than latest Miocene or Pliocene in its lower ones.

In an attempt to shed light on the paradox, E. D. McKee and the Museum of Northern Arizona sponsored studies on critical areas at and near the mouth of the Grand Canyon. Results by Lucchitta (1966) and Young (1966) showed no stratigraphic or morphologic evidence of a through-flowing drainage system during the deposition of Miocene interior-basin materials related to Basin–Range deformation. Nor could the lack of evidence for through-flowing drainage be bypassed by looking elsewhere along the course of the lower Colorado River or the southwest margin of the Colorado Plateau in Arizona (Fig. 15.1). In this area, interior-basin deposits are ubiquitous, and the deposits older than Basin–Range rifting indicate drainage northeastward, from what now is the Basin and Range Province onto what now is the Colorado Plateau.

The northeast drainage existed as recently as the emplacement of the Peach Springs Tuff, as 18-million-year-old ignimbrite that flowed onto the Colorado Plateau. Before rifting of the Basin–Range, therefore, drainage was not to the west or southwest (as would be required for a river with a course similar to that of the present Colorado) but in the opposite direction—to the northeast.

The next step in the conceptual journey was the idea of a polyphase history for the river. Hunt (1969) contributed to it by proposing drainage systems initially departing markedly from the present Colorado River, but gradually evolving into this configuration. However, Hunt, as well as Lovejoy (1980), still postulated an ancestral Colorado river flowing westward from the Colorado Plateau even before Basin–Range deformation, a concept not supported by the evidence.

The concept of a polyphase history for the Colorado river was developed fully for the first time by McKee et al. (1967). These authors accepted the antiquity of the upper part of the drainage system, as documented by Hunt, but could not accept a continuation of this drainage westward through the Grand Canyon into the Basin and Range Province. Instead, they proposed that the ancestral Colorado followed its present course as far as the eastern end of the Grand Canyon, but then continued southeastward (not westward) along the course of the present Little Colorado (Fig. 15.1) and Rio Grande rivers into the Gulf of Mexico. In Pliocene time, a youthful stream, emptying into the newly formed Gulf of California and invigorated by a steep gradient, eroded headward and captured the sluggish ancestral river somewhere in the eastern Grand Canyon area. It was then that the river became established in its present course, and the carving of the Grand Canyon began.

The concept is pivotal because it introduces the idea (even though the point is not made explicitly) that drainage systems evolve continually—and do so chiefly through headward erosion and capture and in response to tectonic movements. During this process, the configuration and course of a drainage system may change so much that it becomes difficult and rather arbitrary to continue calling the ancestral drainage system by its present name.

Evidence accumulated since 1967 argues against drainage southeastward along the Little Colorado and Rio Grand rivers, as proposed by McKee et al. (1967). On the other hand, evidence has continued to grow that an ancient river could not have flowed through the western Grand Canyon region (Fig. 15.2) into the nearby Basin and Range Province (Lucchitta 1972, 1975; Young 1970; Young and Brennan 1974).

Analysis of deposits along the course of the lower Colorado River in the Basin and Range Province has confirmed that this part of the river is no older than latest Miocene. It also has shown that the capture of the ancestral Colorado River is documented by the appearance within river deposits (Imperial Formation of Miocene and Pliocene age) in California's Salton trough of coccoliths otherwise found only in the Cretaceous Mancos Shale of the Colorado Plateau (Lucchitta 1972).

An attempt to synthesize current information led Lucchitta (1975, 1984) to postulate that the ancestral Colorado did not flow to the southeast along the valley of the Little Colorado River, as proposed by McKee et al. (1967). Instead, the river cross the Kaibab Plateau along the present course of the Grand Canyon, then continued northwestward along a strike valley in the area of the Kanab, Uinkaret, or Shivwits plateaus (Fig. 15.2) to an as yet unknown destination. After the opening of the Gulf of California, this ancestral drainage was captured west of the Kaibab Plateau by the lower Colorado drainage. According to this concept, the upper part of the Grand Canyon in the Kaibab Plateau part of the canyon in this area and in all of the western Grand Canyon postdates the capture and was carved in a few million years, a process aided by nearly 0.6 miles (0.9 km) of regional uplift since the inception of the lower river (Lucchitta 1979).

This hypothesis is based on (1) the occurrence of gravels of probable river origin in the area of the Kanab, Uinkaret, and Shivwits plateaus and (2) the observation that northwest-trending drainages along strike valleys were common and persistent before canyon cutting, as evidenced by fossil valleys preserved

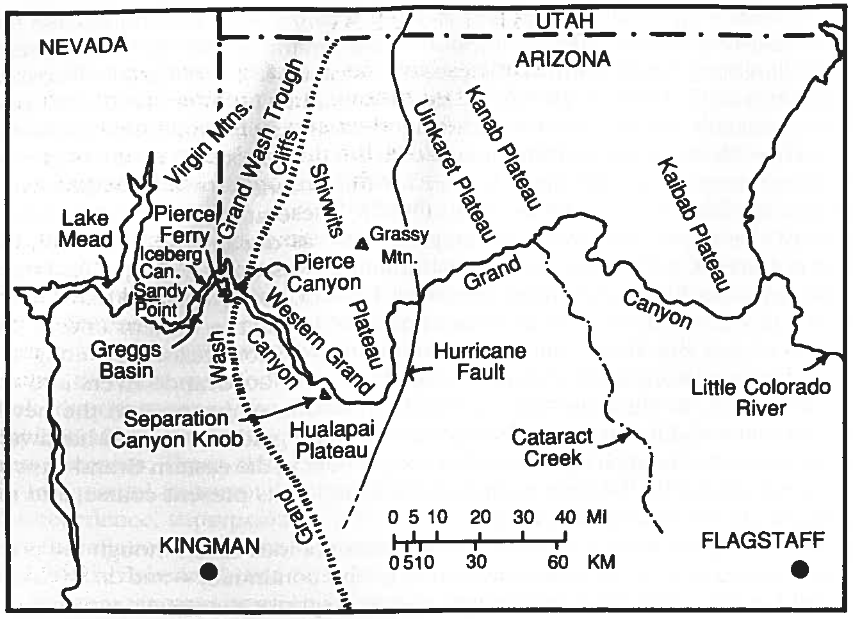

FIGURE 15.2. Map showing selected geologic and geographic features in northern Arizona.

under Miocene lavas in many places in the southwestern Colorado Plateau and by ancient valleys in the plateau country. Examples of such valleys are those of Cataract Creek (Fig. 15.2) and the Little Colorado River (Fig. 15.1), which predate canyon cutting and have not yet been affected appreciably by it.

The old problem of how the Colorado could have crossed the Kaibab Plateau can be analyzed by going backwards in time and restoring rocks removed from this upwarp in the past few million years (Fig. 15.3). This shows that a river such as the ancestral Colorado could have flowed readily across the Kaibab (then lower topographically than its surroundings) in an arcuate racetrack corresponding to the present configuration of the eastern Grand Canyon. The racetrack was localized by north-facing monoclinal flexures that cross the Kaibab and interrupt the general southward plunge of this dome.

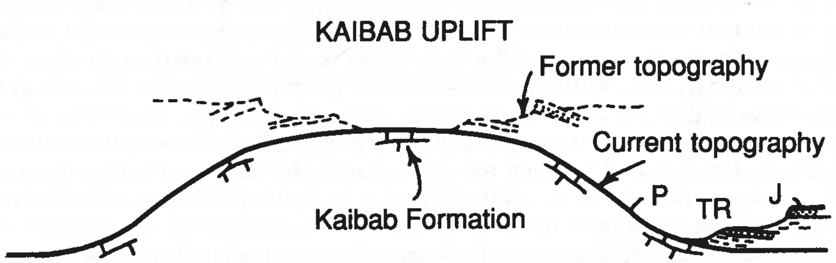

FIGURE 15.3. Cross section showing uplift of the Kaibab Plateau both a few million years ago and currently. P, Permian; TR, Triassic; J, Jurassic.

THE EVIDENCE

The Colorado River and its tributaries have been cutting down vigorously during the last several million years of their history. Such erosion—or at least nondeposition—may have been typical of this river system through most of its life. This characteristic creates great difficulties for the geologist intent on reconstructing the history of the river because such a history can only be pieced together from evidence left behind by the river.

The evidence is of two kinds: (1) river deposits such as gravel, sand, and silt and (2) landforms such as river valleys and canyons. Of the two, the deposits are by far the more useful because most can be attributed unequivocally to a specific river, on whose provenance, direction of flow, and age they provide valuable information. In contrast, a landform such as a valley can result from a river other than the one of interest or even from the action of an entirely different agent, such as a glacier.

Any river (even one that is cutting down) leaves behind deposits. For rivers that are cutting down, such as the Colorado, most of the deposits are removed soon after deposition. This means that deposits with which the geologist can work are few and scattered, especially in the Grand Canyon. Furthermore, the deposits preserved tend to reflect only the most recent part of the river's history.

Because of these factors, study of the Colorado River consists largely of a detective-like piecing together of circumstantial evidence, most of which is negative. In other words, the evidence is more likely to show that the Colorado River did not go through some area at a specified time than to document its existence in some specific place at a specific time. When circumstantial evidence is not negative, it typically attributes to an inferred Colorado River the properties known to be widespread at the time in question. For example, if at some time in the past most drainages followed northwest-trending strike valleys, one can reasonably infer that the Colorado also followed such a valley.

These points may seem too obvious to be worth repeating, but they need to be made once more because many people are taken aback by the lack of hard data pertaining to the history of the Colorado River. It is true that we do not have much direct evidence, given the problems mentioned above, but the circumstantial information is of many different kinds and from different places. Collectively taken, it enables us to construct a solid history that has a good chance of being correct, at least in its major aspects.

The history of the Colorado River and the Grand Canyon is best subdivided into three periods of tectonism that profoundly affected the drainage patterns of their time. These periods are:

1. *Pre-rifting.* The interval between the beginning and the middle of the Tertiary, at which time basin–range rifting got underway along the present course of the Colorado River, and the Colorado Plateau became distinct structurally and morphologically from the adjacent Basin and Range Province.
2. *Rifting.* The time of basin–range extension, with intensity tapering off toward the end of the interval—five to eight million years ago, at most. This was a time of widespread interior drainage in the Basin and Range Province.
3. *Post-rifting.* During this time—between five to eight million years ago and the present—rifting ceased, the Gulf of California opened, and through-flowing drainage became established.

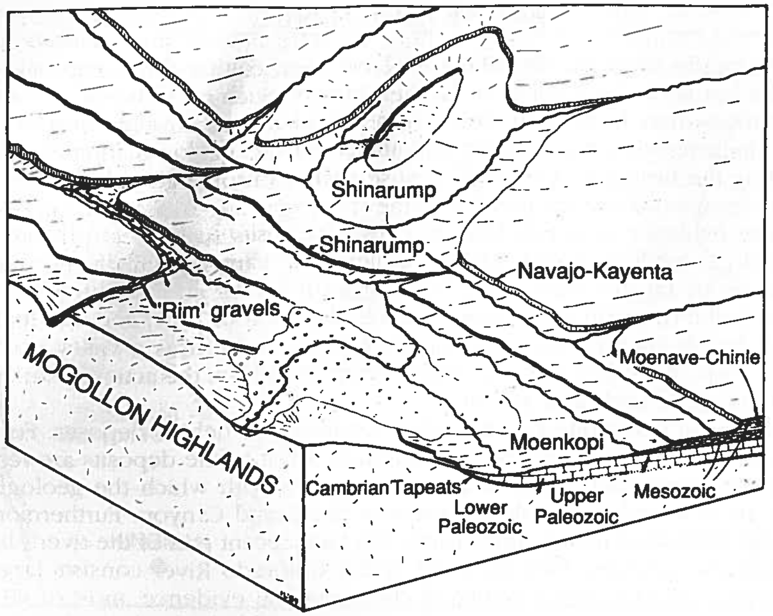

Figure 15.4. Block diagram showing schematically the relation of Mogollon Highlands to the Colorado Plateau before Miocene rifting. Diagram shows scarps composed of hard-over-soft couplets retreating down structural slope, the inferred topography on the Kaibab uplift, and the trellis drainage network. Looking about northwest, Hualapai Plateau would be at left side of diagram, the area of the Little Colorado River at the right side.

Pre-rifting

Before the onset of rifting in late Oligocene to Miocene time, the terrain south and southwest of the Colorado Plateau (Fig. 15.1) was higher than the nearby plateau, both topographically and structurally (Fig. 15.4). It remains high structurally today, even after the rifting. This belt of uplift, often referred to as the Mogollon Highlands, presumably was formed during the orogenic events at the end of the Mesozoic Era and the beginning of the Tertiary period. It caused the gentle northeastward tilting of strata near the southwest margin of the plateau, as witnessed by the widespread "rim gravels" (Finnell 1962, 1966; McKee and McKee 1972; Peirce 1984; Peirce et al. 1979; Peirce and Nations 1986). Along much of the rim, these gravels were deposited on rocks high in the Paleozoic section. They occur locally on Mesozoic rocks as well, yet they contain lasts of Precambrian igneous and metamorphic rocks that could have come only from the belt of uplift south of the rim, where such rocks were exposed.

Along the southwestern margin of the Colorado Plateau, in the area of the Hualapai Plateau, the rim gravels are present within ancient canyons trending northeastward, down the structural slope (Fig. 15.4). Clast provenance, imbrication, and gradient of the canyon floors all indicate derivation from the southwest (Young 1966, 1970). A similar drainage direction is indicated by the 18-million-year-old Peach Springs Tuff, an ignimbrite that flowed northeastward onto what now is the Colorado Plateau from a source area in the Basin and Range Province (Young 1966; Young and Brennan 1974; Glazner et al. 1986). In the

area southwest of the present plateau rim, erosion already had cut down to the Precambrian basement by Miocene time, and Phanerozoic strata were retreating northward from a structural and topographic high near Kingman (Lucchitta 1967, 1972).

At the southern margin of the Colorado Plateau, "rim gravels" are widely distributed (Fig. 15.4) along the Mogollon Rim (Fig. 15.1), which is the physiographic southern edge of the Colorado Plateau. Even though gravel-filled channels are not as prominent there as they are on the Hualapai Plateau, gravel-filled channels do occur, notably along the Little Colorado River. These gravels delineate the probable drainage pattern near the southern margin of the Colorado Plateau for much of Tertiary time (Lucchitta 1984).

For most of its course, the Little Colorado River flows in a mature and subdued valley that trends northwestward—parallel to the regional strike of Mesozoic and Paleozoic units (Figs. 15.1 and 15.5). The valley is at the erosional featheredge of the Triassic Moenkopi Formation on the Permian Kaibab Formation. The northeast side of the valley is in the Moenkopi, capped by the resistant Shinarump member of the Triassic Chinle formation, which dips gently to the northeast and is very resistant to erosion. Because of this resistance, the topographic surface is near the top of the Kaibab over wide areas of the Colorado Plateau.

The hard-over-soft couplet represented by the Shinarump over the Moenkopi is the lowest, stratigraphically, of several such couplets within the Mesozoic section (Fig. 15.6). The scarps formed by the couplets face southwest and with time

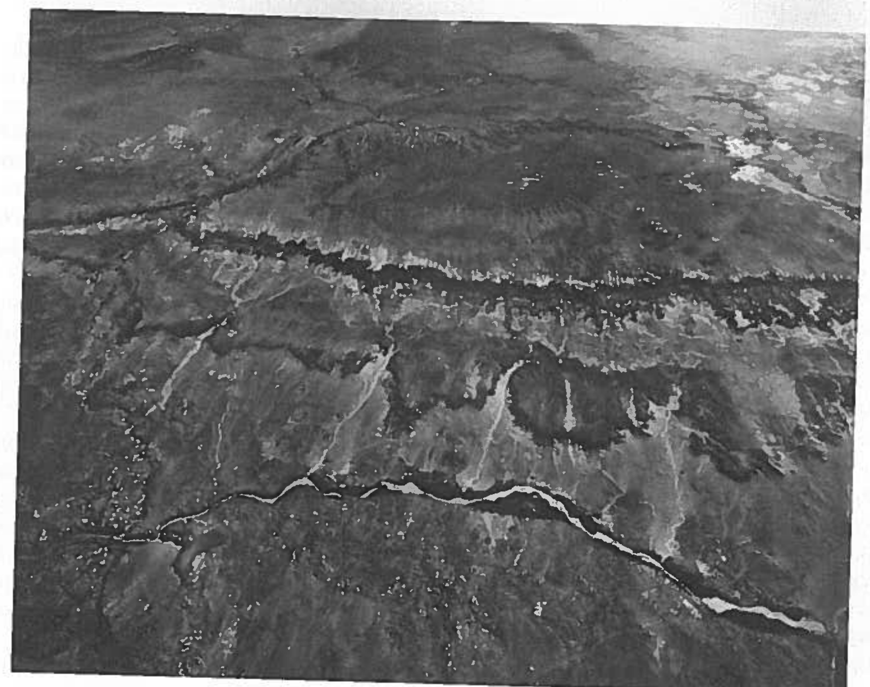

Figure 15.5. High-altitude (U-2) aerial photograph looking north to northeast. The Little Colorado River is in the foreground. Dark mass in middle distance is Black Mesa. Wide valley of the Little Colorado River and scarps forming its northeast side are clearly visible.

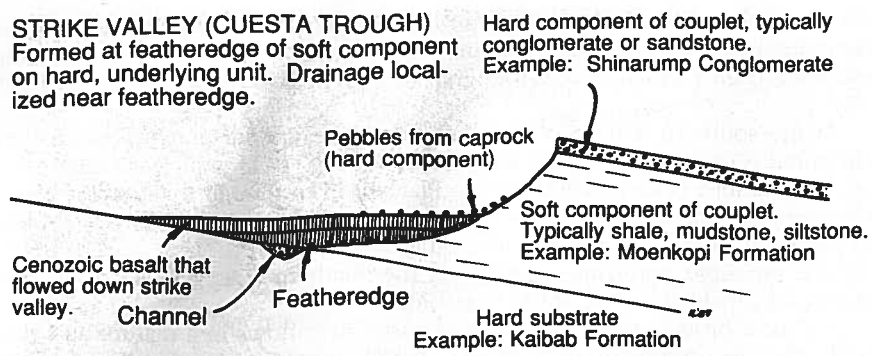

FIGURE 15.6. Diagrammatic representation of typical strike valley formed at foot of hard-over-soft scarp retreating (northeast) down structural slope. Also shown are the channel located near featheredge of slope-forming unit, basalt that in many places flowed down strike valleys, and debris derived from caprock. Looking northwest. Width of valley typically is a few kilometers.

migrate northeastward down the structural slope. The couplets that are stratigraphically highest and youngest have migrated the farthest to the northeast. Those that are lowest and oldest, on the other hand, are found to the south or southwest. These couplets are closest to the belt of uplift (Mogollon Highlands) from which they started. The Grand Staircase described by Dutton (1882) is composed of these couplets.

A northeastward migration can be documented for the valley of the Little Colorado River, where the ancient gravel-filled channels are cut into the top of the Kaibab Formation southwest of the present channel of the river. Both the northwest-trending valleys at the foot of the Shinarump–Moenkopi couplet and the northeastward migration of such valleys with time can be documented on the Shivwits Plateau (Figs. 15.2 and 15.6), where various basaltic lavas of late Miocene age have flowed down successive stands of the valleys (Lucchitta 1975).

Most tributaries of the Grand Canyon are short, steep, and immature (Fig. 15.7). The Little Colorado River, Cataract Creek, and Kanab Creek (Figs. 15.1, 7) have the length and appearance of conventional, mature river valleys. The Little Colorado and the Cataract Creek flow northwestward—parallel to the strike of beds; Kanab Creek also shows the influence of structure on its course by flowing south around the western flank of the Kaibab uplift, which has overprinted and modified the regional northwestern strike.

These three streams illustrate features characteristic of the pre-Grand Canyon drainage. One feature is the control of drainage by structure, which is represented by the gentle dip of beds modified locally by folds and faults. another is the marked effect of lithology, as represented by couplets within the Mesozoic section that differ in resistance to erosion and form strike valleys and scarps. Together, these features have given the region a trellis drainage pattern that consists of northwesterly segments parallel to strike and northeasterly segments trending down the structural slope.

The streams bringing rim gravels onto the Colorado Plateau and ancient drainages such as the Little Colorado must have been functional for a substantial period of time because there is little evidence on the southwestern Colorado Plateau of widespread and long-lasting Tertiary ponding in most pre-rifting time. The conclusion is that the drainages must have been tributary to a master stream that presumably also reflected the effects of structure and stratigraphy (Fig. 15.4).

FIGURE 15.7. High-altitude (U-2) photograph looking west, down much of Grand Canyon. Shivwits Plateau is visible in the distance on the right. The short, steep, and immature tributaries to the Grand Canyon are evident, in contrast to the well-developed Cataract Creek in the middle distance on the left. The confluence of Kanab Creek with the Colorado River is in the middle right of the picture.

The inference is that this was the ancestral Colorado River and that it flowed chiefly in northwest-trending strike valleys. The Colorado, however, did not flow through the western Grand Canyon region. Here, the presence of gravels and the Peach Springs Tuff shows instead that drainage near the edge of the Colorado Plateau was to the northeast.

Additional evidence that the river did not flow through the western Grand Canyon itself is provided by a knob on the south rim of the Grand Canyon near Separation Canyon (Fig. 15.2). The knob is composed of a mid-Miocene basalt flow that caps gravel and colluvium containing upper Paleozoic rocks (Young 1966; Lucchitta and Young 1986). In this area, such rocks crop out only on the Shivwits Plateau to the north. Today, the knob and the Shivwits Plateau are separated by the Grand Canyon, which could not have existed when the colluvium and basalts were emplaced (Young 1966).

Rifting

Mid-to-late Miocene basin–range rifting affected most of the western Cordillera and produced ranges separated by deep structural basins filled with thick sequences of deposits. The basins sank so rapidly that through-flowing drainage was not able to maintain itself, resulting in widespread interior drainage. In western Arizona, rifting began in mid-Miocene time and was essentially over by the end of that epoch.

The southwestern Colorado Plateau was much less affected by rifting than was the Basin and Range terrain to the west and south. Normal displacement occurred on many small faults and several larger ones, such as the Hurricane, but the intensity of faulting did not approach that in the Basin and Range. The most noteworthy effect of the rifting was the foundering of the formerly high areas south and southwest of the plateau. The foundering formed structural basins beyond the edge of the plateau and interrupted the old drainages that had deposited the rim gravels (Peirce et al. 1979; Young 1966; Young and Brennan 1974; Lucchitta 1967, 1979). Ponding and deposition of lake beds occurred locally on the plateau, presumably as a result of minor warping. The most conspicuous such lake beds are the Bidahochi Formation of latest Miocene and Pliocene age and the Willow Springs Formation of Pliocene to Pleistocene(?) age. Such deposits, however, are minor in comparison with the ubiquitous interior-basin deposits of the nearby Basin and Range Province.

Particularly interesting with respect to the history of the Colorado River is the Miocene Muddy Creek Formation, as exposed by the Grand Wash trough and the Pierce Ferry area (Figs. 15.2 and 15.8) at the mouth of the Grand Canyon. No evidence of a river emptying into the Grand Wash trough is present in the lithologies and facies distribution of the Muddy Creek, which, instead, records interior drainage with derivation of clastic material from nearby highlands. Analysis of directions of transport and topographic closure leads to the same conclusion.

West-draining canyons are present in the Grand Wash Cliffs, a prominent, west-facing fault scarp formed during basin-range deformation. The scarp marks the mouth of the Grand Canyon and defines the western edge of the Colorado Plateau (Figs. 15.2 and 15.8). The west-draining canyons are short and steep. Where they debouch into the Grand Wash trough, the washes that carved the canyons have deposited fans of locally derived material. One such fan emerges from Pierce Canyon, whose mouth is only 1.5 miles (2.5 km) north of the Grand Canyon (Figs. 15.2 and 15.8). The fan was deposited across the mouth of the present Grand Canyon. This could not have happened if the Grand Canyon and the Colorado River existed in their present location at the time.

The highest unit of the Muddy Creek Formation in the Grand Wash trough is the Hualapai Limestone, which interfingers downward with clastic rocks of the Muddy Creek. The Hualapai was deposited in a shallow lake, which extended about 25 miles (40 km) west from the Grand Wash Cliffs. As with other lithologies of the Muddy Creek, the Hualapai contains no evidence for a major river emptying into the lake in which the limestone precipitated. Instead, the lake waters were highly charged with calcium carbonate and other salts. Deposits younger than the Hualapai record through-flowing drainage.

The Hualapai was the youngest unit to be deposited in the area near the mouth of the Grand Canyon before the lower Colorado River was established. However, geologists have not yet dated it directly. The Muddy Creek Formation near Hoover Dam contains basalts that are five to six million years old (Anderson 1978; Damon et al. 1978). A tuff about 1600 feet (500 m) below the Hualapai Limestone in the Pierce Ferry area has been dated by fission-track methods at eight million years (Bohannon 1984). Another tuff low within the limestone in an area about 25 miles (40 km) west of the mouth of the Grand Canyon also has yielded an age of about eight million years (Blair 1978). South of Hoover

Figure 15.8. High-altitude (U-2) photograph looking about north from near mouth of the Grand Canyon (right foreground). Lake Mead is in foreground. North-trending valley in center of picture and west of the prominent Grand Wash cliffs is the Grand Wash Trough. Iceberg Canyon is the narrow, north-trending arm of the lake in the lower left. Pierce Ferry is at wide area of the lake in the lower center of the picture.

Dam (Fig. 15.1), the Bouse Formation (an estuarine deposit associated with the opening of the Gulf of California) records the presence of the Colorado River in its present lower course. The Bouse has been dated by the K–Ar method at 5.47 ± 0.2 million years (Damon et al. 1978). In the Pierce Ferry area, basalts intercalated with Colorado River gravels flowed down the valley of the Colorado when it was within 330 feet (100 m) of present grade (Lucchitta 1967, 1972). The flows have been dated by K–Ar methods at 3.8 million years (Damon et al. 1978). This evidence indicates that the end of interior-basin deposition and the establishment of through-flowing drainage along the lower Colorado River in its Basin and Range course occurred between four and six million years. No lower Colorado River existed before that date.

Evidence of various kinds from the western Grand Canyon region leads to similar conclusions. According to Lucchitta (1975), the Shivwits Plateau is capped for the most part by upper Miocene lavas that overlie uppermost Paleozoic and lowermost Mesozoic rocks. A long and narrow finger of the Plateau juts southward for about 20 miles (30 km) into the Grand Canyon (Fig. 15.7). This finger is surrounded on three sides by the Grand Canyon. Relief of nearly 1 mile (1.5 km) within a horizontal distance of a few miles between the top of the plateau and the Colorado River results in an extremely rugged topography of canyons, cliffs, and buttes. In contrast, the basalt capping the plateau, dated by the K–Ar method at 7.5 and 6.0 million years (Lucchitta and McKee 1975; Lucchitta 1975), overlies a remarkably flat and smooth surface with only a few meters of relief. In several places, cone-shaped vent area for the basalt are truncated by the cliff-forming edge of the Shivwits Plateau, yet there are no remnants of Shivwits lava within the Grand Canyon.

Lavas of the Shivwits Plateau locally overlie the erosional featheredge of the Triassic Moenkopi formation on the resistant Permian Kaibab Formation (Fig. 15.6). The featheredge formed northwest-trending strike valleys bounded on the northeast by a scarp capped by the Triassic Shinarump Member. This is a situation common on the Colorado Plateau. The Shinarump shed its pebbles into the valley in a way that preserved the pebbles both beneath and on top of the basalt (Lucchitta 1975). Today, the scarps are gone, and their place is taken by canyons tributary to the Grand Canyon (Figs. 15.6 and 15.9).

On Grassy Mountain, which is on the Shivwits Plateau 15 miles (25 km) northeast of the Grand Canyon (Fig. 15.2), a six-million-year-old basalt overlies remnants of gravels that contain pebbles of igneous and metamorphic rocks. The gravels also contain metamorphosed volcanic rocks similar to rocks of Proterozoic age cropping out south of the present Colorado Plateau margin. The gravels rest on the Moenkopi Formation. The only reasonable source for the pebbles is to the south because in other directions, erosion has cut down only to Paleozoic and Mesozoic rocks. Neither the pebbles nor the arkosic matrix of the gravels shows much sign of weathering; this suggests that the gravels are not reworked from older deposits. The lack of weathering within the gravels, together with the lack of a weathered zone between the gravels and the basalt, suggest that the two are of approximately the same age. The conclusion is that streams flowed northward across the present course of the western Grand Canyon as recently as six million years ago (Lucchitta 1975). These streams presumably were tributary to a larger stream that may well have been the ancestral Colorado River, but the course of this river would have been north or northwest of the western Grand Canyon. This interpretation is supported by other remnants of gravels that

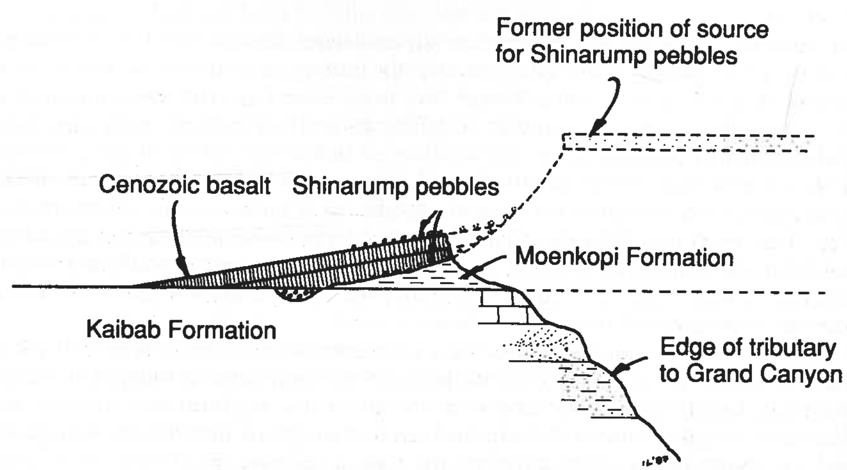

FIGURE 15.9. Diagram showing how scarp previously forming northeast side of Shivwits strike valleys occupied by basaltic lavas now is the site of a deep canyon tributary to the Grand Canyon. Looking approximately northwest.

are composed of exotic lithologies and are found in several places in the Arizona Strip country north of the Grand Canyon.

Collectively, the various features of the southwestern Colorado Plateau indicate that no rugged canyon topography and no Grand Canyon existed in the western Grand Canyon region as recently as six million years ago, the age of the young lavas in the southern Shivwits Plateau (Lucchitta 1975). The Colorado River must have become established in the western Grand Canyon after that date. It may, however, have flowed northwestward through the Strip country long before.

Post-Rifting

The evidence summarized above indicates that the transition from interior drainage to the through-flowing drainage of the lower Colorado River probably occurred between five and six million years ago, or at the Miocene–Pliocene boundary. Miocene, Pliocene, and Quaternary deposits of the Colorado River are distributed widely along its lower course. Included are estuarine deposits (upper Miocene and Pliocene Bouse Formation); probable deltaic deposits (Miocene and Pliocene Imperial chiefly reflecting downcutting; and fine-grained deposits (Pleistocene Chemehuevi Formation) produced by temporary aggradation. Information on these deposits has been summarized by Lucchitta (1972), who interpreted the data in light of the history of the lower Colorado River.

The Bouse Formation is particularly informative. In the Yuma area (Fig. 15.1), the Bouse is distributed widely in the subsurface even well away from the Colorado. Upstream along the river, however, the formation is restricted largely to the river valley. Fossils in the Bouse indicate a brackish estuarine environment that became progressively less salty northward along the river valley (Smith 1970). The Bouse rests on an erosional unconformity with deposits that reflect interior drainage and that locally interfinger upward with gravels of the Colorado River (Metzger 1968; oral communication, 1969).

The evidence indicates that the lower Colorado River became established early in Bouse time (latest Miocene) along its present course and that its prograding deltaic deposits progressively filled the estuary in the southward direction. Eventually, the delta reached the Salton trough of California, where it is represented by the thick Imperial Formation. This formation contains a well-defined horizon above which Cretaceous coccoliths found elsewhere only in the Mancos Shale of the Colorado Plateau make their appearance; the horizon signals the capture of the old, upper Colorado River by the developing and headward-eroding lower Colorado.

A 3.8-million-year-old basalt flow associated with indurated gravels of the Colorado River occurs about 330 feet (100 m) above present river grade at Sandy Point (Fig. 15.2) in the upper Lake Mead area (Longwell 1936; Lucchitta 1967). This basalt is part of the extensive basalts of the Grand Wash (Figs. 15.2 and 15.9) and flowed for several miles along the valley of the Colorado River, which must have been nearly as low then as it is now.

In the western Grand Canyon basaltic flows occur at river level for a long distance. These flows, which have their source in the area of the Uinkaret Plateau (Fig. 15.2), are the intracanyon lavas of McKee et al. (1967) and have been dated at about 1.2 million years. The lavas shows that at the time of their emplacement, the western Grand Canyon was a deep as it is today.

The upper Cenozoic deposits along the lower course of the Colorado thus show that the river came into being five to six million years ago when the Gulf of California opened up. The river extended itself southward, or downstream, by filling it estuary with deltaic deposits. It also extended itself by headward erosion and the integration of former interior drainages. By the time the delta reached the Salton trough area, the head of the river had captured the old upper Colorado River. By 3.8 million years ago, the river was essentially at its present grade in the upper Lake Mead area, and by one million years ago (at the latest), it was at its present grade in the western Grand Canyon. According to these figures, the western Grand Canyon was excavated during the interval between six million years ago (younger Shivwits lavas) and one million years ago (intracanyon basalts). More likely, it began to be cut shortly before five million years ago as it is today. The rapid downcutting almost certainly was aided by the nearly 0.6 mile (0.9 km) of uplift experienced by the western edge of the Colorado Plateau and adjacent Basin and Range Terrain since the time when the Lower Colorado River was established in its present course (Lucchitta 1979).

Much of the course of the lower Colorado River in the western Grand Canyon and upper Lake Mead areas can be explained through structural or topographic features. Thus, the Virgin Canyon section (Fig. 15.2) probably is at the margin of an ancient fan spilling southward from the south Virgin Mountains; the Greggs Basin–Iceberg Canyon section (Figs. 15.2, 8) is along Wheeler fault; the Pierce Ferry section is through the lowest spot in the old Muddy Creek basin; the Hualapai Plateau section is along a strike valley developed at the foot of the upper Grand Wash cliffs (southwest-facing rim of the Shivwits Plateau); and the section along the east side of the Shivwits Plateau is developed near the Hurricane fault.

SUMMARY

A striking conclusion that emerges from the study of the Colorado River is that even a canyon as intricate and immense as the Grand can be carved in a surprisingly short time—five million years, probably substantially less, in spite of the tough and remarkably undeformed rocks that make up its walls. Perhaps the key to understanding this phenomenon is the fact that in canyons the volume of material removed per unit of downcutting is small compared with that for more open valleys. In other words, the rate at which material is carved from a canyon is small in relation to the rate at which the floor of the canyon is lowered (Lucchitta 1967).

Another and perhaps even more striking conclusion is that one can draw a parallel between the development of physical systems, such as drainage networks, and the Darwinian concept of biological evolution based on survival of the fittest and natural selection. In river systems, the external agent that triggers change is tectonism rather than the random mutations of biology. Competition between drainages occurs through changes in gradient, whereby rivers whose gradient is increased are favored, and those whose gradient is reduced are handicapped.

The battles for survival are fought with headward erosion and stream piracy or capture. The ultimate result is a succession of drainage configurations that change with time in response to external forces—chiefly, the deformation of earth's surface. Because any particular configuration is merely a still frame within the movie of evolution, ancestors may bear little resemblance to their descendants.

• 16 •

Hydraulics and Geomorphology of the Colorado River in the Grand Canyon

Susan Werner Kieffer

INTRODUCTION

The Colorado River and its tributaries drain much of the southwestern United States, ultimately emptying into the Gulf of California in Mexico. Only the Mississippi River exceeds the Colorado in length within the United States. The potential use of the water of the Colorado River for irrigation, hydroelectric power, and domestic purposes was recognized more than a century ago. In 1905, severe floods on the Colorado River caused extensive damage in the Imperial Valley of California, and political pressure arose for construction of flood control/storage dams on the river. The development and regulation of the river expanded rapidly until, at the present time, the Colorado River is sometimes referred to as "the world's most regulated river" (National Research Council 1987, p. 18). Additional information about the history of the river's development is summarized in Kieffer et al. (1989), and so further discussion in this chapter focuses on the portion of the Colorado River that lies mainly within Marble and Grand Canyons—that is, between Glen Canyon Dam and Lake Mead.

It is convenient to divide the history of the Colorado River within the Grand Canyon into three (very unequal) time spans: (1) the time of unregulated flow, prior to the finishing of Glen Canyon Dam in 1963; (2) the time of reservoir filling from 1963 to 1983; and (3) the current operational period, when Lake Powell is filled to a capacity that optimizes requirements for water storage and power generation. The hydraulics and sediment transport capacity of the river have been very different in each of these times.

The downcutting of the Grand Canyon was accomplished by a river capable of carrying large sediment loads during the time of unregulated flow. Monitoring at the U.S. Geological Survey's gage station near Phantom Ranch from 1921 to 1962 showed that typical average daily discharges were near 17,000 cubic feet per second (cfs). The mean annual flood was 77,500 cfs, but larger floods were not uncommon. For example, a flood of 300,000 cfs occurred in 1884. Sediment loads were large during this time; the average daily sediment load past the Phantom Ranch gage exceeded 300 tons per day.

During the reservoir-filling period, the average daily discharge was reduced to 11,000 cfs. Sediment that originates upstream of Lake Powell has been trapped in the lake since 1963, and so the average sediment load has been reduced to about 50 tons per day. Much of the sediment now available downstream from

Glen Canyon Dam is contributed by the two large tributary streams, the Paria (near Lees Ferry) and the Little Colorado River. Thus, the Colorado River (whose name in Spanish means "colored red" in reference to its former heavy sediment load) now often runs clear.

Between Lees Ferry (near the Arizona–Utah border) and Diamond Creek (about 220 miles downstream), the water surface of the Colorado River drops from 3116 feet in elevation (944 m) down to 1336 feet in elevation (405 m). The water surface does not maintain a constant gradient between these two elevations, but consists, instead, of a series of pools that are relatively flat and tranquil and rapids that are steep, fast, and turbulent. The bed of the river likewise is not uniform in gradient, but consists of relatively flat sections under the tranquil pools, debris dams (natural wiers) where tributaries enter, and scour holes. These characteristic configurations of the water surface and channel bottom have been created over thousands of years by hydraulic, hydrologic, and geomorphic interactions superimposed on the tectonic forces that drive the uplift of the Colorado Plateau.

The purpose of this chapter is to show how an understanding of hydraulic dynamics in individual stretches of the river near rapids can help us interpret the shape of the river channel and the geomorphic and hydrologic history of floods and erosion over the past thousands to perhaps hundreds of thousands of years.

A DESCRIPTION OF THE MAJOR FEATURES OF RAPIDS

The pool and rapid sequences of the Colorado River are very obvious in high-altitude photographs (Fig. 16.1a) because the rapids occur where the river is constricted by debris from side canyons, and these constrictions, as well as the white water of the rapids, are obvious in the photographs. Photographs taken at lower altitude (Fig. 16.1b) reveal that at the constrictions the water surface changes from a featureless pool into a white, wave-filled chute. At the bottom of the chute, the water squirts out of the constriction into the next quiet region like a jet from a fire hose. Adjacent to the jet are eddies and beaches, sites of some of the most critical ecologic zones and recreational beaches along the river.

These stretches of the river between one relatively tranquil pool (or main-stem segment) and the next are the famous rapids of the Colorado River. They

FIGURE 16.1.(a) (on opposite page.) High-altitude photograph (U.S. Geological Survey Series GS-VFDC-C, #6-134, taken June 24, 1982) of the Middle Granite Gorge of the Grand Canyon. River flows from bottom of photo toward top. Note that each rapid is associated with a constriction in the river at the base of a side canyon (or two side canyons in the case of Deubendorff Rapids; see text). The alternation of quiet pools with high-velocity rapids, visible in four places in this photographs, has been called the "pool and rapid" sequence of the Colorado River (Leopold 1969). (b) A lower altitude aerial photograph centered on Deubendorff Rapids, showing the transition of the river from a quite pool on the left into a foaming jet with standing waves in the constricted region, and the jet emerging like fire hose (labeled "tail-waves") into the next quiet region of the river. (Photograph by U.S. Bureau of Reclamation, 1984.)

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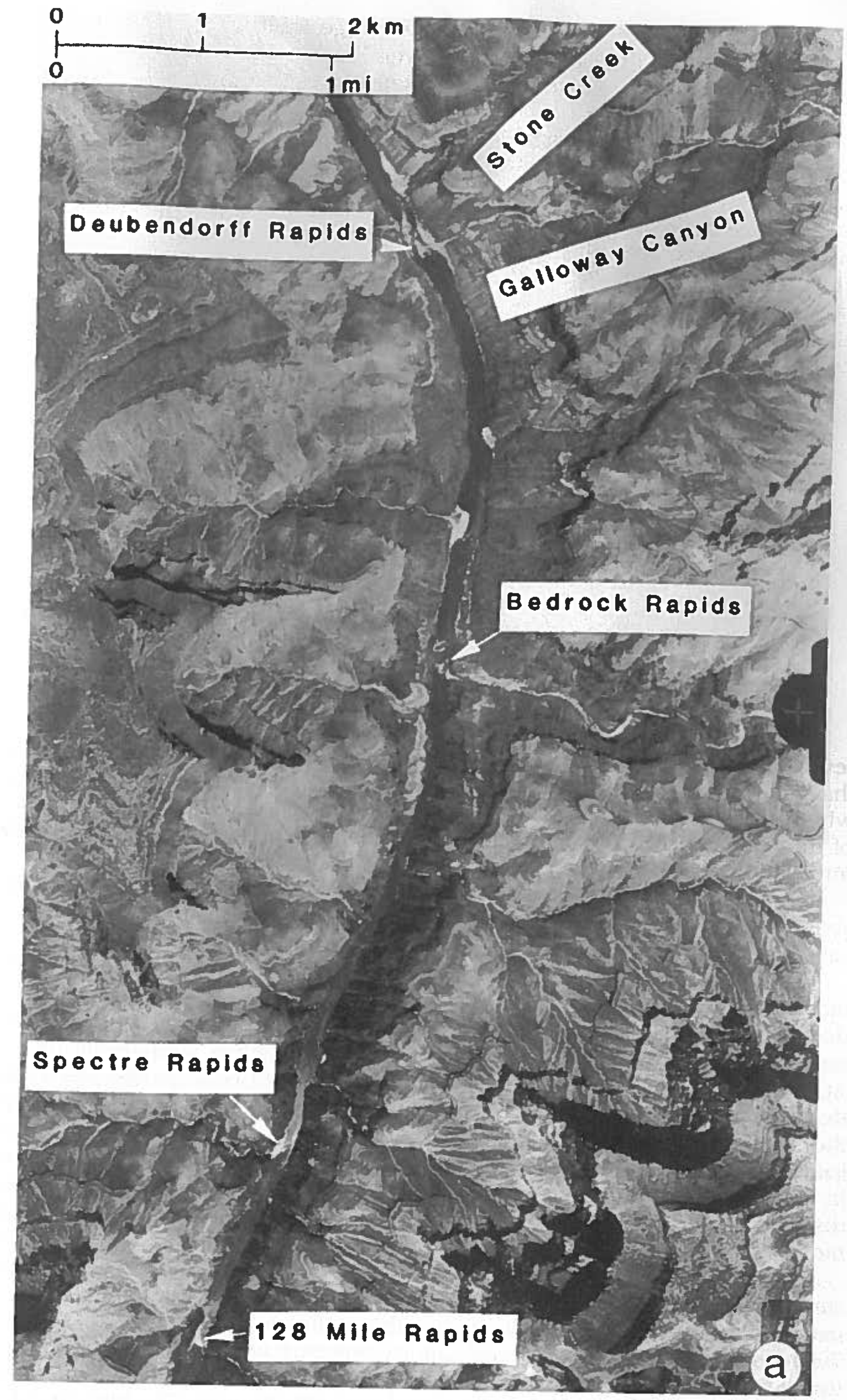

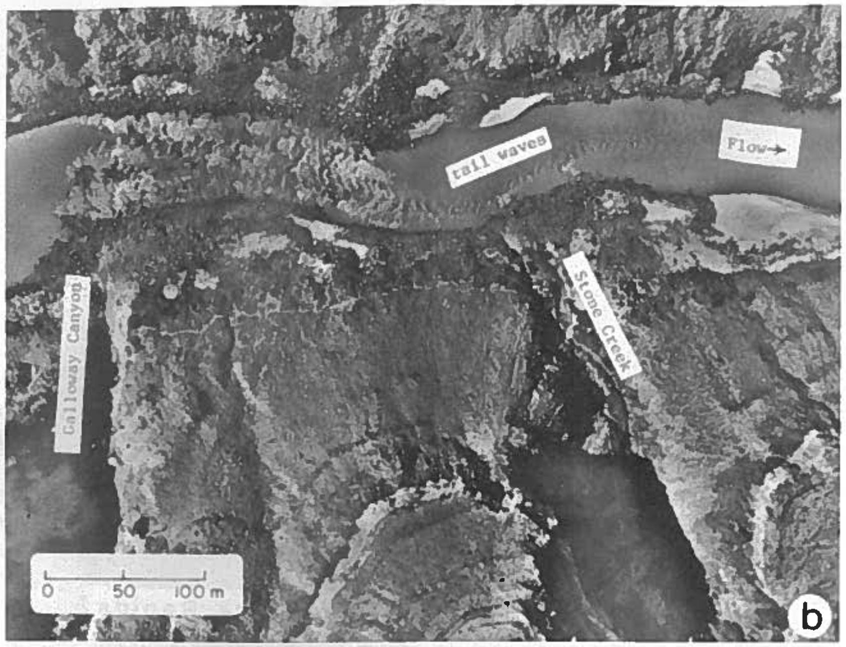

Figure 16.1. *(Continued)*

reveal many fundamental fluid mechanical phenomena on a grand scale: waves that stand still while water flows through them; zones of smooth, tranquil flow where large boulders protruding through the water cause hardly a ripple; zones of turbulent, aerated flow where the large boulders cause major hydraulic features; and even zones where the water flows backwards.

Common features of rapids are illustrated in Figures 16.2 and 16.3. In describing the rapids, we use a mixture of terms from hydraulics and from the vocabulary of those who have explored and described the river.

Rapids in the Grand Canyon typically form where a debris fan from a tributary canyon constricts the river. Above each rapid, the river is wide, relatively deep, and tranquil. In this chapter, the term "pool" is reserved specifically for these tranquil sections of the river immediately upstream of a rapid (Fig. 16.2a). At most discharges, a pool is a hydraulic backwater (a concept discussed later in more detail). Conceptually, a pool can be thought of as a pond formed by the debris-fan dam. At discharges between about 7000 and 30,000 cfs, water velocity is low in the pools, typically less than 1 foot per second (0.3 m/s). Water in the pools is deep, typically more than 15 feet (5 m) at the low end of this discharge range and more than 30 feet (9 m) at the higher end of this discharge range (Kieffer 1987).

At the downstream end of a pool, water accelerates gradually in the constriction (Figs. 16.2b, c; 16.3a, b, c), reaching velocities more than an order of magnitude higher than the velocities in the pool, even at discharges as low as 7000 cfs (Fig. 16.3d). Velocities of this magnitude have been found at most of the major rapids where velocity measurements were performed (they were not performed at Cremation–Bright Angel and 24.5-Mile Rapids). The highest velocities measured were approximately 33 feet per second (10 m/s) at Hermit Rapids (Kieffer 1988, Map 1897 F). In the converging portion of the channel (where rapids are found), standing waves (or laterals) bound a tongue of smooth, accelerating water, upon which may stand smooth, undulating, nonbreaking waves called rollers. In the constriction and the diverging portion of the river channel, crisscrossing lateral waves typically intersect to form high-amplitude breaking waves or haystacks.

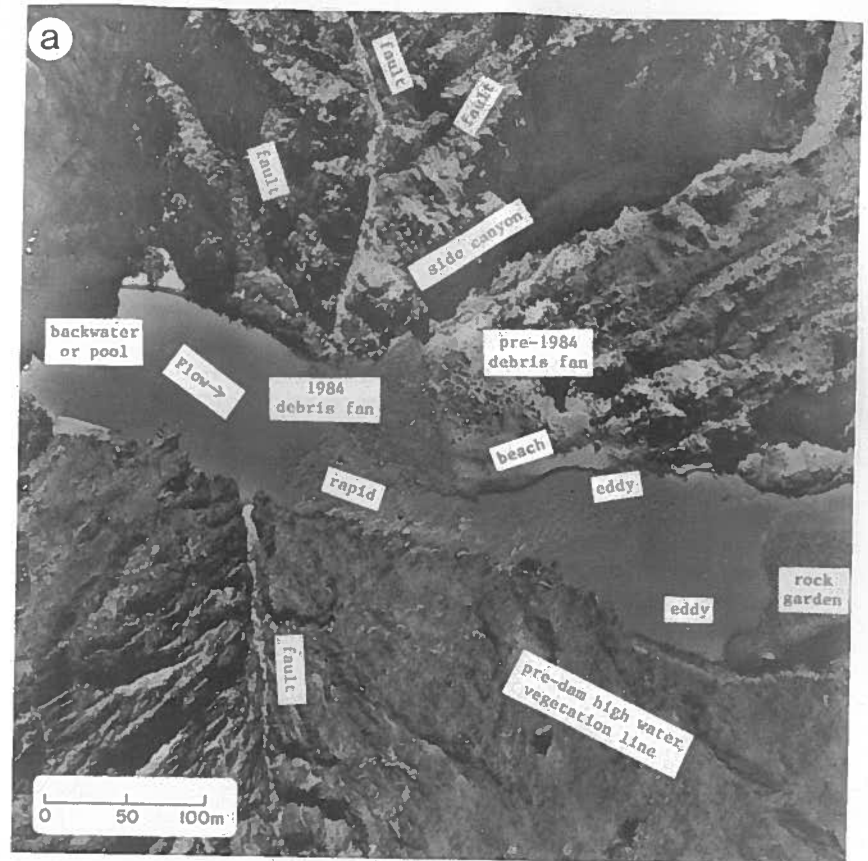

Figure 16.2. (a) Aerial photograph of Granite Rapids at a discharge of 5000 cfs, showing the geomorphic features common to many rapids. Terms defined in text. [Photograph by U.S. Bureau of Reclamation, 1984. Regional faults from Dolan et al. (1978).]

Below each rapid is a zone in which the depth is intermediate between that of the shallow constriction and the deep pool. In this zone, the so-called "runout" of the rapid, water velocity still is relatively fast, typically 10 to 15 feet per second (3–5 m/s), but it decelerates toward the ambient conditions in the next pool, the so-called "tailwater" conditions in hydraulics. Strong vertical motions occur in the water of this region. The shear that results can give rise to turbulent boils with as much as 1 foot (0.3 m) of superelevation, indicating a vertical velocity of at least 8 feet per second (2.4 m/s) (Leopold 1969).

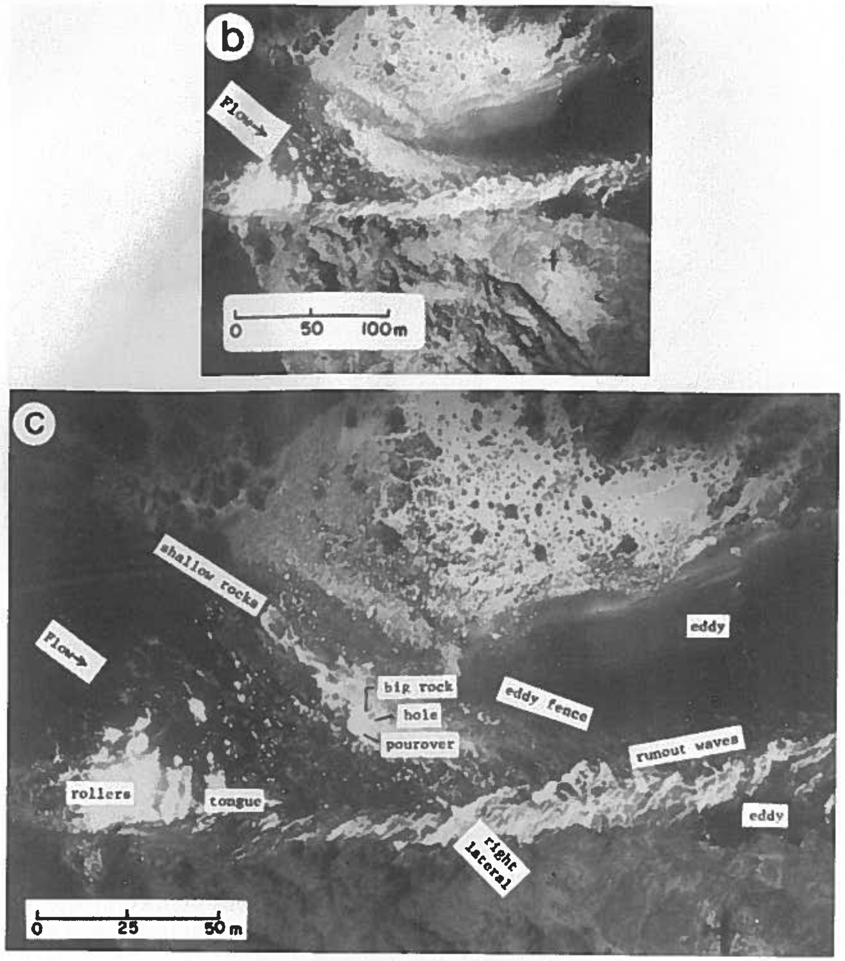

FIGURE 16.2. (*Continued*) (b and c) Aerial photographs of Granite Rapids at 30,000 cfs showing typical wave structures in rapids. (Photographs by U.S. National Park Service.)

Obstacles in the bed of the channel (such as rocks or bedrock protrusions) also cause a variety of wave patterns—including holes, curlers, rooster combs, and sculpted waves. A definition of these less commonly known terms follows (refer also to Fig. 16.2):

1. *Tongue:* Smooth water between the first two strong lateral waves (right and left) at the top of a rapid.
2. *Roller:* A wave that stands oblique, often perpendicular, to the current and breaks back onto the current; the term "nonbreaking roller" is used to indicate the smooth, rolling waves often found on the tongue.
3. *Lateral:* A wave standing oblique to the current near the top of a rapid, usually emanating from shore.
4. *V-wave:* The composite wave formed when opposing laterals intersect.

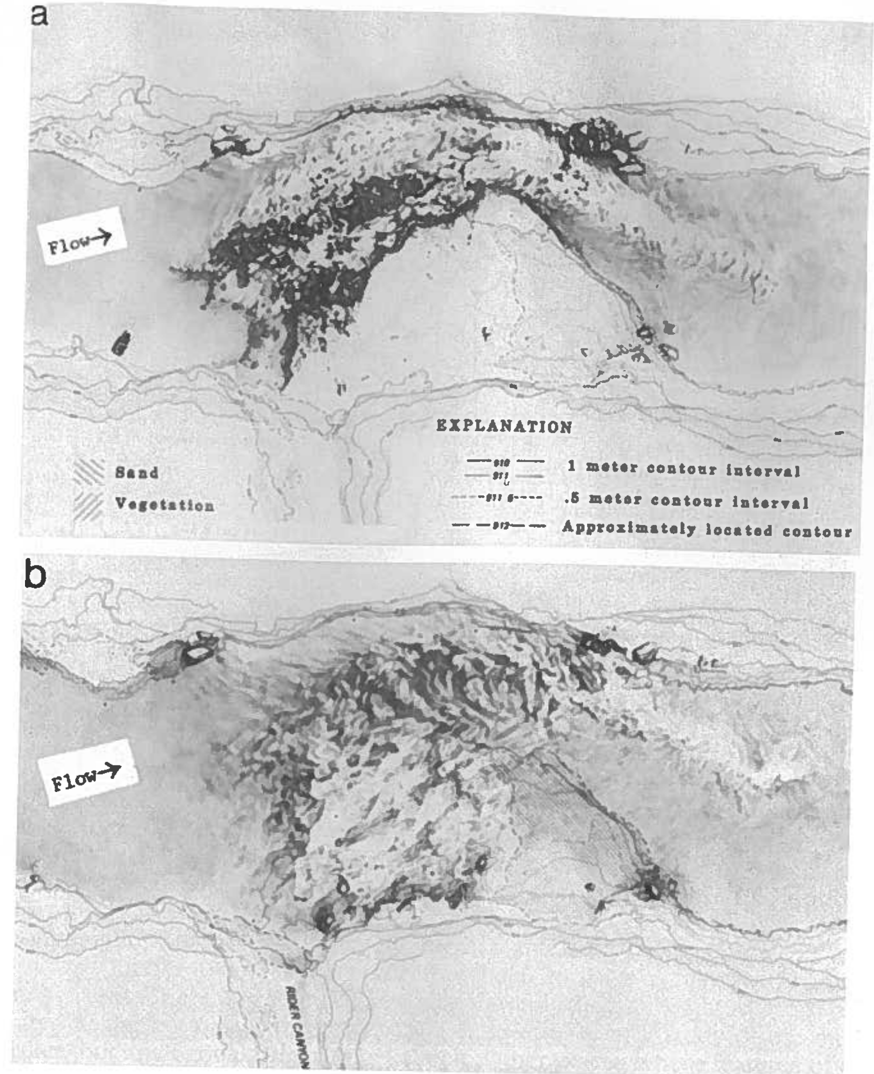

FIGURE 16.3. (a–d) Parts of preliminary hydraulic map I-1897-A, House Rock Rapids (Kieffer 1988). (a–d) Map views; contour intervals indicate with solid lines are 1 meter; dashed lines on beach in (a) are 0.5-m contours. (a) Topographic contours and standing waves at 5000 cfs; (b) the same at 30,000 cfs; (c) water-surface profile at 5000 cfs; (d) water-surface velocities and streamlines of floats at 5000 cfs. In (a)–(d), flow direction is from left. In (d), numerals indicate velocities (in m/s) along the streamlines between the adjacent dots. Trajectories of the floats were determined from analysis of movies taken from the camera station indicated. The boat, shown only for scale, is a standard commercial motor raft that is 10 m (33′) long. These maps are for schematic illustration of hydraulic features only and are not intended for navigation purposes.

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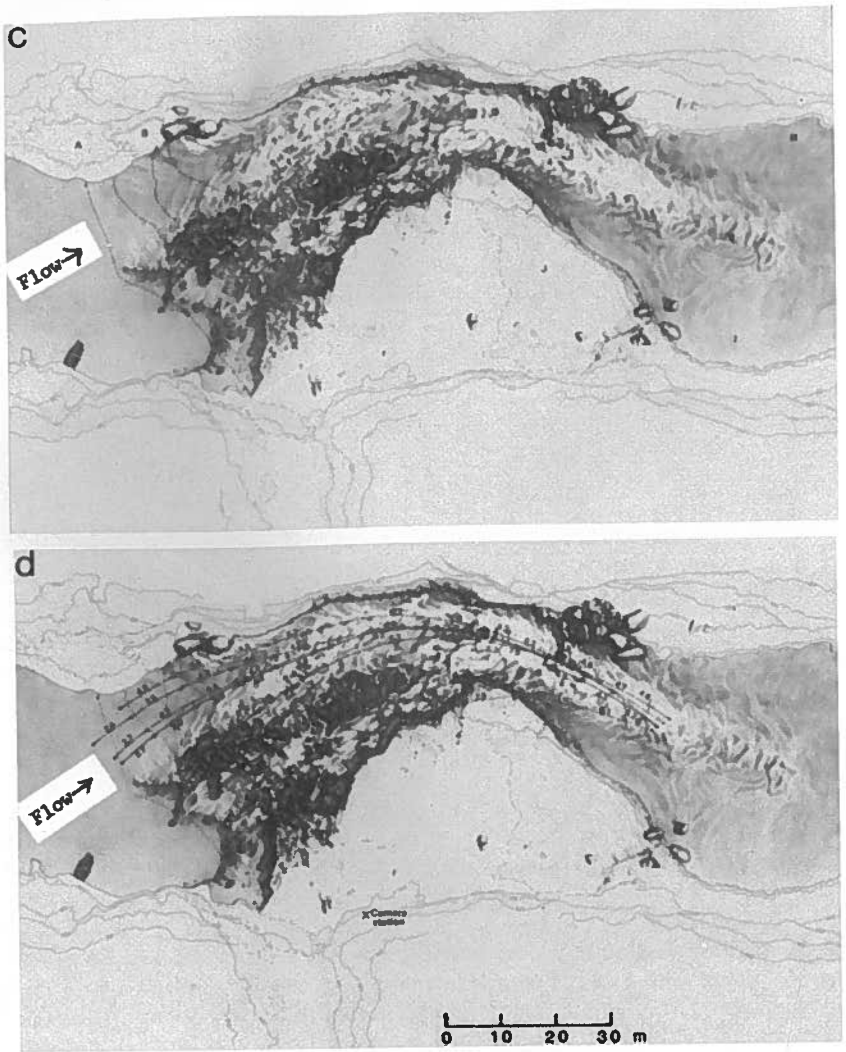


FIGURE 16.3. (*Continued*)

5. *Eddy fence:* The shear zone between two currents with different velocity magnitudes or directions.
6. *Pourover:* A zone where water "pours over" an obstacle, obtaining a large, downward component of velocity.
7. *Hole:* A trough in a standing wave, usually deep.
8. *Haystack:* A pyramidal wave (shaped like a haystack), usually breaking on top and sending spray in all directions.
9. *Rooster comb:* A haystack elongated in the downstream direction.
10. *Runout:* A zone of standing, generally nonbreaking (or weakly breaking) waves at the bottom of a rapid; more or less synonymous with "tailwaves."
11. *Rock garden:* An area of rocks in the channel within or downstream of the diverging section of a rapid.

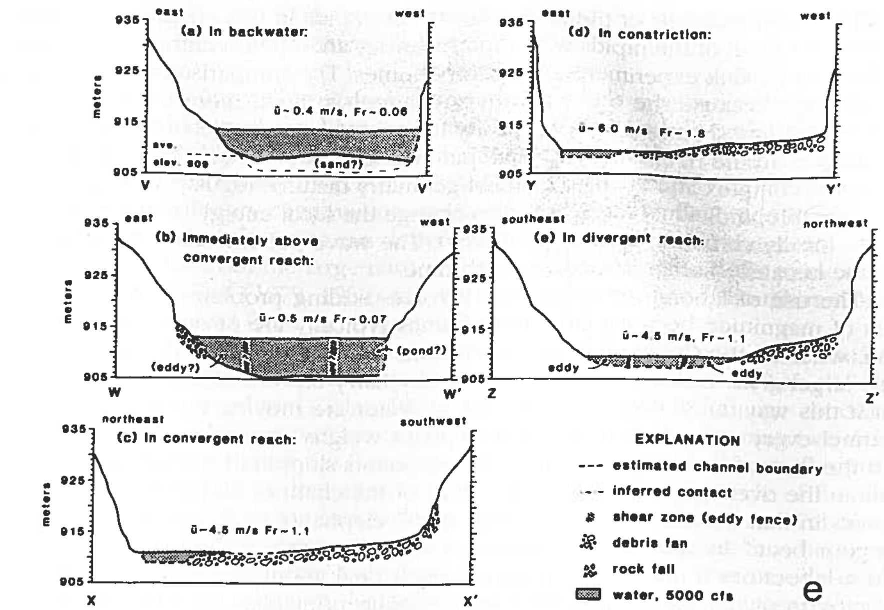


FIGURE 16.3. (e) Location of cross sections is approximately described above each cross section and view through cross section is downstream. Average water velocity is given by *u*, and Froude number is given by Fr.

To visualize some of these features and to understand the hydraulics of the rapids described and interpreted in the rest of this chapter, a reader may rely on, or create, intuitive mental background pictures of the rapids by several methods. An excellent description of the river, its rapids, and the eddies and whirlpools below then can be found in Beer (1988), an account of two men who swam the whole length of the river. Alternately, many features of the rapids can be seen on video (Kieffer 1986). If the reader has been to (or can hike to) Phantom Ranch via the Bright Angel or Kaibab trails, careful study of Bright Angel Rapids located at the bottom of these trails will be useful, as will study of the much smaller, but analogous, Bright Angel Creek. Other places where features of the flow that are discussed here can be seen are in small steep-gradient streams (particularly in the Pacific Northwest), in storm gutters along city streets when water is flowing, and in amusement parks that have water rides!

For the reader who has not been around any rapids, it also may be helpful to spend a few minutes at the kitchen sink—using the faucet to simulate the river and a cookie sheet, some curved dishes or pie pans, and a few utensils to create miniature rapids. Place a cookie sheet at an angle under the faucet, and turn on the water, watching the change in flow patterns as the discharge through the faucet is increased. Try confining the flow into a channel with a few straight-edged implements (yardsticks, knives, etc.), and then try making a converging–diverging flow by placing curved implements on the cookie sheet. As the flow changes characteristics, try placing the tip of a spoon in various flow regions to simulate an obstacle (like a rock) in a rapid, and note the different wave behaviors that can be obtained. This experiment has the potential to be messy, as well as instructive!

The interpretations of the hydraulics of the rapids in this chapter are based on a comparison of the rapids with slightly fancier and better-controlled versions of the kitchen-sink experiment—laboratory flumes. The comparison is only semiquantitative because the Colorado River channel is much more complex than laboratory flumes, and we have relatively little data about the details of the channel shapes. In the river channel, wave patterns are irregular on a microscale because the complex and detailed channel geometry disturbs regular wave patterns (for example, individual large boulders change the local energy of the flow and create locally complex hydraulic patterns). The wave patterns also are irregular in time because discharge changes with time.

The use of laboratory flume data involves scaling problems of several orders of magnitude because laboratory flumes typically are of meters in dimension, whereas the Colorado River dimensions are one or two orders of magnitude larger. This is descriptively conveyed by Larry Stevens (1985, p. 25): "Let's put it this way; at 32,000 cfs, 1000 tons of water are moving through the river channel every second. If an average elephant weighs about 5 tons, this means that the flow of the river is equal to 200 elephants skipping by every second. A hole in the river may take up about a third of the channel, so the hydraulic dynamics in that hole are about the same as 67 elephants jumping up and down on your boat." In spite of the difference in scale between the Colorado River and a laboratory flume, we can learn a great deal about the hydraulics of the river by applying basic open-channel hydraulic principles. In turn, increased knowledge of the properties of the river and its interaction with the channel margins can help us better manage the river corridor (as in the current U.S. Bureau of Reclamation Glen Canyon Environmental Studies project), understand potential hazards for recreational boating, and interpret the geomorphology of the river channel.

LOCATION OF THE RAPIDS AND DESCRIPTION OF CHANNEL MORPHOLOGY

Many geologists have speculated on the origin of the pool–rapid–runout sequence (Leopold 1969; Dolan et al. 1978; Howard and Dolan 1981). Important factors in determining the origin and location of the rapids are gradient and variations of gradient along the river, spacing of the rapids, and the relation of the rapid spacing to the location of structural controls. We can divide the channel of the river into stretches that have different geomorphic and hydraulic characteristics (Howard and Dolan 1981). The frequency and, to some extent, the magnitude of the rapids depends on their location in these stretches. Typical stretches are as follows:

1. A wide valley with a freely meandering channel (for example, miles 67–70, near Tanner Rapids)
2. Valleys of intermediate width with tributary fan deposits (in these valleys, the river usually has cut into soft sandstones or limestones; for example, the few miles just downstream of the Little Colorado River near mile 61.5)
3. Narrow valleys in fractured igneous and metamorphic rocks (for example, "Granite Narrows" through miles 77–112; Fig. 16.1)
4. Narrow valleys of roughly uniform width and few constrictions in massive Muav Limestone (for example, miles 140–165)

5. Nearly flat stretches where the channel bottom is sandy (for example, miles 1–10 and parts of Marble Canyon)

The rapids occur almost exclusively where floods in tributary canyons, controlled by local or regional faulting or jointing, have delivered large boulders into the river channel (Dolan and Trimble 1978) (see the example of Fig. 16.2a, showing the faults and debris fan at Monument Creek where Granite Rapids is formed). Because the tributary canyons are much steeper than the Colorado River channel, floods in the tributaries can deliver boulders into the main channel that may be too large for even the large natural floods of the Colorado River to move [the discussions of Graf (1979, 1980) are relevant to this problem, though not based specifically on data from the Grand Canyon].

Debris fans from the tributary canyons can form on one or both banks of the Colorado River at the tributary junctions because the controlling faults cross the canyon (Howard and Dolan 1981). If meteorologic and drainage conditions are conducive, floods can occur in opposing tributaries, thus forming two debris fans. The relative size of the fans depends on the availability of movable material in the contributing drainages, on the frequency and magnitude of floods in each drainage, and on the tributary gradient (for a recent study, see Webb et al. 1987). It is, however, more common to have one of the debris fans be significantly larger than the other (see Figs. 16.1 and 16.2). The river usually erodes through the weaker parts of the debris fan, but the erosion may extend into the bedrock wall at the distal end of the debris fan if this material is easily eroded. This process can result in the formation of a pronounced change in the course of the river, and many rapids occur on local curves of the river (Fig. 16.1b).

In spite of the variations in the nature of bedrock along the course of the river discussed above and the structural control exerted by faults and joints on the location of the rapids, the river channel at the rapids is remarkably uniform in shape where it cuts through the tributary debris fans (Fig. 16.4). The unconstricted channel is about 300 feet (90 m) wide at 10,000 cfs (the width varies with discharge but this effect is not significant at the level of overall generality

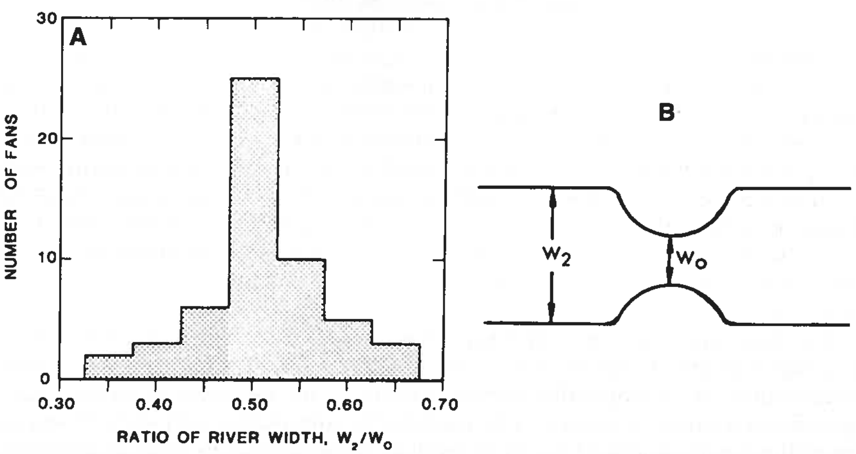


FIGURE 16.4. (a) Histogram of constriction of the Colorado River as it passes through 59 of the largest debris fans in the 248-mile (400-km) stretch below Lees Ferry (Kieffer 1985). (b) Idealized sketch of converging–diverging geometry of a rapid, showing widths W_o and W_2 used to define the "constriction" of the rapid in (a).

considered in this chapter). At each rapid, the channel narrows appreciably, typically to about one-half of the unconstricted width. Figure 16.4 shows the ratio of surface width of the river immediately upstream of a rapid to the width in the narrowest part at 54 major debris fans; this ratio is termed the "constriction" in this chapter. The remarkable feature of the histogram of Figure 16.4 is the uniformity of this shape parameter—sharply peaked with nearly half of the rapids at the value 0.5. Why is the channel so uniform in shape at so many rapids, and why is this value specifically 0.5? The discussion of hydraulic–geomorphic interactions presented in this chapter will suggest an answer to this question.

The values of constriction plotted in Figure 16.4 were measured on 1973 air photographs (Fig. 16.2a is taken from this 1973 series of photographs). Because Glen Canyon Dam was operating under daily fluctuating flow conditions when the 1973 air photograph series was taken, the discharge at different rapids along the photograph series varied from about 7000 to 30,000 cfs. In the analysis that follows in this chapter, I need to use an average width for the river at places above, in, and below the rapids. This average width must be one that, together with an average depth and an average flow velocity, satisfies the requirement for conservation of discharge. A problem is that the surface width of the water measured from the photographs and shown in Figure 16.4 is not identical to the average width of an idealized channel. For example, in the histogram of Fig. 16.4, Crystal Rapids is the point at the furthest left, at a constriction of 0.33. However, idealization of the channel shape to a rectangular cross section suggests that the average channel constriction at Crystal was about 0.25 in 1973. This average value, rather than the value based on surface widths, is used in the hydraulic discussions in this chapter.

At this time, we cannot make a histogram like Figure 16.4 based on actual channel cross sections or on average widths because we lack detailed surveys of the river channel. However, the author has prepared a series of maps of the ten largest rapids along the river (House Rock, 24.5-Mile, Hance, Cremation–Bright Angel, Horn Creek, Granite, Hermit, Crystal, Deubendorff, and Lava Falls; Kieffer 1988). These maps have a topographic base sufficiently accurate to allow the value of constriction and its dependence on river stage to be determined more accurately. They also have additional data that allow visualization of hydraulic features in the rapids. Figure 16.3 is an example of parts of the map of House Rock Rapids.

Each hydraulic map contains the following: (a) topographic contours of the channel (Fig. 16.3a–d); (b) hydraulic information at two or more discharges (compare Figs. 16.3b and 16.3c); (c) water-surface elevations at different discharges, that is, rating curves and water surface profiles, shown implicitly by comparison of the shorelines in Figures 16.3b and 16.3c, and explicitly in figure 16.3b; (d) velocity and streamline data at one or two discharges (Fig. 16.3d); (e) channel cross sections, showing both the shape of the channel and the hydraulic conditions of the water (Fig. 16.3e); and (f) a detailed discussion of the characteristics of the rapid and interpretation of the data (not shown in Fig. 16.3).

Careful study of the channel shape in three dimensions reveals that at rapids the channel is constricted both laterally and vertically. In map view (for example, as in the air photographs of Figs. 16.2a and 16.2b, or on the topographic maps of the channel in Figs. 16.3a and 16.3b), the lateral constriction can be seen easily if the discharge is low enough that the river does not cover the debris fans. In vertical cross sections that are perpendicular to the flow direction (such as those of Fig. 16.3e) or those that are parallel to the flow directions (such as the fathometer tracings of Fig. 16.5), we can see that the channel of the Colorado River is constricted also by a vertical bulge in the bed, caused by the debris fan.

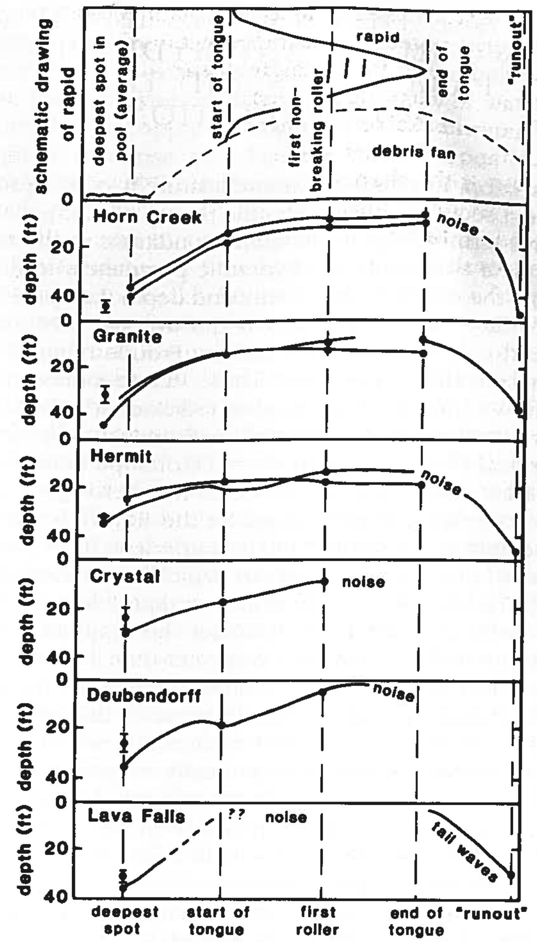


FIGURE 16.5. Summary of fathometer tracings obtained by the author, Julie Craf (U.S. Geological Survey, personal communication, 1986), and Owen Bayham (river guide, personal communication, 1986). No horizontal dimensions are implied. The top drawing is a schematic map view of a typical rapid showing the maximum depth measured in the upstream pool, the start of the tongue, the location of the first nonbreaking rollers, the end of the tongue, and the runout of the rapid. Where known, discharges are indicated. Similar profiles occur in every rapid for which fathometer data were obtained.

Detailed studies of the rapids show that the constrictions typically begin underwater well upstream of the location of the first waves in the rapid. The water gets shallower and faster as the constriction gets tighter, and the most complex and largest waves in a rapid occur in the shallowest and narrowest part of the channel. The channel deepens again under the runout or tailwaves. Although the water surface drops several meters in most rapids, the elevation of the channel bottom is often the same upstream and downstream of the rapid to within experimental accuracy (~1 m). Thus, the rapids are not formed by sudden drops in channel bed elevation, but rather are a result of vertical and lateral constriction.

WHY ARE THERE WAVES IN RAPIDS? TWO IMPORTANT HYDRAULIC PARAMETERS AND THEIR INTERPRETATION

In the previous section, I defined the nomenclature used to describe the rapids. In this and the next sections, I use hydraulic theory of open-channel flow to relate the descriptive terminology to hydraulic conditions in the rapids.

We can calculate two important hydraulic parameters in the pools, rapids, and runouts below the rapids from velocity and depth data presented above: the Reynolds number ($Re = uD/\mu$, where u is the average flow velocity, D is the average depth, and μ is the viscosity), and the Froude number ($Fr = u/(gD)^{1/2}$, where g is the acceleration of gravity). These dimensionless numbers indicate the state of the flow. The Reynolds number indicates whether the flow is laminar ($Re < 10^4$) or turbulent ($Re > 10^4$), and the Froude number indicates whether the flow is subcritical ($Fr < 1$), critical ($Fr = 1$), or supercritical ($Fr > 1$).

Under most flow conditions, the Reynolds number is greater than 10^5, indicating turbulent flow. For example, consider the Reynolds number of a backwater—where the flow looks tranquil and nonturbulent. If the discharge is greater than a few thousand cfs, the water velocity typically is greater than 0.3 feet per second (10 cm/s). The depth typically is greater than 3 feet (100 cm). Water viscosity is on the order of 0.01 cm^2/s. Even for these apparently tranquil backwater conditions, the Reynolds number is greater than 10^5, and the flow thus is turbulent. Appearances are deceptive! In other parts of the river, u increases by up to an order of magnitude, and D can decrease by the same amount, but we generally find Reynolds numbers well in excess of 10^4. Such high Reynolds numbers imply that turbulent conditions exist nearly everywhere in the river. The turbulence has important implications for the mixing of sediment and nutrients. However, because the river is turbulent nearly everywhere, variations in the Reynolds number cannot explain the dramatic differences we find in the water structure of the pools, rapids, and runouts.

The Froude number is the dimensionless number that we need to look at to investigate differences in stability of waves and hydraulic regimes. It characterizes two states of flow that are dramatically different in energy balance and wave stability. The numerator of the Froude number depends on velocity—that is, on momentum or the square root of the kinetic energy. The denominator depends on potential energy—that is, on depth. The Froude number, therefore, is a measure of the relative importance of the kinetic and potential energies. The denominator also defines a characteristic velocity, called the critical velocity [$c = (gd)^{1/2}$], that depends only on water depth. The stability of standing waves depends on the ratio of the fluid velocity (numerator) to this characteristic velocity (denominator).

Dramatic changes in flow regime occur as the Froude number changes from less than one (subcritical) to greater than one (supercritical). These changes are similar to those that occur in transitions from subsonic to supersonic flow in gas dynamics, and this analogy may be useful for the reader in thinking about the standing waves in the rapids.

Changes in Froude number near a rapid are dramatic. In a typical backwater, $u \sim 1$ m/s and $D \sim 10$ m, so $Fr \sim 0.1$ or even less (a subcritical condition). In a rapid, on the other hand, $u > 5$ m/s, $D < 3$ m, so $Fr \sim 1$ or $Fr > 1$ (critical and supercritical conditions). Measurements show that Froude numbers exceeding 2 can occur in the rapids. Figure 16.3e shows Froude numbers for different parts of House Rock Rapids. Measurement of the Froude numbers in different parts of the rapids therefore suggests that the dramatic change in flow regime from backwater to rapid is caused by differences in the balance of kinetic and potential energy. These differences change the stability of waves in the channel. The general principles that apply to analysis of shallow-water flow in these different flow regimes comprise the classic theory of open-channel hydraulics (Bakhmeteff 1932; Ippen 1951; Ippen and Dawson 1951; Rouse et al. 1951; or Chow 1959).

At this point, the reader can develop an intuitive feeling for the significance of Froude numbers for different flow regimes and about the concept of standing waves by returning to the kitchen sink and putting a flat plate on the bottom of the sink under the faucet (this experiment is described in Thompson 1972, p. 525). As the faucet is turned on and the downgoing jet of water strikes the plate, water spreads laterally toward the edges. Depending on the rate of discharge from the faucet and the position of the plate, an inner ring of water that is moving rapidly outward in a radial direction surrounds the impact point where the jet hits the plate. If the reader has produced the correct experimental conditions, a ridge of water surrounds this inner ring a few inches radially outward from the jet. Beyond this ridge, the water is deeper and moves more slowly toward the edge of the plate. The ridge or ring where the water depth and velocity change is a standing wave that separates an inner region of supercritical flow from an outer region of subcritical flow. This standing wave, which is called a "hydraulic jump," is circularly symmetric about the descending jet of water. In a linear channel, however, hydraulic jumps commonly are linear, though they may stand perpendicular or oblique to the flow direction.

If you insert a probe, such as a finger, spoon, or small pebble, into the two different regions of flow on the plate, very different wave phenomena occur. In the inner, supercritical region, standing waves will be formed as "wakes" to the object, whereas, in the outer, subcritical region any waves formed by the object migrate upstream or downstream through the fluid. Thus, they are traveling, not standing, waves. Wave behavior in the critical region is highly unstable (by analogy, it is well known that wave instability makes maneuvering aircraft in the transonic regimen difficult, but that maneuvering becomes more stable under supersonic conditions). In all of these regimes, eddies caused by shear in the fluid can form downstream of an object immersed in the flow. These eddies must be distinguished from standing waves.

Hydraulic jumps are the basic standing waves in the Colorado River, and they occur in a variety of geometries. Large, oblique hydraulic jumps emanate from shore and are oriented downstream at an angle to the current (see Fig. 16.1b, or 16.2b and 16.2c). Normal hydraulic jumps stand perpendicular to the current. Finally, miscellaneous hydraulic jumps of various geometries stand around rocks and obstacles on the bed of the channel in regions of supercritical flow in the rapids.

The same change of conditions that gave rise to standing waves in the inner ring under the kitchen faucet and yet produced no standing waves in the outer, deeper water occur in the rapids of the Colorado River. The contrast in stability of standing waves between supercritical and subcritical flow explains why there are strong waves in rapids where the flow is supercritical, but there are no standing waves in backwaters where the flow is subcritical. Just as you can manipulate the strength, stability, and position of the circular hydraulic jump under the faucet by increasing or decreasing the flow rate through the faucet or by changing the position or angle of the plate, so the behavior of waves in the rapids depends on the flow rate of the Colorado River, the shape of the channel, and the gradient of the bed. Using these concepts, I now show how the different parts of a rapid and the waves can be semiquantitatively explained in terms of flume hydraulics.

THE PIECES AND PARTS OF A RAPID: BACKWATERS (POOLS)–RAPIDS–RUNOUTS

Backwaters (Pools)

Pools and backwaters form upstream from a rapid if changes in channel shape produce local conditions in which the given discharge cannot be accommodated in the channel cross section without a transition from subcritical to supercritical flow. To clarify this, consider a specific example of flow at 30,000 cfs. If the main channel is 100 feet wide (30 m) and 30 feet (9 m) deep, the flow per unit area is 10 ft/s (3 m/s); that is, the flow velocity required to accommodate the discharge is 10 ft/s. The flow is subcritical because the critical velocity is 17.1 ft/s (5.2 m/s) and the Froude number is 0.6. If the channel narrows to 30 feet (9 m) in width and maintains the same depth, the flow per unit width or flow velocity would have to increase to 33.3 ft/s (10 m/s) to accommodate the discharge. The flow would be supercritical with a Froude number of 1.9. Conservation of energy would not allow the fluid to accelerate in this simple way because of the change in flow regime. Instead, water would pond upstream of the rapid in a backwater to increase the depth and, therefore, potential energy of the flow, and the river would adjust itself so that the Froude number would just equal unity in the constriction.

In effect, the cross-sectional area of the whole backwater–rapid system is increased by ponding in the backwater or "pool" above a rapid (see the cross sections in Fig. 16.3e). The potential energy of the deepened pool is available for conversion to velocity in the constriction (compare depth and velocity in cross sections (a) and (b) with those in (c) and (d) in Fig. 16.3e). Within the backwater itself, the increased depth caused by the constriction reduces the velocity compared to that in an unconstricted channel of the same diameter (that is, compared to a normal main-channel flow that enters the backwater). As passengers on a raft float from the main-channel current into a backwater above the rapid, they often lose the sense of "floating" and may feel the need to row across the tranquil pond—especially if there is any upstream wind to halt all progress! For example, the backwater above Crystal Rapids, which extends approximately a mile back to Boucher Rapids, is known affectionately by river runners as "Lake Crystal." Velocities of only a few tenths of a foot per second and Froude numbers as low as 0.01 can occur in the backwaters, indicating conditions dominated by potential energy.

The Converging Section of Rapids: Tongues, Nonbreaking Rollers, and Oblique Lateral Waves

In this section, I show how features in laboratory flumes of relatively simple geometry and well-controlled discharges (Figs. 16.6, 16.7, and 16.8) relate to the more complex and variable patterns of waves in the rapids of the Colorado River (Figs. 16.9 and 16.10).

A simplified illustration of hydraulic features in flumes with converging-diverging geometries is given in Figure 16.6. The top part of Figure 16.6a is a map view of subcritical flow in an idealized converging-diverging laboratory flume, and it is singularly uninteresting because there are no standing waves. The two lines below the map view of the flume are (top) water-surface elevation and (bottom) velocity profile. The increase in velocity and decrease of water-surface elevation in the constriction in subcritical flow conditions is exactly analogous to the well-known venturi effect that gas flow shows in converging–diverging nozzles. Figure 16.6b is a map view of the same flume with supercritical flow conditions. This case is much more interesting than the subcritical case because patterns of oblique, criss-crossing waves occur in the converging

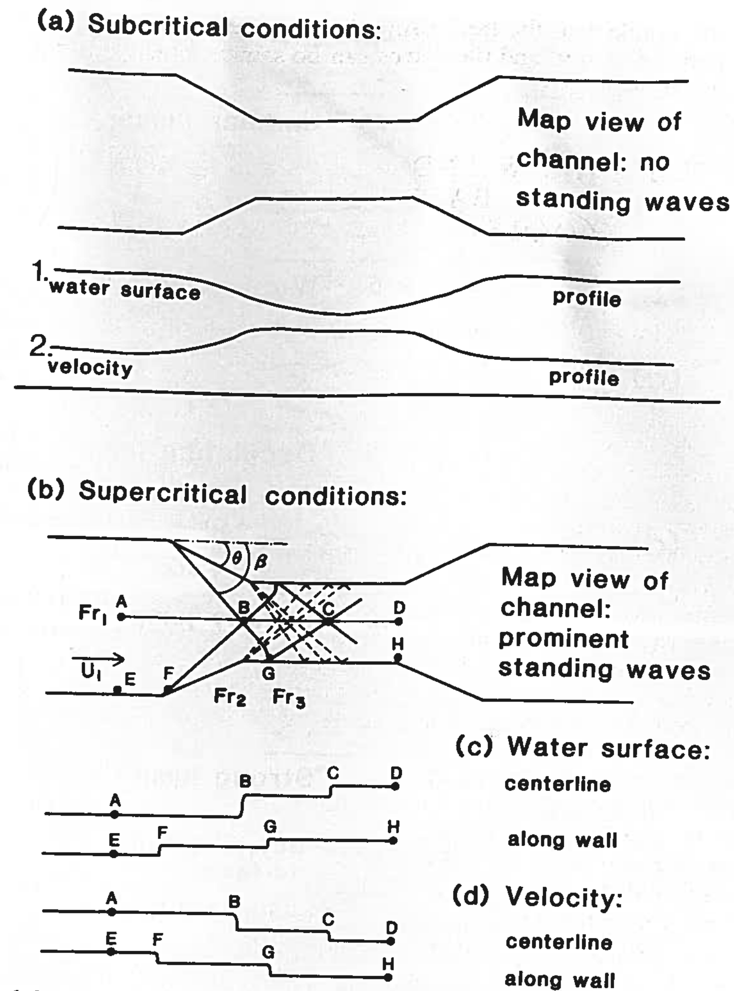


FIGURE 16.6. Comparison of the flow fields in subcritical and supercritical flow [compiled from Chow (1959) and other basis hydraulics text]. (a) Schematic map view of subcritical flow through a constriction. Schematic profiles (1) and (2) show how the water surface elevation and velocity change along the channel. (b) Schematic map view of supercritical flow through the same constriction. (c) Changes in water surface elevation and (d) local velocity through the channel, respectively. The profiles of these quantities are along the paths A-B-C-D and E-F-G-H.

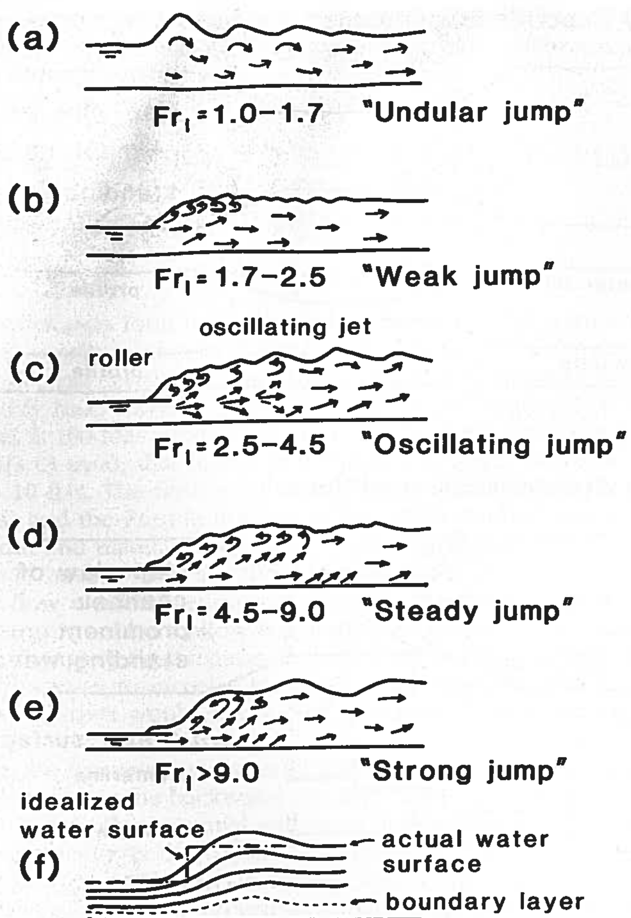


FIGURE 16.7. (a–e) Idealized cross sections of hydraulic jumps with the entering flow at different Froude numbers as shown. (From Chow 1959, p. 395.) (f) Schematic cross section of idealized and actual hydraulic jumps. (After Ippen 1951, p. 339.)

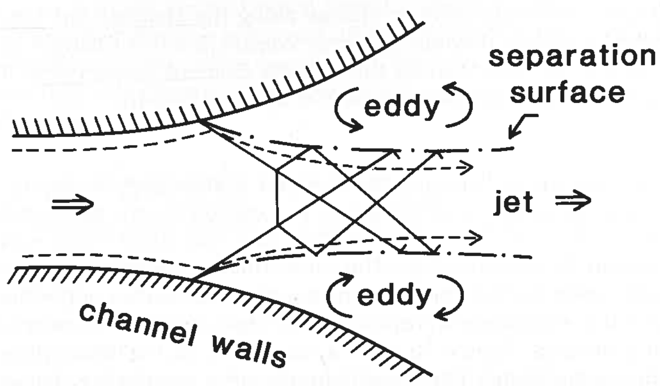


FIGURE 16.8. Illustration of the structure of a supercritical jet emerging from a constriction. (From Chow 1959, p. 471; originally from Homma and Shima 1952.)

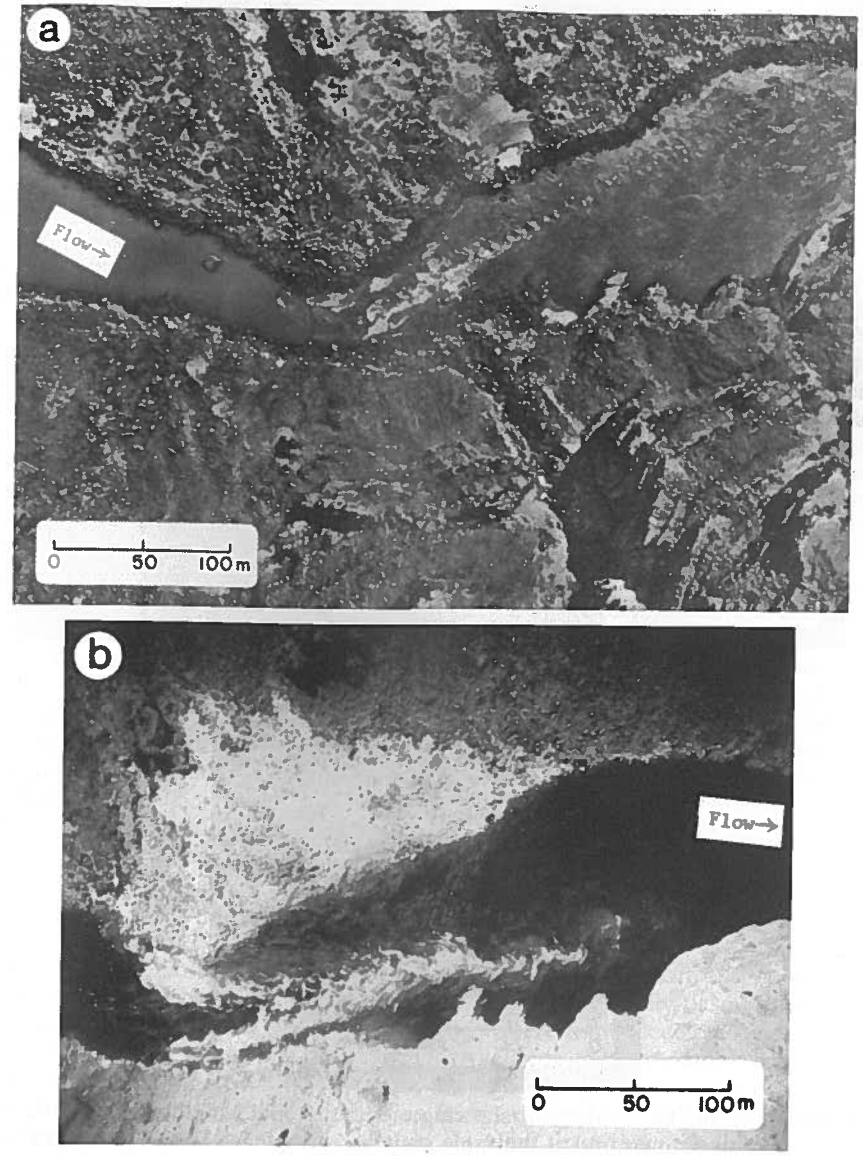
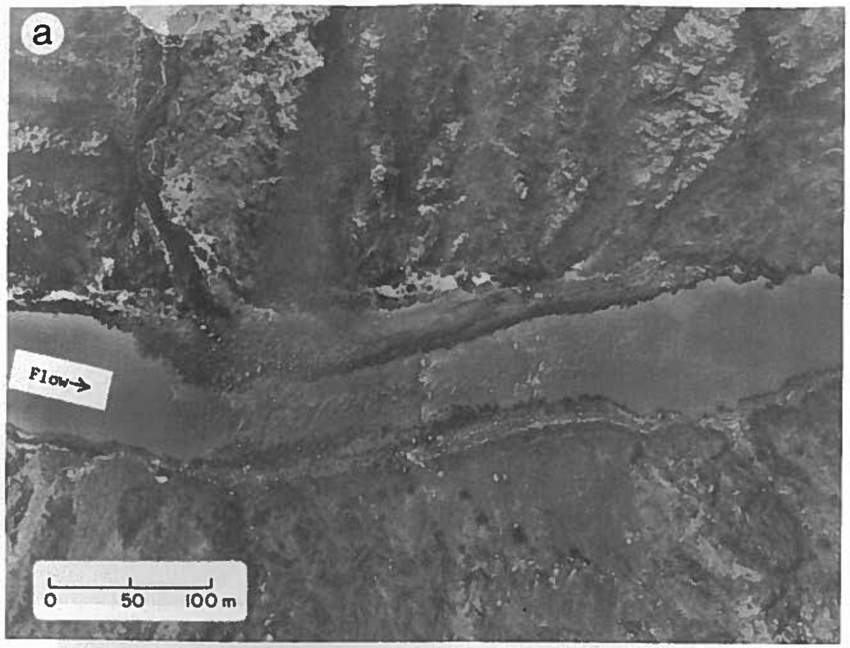


FIGURE 16.9. 24.5-Mile Rapids at discharges of (a) 5000 cfs and (b) 30,000 cfs. Note the dramatic change in the orientation of the tailwaves and the sizes of the eddies on each side of the tailwaves with discharge. [Photograph (a) by U.S. Bureau of Reclamation, 1984; photograph (b) by National Park Service, 1986.]

and constricted sections. Such waves also can occur in the diverging section but for simplicity are not shown here. The wave patterns (specified by the angle β) depend on the channel shape (specified by the angle ϕ). The height of the waves can vary across the channel—particularly where waves intersect each other. At the points of intersection, the wave heights can either be added to or subtracted from each other, depending on the type of wave.



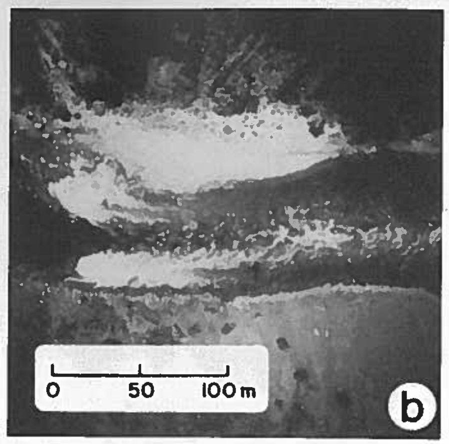


Figure 16.10. (a) Hermit Rapids, showing the tongue and lateral waves at 5000 cfs. (b) View of the same rapid at the same scale showing these features at 30,000 cfs. Note difference in tongue lengths, angle of lateral waves to shore, and nonbreaking rollers on the tongue. [Photograph (a) by U.S. Bureau of Reclamation, 1984; photograph (b) by National Park Service, 1986.]

Smooth water from the backwater extends farthest downstream into a rapid along the tongue in the converging part of the rapid (compare the tongue of Granite Rapids in Fig. 16.2 with the centerline A-B in Fig. 16.6b). A characteristic of most tongues is the presence of smooth, nonbreaking rollers (Fig. 16.1c). The length of the tongue and the angle of the oblique waves change with discharge and channel shape.

The converging part of a rapid is a region in which the flow changes in a complex way from subcritical conditions in the backwater to supercritical conditions in the constriction downstream; this region is often referred to as the "top" or "head" of the rapid. The measurements of water surface velocities and depths in the rapids suggest that the Froude number of the flow reaches a value of unity near the beginning of the tongue. For this reason, I interpret the nonbreaking rollers seen on most tongues as undular hydraulic jumps typically found when the Froude number is in the range of 1 to 1.7 (Fig. 16.7a).

In the supercritical region of the rapid (to the sides of and downstream of the tongue), complex waves appear. The stronger breaking waves (the oblique lateral waves and the haystacks) suggest higher Froude numbers than do the undular waves on the tongue. Values greater than 2 have been measured (Fig. 16.3e). The measured values of Froude number are probably lower limits for several reasons. The Froude number depends on water velocity, and, properly defined, it must be an average velocity for the whole water depth. When floats are used to measure velocities, several effects cause the measured velocities to be lower than the average velocity: drag between the water surface and air, drag between the float and air, and "surfing" and bouncing of the floats on the waves. It seems probable that the floats we used to measure surface velocities traveled appreciably slower (10–20%?) than the mean water velocity. Nevertheless, the measurements are internally consistent, and they show that the velocities and Froude numbers increase continuously from the backwater into the constriction and, sometimes, on into the diverging region of the rapid (Fig. 16.3d).

The water's energy is dissipated in the converging and constricted part of a rapid by wave action, bottom friction, and air entrainment. The noise of the rapids is part of the mechanism by which potential energy that was stored in the water in the backwater (and that was converted to kinetic energy in the converging section of the rapid) is dissipated as the water begins to decelerate back toward tailwater conditions.

The Diverging Section: Haystacks, Tailwaves, Eddies, and Beaches

Some very nonintuitive phenomena occur in the diverging part of a constricted rapid because the flow is supercritical. Our intuition, typically based on subcritical flow, would lead us to believe that the water would decelerate as soon as the river channel widens (Fig. 16.6a). Flow velocity measurements show, however, that the velocity can increase not only into the constriction, but well into the diverging part of the channel (this is not dramatic at House Rock Rapids shown in Fig. 16.3, but can be seen in the data from the other rapids in the maps of Kieffer 1988). The flow velocity and the Froude number both increase until a hydraulic jump is encountered—the hydraulic purpose of the jump is to return the flow toward ambient main-channel conditions appropriate to the downstream reach. This process is nonlinear; the flow does not adjust smoothly to the convergences and divergences of the channel. Instead, changes occur discontinuously through standing waves (hydraulic jumps). When sufficient energy has been dissipated by the mechanisms mentioned above, the water returns to subcritical conditions and must adjust its depth and velocity to the tailwater conditions downstream.

Another nonintuitive aspect of the flow in the rapids is the influence—or, rather, lack of influence—of the downstream tailwater conditions on the upstream flow conditions. In a subcritical constriction, the velocity and depth of water in the constriction and even upstream of the constriction are influenced by the depth in the downstream tailwater—that is, the flow adjusts continuously and smoothly to changes in channel geometry. In contrast, in supercritical flow, the tailwater conditions can only influence flow conditions in the lower part of the divergence below the hydraulic jump where flow returns to subcritical conditions. Flow in the converging and constricted parts of the rapid is uninfluenced by conditions downstream of the major waves.

Channel expansion typically is very abrupt in the diverging section (Fig. 16.2). When changes in channel shape are abrupt, flow streamlines generally cannot follow channel boundary curvatures (a condition known as rapidly varying flow in hydraulics). Because the streamlines cannot follow the channel boundary, the narrow high-speed jet formed in the constriction squirts into the divergent section with nearly the same diameter that it acquired in the constriction. The jet in the divergent reach is marked by a train of gently breaking to nonbreaking tailwaves (Fig. 16.2c). Measurements show that tailwaves generally are regions where the Froude number is very close to 1 (see Fig. 16.3d). Thus, we can interpret the tailwaves as nonbreaking, undular hydraulic jumps, and we note that they have a structure similar to, but not identical with, the nonbreaking rollers on the tongue in the converging part of the rapid where the Froude number was also estimated to be 1. A major difference between the tongue and the tailwaves is the width of the main current: in the converging part of the rapid, nearly the whole discharge is gradually funneled into the tongue, and the flow accelerates until $Fr \sim 1$ conditions are obtained. In contrast, in the diverging part of the rapid, the flow does not expand back to the full width of the channel.

The region between the jet and the channel boundary fills with recirculating flow of relatively low velocity—an eddy (Fig. 16.8). Eddies always are found on at least one side of a strongly diverging river channel below a rapid, and they often are found on both sides (Figs. 16.1b and 16.2c). These recirculating zones, in turn, allow sand deposition. For this reason, many important ecologic sites and pleasant camping sites are downstream of the rapids. Details of the expansion of the constricted jet back to ambient conditions in the tailwater are important in determining sediment transport to these beach sites (Fig. 16.2c).

The shape of the main jet of water emerging from the constriction can be defined quite well because separation surfaces between the jet and the recirculating zones are manifested as strong shear zones (eddy fences) and "boils" (small whirlpools). The shape of the jet in the tailwater cannot be predicted accurately from theory. However, laboratory data suggest that the distance downstream that the jet will maintain constant diameter is proportional to the diameter of the constriction in which it was formed. A jet typically will maintain constant diameter for a distance downstream of several constriction diameters. In detail, this relation depends also on the Froude number of the jet in the constriction and divergence. For example, with a Froude number of two, an ideal laboratory jet would maintain constant diameter for roughly three constriction diameters downstream. As an example, at House Rock Rapids, the constriction diameter is about 30 m when the discharge is 5000 cfs (Fig. 16.3a), and the jet maintains a strong identity (as evidenced by tailwaves) for at least 100 m downstream. Within this distance, the jet velocity appears to remain constant at 10 to 15 ft/s (3–5 m/s), and the Froude number appears to be near unity. The length and orientation of the jet and tailwaves can change dramatically with discharge (Fig. 16.9), a fact that has important implications for the formation and stability of beaches under different discharge conditions.

Much of the energy that must be dissipated in the rapid (the excess backwater head and the vertical elevation drop) is expended in the waves and in bottom friction. Nevertheless, the fact that the flow often has a high velocity at the bottom of the rapid indicates that not all of the water's excess energy has been dissipated. Additional dissipation occurs through mixing with the relatively stagnant water of the eddies that bound the jet in the tailwater. The motion of the jet induces circulation in the eddies, and the two flows (jet and eddy) mix in a zone that expands around the separation line (Landau and Lifschitz 1959, p. 131).

In the photographs and airbrush illustrations of jets in this chapter (Figs. 16.1, 16.2, 16.3, 16.9, 16.10), you can see the narrowing of the tailwaves with distance downstream. The converging lines that bound the tailwaves can be taken as evidence of the extent to which the mixing zone has extended into the main current (the jet). To a first approximation, one can say that the jet has decelerated to and been mixed back into main channel (or tailwater) conditions by the end of the tailwaves. The mixing not only decelerates the jet, but it allows suspended sediment carried in the high-velocity main channel current to be transported laterally toward the channel boundaries, into the recirculating eddy, and, ultimately, onto the beaches. Flow in the recirculating zone typically is very slow, about 1 f/s (0.3 m/s). Thus, sediment, if available, can be deposited within these zones (the rate of sedimentation will depend on particle size and density, fluid velocity and density, and other factors). Detailed studies of this zone are in Schmidt and Graf (1987).

Below the eddy-beach system, another pool and rapid sequence begins, and the hydraulics described here are repeated again, each time like a theme and variations, along the length of the Grand Canyon. Each rapid is unique, yet all are similar.

With this background, we can ask how the characteristic configuration of backwaters, rapids, and fast-water runouts has been created and evolved over thousands of years of hydraulic, hydrologic, and geomorphic interactions superimposed on the tectonic forces that drive the uplift of the Colorado Plateau and the downcutting of the Grand Canyon. A glimpse into these processes was offered by a series of unique events in 1983.

THE 1983 FLOODS: A UNIQUE WINDOW TO HYDRAULIC–GEOMORPHIC INTERACTIONS

In 1983, unusually high releases from Glen Canyon Dam, which controls the flow through the Grand Canyon, provided the opportunity to observe how channel geometry at a rapid changes as the discharge history of the river changes. Most of the interesting events were at Crystal Rapids (Fig. 16.11). This rapid was relatively insignificant before 1966 (for example, it was not mentioned in Powell's 1875 report). Figure 16.11a shows the configuration of the rapid in 1963, the year that Glen Canyon Dam was closed. Three years later, a large storm over the north rim of the Grand Canyon caused flash flooding in a number of tributaries (Cooley et al. 1977). In particular, debris poured down Crystal Creek and Bright Angel Creek, which is 10 miles (16 km) upstream from Crystal Creek.

The debris flow down Crystal Creek made Crystal Rapids one of the most dangerous rapids on the river for recreational boating. Large boulders carried in the debris flow tightly constricted the river channel at Crystal Rapids (Fig. 16.11b). Discharges at Glen Canyon Dam between 1966 and 1983 varied between 3000 and 35,000 cfs. These discharges were sufficient to cut a narrow channel through the distal (south) end of the debris flow and to cause occasional shifts of boul-

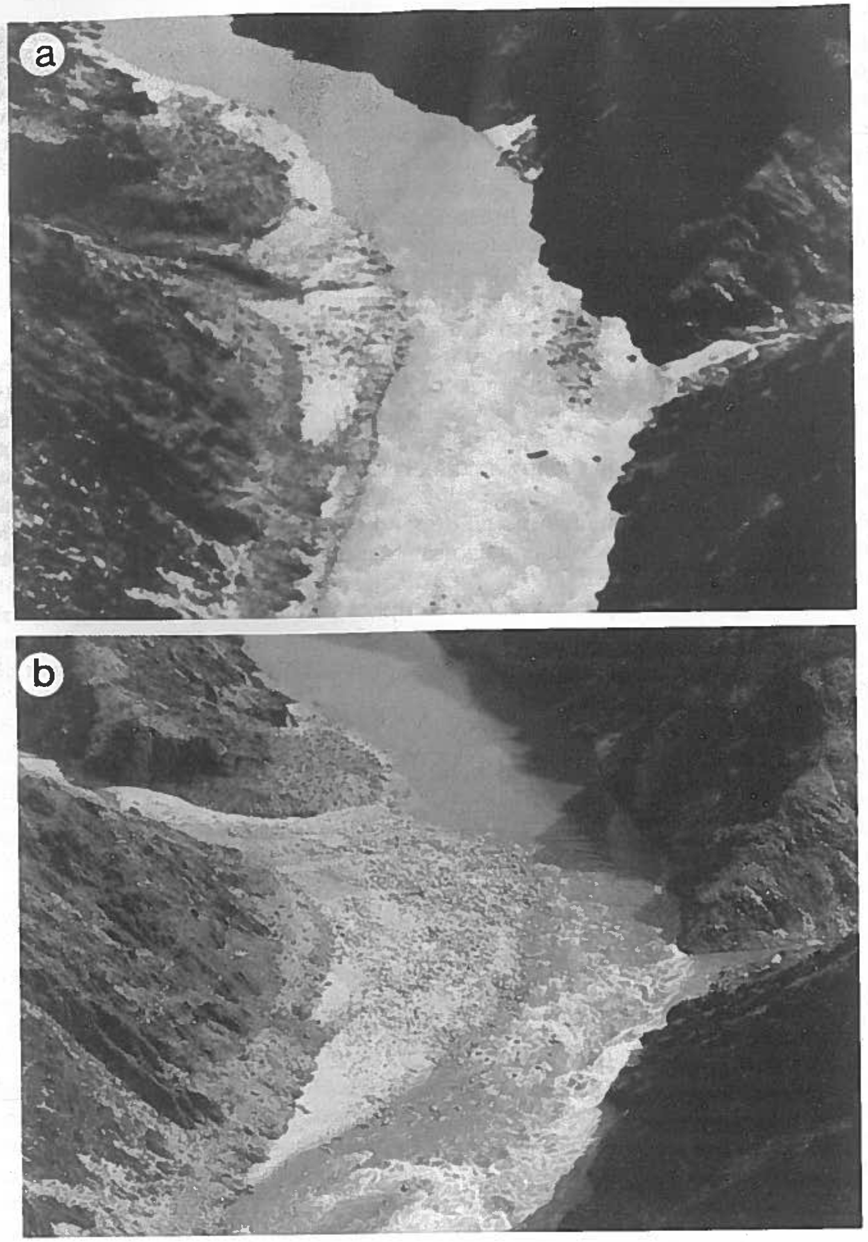


FIGURE 16.11. Crystal Rapids at three different times. (a) March 1963, discharge approximately 5000 to 6000 cfs. (Photographs by A.E. Turnber, Bureau of Reclamation.) (b) Same view, March 1967, approximately three months after a debris flow in Crystal Creek. Discharge is 16,000 cfs. (Photograph by Mel Davis, Bureau of Reclamation.) Note that even though the stage at 16,000 cfs is higher than at 6000 cfs, the debris fan is larger in (b). (c and d) Pair of aerial photos, showing the configuration of the debris fan and the rapid after the 1983 flood. Discharge approximately 5000 cfs in both photographs. The top photograph is from (a). About 10–15 m of lateral erosion has taken place along the shore adjacent to the rock indicated by the arrow. (Bottom photograph by Bureau of Reclamation, October 1984.)

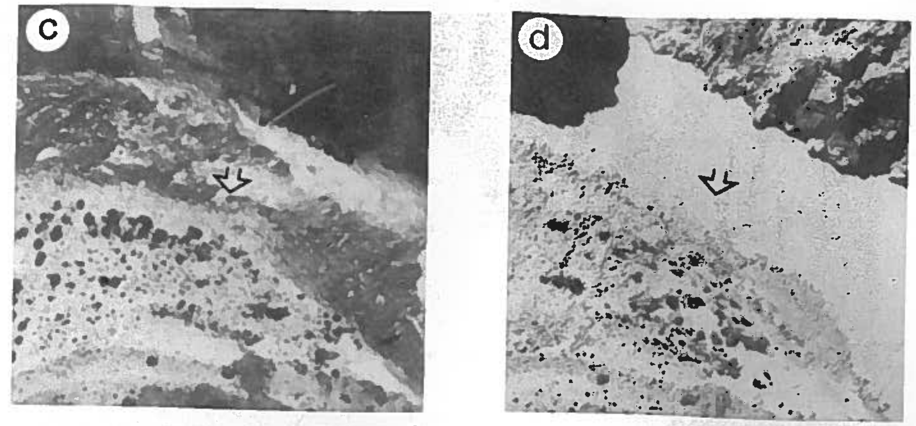


FIGURE 16.11. *(Continued)*

der positions within the channel. Overall, however, the channel at Crystal remained tightly constricted, and it retained the general shape documented by the 1967 photograph shown in Figure 16.11b. The shape parameter or "constriction" defined in Figure 16.3 (but modified as discussed in the text) was about 0.25 in 1973.

During the spring of 1983, rapid snowmelt at the headwaters of the Colorado River forced operators of Glen Canyon Dam to increase discharges to 92,000 cfs to keep Lake Powell from overtopping the spillways of the dam. This discharge was nearly three times larger than any discharge through Crystal Rapids since the 1966 debris flow, and it was comparable to the annual Colorado River floods prior to Glen Canyon Dam. Thus, a relatively young debris fan was subjected to its first "flood." The changes in Crystal Rapids (compare Fig. 16.11c with 16.11d) during these high flows provided an opportunity to observe some of the dynamic processes that contour the river channel over geologic time.

During the 1983 flood, the channel at Crystal Rapids was widened, and the shape parameter increased significantly from 0.25 to 0.4. This corresponds to an increase of 35 to 50 feet (10–15 m) in width, and this increase dramatically altered the hydraulic characteristics of the rapid. Nevertheless, the rapid is still quite different from the other rapids that have constrictions of 0.5 (remember from Fig. 16.3 that most rapids have constrictions of 0.5). Local gradients within Crystal Rapids are steeper than those generally found at other rapids, waves are larger, and the dependence of wave structure on discharge is more variable. By watching the evolution of the rapid toward the configuration of the more mature rapids, I have worked out the following ideas on the hydraulic–geomorphic interactions between the Colorado River and the debris dams that episodically block its course.

EROSION OF THE DEBRIS FANS BY THE RIVER

The boulder deposits that constrict the river to form rapids are emplaced by debris flows from steep tributary canyons. The steep gradient of these canyons allows these streams to carry large boulders into the main channel (which has a much smaller gradient than the tributaries). These boulders cannot be moved by typical main-stem flows in unconstricted reaches of the river. Although the ini-

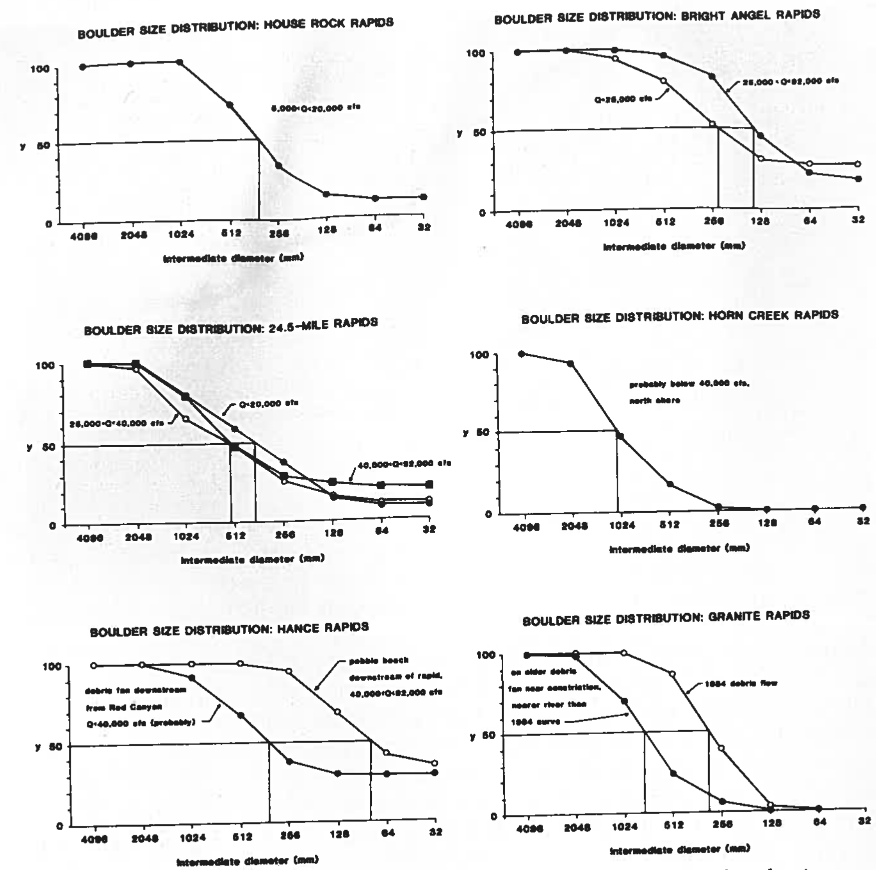


FIGURE 16.12. (Part I opposite page, Part II above) Particle size distributions measured at the places indicated at rapids (Kieffer 1987). The twelve graphs are arranged in downstream order. The ordinate, *y*, is the percent of particles smaller than the given (intermediate) diameter. The horizontal and vertical lines in each graph are to guide the reader's eye to the median diameter of the particles at the rapid.

tial size distribution of material in the debris flows is quite broad (Webb et al. 1987), the smaller material, up to large cobble size, is washed away by the river at relatively low rates of discharge. Excellent descriptions of sediment transport through the canyon can be found in Howard and Dolan (1976, 1981), and the U.S. Bureau of Reclamation Glen Canyon Environmental Studies (1987) contains recent detailed studies of sand and silt transport in the main channel. Boulders 3 to 10 feet (1–3 m) in diameter are common on the debris fans (Fig. 16.12). These boulders resist erosion (either by chemical or mechanical abrasion processes or by movement at low discharges), and so they stabilize the debris fans with the geometry shown in Figure 16.3. Because at least half of the channel generally is cleared of all but the very largest boulders at mature rapids, it is obvious that the boulders can be moved under some conditions.

Although hydrologists have documented the transport of sand and silt-sized sediment past the Lees Ferry and Grand Canyon (Bright Angel) gage stations, we know very little about the mobility of large particles in the vicinity of rapids.

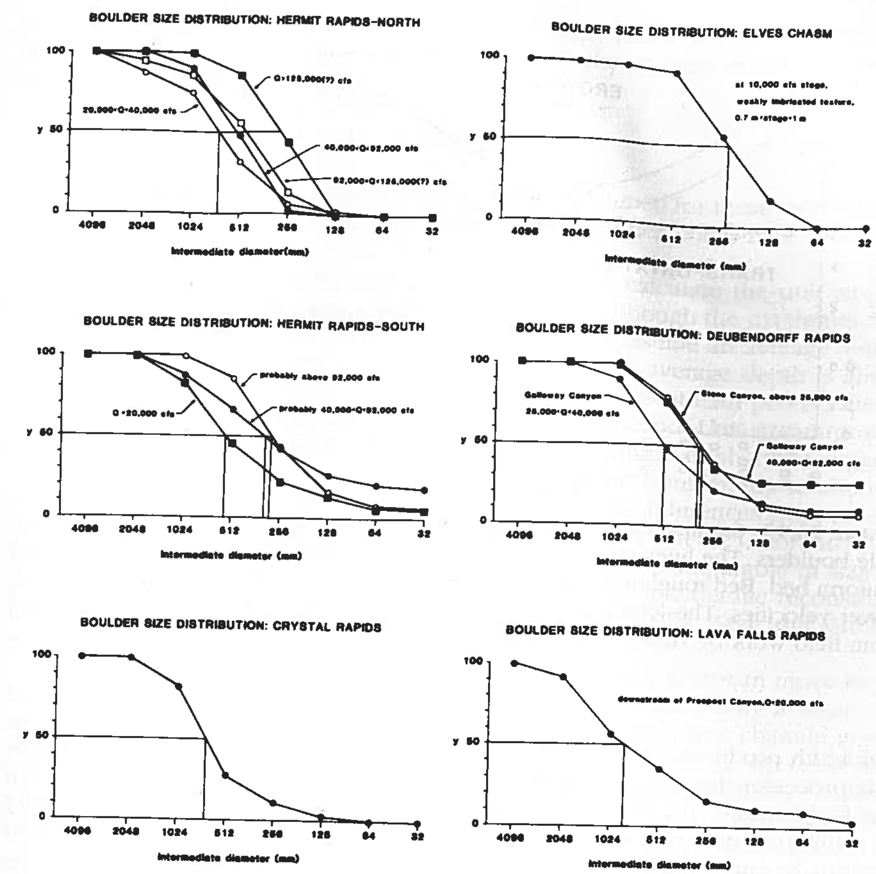


FIGURE 16.12. (*Continued*)

Specifically, quantitative modeling of the movement of boulders in the Grand Canyon has not been done. Laboratory experiments on much smaller particles and theoretical studies have shown that the size and amount of material transported are related to water velocity, depth, and, therefore, rate of discharge. Graf (1979, 1980) analyzed the stability of boulders in the Green River in Utah and concluded that the largest boulders were stable, even against motion during the largest floods. Because the Colorado has many features in common with the Green River, it generally has been assumed that large boulders, once emplaced, also are relatively stable in the rapids in the Grand Canyon.

The ability of the river to clear out debris emplaced in the channel (its "competence") is proportional to variations in velocity and water depth. The discussion in the first half of this chapter demonstrated that changes in these parameters of more than an order of magnitude occur within a rapid. There are, therefore, substantial differences in transport capacity from one section of a rapid to another. The photographic documentation of changes in Crystal Rapids shown in Figure 16.11 and the statistical analysis of rapid shapes shown in Figure 16.4 show that the water can move large boulders in the channel—at least until the river has cleared itself to about one-half of the characteristic unconstricted width. This capacity of the river to clear out large debris from at least half of the chan-

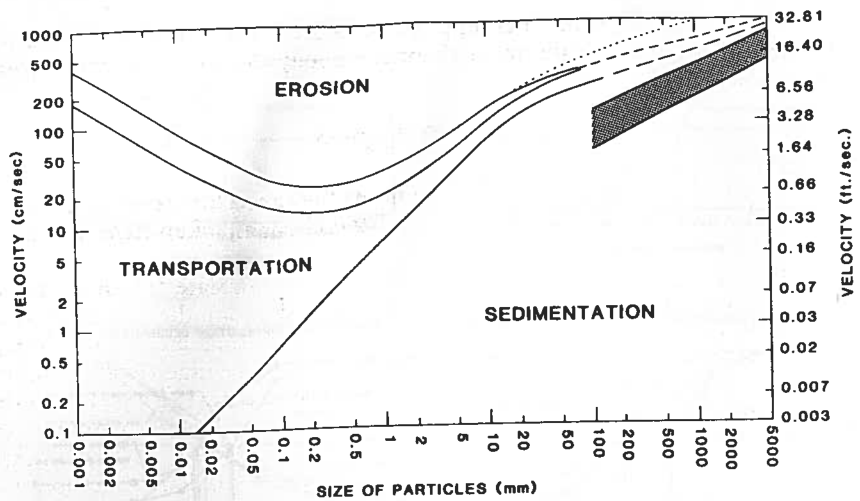


FIGURE 16.13. Summary of the relations between stream velocity and size of movable boulders. The lines on top represent the Hjülstrom criterion is for particles in a uniform bed. Bed roughness and particle shape appear to cause particles to move at lower velocities. The lower stippled area to the right represents criteria developed from field work by Helley (1969).

nel width produces the characteristic "nozzle-like" geometry seen at the rapids. The processes that accomplish this erosion involve a complex feedback between the hydraulics of the river (discharge as a function of time; depth and velocity as a function of position in the river) and the nature of the material in the bed (grain size and position in a bed of variable cross section).

One criterion widely used for predicting the transport of smaller material in a river is the Hjülstrom relation (Hjülstrom 1935; Strand 1986), which relates water velocity to the size of the largest particles that can be transported (Fig. 16.13). This Hjülstrom relation, extrapolated to large boulder sizes in Figure 16.12, suggests that a velocity of 20 f/s (6 m/s) would be capable of moving a boulder 1 to 2 ft (0.5 m) in diameter and that it would be possible to move material up to 3 to 7 ft (1–2 m) in diameter. These values depend on how the Hjülstrom curves are extrapolated.

As discussed in the previous section, water in the highest-velocity part of the rapids can have velocities exceeding 30 ft/s (10 m/s) even at the lower end of the range of discharges from Glen Canyon Dam generators (about 3000 cfs). By the Hjülstrom criterion, flows at these discharges should be capable, therefore, of clearing the channel of boulders in the 3- to 7-ft (1- to 2-m) size range.

A second criterion for boulder transport is the concept of unit stream power, originally introduced by Bagnold (1966, 1980) and recently applied to paleohydrogeologic problems by O'Conner et al. (1986). Unit stream power is the stream power (rate of expenditure) per unit area, Ω. Bagnold (1966) defined it as

$$\Omega = xQS_f/w = \beta u$$

where x is the specific weight of the fluid (assumed to be 9800 N/m^3 for clear water), Q is the discharge (specifically, that component of the discharge carried by the main current), S_f is the friction slope, w is the channel width, β is the total channel shear, and u is the main channel velocity. A more convenient form of this equation is

$$\Omega = xn^2u^3/R^{1/3}$$

where n is the Manning coefficient of roughness (assumed for these calculations to be 0.035) and R is the hydraulic radius of the channel, taken here to be the average depth.

For example, the reader can use Figure 16.3 to calculate the unit stream power in House Rock Rapids at 5000 cfs discharge. Although the maximum water surface velocity reaches 25 ft/s (7.5 m/s), we will assume an average velocity of 21 ft/s (6.5 m/s) for a conservative estimate. The average depth is about 3 feet (1 m) in the narrowest part of the rapid. The unit stream power, therefore, is on the order of 3300 Newtons per meter per second. Unit stream power and sediment transport relations (summarized in O'Conner et al. 1986) suggest that a river with this unit stream power could transport boulders up to about 7 feet (2 m) in diameter. This conclusion agrees with the inferences from an extrapolation of the Hjülstrom diagram. Both criteria suggest that the Colorado River within the Grand Canyon is capable of moving boulders comparable in size to those moved during the largest floods known from paleohydraulic reconstruction techniques (for example, the Missoula flood, Baker 1973, 1984; and Katherine Gorge floods in Australia, Baker and Pickup 1987).

These two criteria also lead to the conclusion that the river can move large boulders even at low discharges in which the flow occupies only a relatively small part of the total river channel (note how little of the river channel is occupied by water at the 5000 cfs discharge conditions shown in the cross sections of Fig. 16.3). Another way to state this is to say that the relatively shallow, high-velocity flow characteristic of high-gradient rivers can move rather large boulders. Thus, that part of the channel of Crystal Rapids that was exposed to discharges up to 35,000 cfs was depleted of small to relatively large material prior to the 1983 flood. During larger floods, the deepest part of the main channel is cleared of even larger boulders. However, even the parts of the channel exposed to relatively shallow overflow during the floods have boulders removed, as shown by the size distributions of debris fan material that has been exposed to main-stem floods (Fig. 16.12). Preliminary measurements by the author of boulder size distributions above the river shoreline that correspond to the 92,000 cfs discharge of 1983 showed depletion of small boulders, suggesting larger floods in the past. These observations suggest that a detailed study of boulder size distributions at different places on the debris fans and talus slopes along the Colorado could be used to infer flood histories.

There are, however, serious complications in interpreting boulder size distributions in terms of simple erosion by the main stem because each debris fan has a complicated erosional and depositional record. If a debris fan had a wide variety of particle sizes when it was emplaced, and if it was not graded in size laterally, then erosion of this fan by a large flood would be expected to remove larger particles low on the fan (where the flow is the deepest and fastest) and to remove progressively smaller particles higher on the fan. The size distributions measured on the north and south banks of Galloway Canyon at Deubendorff Rapids (Fig. 16.12) are consistent with such a simple emplacement and erosion model. However, material removed from the upstream parts of the fan may be deposited on the downstream sides in the recirculating zones discussed in the first part of this chapter; thus, the size distribution depends not only on the vertical elevation on the fan, but on the relative upstream–downstream location (for example, Hance Rapids, Fig. 16.12).

Boulder size distributions on other debris fans are rarely this simple to interpret. In addition to the erosional and depositional processes mentioned above, complexities arise because (a) initial particle size distributions in the debris flow are not known; (b) winnowing or piping of fine particles may be important; and (c) the initial history of the damming and breaching of the fan generally is unknown. For example, photographs of the Crystal Creek debris fan taken in 1967 (Fig. 16.11b) show that it had a surface veneer of boulders nearly all of the way up to the mouth of Crystal Creek—far above any stage reached by Glen Canyon Dam discharges between 1966 and 1967. Because of this absence of known floods prior to the time the 1967 photograph was taken, I speculate that the 1966 debris dam caused ponding of water to this level and that the overflow of this original pond removed a substantial number of boulders all the way to the top of the debris fan. In Kieffer (1987) I discuss in detail the implications of each size distribution shown in Figure 16.12.

These many erosional and depositional processes are important in the details of the channel shape and particle size distribution where the river cuts through a debris fan to form a rapids. The geometry of the channel is the result of a balance between the local increase in the river's erosive power within the constricted zone and its decreased transport capacity in slower parts of the channel. Water accelerates from velocities on the order of a fraction of a foot per second (a few tenths of a meter per second) in the backwater above a rapid to 15 to 25 ft/s (4 to 6 m/s) on the tongue of the rapid, to values greater than 30 ft/s (10 m/s) in the constriction and part of the diverging section. Velocities of about 15 ft/s (4 m/s) then are maintained through the tailwaves. Thus, in the constriction and the diverging section of the rapid immediately below the constriction, velocities are more than adequate to clear the channel of boulders.

The position of the constriction in a rapid can change as the rate of discharge changes because of variations in topography. The position and shape of the constriction also changes with time as the river channel evolves in response to floods. An excellent example of how the position of a constriction depends on discharge can be seen at Horn Creek Rapids. At low discharges (less than 30,000 cfs), the constriction is near the top of Horn Creek Rapids. At 92,000 cfs, this constriction is nearly drowned out by the backwater from a constriction about 1000 feet (several hundred meters) downstream (Kieffer, 1988, I-map 1897-E).

The position and shape of a constriction in a rapid also changes as a result of tributary flash floods, as at Crystal and Bright Angel Rapids in 1966. In 1983, the position of constriction at Crystal Rapids again changed, moving upstream in response to high discharges. These changes are examples of why observations of rapids over a large range of discharges and, by implication, a long period of time are necessary for compiling data on channel processes and their rates.

Because there is an increase in erosive power in the constriction and in the diverging section of a rapid, boulders move downstream until the channel widens and deepens sufficiently for the water to decelerate (typically by a transition from supercritical to subcritical flow). Large boulders are moved out of the constrictions of rapids, transported through all or part of the divergent sections, and then deposited hundreds of feet (in some instances, up to 1 km) downstream to form the "rock gardens" or cobble bars found below rapids (Fig. 16.2a). A rapid, therefore, evolves into two parts: the original debris deposit (reworked, at least on the surface, by overflow) and the rock garden (or cobble bar), usually found downstream of the initial deposits.

Crystal Rapids, 1966–1987

In 1983, we were able to document changes in the hydraulic behavior and channel shape at Crystal Rapids; the following summary is taken from Kieffer (1985). The rapids in the Grand Canyon were exposed to discharge levels three times that which had occurred since 1963, and this was a particularly significant event for Crystal Rapids because of the large debris fan emplaced in 1966. In addition to their geomorphic significance, the hydraulic events during 1983 had a significant effect on commercial and private rafting in the Grand Canyon. About 10,000 people each year navigate the 250-mile (400-km) stretch through the Grand Canyon. The debris flow in 1966 and the flood of 1983 both emplaced boulders and caused waves and eddies in Crystal Rapids that have made this area difficult to navigate.

Although the geometry of the river channel prior to the studies of Kieffer (1988) is largely unknown, studies of air photographs taken in 1973 and calculations of plausible cross-section shapes (Fig. 16.14) suggest that the constriction of the channel was about 0.25—that is, the surface width of the rapid in its narrowest part was only 1/4 of the width of the main channel upstream of the rapid. The water-surface elevation dropped about eight feet (2.5 m) between the head of the rapid and a large obstacle known as "The Crystal Hole" that was several hundred meters downstream from the entrance to the rapid (locales are shown on Fig. 16.14). The water surface dropped about another 8 feet (2.5 m) through the rock garden below the rapid (Leopold, personal communication, 1984). Because the Crystal Creek flood of 1966 appears to have been large enough to have strewn boulders across the entire river channel, it appears that either the initial breaching event (when the river cut through this debris dam) or power plant releases up to 35,000 cfs from Glen Canyon Dam between 1966 and 1973 were sufficient to move rocks out of the distal end of the debris flow into the rock garden. The result was that by the time the 1973 air photographs on which the data of Figure 16.3 were based, the river had carved a channel with the value of constriction equal to 0.25.

A significant clue to the channel hydraulics in 1983 was the comparison of the waves in the mature rapids with the behavior of the wave that occupied the region of the Old Crystal Hole when discharges in late June and early July 1983 increased above earlier maximum discharge levels. Waves in the more mature, less constricted rapids than Crystal disappeared at discharges on the order of 90,000 cfs, unless the topography covered by the water at higher stages formed a different constriction in three dimensions. This phenomenon of the disappearance of local waves within rapids is the so-called "drowning out" of a rapid at high water. Instead of drowning out, the wave associated with the Old Crystal Hole in Crystal Rapids increased in height as discharges increased. At discharges in the range 50,000 to 70,000 cfs, this wave formed a formidable barrier across the river: it was more than 15 to 20 feet in height (5–6 m) and spanned nearly the entire navigable channel (Fig. 16.15).

The size, location, and even the sound of this wave changed with discharge. In the years prior to the 1983 flood, the trough-to-crest height had been about 10 feet (3 m) at 20,000 cfs and about 3 feet (1 m) higher at 30,000 cfs. Between 1966 and 1983, the wave was associated with a large rock in this location rather than with its critical position in the neck of the constriction. As discharges rose to around 50,000 to 60,000 cfs in June 1983, boatmen and passengers reported that the wave surged to a height between 15 and 30 feet (5 and 9 m); it was verified photographically at about 15 to 20 feet (5–6 m) (Fig. 15). Perhaps most interestingly, as discharges reached 92,000 cfs in early July, river observers noted

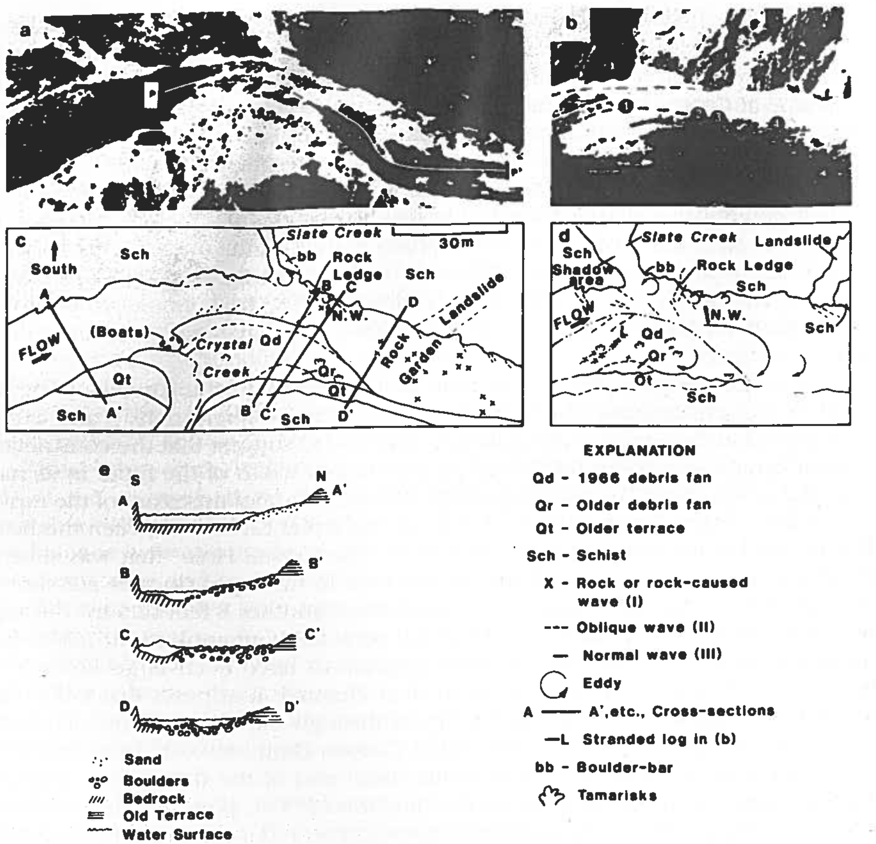


Figure 16.14. (a) Crystal Rapids on June 16, 1973. (U.S. Geological Survey Water Resources Division air photograph.) (b) View of part of the rapid at approximately the same scale during the high flow of 1983. (c) and (d) Keys to features on (a) and (b). (e) Schematic cross sections along lines A-A′, B-B′, C-C′, and D-D′. Relative widths correct; vertical scale exaggerated. Rise of debris fan from the river level in (a) to the old alluvial terrace indicated by Qt is about 5.5 m. The stage at 92,000 cfs is just at the base of this terrace. Assumed boundaries for deep channel are shown by a light dashed line in (a) and (c). P-P′ was the preferred navigation route through the rapids prior to 1983. The Crystal Hole, formed by a normal wave (hydraulic jump), is indicated by N.W. This wave is not easily visible in (a) because of photo scale. The white line in (b) indicates the path of kayaks whose velocities were measured at approximately 9 m/s.

that the wave height decreased to 10 to 15 feet (3–4.5 m). At discharges over 50,000 cfs, the wave appeared to be located about 100 feet (30 m) downstream from its pre-1983 position at 30,000 cfs (compare the position of the wave, labelled N.W. in Figs. 16.14c and 16.14d). Observers reported that at 50,000 to 60,000 cfs the wave emitted a low roar like a jet engine, but it did not generate the same loud roar at 92,000 cfs, though loud booms were clearly audible every few seconds.

After the 92,000 cfs discharges of 1983, surface wave patterns within the rapid altered dramatically, and the local gradient within the rapid changed (Fig. 16.1c).

Figure 16.15. Photograph of the wave, interpreted as a normal hydraulic jump, that formed across much of the main channel in late June 1983, when discharges were raised to about 60,000 cfs. (Photograph taken June 25, 1983; copyrighted by Richard Kocim, reprinted with permission.) Pontoons on the raft are each 1 m in diameter; midsection is about 3 m in diameter. More than 30 passengers were on board; one head is visible on the lower side of the raft. From the scale of the raft, the trough-to-crest height of this wave can be estimated at more than 5–6 m.

This new hydraulic situation has persisted. The drop in elevation of 5 to 10 feet (2–3 m) that was spread between the top of the rapid and the old Crystal Hole is concentrated now in a narrow zone of only a few tens of meters near the top of the rapid. As a result, the oblique waves on the right side of the tongue at the entrance to the rapid have increased dramatically in height. The change in bed slope and water-surface gradient at the head of the rapid suggests that about 30 feet (100 m) of headward erosion occurred during the high discharges.

In addition, the channel widened by 30 to 50 feet (10–15 m) at its narrowest point during this flood (compare Figs. 16.11c and 16.11d). This means that the constriction value changed to 0.40—approaching the 0.5 value typical of the older, more mature debris fans. Widening occurred in and downstream of the constriction in the zone of supercritical flow. Widening did not occur solely during peak flows, but it appears to have begun as soon as the discharges exceeded the controlled flows of the prior two decades. This conclusion is unsupported by direct measurement. However, a very similar series of events—including a debris flow in 1966—occurred at Bright Angel Creek, 10 miles (16 km) upstream from Crystal Rapids and in a similar geologic setting. There, as the discharge rose through the range of 60,000 to 70,000 cfs, the bed at Bright Angel gage station was scoured by about 8 feet (2.4 m). Presumably, a scour of similar magnitude occurred at the same discharges at Crystal because of the similarity in channel morphology and material at the two locations.

Water velocity varies throughout the length of a rapid because of geometry and gradient changes. Unfortunately, no systematic measurements of water velocity could be made at Crystal during the flood; however, on June 27, when the discharge peaked at 92,000 cfs, kayaks were filmed going through the rapid (approximately along the white line shown in Fig. 16.14a). Analysis of the films showed maximum velocities of 28 to 32 ft/s (8.5–9.8 m/s). Although there can be no rigorous correlation of these velocities with average water velocity, the kayaks appeared to be moving with the current. If the average water velocity was even close to these values, the river was capable of moving boulders several meters in diameter (Fig. 16.13). Large, moving boulders presumably were the source of the loud, booming noises heard by the author.

The velocities measured, the depths inferred from measurements of stage, and the behavior of the large wave that developed in Crystal Rapids all indicated conditions of supercritical flow during the flood. The flow was forced into supercritical conditions by the geometry of the converging–diverging channel. The large wave that stood across the channel had characteristics generally associated with a "normal hydraulic jump." Shallow-water flow theory allowed me to analyze the relation between discharge and backwater energy—conventionally expressed as depth (Fig. 16.16a), wave height (Fig. 16.16b), velocity in the constriction (Fig. 16.16d), velocity in the diverging supercritical section of the rapid and in the diverging subcritical section below the hydraulic jump (Fig. 16.16c), and the change in water velocity through the hydraulic jump (Fig. 16.16e).

Figure 16.16b shows that the measured wave height increased with discharge until the discharge reached about 60,000 cfs; the calculations indicate that this behavior would be expected for a channel with a pre-flood constriction of 0.25. At higher discharges, the model calculations suggest that the wave height should have continued to increase, but, instead, it decreased. The curves in Figure 16.16b suggest that changes in the shape parameter caused the decrease in wave height. In particular, it appears that by the time discharges reached about 60,000 cfs, the channel had begun widening and that by the time of peak discharges, the constriction had changed to about 0.40.

Although the parts of Figure 16.16 look complicated, a careful study of these figures reveals the complex interactions that were occurring. The meteorological situation forced the engineers to increase the discharges constantly, causing water velocities to increase. Channel widening occurred as a result of erosion at the high velocities. The channel widening alone would have allowed a decrease in velocity, but the increasing water level of Lake Powell necessitated further increases in discharge.

During this flood event, the Colorado River continued to contour the channel at Crystal Rapids into a shape in which the velocities in the most highly constricted portion of the channel were equal to the threshold velocity for the transport of the major boulders. Material removed from the constriction was transported several hundred meters downstream into the area of the rock garden. This section of the river was modified substantially by the 1983 flood.

For nearly a year following the peak flows in June and July of 1983, high discharges prevented direct observations of the effects of the flood. By the time that Crystal could be examined again at low water (October 1984), the record of the 1983 events was partially overprinted by sustained discharges at 60,000 and 25,000 cfs. Nevertheless, field evidence indicated that the erosion postulated on the basis of the open-channel hydraulics theory did indeed occur (Figs. 16.11c and 16.11d).

Observers found that the eroded section of the channel at Crystal Rapids had a fresh cutbank in the boulders. Similar cutbanks have been observed after the 1984 debris flows down Monument Creek at Granite Rapids and at Elves Chasm in 1984. These cutbanks are evidence of the action of the river contouring its own channel, and if they can be related to specific discharges, they provide valuable clues about the geomorphic evolution of the debris fans.

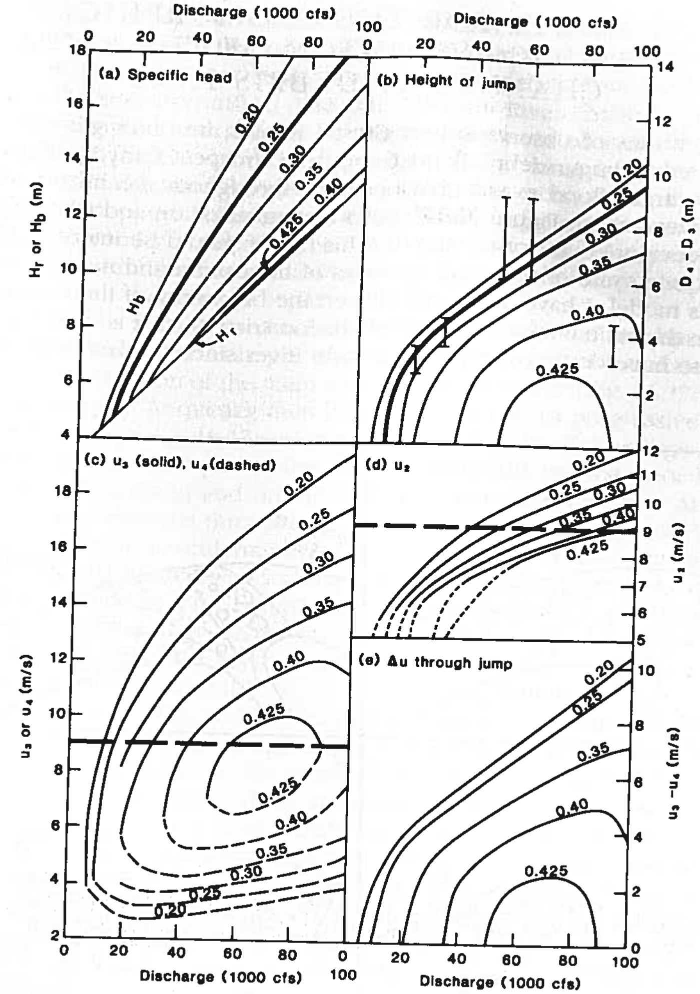


Figure 16.16. Model calculations for flow through Crystal Rapids (from Kieffer 1985). (a) Specific head (Hr) measured at Bright Angel Creek vs. discharge, with backwater heads (Hb) calculated for Crystal Rapids for the constrictions, w2/w0 (0.20, 0.25, . . .) indicated. The specific head can be thought of as the normal depth of the river, and the backwater head as the depth cause by the constriction. (b) Calculated height of the hydraulic jump for constrictions indicated. Bars denote observed values. (c) Calculated values of flow in the supercritical region of the diverging section of the rapid (u3) and in the subcritical region of the diverging section (u4). These two sections are separated by the hydraulic jump. Dashed line at 9 m/s indicates the velocity at which larger boulders at Crystal Rapids were assumed to be transportable by the current. (d) Calculated values of velocity in region 2 (the constriction) for constrictions indicated. Flow subcritical where dashed. (e) Velocity change through hydraulic jump that separated regions 3 and 4.

A MODEL FOR THE GEOMORPHIC-HYDRAULIC EVOLUTION OF THE RIVER CHANNEL AT DEBRIS FANS

The two decades of observations at Crystal Rapids are but a glimpse into the history of much larger debris flows (e.g., from Prospect Canyon at Lava Falls) and much larger flood events that occurred throughout the history of Grand Canyon downcutting. Figure 16.17 shows a generalization and extrapolation of the ideas developed at Crystal Rapids. This figure should be interpreted to represent but one cycle of recurring episodes of deposition and modification.

In this model, I have arbitrarily chosen the beginning of the sequence as a time when the main channel is relatively unconstricted—that is, major floods are assumed to have occurred on the Colorado River since the last time this tributary canyon had a major debris flow (see Fig. 16.17a). Between Figures 16.17a and 16.17b, it is assumed that unusual weather, climate, or an accumulation of debris within the catchment basin caused a large debris flow to emerge suddenly from the tributary and to dam the Colorado River channel. The flow in the river is disrupted and ponded by the emplacement of a debris dam (Fig. 16.17b). A "lake" forms behind the debris dam, and at some time, depending on the height of the dam and the discharge in the main stem, a waterfall forms as this dam is overtopped. Schuster and Costa (1986) suggest that the dam would be breached (typically at the distal end of the debris fan) nearly instantaneously (hours to days)—and perhaps catastrophically. The evolution of a rapid from a lake and waterfall then would begin.

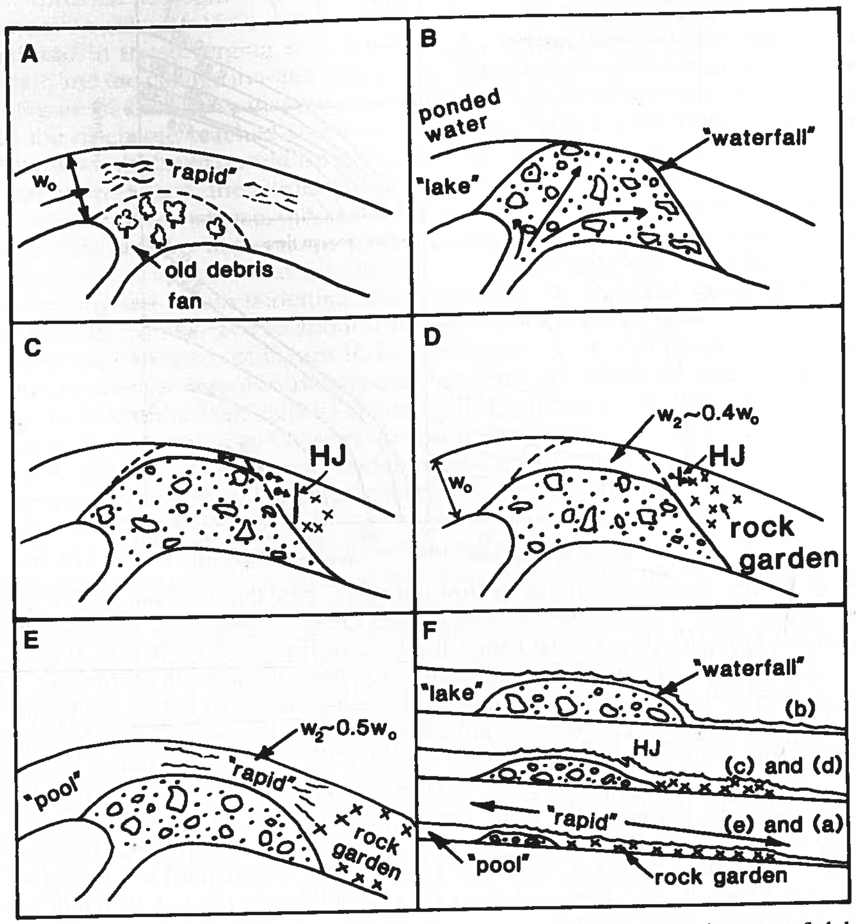


FIGURE 16.17. Schematic illustration of the emplacement and modification of debris fans, the formation and evolution of rapids, and the formation of rock gardens. (a) Initial channel geometry. (b) Side-canyon flood. (c) Erosion: small floods. (d) Erosion: moderate flood, supercritical flow. (e) Erosion: large flood, subcritical flow. (f) Longitudinal cross sections. See text for further explanation.

Unless there is a major breach in the debris dam with the first breakthrough of the ponded water (for example, more than 50 percent of the material is removed), the constriction of the main stem initially is severe (Fig. 16.17b). Floods of differing size and frequency then erode the channel to progressively greater widths (as shown in Fig. 16.17c–e). Small floods (Fig. 16.17c) enlarge the channel slightly (again, a nonquantitative term that depends on the prior history of the fan during breaching and on the material composing the fan). At first, the depletion of fine material through the debris fan may undermine the positions of large boulders. This undermining can cause rather dramatic changes in the scale of individual boulders and waves within the rapid. Indeed, it may be the reason that changes in wave behavior in Crystal Rapids seemed rather frequent in the late 1960s and early 1970s. As discussed above, the movement of a single, large boulder can cause dramatic changes in local wave structure in a rapid within regions of supercritical flow.

Moderate floods (Fig. 16.17d) enlarge and widen the channel further. The channel may become wide enough that at low discharges the flow is weakly supercritical or even subcritical. For example, the 1983 discharges of 92,000 cfs at Canyon Rapids (considered high by standards of Glen Canyon Dam discharges) widened the constriction at Crystal sufficiently that the strength of the Old Crystal Hole was diminished during subsequent lower flows at which it previously had been a substantial obstacle. On the other hand, the lateral waves became stronger because of headward erosion, which created additional potential energy at the top of the rapid. These waves, rather than the Old Crystal Hole, now form the main rafting hazard, though the hole is still a strong wave at discharges above about 30,000 cfs. Lateral widening, vertical scouring, and headwall erosion of the underwater debris dam occur simultaneously (Fig. 16.17f). Large floods further widen the channel at the debris fans and erode upstream through the fans (Fig. 16.17e). This cycle (a–e) can be repeated over and over through geologic time as floods in tributaries and on the main stream occur.

In summary, the local geometry of the river channel is subject to change if a substantially larger discharge is put through a rapid. This discharge can occur from flooding caused by meteorological events, from flooding caused by local emplacement and the breaching of natural dams in the river (debris of laval flows), or, now that Glen Canyon Dam is in place, from operational procedures at the dam. When the local gradient in the channel changes from such flood events, new waves can arise and old waves can become small or disappear.

The Rapids: Past, Present, and Future

The shape of the river channel at a debris fan at any instant of geologic time reflects contouring by different flood events—including any flood event that may have accompanied the emplacement and breaching of the debris dam. Even discharges as low as several thousand cfs appear to have sufficient velocity to clear the channel of large boulders, though such discharges are typically only a few percent of the total channel cross-sectional area. Fine-grained, transient sediment may partially mask the larger scale erosion (e.g., Howard and Dolan 1981, Fig. 7).

We now return to the questions raised earlier in this chapter. "Why is the shape of the channel, i.e., nozzle, eroded through the debris fans so uniform, and why is the value of constriction specifically 0.5?" The answer lies in Figure 16.16d (and in extrapolations of this figure to higher discharges found as Fig. 13 in Kieffer's 1985 publication).

The fact that the Colorado River is less constricted at most of the tributary debris fans than it is at Crystal Rapids suggests that discharges higher than 92,000 cfs have occurred in their history. We know this to be true: A flood of 220,000 cfs occurred in 1921, and a flood estimated at 300,000 cfs occurred in 1884. We can reasonably assume that even larger floods have occurred since many of these debris fans formed, a time that may exceed 10^4 years (Hereford 1984).

Because of the higher velocities associated with higher discharges, larger floods will make the channel at a debris fan wider. Extrapolation of the calculations done for Crystal Rapids, based on a threshold transport velocity of 30 ft/s (9 m/s), suggest that floods as large as 400,000 cfs are required to open the channel up to the constriction of 0.5. The accuracy of this estimate cannot be stated because we know too little about the threshold velocity for erosion (or other similar criterion). There are other variables to consider as well: (1) the reliability of our extrapolations using standard power-law functions of the dependence of depth, velocity, and head-on discharge; (2) our lack of knowledge on the rate at which vertical cutting and headward erosion occurred (we are assuming that the channel comes to an equilibrium shape during each flood); and (3) our inability to consider the true geometry of the river channel. As further data become available, we will be able to construct more accurate models.

Despite these uncertainties, we know that discharges an order of magnitude greater than discharges from the power plant at Glen Canyon Dam (and approximately a factor of five greater than the 1983 flood levels) contoured much of the river channel at the debris fans and gave the rapids their characteristic configuration. Without floods of this magnitude in the future, the character of the rapids will change as tributaries flood. The change will be toward more highly supercritical conditions as the constrictions become narrower, both laterally and vertically.

• 17 •

Late Cenozoic Lava Dams in the Western Grand Canyon

W. K. Hamblin

INTRODUCTION

"We have no difficulty as we float along, and I am able to observe the wonderful phenomena connected with this flood of lava. The canyon was doubtless filled to a height of 1,200 to 1,500 feet, perhaps by more than one flood. This would dam the water back; and in cutting through this great lava bed, a new channel has been formed, sometimes on one side, sometimes on the other. . . .

What a conflict of water and fire there must have been here! Just imagine a river of molten rock running down a river of melted snow. What a seething and boiling of waters; what clouds of steam rolled into the heavens!"

J. W. Powell, Aug. 25, 1869

From the time Powell first viewed the remnants of basalt adhering to the walls of the inner gorge in the western Grand Canyon over 100 years ago, relatively few people have had the opportunity to see and study this isolated area. But more and more visitors are discovering the viewpoint at Toroweap, where they are privileged to see one of the most spectacular displays of volcanism in North America (Fig. 17.1).

The volcanic features of this area are much more complex than one might first imagine. What Powell observed during his epic trip down the Colorado River was only a small fraction of the region's volcanic phenomena. Over 150 lava flows have poured into the canyon during the last 1.5 million years, and they have left an incredible record of volcanic events and their influence on the Grand Canyon. Some flows were extruded on the Uinkaret Plateau and cascaded over the north rim of the canyon into Toroweap Valley and Whitmore Wash. Others were extruded within the canyon itself and spread out over the Esplanade Platform before forming spectacular frozen lava falls that plunged over the rim of the inner gorge into the Colorado River 3000 feet (900 m) below. In several places, volcanoes are perched precariously on the very rim of the canyon, and remnants of others cling to the steep walls of the inner gorge. In addition, the dikes, sills, and volcanic necks exposed in the canyon are all associated with the complex sequence of recent volcanic events in the Uinkaret Plateau.

The spectacular lava falls that spill over the Esplanade into the inner gorge cap remnants of an older sequence of flows that formed huge lava dams, some of which were over 2500 feet (600 m+) high. One was more than 84 miles (135 km) long. The lava dams backed up the water of the Colorado River to form

FIGURE 17.1. Volcanic features in the western Grand Canyon. View looking northeast at the cascades and remnants of lava dams near the mouth of Toroweap Canyon. Vulcan's Throne is perched on the rim of the inner gorge. Recent extrusions of basalt flowed across the Esplanade platform and cascaded into the inner gorge, where they cap large remnants of major complex lava dams. Smaller remnants of other dams can be seen high on the north wall of the canyon.

temporary lakes upstream. Several of these lakes extended upstream through the Grand Canyon into Utah, slightly beyond the present extent of Lake Powell. As the lake behind the barrier overflowed, a new gorge was eroded through the lava dam, leaving only remnants of the basalt clinging to the walls of the canyon. Later eruptions formed new dams, which subsequently were breached and largely destroyed by the overflow of the Colorado River.

At least 13 major lava dams were formed in the Grand Canyon during the last million years. These dams, together with remnants of the sediment deposited in the lakes behind the dams, provide a fascinating record of this unusual and most recent series of events in the history of the Grand Canyon.

METHODS

The volcanic phenomena in the western Grand Canyon region are exceptionally well exposed, but they present some unusual problems to the field geologist because of the extremely rugged and largely inaccessible nature of this part of the canyon. Most of the basalt remnants exist as thin slivers clinging to the vertical cliffs of the inner gorge. This means that they are inaccessible for the most part and cannot be reached by trails from the canyon rim.

To study the exposures of lava throughout their 84-mile (135-km) extent along the river, we had to float equipment and supplies 180 miles (288 km) downstream from Lees Ferry at the beginning of each field season and establish a series of temporary base camps along the river. This allowed us to enter and leave the canyon and work slowly downstream so that our river operations could last the entire season.

However, working from the river in this manner presented its own problems. Although the basalts are almost 100 percent exposed and stand out in stark contrast to the tan and reddish Paleozoic strata, many of the exposures are largely inaccessible because they exist as vertical cliffs hundreds to thousands of feet above the river. In an attempt to solve this problem, we photographed the entire canyon wall from view points on the opposite side of the river, using a handheld aerial camera. Enlarged prints were made and fitted together to produce a photo mosaic (cross section) on which we could plot all of our measurements and geologic observations. Elevations of the contacts between flow units were measured from the river using a theodolite.

We plotted our original data on enlarged vertical aerial photographs, and later we transferred our mapping to 7.5-minute topographic maps as the maps became available.

We also established several base camps on both the north and south rims of the Esplanade to study the more complicated areas between Toroweap and Whitmore Wash. We obtained additional data by using light aircraft and a helicopter.

Although some tantalizing questions about the details of certain relationships among the lava flows and the significance of various flow units remain, the study we made has established a significantly large database. And from this base, reasonably safe interpretations can be made about the late Cenozoic history in the western Grand Canyon.

Methods of Determining Relative Ages

The relative age of most of the flow remnants in the canyon is expressed clearly by superposition or by juxtaposition. The process by which juxtaposition of the flow remnants is produced is shown in the series of diagrams in Figure 17.2. The idea is simple. The first intracanyon flow entered the canyon from cascades or from centers of extrusion in the canyon itself. The lava partially filled the canyon, causing a lake to form upstream. Eventually, the backwater overflowed the barrier and eroded most of the basalt, leaving only thin vestiges of the lava flow adhering to the canyon walls in place of the once continuous flow. A subsequent flow entering the canyon would be juxtaposed against the older.

The remnants of lava flows in the Grand Canyon may be recognized and correlated throughout the canyon on the basis of several criteria. Although the petrography of some flows are similar, others are unique. Therefore, some flows can be recognized without difficulty on the basis of petrographic characteristics

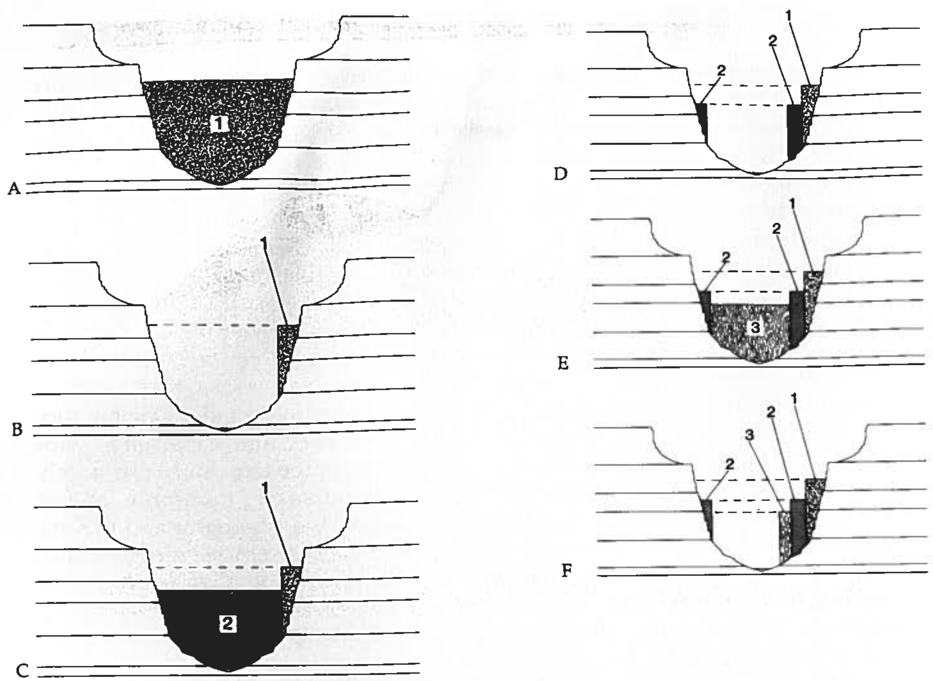


Figure 17.2. Diagrams showing the development of juxtaposed lava flows in the Grand Canyon. (A) A lava flow partly fills the canyon. (B) Erosion leaves small remnants of flow 1 adhering to the canyon wall. (C) Flow 2 refills the canyon. (D) Erosion removes most of flow 2, leaving remnants juxtaposed against flow 1 and against the canyon wall. (E) Flow 3 fills the canyon. (F) Erosion of flow 3 leaves remnants of flows 1, 2, and 3 stacked side-by-side according to relative age.

alone—even in a small outcrop. Diabase flows, for example, are readily distinguished from the dense, black, aphanitic basalt that characterizes other flow units.

In addition to petrographic characteristics, the internal structure of a number of flows is a distinguishing characteristic. One flow can be recognized by its abnormally thick joint columns in the basal colonnade. These range from 7 to 18 feet (2 to 5 m) in diameter. Others have unique jointing in the entablature, not only in size but also in the geometry of the columns. Stratigraphic sequences also are useful in correlation because river gravels of a specific lithology and thickness may occur within a specific sequence of flows. In addition, elevation and gradients of the top of all flow units were measured carefully with a theodolite. This provided an important means of correlating flow units on the basis of geomorphic relations.

Using these methods, we were able to map remnants of 13 major lava dams that were formed and subsequently destroyed in the Grand Canyon during the last one to two million years.

CHARACTERISTICS OF LAVA DAMS

It is apparent from the sequences of basalt preserved in the inner gorge of the Grand Canyon that four different types of dams were constructed during the

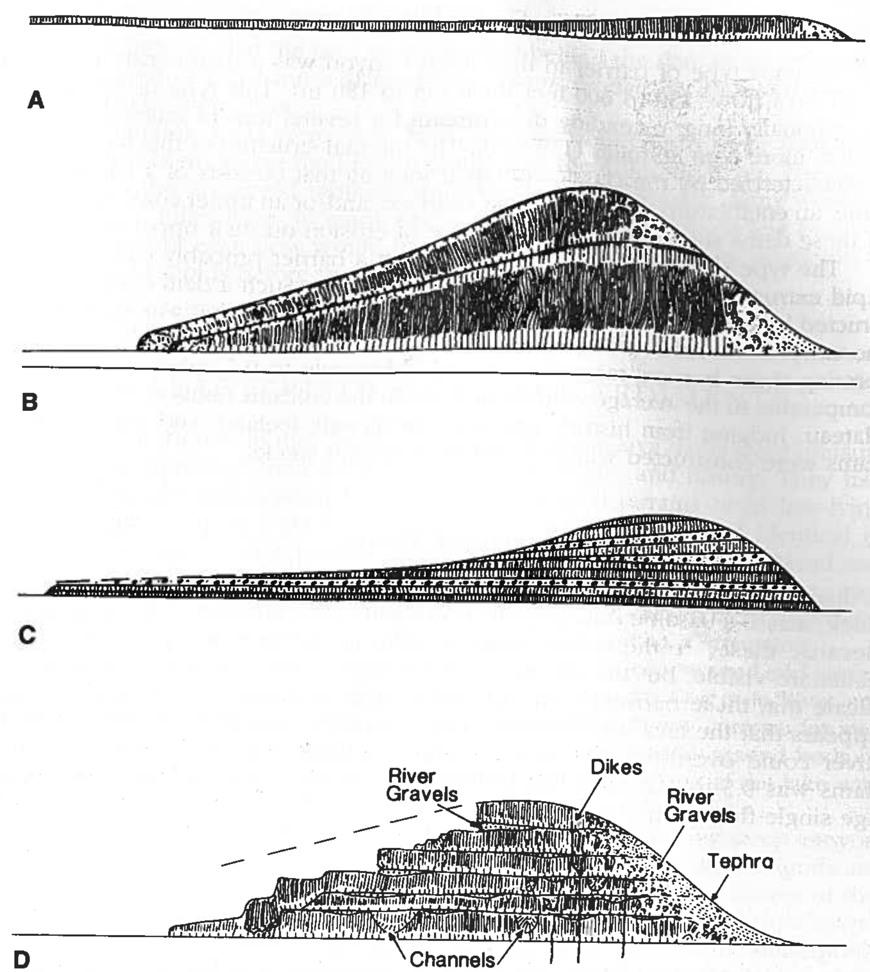


Figure 17.3. Types of lava dams in the Grand Canyon. (A) A simple dam formed by a single flow 150 feet to 600 feet thick. (B) High dam formed by several massive flows over 800 feet thick. (C) Compound dam composed of numerous flows 10 to 30 feet thick (3 to 9 m). (D) Complex dam built from multiple flow units 50 to 200 feet thick.

period of Late Cenozoic volcanic activity in the area. The nature of these structures is shown in Figure 17.3. Lava dams are simply intracanyon lava flows that attained a thickness sufficient to form a significant barrier to the flow of the Colorado River. They were thick, at least in part, because of the restriction of narrow canyon walls. These lava dams were asymmetrical structures, steep on the upstream margin and gently inclined downstream. The length of the dams ranged from 20 to more than 86 miles (32 to 137 km), and the height ranged from 150 to more than 2000 feet (45 to 600 m). The rate, volume, and viscosity of the lava extruded are primarily responsible for differences in the natures of the dams.

Thin Single-Flow Dams

The simplest type of barrier in the Grand Canyon was a dam constructed of a single lava flow, 150 to 600 feet thick (45 to 180 m). This type of barrier was exceptionally long, extending downstream for several tens of miles [in one instance, more than 86 miles (137 km)]. The internal structure of this type of flow is characterized by the classic columnar jointing that consists of a basal colonnade, an entablature, an upper clinker surface, and/or an upper colonnade. Most of these dams show little or no evidence of erosion on their upper surfaces.

The type of eruption that produced such a barrier probably consisted of a rapid extrusion of a moderate amount of fluid lava. Such a dam could be constructed by lava extruded within the canyon, or from lava that was extruded on the canyon rim and subsequently cascaded into the river. The volume of lava forming these barriers ranged from 0.03 cubic mile to 0.5 cubic miles. This is comparable to the average volume of flows in the volcanic fields on the Uinkaret Plateau. Judging from historic eruptions of Hawaii, Iceland, and Mexico, these dams were constructed within a period of several weeks.

Massive Dams

A distinctively different type of dam was formed in the canyon by abnormally thick, massive basaltic flows. These flows were over 800 feet (240 m) in depth. Because these are the oldest dams to form in the canyon, only a few remnants are visible, but the elevation of the upper surfaces of the remnants indicate that these barriers were not more than 20 miles (32 km) long. It also appears that the lava was extruded faster than the backwater of the Colorado River could overflow. The best estimates indicate that the volume of these dams was 0.5 to 1.2 cubic miles. This is only twice the volume of the average single-flow dam.

Compound Dams

Compound dams were constructed by a sequence of numerous flow units 10 to 30 feet thick (3 to 9 m), all of which were deposited in relatively rapid succession. These dams ranged from 300 to 900 feet (90 to 270 m) in height. The overall geometry of the compound dams was much like the simple single-flow dams, consisting of a steeply inclined front (upstream slope) and a long, gently inclined downstream slope. The volume of these dams ranged from 0.06 to 0.23 cubic miles.

Complex Dams

Complex dams were built from a series of multiple flow units 50 to 200 feet (15 to 60 m) thick. The general characteristics of each individual flow unit are similar to those of the single-flow dams, but the upper colonnade or clinker surface on flows within the complex dam often is eroded. Locally, deep channels are cut in the flow. These channels are filled with younger flows, ash, and sediment. Lenses of river gravel, sand, and, in some cases, ash are found separating the major units of basalt. The upstream slope of the complex dams consists of flow units that are inclined in a steep angle upstream. These pass rapidly into rubble, agglomerate, and tephra.

The erosional surfaces on the flows, together with interstratified river gravels, clearly indicate that the lake which formed behind the dam overflowed during the period of extrusion and most flows were eroded, to some extent, between periods of eruption.

The complex dams were quite high, ranging from 600 to more than 1400 feet (180 to 420 m) above the present gradient of the Colorado River. The steep gradient on the preserved remnant suggests that the complex dams were no more than 12 miles long. In many cases, it may have been much less because the lower end of each flow would have been eroded by upstream migration of waterfalls and rapids before the subsequent flows were extruded.

CHARACTERISTICS OF LAKE SEDIMENTS

The lakes that formed in the Grand Canyon behind the lava dams were unusual from the standpoint of their geomorphic setting, origin, and history. They had little in common with natural lakes formed in glacial terrains or in low-lying coastal regions. These were lakes formed in a deep canyon and identical in nearly every respect to the present man-made reservoirs such as Lake Mead and Lake Powell. A considerable amount of hydrologic data is available for these two reservoirs and provides the best insight into the nature of the sediment deposited in the lakes behind the lava dams, their history, and ultimate destruction.

When the Colorado River was blocked by a lava dam, coarse sand and gravel were deposited as deltaic sediments at the point where the Colorado River entered the lake. Fine-grained sediments were carried by turbidity currents far into the deeper part of the reservoir, where they were deposited as graded beds of silt and mud. In the absence of large tributaries, the main filling of the lake was accomplished primarily by deposition from the Colorado River.

The lakes formed in the Grand Canyon were surrounded by steep canyon walls. Large beaches rarely developed along the shore, and without significant input from the tributaries, the major process operating along the shores of the lakes was mass movement. Slope wash, rock falls, and the general downslope migration of colluvium (which presently is the major process operating along the canyon walls) would have continued during the lifetime of the lakes. A thin mantle of coarse, subaqueous slope debris was deposited close to the canyon walls contemporaneous with the deposition of lake silts. Each tributary continued to transport and deposit this material into the lakes at the heads of tributary bays.

Beyond the zone affected by slope processes, the major types of sediment deposited were mud and silt. Gravel would be a significant facies in some major tributary channels as a result of flash floods, and some of the larger tributaries near the upstream reaches of the lake probably constructed small deltas of sand and gravel. However, the major delta of sand and gravel formed in the area where the Colorado River emptied into the lake. Deltaic deposits prograded downstream over the silts deposited by turbidity currents.

After the dam was destroyed, most of the unconsolidated, water-saturated sediment was flushed rapidly out of the main canyon, leaving only minor remnants of sediment close to the valley walls. This material would be dominantly slope wash and colluvium with silt filling the space between the larger particles. Therefore, most of the preserved sediment near the canyon walls was not composed of typical lake silts. Instead, it was composed of colluvium with intercalated laminated silt. This type of deposit is difficult to distinguish from the present accumulation of colluvium.

DEVELOPMENT AND DESTRUCTION OF LAVA DAMS

The hydrologic data from the Bureau of Land Management's Lake Mead Survey (1963 and 1964) provide the basic information from which we are able to calculate the rates at which the lakes behind the various lava dams were filled with water and, subsequently, sediment. These data also provide some indication of the time necessary for a lava dam to be eroded away completely (Table 17.1).

Rates of Formation of Dams

Although the formation of a lava dam in the Grand Canyon was a significant event that dramatically changed canyon morphology, the time needed to create a lava dam was remarkably short by any standard and certainly would be considered instantaneous in a geologic time frame. Observations of basaltic eruptions in historic times indicate that most basaltic extrusions occur in a matter of days or weeks. The major flows in the Grand Canyon, most of which were 100 to 200 feet (30 to 60 m) thick, probably moved tens of miles down the Colorado River in a matter of days.

This conclusion is supported by the fact that the upper colonnade and/or a clinkery upper surface of the flow often is preserved, essentially unmodified by erosion. This indicates that the flow was extruded in a period of time less than that required for the lake impounded behind the dam to overflow. If extrusion occurred during a longer period of time, the lake behind the dam would overflow, and erosion would modify the upper surface features of the basalts quickly.

We can calculate quite precisely the time required for the backwater behind each dam to overflow. Based on present discharge rates of the Colorado River, the lake formed by backwaters behind a lava flow 100 feet (30 m) high would overflow in 17 days. The construction of a single-flow lava dam 2000 feet (600 m) high and tens of miles long could be completed within a few weeks. Therefore, it is clear that any lava flow retaining an uneroded upper surface was extruded in a matter of a few days.

Table 17.1. Data on the Geometry, Age, and Hydrology of the Lava Dams in the Western Grand Canyon

Dam	*Elevation*	*Height above River*	*Radiometric Date*[a]	*Volume of Lava*	*Lake Length*	*Water Fill Time*	*Sediment Fill Time*
1 Prospect	4000	2330	0.68 ± 0.05	4.0		23 yrs	3018 yrs
2 Lava Butte	3365	1730	0.58			10 yrs	382 yrs
3 Ponderosa	2800	1130	0.61 ± 0.02	2.5		1.5 yrs	163 yrs
4 Toroweap	3093	1443	0.56 ± 0.07	3.7		2.62 yrs	345 yrs
5 Esplanade	2600	960		1.8		287 days	92 yrs
6 Buried Canyon	2480	850	0.91 ± 0.07	1.7		231 days	87 yrs
7 Whitmore	2500	900		3.0	100	240 days	88 yrs
8 "D" Flows	2295	635	0.58 ± 0.03	1.1	74	87 days	31 yrs
9 Lava Falls	2260	600		1.2		86 days	30 yrs
10 Black Ledge	2033	373		2.1	53	17 days	7 yrs
11 Gray Ledge	1813	203	0.79 ± 0.13	0.3	37	2 days	10.3 mos
12 Layered Dbs	1938	298	0.62 ± 0.05	0.3	42	8 days	3 yrs
13 Massive Dbs	1826	226	0.14 Ma	0.2		5 days	1.4 yrs

[a]Dates from Dalrymple et al. (1998).

Dams built from multiple flow units required more time and involved cycles of partial erosion between periods of extrusion. The short time necessary for the complete erosion of a dam, however, puts definite time constraints on the development of any barrier to the Colorado River. Regardless of size and history of eruption, the formation of every lava dam in the Grand Canyon was instantaneous from the perspective of a geologic time frame.

Rates of Reservoir Fill

The time required for reservoirs behind these dams to become filled completely with water and sediment also was extremely rapid. The hydrologic data from the various lakes are summarized in Table 17.1. These data indicate that the lakes formed behind the smaller barriers—those 150 to 400 feet (45 to 120 m) high—would overflow in 2 to 17 days. Lakes formed behind the higher dams (200 to 1000 feet or 60 to 330 m high) overflowed in 22 years. Thus, the lava dams were subjected to erosion soon after they were formed, even before the interior of the lava was completely cool.

The volume of sediment carried by the Colorado River has been measured for many years by the Bureau of Reclamation. These data indicate that reservoirs behind the lava dams silted up in only a few hundred years at most. Many of the smaller reservoirs silted up in a few months . Thus, the sediment load of the Colorado River was soon transported over the dam, causing normal erosion by abrasion of the river channel after the dam was formed. The data in Table 17.1 indicates that a reservoir formed behind the dam 150 feet (45 m) high would be filled with sediment in 10.33 months. A dam 1150 feet (345 m) high would be full of sediment in only 345 years. The highest dam would be filled with sediment in 3000 years. Thus, every phase of the construction of the dam and the formation of the reservoir or lake behind it, and the ultimate filling in of the lake with sediments, occurred in a very short time.

Erosion and Destruction of Dams

Although we do not know the precise details in which lava dams were eroded, some boundary conditions can be established, based on downcutting of major stream systems. Normal downcutting of the stream channel by abrasion was undoubtedly a significant process of erosion. It began as soon as water overflowed the dam and reached maximum efficiency when the lake silted up and a normal sediment load was transported over the dam. Another important process was the migration of rapids and waterfalls that initially formed on the downstream end of the flow (Rogers and Pyles 1979). Two important characteristics of the intracanyon flows facilitated the migration of waterfalls (Fig. 17.4):

1. Intracanyon flows were deposited directly on the sand and gravel bed in the channel of the Colorado River. This layer of unconsolidated sediment would be eroded easily by undercutting at the plunge pool below the waterfall.
2. The vertical columnar jointing in the basalt constituted an all-pervasive structural weakness throughout the flow. The hexagonal columns produced by the jointing impart a low cohesive strength to the rock body so that the columns would readily topple into the plunge pool beneath the waterfall.

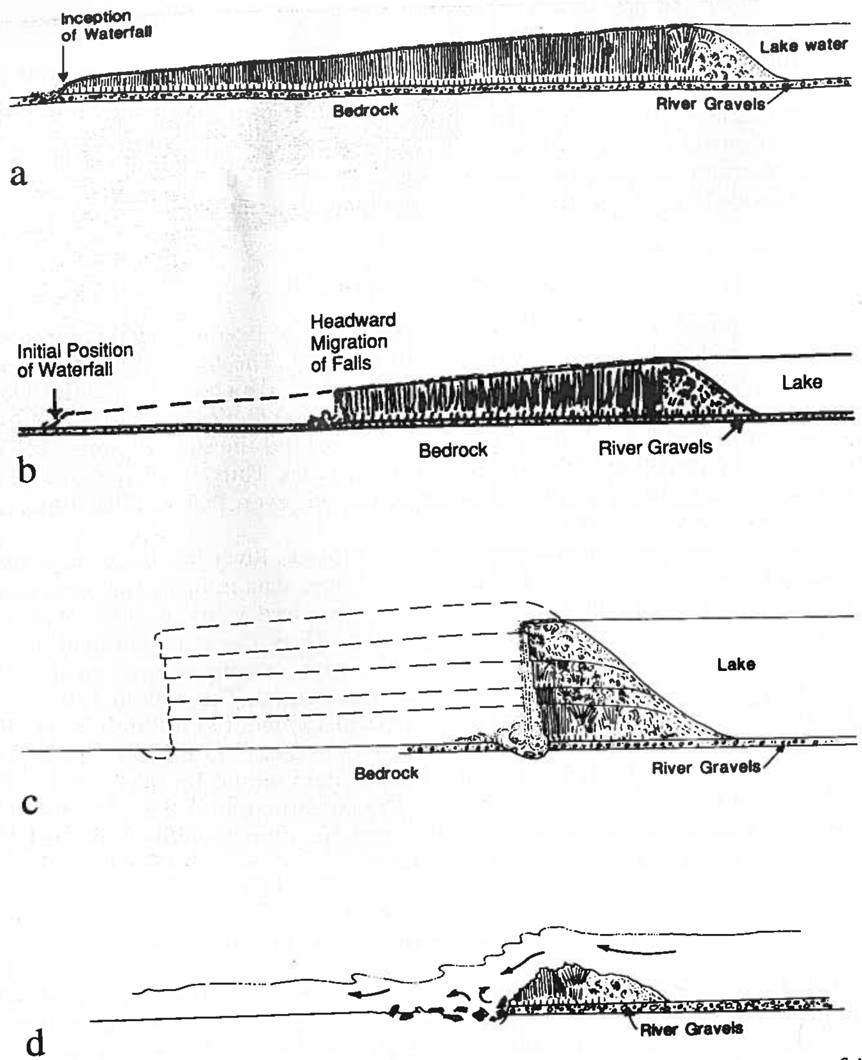


FIGURE 17.4. Diagrams showing erosion of a lava dam by headward migration of waterfalls. (a) As soon as the backwater overflowed the lava dam, a small waterfall would form at the downstream end of the lava flow. (b) Upstream migration of the falls would be accelerated by undercutting of the unconsolidated river sediments beneath the flow and by the weakness of the rock resulting from vertical columnar jointing. (c) As the waterfalls migrated headward, the stability of the dam could be jeopardized by the pore pressure in the columnar jointing. (d) Some dams may have failed catastrophically. This event could have been minimized if contemporaneous downcutting lowered the level of the overflow before the falls migrated to the head of the dam.

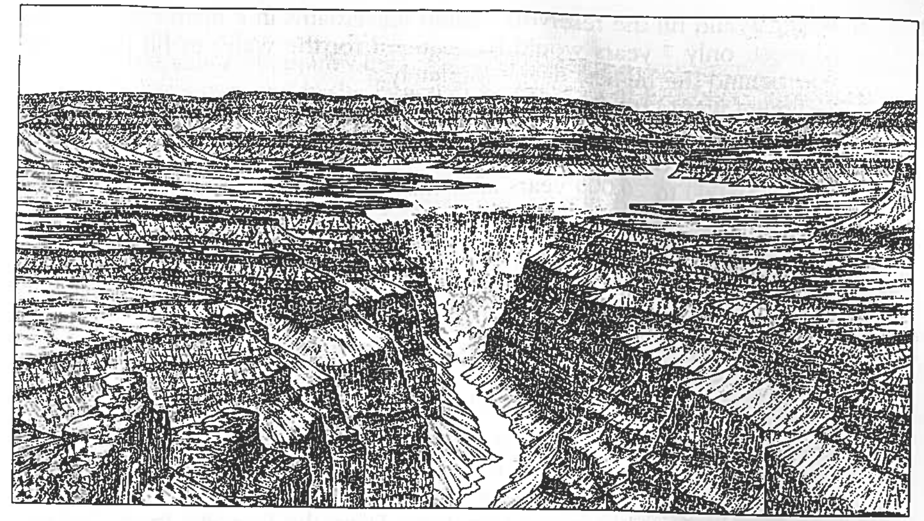


FIGURE 17.5. Waterfalls over a lava dam at Toroweap Canyon. Exceptional high waterfalls [up to 2460 feet (750 m) high] would form as headward erosion destroyed the dam. The higher dams would be 15 times as high as Niagara Falls with more than twice the discharge.

In addition, hydraulic plucking of the columns probably occurred in a river with a discharge as high as that of the Colorado. The combined effects of the unconsolidated sand and gravel substratum, and columnar jointing in the basalt, resulted in rapid upstream migration of the waterfalls. As the waterfalls migrated upstream and approached the head of the dam, they became higher and higher. The process of erosion then was accelerated by the increase in potential energy for undercutting and scouring. Erosion of the highest dams must have presented a spectacular scene. As the waterfalls advanced toward the highest section of the dam, a tremendous scour hole must have been generated at the base of the falls. Exceptionally high waterfalls—those over 2500 feet (759 m)—would form on the highest dams (Fig. 17.5). As erosion proceeded headward, the stability of the dam almost certainly was jeopardized by the enormous pore pressure in the columnar joints. At some critical point, dams may have failed as catastrophic events, with a rapid discharge of a tremendous volume of water and saturated sediment that had accumulated in the lake behind the dam.

Documentation of the rate of waterfall migration in some areas provides important insight into possible time frames of the destruction of the lava dams. Niagara Falls, for example, has migrated a distance of more than 11 miles (18 km) in 8000 years, an average rate of three feet per year. If this figure is typical for the migration of a waterfall on a large river, the lava dams in the Grand Canyon, which generally were less than 20 miles long, would take a maximum of 20,000 years to be completely destroyed by headward migration of waterfalls alone. It also seems safe to conclude that the time interval for the various phases of the buildup and destruction of the dam would be on the following orders of magnitude:

1. Single-flow dams would be formed in a matter of several days, whereas the higher compound lava dams would take up to several years for construction.

2. Water would fill the reservoir behind these dams in a matter of months. At most, only 7 years would be required for the water to fill the reservoir behind the highest dam completely.
3. Sediment would fill the small reservoirs within 1 to 7 years. In the higher reservoirs, it would take between 100 and 1000 years to fill the lake completely with sediments. Most of the dams probably would be destroyed within 10,000 to 20,000 years after they were formed.

It is apparent from these observations that the 13 lava dams occupied the canyon for a relatively brief period of time, probably no more than a total of 240,000 years.

THE LAVA DAMS

In the Grand Canyon the oldest lava dams whose remnants are still preserved are (a) two huge structures formed by thick, massive flows and (b) a multiple flow structure composed of numerous flows 50 to 150 feet thick. These oldest known flows are referred to as the Lava Butte Dam, the Prospect Dam, and the Ponderosa Dam.

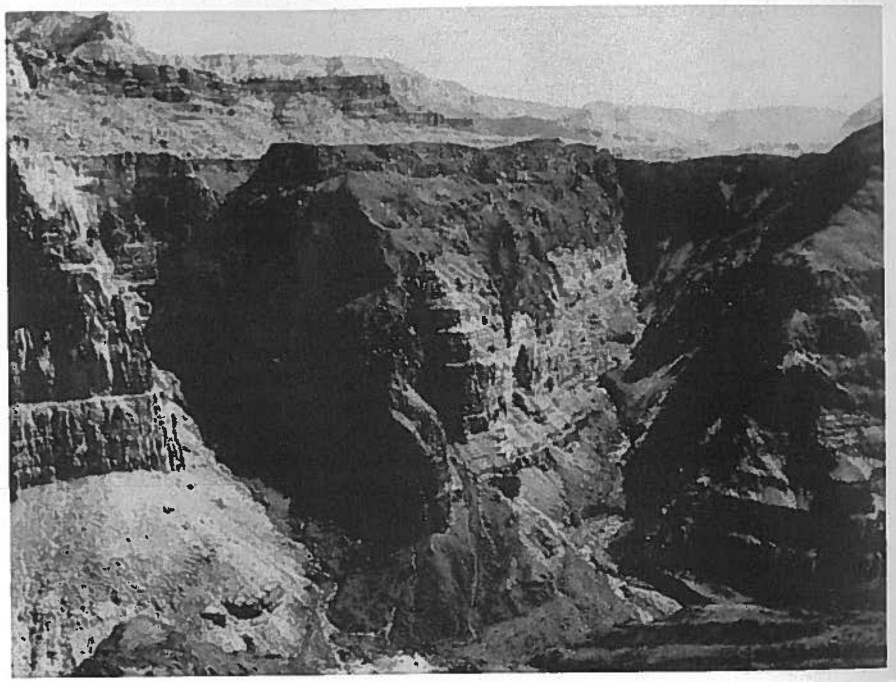


Figure 17.6. (a) Photograph of the Prospect Dam. Although only one remnant of this dam remains, the thick flows suggest that the dam did not extend downstream for more than a few miles. The absence of an erosional surface between flows suggests that extrusion was rapid and the barrier was completely constructed before the backwater in the lake overflowed.

(continues)

The Prospect Dam

The only remaining remnant of the Prospect Dam is a sequence of exceptionably thick flows exposed in the large alcove just east of Prospect Canyon. Here, a vertical cliff of basalt almost 2000 feet (600 m) high extends from the talus slopes near the Colorado River up to the level of the Esplanade (Fig. 17.6). Unfortunately, the boundaries between the flow units are difficult to see from viewpoints on the river or from viewpoints on the north rim near Vulcan's Throne. Observations made from a helicopter, however, reveal that the Prospect flows are extremely thick and that there are no more than three or four major flow units, each of which is more than 800 feet (240 m) thick. The Prospect flows, therefore, are some of the thickest flows in the canyon.

The oldest units in Prospect Valley are a sequence of bulbous or elliptical basalt bodies with intersperse tephra. These flows are overlain by the thick, massive Prospect flow.

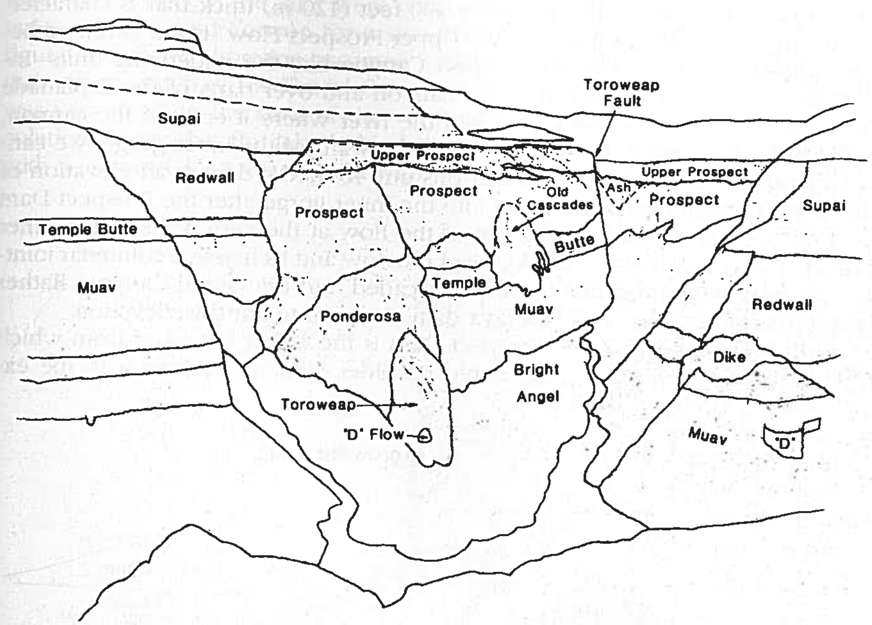


Figure 17.6. (*Continued*) (b) A cross-sectional drawing of the Prospect Dam. The structure of the dam consisted of several units: (1) the elliptical units formed by basalt interacting with backwater; (2) hydroexplosive tephra at the front of the dam; (3) the main flows with long, sinuous columnar jointing, (4) the basalt cap formed shortly after overflow of the lake. The Upper Prospect Flow consists of a horizontal unit nearly 400 feet (120 m) thick which once spread across the top of Prospect Canyon and across part of the Esplanade. It formed a cap rock covering the older flows and cinder cones that filled the ancient canyon. Headward erosion along the Toroweap Fault has re-excavated Prospect Canyon and formed a steep, V-shaped gorge over a mile long. Note the offset of the Upper Prospect flow near the apex of the canyon produced by recurrent movement along the Toroweap fault. Recent movement also has displaced the alluvium covering the Upper Prospect flow.

The elliptical structures probably were produced by the interaction of lava with the water of the Colorado River. Typical pillow structures would not be expected to form in abundance from a flow entering a river because most of the lava would not be covered completely with water. Only the upstream margins of the flows would interact directly with the river. The downstream segment of the flow would remain essentially dry until the backed-up river water overflowed the barrier.

Above the elliptical units are several exceptionally thick and massive flows that constitute the main body of basalt. These flows form a sheer vertical cliff over 1000 feet (300 m) high. The internal structure of these flows consists of long, slightly sinuous, columnar joints that tend to radiate out from a central point like huge shocks of wheat hundreds of feet high and only tens of feet wide. The only obvious stratigraphic break visible in this sequence of flows appears midway up the cliff at the approximate elevation of the contact between the Muav and Temple Butte limestones. A cross section showing the relationships between the flows in the Prospect alcove is shown in Figure 17.7.

On top of the sequence of thick flows exposed in the alcove east of Prospect Canyon is a single flow approximately 400 feet (120 m) thick that is characterized by massive columnar joints. This "Upper Prospect Flow" is not confined between the walls of the ancient Prospect Canyon like the underlying units. Instead, it spreads across the top of the canyon and over parts of the Esplanade (Fig. 17.6). This flow probably blocked the river where it entered the canyon, but because the flow is eroded back from the walls of the inner gorge, we cannot determine with certainty whether this unit formed a dam at an elevation of 4000 feet (1200 m) or if it cascaded into the inner gorge after the Prospect Dam was formed. The horizontal position of the flow at the very walls of the inner gorge, plus the considerable thickness of the flow and its massive columnar jointing, gives no suggestion that the flow cascaded into the Grand Canyon. Rather it supports the conclusion that a lava dam once existed at this elevation.

In the Grand Canyon the Prospect Dam is the oldest lava dam from which we have found remnants. There could be older dams associated with the extrusions capping Mount Trumball and Mount Dellenbaugh, but we found no evidence of their existence in the inner gorge. The old age of the Prospect Dam is indicated by the degree to which the flows in Prospect Canyon have been eroded. The steep, V-shaped gorge of Prospect canyon, which is over a mile long, has been eroded in the Prospect flows. This face shows that the Prospect flows are much older than the basalt that fills similar major tributaries to the Grand Canyon, such as Toroweap Valley and Whitmore Wash, both of which are filled with a sequence of flows only slightly modified by erosion. The flows in Toroweap Valley and Whitmore Wash, therefore, almost certainly are much younger than those that filled Prospect Canyon.

From the elevation of the Prospect Dam remnant, we are able to reconstruct the shoreline of the lake it formed. Prospect lake was the largest and deepest lake to form in the Grand Canyon. It extended all the way up through the Grand Canyon beyond Lake Powell and upstream to beyond Moab, Utah (Fig. 17.8). The rather extensive terrace deposits of gravel, sand , and silt in the Bull Frog area of Lake Powell and at Moab occur at elevations of approximately 4000 feet (1200 m) and probably are remnants of the Colorado River delta formed in the lake. The waters in the lake were deep enough to inundate most of the tributaries in the Grand Canyon for a significant distance. This meant that the configuration of the shoreline assumed a distinct dendritic pattern. The shoreline throughout much of the Grand Canyon in the vicinity of the park headquarters was very close to the base of the vertical cliffs of the Redwall Formation. Calculations show that the lake behind the Prospect Dam required 22 years to fill with water and 3018 years to fill with sediment.

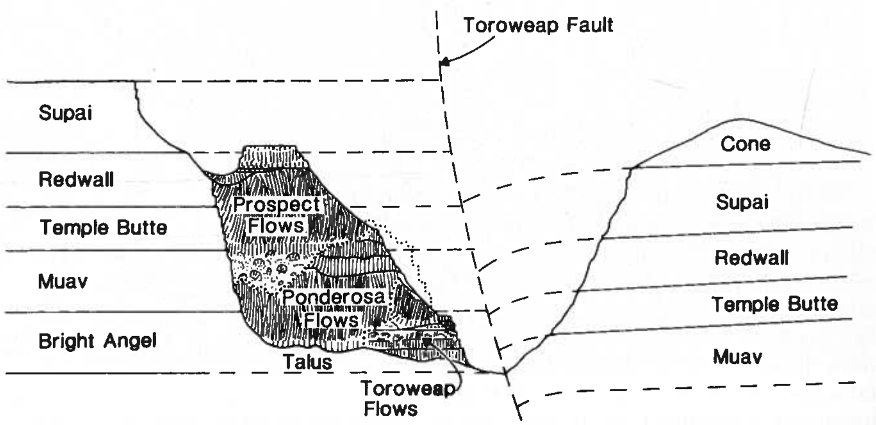


FIGURE 17.7. The flows in Prospect Canyon as seen from the west side of Vulcan's Throne. Over a mile of headward erosion has occurred in Prospect Canyon along the Toroweap fault, re-excavating a sharp, V-shaped gorge in the basalt and underlying Paleozoic rocks (Fig. 17.6). The thick, massive Prospect flow forms a sheer, vertical cliff almost 2000 feet (600 m) high near the left margin of the drawing.

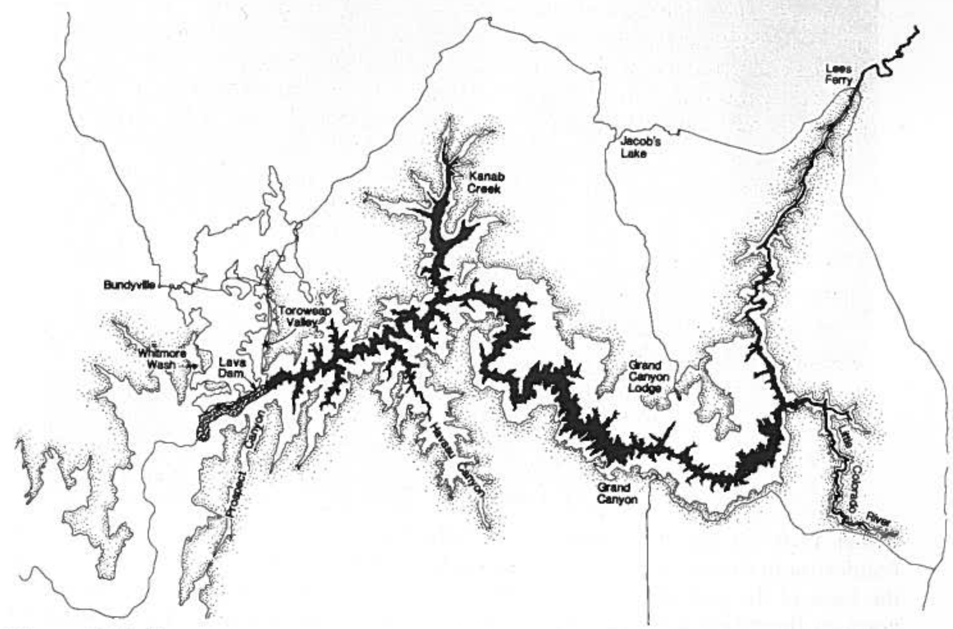


FIGURE 17.8. The Prospect Lake. The Prospect Dam was the highest lava dam formed across the Colorado River. The lake formed behind the dam that extended all the way through the Grand Canyon and up into Utah. It formed more than 1.2 million years ago.

Lava Butte Dam

Several remnants of a sequence of thin flows interbedded with river sediment are preserved high on the north wall of the inner gorge at mile 180.8 about 4 km downstream from Vulcan's Throne. One remnant caps an isolated butte that stands approximately 130 m above the surrounding area. If this remnant was formed by topographic inversion—that is, by erosion along the flow margins—it suggests that the Lava Butte Dam is one of the oldest dams preserved in the canyon. The remnants are not juxtaposed with other flows, however, so its relative age cannot be determined. The Lava Butte Dam reached a height of 560 m above the present river, but there are insufficient remnants to permit an accurate estimate of the original lava volume.

Figure 17.9. (a) The Ponderosa flow at Mile 181.6, as seen from the south rim. The Ponderosa in this area is 800 feet (240 m) thick and extends from near river level to the base of the cascades. Much of the lower part of the Ponderosa is obscured by younger flows that are juxtaposed against it. The Ponderosa Dam was formed from a single, thick flow in which the classic three-part columnar jointing is well-developed. It probably was a relatively short dam and did not extend downstream much beyond Mile 190.

(continues)

The Ponderosa Dam

The Ponderosa Dam was similar to, but somewhat smaller than, the Prospect Dam. It was constructed by a single flow at least 1000 feet (300 m) thick. The top of the barrier was at an elevation of 2800 feet (840 m), or 1130 feet (339 m) above river level. The best exposures are at the west end of the Esplanade Cascades on the north side of the river (Mile 181.6) and in the alcove east of Prospect Canyon (Fig. 17.9).

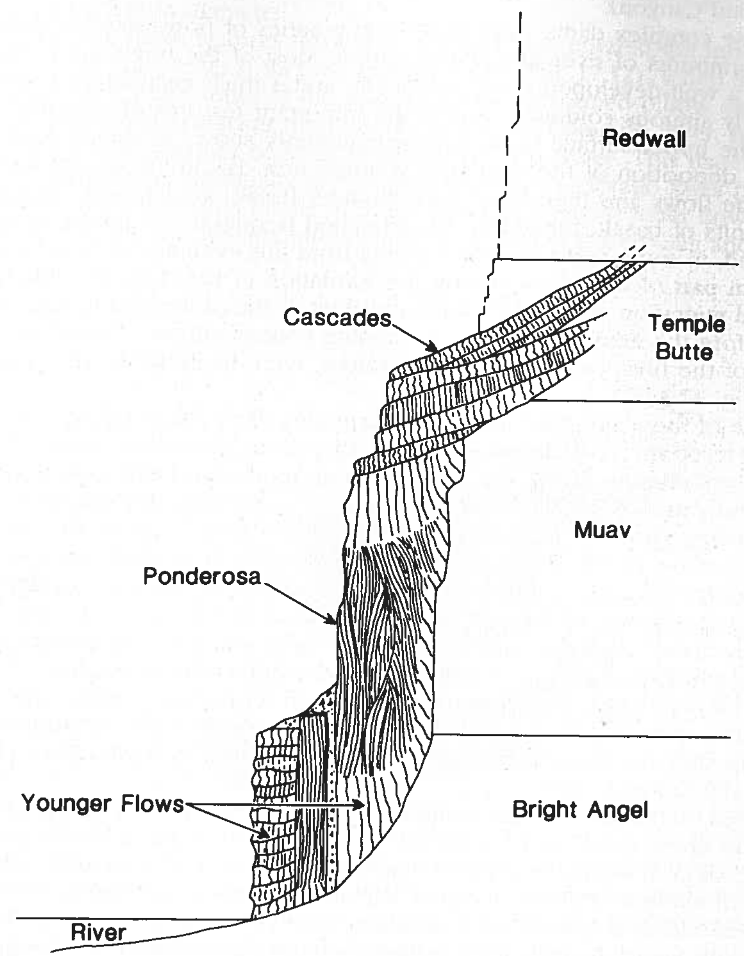


Figure 17.9. *(Continued)* (b) Diagram showing the juxtaposed relations of Ponderosa and younger flows. The Ponderosa is juxtaposed against the Paleozoic rocks and is capped by younger flows and cascades. Remnants of two younger dams are juxtaposed against the Ponderosa.

The lake that formed behind the Ponderosa Dam extended through the park headquarters area of the Grand Canyon upstream to somewhere near Nankoweap Rapids. It took only a little more than a year (450 days) for the Ponderosa Lake to fill and overflow. Within 162 years, the lake was full of sediment.

Complex Dams

After the destruction of the massive barriers of the Prospect and Ponderosa dams, three smaller dams were built. These were complex dams built from multiple units 50 to 200 feet (15 to 60 m) and represent a different style of extrusion. They are referred to as the Toroweap, Esplanade, and Buried Canyon dams after large exposures preserved on the north side of inner gorge at Mile 179 (mouth of Toroweap Canyon), Mile 181 (below the Esplanade Cascades), and Mile 183 (the Buried Canyon).

These complex dams were built from a series of flows interbedded with various amounts of river gravel and tephra. Most of the flows are characterized by a well-developed basal colonnade and a thick entablature consisting of slightly sinuous columns. One of the important features of complex dams is that the upper surface of each flow commonly shows evidence of erosion prior to deposition of the overlying younger unit. Locally deep channels are cut in the flows and then filled with younger basalt, gravel, and tephra. The major units of basalt, therefore, are separated frequently by lenses of gravel, sand, and, in some cases, tephra. Judging from this evidence of erosion in the upstream part of each flow during the formation of the dam, it is likely that waterfall migration destroyed a large segment of the downstream part of each flow before the extrusion of the succeeding younger flows. The downstream profile of the final complex dams, therefore, were likely steep and relatively short (Fig. 17.3d).

Each of the major flow units in the complex dams ranged from 150 to 200 feet (45 to 60 m) in thickness, and each caused an immediate barrier to form across the Colorado River. The small lakes accumulating behind these barriers rose rapidly and overflowed within a matter of a few days. (Calculations based on discharge rates of the Colorado River indicate that a barrier 100 feet high would overflow in only 2.3 days.) The overflow of the river then began to erode and modified the blocky rubble on the upper surface of the flow, together with much of the upper colonnade. Small waterfalls that formed at the end of the flow immediately began to migrate upstream. Subsequent extrusion of a younger flow, therefore, probably overlapped the eroded end of the older unit. The flows in the complex dams, however, must have been formed by a fairly rapid series of volcanic extrusions with only short time intervals between extrusions—time sufficient only for the upper segment of the overflows to be modified by erosion of the Colorado River.

Based on the present rate of discharge and the current sediment load of the Colorado River, overflow of a barrier 200 feet (60 m) high probably occurred within 3 days. It seems most likely that the small lake was completely silted up within 10 months. Without question, the small lakes formed behind each lava flow were filled completely with sediment prior to the next extrusion. Thus, by the time the complex dams were constructed, the lake behind it was completely silted up. Because a dam formed by a single flow 150 to 200 feet (45 to 60 m) would be completely eroded within 20,000 years, the construction of the complex dams must have been completed in less than 100,000 years.

The Toroweap Dam

Remnants of the Toroweap Dam occur on the north wall of the inner gorge beneath Vulcan's Throne. They are largely obscured from viewpoints at the Toroweap Campground, but from the south rim (at the mouth of the Prospect Canyon) and from the river, the thick sequence of flows can be seen adhering to the canyon wall (Fig. 17.10).

Excellent exposures of the upstream part of the Toroweap Dam are found on the north wall of the inner gorge, just upstream from the mouth of Toroweap Canyon (Fig. 17.11). Here, a detailed cross section of the internal structure of the Toroweap Dam is exposed to reveal the structure of the dam at points where the lava made contact with the waters of the Colorado River. Each of the major units shows three distinct types of structures. The front of the flow is characterized by a series of billowy, elliptical structural forms, many of which are 20 or 30 feet (6 or 9 m) in diameter. The size of the elliptical structures grade down to less than 3 feet in diameter.

We believe these features formed as a single flow entered the inner gorge and interacted with water from the Colorado River (Fig. 17.12). When the flows first entered the river, some of the material probably formed pillow structures. Because the flow was thicker than the depths of the Colorado River, however, much of the flow was not covered immediately with water. This material proceeded downstream, essentially dry. Backwater of the Colorado River probably overflowed the barrier in a matter of a few days. As a result, water flowed over the top of the hot lava, greatly influencing the nature of the cooling and the in-

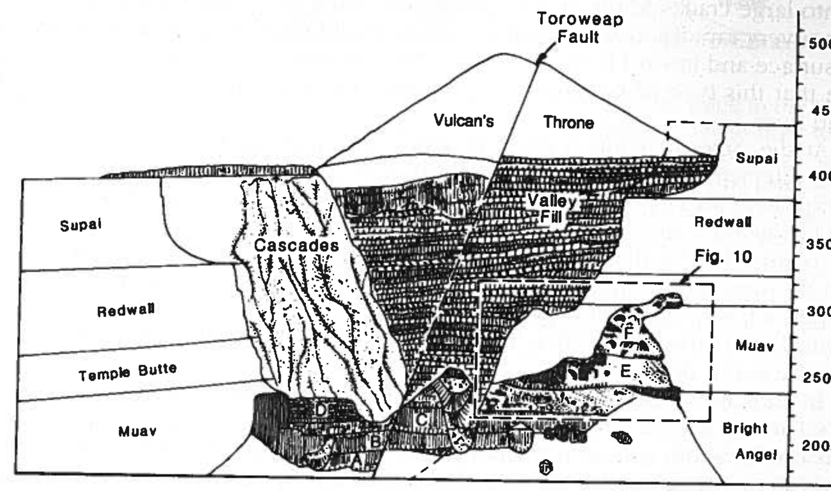


FIGURE 17.10. Diagram showing the relationships between the major flows at the mouth of Toroweap Valley. Remnants of the Toroweap Dam are exposed from river level to an elevation of approximately 3000 feet (900 m). They extend far beyond the mouth of Toroweap Valley and are juxtaposed against the Bright Angel, Muav, and Temple Butte formations. Toroweap Valley is filled with a sequence of thin flows to the level of the Esplanade. Recent cascades cover the western part of this sequence. Vulcan's Throne rests upon the flows t hat fill Toroweap Valley. The Toroweap fault displaces the entire sequence, including Vulcan's Throne.

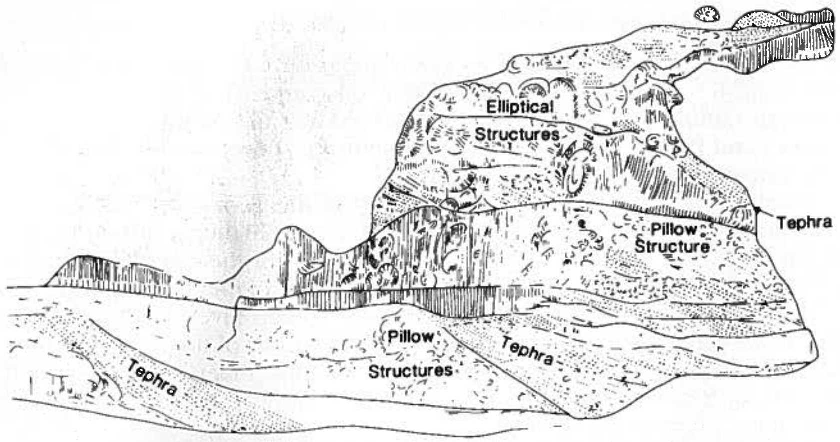


Figure 17.11. Sketch showing the internal structure of the Toroweap Dam, exposed in remnants high on the north wall of the inner gorge at Mile 178.4. The individual flows are 150 to 200 feet (45 to 60 m) thick and are characterized by basal colonnade and entablature jointing. Eastward, the flows develop elliptical structures that, in turn, grade into pillow basalts and stratified tephra. The tephra is inclined upstream. These structures suggest that this area is near the head of the lava dam.

ternal columnar jointing. Water from the overflow of the river would seep easily into large cracks in the lava, causing part of the flow adjacent to the fracture to cool very rapidly. Cooling in these cases would proceed inward from the upper surface and inward from the walls of the vertical cracks. Many geologists believe that this type of cooling produced the elliptical, pseudo-pillow structures found here.

At the point of contact with the river water, the lava was quenched, and water interacted with the basalt, causing it to contract and explode. The hydroexplosive activity resulting from the lava making contact with the water probably produced a small amount of tephra that accumulated in the steadily rising lake behind the barrier. This material was deposited at the angle of repose to form the upstream dipping layer of tephra. Undoubtedly, single-flow units developed lobes or offshoots that were diverted upstream a short distance over the elliptical structures and tephra. Thus, several minor flow lobes likely developed from the major flow unit.

In addition to the large remnant of the Toroweap Dam preserved below Vulcan's Throne, several small, isolated remnants are preserved high on the canyon walls on the south side of the inner gorge. These remnants, some 1400 to 1600 feet (420 to 480 m) above the present level of the river, can be observed easily from the Toroweap Campground.

The Toroweap Dam was at least 1443 feet (433 m) high, making it one of the higher dams in the Grand Canyon. Its history includes five major periods of extrusion, each followed by short periods of backwater overflow and erosion. Remnants of the Toroweap sequence are juxtaposed against the Ponderosa near the mouth of Prospect canyon, indicating that it is younger than both the Ponderosa and Prospect dams. Radiometric dates obtained from the oldest units (flow A) indicate that the Toroweap Dam is 1.16 to 1.18 million years old (McKee et al. 1967). More recent measurements indicate that the Toroweap Dam is only

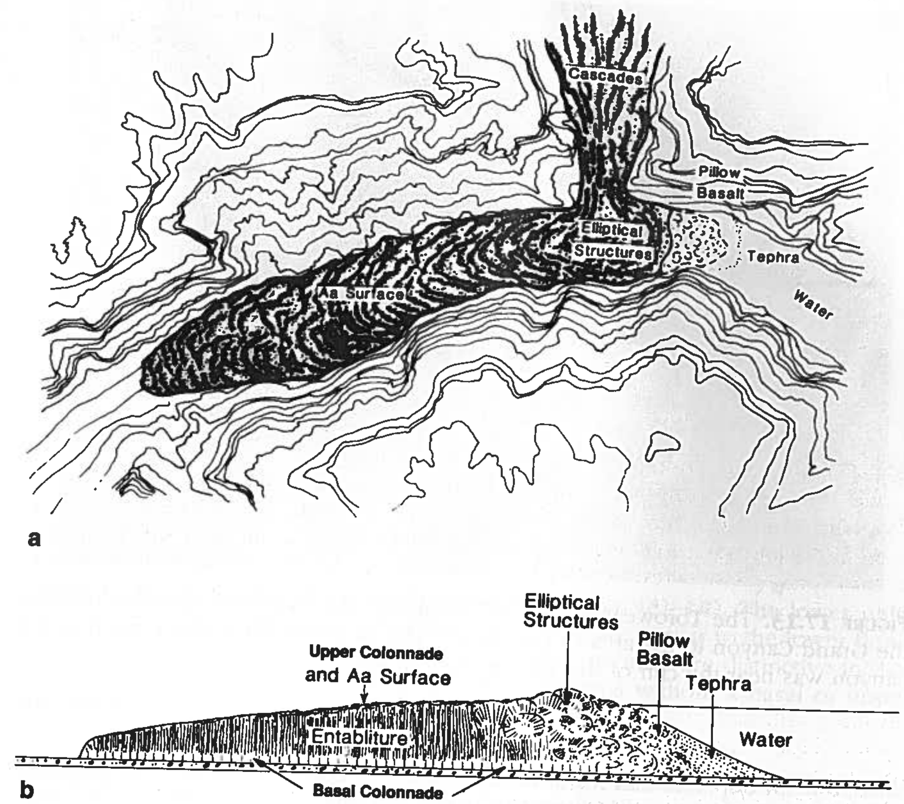


Figure 17.12. The structure of a hypothetical flow 200 feet (60 m) thick in the Grand Canyon. a. Top view, b. Lateral view. Only the upstream margin of the flow would interact with water from the river. This would produce hydroexplosive tephra, which would be deposited at the upstream margin of the dam. Some pillow structures would probably be produced where lava came into direct contact with the river. Most of the flow would move downstream essentially dry until the backwater overflowed the barrier. The Colorado River would overflow the barrier of a single flow within two days and would influence the cooling of the lava. The overflow may have had an effect on the development of the entablature in the downstream part of the flow.

0.56 ± 0.07 (Dalrymple et al., 1998). This indicates that the Toroweap Dam is older than the one formed in the Esplanade and Buried Canyon.

Toroweap Lake was one of the larger lakes to form in the Grand Canyon. It extended all the way through the present National Park Visitor's area and upstream into the vicinity of Lees Ferry (Fig. 17.13). Throughout much of the park region, the lake's depth was approximately 750 feet (225 m). In the region of the visitor's center, the lake essentially flooded all of Granite Gorge—with the shoreline occurring very close to the vertical wall of the Tapeats Sandstone.

The gravel, sand, and silt that formed the terrace deposits at Lees Ferry at an elevation of 3600 feet (1080 m) probably were deposited as a delta built by the Paria and Colorado rivers where they emptied into the Toroweap Lake. Like-

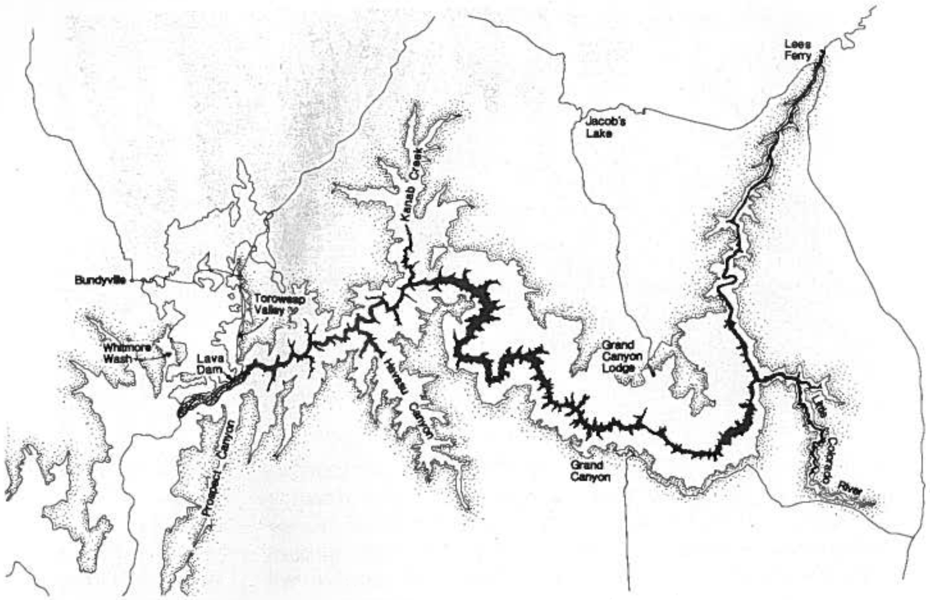


FIGURE 17.13. The Toroweap lake. The Toroweap lake extended upstream through the Grand Canyon to the area of Lees Ferry. The shoreline throughout much of the canyon was near the cliff of the Tapeats Sandstone.

wise, the silt deposits that form the main floor of Havasu Canyon probably represent a major remnant of Toroweap Lake deposits.

If the Toroweap Dam formed instantaneously, the lake behind it would overflow in 2.6 years. It would be completely full of sediment in 345 years. Because the Toroweap Dam was constructed over a period of time, the lake undoubtedly was essentially full of sediment by the time the dam was completed. After the extrusion that built the Toroweap Dam terminated, erosion by headwater migration of the waterfall and the downcutting of the stream channel probably destroyed the dam in less than 10,000 years.

The Esplanade Dam

On the north side of the river between Miles 181 and 182, a sequence of flows that formed a complex dam at an elevation of 2600 feet (780 m), or 960 feet (288 m) above river level, is preserved beneath the Esplanade cascades (Fig. 17.14). These flows can be seen from viewpoints west of the present Toroweap Campground and resemble in some respects the Toroweap sequence. Both are preserved beneath major cascades.

The Esplanade sequence also is similar in some respects to that preserved in the Buried Canyon at Mile 184. These similarities have led some geologists to consider the Toroweap, Esplanade, and Buried Canyon lavas as part of a single complex lava dam. However, the stratigraphic sequence in each dam is distinctly different.

Several small remnants of the Esplanade Dam are preserved in tributary canyons on the south wall of the inner gorge between Mile 183.5 and Mile 184.

FIGURE 17.14. Remnants of the Esplanade Dam at Mile 181–182. The lower units exposed near the lower right part of the photograph are similar to the lower flows of the Toroweap Dam. The upper units below the cascades are distinctive in that the columnar jointing consists of rough, thick columns without a basal or upper colonnade.

They provide important documentation concerning the height of the Esplanade Dam. All are preserved in "hanging valleys" at an elevation of 2600 feet (780 m).

The lake that formed behind the Esplanade Dam was similar to, but slightly smaller than, the Toroweap Lake and extended upstream only to Hance Rapids. It was exceptionally narrow throughout its entire extent and did not extend any appreciable distance up the tributaries. It probably was full of sediment by the time the youngest flow was extruded. Headward migration of waterfalls undoubtedly destroyed the dam within 10,000 years.

The Buried Canyon Dam

One of the most remarkable exposures of basalt in the Grand Canyon is on the north side of the river at Mile 183, halfway between Toroweap Valley and Whitmore Wash. Here, a segment of the Grand Canyon was filled with a sequence of older basaltic flows, which preserved a buried canyon (Fig. 17.15). The basalts here are not remnants clinging to the wall of the canyon as is the case in most areas. In this region, the basalts fill the entire lower part of the original inner gorge, just as they did when they clogged the canyon and formed a dam. The floor and walls of the original canyon are exposed in a perfect cross section. This unique exposure is preserved because the Colorado River shifted to the outside of a slight meander bend at the time it overflowed the lava dam. Subsequent downcutting formed a completely new canyon, leaving a section of the original gorge filled with basalt preserved on the inside of the meander.

The Buried Canyon contains nine major flow units totaling 650 feet (195 m) in thickness. The top of the highest flow stands at an elevation of 2480 feet (744

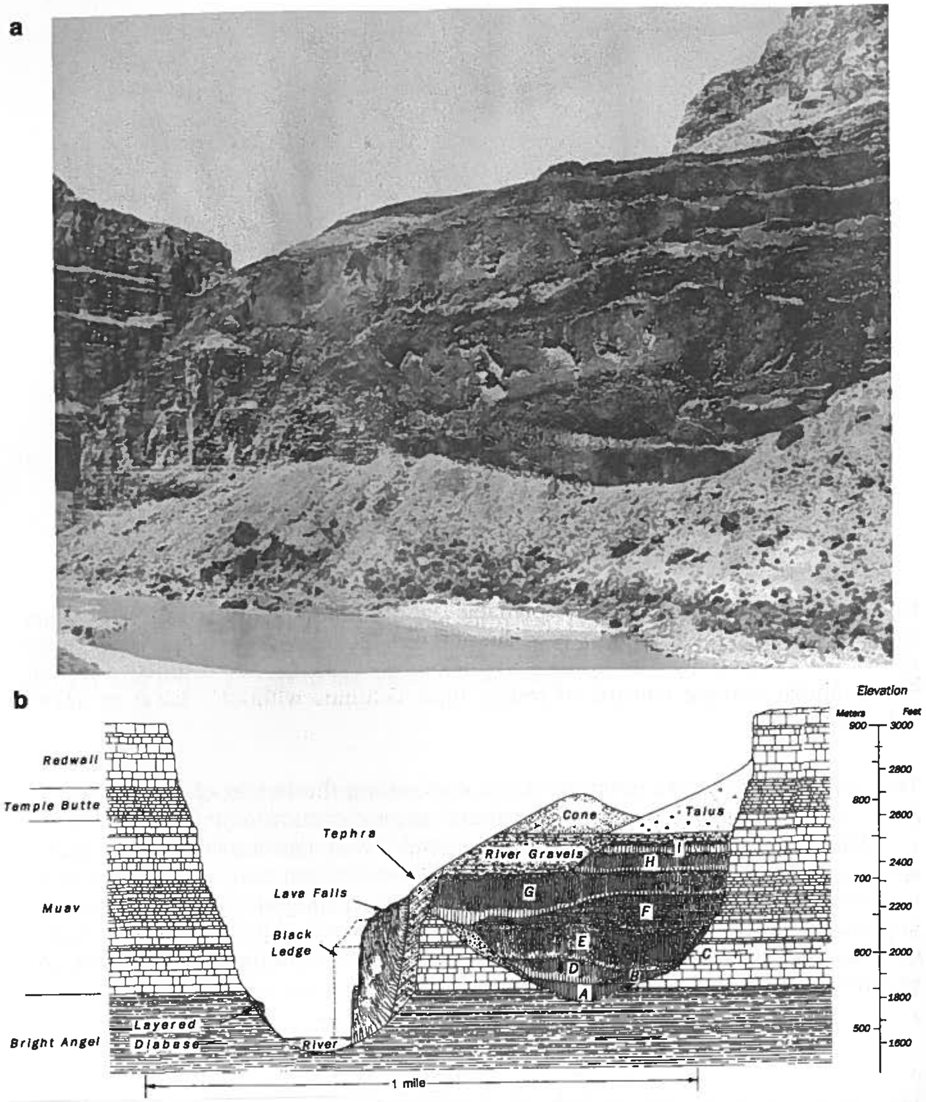


FIGURE 17.15. (a) The Buried Canyon Dam at Mile 183. (Note the boat near the sandbar for scale.) This sequence of flows filled the "original" inner gorge of the canyon. It is preserved because the Colorado River shifted southward around a meander bend and cut a new gorge in the Paleozoic strata instead of eroding the basalt. The Buried Canyon Dam was formed by a series of flows 150 to 200 feet (45 to 60 m) thick, all of which show evidence of erosion prior to the extrusion of the next younger lava. (b) Cross section showing the major flows in the Buried Canyon Dam. The nine major flow units shown in this section are separated by two major erosional surfaces. The entire sequence of basalt is overlain by 165 feet (50 m) of river gravels capped by a recent cinder cone.

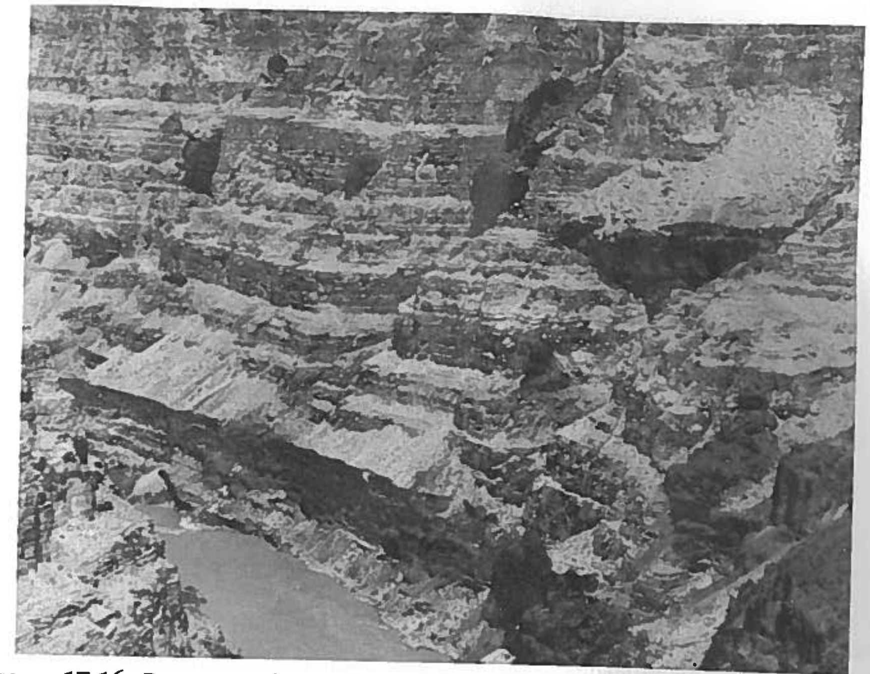


FIGURE 17.16. Remnants of the Buried Canyon Dam preserved downstream at Mile 184–185.

m), or 860 feet (258 m) above the present river level. The flows are overlain by a deposit of river gravels 160 feet (48 m) thick and are capped by a younger volcanic cone.

In addition to the remnants of the dam preserved in the Buried Canyon at Mile 184, a number of small remnants are preserved as "hanging valleys" in small tributary valleys on both sides of the river between Miles 184 and 185 (Fig. 17.16). These remnants represent segments of the higher basalts that formed the dam and flowed from the inner gorge up the small tributaries.

Several flows in the Buried Canyon sequence were dated by Brent Dalrymple from the U.S. Geological Survey. They range in age from 0.89 to 1.14 million years old. Based on these figures, geologists calculate that the buildup of the dam took place over a period of approximately 250,000 years. Large erosional channels in the basalt sequence indicate that the dam was built and partly destroyed four times during this period of time.

The lake behind the Buried Canyon Dam was similar to that formed by the Toroweap Dam and Esplanade Dam (Fig. 17.13).

The Whitmore Dam

One of the most obvious expressions of the juxtaposition of the remnants of a sequence of ancient lava dams in the Grand Canyon is in the vicinity of Whitmore Wash from Mile 187 to 190. Large remnants of a sequence of relatively thin lava flows can be seen filling Whitmore Wash and adjacent canyons when

Figure 17.17. The Whitmore Dam as seen from the air above Mile 192. View looking northeast. Large remnants of the Whitmore Dam are preserved on both sides of the river and form wide terraces. The source of the flows was from cascades into Whitmore Wash.

the area is viewed from the air above Mile 190 (Fig. 17.17). Downstream, this sequence is preserved as prominent terraces that outline large segments of the original dam on both sides of the canyon. What makes these exposures exceptionally striking is the way they are preserved. The remnants of the dam are not preserved in protected alcoves (which is more typical) but, instead, form large terraces—some more than one-quarter mile wide and over a mile long. Thus, they are large enough to constitute a significant geomorphic feature in the canyon, recognizable in aerial photographs and on topographic maps. This is the only dam in which relatively large segments of the original upper surface are preserved.

Whitmore Dam is distinctive because it was composed of numerous thin-flow units, most of which range from 10 to 20 feet (3 to 6 m) in thickness (Fig. 17.18). More than 40 individual flow units fill the valley of Whitmore Wash. Little or no sediment separates the flow units in the exposures at the mouth of Whitmore Wash or in the adjacent valleys, indicating that the flows were extruded in rapid succession.

The source of the flows that built the Whitmore Dam is very clear. As Figure 17.17 shows, lavas were extruded from eruptive centers near the southern tip of the Uinkaret Plateau and formed eruptive centers on the Esplanade. The lavas flowed westward and cascaded into Whitmore Wash. They then moved due south where the river makes an abrupt turn to the south. There was, therefore, no real deviation in the trends of the lava flows as they moved from Whitmore Wash into the Colorado River.

Because the dam was built up by numerous thin flows, the water of the Colorado River immediately overflowed each lava barrier. At various times, it

Figure 17.18. Remnants of the Whitmore Dam preserved high on the north canyon wall about one-half mile downstream from Whitmore Wash. The Whitmore Dam was distinctive in that it was composed of numerous, thin flows. Remnants of younger, lower dams are juxtaposed against the canyon wall in the central part of this photograph.

also deposited river gravels. In this particular situation, there was a strong tendency for the river to impinge against the south wall of the canyon because the lavas were moving into the canyon from the north. Subsequent flows were deposited on the river gravels near the south wall, whereas lavas deposited near the north and west walls were superimposed on one another with interlayers of gravel.

Whitmore Dam was 900 feet (270 m) high and over a half-mile wide, making it the widest lava dam in the Grand Canyon. The head of the dam was right at the confluence of the Whitmore Wash and Colorado River. If it had been constructed instantaneously, only 230 days would be required for the lake behind it to fill with water and overflow. However, the interbedded gravels in the Whitmore basalts indicate that the lake was filled with water and sediment and that it was overflowing at the same time that the lavas were being extruded.

Because the lake behind Whitmore Dam was close to the elevation of the Esplanade Dam, its configuration and shoreline probably were similar to that shown in Figure 17.13. The lake, extending upstream through the Grand Canyon to Hance Rapids, was 100 miles (160 km) long, and throughout most of its extent, it was very narrow. In Havasu Creek, the lake extended up to the base of the Mooney Falls. The level surface of lake silts between Mooney Falls and Beaver Falls most likely are deposits formed in Whitmore Lake.

D Dam

A number of small, isolated remnants of a sequence of thin basalt flows are preserved on both sides of the canyon near Lava Falls Rapids at Mile 179.5. These flows clearly are younger than the Toroweap sequence but are older than a series of younger dams, most remnants of which are further downstream. They represent a barrier across the canyon 635 feet (190 m) high. The sequence of lava originally was described by McKee and Schenk (1942) and is one of the most distinctive series of flows in the canyon. It consists of a series of flow units 5 to 15 feet (2 to 5 m) thick that are separated by thin layers of clinkers and ash. Remnants of the D flow are relatively small and inconspicuous; unless one is consciously looking for them, they may be missed entirely.

The Younger Dams

Five younger dams ranging in height from 200 to 600 feet (60 to 186 m) also were formed in the Grand Canyon. Most of these dams were constructed from a single flow that extended downstream several tens of miles, but one was formed from multiple, thin diabase flows. Numerous remnants of these younger dams are preserved on both sides of the canyon in the areas of least vigorous erosion. The size of the remnants varies from quite small to more than a mile long. Their geomorphic and stratigraphic relationships, for the most part, are indicated by juxtaposition. The best exposures for demonstrating the age sequence of the younger intracanyon flows are in the vicinity of Whitmore Wash, where five major intracanyon flows are exposed (Fig. 17.18).

The lakes formed behind the younger dams were barely more than 1/8 mile wide, occupying little more than the present river channel (Fig. 17.19). They extended upstream to the middle Granite Gorge, but did not reach the Grand Canyon Visitors Area. The lakes were long, narrow, and deep channels. The largest filled with water in 17.5 days and was silted up completely in 6.5 years.

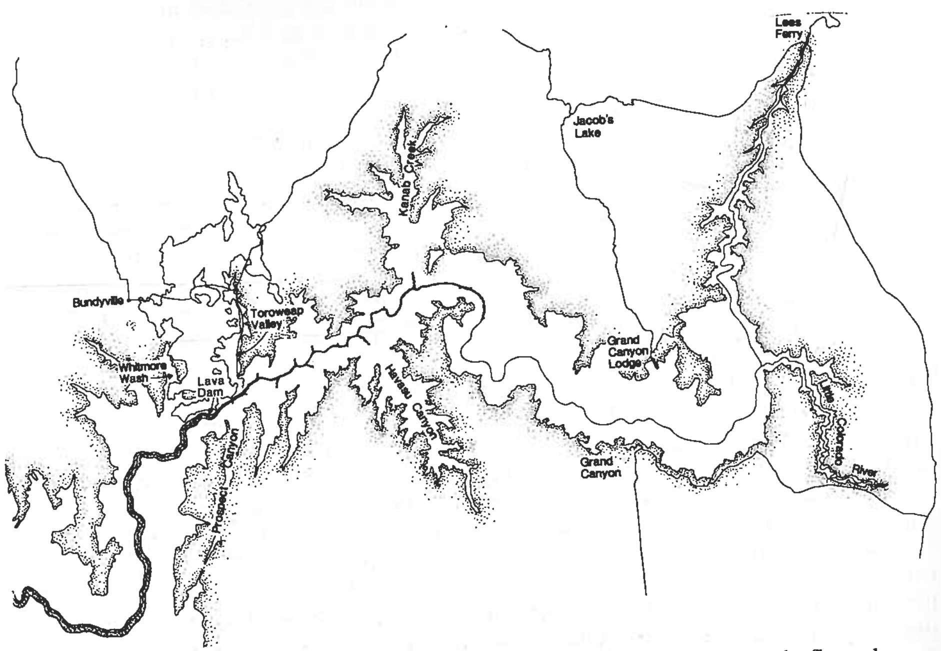


FIGURE 17.19. Lakes formed in the Grand Canyon by younger, single-flow dams.

LAVA CASCADES

Two major lava cascades, the "Toroweap Cascades" and the Esplanade Cascades" (Fig. 17.1), can be seen from the vicinity of Vulcan's Throne. In addition, a number of smaller "frozen lava falls" are found in the inner gorge east of Whitmore Wash.

The Toroweap Cascades, which spill over the rim of the inner gorge just west of Vulcan's Throne, originated from centers of intrusion high on the southern tip of the Uinkaret plateau just south of Mt. Emma. They flowed over the outer rim of the canyon, across the Esplanade, and then cascaded over the rim of the inner gorge into the Colorado River 3000 feet (900 m) below. The cascades represent some of the most recent volcanic activity within the canyon and clearly are equivalent to the younger flows on the Uinkaret Plateau. They are younger than Vulcan's Throne and likely represent volcanic eruptions within the last 10,000 to 20,000 years. The cascades are relatively thin flows, rarely more than 30 feet thick, and on the Esplanade they retain many of their original surface features such as pressure blisters. On the steep slopes, the cascades are jumbled and poorly defined. Near the western margins of the Toroweap Cascades, a relatively thick sequence of steeply inclined flows indicates that considerable volumes of lava entered the Grand Canyon from Toroweap Valley.

The Esplanade Cascades, which cap the Esplanade Dam, are similar to the Toroweap Cascades in that they represent the most recent volcanic activity within the area. The source of the cascades clearly was from centers of eruption on the Esplanade. The lava spread over the relatively flat surface of the Esplanade and spilled over the rim of the inner gorge at several places, both in this area and to the west of it.

Intrusions

The lava cascades between Toroweap Valley and Whitmore Wash are so striking when seen from the air, or from the vicinity of Vulcan's Throne, that there is a tendency to conclude that the cascades were the source of all intracanyon flows. There are, however, numerous dikes and cones within the inner gorge—indicating that much of the lava forming the lava dams may have been extruded within the inner gorge of the canyon. One of the largest, and certainly the most spectacular, evidences of volcanic events within the canyon is a huge mass of basalt referred to as Vulcan's Forge or Lava Pinnacle. It is located near the center of the Colorado River at Mile 177.9 (Fig. 17.20). Another volcanic neck similar in size to Vulcan's Forge is exposed on the south wall of the canyon at Mile 180.2. It is exceptional because the face of the canyon wall cuts across the neck and exposes a vertical cross section of the structure nearly 700 feet (210 m) high. The neck is surrounded by a white alteration zone that makes it exceptionally conspicuous from viewpoints on the Esplanade about a mile west of Vulcan's Throne.

Several systems of thin dikes are exposed on both the north and south walls of the Grand Canyon. These systems, however, are relatively obscure because they are only 3 to 5 feet (1 to 2 m) wide and follow a vertical joint system to

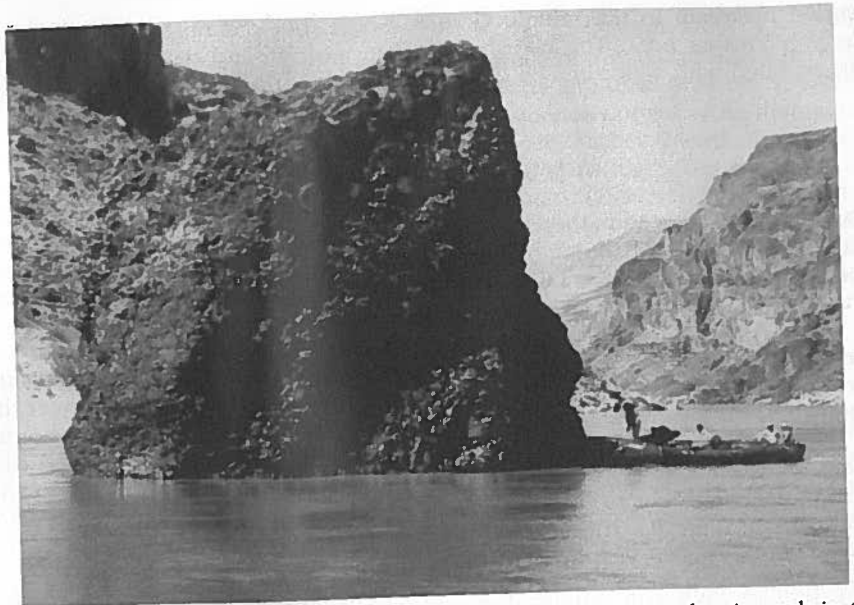


FIGURE 17.20. Vulcan's Forge at Mile 178. Vulcan's Forge is a volcanic neck in the middle of the Colorado River a short distance upstream from Vulcan's Throne.

the top of the Esplanade. They extend upward with little variation in width from the bottom of the canyon up to beyond the base of the Redwall Limestone. Several additional thin dikes intrude the older intracanyon flows and laminated ash in the vicinity of Lava Falls Rapids and downstream at Mile 182.5.

The most prominent dike, however, is located high on the south rim of the Grand Canyon near the mouth of Prospect Valley (Fig. 17.21). This dike is 30 to 40 feet (10 to 12 m) thick and forms a high wall projecting above the surrounding country rock. In contrast to the smaller dikes, the Prospect dike trends in an east–west direction, parallel to the Colorado river. A dike of similar size, but with a much less imposing topographic expression, is exposed along the rough trail descending the cascades from near Vulcan's Throne to the river below. It also trends in an east–west direction but is only slightly more than 10 feet (3 m) thick.

Cones

Most of the cinder cones found within the inner gorge are associated with the Toroweap fault. Vulcan's Throne is the most impressive, but a similar large cone is positioned on the canyon rim at the mouth of Prospect Canyon. In addition, remnants of five cones are found down in the canyon along the Toroweap fault zone. Two of these occur below Vulcan's Throne, and two are found along the Toroweap "trail" to the Colorado river. Also, a large cinder cone, located along the fault, is buried beneath the lavas in Prospect Canyon and now is exposed at the head of the new valley excavated along the trace of the Toroweap fault.

Several large remnants of cinder cones also adhere to the wall of the inner gorge in the vicinity of the Esplanade Cascades. The largest of these, called "Bill's Cone," is approximately the same size as Vulcan's Throne. Although much of

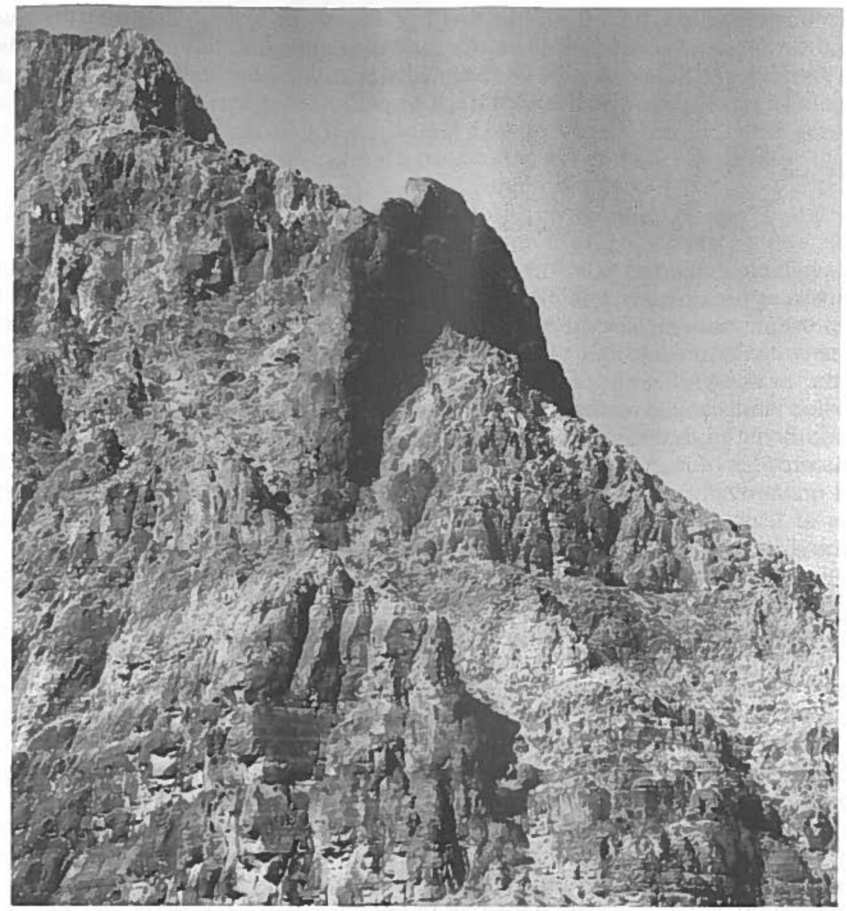


FIGURE 17.21. Large dike just west of the mouth of Prospect Canyon.

this cone has sloughed off into the canyon, the upper part of the cone and crater appears to be relatively unmodified by erosion. Small remnants of older cones adhere to the canyon wall in the same vicinity.

RATES OF EROSION

The series of lava flows within the Grand Canyon provide an unusually well documented record of rates of erosion, slope retreat, and the adjustment of streams to equilibrium. This is possible because the relative ages of the 12 lava dams have been determined by juxtaposition, and radiometric dates provide a series of time benchmarks for the major extrusions. In addition, we have measured the thickness, length, and volume of the lava dams so that we can calculate the rates at which erosion has cut through the dams. What we found is that the Colorado River was able to erode through the lava dams at an astounding rate. When we consider the total thickness of all the rock through which the river has cut its channel in the last million years or so, it is clear that the river has the capacity to erode through any rock type almost instantaneously and that the profile of the Colorado River essentially is at equilibrium. In addition, when equilibrium is upset, adjustment back to equilibrium is extremely rapid. The rationale is as follows.

Rates of Downcutting

All available evidence indicates that prior to the extrusions of lava into the Grand Canyon approximately a million years ago, the Colorado River had cut down to its present gradient. The size and shape of the canyon walls at the time the lava dams were formed were essentially the same as that which we see today. One of the most significant conclusions of our studies, with respect to processes of erosion and canyon cutting, is that after each lava dam was formed, the Colorado River eroded through the dam, down to its original profile *but no farther.* This process of reexcavating the canyon took place at least 13 times during the last million years or so. The cumulative thickness of basalt in these 13 lava dams was at least 11,300 feet (3390 m). Thus, the Colorado River actually has cut through a cumulative vertical thickness of nearly two miles (3 km) of rocks to remove the lava dams. In all probability, the actual downcutting through the basalts took place in less than a million years because there were large time intervals between periods of dam destruction when the Colorado River was not influenced by the presence of lava. The best estimates (based on the time necessary for the destruction of the lava dams) would be that actual downcutting through the 11,300 feet (3390 m) of basalt which formed the dams took place in approximately 200,000 years. During the intervening time, the Colorado River was flowing at its normal gradient with neither large-scale deposition nor erosion taking place.

The fundamental question of the age of the Grand Canyon (how long has it taken the Colorado River to cut through a sequence of Paleozoic strata one mile deep) is not only a question of how fast could the river erode, but of how fast the region was uplifted. Based on the rates at which the river has been able to cut through the sequence of lava dams (or the Paleozoic strata when its channel was displaced beyond the lava flows), it appears that the Colorado River has the capacity to cut down faster than tectonic processes can produce uplift.

This conclusion is supported by the fact that recurrent movement on the Toroweap fault has displaced the strata more than 150 feet in slightly more than one million years. There is no indication of rapids or waterfalls associated with the fault scarp, indicating that the escarpment produced by recurrent movement along the fault is erased by erosion immediately after it forms.

Rates of Slope Retreat

In addition to providing an insight into the capacity of a river to downcut and to maintain a profile of equilibrium, the remnants of the lava dams provide important documentation concerning rates of slope retreat and the relationship between slope profiles, stream gradient, and tectonic uplift.

Throughout the sections of the western Grand Canyon where remnants of lava dams still remain, one fact is completely clear: The remnants of lava dams are preserved only in the more protected parts of the canyon—that is, on the insides of meander bends, in protected alcoves, and as hanging valleys in the mouths of minor tributaries. Those remnants clinging to the canyon walls are preserved only as thin slivers, commonly only a few tens of feet wide. Except for the youngest flows, which may be 200 to 400 feet (60 to 120 m) wide in the broader sections of the canyon below Whitmore Wash, the remnants of the dams remain as thin sheets plastered against the canyon walls.

It is important to note that actual downcutting of the river by abrasion along the river channel would produce a vertical gorge only 200 to 250 feet (60 to 75 m) wide (the average width of the Colorado River channel). Slope retreat has been responsible for widening the rest of the inner gorge. At the present time, the inner gorge at the level of Temple Butte Limestone is roughly 2000 feet (600 m) wide. This means that approximately 900 feet (270 m) of slope retreat occurred on each side of the canyon after each lava dam was formed. At the base of the Muav Limestone, the canyon is approximately 1000 feet (300 m) wide; at that level, 400 feet (120 m) of slope retreat on each side of the river has occurred.

With the destruction of each lava dam and the rapid reestablishment of the river gradient to its original profile, there also was rapid and contemporaneous retreat of the slopes back to their original profile. The process occurred so fast that the original slope profile of the canyon was reestablished before the formation of the next lava dam.

The sequence of juxtaposed remnants near the southern end of the Esplanade sequence serves as an important example. Here, remnants of four lava dams are in juxtaposition (Ponderosa, Esplanade, and two younger dams). Remnants upstream and downstream from this site indicate that five additional units were once present in this area. After each lava dam was breached, there was an average distance of 700 feet (210 m) of slope retreat back to the original canyon wall. The cumulative distance of slope retreat that actually occurred in this area of the canyon was 4900 feet (1470 m), a rate of nearly one mile (1.6 km) per million years. If we consider all of the late Cenozoic lava dams in the canyon, the cumulative distance of slope retreat at the level of the Redwall Limestone would be 11,600 feet (3480 m). At the top of the Bright Angel, the distance would be 5500 feet (1650 m).

In each case, after a lava dam was eroded, the basalt retreated to within a few feet of the original canyon wall. Then, the processes of slope retreat essentially stopped. In many places, the processes of slope retreat completely removed the basaltic flows, *but the process of slope retreat did not enlarge the canyon and go back beyond the original canyon walls.*

Several important conclusions are obvious from these facts: (1) The energy system that causes slope retreat in the Grand Canyon has the capacity to erode canyon walls at a rate of one to two miles per million years, or more. (2) The slopes recede rapidly to a profile of quasi-equilibrium. They then recede at a very slow rate. (3) Slope retreat is balanced delicately with the stream gradient. Renewed downcutting of the stream channel is accompanied by renewed slope retreat. (4) The profile of the Grand Canyon apparently is in a state of quasi-equilibrium. (5) Periods of major rapid slope retreat are initiated by tectonic uplift.

We can conclude from these studies that erosion of the Grand Canyon did not take place at an imperceptibly slow and constant rate. Instead, it was governed by the style and degree of tectonic uplift. Inasmuch as movement on the major faults in the Colorado Plateau occurs in pulses, the actual processes of erosion also must occur in pulses; and, inasmuch as a tectonic disturbance produces only a few feet of displacement at a time, erosion back to equilibrium occurs in a series of small pulses separated by relatively long periods of quiescence.